Regional Climate Studies

The BACC II Author Team

Second Assessment of Climate Change for the Baltic Sea Basin

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Regional Climate Studies

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Second Assessment of Climate Change for the Baltic Sea Basin

Baltic Earth
Earth System Science for the Baltic Sea Region

Springer Open

The BACC II Author Team
c/o International Baltic Earth Secretariat
Helmholtz-Zentrum Geesthacht GmbH
Geesthacht
Germany

ISSN 1862-0248 ISSN 1865-505X (electronic)
Regional Climate Studies
ISBN 978-3-319-16005-4 ISBN 978-3-319-16006-1 (eBook)
DOI 10.1007/978-3-319-16006-1

Library of Congress Control Number: 2015933368

Springer Cham Heidelberg New York Dordrecht London

Printed on acid-free paper

Springer International Publishing AG Switzerland is part of Springer Science+Business Media
(www.springer.com)

Preface

This is the second assessment report addressing the state of knowledge concerning climate, climate change and climate impacts within the Baltic Sea region. It is not intended as a report about the 'best' knowledge, but as a description of what the scientific community at the time of the assessment accepts as being a representative description of climate change, the factors driving the change, and the interlinkage of phenomena, processes and sub-systems—while recognising the tentative nature of all scientific assertions, namely that future insights and observations may call for revisions of our present knowledge. The BACC-process was established to document the convergence, divergence and gaps in knowledge of scientific understanding with respect to climate change and climate impacts in the Baltic Sea region.

This assessment report is organised in the same way as the previous one, such that the report starts with a summary of the assessment as a whole before continuing with detailed chapters on each topic. Thus, it is possible for readers to gain a general overview of the entire results, turning to the detail of specific chapters as needs dictate.

The good news is that the conclusions of the first BACC assessment published in 2008 are mostly still valid. Although the scientific community has not revised its understanding of the climatic issues within the Baltic Sea region since 2008, many new and additional aspects have appeared in the scientific literature in recent years.

An organisational detail adds to the credibility of the continuity, namely that different people undertook the second assessment. In fact, all lead authors of the first BACC assessment were replaced by new members of the scientific community for BACC II. This was despite most lead authors of BACC I having done an excellent job and was to ensure the independence of the BACC II process from its predecessor. However, many of the BACC I lead authors did serve on the BACC II Scientific Steering Committee and helped to establish the chapter structure as well as to select lead authors for the report.

Some of the chapters were rather broad and it would have been very demanding if not impossible for an individual to draft the entire chapter. In such cases, the lead authors asked colleagues to team up in the writing. Nevertheless, the lead authors are responsible for the chapters as a whole.

Admittedly, the well-intended planning of the BACC-process for ensuring a representative description of the current breadth of knowledge without involving the author's own unpublished research or personal preferences was not completely successful. But the difficulties were few, and the overall value of the report was not compromised.

It is more than ten years since the BACC-process was initiated—originally as an independent effort in the scientific community and later embedded within the framework of the Baltic Sea region research programme BALTEX. Indeed, the logistics of the BACC I process were organised by the International BALTEX Secretariat, and after the re-launch of the programme as Baltic Earth, the organisation of the new BACC effort remained in the experienced hands of the Secretariat, with Dr. Marcus Reckermann as the key figure for making BACC II a reality.

Of course the process required some funding; for printing and for making the assessment freely accessible online, for linguistic editing, and for the working time of the International

Baltic Earth Secretariat. This funding was provided by the Helmholtz-Zentrum für Material- und Küstenforschung in Geesthacht, Germany.

We have been the chair people of both the BACC I and BACC II processes. This has been quite a challenge, but nevertheless a rewarding one, which we think helped to further consolidate the Baltic Sea region climate scientific community. In this time we have become ten years older and it is now time for younger colleagues to take over. We would like to see BACC III organised with new chair people from within the Baltic Earth programme.

Hans von Storch
Anders Omstedt

Acknowledgments

First and foremost, we are grateful for the expertise and many hours of hard work which the lead and contributing authors have put into this project. Their phenomenal efforts and excellent teamwork have combined to create a comprehensive assessment of climate change and its impacts in the Baltic Sea region that will be useful for many years to come.

Furthermore, we would like to thank the reviewers, whose work has been crucial in ensuring the high scientific standard of this assessment report. A total of 31 anonymous reviewers from the Baltic Sea countries, Canada, France, the Netherlands, Norway, Spain, Switzerland, the UK and the USA accepted the onerous task of reviewing the manuscripts and suggesting improvements. In addition, several anonymous scientists helped in shaping the manuscripts and tracing inconsistencies between chapters, we thank them for this.

The review process was initiated by three independent review editors, who helped in identifying and contacting relevant reviewers. Special thanks go to Jouni Räisänen (University of Helsinki, Finland), Jens Hesselbjerg Christensen (Danish Meteorological Institute, Copenhagen, Denmark) and Wolfgang Fennel (Baltic Sea Research Institute, Warnemünde, Germany).

The work of the BACC II Science Steering Group is greatly appreciated, particularly in terms of the advice given to the group of authors and the editors, for providing material, for reviewing early drafts of the manuscript and for suggesting reviewers. The BACC II Science Steering Group members are Mikko Alestalo (Finnish Meteorological Institute, Helsinki, Finland), Phil Graham (Swedish Meteorological and Hydrological Institute, Norrköping, Sweden), Hans-Jörg Isemer (Helmholtz-Zentrum Geesthacht, Germany), Sirje Keevallik (Tallinn University of Technology, Estonia), Maris Klavins (University of Latvia, Riga, Latvia), Maria Laamanen (HELCOM Secretariat, Helsinki, Finland), Juha-Markku Leppänen (Finnish Environment Institute, Helsinki, Finland), Anders Omstedt (University of Gothenburg, Sweden), Rajmund Przybylak (Nicolaus Copernicus University Toruń, Poland), Marcus Reckermann (Helmholtz-Zentrum Geesthacht, Germany), Benjamin Smith (Lund University, Sweden), Hans von Storch (Chair; Helmholtz-Zentrum Geesthacht, Germany), Heikki Tuomenvirta (Finnish Meteorological Institute, Helsinki, Finland), Timo Vihma (Finnish Meteorological Institute, Helsinki, Finland), Valery Vuglinsky (State Hydrological Institute, St Petersburg, Russia) and Ilppo Vuorinen (University of Turku, Finland).

As was the case for the first BACC assessment, the BACC II material has been used by the Baltic Marine Environment Protection Commission (HELCOM) for its Thematic Assessments. This fruitful collaboration with HELCOM was coordinated by Maria Laamanen (HELCOM Secretariat), while Janet Pawlak (MEC Consulting ApS, Charlottenlund, Denmark) summarised the BACC II material for HELCOM and we greatly appreciate her work.

This tremendous effort would not have been possible without the technical support of some individuals who deserve particular mention. From the International Baltic Earth Secretariat (Helmholtz-Zentrum Geesthacht, Germany) Silke Köppen helped with technical editing and obtaining permissions and Beate Gardeike improved the figures whenever this was considered necessary. Finally, Carolyn Symon (Environmental Editing Ltd, UK) is thanked for a dedicated final language check of the text.

The glossary was compiled with the help of anonymous scientists, and partly using the glossaries of the American Meteorological Society, IPCC and Wikipedia.

Permission

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Abbreviations and Acronyms

AAR	Accumulation Area Ratio
AD	Anno Domini (years after Christ)
ADCP	Acoustic Doppler Current Profiler
AEROCOM	Open international initiative of scientists interested in the advancement of the understanding of the global aerosol and its impact on climate
AIS	Automatic Identification System (for shipping)
Al	Aluminium
ALCC	Anthropogenic land-cover change
AMIP	Atmospheric Model Intercomparison Project
AMOC	North Atlantic Meridional Overturning Circulation
AMS	Aerosol Mass Spectrometer
amsl	Above mean sea level
ANC	Acid neutralising capacity
AO	Arctic Oscillation
AOGCM	Atmosphere-Ocean general circulation (or climate) model
AR4	Fourth Assessment Report of the IPCC
ARPEGE	Global circulation model by CNRM, France
ASL	Absolute sea level
ASTRA	Developing Policies and Adaptation Strategies to Climate Change in the Baltic Sea Region
A_T	Alkalinity
AV	Added Value
AVHRR	Advanced Very High Resolution Radiometer
BACC	BALTEX Assessment of climate change for the Baltic Sea region
BALTEX	The Baltic Sea Experiment
Baltic-C	Building predictive capability regarding the Baltic Sea organic/inorganic carbon and oxygen systems (a BONUS+ programme)
BALTIMOS	Coupled regional climate model system for the Baltic Sea region
BASYS	Baltic Sea System Study
BED	Baltic Environmental Database
BIFROST	Baseline Inferences for Fennoscandian Rebound Observations, Sea level and Tectonics
BODC	British Oceanographic Data Centre
BONUS	Joint Baltic Sea Research and Development Programme for the Baltic Sea region by EU and members states
BOOS	Baltic Operational Oceanographic System
BP	Before Present
BSAP	Baltic Sea Action Plan by HELCOM
BSi	Biogenic silicon, bound in diatom shells
BSS	Brier Skill Score
C	Carbon

C4MIP	Coupled Climate Carbon Cycle Model Intercomparison Project
Ca	Calcium
CAFE	EU Clean Air for Europe
cal. yr BP	Calibrated or Calendar years before present
CAMx	Open-source modeling system for multi-scale integrated assessment of gaseous and particulate air pollution
CCA	Canonical correlation analysis
CEI	Climate Extreme Index
CH_4	Methane
CLCC	Climate-induced land-cover change
CLM	Community Land Model
CLRTAP	Convention on Long-range Transboundary Air Pollution
CMIP3, 5	Coupled Model Intercomparison Project, phase 3 or 5 (of the WCRP)
CMT	Chemical transport model
CNRM	Centre National de Recherches Météorologiques, France
CO	Carbon monoxide
CO_2	Carbon dioxide
CO_3^{2-}	Carbonate
CORDEX	Coordinated regional climate downscaling experiment
COSAM	Intercomparison of largescale sulfur models
C_T	Total inorganic carbon
CTM	Chemical transport model
CW	Contemporary Warm Period
DAS	Data Assimilation System
DBS	Distribution-based scaling
DC	Delta change
DCF	Delta change factors
DEHM	Danish Air Pollution Model
DIAMIX	DIApycnal MIXing project
DIC	Dissolved inorganic carbon
DIN	Dissolved inorganic nitrogen
DJF	December January February (denoting the winter months)
DJFM	December January February March (denoting the winter months)
DMI	Danish Meteorological Institute
DOC	Dissolved organic carbon
DOM	Dissolved organic matter
DON	Dissolved organic nitrogen
DOP	Dissolved organic phosphorus
DSi	Dissolved silicon
DSL	Dynamic sea level
ECHAM4, 5	General Circulation Model developed by the Max Planck Institute for Meteorology in Hamburg, based on ECMWF models
ECHO-G	Coupled climate model consisting of the atmospheric model ECHAM4 and the ocean model HOPE (developed by the Max Planck Institute for Meteorology in Hamburg)
ECMWF	European Centre for Medium-Range Weather Forecasts
ECOSUPPORT	Advanced modeling tool for scenarios of the Baltic Sea ECOsystem to SUPPORT decision making (a BONUS+ programme)
EDGAR	Emission Database for Global Atmospheric Research of the Netherlands Environmental Assessment Agency, The Netherlands and the EU Joint Research Centre (JRC)y
eDIP	Excess inorganic dissolved phosphorus

EEA	European Environmental Agency
EMEP	European Monitoring and Evaluation Programme
EMHI	Estonian Meteorological and Hydrological Institute
EMULATE	European and North Atlantic daily to MULtidecadal climATE variability
ENSEMBLES	EC Sixth Framework Programme Integrated Project (2004–2009) to develop an ensemble prediction system for climate change
ENSO	El Niño/Southern Oscillation
E-OBS	European Climate Assessment and Dataset information
EOF	Empirical Orthogonal Functions
ERA-40	Reanalysis of the global atmosphere and surface conditions for 45-years, over the period from September 1957 through August 2002 by ECMWF
ERA-Interim	Global Atmospheric Reanalysis from 1979, continuously updated in real time
EU	European Union
EUCAARI	European Integrated project on Aerosol, Cloud, Climate, and Air Quality Interactions (EU FP6 project)
EUREF-EPN	European Terrestrial Reference System Permanent Network
EURO-CLIMHIST	European Climate Historical database European Centre of Historical Climate
EUROSION	European initiative for sustainable coastal erosion management
FAO	Food and Agriculture Organization of the UN
FAR	Fraction of attributable risk
Fe	Iron
FIMR	Finnish Institute of Marine Research
FinnRef	Finnish National network of reference stations for GPS
FMI	Finnish Meteorological Institute
FTE	Full-time equivalents
GAINS	Greenhouse Gas and Air Pollution Interactions and Synergies model by IIASA
GCM	General circulation model
GEV	Generalised Extreme Value
GFDL-CM2.1, R30	Coupled Atmosphere-Ocean models of GFDL, NOAA, USA
GHG	Greenhouse gas
GIA	Glacial isostatic adjustment
GISS-ER	Climate ocean model by Goddard Institute for Space Studies of NASA, USA
GLIMS	Global Land Ice Measurements from Space
GLUE	Generalized Likelihood Uncertainty Estimation
GLUES	Global Land Use and technological Evolution Simulator
GMSL	Global mean sea level
GPD	Generalised Pareto Distribution
GPS	Global Positioning System
GRACE	Gravity Recovery And Climate Experiment (satellite)
GrIS	Greenland Ice Sheet
H^+	Proton
H_2CO_3	Carbonic acid
H_2S	Hydrogen sulphide
HadAM3H	Hadley Centre Regional Climate Model
HadCM2, 3	Hadley Centre Coupled General Circulation Model, Version 2 or 3

HadGEM2-ES	Coupled Earth System Model being used by the Met Office Hadley Centre for the CMIP5 centennial simulation
HBV	Hydrological model of SMHI
HCO_3	Hydrogen carbonate
HELCOM	Helsinki Commission; Baltic Marine Environment Protection Commission
Hg	Mercury
HiResAFF	HIgh RESolution Atmospheric Forcing Fields
HIRHAM5	Regional Climate Model by the DMI, Denmark
HIROMB-SMHI	High Resolution Operational Model for the Baltic, a version by the Swedish Meteorological and Hydrological Institute
HNO_3	Nitric acid
HPO_4^{2-}	Hydrogen phosphate
Hs	Significant wave height
HTC	Selyaninov Hydrothermic Coefficient
HTM	Holocene Thermal Maximum
HYDE	History Database of the Global Environment
HZG	Helmholtz-Zentrum Geesthacht
ICES	International Council for the Exploration of the Sea
ICP Forest	International Co-operative Programme on Assessment and Monitoring of Air Pollution Effects on Forests
ICTP	International Centre for Theoretical Physics, Italy
IIASA	International Institute for Applied Systems Analysis, Austria
IMO	International Maritime Organization
INTAS-SCCONE	International Association for the promotion of co-operation with scientists from the New Independent States of the former Soviet Union—Snow Cover Changes Over Northern Eurasia
IPCC	Intergovernmental Panel on Climate Change
IUFRO	International Union of Forest Research Organizations
JJA	June July August (denoting the summer months)
JONSWAP	Joint North Sea Wave Observation Project oceanic wave spectrum
K	Potassium
ka	Kilo years (1000 years)
KK10	Historical land-use and population datasets by Kaplan and Krumhardt
KNMI	Royal Netherlands Meteorological Institute
k_{sp}	Solubility product
LAI	Leaf area index
LANDCLIM	Land Cover-Climate Feedbacks Network
LBC	Lateral boundary condition
LIA	Little Ice Age
LOVE	Local Vegetation Estimates
LPJ-GUESS	Process-based dynamic vegetation-terrestrial ecosystem model of Lund University, Sweden
LRA	Landscape Reconstruction Algorithm
LSL	Local sea level
LUCID	Land-Use and Climate, Identification of Robust Impacts project
MAM	March April May (denoting the spring months)
MAST	EU Marine Science and Technology programme
MATCH	Ad-hoc group for modelling and assessment of historic contributions to climate change
MBI	Major Baltic Inflow
MCA	Medieval Climate Anomaly, see also MWP

MEAD	Marine Effects of Atmospheric Depostion model, EU FP5 project
MESAN	Mesoscale Analysis System
Mg	Magnesium
MIB	Maximum sea-ice extent in the Baltic Sea
MIB	Maximum ice extent in the Baltic Sea
MICORE	Morphological impacts and coastal risks induced by extreme storm events
MIROC3.2	Model for Interdisciplinary Research on Climate; coupled climate model by Center for Climate System Research, Japan
Mn	Manganese
MODIS	Moderate Resolution Imaging Spectroradiometer
MOS	Model output statistics (a statistical downscaling technique)
MPI	Max-Planck-Institute
MPI-OM	MPI Ocean Model
MSC-W	Chemical transport model Meteorological Synthesizing Centre—West
MSFD	EU Marine Strategy Framework Directive
MSL	Mean sea level
MUSTOK	Model based investigation of extreme water levels at the German Baltic Sea coast
MWP	Medieval Warm Period
N	Nitrogen
N_2	Molecular nitrogen
N_2O	Nitrous oxide
Na	Sodium
NAO	North Atlantic Oscillation
NAP	Non Arboreal Pollen (pollen from herbaceous plants)
NASA	National Aeronautics and Space Administration
NCEP/NCAR	National Centers for Environmental Prediction/National Center for Atmospheric Research
NE	North East
NECA	NO_x emission control area
NEC-II	EU directive on national emission ceilings
NEU	EU NitroEurope project
NH_3	Ammonia
NH_4^+	Ammonium
NMVOC	Non-methane volatile organic compounds
NO_2	Nitrogen dioxide
NO_3^-	Nitrate
NOAA	National Oceanic and Atmospheric Administration, USA
NO_X	Nitrogen oxides
NPP	Net primary productivity
NPV	Net present value
Nr	Reactive nitrogen
NSRP	Neyman-Scott Rectangular Pulse
nss	Non-sea-salt
nss-sulphate	Non-sea-salt sulphate
NW	North West
O	Oxygen
O_3	Ozone
OECD	Organisation for Economic Co-operation and Development
OPYC3	Ocean General Circulation Model by MPI Hamburg, Germany
OWS	Ocean Weather Ship

P	Phosphorus
PBO	Preboreal oscillation
PCM	Parallel Climate Model, USA
pCO_2	CO_2 partial pressure
PET	Potential evapotranspiration
PgC	Petagrams of carbon, i.e., 10^{15} g of carbon
PGV	Area potentially covered by vegetation on the globe
PM	Pierson-Moskowitz oceanic wave spectrum
PM	Particulate matter
PM_{10}	Particles with an aerodynamic diameter equal to or less than 10 μm
$PM_{2.5}$	Particles with an aerodynamic diameter equal to or less than 2.5 μm
POC	Particulate organic carbon
POD	Perturbation of observed data (a statistical downscaling technique)
POD	Phyto-toxic ozone dose
POP	Particulate organic phosphorus
PP	Perfect prognosis (a statistical downscaling technique)
PP	Probability Plots
ppb(v)	Parts per billion by volume
ppm	Parts per million
PROBE-Baltic	Process-based ocean modelling system of Earth System Sciences Centre, Gothenbirg University, Sweden
PRUDENCE	Project to provide a series of high-resolution climate change scenarios for 2071–2100 for Europe (ended 2004)
PSMSL	Permanent Service for Mean Sea Level
QQ	Quantile Plots
RACMO2	KNMI regional atmospheric climate model Phase 2
RADS	Radar Altimeter Database System
RCA3	Rossby Centre Regional Climate model of SMHI, Sweden
RCAO	Rossby Centre Atmosphere–Ocean–Sea Ice model
RCM	Regional climate model
RCO	Rossby Centre Ocean Model
RCP	Representative Concentration Pathway
RegCM3	Regional Climate Model by ICTP
REMO	Regional Climate Model by MPI Hamburg, Germany
REVEALS	Regional Estimates of Vegetation Abundance from Large Sites
RF	Radiative forcing
RLR	Revised local reference
RMSE	Root-mean square error
ROSHYDROMET	Russian Federal Service for Hydrometeorology and Environmental Monitoring
RSL	Relative sea level
RV	Reproductive Volume
S	Sulphur
SAPOS	Satellite Positioning Service of the German State Survey
SAR	Wide-swath synthetic aperture radar
SATREF	Norwegian National network of reference stations for GPS
SCD	Snow Cover Duration
SCE	Snow-cover extent
SECA	SO_x Emission Control Area
Si	Silicon

SINCOS	Sinking Coasts—Geosphere, Ecosphere and Anthroposphere of the Holocene Southern Baltic Sea (research unit of the DFG, German Research Foundation)
SLP	Sea-level pressure
SMB	Surface Mass Balance
SMHI	Swedish Meteorological and Hydrological Institute
SO_2	Sulphur dioxide
SO_2	Sulfide
SO_4^{2-}	Sulphate
SON	September October November (denoting the autumn months)
SO_X	Sulphur oxides
SPI	Standardized Precipitation Index
SRES	Special Report on Emissions Scenarios (of the IPCC)
SSC	Science Steering Committee
SSH	Sea surface height
SST	Sea surface temperature
STARDEX	Statistical and Regional dynamical Downscaling of Extremes for European regions, EC funded project 2002–2005
STEAM	Ship Traffic Emissions Assessment Model
SVD	Singular value decomposition
SWE	Snow Water Equivalent
SWEPOS	Swedish National network of reference stations for GPS
SYKE	Finnish Environment Institute
TIC	Total inorganic carbon
TN	Total nitrogen
TOC	Total organic carbon
TOF-MS	Time-of-Flight Mass Spectrometer
TON	Total organic nitrogen
TP	Transitional Period
TP	Total phosphorus
TUFLOW	Flood and Coastal Simulation Software
UAA	Utilised agricultural area
UCLM	Universidad de Castilla-La Mancha, Spain
UHSLC	University of Hawaii Sea Level Center
UNECE	United Nations Economic Commission for Europe
UNEP	United Nations Environmental Programme
UNFCCC	United Nations Framework Convention on Climate Change
UV	Ultraviolet
Vd	Deposition velocity
VOC	Volatile organic compounds
W	Watt
WAM	Ocean wave model by HZG
WASA	Group on “Waves and Storms in the North Atlantic”
WCRP	World Climate Research Programme
WFD	EU Water Framework Directive
WGMS	World Glacier Monitoring Service
WIBIX	Winter Baltic Climate Index
WSA	Wasser- und Schifffahrtsämter (Regional Waterways and Shipping Offices in Germany)
WSFS	Watershed Simulation and Forcasting System of SYKE, Finland
YOLL	Amount of life years lost
$\delta^{13}C$	Delta Carbon isotope 13

$\delta^{15}N$	Delta Nitrogen isotope 15
$\delta^{18}O$	Delta Oxygen isotope 18
Ω	Saturation state (for calcium carbonate)
ΔpCO_2	CO_2 partial pressure difference
µatm	micro atm (pressure unit)
^{10}Be	Beryllium-10
^{14}C	Carbon-14, radiocarbon
20CR	20th Century reanalysis (dataset)

Introduction and Summary

1

Hans von Storch, Anders Omstedt, Janet Pawlak, and Marcus Reckermann

Abstract
The Baltic Earth Assessment of Climate Change (BACC) in the Baltic Sea region is an effort to establish what scientifically legitimised knowledge about climate change and its impacts is available for the Baltic Sea catchment. Observed past and projected future changes in atmospheric, hydrological, and oceanographic conditions are assessed, as well as the observed and potential impacts on the natural and socio-economic environments. The BACC programme focuses purely on the science and does not draw conclusions about the political, economic, or management consequences of the scientific knowledge. This report (the BACC II assessment, the follow-up to the BACC I assessment in 2008) documents the consensus and dissensus on climate knowledge up to about 2012. More than 180 researchers contributed in various functions to this peer-reviewed assessment. The process was overseen by a scientific steering committee and coordinated by the International Baltic Earth Secretariat.

Keywords
Baltic Sea • Climate Change

1.1 Overview

1.1.1 Background

The Baltic Earth[1] Assessment of Climate Change (BACC) in the Baltic Sea region is an effort to establish what scientifically legitimised knowledge about climate change and its impacts is available for the Baltic Sea catchment. Observed past and projected future changes in atmospheric, hydrological, and oceanographic conditions are assessed, as well as the observed and potential impacts on the natural and socio-economic environments.

The assessment has been used by the intergovernmental Baltic Marine Environment Protection Commission (HELCOM) as a basis for its deliberations. A division of labour has been established between BACC and HELCOM: the BACC programme focuses purely on the science and does not draw conclusions about the political, economic or management consequences of the scientific knowledge, while HELCOM accepts the BACC findings and uses them in its work within the political negotiation processes.

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[1] Baltic Earth is the successor to BALTEX, the scientific network dealing with meteorological, oceanographic, and hydrographic issues in the Baltic Sea basin. BALTEX was concluded in June 2013, and Baltic Earth has been established with a broader agenda of bringing together scientists interested in the dynamics, the past and possible futures of the Baltic Sea basin (www.baltic-earth.eu).

The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_1

HELCOM used the outcome of the present BACC II assessment in its Thematic Report on Climate Change published in 2013 (HELCOM 2013), once BACC had reached final agreement on its findings, but before various editorial tasks were complete.

The first BACC assessment report was published in 2008 (BACC Author Team 2008) and was accompanied by a Thematic Report by HELCOM published the previous year (HELCOM 2007). For the first BACC book, approximately 80 scientists from 12 countries researched and assessed the available literature on climate change in the Baltic Sea basin.

BACC II follows the BACC I format in documenting the consensus and dissensus on climate knowledge up to about 2012. New aspects in BACC II are a consideration of the long-term and recent historical past to provide a framework for evaluating the severity and unusualness of the current climate change, a detailed analysis of sea-level changes in the Baltic Sea, a description of climate change effects on urban complexes, and a reflection on detection and attribution of regional climate change in the Baltic Sea region. For the present assessment, more than 180 researchers contributed in various functions, from the lead and contributing authors and reviewers to the overseeing of the process by the Scientific Steering Committee and coordination by the International Baltic Earth Secretariat.

1.1.2 Overall Summary

The key findings of the BACC I assessment were as follows:

- The Baltic Sea region is warming, and the warming is almost certain to continue throughout the twenty-first century.
- It is plausible that the warming is at least partly related to anthropogenic factors.
- So far, and as is likely to be the case for the next few decades, the signal is limited to temperature and to directly related variables, such as ice conditions.
- Changes in the hydrological cycle are expected to become obvious in the coming decades.
- The regional warming is almost certain to have a variety of effects on terrestrial and marine ecosystems—some will be more predictable (such as the changes in phenology) than others.

The key findings of the BACC II assessment—the present assessment—are as follows:

- The results of the BACC I assessment remain valid.
- Significant additional material has been found and assessed. Some previously contested issues have been resolved (such as trends in sea-surface temperature).
- The use of multi-model ensembles seems to be a major improvement; there are first detection studies, but attribution is still weak.
- Regional climate models still suffer from biases related to the heat and water balances. The effect of changing atmospheric aerosol load to date cannot be described; first efforts at describing the effect of land-use change have now been done.
- Data homogeneity is still a problem and is sometimes not taken seriously enough.
- The issue of multiple drivers on ecosystems and socio-economics is recognised, but more efforts to deal with them are needed.
- In many cases, the relative importance of different drivers of change, not only climate change, needs to be evaluated (e.g. atmospheric and aquatic pollution and eutrophication, overfishing, and changes in land cover).
- Estimates of future concentrations and deposition of substances such as sulphur and nitrogen oxides, ammonia/ammonium, ozone, and carbon dioxide depend on future emissions and climate conditions. Atmospheric warming seems relatively less important than changes in emissions. The specification of future emissions is plausibly the biggest source of uncertainty when attempting to project future deposition or ocean acidification.
- In the narrow coastal zone, the combination of climate change and land uplift acting together creates a particularly challenging situation for plant and animal communities in terms of adaptation to changing environmental conditions.
- Climate change is a compounding factor for major drivers of changes in freshwater biogeochemistry, but evidence is still often based on small-scale studies in time and space. The effect of climate change cannot yet be quantified on a basin-wide scale.
- Climate model scenarios show a tendency towards future reduced salinity, but due to the large bias in the water balance projections, it is still uncertain whether the Baltic Sea will become less or more saline.
- Scenario simulations suggest that the Baltic Sea water may become more acidic in the future. Increased oxygen deficiency, increased temperature, changed salinity, and increased ocean acidification are expected to affect the marine ecosystem in various ways and may erode the resilience of the ecosystem.
- When addressing climate change impacts on, for example, forestry, agriculture, urban complexes, and the marine environment in the Baltic Sea basin, a broad perspective is needed which considers not only climate change but also other significant factors such as changes in emissions, demographic and economic changes, and changes in land use.
- Palaeoecological ‘proxy’ data indicate that the major change in anthropogenic land cover in the Baltic Sea catchment area occurred more than two thousand years ago. Climate model studies indicate that past

anthropogenic land-cover change had a significant impact on past climate in the northern hemisphere and the Baltic Sea region, but there is no evidence that land-cover change since AD 1850 was even partly responsible for driving the recent climate warming.

1.1.3 The BACC Process

The overall format of the BACC process is similar to the process for assessments undertaken by the Intergovernmental Panel on Climate Change (IPCC), with author groups for the individual chapters, an overall summary for policymakers (here: the executive summary in Sect. 1.2) and a peer-reviewed process. The assessment is a synthesis of material drawn entirely from the available scientifically legitimate literature (e.g. peer-reviewed literature, conference proceedings, and reports of scientific institutes). Influence or funding from groups with a political, economic, or ideological agenda is not allowed; however, questions from such groups are welcome. If a consensus view cannot be found in the above-defined literature, this is clearly stated and the differing views are documented. The assessment thus encompasses the knowledge about what scientists agree on but also identifies cases of disagreement or knowledge gaps. The assessment is evaluated by independent scientific reviewers.

The BACC process was overseen by a Science Steering Committee (SSC) and managed by the International Baltic Earth Secretariat. The role of the BACC II SSC was to formulate a plan for the BACC II assessment and the outline topics to be addressed. The SSC was responsible for selecting Lead Authors for each specific topic. It was also responsible for overseeing the review process.

The International Baltic Earth Secretariat was the focal point regarding the logistics of the review process, meetings, and technical writing. Of particular importance was the full availability of information, including text material for all participants in the BACC II writing process and representatives of HELCOM. A dedicated BACC II website is available at www.baltic-earth.eu/BACC2/, including a password-protected page that was used for the exchange of internal information and documents (e.g. draft chapters).

Lead Authors were invited by the BACC II SSC to prepare chapters on the basis of currently available scientific knowledge. The Lead Authors are responsible for the overall quality and content of their chapter. This applies even if they formed a writing team to help draft the chapter. The essence of the Lead Authors task was the synthesis of material drawn entirely from the available scientifically legitimate literature (e.g. peer-reviewed literature, conference proceedings, and reports of scientific institutes). It took about four years to complete the assessment, which concludes with the publication of this book. The process started in April 2010, when the BACC II SSC was formed and the BACC II Lead Authors were nominated at a meeting in Lund, Sweden. In June 2010, the BACC II programme was presented and discussed by the scientific community at the 6th BALTEX Study Conference in Międzyzdroje, Poland. The first meeting of Lead Authors took place in Gothenburg, Sweden in November 2010; the second meeting in Hamburg, Germany in March 2011. In 2011, the assessment work was begun by the Lead and Contributing Authors, and in February 2012, the third meeting of Lead Authors took place in Copenhagen, Denmark. In May 2012, BACC II was presented at the BONUS Workshop at the European Maritime Day in Gothenburg and the first drafts of the BACC II chapters became available in mid-2012. The chapters were reviewed at a dedicated BACC II Review/Stakeholder Conference in Tallinn, Estonia, in September 2012. The external peer review was completed by the end of 2012. By March 2013, the chapters had been revised according to the reviewers' comments. The final text was published as a print and open access book in early 2015.

1.1.4 Important Terminology

The BACC II assessment uses the following terminology as defined here.

- *Climate change*: Deterministic response to changes in external climate forcing.
- *Internal variability*: Random fluctuations independent of external forcing.
- *Detection*: The observed climate change is unlikely due to internal variability alone = Detection of (some) external forcing.
- *Attribution*: Mix of plausible external forcing mechanisms that best explain the detected change.

1.1.5 Annexes

This report features two annexes. Annex 1 provides a short mathematical definition of the scientific concept of 'Detection and Attribution'. Annex 2 deals with the impact of the BACC I report within the regional scientific community and among decision-makers with responsibility for the German Baltic Sea coast. In short, it is a peer assessment of the BACC I report and a policy decision-makers' assessment of the state of the region, their sources of scientific information, and the utility of the information contained in the report. While the scientific peer review indicated overall satisfaction, it was clear that the report would require significant changes to be of maximum use for decision-makers. However, this is not surprising since the intended audience for the book was primarily the scientific community.

1.2 Executive Summary

In addition to this introductory chapter, the report comprises six major sections.

- Part I: Long-term climate change (Chaps. 2 and 3)
- Part II: Recent climate change (past 200 years) (Chaps. 4–9)
- Part III: Future climate change (Chaps. 10–14)
- Part IV: Environmental impacts of climate change (Chaps. 15–20)
- Part V: Socio-economic impacts of climate change (Chaps. 21 and 22)
- Part VI: Drivers of regional climate change (Chaps. 23–25)

1.2.1 Long-term Climate Change: From the Holocene to the Little Ice Age

The first two chapters of this report provide a broad overview of the geo-historical development of conditions in the past millennia, based on proxy data from sources such as fossil pollen and insects from dated continental and lake sediments, tree ring widths and density, and some written records or diaries.

The Baltic Sea is a young sea. It was formed after the glaciers that covered this area for over 100,000 years began to recede about 18,000 years ago, mainly as a result of changes in the orbital configuration of the Earth's rotation around the sun. In the Baltic Sea area, this last Ice Age ended about 11,000 years ago. During the following warmer Holocene, the Baltic Sea basin underwent a number of major changes owing to the gradual melting of the Fennoscandian ice sheet, which caused a rise in sea level and the slow isostatic uplift of the land readjusting to the disappearance of the heavy ice sheet, thus decreasing relative sea level. The interplay between these two processes led to a series of transitions in the Baltic Sea basin, varying between a freshwater lake during periods of isolation from the North Sea and a brackish sea when water mass exchange with the saline North Sea occurred.

The melting of the ice sheets in North America had an influence on oceanic circulation, and it is believed that large volumes of cold water from the melting ice sheets entering the North Atlantic Ocean exerted an abrupt cooling around 8200 years ago that lasted several centuries in the Baltic Sea area. Changes in the orbital configuration of the Earth also modulated the incoming solar insolation of the boreal high latitudes and thus strongly influenced the regional energy balance. The summer solar insolation peaked at around 7000–6000 years ago and subsequently decreased. After the melting of the remnants of the Fennoscandian ice sheets, a relatively stable period occurred around 7500–5500 years ago, with summer temperatures 1.0–3.5 °C higher than at present. Since then, a continued decreasing trend in summer solar insolation due to astronomical factors has resulted in a more unstable climate and a progressive millennial cooling.

Other external and internal factors influencing the climate of the Baltic Sea basin during the Holocene include variations in the concentrations of stratospheric aerosols induced by volcanic activity, changes in the concentrations of greenhouse gases (GHGs) in the atmosphere due to natural factors, changes in the surface albedo of the sea and changes in the vegetation of the surrounding land, and salinity-induced changes in the intensity and type of circulation in the sea.

In the Baltic Sea basin and surrounding areas of Europe, relatively stable climate conditions prevailed in the tenth and eleventh centuries, typified by warm, dry summers. Investigations in Fennoscandia indicate that a so-called Medieval Warm Period (MWP) occurred between AD 900 and 1350; at that time, warm-season (May–September) temperatures exceeded the contemporary warming at the end of the twentieth century by about +0.5 °C. An exceptionally warm period occurred from AD 1200–1250.

The climate of the past 500 years has been characterised by centennial-scale variability and the modulation of inter-annual and decadal signals, often accompanied by rapid shifts. In the latter half of the sixteenth century, the temperature dropped, initiating the Little Ice Age. In the recently modelled and reconstructed northern Scandinavian summer temperatures since 1500, the longest cool period prevailed during the first half of the seventeenth century and at the beginning of the eighteenth century, as well as during the first years of the nineteenth century. During these main historical climatic periods, climatic conditions were not uniform and shorter warm/cool and wet/dry fluctuations were observed depending on regional factors.

1.2.2 Recent Climate Change: The Past 200 Years

The six chapters of this section describe the observed climatic changes in the atmosphere, on land (river run-off, cryosphere) and at sea (circulation and stratification, sea ice, sea level and wind waves) for properties in the Baltic Sea drainage basin over the past 200–300 years. The Baltic Sea has a dense observation network covering an extended time period, and measurements of environmental conditions have been made with increasing accuracy over this period. Continuous time series exist since the middle of the eighteenth century for a few stations, and a denser network of stations has been operational since the mid-nineteenth century.

1.2.2.1 Recent Changes in the Atmosphere

The Baltic Sea region features very variable weather conditions due to its location in the extra-tropics of the northern hemisphere. The region is controlled by two large-scale pressure systems over the north-eastern Atlantic Ocean—the Icelandic Low and the Azores High. In general, westerly winds predominate, although any other wind direction is also frequently observed. As the climate of the Baltic Sea basin is to a large extent controlled by the prevailing air masses, atmospheric conditions will therefore be controlled by global climate as well as by regional circulation patterns, and the atmospheric parameters are strongly interlinked.

Large-Scale Atmospheric Circulation

The large-scale circulation patterns in the Baltic Sea region are influenced by the North Atlantic Oscillation (NAO), which is the dominant mode of near-surface pressure variability over the North Atlantic Ocean and neighbouring land masses, particularly during winter. In its positive phase, the Icelandic Low and Azores High are strengthened, resulting in a stronger than normal westerly airflow on a more northerly tract over the eastern North Atlantic and Europe; this brings warm, wet winters to all of Europe except for the southern part. When both the Icelandic Low and the Azores High are weak, termed a negative NAO, the westerly airflows are also weak, resulting in colder, drier winters in northern Europe. The NAO shows considerable seasonal and interannual variability, with prolonged periods of positive or negative phases.

Although the NAO shows long-term variability, from the mid-1960s to the mid-1990s it was in a generally positive phase, with stronger westerly winds, mild and wet winters, and increased storminess in the Baltic Sea area. After the mid-1990s, there was a trend towards more negative NAO values, resulting in weak westerly airflows and weather types that appear to be more persistent than in earlier decades.

Wind

The wind climate is generally related to large-scale variations in the atmospheric circulation of the North Atlantic, including the NAO in winter. Overall, reconstructions for the past 200 years show that storminess in northern Europe is dominated by large multi-decadal variations rather than by long-term trends. However, during the latter half of the twentieth century, large changes were observed in the wind climate over the North-east Atlantic and northern Europe. An unusually calm period occurred during the early 1970s; this coincided with a period with strongly negative NAO values as well as a very high frequency of Euro-Atlantic atmospheric blocking in winter, preventing or weakening westerly flow and leading to low wind speeds and fewer storms over Scandinavia. This was followed by a strong increase in annual and winter-to-spring storminess with unprecedented high winter storminess in the 1980s and early 1990s, during which the NAO index reached very high values. In the first decade of the twenty-first century, wind speeds returned to average values in the Baltic Sea area.

A northward shift in storm tracks and increased cyclonic activity have been observed in recent decades with an increased persistence of weather types. No long-term trend has been observed in annual wind statistics since the nineteenth century, but considerable variations on (multi-) decadal timescales. An anthropogenic influence cannot be excluded since the middle of the twentieth century. The pattern in wind and wave heights over the northern hemisphere with a north-easterly shift of storm tracks appears to be consistent with combined natural and external forcing.

Temperature

There has been a clear increase in surface air temperature in the Baltic Sea basin since the beginning of the observational record in 1871. This has occurred with large multi-decadal variations dividing the twentieth century into three phases: warming until the 1930s, followed by cooling until the 1960s, and then another distinct warming during the final decades of the century. Linear trends in the annual mean temperature anomalies from 1871 to 2011 were 0.11 °C per decade north of 60°N and 0.08 °C per decade south of 60°N in the Baltic Sea basin. All seasonal trends are positive and statistically significant except for winter temperature north of 60°N. These trends are shown in Fig. 1.1.

An analysis of temperature trends from 1970 to 2008 in the Baltic Sea area showed the strongest increase in the Gulf of Bothnia in autumn and winter (+0.5 to +0.6 °C per decade), while significant changes occurred during spring and summer in the central and southern parts of the Baltic Sea area (+0.2 to +0.3 °C per decade). During the past decade, the warming has continued during spring and summer in the southern parts and during autumn and spring in the northern parts, but the winters of 2009/10 and 2010/11 were very cold.

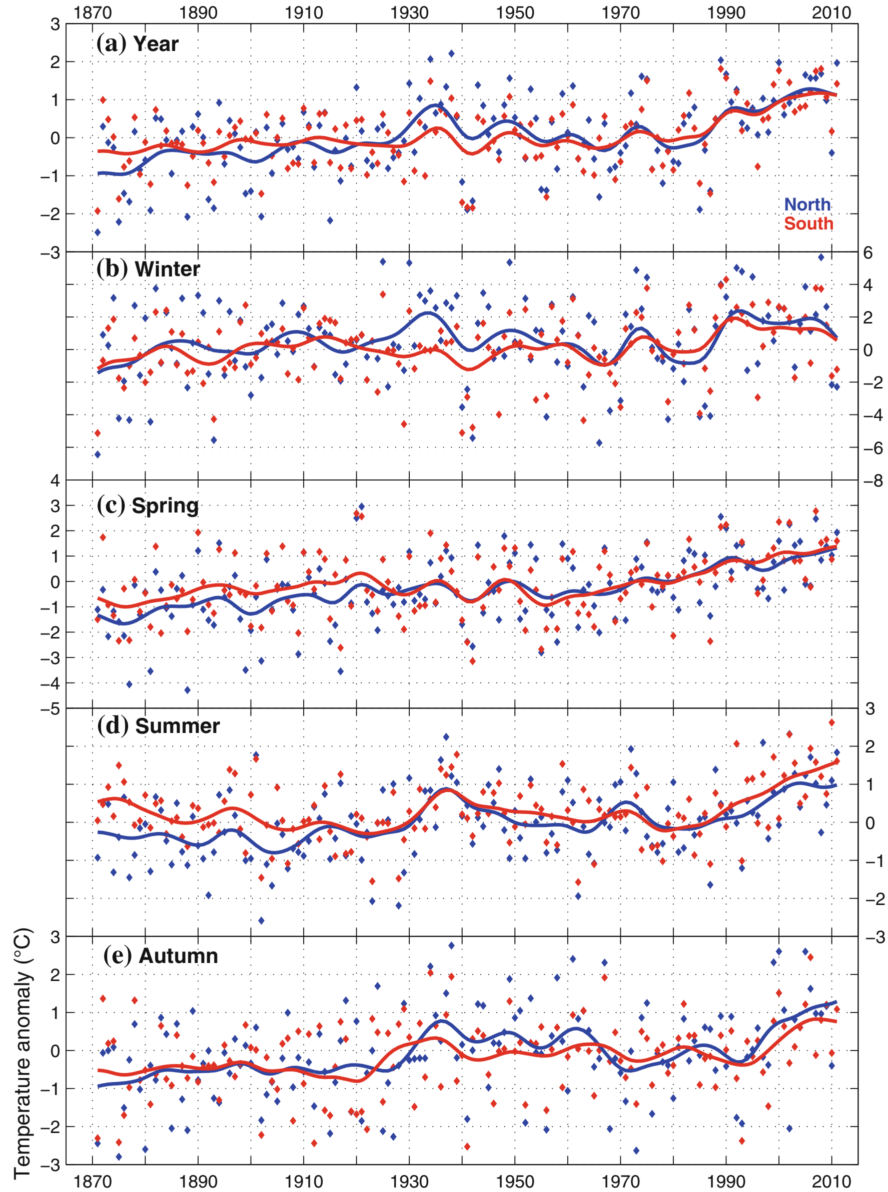
Seasonal changes are also evident: the duration of the growing season has increased, and the duration of the cold season has decreased. For instance, in Tartu, Estonia, the number of deep winter days (with snow cover) decreased by 29 over the twentieth century, while the growing season increased by 13 days. For the Baltic States and Poland, there has been a significant increase in the number of hot days and nights and the number of heat waves, as well as a significant decrease in the number of frost days.

Precipitation

No long-term trend was observed for precipitation, but there is some indication of an increased duration of precipitation periods and possibly an increased risk of extreme precipitation events.

The amount of precipitation in the Baltic Sea area during the past century has varied between regions and seasons,

Fig. 1.1 Annual and seasonal mean surface air temperature anomalies (relative to 1960–1991) for the Baltic Sea basin 1871–2011, calculated from 5° by 5° latitude, longitude box average taken from the CRUTEM3v data set (Brohan et al. 2006) based on land stations (from *top* to *bottom* **a** annual, **b** winter (DJF), **c** spring (MAM), **d** summer (JJA), **e** autumn (SON)). *Blue* comprises the Baltic Sea basin north of 60°N, and *red* south of 60°N. The *dots* represent individual years, and the *smoothed curves* (Gaussian filter, $\sigma = 3$) highlight variability on timescales longer than 10 years



with both increasing and decreasing precipitation apparent. A tendency of increasing precipitation in winter and spring was detected during the latter half of the twentieth century. Comparing the annual mean precipitation during 1994–2008 with that of 1979–1993 showed less precipitation in the northern and central Baltic Sea region and more in the southern region. The pattern also varied by season (Fig. 1.2).

Mostly, positive trends in daily precipitation and precipitation totals were detected at 116 stations across Europe, including the Baltic Sea region, during 1950–2000. Although most of the trends are not statistically significant, some positive trends were identified for winter and spring, while negative trends mostly occurred in summer and, in some cases, in autumn.

Cloud Cover and Sunshine Duration

Generally, negative trends in cloudiness and coincident positive trends in sunshine duration and solar radiation in the major parts of the Baltic Sea basin have been detected over recent decades. However, the data are highly variable, and the number of investigations to date is not sufficient to enable more detailed and reliable conclusions.

1.2.2.2 Recent Changes in Hydrology and the Terrestrial Cryosphere

The Baltic Sea can be considered a large, semi-enclosed brackish water estuary draining into the North Sea via the Danish Straits. The inflow from rivers to the Baltic Sea is an

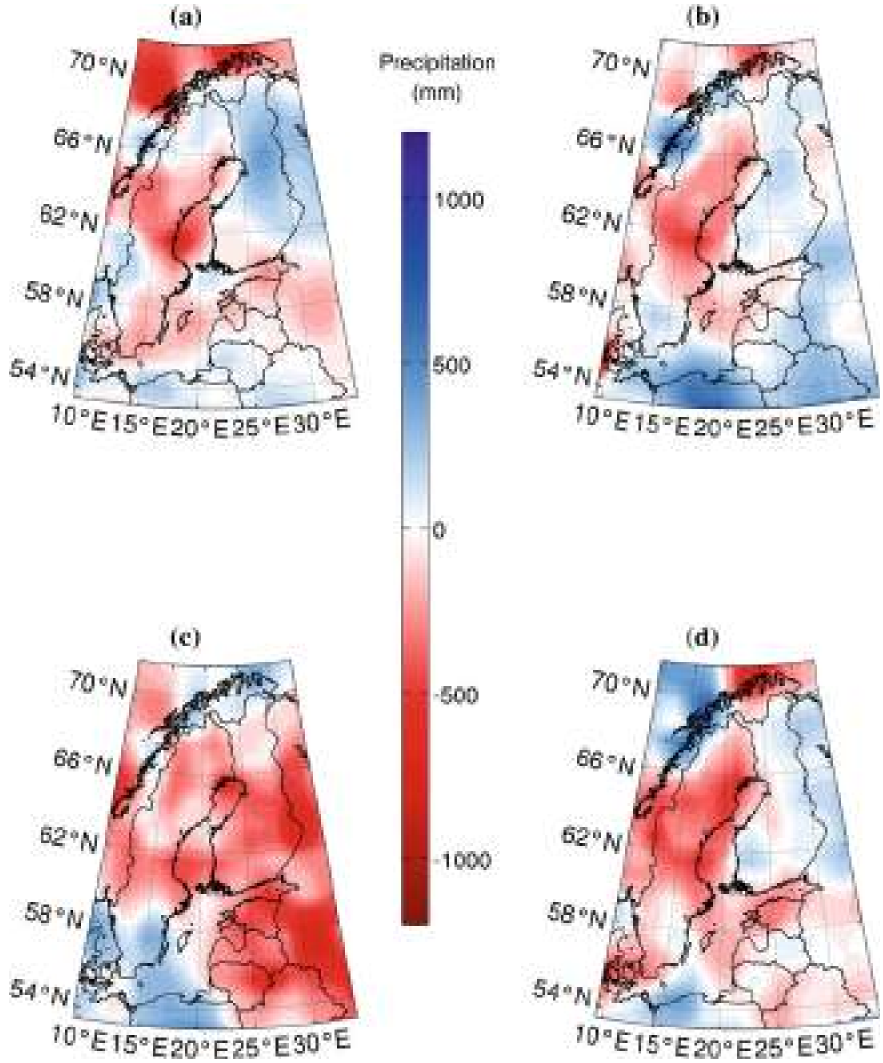


Fig. 1.2 Change in total precipitation between 1994–2008 and 1979–1993 by season based on SMHI data (Lehmann et al. 2011). **a** DJF, **b** MAM, **c** JJA, **d** SON

important variable for both the physical, chemical, and ecological processes of the sea. In winter, much of the precipitation is stored as snow. Thus, in the north, discharges are lowest towards the end of winter before snowmelt. The highest discharges are recorded in spring or early summer owing to snowmelt. Water levels and discharges usually decrease during summer when evaporation and evapotranspiration are greatest and normally greater than precipitation. In some regions, river discharge is strongly influenced by hydropower, damping the seasonal cycle.

River Run-Off
Although decadal and regional variability is large, no statistically significant long-term change has been detected in total river run-off to the Baltic Sea during the past 500 years. An analysis of streamflow in a large number of rivers and streams in the Nordic countries over three periods from 1920 to 2002 generally showed that trends towards increased streamflow dominated annual values as well as the winter and spring seasons, while no trend was found for autumn. An indication of earlier snowmelt floods was also apparent. Although these trends in streamflow result from changes in both temperature and precipitation, temperature was shown to have a stronger effect. As an example, Fig. 1.3 shows annual anomalies and long-term variations in precipitation, temperature, water resources, and flood magnitude in Sweden for the period 1901–2010 in relation to the reference period 1961–1990. A decrease in annual discharge from southern catchments of the Baltic Sea of about 10 % has been observed from the rivers Nemunas (Lithuania), Neva (Russia), Vistula, and Oder (Poland) over the past century.

River Ice
The river ice regime is considered to be a sensitive indicator of climate change. A study of ice in the River Daugava (Latvia), which has a data series starting in 1530, showed a pronounced decreasing trend for the past 150 years with an even clearer trend for the past 30 years. This indicates a

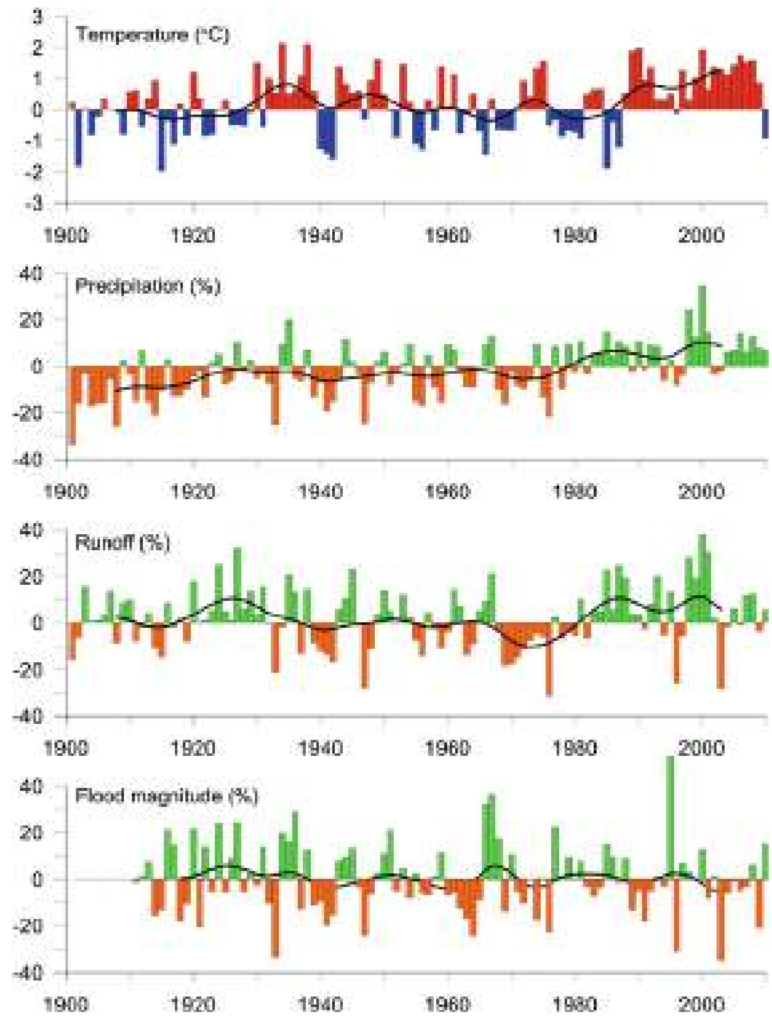


Fig. 1.3 Annual anomalies (relative to 1961–1990) and long-term variability in precipitation, temperature, water resources, and flood magnitude in Sweden for 1901–2010. For flood magnitude, the years before 1911 were omitted due to data scarcity (Hellström and Lindström 2008)

reduction of ice-cover duration and a shift to earlier ice break-up. The ice-cover duration has declined by 2.8–6.3 days per decade during the past 30 years. Although regional variations exist, similar observations have been made for other rivers flowing into the Baltic Sea. Both the ice regime and the seasonal river discharge are strongly influenced by large-scale atmospheric circulation processes over the North Atlantic, and this is shown by a close correlation with the NAO.

Snow

Snow is the origin of a significant fraction of run-off in the Baltic Sea basin. Although there is large interannual variation in the extent of snow cover in the Baltic Sea drainage basin, a decreasing trend in snow-cover extent since the 1970s has been observed in Fennoscandia, with some regional exceptions. The duration of annual snow cover in western Scandinavia and in the south-western East European plain has also became shorter over the past century. In northern and eastern parts of the drainage basin, and in mountain regions where both precipitation and temperature control snow amount, colder average winters have led to an increase in annual snow depth. There was a tendency towards weaker spring floods in Latvia from 1951 to 2006 and stronger winter flows due to changes in snow-cover duration and snow amount.

Frozen Ground

The current understanding of the spatial patterns of freeze/thaw cycles in the Baltic Sea region remains poor and has not been subject to systematic study. Some warming trends and some decreases in the duration and depth of seasonally frozen ground have been seen. Warming trends have been observed in the European permafrost, as well as a northward shift in the southern boundary of near-surface permafrost in European Russia.

1.2.2.3 Recent Changes in Baltic Sea Hydrography

The Baltic Sea is well stratified, with a seasonal cycle of temperature superimposed on the more or less permanent two-layer salinity stratification. While temperature and sea ice respond rapidly to the changes in atmospheric heat

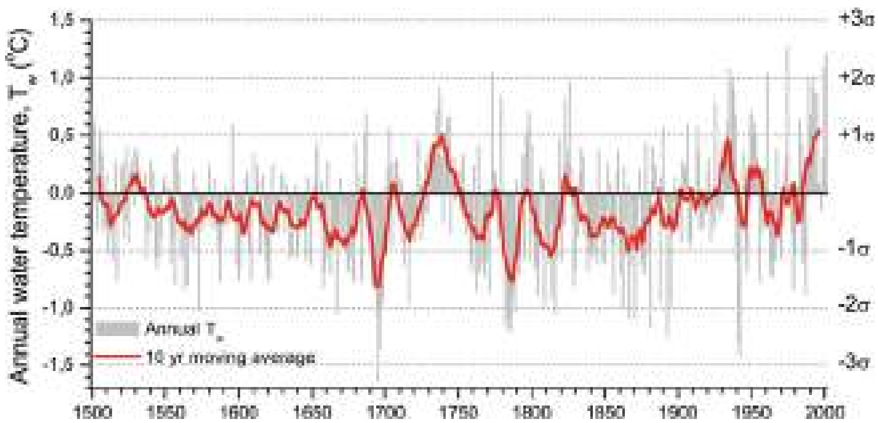


Fig. 1.4 Anomalies of the annual and decadal moving average of the modelled Baltic Sea spatial mean water temperature over the period 1500–2001. The dotted horizontal lines are the standard deviations of water temperature during the reference period 1900–1999 (Hansson and Omstedt 2008)

fluxes, variations in salinity are governed mainly by lateral transport processes, resulting together with mixing in response times of many decades. Cold waters, formed during winter, extend down to the halocline which has a typical depth of 60–80 m in the Baltic Proper and somewhat less in the southern basins. During summer, when a seasonal thermocline develops in the surface waters at depths of about 15–20 m, the underlying cold intermediate layer generally retains a 'memory' of the severity of the previous winter. This layer is often referred to as 'old winter water'. Deeper waters below the halocline are formed mainly by lateral advection of saline waters of North Sea origin that entrain and mix with ambient waters during their passage into and through the Baltic Sea. Below 100 m depth, the range in temperature variation within the Gotland Deep is only 5 °C (range 3–8 °C), compared to a range in surface temperature of up to 25 °C.

Water Temperature

A recent warming trend in sea-surface waters has been clearly demonstrated by in situ measurements, remote sensing data, and modelling results. In particular, remote sensing data for the period 1990–2008 indicate that the annual mean sea-surface temperature has increased by up to 1 °C per decade, with the greatest increase in the northern Bothnian Bay and large increases in the Gulf of Finland, the Gulf of Riga, and the northern Baltic Proper. Although the increase in the northern areas is affected by the recent decline in the extent and duration of sea ice, warming is still evident during all seasons and with the greatest increase occurring in summer. The least warming of surface waters (0.3–0.5 °C per decade) occurred north-east of Bornholm Island up to and along the Swedish coast, probably owing to an increase in the frequency of coastal upwelling. The latter is explained by the change in atmospheric circulation. Comparing observations with the results of centennial-scale modelling, recent changes in sea-water temperature appear to be within the range of the variability observed during the past 500 years (Fig. 1.4). Nonetheless, the twentieth century can be interpreted as the warmest, except for the warm anomaly around the 1730s.

Salinity

The salinity and stratification of the deep waters, governed by interacting circulation, transport, and mixing processes, are strongly linked to the major Baltic inflows of North Sea water that occur sporadically and bring high-saline water into the deep layers of the Baltic Sea. These major inflows are often followed by a period of stagnation during which saline stratification decreases and oxygen deficiency develops in the bottom waters. Major inflows normally occur during winter and spring; they bring relatively cold and oxygen-rich waters to the deep basins. Since 1996, several large inflows have occurred during summer. These inflows have transported high-saline, but warm and low-oxygen water into the deep layers of the Baltic Sea. Overall, a clear trend in salinity cannot be detected.

Sea Ice

Sea-ice conditions in the Baltic Sea have been systematically monitored for more than a century. The annual maximum ice extent of the Baltic Sea (MIB) is the most widely used indicator of sea-ice changes because it integrates winter period weather over the entire basin. A reconstruction based on various observational methods shows that the MIB displays large interannual variability owing to large-scale atmospheric circulation associated with the NAO (Fig. 1.5). A larger MIB occurs during negative NAO phases, while a smaller MIB occurs when the NAO is in a positive phase. All sea ice related parameters display large interannual variability, but a change towards milder ice winters has been observed over the past 100 years: in particular, the annual

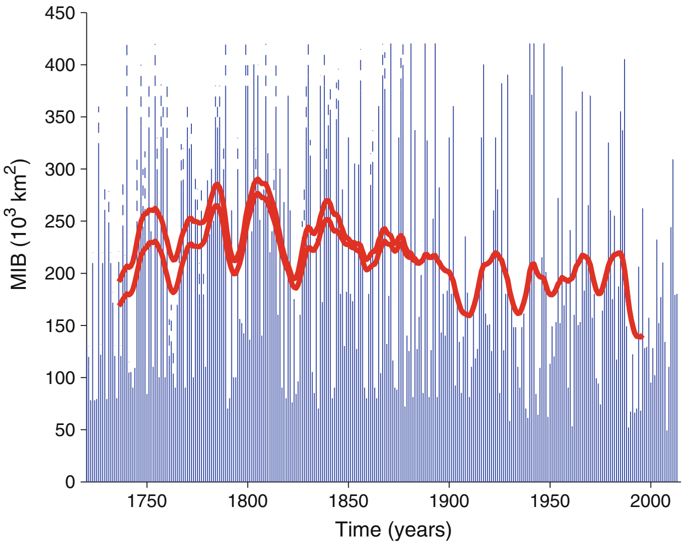


Fig. 1.5 Maximum ice-cover extent in the Baltic Sea (MIB), 1720–2012. The *dashed bars* represent the error range of the early estimates. The 30-year moving average is indicated by *two lines* representing the error range early in the series, converging into *one line* when high quality data became available

maximum ice extent has decreased and the length of the ice season has become shorter. However, a winter without any ice formation in the Baltic Sea is far from the present climatology. According to some 300 years of records of annual maximum sea-ice extent, the northernmost sub-basin, the Bothnian Bay, has been entirely ice-covered even during the mildest winters and the length of the ice season near the coast has been 150 days at a minimum.

There has also been a large change in the length of the ice season during the past century. In the Bothnian Bay, which has the longest ice season, the trend is −18 days per century. Greater changes have been observed in the eastern Gulf of Finland, where ice also forms every winter; over the past century, the length of the ice season decreased by 41 days per century, while in the past 50 years, it further decreased to −62 days per century. In the southern Baltic Sea, the length of the ice season decreases from east to west and from the inner waters towards the sea areas. A weak trend towards a smaller number of days with ice has been found for the last 30-year period.

The changes in ice conditions are consistent with the observed increase in temperature, but some sea-ice changes could also be caused by shipping. Ship-induced waves can prevent the formation of a permanent ice cover in the autumn as well as enhance the break-up of the ice cover during spring; thus, the increase in the size of vessels and the intensity of shipping activity in the Baltic Sea could also affect local ice conditions.

Sea Level

The overall change in mean sea level at the Baltic Sea coasts results from the combined effects of post-glacial rebound, the increase in global ocean mass, thermal expansion of sea water, and the contributions of regional factors that may cause an overall change and/or a redistribution of sea level within the Baltic Sea. The glacial isostatic adjustment exerts a strong influence in the Baltic Sea area, with a maximum uplift of the Earth's crust in the Gulf of Bothnia of approximately 10 mm year^{-1} and subsidence in parts of the southern Baltic Sea coast of about 1 mm year^{-1}. Thus, relative sea level is decreasing in the northern Baltic Sea region where the continental crust is rising, while sea level is rising in the southern Baltic Sea region where the continental crust is sinking (Fig. 1.6). In addition, many climate factors also influence relative sea level, including changes in water density (driven by changes in water temperature and salinity), changes in the total volume of the Baltic Sea, and meteorological factors.

Analysis of individual records of landlocked tide gauge measurements corrected for the vertical land movements indicates that Baltic Sea sea level may have risen during the twentieth century at rates of around 1.5 mm year^{-1}, which are close to the rate of global sea-level rise. However, uncertainty is large because the margin of error of the estimates of land uplift may be locally of the order of 1 mm year^{-1}. An accurate estimate of absolute, climate-induced, Baltic Sea sea-level rise over the twentieth century is still not available, but is unlikely to deviate much from the global average. In more recent decades, as satellite altimetry data have become available, the basinwide rate of sea-level rise may be around 5 mm year^{-1} (with an uncertainty of roughly ±3 mm year^{-1}) with the central estimate thus higher than the recent global mean of 3.2 mm year^{-1}.

Storm Surges

Extreme sea level is caused by storm surges driven by the passage of atmospheric cyclones. There is some evidence that the intensity of storm surges may have increased in recent decades in some parts of the Baltic Sea, and this has been attributed to long-term shifts in the tracks of some types

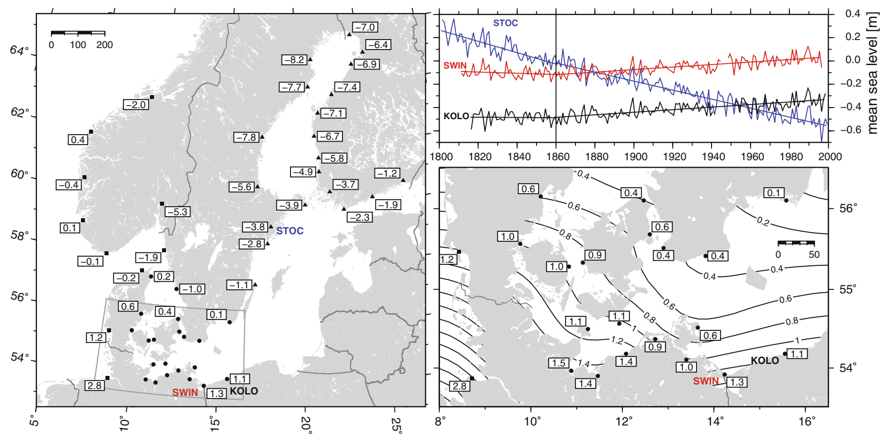


Fig. 1.6 Maps of secular (100 years) changes in relative sea level (RSL), based on tide gauge measurements of the entire Baltic Sea region (*left panel*), and in more detail, the southern Baltic Sea coast (*right panel below*) together with changes in the linear trend of the (arbitrarily shifted) annual RSLs at Stockholm, Swinoujscie (SWIN), and Kolobrzeg (KOL) between the period before and after 1860. The symbols represent the affiliation to different reference stations (*dots* Warnemünde, *triangles* Stockholm, *squares* Smögen) (redrawn from Richter et al. 2011)

of cyclone rather than to long-term change in the intensity of storminess. Analyses of storm surges have focused on local records, however, and there is no systematic basinwide analysis of change in storm surges yet available.

Waves

Analyses show no significant change in average wave activity in the Baltic Sea basin. However, extensive spatial patterns of changes within the basin exist, possibly leading to long-term variations in areas with the greatest wave intensity. Regional studies have even revealed different trends in average and extreme wave conditions that are probably due to systematic change in wind direction.

1.2.3 Future Climate Change

Regional climate models have been used extensively since the first assessment of climate change in the Baltic Sea region published in 2008, not least for studies of Europe (which includes the Baltic Sea catchment area). Therefore, conclusions regarding climate model results have a better foundation than was the case for the first Baltic Sea assessment.

1.2.3.1 Models and Methodology

Projections of future climate change make use of general circulation models (GCMs) that describe climate based on a set of grid points regularly distributed in space and time. The grid scale (i.e. the difference between two neighbouring points) of present-day GCMs is in the range 100–300 km. However, many important processes, such as cloud formation, convection, and precipitation, occur on much smaller spatial scales; thus, these sub-grid processes need to be approximated using simplifying algorithms termed parameterisations. To obtain estimates of regional climate, the results of GCMs are downscaled (a process linking large-scale features to small-scale features) with regional climate models (RCMs). RCMs have much higher resolution than GCMs and can better describe local features while still remaining able to simulate the atmospheric state in a realistic manner.

Climate models describe statistical features of states of the atmosphere over a long period of time. Climate variations are caused by changes in the environment, including the ocean, vegetation, ice, solar activity, and the composition of the atmosphere, with the strongest emphasis on the concentration of GHGs and aerosols. While some of these changes can be well estimated, others including changes in land cover and GHG concentrations are very difficult to predict. Thus, scenarios of possible future changes in world population and economic activity are developed and used to project how the climate could change in the future.

There are many sources of uncertainty in climate model results. These include uncertainty related to future changes in land cover and atmospheric GHG concentrations, the amount and accuracy of input data, and the chaotic nature of weather. Many sub-grid scale processes must be represented in models in a simplified form and are not well described by the models. For example, representations of cloud formation, the optical and radiative features of clouds, and the creation of atmospheric precipitation still carry considerable model error. The skill of methods for describing regional climate futures is also limited by natural climate variability.

Non-GHG drivers, such as aerosols and changes in land cover, are not fully represented in RCMs. This can be a source of major uncertainty in projections of future climate as a large part of the simulated multi-decadal variance in North Atlantic sea-surface temperature depends on levels of aerosols. Natural climate variability limits the skill of future climate predictability in many regions. In locations where the amplitude of natural variability is high, predictability is low, and vice versa. The uncertainty of future climate projections is largely a consequence of the chaotic nature of large-scale atmospheric circulation patterns, and improving models or GHG scenarios cannot eliminate this uncertainty. Scenarios, or projections, represent possible future developments and provide decision-makers with a variety of perspectives to consider when developing plans for the future. Scenarios are not predictions, at least when using IPCC terminology, even though the two terms are frequently confused. For example, the term "prediction" is often used when possible developments are meant.

1.2.3.2 Projections of Future Climate Change

Most regional climate change information from global models has originated in recent years from the World Climate Research Programme's Coupled Model Inter comparison Project phase 3 (CMIP3) multi-model data set. In CMIP3, about 20 different coupled atmosphere–ocean GCMs were used in a number of different simulations—on natural variability, on the effect of observed increases in anthropogenic forcing during the twentieth century, and on the expected effect of a number of possible increases in anthropogenic forcing (scenarios provided by the IPCC Special Report on Emissions Scenarios; SRES[2]) for the twenty-first century.

Downscaling of global climate model results has been undertaken with a number of RCMs. Such RCM simulations are generally conducted for the atmosphere only, using sea-surface temperature data from the driving GCM. For the case of the Baltic Sea, this is a severe limitation, because sea-surface temperature data taken from global models do not describe the Baltic Sea adequately. Still, there are some clear improvements since the BACC I assessment of 2008: Models now operate at higher horizontal resolution, and the simulations cover a larger degree of the uncertainty range including a wider range of emission scenarios (sampling the uncertainty in forcing), more climate models (addressing model uncertainty), and ensemble members (addressing natural variability).

Atmosphere

Temperature

Although there is a large spread between different GCMs, a clear increase in temperature is projected for all seasons (Fig. 1.7). While the pattern of highest warming in the north in winter is similar for all models, there is a spread in the magnitude of change. Changes increase with time and/or rising emissions of GHGs.

The land is expected to warm more quickly than the sea. This is particularly the case in winter when retreating snow and sea-ice cover are expected to enhance absorption of sunlight and increase heat storage in the soil, leading to higher temperatures. Cold extremes in winter and warm extremes in summer are expected to change more than the average conditions; implying a narrower (broader) temperature distribution in winter (summer).

Precipitation

Precipitation is projected to increase across the entire Baltic Sea region during winter, while in summer, increases are mainly projected for the northern half of the basin only. For the southern part of the Baltic Sea, there is a large spread between the different models including both increases and decreases, and thus, little clear change in precipitation is projected (Fig. 1.8).

Extremes of precipitation are also projected to increase. Some projections for the Baltic Sea area that show considerable decreases in average summer precipitation also show an increased likelihood of very extreme precipitation. In addition to increased intensity of extreme precipitation events, RCM simulations also indicate an increased frequency of such precipitation extremes.

Wind

Projections for future changes in wind speed are highly dependent on changes in the large-scale atmospheric circulation simulated by the GCMs. The results diverge, and it is not possible to estimate whether there will be a general increase or decrease in wind speed in the future. A common feature of many model simulations, however, is an increase in wind speed over oceans that are ice-covered in the present climate, but not in the future. Future changes in extreme wind speed are uncertain. Simulations of extremes of wind speed show an even wider spread than those for mean wind speed.

Snow

Simulations for the Baltic Sea area clearly show that the volume of snow in the region may decrease considerably in the future, even though a very few Scandinavian mountain areas may experience slight and statistically insignificant increases. There is a possibility that in extreme years the maximum amount of snow could be greater than in extreme years in the climate of the recent past, even if the total annual amount of snow is reduced. The southern half of the Baltic Sea catchment area is projected to experience on average significant reductions in the amount of snow, with median reductions of about 75 %.

[2] www.ipcc.ch/ipccreports/sres/emission.

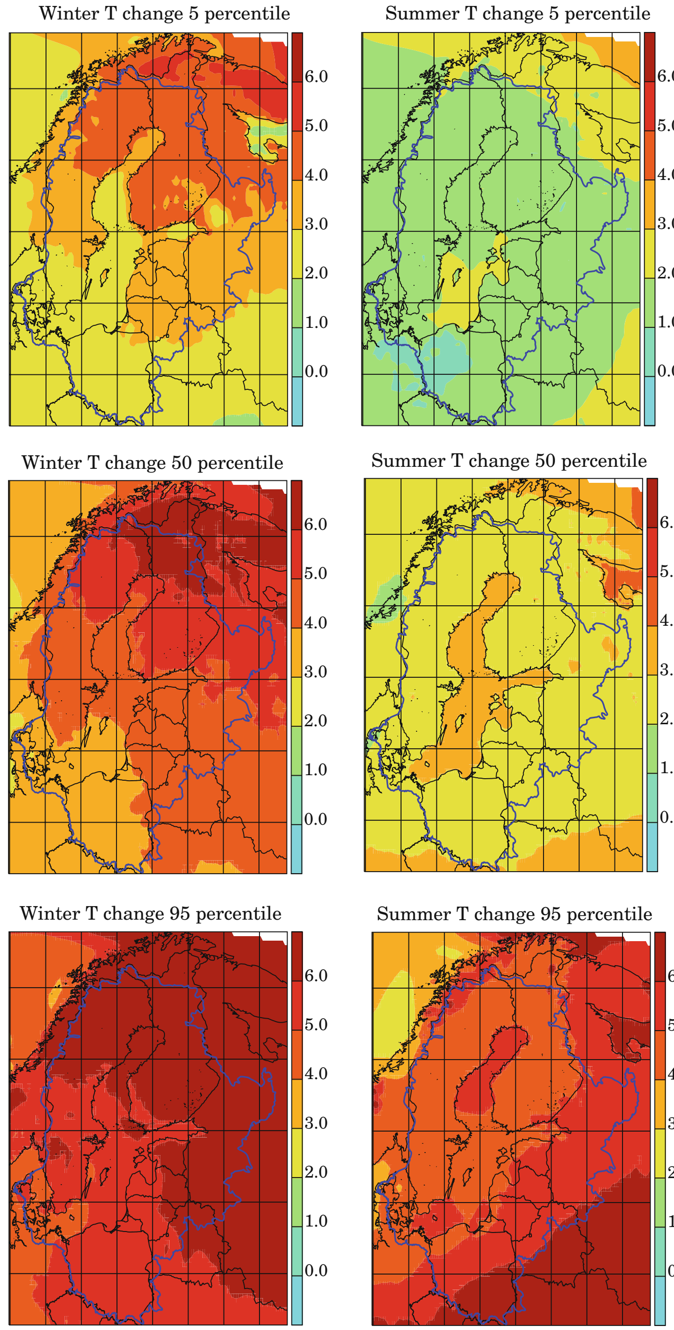


Fig. 1.7 Projected change in surface air temperature for 2071–2099 relative to 1961–1990 using the SRES A1B scenario as simulated by 13 RCM models from the ENSEMBLES project. *Left column* winter (DJF), *right column* summer (JJA). *Upper row* 5th percentile (corresponding to the lowest model result), *middle row* 50th percentile (corresponding to the median model result), *lower row* 95th percentile (corresponding to the highest model result). The *blue line* indicates the Baltic Sea catchment area

Hydrology

Future run-off will be influenced by an increase in evaporation owing to higher air temperatures as well as the increase in precipitation expected across much of the Baltic Sea area. A decrease in river run-off is possible despite increased precipitation, if increasing evaporation is the more important factor; but this remains contested due to the low numbers of simulations and the large uncertainties involved.

The annual cycle of run-off is expected to change considerably. For areas presently characterised by spring floods due to snow melt, the floods are likely to occur earlier in the year and their magnitude is likely to decrease owing to less snowfall and a shorter snow accumulation period. As a consequence, sediment transport and the risk of inundation are likely to decrease. In the southern part of the Baltic Sea area, increasing winter precipitation is projected to result in increased river discharge during winter. In addition, groundwater recharge is projected to increase in areas where the infiltration capacity is not currently exceeded, resulting in higher groundwater levels. Decreasing precipitation

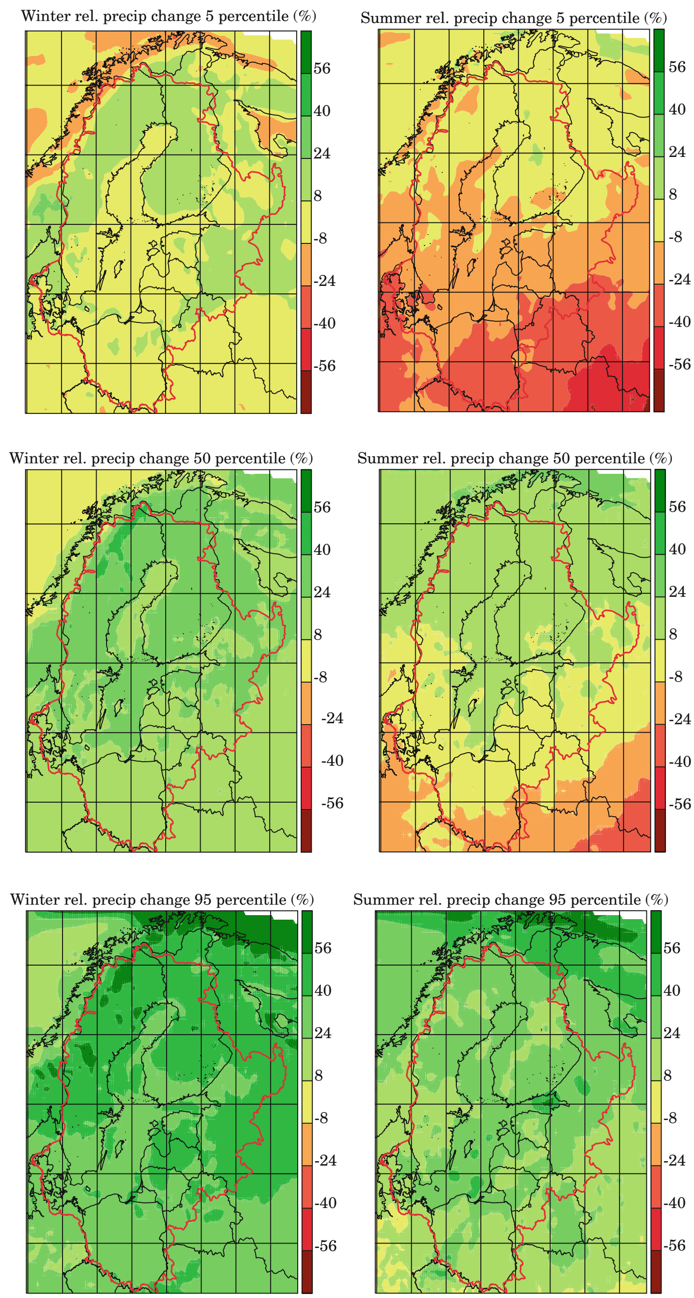


Fig. 1.8 Projected change in average precipitation for 2071–2099 relative to 1961–1990 using the SRES A1B scenario as simulated by 13 RCM models from the ENSEMBLES project. *Left column* winter (DJF), *right column* summer (JJA). *Upper row* 5th percentile (corresponding to the lowest model result), *middle row* 50th percentile (corresponding to the median model result), *lower row* 95th percentile (corresponding to the highest model result). The *red line* indicates the Baltic Sea catchment area

combined with rising temperature and evapotranspiration during summer is projected to result in a drying of the root zone which would drive increasing irrigation demands in the southern part of the Baltic Sea area.

Baltic Sea Hydrography

Over the past five years, there has been a considerable increase in the number of scenario simulations with respect to the Baltic Sea. Scenario simulations based on three coupled physical–biogeochemical models of the Baltic Sea, forced with atmospheric RCM data downscaled from two GCMs and two GHG emissions (the SRES A1B and A2 scenarios), are discussed below.

Water temperature

The greatest changes in water temperature are projected to occur in the Bothnian Bay and Bothnian Sea during summer and in the Gulf of Finland in spring. Using the SRES A1B and A2 scenarios, the summer sea-surface temperature is expected to increase by about 2 °C in the southern parts of the Baltic Sea and by about 4 °C in the northern parts. At least some of the greater change in the northern Baltic Sea is caused by the ice-albedo feedback owing to the decline in sea ice in winter. The surface water layer is projected to warm more than the deep water in all sub-basins of the Baltic Sea.

Salinity

The scenario simulations indicate that salinity may decrease and that changes in sea-surface salinity could be greatest in the region of the Danish Straits, especially in the Belt Sea, and small in the northern and eastern Baltic Sea, with the smallest change in the Bothnian Bay (Fig. 1.9). Changes in sea-surface salinity are projected to be reasonably uniform across seasons.

In the Bornholm Basin and Gotland Basin, the reductions in salinity with depth are nearly constant and were 1.5–2 g kg^{-1} in the ensemble mean; changes in the deep water were greater than in the surface layer in these sub-basins. In more weakly stratified basins, such as the Gulf of Finland and Bothnian Bay, there were greater differences in salinity changes in the surface and bottom layers, causing a reduction in vertical stability. The changes in salinity are due to changes in run-off, and as climate models have severe biases with regard to the water balance, it is still unclear whether Baltic Sea salinity will increase or decrease.

Studies of both past and future climates suggest that increased total run-off would increase the ventilation of the upper halocline due to weakened stratification causing improved oxygen conditions in the upper deep water. However, this is uncertain owing to biases in model results for the hydrological cycle.

Sea ice

The future reduction of sea-ice cover in the Baltic Sea mainly depends on the projected changes in air temperature during winter. All simulations indicate a drastic decrease in sea-ice cover in the Baltic Sea in the future, in agreement with earlier studies. However, even under a warmer future climate, sea ice is likely to occur in future winters in the northern Baltic Sea.

Storm surges

Projected changes in sea-level extremes caused by changes in the regional wind field indicate that at the end of the twenty-first century, the greatest changes in mean sea-surface height would occur during spring, amounting to up to 20 cm in coastal areas of the Bothnian Bay. The maximum change in annual mean sea-surface height is expected to be 10 cm without accounting for the large-scale sea-level rise or land uplift in the Baltic Sea area. Overall, sea-level rise has a greater potential to increase storm surge levels in the Baltic Sea than increased wind speed.

Sea level

Sea levels are rising owing to global warming, primarily as a result of the loss of ice masses on land and the thermal

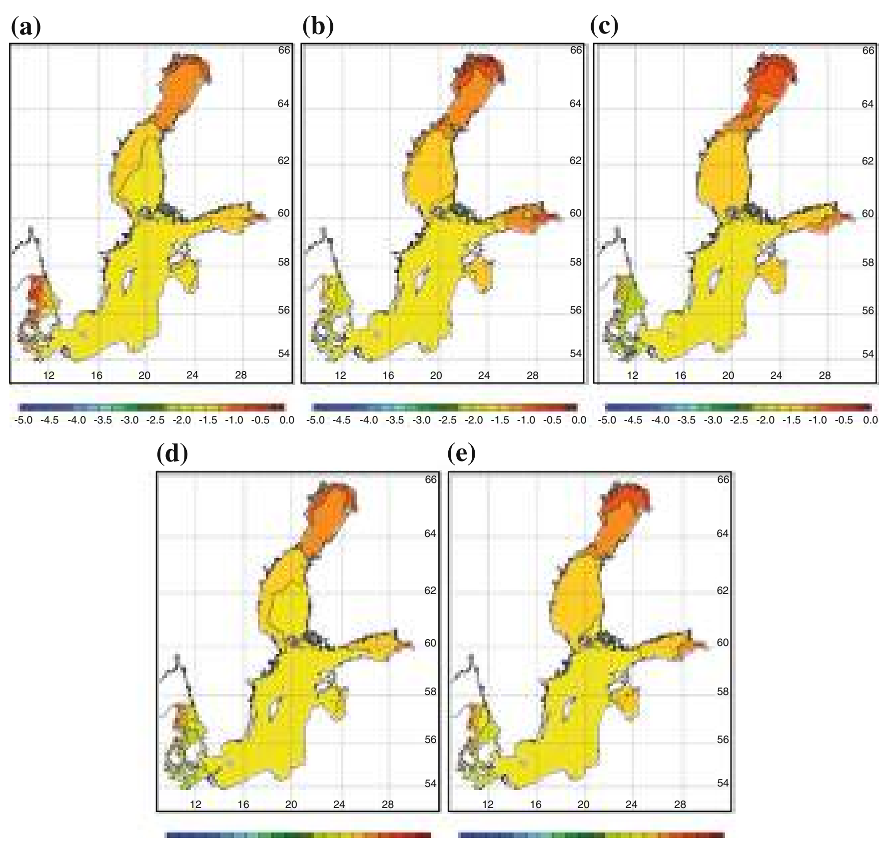


Fig. 1.9 Projected change in seasonal **a** DJF, **b** MAM, **c** JJA, **d** SON and **e** annual mean ensemble average sea-surface salinity for 2069–2098 relative to a baseline of 1978–2007. See Meier et al. (2012)

(steric) expansion of sea water. Sea-level rise varies in complex spatial patterns depending on several factors. In addition, relative sea level in the Baltic Sea is influenced by the large ongoing glacial isostatic adjustment resulting from the loss of the Fennoscandian ice sheet at the end of the last glacial period.

The magnitude of future sea-level rise is highly contested in the scientific community, and several years may be needed to reconcile the various estimates. More data extending over longer periods are needed, particularly for estimating the contributions from Greenland and Antarctica. The IPCC AR5 report summarises the state of knowledge regarding the global issue, with the change in the Baltic Sea strongly conditioned by these global changes.

There is considerable uncertainty in projections of sea-level rise over the twenty-first century and disagreement over the level of confidence assigned to different modelling approaches. BACC II has compiled a mid-range sea-level rise scenario through an assessment of process model projections and uncertainties. The mid-range scenario, based on the SRES A1B scenario, projects a rise of 0.7 m (±0.3 m) for sea level in the Baltic Sea until the end of the twenty-first century. It should be noted, however, that this is one projection from the full range of IPCC projections (0.26–0.82 m for AR5). This potential local sea level rise is partly compensated by vertical land movement which varies between 0 m per century in Denmark and roughly 0.8 m per century in the Bothnian Bay.

1.2.4 Environmental Impacts of Climate Change

This section describes the environmental impacts of climate change on the coastal and marine environments of the Baltic Sea basin as well as on atmospheric chemistry. It is shown that most of the observed environmental changes are due to several interrelated factors of which climate change is but one.

1.2.4.1 Atmospheric Chemistry

The main air pollutants addressed in BACC II are the acidifying compounds (sulphur and nitrogen oxides, as well as ammonia/ammonium) and ozone. In general, the main driver of changes in atmospheric concentration and deposition with time is changes in emissions rather than impacts of meteorological changes. The dramatic increase in emissions after the 1940s, in Europe and North America, resulted in substantial changes in reactive nitrogen (N_r) and sulphur deposition, and in ozone levels. Indeed, deposition of N_r species is now of particular concern for the Baltic Sea region. Reductions in emissions in Europe starting around the 1980s have resulted in significant reductions in sulphur and oxidised N_r compounds in the European atmosphere. Emissions of reduced N_r compounds have not declined to the same extent, and indeed in some areas ammonia (NH_3) emissions are increasing.

For the future development of air pollution in Europe, some climate-induced changes are potentially important, however. For example, potential increased shipping activity and new shipping routes within the Arctic may lead to increased nitrogen deposition in environmentally sensitive areas and even to increased phyto-toxic ozone uptake. A new understanding is also that higher temperatures may increase NH_3 emissions from evaporative sources over land by very substantial amounts (e.g. 20–50 %); a process that is not yet included in NH_3 emission inventories. In summary, while it seems likely that air pollution impacts from sulphur and oxidised nitrogen will be substantially reduced in future compared to recent years, and the situation for ozone and reduced nitrogen is still unclear and strongly dependent on policy developments, that is, future emission control measures, both at regional and (for ozone) hemispheric scale.

1.2.4.2 Coastal Ecosystems, Birds, and Forests

The ecosystems of the extensive coastal zone of the Baltic Sea are relatively unstable, owing particularly to land uplift and the effects of sea-level rise and drainage basin processes. Furthermore, human influences from the often densely populated coastal areas exert a wide range of impacts on this area. Land use varies greatly within the Baltic Sea drainage basin. The most notable contrast is between the agricultural south and the forested north. Some of the ecosystem responses are specific to the relatively simply structured brackish water ecosystems of the Baltic Sea. The high diversity of the regional geography and distribution of the different habitats must be considered when assessing ongoing and possible future change patterns. The combined effects of climate change and land uplift on coastal ecosystems have been little studied and need particular emphasis in the future.

Warmer terrestrial ecosystems and warmer coastal sea water affect the northward migration of terrestrial and aquatic species and result in longer reproductive periods for coastal fauna and flora. The biodiversity of the Baltic Sea is particularly sensitive to changes in salinity, which can have a cascading effect on food webs and interaction between aquatic and terrestrial ecosystems. The effects of climate change on salinity and water temperature can facilitate invasion by non-indigenous aquatic bird species, such as cormorants, which can cause major changes in coastal bird communities. The climate-mediated changes can also facilitate the invasion of mammalian predators which can cause major changes in coastal and archipelago ecosystems.

The positive impacts of climate change on forest growth are expected to continue. In relative terms, boreal forest stands may benefit more from climate change than temperate forest stands.

The Baltic Sea drainage basin is likely to undergo a change in the species composition of natural vegetation, with a predominantly northward shift of the hemi-boreal and temperate mixed forests. Projected losses of species are greater in the southern part of the Baltic Sea basin than in the north.

Terrestrial carbon storage is likely to increase in the Baltic Sea catchment area. However, land-use change can play an important role in the terrestrial carbon cycle and have both positive and negative impacts on carbon storage.

1.2.4.3 Freshwater Biogeochemistry

The effect of climate change on freshwater biogeochemistry and riverine loads of biogenic elements to the Baltic Sea is not straightforward and difficult to disentangle from other human drivers such as atmospheric deposition, forest and wetland management, nutrient loads from agriculture, municipalities and industry, and hydrological alterations by, for example, hydroelectric dams. Climate change is a compounding factor for all major drivers of freshwater biogeochemistry. The evidence for assessing the effect of climate change is still often based on small-scale studies in both time and space; however, qualitative assertions are possible.

The change in the seasonal distribution of the discharge regime of major boreal rivers has potentially large impacts on the redistribution of organically bound carbon and nutrients from land to sea. Areas with a mean annual temperature around 0 °C (i.e. around 61°N) are most sensitive to further warming.

Atmospheric deposition in the past decades probably had a stronger effect on freshwater biogeochemical conditions in the Baltic Sea drainage area than climate. However, this pattern may change as atmospheric deposition decreases.

Over the short term, climate change is unlikely to affect the spatial distribution of wetlands, except for palsa mires that cover too small an area in the boreal watersheds to be significant for element fluxes to the Baltic Sea. Results from small-scale field and modelling studies indicate that increased temperature and precipitation could increase the transport of dissolved organic matter to the Baltic Sea significantly. However, the large-scale impacts on the Baltic Sea basin are still unknown. Even a northward shift in boreal forest (i.e. Norwegian spruce) with climate change may alter quite significantly the biogeochemistry of the northernmost rivers over the long term.

Agricultural practices and urban sources have significantly increased nitrogen and phosphorus concentrations in the rivers draining the cultivated watersheds of the southern Baltic Sea catchment. Changes in climate are not uniform across the cultivated southern catchment area; in the southwestern part (i.e. Denmark and western parts of Germany), precipitation has increased since the 1980s and farmers are currently adapting to a warmer and wetter climate by selecting heat-demanding and nutrient-demanding crops like maize. Transitional countries such as Poland and the Baltic States may increasingly use fertilisers as a result of the European Common Agriculture Policies and change to a more meat-eating lifestyle which may lead to an increased nutrient flux to the Baltic Sea. Initial studies (that still need further scientific elaboration) indicate that total nitrogen fluxes to the Baltic Sea may increase by up to 70 % as a result of changes in water discharge and changes in lifestyle (including change in the demand for animal protein).

1.2.4.4 Marine Biogeochemistry

Biogeochemical processes in the marine environment of the Baltic Sea are mainly controlled by the biological production and decomposition of organic matter taking place in the context of the hydrography of the region. Carbon, nitrogen, phosphorus, and oxygen are the major elements in these processes, and their distributions and concentrations strongly influence the ecosystem of the Baltic Sea. Enhanced anthropogenic nutrient inputs via rivers and atmospheric deposition during the past century have resulted in major changes in the biogeochemistry of the Baltic Sea. Although the implementation of reduction measures since about 1980 has decreased inputs to a level that is comparable to that in around the 1960s, this is only reflected in a decrease in the nitrate concentrations in the winter surface water of the Baltic Proper. This is not the case for phosphate concentrations, which can partly be explained by the enhanced recycling of phosphate due to increased areas of anoxic water.

The increase of atmospheric carbon dioxide (CO_2) from 280 ppm during the pre-industrial era to a level of almost 400 ppm in 2010 has led to a corresponding increase in the mean $p\mathrm{CO_2}$ in surface water. In marine systems, this would be expected to cause a decrease in the pH by 0.15 units; however, the increase in alkalinity in the central parts of the Baltic Sea over the past 60 years has diminished this decrease by roughly 0.03 units.

There is still limited knowledge of the impact that future changes in climate and other anthropogenic drivers may have on the biogeochemical cycles of the Baltic Sea. Various factors may influence these cycles in different ways. Changes in precipitation and run-off patterns will influence the inputs of nutrients, alkalinity, and organic matter to the Baltic Sea. Higher temperatures will decrease the solubility of oxygen in sea water as well as accelerate many biological and biogeochemical processes. Future warming is expected to increase hypoxia given that temperature controls the stratification of the water column, the respiration of organisms, and the solubility of oxygen. Increasing areas of hypoxia and anoxia are anticipated owing to the increased nutrient inputs due to increased run-off, the reduced oxygen flux from the atmosphere due to higher temperatures, and the intensified biogeochemical cycling including mineralisation of organic matter.

Model simulations of scenarios concerning the future biogeochemistry of the Baltic Sea have estimated that, taking into account climate change, the implementation of nutrient reductions according to the Baltic Sea Action Plan will result in a slight decrease in the deep-water area covered with hypoxic and anoxic waters. In contrast, the ‘business-as-usual’ nutrient input scenario yielded an approximate doubling of the anoxic area. Simulations of the possible future development of the pH in the Baltic Sea (Fig. 1.10) showed that the rising atmospheric CO_2 concentrations mainly control future pH changes in the surface water, while eutrophication and enhanced biological production would mainly enhance the seasonal cycle of pH. All CO_2 emissions scenarios studied yielded a significant decrease in pH by the end of this century.

1.2.4.5 Marine Ecosystems

Increase in sea-surface temperature has been suggested to change seasonal succession and induces dominance shifts in primary producers in spring. Shifts in dominant species may affect the biogeochemistry and functioning of the pelagic ecosystem in the following summer. As increasing temperature and stratification also favour cyanobacteria, rotifers, and small cladocerans, the plankton community is projected to shift towards smaller sized organisms. Mesocosm studies also suggest that climate change may influence the seasonal succession of phytoplankton and zooplankton, potentially increasing the temporal mismatch between these groups in spring. Such changes may have negative consequences on zooplankton production and thus food conditions of planktivorous fish. A climate-induced decrease in salinity together with poor oxygen conditions in the deep basins would negatively influence the main Baltic Sea piscivore, cod. Several studies have confirmed that this causes cascading effects on clupeids and zooplankton. It is less clear whether the effects cascade from zooplankton to phytoplankton.

Reduced duration and spatial extent of sea ice would cause habitat loss for ice-dwelling organisms, affect the ice-modulated land–ocean interactions, and probably induce changes in nutrient dynamics within and under the sea ice.

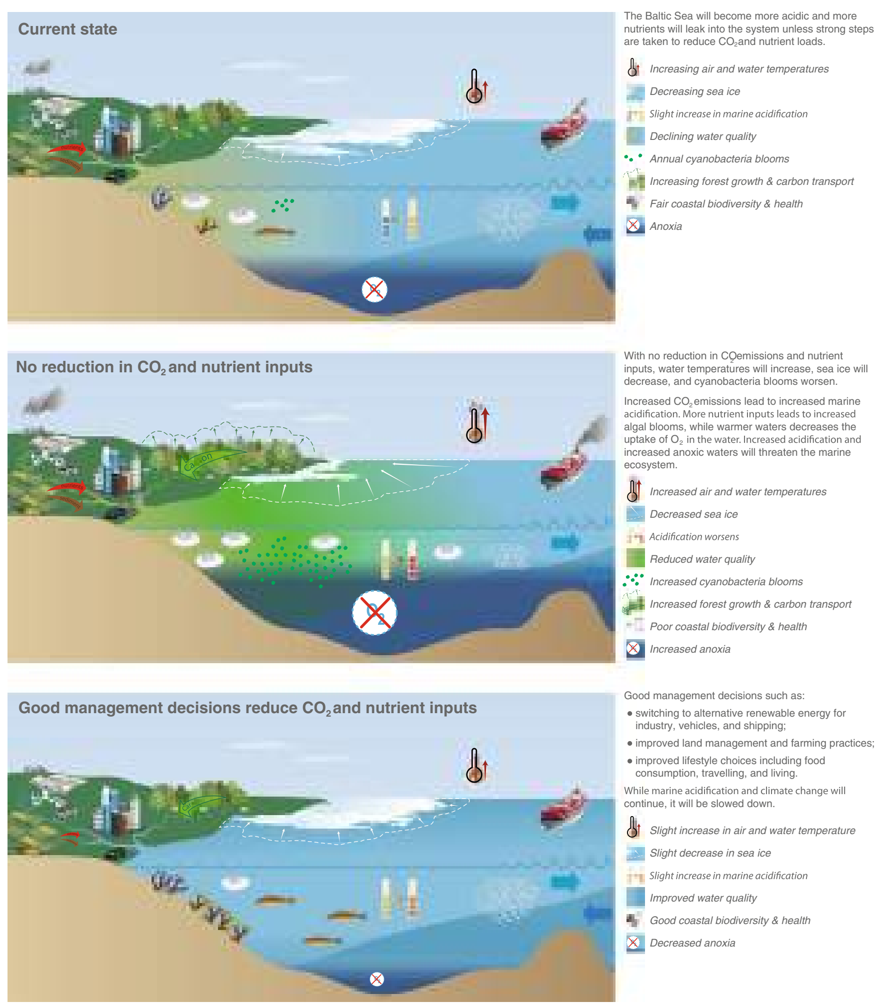


Fig. 1.10 Schematic of possible developments in the Baltic Sea area illustrating the findings of Meier et al. (2011), Omstedt et al. (2012), and others. Created by the Integration and Application Network, University of Maryland Center for Environmental Science, USA under guidance from A. Omstedt

There are, however, no estimates of the effects of declining sea ice on the overall productivity and pelagic–benthic coupling of the Baltic Sea ecosystem.

Modelling efforts suggest that climate change could worsen eutrophication by increasing freshwater discharge and thereby nutrient loads from land. An increase in sea-surface temperature would probably also favour cyanobacteria that binds nitrogen from the atmosphere and increases the supply of nitrogen to the nitrogen-limited phytoplankton. Summer primary production and sedimentation would then increase, worsening oxygen conditions and inducing the release of phosphorus from sediments. On the other hand, increasing the supply of freshwater and associated dissolved organic carbon may also reduce phytoplankton productivity, at least in the Gulf of Bothnia. Thus, it is clear that the effects of climate change on the productivity of the marine ecosystem vary from basin to basin.

Some of the most profound effects of the projected salinity decline involve losses in functional diversity that would accompany the loss of marine elements in the fauna. Also, the potential increase in primary production and sedimentation of organic matter in the northern Baltic Proper, as well the climate-driven decrease in trophic efficiency, as suggested for the Gulf of Bothnia, are potentially important factors for benthic communities. Acidification associated with high concentrations of CO_2 in the sea may also have severe implications for calcifying organisms such as bivalves. Key physiological processes including growth, metabolic rate, reproduction, and activity are also likely to be affected, thus potentially affecting the abundance, diversity, and functioning of benthic communities.

Human-induced pressures, such as overfishing and eutrophication, may erode the resilience of the Baltic Sea ecosystem, thereby making it more vulnerable to climatic variations. The Baltic Sea communities, that are poor in both species and genetic diversity, may therefore be particularly vulnerable to external forcing factors caused by the climate change.

1.2.4.6 Coastal Erosion and Coastline Changes

The Baltic Sea features a large variety of shorelines, from bedrock-dominated coasts to soft depositional shores. The response to climate-related changes differs for the various shore types in the Baltic Sea which include chalk cliffs, rocky shores, barrier islands with coastal lagoons, sandy beaches, flat clay shores, and esker shores.

The response of the coasts to climate change varies over different timescales, owing to changes in local dynamics or geo-morphological conditions as well as to the intensity of the driving forces. The main drivers of change in coastal geomorphology are geological structural resistance, changes in sea level, long-shore currents, and storm surges. These factors are responsible for both coastal erosion and accumulation, as well as for the emergence and variability of beaches. Identifying the contribution of climatic change to geomorphic changes can be difficult and will also vary regionally. Coasts are particularly vulnerable to extreme events.

The impact of wave energy on a shore depends on its exposure to the open sea and thus the potential for wave formation. The formation of sea ice also influences the shores of the Baltic Sea. Ice usually forms first in shallow inland bays, and the contact of the sea ice with the shore can cause local erosion. Simulations that project warmer winters indicate significant changes in ice conditions in the Baltic Sea, with the consequence that the interaction of the shores with sea ice may become less important, while the impact of wave energy may become more important. These changes would have a significant impact on coastal morphodynamics and ecosystems.

Maritime activities and shore protection also cause physical stress on the shoreline, particularly on shallow coasts and archipelagos. The low coasts of the Baltic Sea will be strongly affected by sea-level rise. Soft cliffs are also eroding due to heavy rain and storm surges. The combination of high water levels with strong wind can result in severe damage to soft coastal cliffs. Coastal erosion reduces the habitats for plant and animal communities, such as dune environments. In summary, the effects of climate change would include losses of sediment for coastal rebuilding, losses of valuable natural habitats, economic value and property, coastline changes due to extreme storm events, and increasing costs to society in terms of coastal protection measures.

1.2.5 Socio-Economic Impacts of Climate Change

The Baltic Sea basin is home to a diverse range of human activities that may all be affected by climate change in one way or another. The socio-economic impacts of climate change on the scale of the Baltic Sea basin have been the subject of a limited number of international and national studies. Two aspects are discussed here: impacts on the managed rural landscape (i.e. agriculture and forestry) and impacts on urban complexes (i.e. cities and towns).

1.2.5.1 Forestry and Agriculture

Climate change affects the vulnerability and productivity of agricultural and forestry systems predominantly through changes in precipitation and temperature patterns, and by changes in the frequency and intensity of risk factors for damage such as droughts, floods, storms, and biotic disturbances like pest infestations. In addition to changes in environmental factors, changing energy policies may influence agricultural and forestry systems through changes in demand for biomass for use as a biofuel. Effects differ with location, with growing conditions tending to improve in the northern boreal zone, with reduced precipitation and higher temperatures tending to result in deteriorating growing conditions in the southern temperate zone. Changing growing conditions are likely to cause shifts in forest structure and diversity. The importance of adapting management practices to altered conditions is clear and may allow increased yields and economic benefits as well as climate mitigation through substitution of fossil fuel energy with bioenergy. Evidence suggests that this is particularly the case for the northern parts of the Baltic Sea basin, while in the south, the potential for improved growing conditions might be counteracted by water stress and reduced growth in sensitive species such as Norway spruce. The need for management adaptation is especially clear in the south, in terms of change in thinning regimes, rotation periods, and species selection.

Conclusions on socio-economic impacts cannot be generalised because potential yield increases as well as loss risks from more unfavourable conditions must be considered. On the other hand, investment in better transport infrastructure in the north and the higher risk of storm damage, with market distortion, risk of species die-back, and more frequent bark beetle damage necessitating costly salvage cuttings, would be a considerable burden to forest management, increasing the need for planting where natural regeneration of current species is no longer suitable. A general decrease in tree age at harvesting may also decrease risk, irrespective of whether clear-cutting or selective cutting is practised.

Overall, the results highlight the importance of adaptive forest management strategies in the Baltic Sea basin and show positive benefits for forest management and conserving biodiversity. This could be of particular importance as management practises become more intensive, increasing the need to consider other aims (such as biodiversity and carbon mitigation).

It is clear from several studies that the effects of climate change on agricultural production in the Baltic Sea basin are likely to be mainly positive for crop yield, especially for winter crops. However, increasing climate variability will lead to a need for adaptation measures. These are manifold and will differ among region and crop species.

1.2.5.2 Urban Complexes

Urban complexes, that is, cities and towns, are characterised by high concentrations of buildings and built-up areas with consequent soil sealing, high concentrations of people and infrastructure as well as specific economic and cultural roles and activities. As every particular urban complex is characterised by a specific mix of social, ecological, and economic interdependencies and its specific settlement and building structure, it is difficult to generalise scientific findings on urban complexes.

Climate change impacts are determined not only by the specific structures of these urban complexes, but also by the vulnerability of the urban society, its socio-economic and institutional structure as well as infrastructure and its capacity to cope with impacts. Urban areas are usually characterised by higher temperatures than the surrounding countryside. Urban cold islands, where built-up areas are colder than areas outside the city core, also appear. These effects are not unique over the entire urban area and depend on urban land use. Moreover, climate change is not the only driver of change in urban complexes as they are also influenced by, among others, demographic change, changes in land use, and political and economic changes, which are interacting with climate change impacts.

For urban complexes in the Baltic Sea catchment, several climate change impacts are expected in very different fields such as on urban services and technical infrastructure, on buildings and settlement structures, on the urban economy, and on the urban population. The impacts differ based on the location of the urban complexes, whether they are in the northern or southern part of the catchment and directly at the Baltic Sea coast or inland. Specifically, sea-level rise is a main impact for coastal cities, especially at the southern Baltic Sea coast, threatening urban infrastructures and settlements and in some cases also the urban drinking water supply through the intrusion of saltwater into coastal aquifers. Changing precipitation patterns, heavy precipitation, and storm surges as well as rapid snow melting events can cause severe further problems for urban infrastructure and urban settlements through surface floods, rapid surface runoff, and changes in water quality and availability. Transport is especially vulnerable to flooding (through storm surges and heavy precipitation events) and sea-level rise. Heat stress and changes in air quality are expected to affect the urban population directly.

1.2.6 Drivers of Regional Climate Change: Detecting Anthropogenic Change and Attributing Plausible Causes

A new issue in BACC II is the question: Is there evidence that recent change is beyond natural variation, so that it may be concluded that external drivers are at work (‘detection’); and if so, which mix of such external drivers is most plausibly responsible for this change (‘attribution’). The drivers considered are global climate change, changing atmospheric aerosol loads and changing land use. Although there is a small amount of literature available on the impact of global climate change on the Baltic Sea region, for the other two issues scientific literature exists on mechanisms, but virtually nothing on quantitatively linking changing aerosol loads and changing land use to changes in the regional climate.

1.2.6.1 Regional Evidence of Global Warming

There is some indication of an emerging anthropogenic signal, which is detectable in thermal quantities such as seasonal temperature, but evidence for detection of changes in non-thermal quantities such as circulation and precipitation is weak.

Although human influence (mainly increasing concentrations of GHGs) has been identified as a cause of the recent warming in the Baltic Sea area, there are caveats. The causes of recent circulation changes in the Euro-Atlantic sector are not yet understood, and therefore, attribution of changes especially in winter and spring is to be treated with caution. Furthermore, quantification of the contribution of individual drivers has not been accomplished. Better understanding of the regional effects of natural drivers and the effect of anthropogenic aerosols may help to achieve quantified attribution statements. Finally, detection and attribution efforts are often subject to selection and publication biases. That is, time series with ‘interesting’ behaviour are preferentially studied and positive findings (of a detectable anthropogenic effect) are more likely to be published. To avoid these biases, systems to routinely issue statements on the contribution of external forcing to the observed climate across a range of pre-specified variables and regions are needed.

1.2.6.2 Aerosols

Scientific understanding of aerosol effects on the global and regional climate is still accumulating. Targeted analyses on regional aerosol effects in the Baltic Sea region are rare, and commonly used regional climate models are generally unable to simulate aerosol–climate interactions.

Anthropogenic aerosol emissions are mainly concentrated in heavily populated and industrialised regions. The Baltic Sea area contains such regions, which implies a considerably higher regional climate forcing. Major reductions in emissions of sulphur and nitrogen compounds have already been achieved, black carbon emissions will probably decrease far more, as will organic compounds emitted from the same combustion sources. As a result, the total climate effect from the expected changes in aerosol emissions will probably be minor, while emission reductions strongly affecting ozone and methane concentrations may reduce climate warming significantly. However, ozone and methane concentrations over Europe and the Baltic Sea area are strongly dependent on emissions over the northern hemisphere and the world as a whole, respectively.

Analyses on regional aerosol effects in northern Europe are rare, and the commonly used regional climate models are mostly unable to simulate aerosol–climate interactions. However, recent modelling efforts investigating the influence of European aerosol emissions indicate an effect on large-scale circulation over Europe that is very likely to have affected the climate in the Baltic Sea region. To what extent is still not known. Development of the modelling capability and targeted analyses are urgently needed to reduce uncertainties related to the effect of changes in aerosol concentration on regional climate.

1.2.6.3 Land Cover

Anthropogenic land-cover change is one of the few climate drivers for which the net direction of the climate response (warming or cooling) over the past two centuries is still not known with certainty. The major uncertainty is due to the often counteracting temperature responses to biogeochemical versus biogeophysical effects, but also to the difficulty of quantifying the counteracting effects of changes in albedo and hydrological cycle (both biogeophysical effects), as well as obtaining precise land-cover data for the past.

In particular, land-cover change affects the exchange of heat between the land and the atmosphere (biogeophysical effect) and the sequestration of CO_2 (biogeochemical effect). These two effects often have opposite consequences on climate (mainly temperatures) for the same vegetation change. However, these are only two of many biogeophysical and biochemical effects. Different biogeophysical effects may also have opposite results on the climate between them. Depending on the respective size of all biogeophysical and biochemical effects, the net result will be a warming, a cooling, or no change. While no quantitative studies are available for the Baltic Sea region on the effect of changing anthropogenic land-cover since the beginning of industrialisation, there is ample evidence that important land-use changes have taken place in the past in this region. The most substantial deforestation of the region occurred about 2500–2000 years ago as indicated by palaeoecological evidence; later, deforestation gradually increased until the nineteenth century. The net effect of this long-term deforestation on past climate, at the global or regional scale, is debated but still unknown, as is the net effect on climate of the reforestation of the nineteenth and twentieth centuries (by spruce and pine plantations) in the southern part of the Baltic Sea region.

A study on land cover–climate interactions in Europe 6000 and 200 years ago, using a regional climate model, indicates that anthropogenic land-cover changes between these two time frames had significant effects on climate and that the effects differed between seasons and regions. Other modelling studies show that simulated large-scale afforestation of the northern hemisphere leads to an albedo-induced global mean warming that would offset the cooling due to the reduction in atmospheric CO_2 through sequestration by trees.

Studies on land cover–climate interactions in the northern hemisphere and reconstructions of past human-induced land-cover change in the Baltic Sea region suggest there is no evidence that anthropogenic land-cover change would be one of the factors responsible for the recent warming in the Baltic Sea region.

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Part I

Long-term Climate Change

Introduction

Part I addresses climate change in the Baltic Sea area over the past 12,000 years. Chapter 2 describes the climatic conditions of the Holocene, largely reconstructed on the basis of various types of proxy data. These include lake and bog sediments, pollen data and isotopic analysis. Chapter 3 describes climatic variability over the past millennium, based mainly on tree-ring analysis, historical documents that report extreme weather events and regional climatic model simulations. Climate change on the millennial timescale was not addressed in the first assessment of climate change in the Baltic Sea area (BACC Author Team 2008).

2 Climate Change During the Holocene (Past 12,000 Years)

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Abstract

This chapter summarises the climatic and environmental information that can be inferred from proxy archives over the past 12,000 years. The proxy archives from continental and lake sediments include pollen, insect remnants and isotopic data. Over the Holocene, the Baltic Sea area underwent major changes due to two interrelated factors—melting of the Fennoscandian ice sheet (causing interplay between global sea-level rise due to the meltwater and regional isostatic rebound of the earth's crust causing a drop in relative sea level) and changes in the orbital configuration of the Earth (triggering the glacial to interglacial transition and affecting incoming solar radiation and so controlling the regional energy balance). The Holocene climate history showed three stages of natural climate oscillations in the Baltic Sea region: short-term cold episodes related to deglaciation during a stable positive temperature trend (11,000–8000 cal year BP); a warm and stable climate with air temperature 1.0–3.5 °C above modern levels (8000–4500 cal year BP), a decreasing temperature trend; and increased climatic instability (last 5000–4500 years). The climatic variation during the Lateglacial and Holocene is reflected in the changing lake levels and vegetation, and in the formation of a complex hydrographical network that set the stage for the Medieval Warm Period and the Little Ice Age of the past millennium.

2.1 Introduction

The evolution of the coastline in the Baltic Sea area since the end of the last deglaciation has been the object of study for more than one hundred years. Despite this, some questions remain concerning both the chronologies of the transgression–regression phases of the prehistoric Baltic Sea basins and their spatial characteristics. Nevertheless, the long-term change in environmental conditions does provide information about the magnitude of natural climate and environmental variability, both that resulting from external climate drivers and that internally generated. This knowledge is important not only for understanding the mechanisms driving variability and the trend in climate change over centennial timescales,

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_2

but also to estimate the extent to which future climate could deviate from the monotonic long-term global trend caused by increasing concentrations of atmospheric greenhouse gases. At regional scales, deviations from the global average can be substantial and should be accounted for when designing regional policy for adaptation to climate change.

There are four main stages to the Lateglacial and Holocene history of the Baltic Sea basin:

- *Baltic Ice Lakestage*. This covers the deglaciation to ca. 11,550 cal year BP (calendar years before present)
- *Yoldia Sea stage*. This covers ca. 11,700–10,700 cal year BP (a brackish water basin in the first part of this stage and freshwater basin during the second; some studies place the end of the Yoldia Sea stage at 11,100 cal year BP)
- *Ancylus Lake stage*. A freshwater basin ca. 10,700–9500 cal year BP
- *Littorina Sea stage*. A brackish water basin, ca. 9500 cal year BP to present (Hyvärinen et al. 1988; Björck 1995, 1999; Andrén 2003; Andrén et al. 2002; Heinsalu and Veski 2007; Zillén et al. 2008). From colonisation by freshwater molluscs, the Limnea Sea has been dated at 4400 cal year BP and located in northern Estonia (Saarse and Vassiljev 2010).

The Baltic Sea basin and the Atlantic Ocean have exchanged water masses over almost their entire history. The exchange has been modulated both by glacio-isostatic land uplift and by eustatic (water volume) changes in sea level. Approximately 250 years after the final drainage of the Baltic Ice Lake, seawater flowed into the Baltic Sea basin through the Närke Strait in Billingen following a rapid rise in sea level resulting from the melting of the Scandinavian ice sheet (Yu 2003). This caused the Baltic Sea basin to become a brackish basin between 11,700 and 10,700 cal year BP (Yoldia Sea stage). Subsequent climate warming caused rapid melting of the continental ice sheets and pronounced isostatic uplift led to the isolation of the Baltic Sea basin from the Atlantic Ocean, turning it into a large freshwater lake (Ancylus Lake). The culmination of the transgression phase of this lake is dated to ca. 10,700 cal year BP (Yu 2003). The Ancylus Lake stage lasted until approximately ca. 9500 cal year BP. The modern Baltic Sea basin is part of the Littorina stage during which sea level and salinity have varied considerably (Miettinen et al. 2007).

Climate change in the Baltic Sea basin during the Holocene has been the result of various external and internal factors: changes in incoming seasonal solar radiation due to slow changes in the Earth's orbit, variations in the concentration of stratospheric aerosol caused by volcanic activity, change in the greenhouse gas content of the atmosphere due to natural factors, change in surface albedo of the sea-lake itself and in the surrounding land vegetation, and change in the intensity and type of circulation due to changes in basin salinity. The widely varying lake levels and vegetation changes and the formation of a complex hydrographical network together indicate a complex pattern of climatic change in the Baltic Sea region through the Lateglacial and Holocene. These environmental changes also affected the stages of human migration in this territory during this period.

This chapter summarises the climatic and environmental information that can be inferred from proxy archives of the Baltic Sea basin. These cover approximately the past 12,000 years and include different types of proxy data, mostly from continental and lake sediments, insect remnants, and isotopic data. There is also a brief discussion on the sources and mechanisms of externally driven and internally generated climate variability at timescales relevant for the Holocene.

2.2 Causes of Climate Variability During the Holocene

Climate variability in the Baltic Sea basin over the past 12,000 years has been caused by changes in external climate drivers, or internally generated by nonlinear dynamics and interactions among the different components of the climate system. During the deglaciation that began in the Last Glacial Maximum and later in the Holocene, the main external climate drivers were the modulation of the orbital parameters of the Earth, change in the solar irradiance, volcanic activity, change in the properties of land cover (see Chap. 25), and the concentration of greenhouse gases in the atmosphere. Distinguishing between external and internal drivers is a matter of perspective. For instance, the effect of greenhouse gases and changes in the land surface are usually considered external climate forcers. However, if changes in the biosphere and the dynamics of the carbon cycle are included as interactive components of the climate system, these would constitute a source of additional internal variability. This apparent contradiction can be illustrated using the example of large-scale circulation patterns. From a regional perspective, they may be considered external climate forcers, whereas changes in the circulation patterns may in fact be due to changes in the external radiative forcing or simply the result of internal climate variability, partly also due to processes originating within the Baltic Sea region itself.

2.2.1 External Climate Forcing

According to present knowledge, the externally forced climate variability in the Baltic Sea basin is most likely to be due to orbital forcing at millennial timescales, to changes in solar irradiance at multi-decadal or centennial timescales, and to volcanic activity at annual to multi-decadal timescales. The multi-decadal climate response to volcanic forcing arises because, although the climate effect of a single volcanic eruption may just last a few years, there exist multi-decadal periods where the frequency of eruptions has been considerably greater than average, for reasons yet unknown. A series of eruptions can then tip the climate system to a colder state giving rise to a longer-lasting climate response (Borzenkova 2012; Miller et al. 2012).

2.2.1.1 Astronomical Conditions

The only source of external climate forcing that can be accurately calculated is the Earth's orbital forcing (see Fig. 2.1 for an illustration of the Late Holocene trends). For longer periods, the evolution of solar insolation deviates too strongly from linearity to be represented by linear trends.

Variations in the position of the point of the orbit closest to the sun (perihelion), the obliquity of the Earth's axis and the eccentricity of the orbital ellipse, redistribute the incoming solar energy through the seasons and across latitudes. Although it is acknowledged that orbital variations are the main cause of the recurrence of glacial and interglacial periods, orbitally induced temperature variations can be reinforced by other factors, such as natural variations in atmospheric concentrations of greenhouse gases. The precise mechanisms that result in the interglacial successions, and thus caused the last deglaciation that extended from 20,000 to 10,000 cal year BP, are still not completely understood, and climate models are not yet able to reproduce the millennial-scale evolution of the rise in global sea level since the Last Glacial Maximum as reconstructed from proxy records (Brovkin et al. 2012). Since obliquity and eccentricity vary over longer timescales, the most important orbital factor during the relatively short Holocene period was the shift in the perihelion, from July 10,000 years ago to its current position at the beginning of January. Thus, northern high-latitude summers have received diminishing amounts of solar insolation over the past few millennia, with a linear trend more negative than −5 W m^{-2} per thousand years at the latitudes of the Baltic Sea region (about 60°N). In contrast, winter, spring, and autumn have received increasing levels of insolation over the Holocene. Since most proxy records reflect summer mean temperature or the temperature in the biological growing season, the millennial trends or the changes between mid-Holocene and present reflected in the proxy records should be interpreted based on biological knowledge about that particular proxy.

2.2.1.2 Solar Activity

Solar irradiance, the energy of the sun reaching the upper layers of the atmosphere (the solar 'constant'), also varies over all timescales due to internal solar dynamics. Past solar irradiance can be approximately reconstructed and dated by analysing the concentrations of cosmogenic isotopes, beryllium-10 (^{10}Be) in polar ice cores and carbon-14 (^{14}C) in tree rings, and over the past 400 years from the reported number of sun spots (see for instance, Crowley 2000). Although the general shape of the time evolution for solar irradiance is generally agreed upon, the magnitude of its variations at centennial timescales is still contended. Some authors suggest values of the order of 0.5 % of the total solar constant (Shapiro et al. 2011), while others interpret the isotopic record to indicate typical changes of the order of 0.1 % (Schmidt et al. 2011). The recent reconstruction by Shapiro et al. (2011), displaying amplitude of centennial variations of the order of 0.5 %, has been critically assessed by the climate research community.

Figure 2.2 illustrates the wide range of uncertainty between two multi-millennial reconstructions of total solar irradiance by displaying two extreme estimates, both very recent. Thus, it is still unclear whether cooling or warming

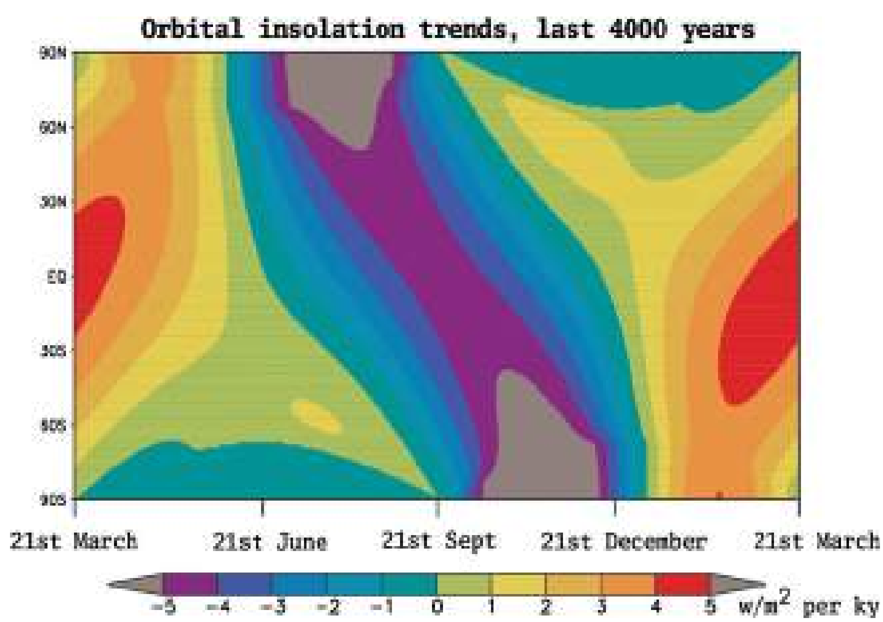

Fig. 2.1 Linear trends in solar insolation in the boreal mid- and high latitudes as a function of season and latitude over the past 4000 years, calculated from data by Laskar et al. (2004)

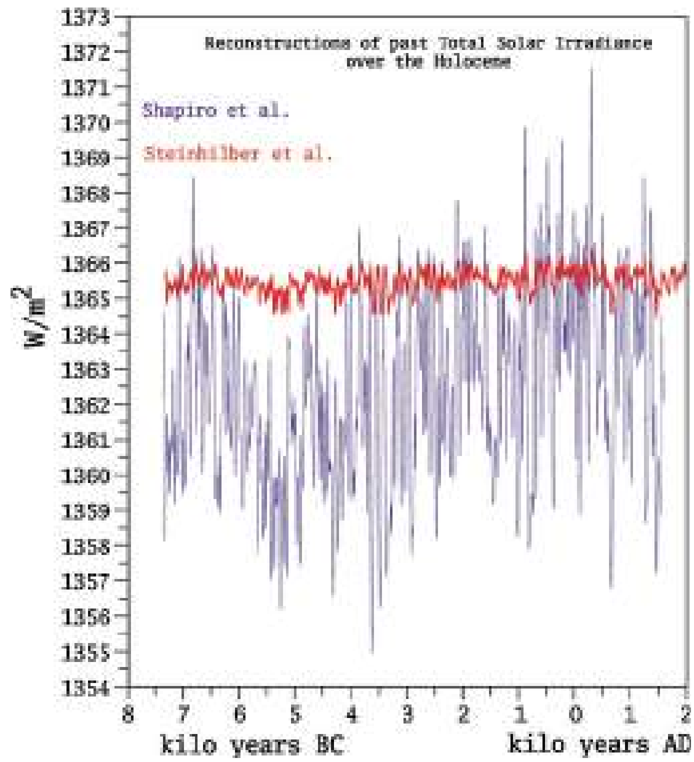

Fig. 2.2 Reconstructions of past total solar irradiance based on ^{10}Be concentrations in polar ice cores and different assumptions about solar physics. Adapted from Steinhilber et al. (2009) and Shapiro et al. (2011)

events can be confidently attributed to changes in solar irradiance alone, even when they are coincidental with periods of markedly different solar irradiance, or whether reinforcing mechanisms must be invoked (Shindell et al. 2001). Over the past millennium, for which more reliable reconstructions of volcanic activity exist than for previous periods, this situation is compounded by coincidences between solar minima and clustering of volcanic eruptions, as for instance during the Late Maunder Minimum and the Dalton Minimum. It is not known whether this coincidence also occurred prior to the past millennium.

2.2.1.3 Volcanic Eruptions

Past volcanic forcing can be reconstructed for recent centuries and, but with greater uncertainty, for the past few millennia by analysing the acidity of ice layers in polar ice cores (see Schmidt et al. 2011 for a thorough discussion). Volcanic eruptions produce vast amounts of sulphate aerosols which are transported through the stratosphere and deposited on snow and ice. The intensity, location, and seasonality of the eruptions (all important factors for estimate their climate effect) can only be indirectly inferred, which increases the uncertainty of estimating past volcanic forcing. Over the pre-industrial period, volcanic forcing is likely to have been the major factor driving external climate variability at multi-decadal and centennial timescales (Hegerl et al. 2003; Borzenkova et al. 2011; Borzenkova 2012; Miller et al. 2012). For periods further back in time, the lack of data precludes a robust inference.

2.2.1.4 Greenhouse Gases

Atmospheric concentrations of carbon dioxide (CO_2), methane (CH_4) and nitrous oxide (N_2O)—greenhouse gases—have varied strongly since the Last Glacial Maximum and, albeit in a more subtle way, during the Holocene. Atmospheric CO_2 levels rose from 180 to 280 ppm between the glacial state and the start of the Holocene 10,000 cal year BP (Borzenkova 1992, 2003; Blunier and Brook 2001; Flückiger et al. 2002). Since then, there has been a slight, and intriguing, positive trend, the causes of which are unclear (Ruddiman et al. 2011).

Several factors, such as reforestation, coral growth and ocean geochemistry (carbonate compensation), are likely to be contributed to the changes in Holocene CO_2 concentrations (Joos et al. 2004). However, this small trend in rising Holocene CO_2 levels could only have had a very small effect on climate, if any at all. (Only from about the year 1750, the point at which atmospheric greenhouse gas concentrations rose sharply, could they have made a significant contribution to climate change.)

2.2.2 Climate Modelling of the Holocene in the Baltic Sea Basin

Ideally, the external forcing mechanisms described in this chapter could be used to drive a global climatemodel that would simulate the Earth's climate over the Holocene.

To date, this has only been possible using simplified climate models, since the computing requirements needed to reconstruct a period of several thousand years are considerable. These 'models of intermediate complexity' cannot adequately represent the Baltic Sea basin. Although the first simulations with comprehensive ocean–atmosphere global models covering the past 7000 years have recently been completed (Hünicke et al. 2010), the resolution of these models is still insufficient for a realistic representation of the Baltic Sea. For example, in these models the North Sea–Baltic Sea gateway is either closed or is 700 km wide. To properly simulate the climate of the Baltic Sea area, regional atmosphere–ocean models with a high spatial resolution are required (see Chap. 10). Such models, however, have not yet been applied for the long timescales of the Holocene. This is a task that could be envisaged for the coming decade, as it is not presently feasible. However, some ongoing projects will soon be able to simulate particular time slices (e.g. a few hundred years), of the Baltic Sea climate in the Holocene. A regional simulation of the past millennium was recently performed at the Swedish Meteorological and Hydrological Institute (Schimanke et al. 2012, see Chap. 3); this time span will hopefully be expanded back into the past.

In addition to the external climate drivers, nonlinear mechanisms within the different components of the climate system give rise to internal climate variability over all timescales. At the timescales relevant for the Holocene as a whole, multi-centennial and millennial, interaction between the hydrosphere and ocean dynamics are believed to have been the main factors modulating the climate variability superimposed on the millennial trends caused by the orbital forcing. Perhaps the best known episode of multi-centennial climate variability in the Holocene is the '8.2 ka cold event'; a sudden cooling that lasted for about 200–300 years, and that has left its imprint on many proxy records around the North Atlantic basin including northern Europe, as reported in several comprehensive reviews (Alley and Ágústsdóttir 2005; Rohling and Pälike 2005). A series of modelling studies (Renssen et al. 2001) supported the idea that the 8.2 ka cold event was caused by a sudden slowdown of the North Atlantic Meridional Overturning Circulation (AMOC), itself caused by a sudden release of melt water from the remnants of the Laurentide Ice Sheet into the North Atlantic Ocean. However, a large uncertainty concerning the volume of melt water which may have been discharged, its timing and the discharge rate still remains. According to this mechanism, freshening of the surface waters in the North Atlantic Ocean hindered oceanic convection and the deep-water formation that is typical of the present climate, thus reducing the amount of heat flux from the ocean to the atmosphere and also slowing the northward heat advection by the North Atlantic currents. Oceanic convection would have resumed once the anomalously fresher seawater in the high-latitude North Atlantic Ocean had been re-distributed through the world ocean, which would have required a period of a few hundred years (Clark 2001; Clark et al. 2002; Clarke et al. 2004).

2.3 Palaeoclimatic Reconstructions Over the Holocene

2.3.1 Sources of Palaeoclimatic Data

Pollen records from mire deposits, lake sediments and sea sediments have been used to reconstruct climate variability during the Late Glacial and the Holocene periods. New quantitative information has been obtained from pollen and chironomid data over the past few years due to more detailed analysis and ^{14}C dating of sediments. The new insights are based on calibration of the modern pollen, diatom and beetle fauna data to climate parameters (annual and seasonal air temperature and annual precipitation) using different types of transfer functions (Birks 2003). However, the results from earlier investigations are still relevant.

The most recent and carefully investigated sediment sequences used for climate reconstruction stem from the following lakes: Svētiņu (Veski et al. 2012), Kurjanovas (Heikkilä et al. 2009), Busnieks (Grudzinska et al. 2010; Ozola et al. 2010), mires: Cena (Kalniņa 2007, 2008), Eipurs, Dzelve (Kuške et al. 2010a), Rozhu (Kuške et al. 2010b) and from archaeological sites at Lubans (Kalniņa et al. 2004), Sarnate (Kalniņa et al. 2011) and Priedaine (Cerina et al. 2010) (see Table 2.1). Some of the pollen records have been re-interpreted using the model REVEALS. This model estimates the relative species composition in a region from the pollen counts found in lake sediments layers, describing the dispersal and deposition of pollen and taking into account inter-species differences in pollen productivity (Sugita 2007; Ozola and Ratniece 2012).

2.3.2 Methodology for Palaeoclimatic Reconstructions

There are three established ways of using palaeobotanical data for palaeoclimatic reconstructions. All three involve the use of plant species (or sometimes genera) as climate indicators together with statistical methods to infer climate by mapping vegetation composition on climatic zones. Because these methods all extrapolate the ecological requirements of modern plants back into the past, this necessarily introduces a certain degree of uncertainty.

The first method was published by Iversen (1944), who examined the modern distributions of *Ilex aquifolium* (holly), *Hedera helix* (ivy), and *Viscum album* (mistletoe) in

Table 2.1 Location of sites dated by ^{14}C used in this study

Site No.	Location	Lat (N)	Lon (E)	Source
1	Toskaljavri, Finland	69°12	21°28	Seppä et al. (2002a)
2	Tsuolbmajavri, Finland	68°41	22°05	Kornola et al. (2000), Seppä and Birks (2001)
3	Tibetanus, Sweden	68°20	18°42	Hammarlund et al. (2002)
4	L. Spåime, Sweden	63°07	12°19	Velle et al. (2005)
5	Nautajärvi, Finland	61°48	24°41	Ojala et al. (2008)
6	Klotjärnen, Sweden	61°49	16°32	Giesecke (2005), Giesecke et al. (2008)
7	Laihalampi, Finland	61°29	26°04	Heikkilä and Seppä (2003)
8	Mentilampi, Russia	61°22	29°15	Davydova et al. (1998)
9	Holtjärnen, Sweden	6°39	15°56	Giesecke (2005)
10	Medvedevskoye, Russia	60°13	29°54	Subetto et al. (2002)
11	Pastorskoye, Russia	60°13	30°02	Subetto et al. (2002)
12	Arapisto, Finland	60°35	24°05	Sarmaja-Korjonen and Seppä (2007)
13	L. Gilltjärnen, Sweden	60°05	15°50	Antonsson et al. (2006)
14	Vääna Lagoon, Estonia	59°22	24°25	Saarse and Vassiljev (2010)
15	Flarken, Sweden	58°33	13°44	Seppä et al. (2005)
16	Trehörningen, Sweden	58°33	11°36	Antonsson and Seppä (2007)
17	Igelsjön, Sweden	58°28	1°44	Hammarlund et al. (2003, 2005)
18	Lake Vättern, Sweden	58°48	14°36	Björck et al. (2001)
19	Pulli, Estonia	58°26	24°35	Veski et al. (2005)
20	Löpe, Estonia	58°26	24°35	Veski et al. (2005)
21	Cena Mire, Latvia	57°69	22°49	Kalniņa (2008)
22	Sarnate site, Latvia	57°10	21°47	Kalniņa et al. (2011)
23	Busnieks Lake, Latvia	57°23	24°51	Grudzinska et al. (2010)
24	Bazhi Mire, Latvia	57°69	22°48	Pakalne and Kalniņa (2005)
25	Engure Lake, Latvia	57°18	23°15	Kalniņa et al. (2012)
26	Dzelve Mire, Latvia	57°23	24°51	Kuške et al. (2010a)
27	Eipurs Mire, Latvia	57°24	24°62	Kuške et al. (2010a)
28	Roûge, Estonia	57°44	26°54	Veski et al. (2004)
29	Rozhu Mire, Latvia	56°86	26°88	Kuške et al. (2010b)
30	Svētiņu Lake, Latvia	56°45	27°09	Veski et al. (2012)
31	Priedaine site, Latvia	56°97	23°91	Cerina et al. (2010)
32	Eini site, Latvia	56°86	26°88	Kalniņa et al. (2004)
33	Viki Mire, Latvia	56°52	22°91	Kuške et al. (2010b)
34	Petrašiūnai, NE Lithuania	55°51	25°25	Stančikaitė et al. (2009)
35	Kašučiai Lake, western Lithuania	55°59	21°18	Stančikaitė et al. (2008)
36	Juodonys, Lithuania	55°44	25°26	Stančikaitė et al. (2009)
37	Hańcza, NE Poland	54°16	22°49	Lauterbach et al. (2011)
38	Pamerkiai, SE Lithuania	54°18	24°44	Stančikaitė et al. (2008)
39	L. Sumenko, northern Poland	54°11	17°48	Tylmann et al. (2011)
40	Biebrza Upper Basin, NE Poland	53°44	23°23	de Klerk et al. (2007)
41	Kurjanovas, Latvia	56°31	27°59	Heikkilä and Seppä (2010)
42	Zabieniec bog, central Poland	51°51	19°47	Płóciennik et al. (2011)
43	L. Perespilno, Poland	51°26	23°33	Goslar et al. (1999)
44	L. Słone, SE Poland	51°18	22°03	Kulesza et al. (2011)

(continued)

Table 2.1 (continued)

Site No.	Location	Lat (N)	Lon (E)	Source
45	No. 554, Gulf of Riga, water depth, 38.0 m; length of analysed sediments, 7.0 m	57°16	24°06	Kalniņa et al. (2012, 2003, 2012)
46	No. 556, Gulf of Riga, water depth, 32.4 m, length of analysed sediments, 3.55 m	57°06	23°29	Kalniņa et al. (2003)
47	No. 15, Gulf of Riga, water depth, 37.0 m; length of analysed sediments, 21.4 m	57°31	23°04	Kalniņa et al. (1999, 2003)
48	No. 989, Gulf of Riga, water depth, 49.5 m; length of analysed sediments, 5.1 m	57°47	23°26	Kalniņa et al. (1999, 2003)

northern Europe and established a relationship between their occurrence and mean summer and winter temperatures. This approach is still applied to quantitative palaeoclimatic reconstructions (e.g. Zagwijn 1994). The second method, developed by V.P. Grichuk from Iversen's method, is known as the mutual climatic range method (Grichuk et al. 1984). This consists of determining the modern ranges of climatic factors that permit the existence of all the species of plants identified in the composition of a given fossil flora, either from pollen or plant macrofossil data. For each plant species, the modern climate ranges are determined and the intersection of all climate ranges of all species present in a given fossil flora determines the climatic conditions that allowed the existence of all species identified in a given sample.

The third method was also developed by V.P. Grichuk. It consists of reconstructing the main climatic indices from fossil plant data using a concept developed by Szafer (1946). This author proposed to locate a modern analogue of a palaeoflora by comparing present-day ranges of plants. This method has been widely used to reconstruct the Lateglacial and Holocene climate in the Russian territory (e.g. Borisova 1990, 1997). The accuracy of this method for determining mean summer and winter temperatures has been estimated at about ±1 °C. For annual precipitation, the estimated accuracy is about ±50 mm (Grichuk et al. 1984), assuming no change in plant physiology. Other methods for climate reconstructions using pollen data have recently been developed. These are based on multi-variate statistical techniques. A number of different mathematical techniques, ranging from multiple regressions to correlation analysis, can be used to design transfer functions. The calibration of a statistical transfer function for pollen data requires a collection of modern surface pollen samples that are mathematically correlated to modern climate parameters. Transfer functions have been developed for different parts of the Baltic Sea basin and can be applied to climate reconstructions in the Lateglacial–Holocene period (Seppä 1996; Seppä and Birks 2002; Seppä et al. 2002a, b, 2004a, b, 2005, 2008; Birks 2003; Heikkilä and Seppä 2003, 2010; Antonsson 2006; Antonsson et al. 2006, 2008; Antonsson and Seppä 2007; Birks and Seppä 2010). A calibrated pollen–climate model has been recently developed to quantitatively reconstruct the Holocene annual mean, summer and winter air temperatures in northern Europe and in the Baltic Sea region. The model is based on modern pollen data sets from Finland, Estonia, and southern and central Sweden.

Common climatic fields and information-statistical methods have been successfully used to process fossil insect data (Coleoptera and Chironomidae) (Lemdahl 1991, 1997, 1998; Coope and Lemdahl 1995; Coope et al. 1998; Larocque et al. 2001; Walker 2001; Velle et al. 2005; Luoto 2009a, b; Olsson and Lemdahl 2009, 2010). Mean temperature estimates for the Lateglacial–Holocene transition were reconstructed based on coleopteran data from a number of sites in southern Sweden, central Poland and southern Finland (Lemdahl 1997; Coope et al. 1998; Lemdahl and Coope 2007). Continuous beetle records, covering almost the entire Holocene, have been obtained from sites in the uplands of southern Sweden (Olsson and Lemdahl 2009; Olsson et al. 2010) and at three alpine sites in the Abisko area, northern Sweden (Buckland et al. 2011). Fossil Coleoptera (beetles) can serve as a good proxy indicator of Lateglacial and Holocene climate changes in the Baltic Sea basin (Walker 2001). Many palaeoclimatic studies have been carried out in this region showing the presence of Coleoptera assemblages during interstadials (short warmer periods embedded in the glacial periods) that are more thermophilous than modern temperate, boreal or polar assemblages of beetle species. Beetle populations react rapidly to climate change and thus are indicative of contemporary climatic conditions. Coleoptera fauna have been widely employed to generate both temperature estimates and thermal gradients during the Lateglacial period in northern Europe. Chironomids (non-biting midges) are of special interest in palaeoclimatology because their larval head capsules remain well preserved in lake deposits. By using inference models that link present distribution and abundance of chironomids to contemporary climate, the past climate can be quantified from fossil assemblages. Chironomid populations respond rapidly to climate change and occur in a wide variety of modern environmental conditions (Brooks and Birks 2000, 2001;

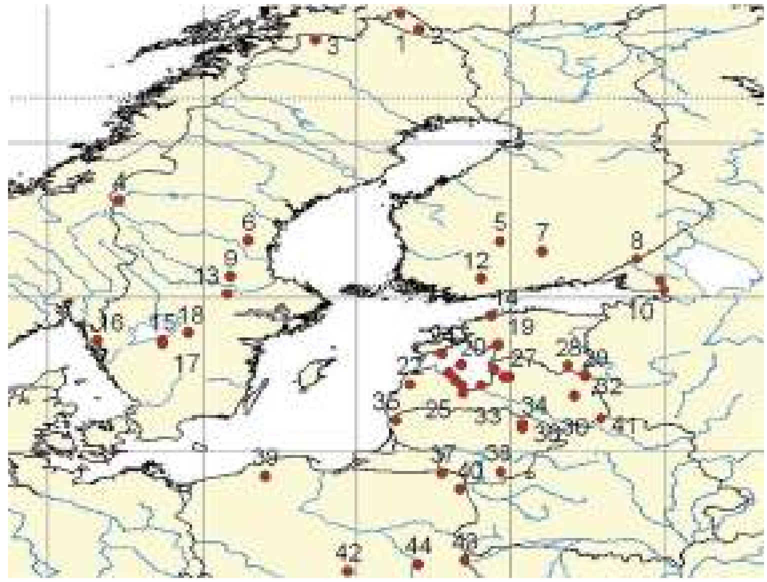

Fig. 2.3 Location of sites discussed in this study. Numbers refer to site numbers in Table 2.1

Larocque et al. 2001; Velle et al. 2005; Antonsson 2006; Luoto 2009a, b). Statistical comparison between modern chironomid assemblages and July temperature enables the calibration of statistical models and this provides a basis for quantitative estimates of summer temperature during the Lateglacial period (Velle et al. 2005; Antonsson 2006). Finally, by comparing pollen-inferred temperatures with independent proxy records (e.g. chironomids and oxygen isotopes), it is possible to create a more comprehensive picture of past climatic patterns in the Baltic Sea basin (Rosén et al. 2001).

Dendroclimatological data are more successfully used to reconstruct climate in those sites where climate is a limiting factor (such as at the latitudinal or altitudinal boundary of the forest). Various dendrochronological records are used as indicators of climate change and have been used to reconstruct temperature and humidity during the warm season. These comprise tree ring width, wood density, distribution of frost-damaged rings (Briffa et al. 2001), and stable isotopes of hydrogen, oxygen and carbon stored in cellulose. Analysis of radiocarbon (^{14}C) from tree rings has been used to establish high-resolution radiocarbon chronologies. The density of wood in tree rings determined by X-ray densitometry is a much more informative characteristic of past climate compared to the conventional data on tree ring width. In recent times, this technique has been widely accepted, and data on tree ring density of coniferous (predominantly pine and black spruce) and oak trees have been used to reconstruct the summer air temperature over the past millennia (Grudd 2008; Esper et al. 2012). Long Holocene chronologies consist of oak and pine tree ring data from the southern part of Germany and cover large portions of the Lateglacial period, extending into the Younger Dryas back to about 12,000 years ago (Friedrich et al. 1999), see also Chap. 3, Sect. 3.3.

The oxygen-isotopic palaeothermometry method is being widely applied not only to organic carbonates of marine origin but also to freshwater lake sediments (algae, lake marl), inorganic carbonate in caves (stalactites and stalagmites), and continental glaciers (mountain glaciers, polar ice sheets) (Mörner 1980; Seppä and Hammarlund 2000; Baldini et al. 2002; Hammarlund et al. 2002, 2003, 2005; Rasmussen et al. 2006; Kobashi et al. 2007, 2008).

Among other indications of past climate change in wide use is the diatomic analysis of lake sediments, data on varve width—which may depend on summer water temperature—and archaeological artefacts. Such artefacts (tools, land cultivation traces, ceramics) are primarily indicative of improved climatic conditions in the Baltic Sea basin, in particular increased summer and winter air temperatures and humidity (Pazdur 2004; Poska et al. 2004, 2008; Dolukhanov et al. 2009a, b, 2010).

To reconstruct past landscapes and climate, the entire set of available proxy evidence is used. This should render these reconstructions more robust. Table 2.1; Fig. 2.3 show key sections of the data used in this review.

2.4 Climate Variability During the Holocene Relevant for the Baltic Sea Basin

The modern Greenland Ice Core chronology (GRIP, NGRIP, and Dye-3) places the Younger Dryas /Holocene boundary at about 11,653 ice years ago (Alley 2000; Rasmussen et al. 2006; Walker et al. 2009). By analysing the ^{14}C isotope content in tree rings, this boundary has been dated to 11,573 cal year BP and by tree ring chronology at ~11,590 cal year BP (Kobashi et al. 2008). This boundary almost coincides with the final drainage of the Baltic Ice Lake at in the Billingen area (central Sweden).

2.4.1 Climate at the Boundary of the Younger Dryas/Holocene

Two rapid warming events dated ~14,700 and ~11,500 cal year BP were clearly greater than background climate variability (Mangerud 1987; Björck et al. 2001; Hoek 2001; Hoek and Bohncke 2001; Hoek and Bos 2007; Hoek et al. 2008; Lowe et al. 2008). Between these intervals, the climate varied between alternating centennial warm and cool phases. During a warm episode around ~14,700 cal year BP, air temperatures in ice-free regions of the Baltic Sea basin increased to values close to modern values. This can be inferred from data of changing vegetation (Hoek 2001) and is consistent with modelling results obtained by Renssen and Isarin (2001) with the general circulation model ECHAM4. In their study, ECHAM4 was driven by different reconstructions of sea surface temperature available at the time of publication, by changed orbital configuration and by reconstructions of vegetation cover. Summer air temperatures in ice-free regions reached 13–15 °C as indicated by pollen of *Hippophaë rhamnoides* (sea buckthorn) and *Typha latifolia* (cat's tail). The latest results show that during the Bølling warming 14,500 cal year BP (GI-1e), a treeless tundra community comprising the shrubs *Betula nana* (dwarf birch), *Dryas octopetala* (mountain avens) and *Salix polaris* (polar willow) thrived in the eastern Baltic Sea area (Veski et al. 2012). This warming was interrupted by a series of cold episodes: the Oldest Dryas (GS-2a), the Older Dryas (GI-1d) and the Younger Dryas (GS-1). During the latter period, the cooling lasted for 700–1000 years and the expansion of arboreal vegetation that had started during the Bølling (GI-1e) and Allerød (GI-1a-c) warming was interrupted. The vegetation cover was again replaced by tundra–steppe vegetation typical of the glacial period. During the warmest period of the GI-1a (Allerød) at 13,000–12,700 cal year BP, a pine forest mixed with deciduous trees of the species *Betula pendula* (silver birch) and *Populus tremula* (common aspen) developed in the eastern Baltic region (Veski et al. 2012).

During the Younger Dryas, the ice sheet still covered a considerable part of Fennoscandia, and sea level was 60–70 m lower than at present. The Baltic Sea basin was filled with freshwater and comprised the Baltic Ice Lake with a dry area where the modern Danish–Swedish straits are located and a land strip 50–80 km wide along the modern coast of the north German and Pomeranian lowlands. The coasts of the Gulfs of Riga and Finland and the Ladoga Lake basin, being subject to a strong glacio-isostatic depression, were submerged (Björck 1995; Mangerud et al. 2007). The cooling at that time has been correlated with the phases of glacial advance that resulted in the formation Salpausselkä I and II ice marginal formation, dated to 12,250 and 11,600 cal year BP, respectively (Rainio et al. 1995; Subetto et al. 2002; Rinterknecht et al. 2004, 2006; Subetto 2009). During the formation of the Salpausselkä marginal formation, the northern lowland part of the Karelian Isthmus formed the bottom of the Baltic Ice Lake, filling with its waters the Baltic and Ladoga basins (Subetto 2009).

The sediments of the periglacial lakes and the Baltic Ice Lake in the Gulf of Riga contain the lowest concentration of pollen and spores. At that time, the periglacial vegetation type emerged on weakly developed soils formed under conditions of severe Arctic-like climate. On the Karelian Isthmus, cold and dry climatic conditions and grass–bush (tundra–steppe) associations dominated until 11,000 cal year BP (Miettinen et al. 2007). The Younger Dryas (GS-1) cool phase strongly affected the floral composition of the vegetation. The pine forest collapsed within 100 years in eastern Latvia (Veski et al. 2012). The landscape again became treeless tundra as indicated by pollen spectra dominated by *Betula* sect. *albae, B. nana, Pinus, Salix, Artemisia,* Poaceae, Cyperaceae, Chenopodiaceae and *Dryas* (Ozola et al. 2010).

On the shores of the Baltic Ice Lake in the territories of modern Lithuania, Latvia and Estonia open tundra landscapes became widespread. Reconstructions based on the transfer function between modern climate and vegetation assemblage indicate that the mean annual temperature during the coldest period of the Younger Dryas (about 10,500–10,700 ^{14}C year BP) was approximately 6 ± 1 °C lower than the present mean annual temperature (Arslanov et al. 2001).

Analysis of relic cryogenic forms indicates that permafrost developed in an ice-free territory of the Baltic Sea basin in the Younger Dryas (Mangerud 1987; Isarin et al. 1997). In addition, there is also evidence of activation of aeolian processes (Isarin et al. 1997; Isarin and Bohncke 1999; Kasse 2002), changes in river channel morphology (Starkel 1999) and lake levels and composition of lake sediments (Subetto 2009).

Air temperature has been estimated by applying the method of arealograms based on palaeofloristic data (pollen and plant macrofossil data) (Borisova 1990, 1997). The results show that during the coldest phase of the Younger Dryas, deviations of the mean January temperature from the modern value were greatest in north-western Europe, that is, in the Baltic Sea basin. The January temperature there attained values 10–13 °C below the modern level. As at present, mean temperature decreased from west to east, from approximately −14 °C at the Jutland Peninsula to −20 °C on the coasts of the Gulf of Finland. Temperature deviations for the warmest month (July) from modern values were much smaller, about −2 °C on the southern coast of the Baltic Ice Lake. For most of this region, the July temperature in the Younger Dryas was about 13–14 °C (Borisova 1997).

In northern Germany, July temperatures were about 12 °C and in central Germany and Poland about 13 °C (Borisova 1990). In Finnish Karelia, they were 7–10 °C, in southern Sweden about 10 °C and in western Poland about 12 °C. In Poland, mean January temperatures were never above −20 °C (Walker 1995).

Analyses of the stable isotope content of Lake Gościąż sediments from the middle part of the Vistula River basin suggest that the July temperature was about 10–13 °C in the Younger Dryas (Starkel 2002). An independent reconstruction based on insect fauna composition (Chironomidae) indicates that the July temperature in south-western Norway near the south-western border of the Scandinavian ice sheet (Velle et al. 2005) was 5–6 °C compared to 11 °C at present. In south-eastern Karelia east of the Onega Lake (beyond the limits of the Baltic Sea basin), the minimum July temperature in the Younger Dryas has been estimated at 4 °C by pollen and macrofossils of *B. nana* versus a modern July temperature of 14 °C (Wohlfarth 1996; Wastegård et al. 2002; Wohlfarth et al. 2002, 2007).

Mörner (1980) was the first to quantitatively estimate air temperature changes at the Younger Dryas/Holocene boundary by oxygen isotope analysis of lake carbonates from southern Sweden. This data set showed a rise in air temperature of about 9 °C at the Younger Dryas/Holocene boundary. Independent estimates of the $^{29}N/^{28}N$ and $^{40}Ar/^{36}Ar$ ratio in Greenland ice cores showed a temperature increase at the Younger Dryas/Holocene boundary of 10 ± 4 °C and that this occurred within less than 50 years (Grachev and Severinghaus 2005).

The first warming in marine sediment (in the Gulf of Riga) is reflected by the dominance of *Pinus* pollen, the presence of *B. nana*-type pollen up to 20 %, and pollen and spores of the periglacial plants *D. octopetala, H. rhamnoides* and *Selaginella selaginoides* (Kalniņa et al. 1999).

The end of the Younger Dryas (11,700 cal year BP) is marked by the rise of *Betula pubescens/pendula* species and *Pinus* pollen curves, accompanied by the abundance of their macrofossils and by a decline of *B. nana, Artemisia*, Chenopodiaceae and *Juniperus* (core No. 554, No. 989 and No. 15 in the Gulf of Riga, Eini Lake, Svētiņu Lake) (see Table 2.1) Within grass pollen, concentrations of Poaceae and Cyperaceae pollen also increased at this time (Kalniņa et al. 1999, 2011, 2012).

2.4.2 Early Holocene Oscillations

In the Early Holocene, a cold and relatively dry climate became suddenly warmer and more humid. The summer temperature in north-western Russia increased from 4 to 10–12 °C (Wohlfarth et al. 2007). In the very first warm phase of the Preboreal around 11,530–11,500 cal year BP, the arboreal vegetation began spreading rapidly in ice-free regions as a result of an abrupt warming (within 50 years or less) and tundra–steppe vegetation dominated by shrubs and grass typical of the Younger Dryas was replaced by open forest associations (Bos et al. 2007). In the earliest warm phase of the Preboreal period, birch and pine started to spread, the former more vigorously than the latter. Periglacial vegetation still occupied large areas in some parts of the Baltic Sea basin. During the initial phase of this warming, the broad-leaved trees appeared in ice-free regions, where *Tilia, Ulmus, Corylus* and *Fraxinus* pollen was first found.

A shift from clastic–detrital deposition to an autochthonous sedimentation dominated by biochemical calcite precipitation was identified in the Early Holocene by accelerator mass spectrometer (AMS) ^{14}C dating ~11,600 cal year BP of Lake Hańcza (north-eastern Poland) sediments (Lauterbach et al. 2011). The accumulation of silt and clay was replaced by organic-rich gyttja at 11,650 cal year BP also in eastern Latvia (Veski et al. 2012).

Within the territory of the present Lithuania, a gradual amelioration of environmental conditions started at about 11,500 cal BP. However, there (Stančikaitė et al. 2004, 2008) and in north-western Russia (Wohlfarth et al. 2007) the change was less pronounced than that recorded in the North Atlantic region (Björck et al. 1996). The early expansion of *Picea* within the local vegetation, even before 11,500 cal BP, is worth noting. Despite a prolonged discussion involving pollen data, macrofossil finds and modern genetic information, the Lateglacial and Early Holocene history of *Picea* in this part of Europe is still under debate (Giesecke and Bennett 2004; Latałowa and van der Knaap 2006). Identification of *Picea* pollen and macrofossil finds suggests the local presence of this tree species during the earliest stages of the Holocene in the north-eastern and northern Lithuania (Kabailienė 1993; Kabailienė et al. 2009; Stančikaitė et al. 2009; Gaidamavičius et al. 2011). Recent evidence of stomata, needles and wood suggests that spruce populations expanded into Latvia during the Younger Dryas (Koff and Terasmaa 2011; Veski et al. 2012).

Moreover, the presence of *Picea abies* suggests a relatively warm climate in this region, with warm summers (~10–13 °C) and moist soil conditions (Giesecke and Bennett 2004) shortly before 11,500 cal year BP. However, changes in the vegetation composition point to the presence of at least two short cold climate episodes between 11,500 and 11,100 cal year BP in north-eastern Lithuania (Stančikaitė et al. 2009).

In Latvia, the earliest warm phase of the Preboreal period is marked by the rise of birch and pine curves in pollen diagrams. However, in several diagrams, especially those from the Gulf of Riga (No. 989, No. 15), pollen and spores of periglacial plant (*B. nana, D. octopetala*) types are still present, albeit more rarely. Their values fluctuate strongly and display a marked spatial variability. Such fluctuations may suggest some climate instability. The pollen spectra characteristic of the Preboreal (PB) can be subdivided into two parts: those corresponding to the Early Preboreal (PB1), characterised by a small increase in *Betula* along with some increase in different grass pollen; and those spectra, mostly found in the second part of the Preboreal (PB2), where the typical Preboreal pollen is much reduced or completely missing. This can be explained by erosion of the sediment or by much reduced sedimentation rates during the Yoldia Sea stage (core No. 15, see Table 2.1).

The warming at the Holocene boundary, at about 11,530 and 11,500 cal year BP was interrupted by a short cold Preboreal oscillation dated using ice core data to 11,430–11,270 cal year BP (Rasmussen et al. 2006; Kobashi et al. 2008). A relatively short cooling phase (about 200–250 years long) occurred approximately 250 years after the final drainage of the Baltic Ice Lake, when seawater flowed into the Baltic Sea basin through the Närke Strait in Billingen due to a rapid rise in sea level resulting from the melting of the Scandinavian ice sheet (see Fig. 2.4). Brackish water entered the Yoldia Sea along the southern coast of the Gulf of Finland and freshwater flowed southwards from the melting ice sheet (Yu 2003).

The coldest part of the Preboreal oscillation is dated to ~11,430–11,350 cal year BP. The Preboreal cool oscillation almost coincides with the short brackish phase of the Yoldia Sea whose end has been dated to 11,200 cal year BP (Heinsalu and Veski 2007), which corresponds to the age 11,190 year BP in the GRIP ice core (Bjorck 1999).

Fig. 2.4 The configuration of the Yoldia Sea stage at the end of the brackish phase at 11,100 cal year BP (Andrén et al. 2011)

At the start of the Preboreal cooling phase, sea level was approximately 50 m lower than at present. Subsequent melting of the continental ice sheets (Laurentide and Scandinavian) caused an increase in ocean volume and sea level rose.

Pollen stratigraphy clearly suggests some climate instability in the Early Holocene. During its coldest phase, forest tundra and open forest landscapes were established over the greater part of the Baltic Sea basin. This vegetation type remained in some regions until 10,700–10,600 cal year BP. Pazdur (2004) characterised the Preboreal climatic conditions in Poland as cold and dry. At this time, birch expansion that had started during the warming at the beginning of the Early Holocene was interrupted by a dry continental phase characterised by open grassland vegetation (Rammelbeek Phase) (Bos et al. 2007).

Within the territory of the present Lithuania, the most prominent climate cooling recorded shortly before 11,100 cal year BP may be correlated with the climate event termed the Preboreal Oscillation (PBO) ca. 11,300–11,150 cal BP, described as a humid and cool interval in north-western and central Europe. Only after 11,100 cal year BP does ongoing forestation of the territory by open forest dominated by birch and pine suggest a climatic improvement that could be interpreted as a delayed Pleistocene/Holocene warming.

At the beginning of the Late Preboreal time, between 11,270 and 11,210 cal year BP, there was a sudden shift to a warmer and more humid climate, and forest vegetation expanded once more. Expansion of pine occurred in the later part of the Late Preboreal. This can also be inferred from the increase in organic matter in lake sediments from central Latvia after 11,200 cal year BP, likely to be due to the reduced inflow of minerogenic matter in a denser forest (Puusepp and Kangur 2010).

As indicated by the Greenland ice core GISP2 isotope data, at the Preboreal/Boreal (PB/BO) boundary air temperatures increased by 4 ± 1.5 °C (Grachev and Severinghaus 2005). At this boundary between 10,770 and 10,700 cal year BP, dense pine woodland started to expand in the southern part of the Baltic Sea basin (Bos et al. 2007). In north-western Russia, boreal forest comprising *Pinus, Picea, Betula, Alnus incana* was present at lower altitudes (Subetto et al. 2002).

At that time, open landscapes typical of the Preboreal cooling changed to poplar–pine–birch closed vegetation in southern Finland. From 11,000 cal year BP onwards, the glacial boundary retreated fast in the south of the modern boreal belt in Finland, and at about 10,700 cal year BP the

Fig. 2.5 The configuration of the Ancylus Lake stage during the maximum transgression at ca. 10.5 cal year BP (Andrén et al. 2011)

region was free of ice, although the ice boundary was still very close. According to estimates made by (Heikkilä and Seppä 2003), between 10,700 and 10,500 cal year BP, the mean annual temperatures were about −3.0 to 0 °C. These temperatures correspond to the modern temperature at 70°N.

This climate warming caused the rapid melting of the continental ice sheets. The subsequent isostatic uplift cut-off the Baltic Sea basin from the Atlantic Ocean and lead to its conversion into a large freshwater basin (Ancylus Lake). The culmination of the transgression phase of this lake is dated to about 10,700 cal year BP (Yu 2003). The Ancylus Lake stage lasted until approximately 9500 cal year BP. The contemporary Baltic Sea basin is part of the last Littorina Stage when significant water level and salinity variations took place (see Fig. 2.5).

Beetle data from southern Sweden dated to between ca. 11,000 and 9000 cal year BP (Olsson and Lemdahl 2010; Olsson et al. 2010) suggest temperatures of the warmest month similar to present values. The relatively low numbers of aquatic species imply relatively dry and continental conditions. Reconstructed temperatures based on coleopteran data from Abisko are also similar to present temperatures (Buckland et al. 2011).

A temperature rise started after 10,000 cal year BP and lasted until the beginning of the cold event at 8200 cal year BP. Over the greater part of southern Sweden, Estonia, Latvia and Lithuania, open boreal woodlands and sparse birch vegetation were established. Summer air temperatures rose by 7–10 °C above the level attained during the Younger Dryas. However, major environmental changes in the area had begun already in 10,300–10,200 cal year BP, when deciduous trees including thermophilous species spread into the region.

According to data from Lake Hańcza in north-eastern Poland (11,600 cal year BP), the shrub pollen content of lake sediments decreased and a shift from clastic–detrital lake deposits to autochthonous sediments dominated by biochemical calcite occurred at the beginning of the Holocene. Warmer conditions ensued, and between 10,000 and 9000 cal year BP pollen spectra showed an increased content of pollen from broad-leaved trees. The organic content of the lake sediments also increased (Lauterbach et al. 2011). Figure 2.6 shows the reconstructed annual mean air temperature for south-central Sweden based on pollen data from Lake Flarken (Seppä et al. 2005).

The transition from the Ancylus Lake to the Littorina Sea in the Baltic Sea basin is marked by evidence of a weak

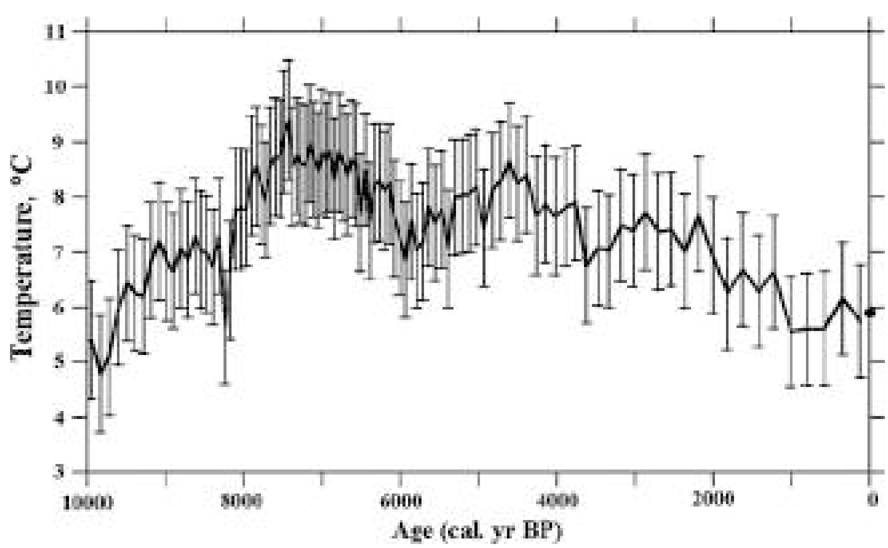

Fig. 2.6 Mean annual temperature reconstructed from pollen data from Lake Flarken (south-central Sweden) during the past 10,000 years (*black line*). Present-day mean annual temperature (5.9 °C) is marked by the point, modified from Seppä et al. (2005)

brackish phase between 9800 and 8500 cal year BP, called the Early Littorina Sea (Andrén et al. 2011). The ensuing brackish–marine stage is defined by evidence of significantly increased seawater influx (Hyvärinen et al. 1988) and an opening of the Öresund Strait, an event dated to around 8500 cal year BP (Björck 1995; Yu 2003).

Between 11,000 and 9500 cal year BP, air temperatures in the Baltic Sea basin slowly raised reaching values of ~0.5 °C lower than at present. A prominent warming started about 9000 cal year BP and lasted until the Holocene thermal maximum (HTM) about 8000–4500 cal year BP, with summer air temperatures 2.5–3.5 °C above modern levels. Davis et al. (2003) reconstructed quantitative characteristics of the climate for the past 12,000 years in western Europe based on analyses of pollen data from more than 500 sections. In northern Scandinavia, summer air temperatures close to the modern levels have been inferred from pollen data (Seppä and Birks 2001) and independent estimates based on chironomid and diatom assemblages (Rosén et al. 2001).

The temperature estimates based on macrofossils (for example, tree-line position) indicate a rapid and stable warming after 10,000 cal year BP across the entire northern Scandinavia (Seppä and Birks 2001). Climate reconstructions for central Sweden using pollen spectra from Lake Gilltjärnen sediments show a rapid increase in summer temperatures from 10.0 to 12.0 °C between 10,700 and 9000 cal year BP. This stable positive trend remained until the sudden cooling at about 8200 cal year BP (Antonsson et al. 2006).

2.4.3 The ‘8.2 ka Cold Event’

The ‘8.2 ka cold event’ was first identified from changes in oxygen isotope composition in ice cores from the Summit site in Greenland. The decrease in air temperature during this event has been estimated at 6 ± 2 °C in central Greenland (Allen et al. 2007; Alley et al. 1997; Muscheler et al. 2004; Alley and Ágústsdóttir 2005; Rohling and Pälike 2005; Thomas et al. 2007). An independent method to estimate both air temperature change and the duration of this cold phase, based on the concentrations of $\delta^{15}N$ in nitrogen (N_2) in ice cores (Leuenberger et al. 1999; Kobashi et al. 2007; Thomas et al. 2007). Kobashi et al. (2007) from the GISP2 ice core data, with a time resolution of ~10 year, showed a complex time evolution of this event.

The duration of the entire cold event was about 160.5 ± 5.5 years, the coldest phase occurring at 69 ± 2 years (Thomas et al. 2007). A drop in air temperature of 3 ± 1.1 °C occurred within less than 20 years. Independent estimates of changes in air temperature during this cooling event have been obtained from isotope data from four Greenland ice cores: NGRIP, GRIP, GISP2, and Dye-3 (Thomas et al. 2007). According to the GRIP core data, the event has been dated to ca. 8190 ice years ago. The data from the GISP2 core revealed two milder cool stages: ca. 8220 and ca. 8160 ice years ago. The initial stage of cooling is more explicitly recorded in the NGRIP core data.

Independent empirical data, such as marine sediment cores with high resolution, lake sediments, clay varves, speleothem isotope data, and pollen diagrams among others, indicate that the cooling was widespread throughout the entire Baltic Sea basin, except for the most northern regions north of 70°N. The high-altitude glaciation intensified across all regions. In central and southern Norway, the snow line (equilibrium line altitude) lowered by 200 m (Nesje and Dahl 2001; Nesje 2009). The altitude of the upper boundary of arboreal vegetation also lowered (Seppä and Birks 2001; Seppä et al. 2007, 2008).

Pollen diagrams from lake and continental sediments provide independent evidence of this cooling. Davis et al. (2003) clearly identified the 8.2 ka cold event across the whole of western Europe, with the lowest air temperatures in areas adjacent to the North Atlantic Ocean (Seppä et al. 2008; Rasmussen et al. 2009).

Pollen data from Estonia and southern Fennoscandia indicate a 1 °C air temperature drop (Fig. 2.7). At Lake Rõuge in Estonia, air temperature decreased by 1.8 °C (Veski et al. 2004), which agrees with estimates from pollen data from lakes in different part of Sweden (Lakes Giltjärnen, Flarken and Trehörningen) (see Fig. 2.8).

Lake sediments from four sites—Lake Rõuge (Estonia), Lake Flarken (Sweden), Lake Arapisto (Finland) and Lake Laihalampi (Finland), all located south of 61° N—indicate that the pollen percentage for the thermophilous deciduous tree taxa *Corylus* and *Ulmus,* decreased from 10 to 15 % in the Early Holocene to 5 % between 8250 and 8050 cal year BP (Seppä et al. 2007). This implies a temperature drop of 0.5–1.5 °C at all these sites. Here, the cold event lasted 200–300 years and ended by a sudden temperature rise. In Estonia, the cold event lasted longer as indicated by a longer period of vegetation re-establishment after a decrease in annual air temperature of at least 1.5–2.0 °C (Seppä and Poska 2004; Veski et al. 2004).

The 8.2 ka cold event has been identified in pollen diagrams from lake sediments from eastern Latvia (Eini, Malmuta) and the Gulf of Riga displaying a decrease in *Alnus*, *Corylus* and also *Ulmus,* and some increase in grass pollen. According to Heikkilä and Seppä (2010), a relatively warm and stable climate was interrupted by cooling at about 8350–8150 cal year BP when mean annual air temperatures dropped by 0.9–1.8 °C. This was reflected in a change in vegetation composition, characterised by a decrease in broad-leaved tree productivity and an increase in the pollen of boreal species (Heikkilä and Seppä 2010). The limit of

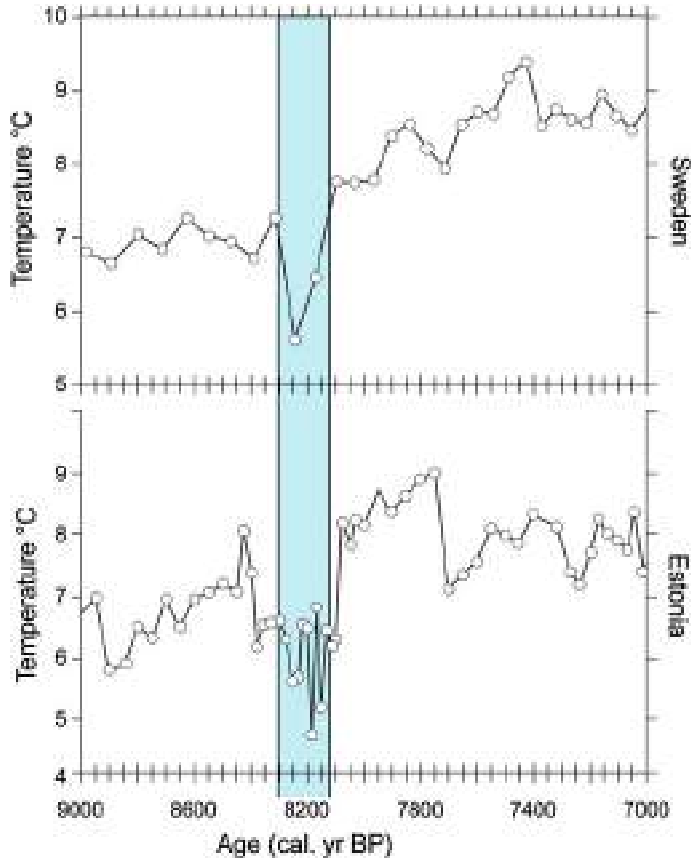

Fig. 2.7 Two climate reconstructions for the cold event about 8200 cal year BP: annual air temperature reconstructed from pollen data from Lake Rõuge, Estonia (Veski et al. 2004) and pollen data from Lake Flarken, Sweden. The *blue bar* highlights the timing of the '8.2 ka cold event' (adapted from Seppä et al. 2005)

detection for *Ulmus* and *Tilia* pollen is reached in the second part of the Boreal period when their pollen almost disappears before its contribution to the pollen assemblages starts rising again at beginning of the Holocene Thermal Maximum.

Re-establishment of *Picea* in the eastern Baltic Sea region (within the territory of the present Lithuania) is dated to 8600–8000 cal year BP. This short climatic deterioration may have limited the expansion of deciduous trees, providing more space for the expansion of spruce. The expansion of spruce may have been a response to a wet and cool climate with colder and snowier winters (Seppä and Poska 2004).

A section containing about 400 layers of varved sediments in west-central Sweden, supported by radiocarbon dating and by dendrochronological data, provides a detailed account of environmental development in this region, identifying an analogue of the 8.2 ka cold event between 8066 ± 25 cal year BP and 7920 ± 25 cal year BP. Snow precipitation increased during this cold period. The estimated duration of the 8.2 ka cold event has been estimated at 150 years (Snowball et al. 2002, 2010; Zillén et al. 2008; Zillén and Snowball 2009). There are no clear changes in the beetle assemblages from southern or northern Sweden that would indicate a cold spell during this time. This may be due to the insufficient time resolution of the data to identify this change.

In Germany, changes in the isotope composition of ostracod valve carbonate suggest this cold event lasts for about 200 years (von Grafenstein et al. 1998). Almost all data indicate that the decrease in air temperature was significantly greater in winter than in summer. This promoted an earlier freezing and later thawing of sea and lake surfaces.

The causes of the cooling during the 8.2 ka cold event and other cold episodes in the Lateglacial/Early Holocene drew the early interest of researchers (Alley et al. 1997; Barber et al. 1999; Leuenberger et al. 1999; Marotzke 2000; Clark 2001; Clark et al. 2002; Clarke et al. 2004; Alley and Ágústsdóttir 2005; Denton et al. 2005; Kobashi et al. 2007; Thomas et al. 2007; Yu et al. 2010). Although some scientists attribute the 8.2 ka cold event to changes in incoming solar radiation (Bos et al. 2007), most studies relate this cooling and other cooling events in the past 13,000 years to changes in the circulation of surface and deep water in the North Atlantic Ocean driven by melt water from the continental ice sheets.

Clark (2001) and Clark et al. (2002) assumed that freshening of the sea surface layer of the North Atlantic Ocean not only disturbed the circulation in the surface layer but also hindered the formation of deep water, thus affecting the intensity and position of the Atlantic 'conveyor belt' itself. Drainage of glacial lakes Agassiz and Ojibway due to

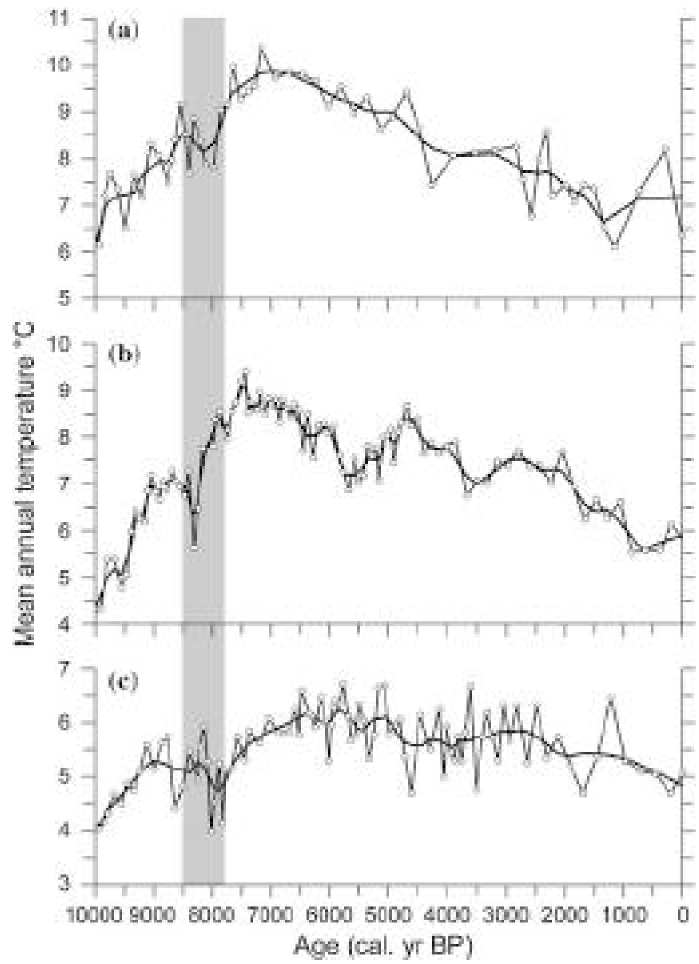

Fig. 2.8 Quantitative reconstructions of mean annual air temperature from pollen data for three lakes (C-Gilltjärnen, B-Flarken, A-Trehörningen). The *grey bar* highlights the timing of the '8.2 ka cold event' (adapted from Antonsson and Seppä 2007)

the melting of the Laurentide Ice Sheet ca. 8470 cal year BP (~7700 ^{14}C year BP), during which about 2×10^{14} m^3 fresh lake water could have been released within less than 100 years, could have exerted a serious impact on the formation of sea ice, and thus significantly changing the timing of its formation in autumn and melting in spring. Sea ice has a higher albedo than open water, and a longer period of sea ice would cause an additional cooling (Barber et al. 1999; Clark 2001; Ganopolski and Rahmstorf, 2001; Fisher et al. 2002; Clarke et al. 2004; Wiersma and Renssen 2006; Widerlund and Andersson 2011).

2.4.4 The Holocene Thermal Maximum

Pollen and chironomid records show that the period between 7500 and 5500 cal year BP was the warmest for the entire Baltic Sea basin area, although temperatures peaked at different times in different parts of the Baltic Sea basin, with amplitude varying from 0.8 to 1.5 °C and higher. The cold event at about 8.2 ka BP was followed by a warm period with air temperature and annual precipitation higher than present. At that time, forest vegetation with thermophilous species flourished. In northern Finland, the warmest climate of the Holocene occurred between 8200 and 5700 cal year BP, with a temperature maximum at ca. 7950–6750 cal year BP. Pollen diagrams of Lake Tsuolbmajavri, northern Finland, suggest July temperatures around 13 °C. They dropped by 1 °C only after about 5750 cal year BP (see Fig. 2.9). According to the increasing *Pinus sylvestris* pollen percentages and by the decrease of lycopod and fern spore values, the mean annual sum of precipitation must have rapidly fallen to below 500 mm year^{-1} ca. 6650–5700 cal year BP (Seppä and Birks 2001), due to still unknown causes.

In the north-western part of Russia, forest vegetation with spruce and deciduous trees predominated (*Ulmus, Tilia, Quercus*). Between 8000 and 4500 cal year BP, summer temperatures were 2.0–2.5 °C and precipitation was 100–150 mm year^{-1} higher than present (Arslanov et al. 1999, 2001). On the Kola Peninsula, the warm period occurred between ca. 8000 and 4200 cal year BP with summer temperatures peaking around 6000–5000 cal year BP when July temperature was 13.6 °C (Solovieva et al. 2005).

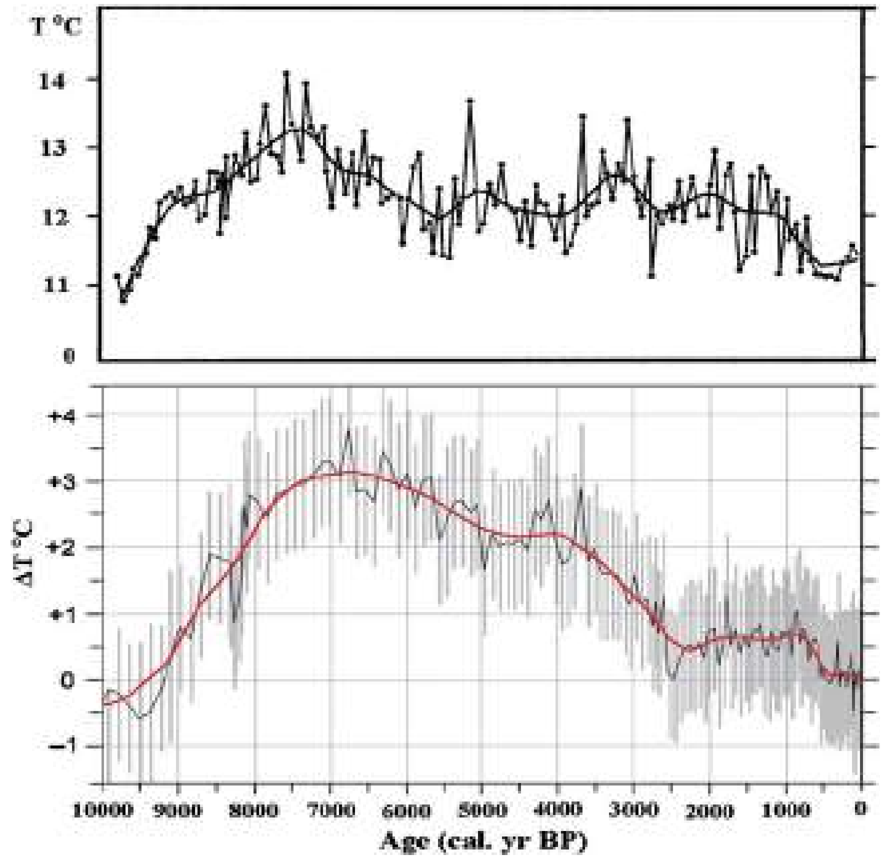

Fig. 2.9 Reconstructed summer temperature in the northern and southern parts of the Baltic Sea basin. *Upper panel* The mean summer (July) air temperature reconstructed from pollen data from Lake Tsuolbmajavri, northern Finland 68° 41′N; *Lower panel* Reconstructed summer temperature anomalies shown as deviations from the modern reconstructed value. Lake Kurjanovas, south-eastern Latvia, 56° 31′N (adapted from Seppä and Birks 2001 and Heikkilä and Seppä 2010)

Giesecke et al. (2008) reconstructed winter (January) temperatures for the past 11,000 years using pollen data from three lakes (two in central and western Sweden and one in southern Finland) (Fig. 2.10). The highest winter temperatures were found to have occurred between 7000 and 6000 cal year BP. Quantitative reconstructions indicate that the climate in Fennoscandia has become increasingly continental over the past 7000 years, and mostly reflecting winter cooling. These reconstructions suggest that the Early Holocene in Fennoscandia was the most oceanic period, but probably with higher temperature variability. Millennial reconstructions of the North Atlantic Oscillation Index since the Holocene Thermal Maximum until present based on Greenland ice core data suggest a long-term negative trend, and thus increased continentality (Olsen et al. 2012). This is supported by a simulation with the coupled atmosphere–ocean model ECHO-G driven by changes in orbital configuration (Hünicke et al. 2010).

Reconstructions of mean annual air temperature by pollen data from Lake Trehörningen in south-western Sweden show that the maximum temperatures 2.5–3.0 °C above the modern values occurred between 7000 and 4000 cal year BP. In northern and eastern Scandinavia, mid-Holocene temperatures were somewhat lower, about 1.5–2.5 °C above present-day values (Antonsson and Seppä 2007). The presence of a number of beetle species during the Mid-Holocene that today are found in central Europe suggests higher temperatures than present in southern Sweden during that period. The mutual climate range (MCR) reconstructions based on stenothermic beetles from Abisko (Buckland et al. 2011) indicate a climatic optimum between 7500 and 6000 cal year BP, with mean summer temperatures about 3 °C higher than at present.

In central Sweden, thermophilous vegetation (oak and linden) expanded after 7000 cal year BP. This type of vegetation requires high summer temperatures and is relatively resistant to low winter temperatures. During the warmest period in the Holocene (between 7000 and 4000 cal year BP), the range of *Tilia* advanced 300 km north compared to its modern position. Over that period, precipitation decreased and lake levels dropped. Near Lake Gilltjärnen (60°N), a maximum mean annual temperature of 6.0–7.0 °C occurred at about 6500 cal year BP, compared to 5.0 °C at present (Antonsson et al. 2006). Reconstructions of summer air temperature in central Sweden using fossil Chironomidae composition show that these temperatures varied within wide limits, from −0.8 to +0.8 °C of its modern values (Brooks and Birks 2000, 2001; Velle et al. 2005). In Estonia, *Ulmus* and *Tilia* spread most widely between 7000 and 5000 cal year BP, which coincides with the maximum in

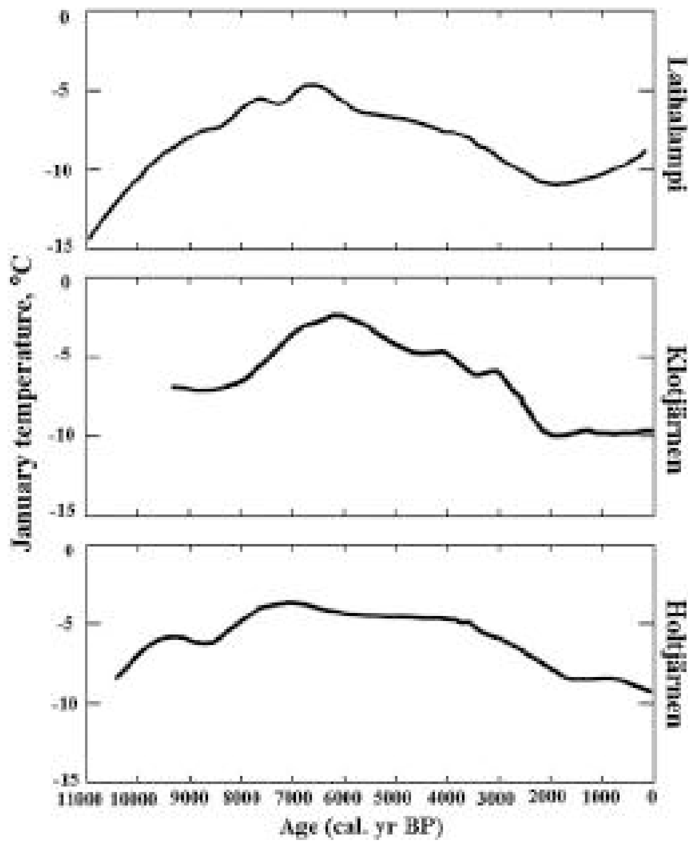

Fig. 2.10 The winter (January) air temperature in parts of the Baltic Sea basin over the past 11,000 years, obtained by pollen from three lakes: Laihalampi 61°29N (southern Finland), Klotjärnen 61°49N (Sweden), Holtjärnen, 60°39N (Sweden) (adapted from Giesecke et al. 2008)

summer air temperature. At that time, these species occupied about 40 % of the Estonian territory compared to 1 % at present (Saarse and Veski 2001).

Reconstructions based on sediment data from lakes Holzmaar and Meerfelder in the Westeifel Volcanic Field (Germany) show that temperatures were 1 °C higher than at present and declined after 5000 cal year BP (Litt et al. 2009).

In Latvia, the warmest climate of the Holocene occurred between about 8200 and 5700 cal year BP with temperatures more than ~2.5–3.5 °C above modern values and a peak warming of about 3.0–3.7 °C at ca. 7950–6750 cal year BP (Heikkilä and Seppä 2010). Temperatures subsequently declined towards present-day values (Figs. 2.8 and 2.9). Deciduous forest appeared here later than in Estonia. *Ulmus* started to spread in the first half of the boreal period and *Tilia* at about 7000–5000 cal year BP (Heikkilä and Seppä 2010). With the beginning of the Atlantic warming, mixed deciduous forest composed of *Ulmus, Tilia, Alnus, Corylus*, and *Picea* expanded (Mūrniece et al. 1999; Ozola et al. 2010). The maximum values of *Ulmus* and *Tilia* pollen are found in the deposits formed in the first part of the Atlantic warming, but the maxima of the *Quercus* pollen occur in the later period. A short climate deterioration in the central part of the Holocene Thermal Maximum has been identified in pollen diagrams from Eini Lake, Malmuta River mouth (Segliņš et al. 1999) and the Gulf of Riga. The Atlantic warming is clearly identified in pollen diagrams from lagoon sediments. During the second part of the Atlantic period, water level in the lagoon basins gradually decreased. The lagoons gradually filled up and lake sediments record the presence of fen peat with a dominance of *Carex* species, such as *Carex elata, C. lasiocarpa, C. teretiuscula, C. aproximate* and *Hypnum* (Cerina et al. 2010).

In Lithuania, the major environmental changes started after 10,300–10,200 cal year BP when deciduous trees, including thermophilous species, spread in this region. In eastern Lithuania, the expanding deciduous taxa, such as *Corylus* (ca. 10,200–10,000 cal year BP), *Alnus* (8200–8000 cal year BP), and broad-leaved species such as *Ulmus* (ca. 10,000 cal year BP), *Tilia* (7700–7400 cal year BP) and *Quercus* (5200 cal year BP), formed a dense mixed forest. *Picea* reappeared at 7300–6800 cal year BP (Gaidamavičius et al. 2011). In north-western Lithuania, immigration of *Corylus* was dated back to 7600–7200 cal year BP, *Alnus*—back to 7300–6900 cal year BP, and *Ulmus*—back to 8100–7500 cal year BP (Stančikaitė et al. 2006). After about 9800–9900 cal year BP, the expansion of deciduous species was accompanied by the spread of *Cladium mariscus* (L.) Pohl, suggesting that minimum mean July temperature increased up to +15.5 °C even in the northern part of the country. Development of deciduous forest requires moist and fertile soil and high humidity. Therefore, its expansion confirmed that such ecological conditions pre-existed there. The Holocene Thermal Maximum, ca. 8000–4500 cal year BP, with dryer and warmer summers reflecting stronger continentality (Seppä and Poska 2004), was generally responsible for expansion of deciduous trees to the eastern Baltic Sea region. These taxa continued to flourish there between about 7500 and 4200 cal year BP (Stančikaitė et al. 2003, 2004, 2006, 2008, 2009; Giesecke and Bennett 2004; Giesecke et al. 2008; Gaidamavičius et al. 2011).

In central Poland, *Tilia* and *Quercus* appeared about 9300 cal year BP. However, they did not become widespread until 8000–7000 cal year BP. At the same time, *Alnus*—and in mountainous regions, *Picea*—spread widely. Then, at about 5000 cal year BP, the climate became wetter and colder (Pazdur 2004; Lauterbach et al. 2011). It should be noted that the first signs of anthropogenic deforestation are recorded at 6500 cal year BP in this region.

For northern Europe as a whole, Davis et al. (2003) dated the well-defined thermal maximum of the Holocene at ca. 6000 cal year BP. For the Baltic Sea region, the highest temperatures occurred around 6500 cal year BP and were 1.5–2.5 °C above present-day values in the north-west area and 1.0–1.5 °C in the north-east.

2.4.5 Late Holocene Cooling

As is clear from Figs. 2.7, 2.8 and 2.9, climate has been relatively unstable across the Baltic Sea basin over the past 4500 years, with alternating warm and cold periods superimposed on a general cooling trend.

Different proxy data indicate a two-stage air temperature decrease in the Late Holocene. The first occurred between 5000 and 4500 cal year BP and the second between 4300 and 3300 cal year BP. During each of these periods, the temperature drop was at least 1 °C. A warming at ca. 3200 and cooling at ca. 2800 cal year BP are revealed by detailed palaeoclimatic reconstructions (Zubakov and Borzenkova 1990; Arslanov et al. 2001; Bjune et al. 2009). Although the general trend of the Late Holocene cooling is probably related to decreased summer solar radiation due to astronomical factors, the causes of the superimposed oscillations are still unclear.

Climate cooling in north-western and northern Russia and other regions of the Baltic Sea basin started at about 4500 cal year BP. A warmer period at about 3500 cal year BP and a colder period at about 2500 cal year BP are apparent, superimposed on the general cooling trend (Arslanov et al. 2001; Subetto et al. 2006).

Analyses of Lake Igelsjön sediments in southern Sweden suggest that relatively warm conditions remained until 4700 cal year BP, followed by an abrupt hydrological shift to cooler and/or wetter conditions by around 4000 cal year BP (Jessen et al. 2005). A cooling trend thereafter, up to pre-industrial times, was punctuated by a series of short rapid warm fluctuations. Pollen diagrams show changes in vegetation composition and a distinct decrease in thermophilous species (primarily *Corylus*). About 2000 cal year BP, the temperatures were already close to modern values (Jessen et al. 2005).

Proxy data reconstructions show that a trend towards cooler and wetter conditions prevailed across Fennoscandia during the past 2000–1500 years (Bjune et al. 2009; Esper et al. 2012). These climate trends are shown by, for example, $\delta^{18}O$ records from lake sediments (Hammarlund et al. 2003; Rosqvist et al. 2007), tree ring-inferred summer temperature and pollen data. The mean July temperature reconstructions based on these data obtained from 11 lakes located in the different landscape zones in the northern parts of Norway, Sweden, Finland, and north-western Russia are shown in Fig. 2.11.

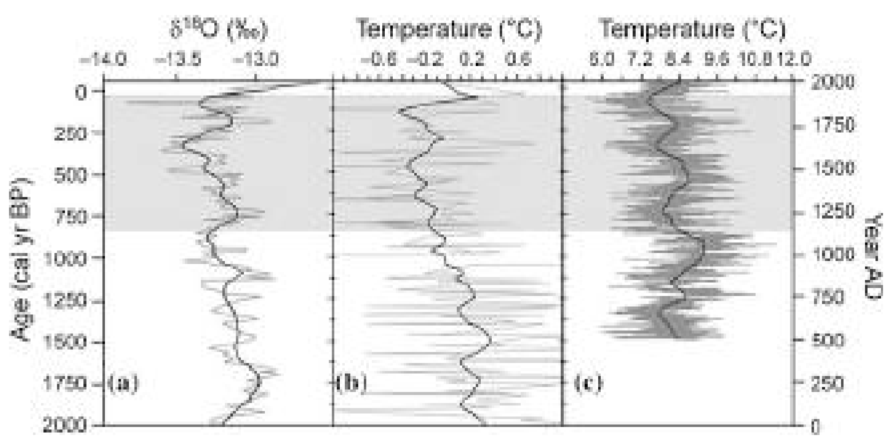

Fig. 2.11 Three proxy climate records covering the past 2000 years from northern Fennoscandia: **a** $\delta^{18}O$ from calcareous lacustrine sediments from Abisko, Sweden (Rosqvist et al. 2007); **b** pollen-inferred mean July temperature deviations from 11 lakes. **c** tree ring-inferred summer (April–August) temperature from the Torneträsk region, Sweden (Grudd 2008), (Bjune et al. 2009)

In Poland, analyses of lake sediments of the Late Holocene, including pollen spectra, indicate that lake levels rose and peat mires developed at this time. Between 5000 and 4200 cal year BP, the climate was relatively warm and dry, but by the end of this period became more humid. The dominant species in forest associations changed slightly. In the mountains, spruce expanded and *Carpinus* and *Fagus* migrated from the south-east into this region. At about 3500 cal year BP, *Abies* spread from the south-west and became a forest-forming species in southern Poland (Kulesza et al. 2011). *Picea* spread to the north-east of Poland and *Fagus* penetrated this region from the west. Pazdur (2004) showed that the Subatlantic period started about 2800 cal year BP.

According to Heikkilä and Seppä (2010), summer air temperature dropped by ~3°C in Latvia during the Late Holocene. Pollen diagrams show changes in vegetation; a significant decrease and almost disappearance of the thermophilous broad-leaved species and *Corylus*. The dominant tree species over the past 3000 years was *Pinus* and *Betula* (Mūrniece et al. 1999; Kalniņa 2008; Kuške et al. 2010b), except in the central part of the Subatlantic period when the *Picea* pollen curve, and to a lesser extent *Alnus* and *Quercus,* rose again. Dwarf shrub (mainly Ericales) and grass pollen (Poaceae, Cyperaceae) concentrations increased during the past 1000 years, reflecting a decrease in forest cover, which resulted from both climatic and human impacts (Ozola et al. 2010; Kalniņa et al. 2012).

Analyses of isotopic strontium concentrations in molluscs can give information about past salinity in the Baltic Sea (Widerlund and Andersson 2011). Data suggest that the difference in surface salinity in the various Baltic Sea basins has been much less in the past that now and that the general trend over the past two millennia is towards freshening, with this trend strongest in the Bothnian Bay. These results would be consistent with a decrease in summer evaporation caused by increasingly colder summers, or with an annual increase in precipitation. Since the seasonality of the orbital forcing is quite different from that which is likely for a GHG-forced future climate, an extrapolation of these trends is not justified.

2.5 Conclusion

Over the Holocene, the Baltic Sea area has undergone major changes due to two interrelated factors. First, the demise of the Fennoscandian ice sheet that had previously accumulated as part of the 100-thousand-year-long glacial cycle. This demise started around 18,000 years ago, but due to the long timescales involved in land-ice dynamics and the slow crustal readjustment to the disappearance of the weight of an ice sheet 3 km deep, its effects are felt throughout most of the Holocene. The most important effect of the melting of the ice sheets is an interplay between global sea-level rise due to the increase in the ocean volume and a regional isostatic rebound (uplift) of the earth's crust. This interplay led to a series of transitions of the Baltic Sea basin as a brackish sea or freshwater lake, with periods of isolation and periods of interchange with the North Sea. The modern configuration of the Baltic Sea as a brackish Mediterranean or inland sea was established around 4500 years ago. Second, changes in the orbital configuration of the Earth. This is thought to be the major trigger for glacial to interglacial transition, but also modulated solar insolation to the high northern latitudes during the Holocene and thus strongly influenced the energy balance of the Baltic Sea area. Summer solar insolation modulated by these changes peaked around 7000–6000 years ago. Its effects were recorded in a phase of warm temperatures, extended vegetation cover and a progressive millennial cooling thereafter.

The instruments used to reconstruct the climate of the Baltic Sea area during the Holocene are multifaceted. Palaeobiological and geochemical records that, at least partly, reflect past environmental conditions can be used to reconstruct past temperature and/or precipitation in some seasons. Among these, the more widely used are pollen and spore assemblages derived from dated lake sediments, tree ring data, and fossil remains that may be indicative of the spatial spread of thermophilous or halophilous species and thus may give information about past temperatures or salinity.

The Baltic Sea area underwent a series of climatic phases during the Holocene. The Younger Dryas occurred in a transitional period between the Last Ice Age and the Holocene and ended around 11,500 years ago. Temperatures then rose, land ice partially disappeared, and the earth's crust rose causing a drop in relative sea level, giving rise to a large freshwater lake (Ancylus Lake). As temperatures and sea level continued to rise (due to the melting of the northern hemisphere ice sheets), a connection was established between the North Atlantic Ocean and Ancylus Lake forming the brackish Littorina Sea. The rise in temperature was interrupted by another climatic cold event, however, about 8200 years ago which is thought to have been caused by freshwater drainage from the North American ice sheet into the North Atlantic Ocean. This event was shorter and less intense than the Younger Dryas, with a drop in temperature of about 2 °C lasting for about 200 years. This event was the last cooling episode of the Early Holocene and was followed by a stable and relatively warm period with summer temperatures 1.0–3.5 °C higher than at present. The period between 7500 and 5500 years ago was the warmest in the Baltic Sea basin area as a whole, although the timing of maximum temperatures was not synchronous in different parts of the region. Thereafter, a trend of northern hemisphere cooling and increased climatic instability began, typical of the Late Holocene interval. Temperatures in the

Baltic Sea region started to drop around 5000–4500 cal year BP coincident with decreased summer solar insolation due to the quasi-cyclical changes in the Earth's orbit. The climatic variation during the Lateglacial and Holocene is reflected in the changing lake levels and vegetation and in the formation of a complex hydrographical network that set the stage for the Medieval Warm Period and the Little Ice Age of the past millennium.

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The Historical Time Frame (Past 1000 Years) 3

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Abstract

This chapter summarises the climatic and environmental information that can be inferred from proxy archives of the Baltic Sea area during the past millennium (1000 years). The proxy archives mainly comprise tree-ring analyses together with historical documents on extreme weather events and weather-related disasters. In addition to the reconstructed climate, climatic conditions are simulated using a regional climate model covering the Baltic Sea area. The chapter focuses on three of the main climatic periods of the past millennium: the Medieval Warm Period (900–1350), the Transitional Period (1350–1550) and the Little Ice Age (1550–1850). During these main historical climatic periods, climatic conditions were not uniform. Shorter warm/cool and wet/dry fluctuations were observed depending on regional factors.

Keywords

Millenium climate • Medieval warm period • Little ice age • Baltic sea basin

3.1 Introduction

This chapter addresses climate variability in the Baltic Sea basin over the past millennium (1000 years). Climate change at the millennial time scale was not discussed in the previous assessment of climate change in the Baltic Sea region (BACC Author Team 2008).

Understanding of millennial climate variability is mainly based on proxy data (see Sect. 3.2). Historical notes in the form of chronicles containing information about extreme weather events, and weather-related disasters are important sources of data about climatic conditions in the past millennium. Together with historical documents, dendroclimatology has provided much of the information on the climatic conditions of the past millennium. Tree-ring width and wood density are the main sources of dendroclimatological data. Other proxy data are of less importance for the past millennium and are more relevant for reconstructing the climatic conditions of the Holocene as a whole. Temperature and precipitation data for the Baltic Sea region, as reconstructed from different proxy data sets were also compared against the results of a regional climate model simulation at the Swedish Meteorological and Hydrological Institute (Schimanke et al. 2012). The chapter also describes the main climatic drivers, of which solar radiation factors, atmospheric circulation patterns

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_3

and volcanic eruptions appear the most important. The most unusual weather conditions in the Baltic Sea area in each century have been described in several publications (e.g. Rojecki 1965; Borisenkov and Pasetsky 1988, 2002; Pfister 1992, 1999; Glaser 2008).

3.2 Data Sources and Methodology

Research on millennial climate variability is mainly based on proxy data. This is because long-term instrumental measurements of meteorological elements (usually limited to atmospheric pressure, air temperature and wind) were recorded at only a few stations and over a relatively short period.

The first non-regular measurements of atmospheric pressure by means of a barometer constructed by Torricelli in 1643 were made in the period 1649–1658 in Clermont-Ferrand, Florence, Paris and Stockholm (von Rudloff 1967). The first regular meteorological measurements were launched in Greenwich (1774). The oldest continuous series of atmospheric pressure data for the Baltic Sea basin and surrounding areas date back to the nineteenth century: Prague (1800), Oslo (1816), Warsaw (1826), St Petersburg (1837), Copenhagen (1842), Stockholm (1844), Berlin (1848), Uppsala (1855), Haparanda (1859) and Helsinki (1882).

The first non-regular attempts at measuring air temperature began between 1654 and 1670 in Florence and Pisa in Italy (von Rudloff 1967); however, the longest homogenised and uninterrupted series of temperature data began in 1659 in central England (Manley 1974). In the Netherlands, direct temperature measurements started in 1705 in De Bilt (Cowie 2007). For the Baltic Sea basin, temperature measurements were initiated around the mid-1700s, with a denser coverage of the southern region in the late eighteenth century. The oldest continuous observational records of temperature are from Sweden: Uppsala (1739—non-regular measurements started in 1722), Lund (1741) and Stockholm (1756); Russia: St. Petersburg (1743); and Denmark: Copenhagen (1798) (von Rudloff 1967). A few temperature series began in the eighteenth century in central Europe: Berlin (1719), Jena (1770), Prague (1775), Warsaw (1779), Vilnius (1781), Wroclaw (1791) and Kraków (1792).

The oldest measurements of precipitation in Europe started in 1715 in Hoofddorp-Zwanenburg in the Netherlands and in 1725 in Padova in Italy. In the Baltic Sea area, the oldest precipitation station is Uppsala (1739) (von Rudloff 1967). In the first half of the nineteenth century, precipitation was also measured in a few other places: Warsaw (1803), Prague (1804), Copenhagen (1805), Jena (1827), Dresden (1828), Helsinki (1844) and Berlin (1847).

Historical notes in the form of chronicles containing information about extreme weather events and weather-related disasters are important sources of data about climatic conditions of the past millennium. In many cases, such notations are very precise as they locate events in space and time, sometimes even with an accuracy of a day. A systematic daily weather diary carried from 1502 to 1540 in Kraków and surroundings by Marcin Biem, a professor of Kraków Academy (Bokwa et al. 2001; Limanówka 2001) is unique on the European scale. A similar weather diary for the north-eastern part of Poland was kept by Jan Antoni Chrapowicki (Nowosad et al. 2007; Przybylak and Marciniak 2010) for the period 1656–1685. Borisenkov and Pasetsky (1988, 2002) compiled information about climate extremes and natural disasters from Russian chronicles. In Switzerland, Pfister (1992) established the European Centre of Historical Climate with the European Climate Historical database—EURO-CLIMHIST (Pfister 1992; Brázdil et al. 2010). Later, Glaser (2008) published the complete climate history for central Europe and Germany, covering the past 1200 years. Brázdil et al. (2005) discussed the state of European historical-climatological research with special attention to data sources, methods and significant results.

Together with historical documents, dendroclimatology has provided a large part of information on climatic conditions of the past millennium. Tree-ring width and wood density (Ljungqvist 2010) are the main sources of dendroclimatological data. Several recent multi-proxy reconstructions were made for the northern hemisphere (Jones et al. 1998, 2001a, b, 2009; Mann et al. 1998, 1999, 2008; Bertrand et al. 2002; Mann and Jones 2003; Jones and Mann 2004; Moberg et al. 2005; Ljungqvist 2010; Ljungqvist et al. 2012). On a regional scale, the most important reconstructions based on tree-ring or multi-proxy data are for Fennoscandia (Briffa et al. 1992; Gouirand et al. 2008; Lindholm et al. 2009, 2011; Esper et al. 2012), Finland (Helama et al. 2002, 2005, 2009b; Ogurtsov et al. 2008; Luoto and Helama 2010), central and northern Sweden (Gunnarson and Linderholm 2002; Linderholm and Gunnarson 2005; Moberg et al. 2006; Grudd 2008), eastern Norway (Kalela-Brundin 1999), Germany (Glaser 2008; Glaser and Riemann 2009), the north-western Baltic Sea (Klimanov et al. 1985), Russia (Klimenko and Solomina 2010), and Poland (Przybylak et al. 2005, 2010; Przybylak 2007, 2011; Szychowska-Krąpiec 2010; Koprowski et al. 2012). Although many of the data sets concern the Alps (Büntgen et al. 2005, 2006), they are well correlated with the central European mountains (Bednarz 1984, 1996; Bednarz et al. 1999; Niedźwiedź 2004; Büntgen et al. 2007, 2012).

Other proxy data were also applied to the millennium temperature reconstructions, for example, peat-bog deposits (Lamentowicz et al. 2008, 2009), lake fossils and sediments from Tsuolbmajavri Lake in northernmost Finland (Korhola et al. 2000), laminated lake sediments in Gościąż Lake in central Poland (Starkel et al. 1996; Ralska-Jasiewiczowa et al. 1998) and borehole temperatures (Majorowicz et al. 2004;

Majorowicz 2010). The reliability of the reconstructions is discussed by Holmström (2011). For northern Sweden, a summer-temperature reconstruction for the past 2000 years was achieved using pollen-stratigraphical data (Bjune et al. 2009). Among other biological proxies, a few reconstructions were based on diatoms (Korhola et al. 2000; Weckström et al. 2006) and chironomids (Korhola et al. 2002). These data are of less importance for reconstructing the climatic conditions of the past millennium and better used for reconstructing the climate of the Holocene as a whole.

Proxy records are clearly useful for helping understand the spatial and temporal variability of climate change, especially over periods shorter than the millennial time frame and which fall outside the instrumental period. Other data sets, such as long-term variability in Baltic Sea ice cover (Koslowski and Glaser 1999), runoff or oxygen conditions (Hansson and Omstedt 2008; Hansson et al. 2011; Hansson and Gustafsson 2011) can be used in combination with climate models to increase understanding.

3.3 General Features of the Millennial Climate

According to the scientific literature, four climatic periods have occurred over the past millennium (Lamb 1977, 1982; Grove 1988; Folland et al. 1990; Brázdil 1996; Crowley 2000; Crowley and Lowry 2000; Bradley et al. 2003; Brázdil et al. 2005; Xoplaki et al. 2005; NCR 2006; Esper and Frank 2009; Jones et al. 2009; Brázdil and Dobrovolný 2010; Büntgen and Tegel 2011; Büntgen et al. 2011b; Ogurtsov et al. 2011; Przybylak 2011; Ljungqvist et al. 2012):

- Medieval Warm Period (MWP 900–1350) or Medieval Climate Anomaly (MCA)
- Transitional Period (TP 1350–1550)
- Little Ice Age (LIA 1550–1850)
- Contemporary Warm Period (CW after 1850).

The dates for the four climatic periods are approximate and may differ slightly from one geographical region to another (Ljungqvist et al. 2012). Some shorter intervals are mainly related to the changes in solar activity or large volcanic eruptions. The Contemporary Warm Period is addressed in detail in Chaps. 4–9.

A comparison of late spring (April–June) precipitation and summer (June–August) air temperature across Europe between the southern Baltic Sea and the Alps (Fig. 3.1) was reported by Büntgen et al. (2011b) for the last 2500 years, together with possible impacts on civilisation.

Gouirand et al. (2008) reported variability in summer temperature for the whole of Fennoscandia (Fig. 3.2) over the past 1500 years. All four periods of the past millennium (MWP, TP, LIA and CW) are visible. The curves for central Europe (Fig. 3.1) and Fennoscandia (Fig. 3.2) show similarities but the deepest cooling in Fennoscandia is observed at the beginning of eighteenth century, while in central Europe summer temperature, negative anomalies are higher in the first decades of the nineteenth century. Also, a cool episode during the MWP in the first half of the twelfth century in Fennoscandia is sharper than in central Europe.

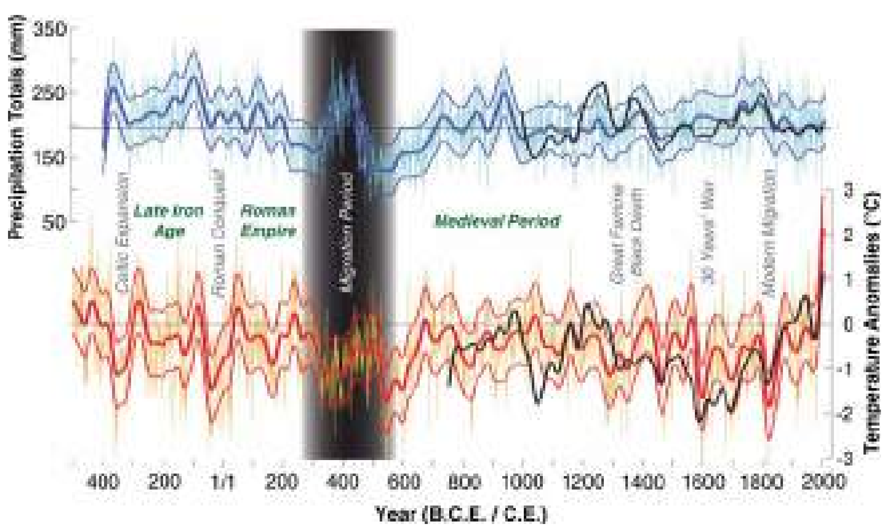

Fig. 3.1 Reconstructed April–June (AMJ) precipitation totals (*top*) and summer (June–August) temperature anomalies (*bottom*) for central Europe with respect to 1901–2000. *Error bars* are ±1 RMSE (Root-Mean Square Error) for the calibration periods. *Black lines* show independent precipitation and temperature reconstructions from Germany (Büntgen et al. 2010) and Switzerland (Büntgen et al. 2006). *Bold lines* are 60-year low-pass filters. Periods of demographic expansion, economic prosperity and societal stability are noted, as are periods of political turmoil, cultural change and population instability. Büntgen et al. (2011b)

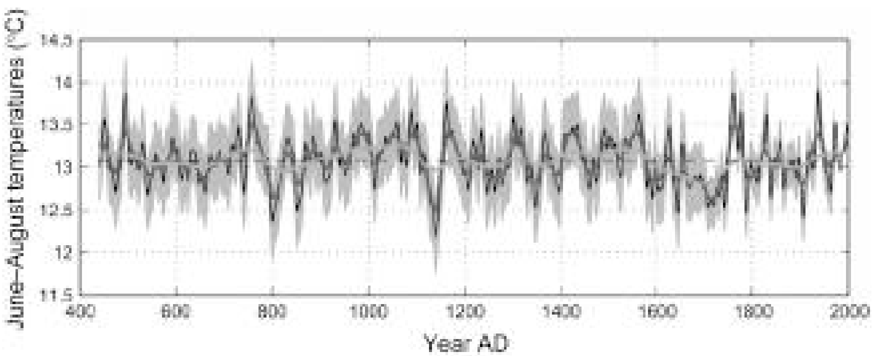

Fig. 3.2 Regional-average summer (June–August) temperatures AD 442–1970 for Fennoscandia created by merging seven reconstructions based on seven networks (Gouirand et al. 2008). The time series is extended to year 2000 with instrumental data. Data are shown as smoothed (Gaussian filtered) temperatures, highlighting variability on timescales longer than 10 years (*thin black*), 30 years (*thick dark grey*) and 100 years (*thick light grey*). The *dashed horizontal line* is the average for the entire period. The uncertainty in reconstructed temperatures (based on the calibration period statistics) is illustrated by ±2 standard errors with *grey shading* (for the 10-year smoothing only)

Mean January–April air temperature for the period 1170–1994 (Fig. 3.3) and 10-year anomalies of winter and summer air temperature from 1401 to 1800 (Fig. 3.4) have been reconstructed for Poland. The climatic history for Poland over the past millennium was reconstructed by Przybylak (2011) and more comprehensively by Przybylak et al. (2010). Tree-ring reconstruction of January–April air temperature indicates three relatively cool periods: 1475–1500, 1600–1660 and 1725–1830 (Fig. 3.3). The peak temperature of an exceptionally warm episode occurring 1661–1675 is slightly lower than indicated by other reconstructions for central Europe (compare Fig. 3.9). Cold anomalies for winter temperature in Poland (Fig. 3.4) suggest an increased annual temperature range during the LIA. A cool period at the final phase of the LIA in the first half of the nineteenth century is also found in a reconstruction based on the full depth of ground temperature in boreholes (Majorowicz et al. 2004).

Climate change during the past millennium over the Baltic Sea region was simulated by the Swedish Meteorological and Hydrological Institute using a regional climate model (Schimanke et al. 2012). The authors used the Rossby Centre Regional model (RCA3) with boundary conditions from the general circulation model ECHO-G. RCA3 includes radiative forcing, changes in orbital parameters, changes in greenhouse gas concentration and atmospheric circulation. The model covers the whole Baltic Sea area and its surroundings and has a horizontal resolution of about 50 × 50 km. Results were presented as 50-year running means; air temperature is largely underestimated (Schimanke et al. 2012). Biases in annual precipitation are about 20 % in the Baltic Sea region and during winter and autumn exceed 50 %.

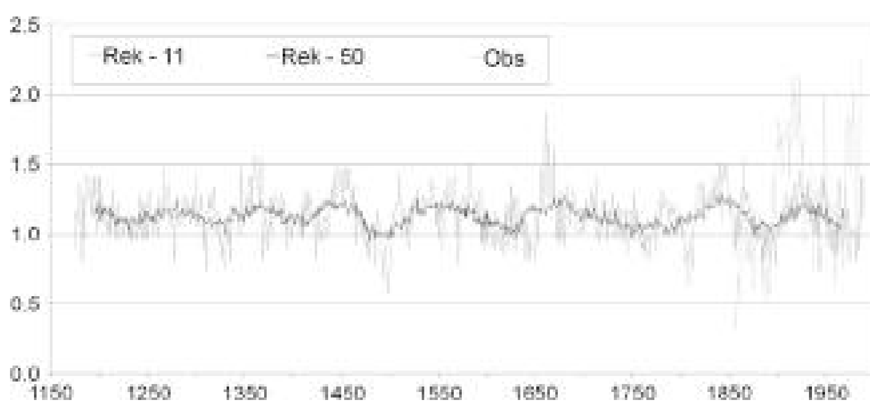

Fig. 3.3 Reconstructed mean January–April air temperature in Poland for the period 1170–1994 using a standardised chronology of Scots pine (*Pinus sylvestris* L.) tree-ring widths (modified after Przybylak et al. 2005; Przybylak 2011). Rek-11 and Rek-50 represent 11- and 50-year running means; reconstruction using areally averaged air temperature from Warsaw, Bydgoszcz and Gdańsk for calibration. Obs: measured mean January–April area-averaged air temperature from Warsaw, Bydgoszcz and Gdańsk (Przybylak et al. 2005; Przybylak 2007, 2011)

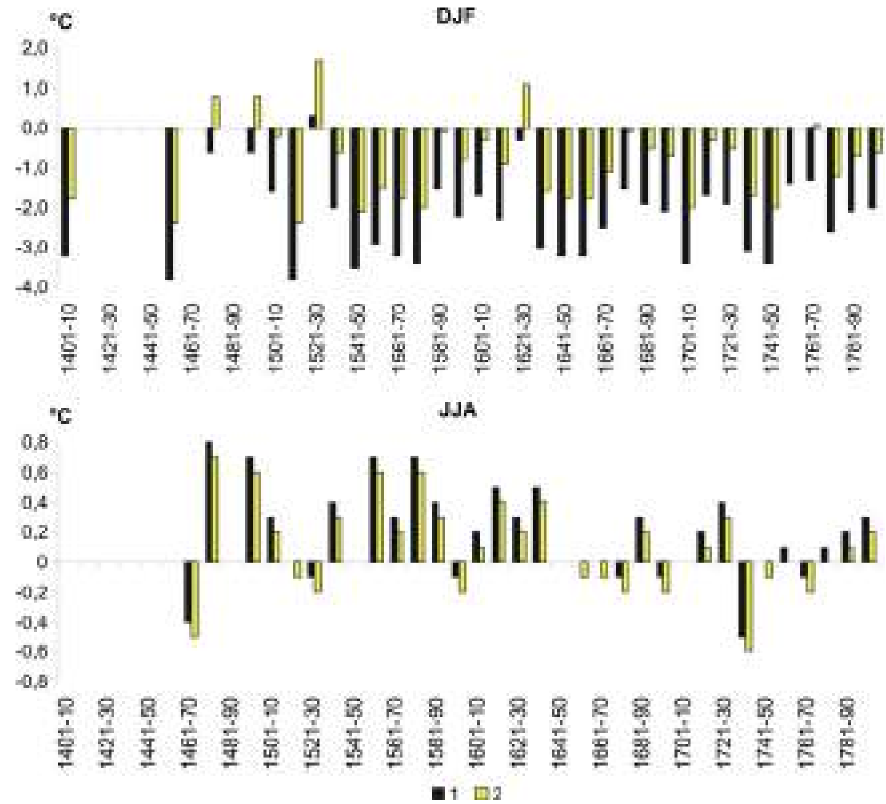

Fig. 3.4 Reconstructed mean 10-year air temperature in Poland from 1401 to 1840 based on historical *sources* winter (*DJF*) and summer (*JJA*). *1* and *2* anomalies with respect to 1901–1960 and 1789–1850 means, respectively. Przybylak (2011)

The 50-year running means for the annual temperature anomalies over Sweden varied from −0.5 °C in the seventeenth century during the LIA to about +0.4 °C in twelfth to thirteenth centuries representing the MWP (Fig. 3.5). The TP is clearly expressed by anomalies varying from about 0 °C through the fourteenth century to about −0.2 °C at 1550.

Winter temperature anomalies are larger ranging from about −0.9 °C in 1700 to +0.8 °C at the middle of the twelfth century (Fig. 3.5). Variability in summer temperature anomalies is less than for the annual anomalies, ranging from −0.3 °C in the early eighteenth century to +0.3 °C during the MWP near the mid-twelfth century.

The variability in annual and winter temperature anomalies for the Baltic Sea region presented in Fig. 3.5 (Schimanke et al. 2012) differs from other reconstructions. Winter temperature anomalies are lowest in the latter half of the seventeenth century (−0.9 °C). This shows some agreement with the reconstructed winter temperatures for the greater Baltic Sea area (Eriksson et al. 2007) discussed in Sect. 3.5. A large discrepancy is visible in the first half of the eighteenth century when the reconstructed winter temperature anomaly is greatest in the LIA, while the model simulation indicates a warm episode near 1800, which Eriksson et al. (2007) suggested was the coolest in the LIA (based on 15-year running means for temperature).

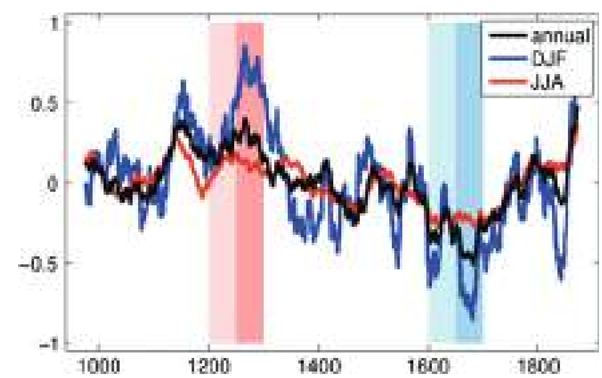

Fig. 3.5 The 2-m temperature anomaly with regard to the preindustrial mean (950–1900) for the winter (*DJF*), summer (*JJA*) and annual mean averaged over the Baltic Sea region. The coloured sections highlight the periods that are defined as MWP (*red*) and LIA (*blue*). The darker colours reflect the 50-year periods which are considered for the analysis of the Baltic Sea. After Schimanke et al. (2012)

Reconstructing past precipitation using proxy measurements is more difficult than for temperature and is only possible for parts of the year. There is no possibility of reconstructing annual precipitation data from proxy data. Büntgen et al. (2011b) reconstructed total precipitation from April to June for central Europe between the southern Baltic Sea and the Alps. Warm and dry summers are typical during

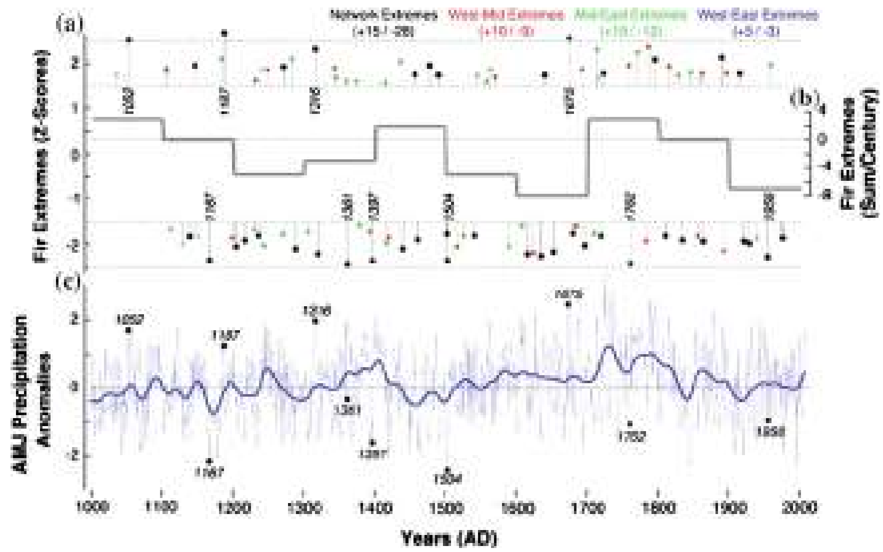

Fig. 3.6 **a** Central European and regional fir TRW (Tree Ring Width) extremes, and **b** their centennial changes over the past millennium (network extremes were double weighted), compared to **c** annual-resolve and 40-year low-passed Central European April–June precipitation variability (Büntgen et al. 2011c)

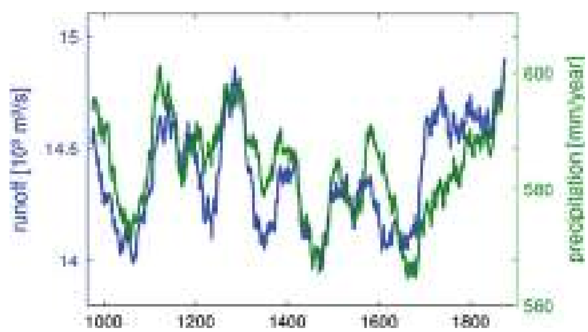

Fig. 3.7 Annual precipitation in the Baltic Sea catchment area (*green*) and statistical estimated runoff (*blue*) as 50-year running means. After Schimanke et al. (2012)

the MWP (900–1350). The onset of the TP in the latter half of the thirteenth century is indicated by cooler and wetter summers. Total precipitation between April and June over the past millennium in central Europe is presented in Figs. 3.1 (Büntgen et al. 2011b) and 3.6 (Büntgen et al. 2011c). Generally, the MWP was drier than the first half of the TP and the LIA in spring and early summer seasons.

Annual precipitation simulated in the Baltic Sea catchment area by a regional climate model (Schimanke et al. 2012) and shown in Fig. 3.7 indicates that the annual totals are larger during the MWP than the TP and the LIA. The annual precipitation data cannot be compared with the reconstructed data for April–June. However, the very dry period in the latter half of the fifteenth century is marked in both precipitation series. The variability of river runoff to the Baltic Sea follows changes in precipitation. However, the simulation of river runoff for the period 1500–1995 by Hansson et al. (2011) presented in Chap. 5, Fig. 5.3 indicates no significant long-term change.

3.4 The Medieval Warm Period (MWP 900–1350)

At the turn of the tenth and eleventh centuries, relatively stable climate conditions with few extremes prevailed in the Baltic Sea basin and the surrounding parts of Europe. Hot and dry summers were noted in 993 and 994, respectively. For example, in Russian chronicles, eight droughts (1000, 1025–1028, 1035, 1037, 1092), four wet summers with rain and floods (1002, 1031, 1034, 1043) and seven severe winters (1034/35, 1043/44, 1047/48, 1056/57, 1058/59, 1066/67, 1076/77) were noted in the eleventh century (Borisenkov and Pasetsky 1988, 2002).

In Europe, the warmest conditions occurred between 1200 and 1250, and the MWP ended about 1350 (Borisenkov and Pasetsky 1988, 2002). Chernavskaya (1996) reconstructed the June temperature changes in European Russia over the past two millennia based on pollen analysis in peat bogs. Data from Polistovo (56.8°N, 38.1°E) suggest an earlier occurrence of the MWP (ninth to tenth centuries) in European Russia than in central Russia (tenth to eleventh centuries). Two periods of strong cooling occurred in the middle of the twelfth century and at the end of the fourteenth century. On the East European Plain, summer temperatures during the MWP were found to be above the long-term average between 900 and 1200 (Klimenko and Solomina 2010). For Fennoscandia (Fig. 3.2), summer temperatures were elevated in the tenth and eleventh centuries (Gouirand et al. 2008). Even warmer summer periods were noted in the late twelfth century, succeeding the extreme cool summers in the mid-twelfth century.

Recent investigations of Fennoscandia by Ljungqvist (2010) showed that the MWP occurred between 800 and 1300. At that time, warm-season (May–September) temperatures exceeded the contemporary warming of the end of twentieth century by about +0.5 °C. The start of the warming was noted between the ninth and tenth centuries, and the peak temperature appeared at the beginning of the second half of the twelfth century. In a winter temperature simulation over the Baltic Sea region (Schimanke et al. 2012) during that time anomalies reached their highest value of +0.8 °C for the MWP (Fig. 3.5). Lower temperatures occurred at the end of the MWP, about 1350. A diatom-based July temperature reconstruction for the past 800 years in northern Scandinavia (Weckström et al. 2006; Holmström 2011) indicates that temperature was about 0.2 °C higher in the latter half of the fifteenth century than in 1970–2000. An exceptionally warm period occurred in 1220–1250 and in the latter half of the fifteenth century (1470–1500) in the TP. The frequency of extreme temperature events in Russia increased in the twelfth century (Borisenkov and Pasetsky 1988, 2002). Winter-simulated temperature indicates the second warm episode of the MWP in the latter half of the thirteenth century (Schimanke et al. 2012) for the Baltic Sea region (Fig. 3.5). At the beginning of the fourteenth century, climatic conditions cooled rapidly. In 1315, a serious famine in northern Europe resulted from a series of very cold winters (1302/03 and 1305/06) and cool and wet summers (1314–1317) across the whole of Europe (Cowie 2007).

There is less information available on precipitation in the MWP (Gouirand et al. 2008; Büntgen et al. 2011b, c). Nevertheless, a regional dendroclimatic precipitation reconstruction from southern Finland showed a uniquely prolonged rainfall deficit coinciding with the MWP (Helama et al. 2009a). The drought was particularly severe between 1000 and 1200. The simulation of annual precipitation for the Baltic Sea catchment using a regional climate model (Schimanke et al. 2012) shows that the driest period was the mid-eleventh century and that two wet periods occurred in the first half of the twelfth century and the latter half of the thirteenth century (Fig. 3.7). As a generalisation, relatively dry periods occurred in Europe in the years: 1272–1291, and 1300–1309, while the wettest conditions were noted in 1312–1322 (Borisenkov and Pasetsky 1988, 2002). In April–June in central Europe (Fig. 3.6), wet conditions were observed in 1052, 1187 and 1316, and the driest in 1167 (Büntgen et al. 2011c). Distinct wet periods with frequent floods were recognised in this region at 1020–1030 and 1075–1100 (Starkel 2001).

3.5 The Transitional Period (TP 1350–1550)

The increase in the intra-seasonal variability of climate at the end of the MWP in the period 1270–1350 is considered to be the beginning of the LIA; however, Brázdil et al. (2005) suggested that the following 200-year period should be treated as transitional between the MWP and the LIA. This period was characterised by a great variability of climatic conditions. At that time, temperature decreased by about 1.2 °C, but cooling occurred until the latter half of the sixteenth century (Borisenkov and Pasetsky 1988, 2002). During this period, the decreasing tendency of mean annual and seasonal temperatures simulated for Baltic Sea region (Fig. 3.5) is clearly visible.

Over the majority of Europe and Russia, very unfavourable conditions for agriculture in the period 1400–1480 were linked to large fluctuations in temperature and precipitation. For example, the summers of 1428, 1434, 1436 and 1438 were hot and dry and the summer of 1435 was cool and dry, whereas the summers of 1432, 1437 and 1439 were extremely wet with flooding (Borisenkov and Pasetsky 1988, 2002). In central Europe (Fig. 3.6), the first part of the TP up to about 1430 was very wet according to the reconstructed April–June precipitation curve (Büntgen et al. 2011c) and followed by very dry conditions with an extremely dry and hot spring and summer in the year 1504 (Glaser 2008).

In Poland (Fig. 3.4), severe winters were detected in four decades: 1401–1410, 1451–1460, 1511–1520 and 1541–1550 (Przybylak 2011). The warmest were winters during the years 1521–1530. Similar thermal conditions in winter based on historical sources were found in Latvia (Jevrejeva 2001) and Estonia (Tarand and Nordli 2001). The summers were relatively warm in two decades: 1471–1480 and 1491–1500.

The first halves of the fifteenth and sixteenth centuries appear relatively warm periods, but there was large variability (Helama et al. 2009b). Climatic variability may be reflected in the proxy instability. According to a diatom-based reconstruction, the warmest 30-year non-overlapping period in northern Scandinavia occurred in 1470–1500 (Weckström et al. 2006); however, according to a dendroclimatic reconstruction in this region, the summers of the late fifteenth century were anomalously cold (Helama et al. 2009b). There were very warm conditions in Fennoscandia in summer at the end of the TP (Gouirand et al. 2008; Fig. 3.2).

3.6 The Little Ice Age (LIA 1550–1850)

In the latter half of the sixteenth century, the temperature dropped. This tendency was particularly clear in the period 1569–1579. Another sequence of extremely wet and cool summers was identified at the end of the sixteenth century (Borisenkov and Pasetsky 1988, 2002). The longest consecutive cold period occurred from the late sixteenth century to the mid-eighteenth century (Gouirand et al. 2008), which is in very good agreement with a regional climate model simulation for temperature in the Baltic Sea region (Schimanke et al. 2012; Fig. 3.5). But a short sequence of very warm summers was observed in the latter part of eighteenth century, just before the prolonged cooling at the end of the eighteenth and during the nineteenth century (Fig. 3.2).

Eriksson et al. (2007) analysed the complex description of winter climate conditions during the LIA for the greater Baltic Sea region (50–70°N, 0–30°E). Their study is based on well-documented time series of ice cover, sea-level pressure and winter surface-air temperatures. Using winter temperature in connection with atmospheric circulation and ice conditions, they found four cold and three warm periods during the LIA (Fig. 3.8). In the latter half of the sixteenth century, a cool phase (1562–1576) passed to a relatively mild period (1577–1591). In the seventeenth century, two phases of cold winters, 1597–1629 and 1663–1706, were

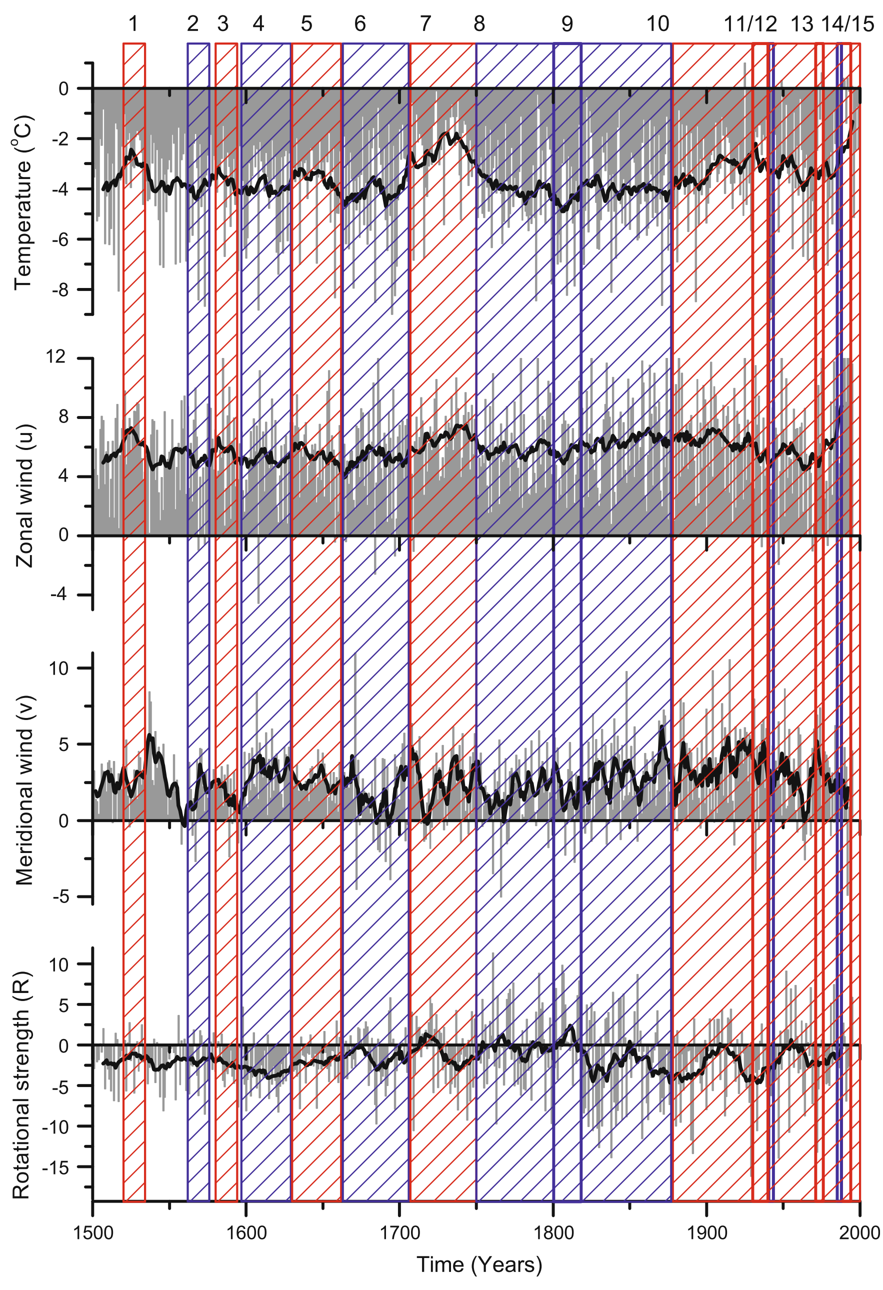

Fig. 3.8 Winter climate conditions in the greater Baltic Sea region since 1500. The *grey colour* shows seasonal winter data from two gridded data sets: (*top* to *bottom*) Baltic Sea mean winter air temperature. *Black line* in all panels is a 15-year running mean. *Blue* and *red* fields cover periods classified as cold and mild, respectively (Eriksson et al. 2007)

divided by the mild period of 1630–1662. The first half of the eighteenth century (1707–1750) which includes the warmest winter (1723/24) is considered to be the major warm period. At this time, the maximum ice extent in the Baltic Sea was similar to present-day conditions, albeit there are many uncertainties related to the observations at that time (Seinä and Palosuo 1996; Hansson and Omstedt 2008). The period 1730–1745 has also been described as particularly variable interannually, swinging from extremely mild to extremely cold winters (Jones and Briffa 2006). The occurrence of cold winters is related to the Late Maunder Minimum (1675–1715) which has also been discussed by Luterbacher et al. (2001) and in Chap. 4, Sect. 4.2.3. The longest cool period in the final phase of the LIA in the Baltic Sea region (1750–1877) coincided with the Dalton Minimum (1790–1840) in solar activity (Eriksson et al. 2007; see also Chap. 4). During the entire LIA, no downward trend in ice break-up date of the river Daugava was detected (see also Chap. 5).

A new reconstruction of the Baltic Sea region climate for the past 500 years was prepared by Brázdil et al. (2010) on the basis of instrumental data and documentary evidence under the MILLENNIUM project. January–April mean temperatures were reconstructed for Stockholm (1502–2008) and central Europe (1500–2007). In central Europe, the coldest conditions were observed in the sixteenth century, while in central Sweden the end of the seventeenth century was cooler (Fig. 3.9). In central Europe, the warmest period was the first half of the eighteenth century, while in Stockholm such conditions occurred at the end of the eighteenth

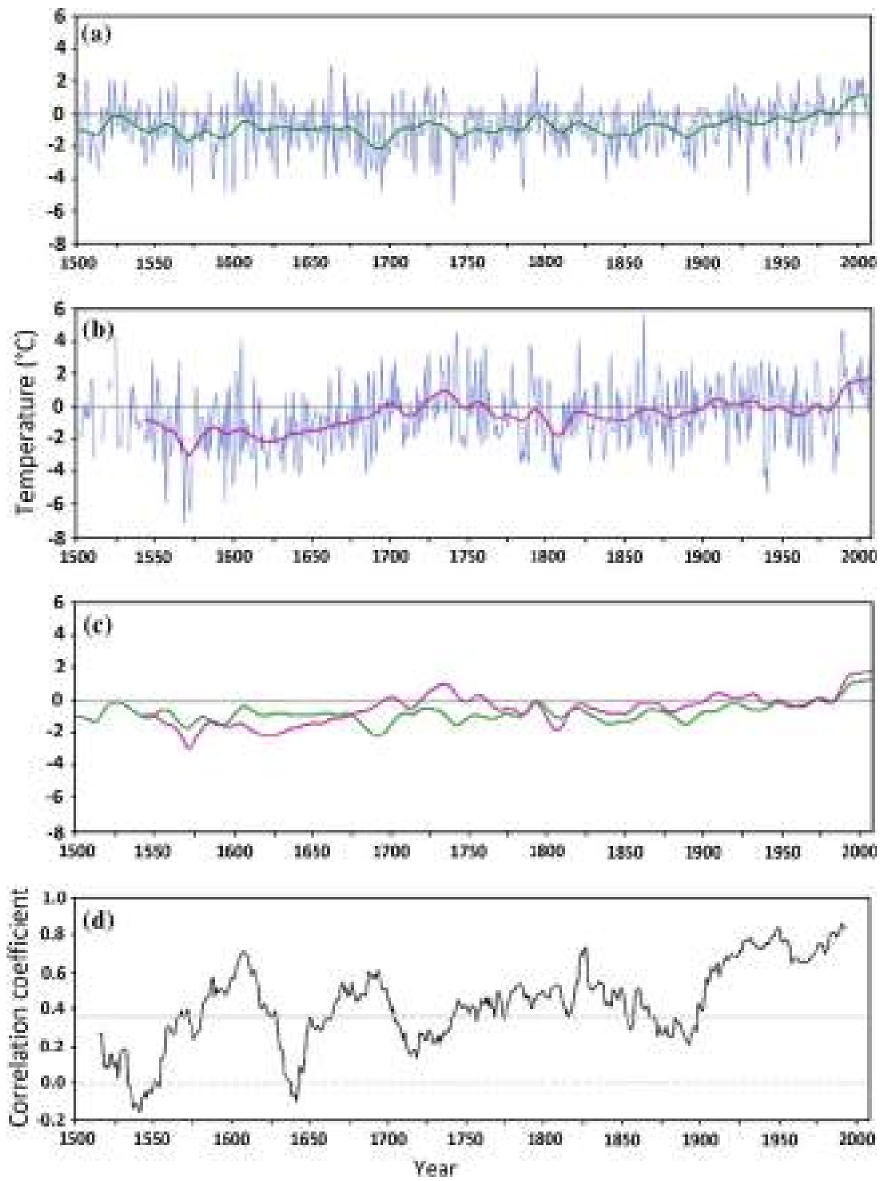

Fig. 3.9 Comparison of reconstructed JFMA (January–April) temperatures for Stockholm (1502–2008) and CEuT (Central European Temperature) (1500–2007) (anomalies from the 1961–1990 mean). Original series of CEuT (**a**) and Stockholm (**b**) are smoothed with a 30-year Gaussian filter (**c**) and compared using 31-year running correlations between unfiltered data (**d**). The *horizontal solid line* in d denotes the critical value of correlation coefficients for $\alpha = 0.05$ for one-tailed t test (Brázdil et al. 2010)

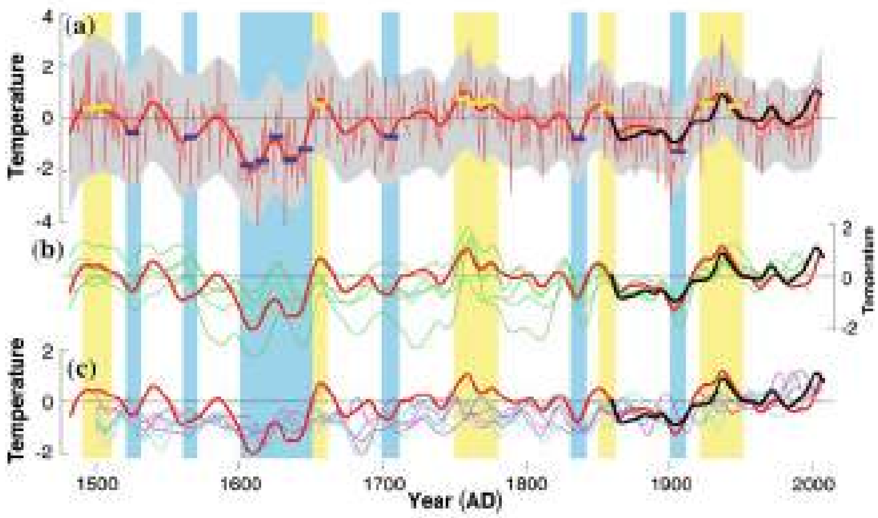

Fig. 3.10 Reconstructed and modelled northern Scandinavian summer temperature variations over the past five centuries (Büntgen et al. 2011a). **a** Actual (*black*) and reconstructed (*red*) JJA (June–August) temperature anomalies (°C) with error estimates (*grey*) and the ten warmest and coldest decades superimposed (*colour boxes*). **b** Comparison of the actual and reconstructed temperatures with five existing (*green*) reconstructions. **c** Comparison of the actual and reconstructed temperatures with CCSM3 surface-air temperature (*pink*) and sea surface temperature (*blue*) model simulations. Mean and variance of the data are scaled against JJA temperature (1860–2006), expressed as anomalies (relative to 1961–1990) and 20-year low-pass filtered

century. In the recently modelled and reconstructed northern Scandinavian summer temperatures since 1500 (Fig. 3.10), the longest cool period covered the first half of the seventeenth century (Büntgen et al. 2011a) and the beginning of the eighteenth century as well as in the early nineteenth century. The warmest conditions occurred just after the middle of the eighteenth century.

There are several climate reconstructions based on the documentary and instrumental data for the period since 1500. Koslowski and Glaser (1999) constructed an ice winter severity index and temperature for the western Baltic Sea region, covering the period of 1500–1759. Temperature series were reconstructed for Stockholm (Moberg et al. 2002; Leijonhufvud et al. 2010), Tallin (Tarand and Nordli 2001), Poland as a whole (Dobrovolný et al. 2010; Luterbacher et al. 2010; Przybylak et al. 2010) and the Tatra Mountains (Bednarz 1984; Niedźwiedź 2004; Büntgen et al. 2007), and several parts of central Europe (Pfister 1992; Brázdil 1996; Luterbacher et al. 2004; Dobrovolný et al. 2010).

Glaser et al. (2010) reconstructed the variability of floods in Europe since 1500. The precipitation conditions are presented in Figs. 3.1 and 3.6. The wettest conditions during the LIA were observed in 1675 (Glaser 2008) and the first half of the eighteenth century, and the driest were the years 1504 and 1762 (Büntgen et al. 2011c). In southern Finland, conditions were markedly wetter during the LIA than the MWP, as inferred from tree rings (Helama et al. 2009a). An exception was a period of transient drought which, as the same dendroclimatic reconstruction indicates, occurred during the first decades of the nineteenth century. A regional model simulation of annual precipitation (Schimanke et al. 2012) shows that the driest period of the LIA was the latter half of the seventeenth century (Fig. 3.7). For central-east Sweden the low May–June precipitation was found by Jönsson and Nilsson (2009) to occur in the periods 1560–1590 and 1694–1751. The clustering of heavy rainfall years is typical for the LIA, causing the floods in central Europe in 1590–1610, 1705–1715 and 1800–1815 (Starkel 2001). They coincide with the relatively cool periods of Spörer, Maunder and Dalton solar activity minima.

In the seventeenth century, the winter of 1657/58 was exceptionally cool. Modelled Baltic Sea ice extent indicates that it was one of the coldest winters for the Baltic Sea since 1500 (Hansson and Omstedt 2008). The sequence of cool summers and winters occurred over the period 1690–1699. In Sweden, climatic conditions favourable for agriculture occurred in 1604–1620 (Borisenkov and Pasetsky 1988, 2002). According to a diatom-based July temperature reconstruction in Finnish Lapland (Weckström et al. 2006), particularly cold 30-year periods were detected between 1640–1670 and 1750–1780. In several reconstructions, the low-frequency variability of LIA temperature is underestimated (e.g. von Storch et al. 2004; Zorita et al. 2007;

Christiansen and Ljungqvist 2011). A new method of temperature reconstruction (Christiansen and Ljungqvist 2011) showed that in the seventeenth century, the lowest temperature anomaly in the cooling northern hemisphere reached a 50-year smoothing value 1.1 °C below the contemporary level (1880–1960). The period 1630–1700 was the coolest consecutive period of the entire past millennium.

At the beginning of the eighteenth century, the winter of 1708/09 was perhaps the coldest winter of the past 500 years (Luterbacher et al. 2004). In Poland, the coolest winters were recorded in the decades 1701–1710 and 1741–1750, while the coldest summers were found in the decade 1731–1740 (Przybylak 2011). Very cold winters were also observed at the end of eighteenth century in 1783/84, 1788/89, 1794/95 and 1798/99 (Borisenkov and Pasetsky 1988, 2002). This cold period continued to the end of the first half of the nineteenth century. In 1815, the Tambora volcanic eruption in Indonesia discharged large amounts of ash into the upper atmosphere, resulting in the famous 'year without a summer' in 1816. This particular year, the summer in western Europe was unusually cold. However, that was not the case in the Baltic Sea basin (Briffa and Jones 1992). In eighteenth century in Poland, precipitation showed large variability. The period 1731–1750 was wetter than normal (Przybylak 2011). But between 1751 and 1766, during a generally dry period with 13 dry years, the wettest was 1755 and the driest 1762.

Dendroclimatological studies have identified several cool and rainy summers in the Carpathian Mountains (southern Poland) in the latter part of the LIA: 1650–1660, 1670–1685, 1690–1719 and 1735–1745 (Bednarz 1984, 1996). The final phase of the LIA in the first half of the nineteenth century was also marked by a sequence of exceptionally cold years between 1812 and 1824. In that period, average winter temperature in Russia and a large part of Europe was lower than normal by as much as 10–12 °C (Borisenkov and Pasetsky 1988, 2002). In central Europe, the winter of 1829/30 was extremely cold, as well as the winters of 1822/23 and 1837/38. In Norway, based on farmer diaries (Nordli et al. 2003), the severely cold spring/summer (April–August) decadal temperatures were found around 1740, and the 1800s and 1830s.

The cooling during the LIA had an important influence on human society. At the turn of the thirteenth and fourteenth centuries, the number of farms in northern Norway decreased due to a drop in temperature (Cowie 2007). In Finland, abandoning of farms in the sixteenth and seventeenth centuries coincided with a long-term summer temperature cooling, implying that the desertion may have resulted from a change in climatic conditions that significantly limited agriculture as a means of subsistence (Holopainen and Helama 2009). In the same region, harvest records show that during the poorest years the amount of grain harvested was less than had been sown (Holopainen and Helama 2009). Very unfavourable weather conditions in 1697 resulted in a failed crop. This caused widespread famine, followed by over a third of the Finnish and a fifth of the Estonian population dying in just a few years (Cowie 2007). Also, the cold winter of 1657/58 permitted the Swedish King Charles X to walk across the frozen Belts and Sound with his army and occupy all of Denmark, except for Copenhagen, with a large loss of land for the Danish (Neumann 1978; Hansson and Omstedt 2008).

There are discrepancies in the dates of both the end of the LIA and the beginning of the CW. The majority of scientists agree on 1850 as being crucial (e.g. Grove 1988). However, some climatologists claim that the LIA did not finish until the last decades of the nineteenth century (e.g. Lamb 1977). In southern Poland (Tatra Mountains) according to summer temperature data, the final episode of the LIA lasted 103 years: 1793–1895 (Niedźwiedź 2004).

3.7 Conclusion

According to the scientific literature, there are four climatic periods of the past millennium: the Medieval Warm Period (MWP 900–1350), the Transitional Period (TP 1350–1550), the Little Ice Age (LIA 1550–1850), and the Contemporary Warm Period (CW after 1850). The MWP started at the beginning of the tenth century with relatively stable climate conditions, and few extremes prevailed in the Baltic Sea basin and the surrounding parts of Europe. The warmest conditions occurred between 1200 and 1250. Two periods of strong cooling were detected in the middle of the eleventh and at the beginning of the fourteenth century. During the MWP in Fennoscandia, warm-season (May–September) temperatures exceeded the contemporary warming of the end of twentieth century by about +0.5 °C. The following 200-year period should be treated as a transitional period between the MWP and LIA. This period was characterised by a great variability in climatic conditions; at that time, temperature decreased by about 1.2 °C. In the latter half of the sixteenth century, the temperature dropped, initiating the LIA. Winter temperatures in combination with atmospheric circulation and ice conditions indicate four cold and three warm periods during the LIA. In the recently modelled and reconstructed northern Scandinavian summer temperatures since 1500, the longest cool period prevailed during the first half of the seventeenth century and at the beginning of the eighteenth century, as well as during the first years of the nineteenth century. During these main historical climatic periods, climatic conditions were not uniform. Shorter warm/cool and wet/dry fluctuations were observed depending on regional factors.

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Part II

Recent Climate Change (Past 200 Years)

Introduction

The six chapters of this part describe the observed climatic changes in the atmosphere, on land (river runoff, cryosphere) and at sea (circulation and stratification, sea ice, sea level and wind waves) for properties in the Baltic Sea drainage basin over the past 200 to 300 years. This period is characterised by the existence of in-situ measurements in contrast to earlier periods for which the information is mainly obtained by proxy data (see Part I). The Baltic Sea area is relatively unique with a dense observational network covering an extended time period. Continuous time series extend back to the mid-eighteenth century for a few stations. A denser network of stations measuring continuously with relatively good accuracy has been developed since the mid-nineteenth century. The start of environmental monitoring by satellites in 1978 has meant increased coverage in space and time, especially in regions not covered by conventional observations, and the inclusion of additional variables has enhanced the possibility for extracting and analysing the data enormously. Data analysis may be subdivided into three periods: an early period with sparse and relatively uncertain measurements, a period with well-developed synoptic stations, and the past 30 + years characterised by the availability of satellite data and sounding systems. It cannot be expected that data spanning such long periods is homogeneous in time. Conclusions concerning long-term trends should only be drawn from homogenised data. It is important to have a good understanding of climate change over the past 200 years, since it is during the latter part of this period that the recent anthropogenic influence can potentially be seen. The focus of the following six chapters is on physical environmental parameters such as patterns of atmospheric circulation, radiation and temperature, and on parameters related to the water cycle (precipitation, hydrology and the cryosphere). Most of these parameters are strongly interdependent.

4 Recent Change—Atmosphere

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Abstract

This chapter describes observed changes in atmospheric conditions in the Baltic Sea drainage basin over the past 200–300 years. The Baltic Sea area is relatively unique with a dense observational network covering an extended time period. Data analysis covers an early period with sparse and relatively uncertain measurements, a period with well-developed synoptic stations, and a final period with 30+ years of satellite data and sounding systems. The atmospheric circulation in the European/Atlantic sector has an important role in the regional climate of the Baltic Sea basin, especially the North Atlantic Oscillation. Warming has been observed, particularly in spring, and has been stronger in the northern regions. There has been a northward shift in storm tracks, as well as increased cyclonic activity in recent decades and an increased persistence of weather types. There are no long-term trends in annual wind statistics since the nineteenth century, but much variation at the (multi-)decadal timescale. There are also no long-term trends in precipitation, but an indication of longer precipitation periods and possibly an increased risk of extreme precipitation events.

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_4

4.1 Introduction

This chapter reports on trends and variability in atmospheric parameters over the past 200–300 years. The focus is on large-scale atmospheric circulation and its changes, as well as on observed changes in surface variables such as wind, temperature and precipitation. Situated in the extra-tropics of the Northern Hemisphere, the Baltic Sea basin is under the influence of air masses from the Arctic to the subtropics. It is therefore a region of very variable weather conditions. From a climatological point of view, the region is controlled by two large-scale pressure systems over the north-eastern Atlantic Ocean—the Icelandic Low and the Azores High—and a thermally driven pressure system over Eurasia (high pressure in winter, low pressure in summer). In general, westerly winds predominate, although any other wind direction is also frequently observed. As the climate of the Baltic Sea basin is to a large extent controlled by the prevailing air masses, atmospheric conditions will therefore be controlled by global climate as well as by regional circulation patterns. The atmospheric parameters are strongly interlinked (i.e. the circulation influences the wind, temperature, humidity, cloudiness and precipitation patterns, and the radiation and cloudiness influence surface temperature).

4.2 Large-Scale Circulation Patterns

The atmospheric circulation in the European/Atlantic sector plays an important role in the regional climate of the Baltic Sea basin (Hurrell 1995; Slonosky et al. 2000, 2001; Moberg and Jones 2005; Achberger et al. 2007). The Baltic Sea region is influenced in particular by the North Atlantic Oscillation (NAO; Hurrell 1995). The NAO influences northern and central Europe and the north-east Atlantic and therefore also the climate in the Baltic Sea basin. The impact of the NAO is most pronounced during the winter season, November to March (Hurrell et al. 2003). While the NAO is defined in relation to conditions within the European/Atlantic sector, it is in fact part of a hemispheric circulation pattern, the Arctic Oscillation (AO; e.g. Thompson and Wallace 1998). See Box 4.1.

Box 4.1 North Atlantic Oscillation

The NAO is the dominant mode of near-surface pressure variability over the North Atlantic and neighbouring land masses, accounting for roughly one-third of the sea level pressure (SLP) variance in winter. In its positive (negative) phase, the Icelandic Low and the Azores High are enhanced (diminished), resulting in a stronger (weaker) than normal westerly flow (Hurrell 1995). For strongly negative NAO indices, the flow can even reverse when there is higher pressure over Iceland than over the Azores.

There is no unique way to define the spatial structure of the NAO. One approach uses one-point correlation maps (Hurrell et al. 2003). These can be used to identify the NAO as regions of maximal negative correlation over the North Atlantic (e.g. Wallace and Gutzler 1981). Points identified by this procedure are situated near or over Iceland and over the Azores extending to Portugal, respectively. Other approaches use principal component analysis, in which the NAO is identified by the eigenvectors of the cross-correlation matrix which is computed from the temporal variation of the grid point values of SLP, scaled by the amount of variance they explain (e.g. Barnston and Livezey 1987) or clustering techniques (e.g. Cassou and Terray 2001a, b). A third option uses latitudinal belts. An index defined this way yields higher correlations with air temperature and precipitation in the eastern Baltic Sea region (e.g. Li and Wang 2003).

The NAO is the first mode of a principal component analysis of winter SLP. The second mode is called the east Atlantic pattern (Wallace and Gutzler 1981) and represents changes in the north–south location of the NAO (Woolings et al. 2008). It is characterised by an anomaly in the north-eastern North Atlantic Ocean, between the NAO centres of action. Negative values mean a southward displacement of the NAO centres of action and lower temperatures (Moore and Renfrew 2012), positive values correspond to more zonal winds over Europe and expected higher temperatures. The third dominant mode is the Scandinavian pattern, also called the Eurasian (Wallace and Gutzler 1981) or blocking pattern (Hurrell and Deser 2009), which in its positive phase is characterised by a high-pressure anomaly over Scandinavia and a low-pressure anomaly over Greenland. This indicates an east–west shift of the northern centre of variability defining the NAO.

As shown in Fig. 4.1, the strongly positive NAO phase in the 1990s can be seen as a component of multi-decadal variability comparable to conditions at

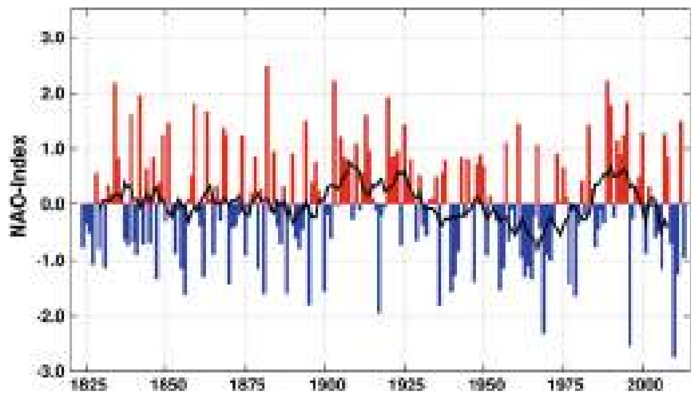


Fig. 4.1 NAO index for boreal winter (DJFM) 1823/1824–2012/2013 calculated as the difference between the normalised station pressures of Gibraltar and Iceland (Jones et al. 1997). Updated via www.cru.uea.ac.uk/~timo/datapages/naoi.htm and renormalised for the period 1824–2013

the beginning of the twentieth century rather than as part of a trend towards more positive values.

The long-term annual variations in the NAO are in good agreement with 99th percentile wind speeds (Wang et al. 2011) over western Europe and the first principal component (PC1) calculated over eight different pressure-based storm indices over Scandinavia (Bärring and Fortuniak 2009), showing large multi-(decadal) variations in atmospheric circulation and related wind climates (Fig. 4.2 and further discussed in Sect. 4.3.2).

4.2.1 Circulation Changes in Recent Decades

From a long-term perspective, the behaviour of the NAO is irregular. However, for the past five decades, specific periods are apparent. Beginning in the mid-1960s, a positive trend has been observed, that is towards more zonal circulation with mild and wet winters and increased storminess in central and northern Europe, including the Baltic Sea area (e.g. Hurrell et al. 2003). After the mid-1990s, however, there was a trend towards more negative NAO indices, in other words a more meridional circulation. These circulation changes are apparently independent of the exact definition of

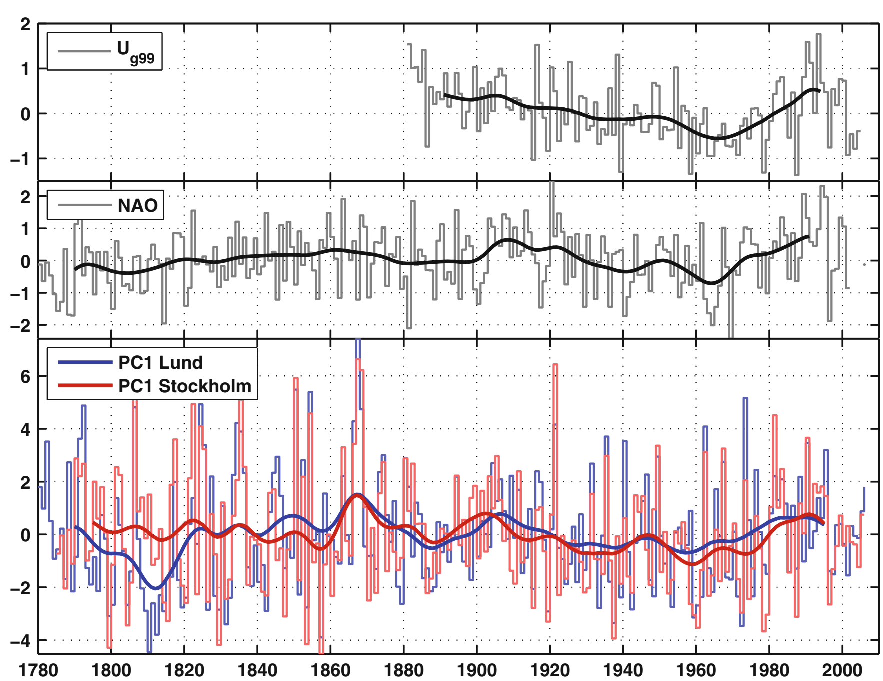


Fig. 4.2 Time evolution of the 99th percentiles of the geostrophic wind index (Alexandersson et al. 1998, 2000, *top*), a reconstructed NAO index (Luterbacher et al. 2002, *centre*) and the first principal components of the Lund and Stockholm storminess indices (PC1) over the Baltic Sea region. *Thick curves* are filtered with a Gaussian filter (σ = 4) to focus on inter-decadal variations (Bärring and Fortuniak 2009)

the NAO (see also Jones et al. 1997; Slonosky et al. 2000, 2001; Moberg et al. 2005).

Kyselý and Huth (2006, see Fig. 4.3) discussed the intensification of zonal circulation, especially that during the 1970s and 1980s. The stronger zonal circulation does not appear isolated, but coincides with changes in other atmospheric modes. In recent winters, the authors noted an intensification of cyclonic activity over Fennoscandia along with more frequent blocking situations over the British Isles. At the same time, less cyclonic activity is observed over the Mediterranean. While there is a general increase in the zonality of the flow in winter, the opposite appears to occur in summer (Kaszewski and Filipiuk 2003; Wang et al. 2009a).

There are also indications (Kyselý 2000; Werner et al. 2000; Kyselý 2002; Kyselý and Huth 2006) that weather types (as defined, for example, by Hess and Brezowksy 1952) are more persistent than in earlier decades. For all weather types (zonal, meridional or anticyclonic), an increase in persistence of the order of 2–4 days is found from the 1970s to the 1990s. This increase in persistence may be reflected in the increase in the occurrence of extreme events.

Getzlaff et al. (2011) and Lehmann et al. (2011, Fig. 4.4) showed intensified cyclonic circulation and stronger westerlies for the 1990s and 2000s compared to the 1970s and 1980s.

Interpretation of circulation changes must be done with care, and reanalysis products are often used (such as the reanalysis from the National Centers for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) NCEP/NCAR, or the reanalysis of the European Centre of Medium Range weather forecasts; ERA). Despite inhomogeneities in the NCEP/NCAR reanalysis data, both before and after the introduction of satellites as a source of environmental monitoring data in late 1978 (the same holds for ERA products), the results in Fig. 4.4 probably mirror real changes. Lack of data over ocean areas before the introduction of satellites might not introduce major problems since several Ocean Weather Ships (OWS) were on duty after the Second World War in the north-east Atlantic. It appears probable that deep cyclones were identified in particular by OWS 'C' (south of Greenland) or OWS 'M' (east of Greenland). For the Barents Sea region, inhomogeneities in the data cannot be completely ruled out. However, periods P3 and P4 (see Fig. 4.4 for definition), both occurring after the transition to satellites, should be directly comparable.

Jaagus (2006) investigated the large-scale circulation over Estonia during the second half of the twentieth century and found a general increase in westerlies, particularly in February and March with a decrease in May. Such an increase

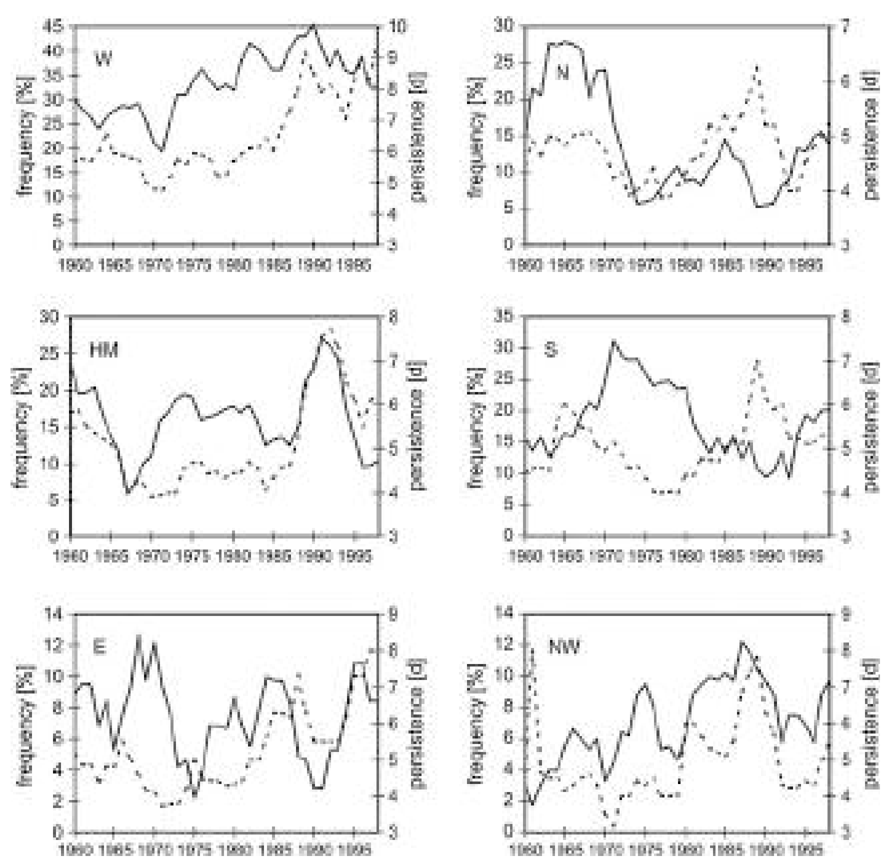


Fig. 4.3 Temporal changes in the relative frequencies of occurrence (in %; *solid curve*) and mean lifetime (in days; *dashed curve*) of groups of large-scale circulation patterns ('Großwetterlagen'; GWL) in winter in the period 1958–2000. Five-year running means are shown. The *capital letters* indicate the circulation pattern: *W* (westerly), *N* (northerly), *S* (southerly), *E* (easterly), *NW* (north-westerly) and *HM* (anticyclonic; 'Hoch Mitteleuropa'). GWL are defined in detail by Hess and Brezowski (1952)and Kyselý and Huth (2006)

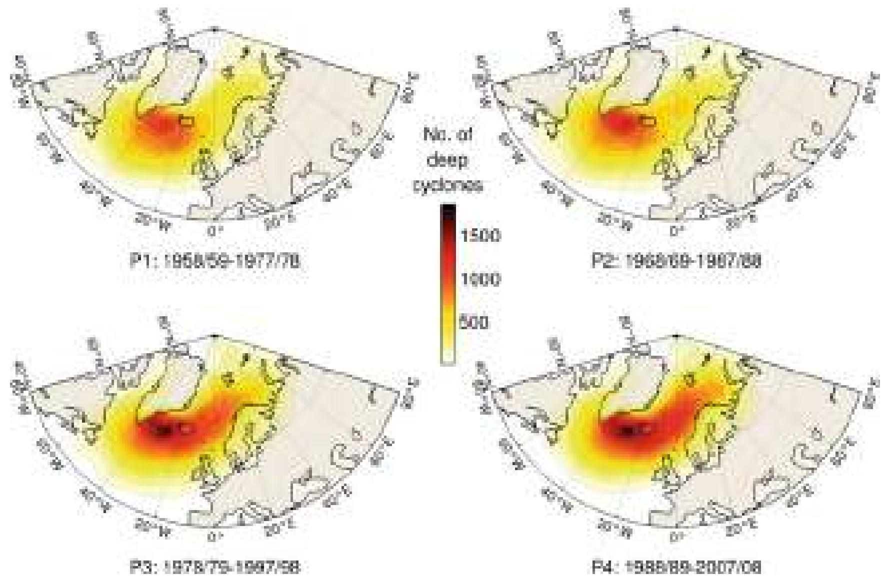


Fig. 4.4 Total number of deep (core pressure < 980 hPa) cyclones counted for four 20-year periods P1–P4, based on NCEP/NCAR reanalysis four times daily SLP data for winter (DJFM). Unit is number of deep cyclones, with an increment of 100 (Lehmann et al. 2011). See discussion of inhomogeneities in Sect. 4.2

may have caused additional coastal erosion along the eastern margin of the Baltic Sea (see Chap. 20), as well as changes in other parameters in the region (Klavinš et al. 2007, 2009; Valdmann et al. 2008; Draveniece 2009; Rivza and Brunina 2009; Avotniece et al. 2010; Klavinš and Rodinov 2010; Lizuma et al. 2010).

4.2.2 Long-Term Circulation Changes

There are a large number of studies discussing the influence of long-term change in atmospheric circulation on surface characteristics of the Baltic Sea region. Early publications, for example, by Tinz (1996), Chen and Hellström (1999), Koslowski and Glaser (1999), Jevrejeva (2001), Omstedt and Chen (2001) and Andersson (2002) agreed that there has been a north-eastward shift in low-pressure tracks, which is consistent with a more zonal circulation over the Baltic Sea basin and the observed trend of a more positive NAO index, at least up to the 1990s (Trenberth et al. 2007). A northward shift in low-pressure tracks is also consistent with model projections of anthropogenic climate change, as pointed out by Leckebusch and Ulbrich (2004), Bengtsson et al. (2006), Leckebusch et al. (2006), Pinto et al. (2007) and, more recently, Lehmann et al. (2011).

Jacobeit et al. (2001, 2003) and Hurrell and Folland (2002) discussed the strong temporal variability in the relationship between the general circulation of the atmosphere and surface climate characteristics over the past 300 years. Their studies suggested that the increased frequency of both anticyclonic circulation and westerly wind types result in a warmer climate with reduced sea-ice cover and a reduced seasonal amplitude in temperature. Their studies concluded that long-term (multi-decadal) climate change in the Baltic Sea region is at least partly related to changes in atmospheric circulation.

Omstedt et al. (2004) made a thorough investigation of the past 200 years of climate variability and changes based on the long Stockholm time series of temperature and sea level as well as ice cover and circulation types based on pressure data (Fig. 4.5; see also Chap. 9 for further discussion). Over the entire period, the authors found positive trends in temperature and sea level, increased frequencies in both westerlies and anticyclonic circulation and negative trends for the amplitude of the seasonal temperature cycle and sea-ice cover. Increased westerlies indicate a stronger than normal zonal flow with a positive NAO index, whereas anticyclonic circulation indicates a north-eastward movement of the low-pressure tracks. This is consistent with the observed upward trend in the NAO index (Hurrell and Folland 2002) and circulation changes as reported by Jacobeit et al. (2003). Eriksson et al. (2007) and Eriksson (2009) extended the analysis of Omstedt et al. (2004) by examining the covariability of long time series from the Baltic Sea region over different timescales during boreal winter. Over a period of 500 years, 15 periods with a clearly distinct climatic signature with respect to circulation patterns, inter-annual variability and the severity of winters were identified (see Chap. 3, Fig. 3.10). The onsets of these periods appear to have been mainly driven by internal perturbations, although volcanic activity and solar variability may also have played a role at certain times. The analysis indicates a clear increase in mean and maximum temperatures beginning at the end of the nineteenth century. The seasonal index (i.e. the magnitude of the annual temperature amplitude) shows a negative trend. Further inspection reveals that the frequency of both westerlies and anticyclonic circulation is considerably higher in the twentieth century than in the nineteenth century.

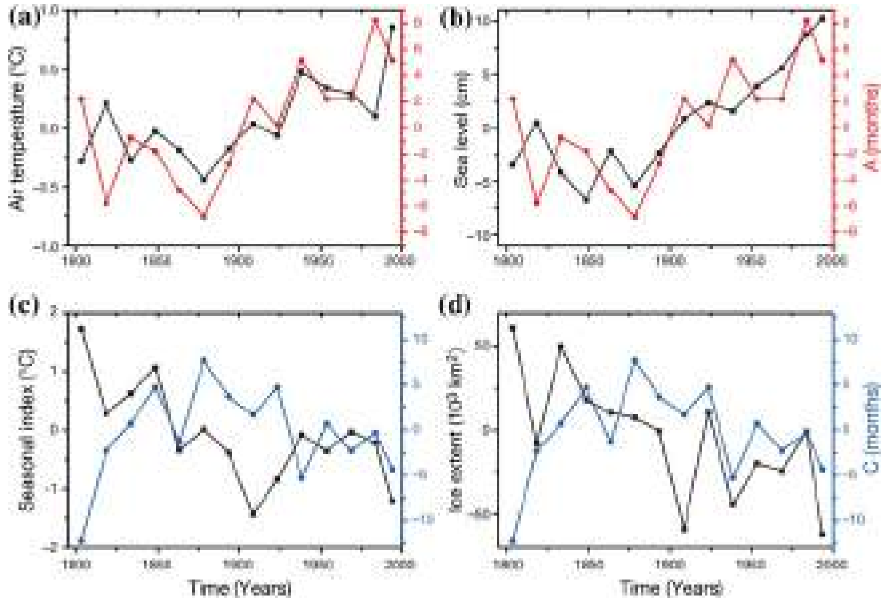


Fig. 4.5 Anomalies in the Stockholm climate record together with the circulation types that describe the vorticity of the atmospheric circulation. *Red* indicates anticyclonic circulation and *blue* cyclonic circulation. **a** Air temperature and anticyclonic circulation, **b** sea level and anticyclonic circulation, **c** seasonal index, defined as the difference between summer (JJA) and winter (DJF) seasonal temperatures, and cyclonic circulation, and **d** ice cover and cyclonic circulation. 15-year averages for 1800–1815, 1811–1825, 1826–1840…, 1961–1975, 1976–1990, 1986–2000 (Omstedt et al. 2004)

Sepp (2009) examined the increase in cyclonic activity and the frequency of westerlies over the Baltic Sea basin during the twentieth century and the tendency for increased cyclogenesis. In recent years, an increase in the percentage of deep cyclones has been observed, while the total number has not changed. There is also a dependence on the NAO: during its positive phase, less, but stronger cyclones form over the Baltic Sea region.

4.2.3 NAO and Blocking

Blocking of the atmospheric flow is frequently observed in the Baltic Sea region. Since blocking situations, once they have developed, are often quasi-stationary and can persist for extended periods, they are often responsible for extreme weather events and have quite early raised the interest of scientists (for example Namias 1947; Rex 1950b; Green 1977). Figure 4.6 gives an example of a blocking pattern over central Europe.

Rex (1950a) subjectively defined a blocking event as a quasi-persistent (more than 10 days) split of the mid-tropospheric flow over more than 45° in longitude. Numerous authors have suggested modifications to this definition, including objective measures based on meridional height gradients. These approaches were reviewed by Barriopedro et al. (2006). Vial and Osborn (2012) discussed the poor performance of models with respect to simulating number,

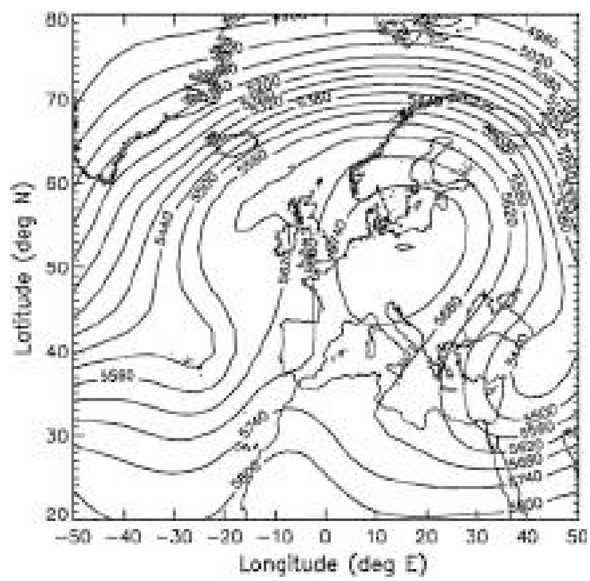


Fig. 4.6 The 500 hPa height field on 6 March 1948, showing a typical blocking situation (Barriopedro et al. 2006)

frequency and spatial extent of blocking situations, a problem that had persisted for many years (d'Andrea et al. 1998).

Rimbu and Lohmann (2011) used south-western Greenland temperature measurements and stable isotope records from ice cores as a proxy for North Atlantic atmospheric blocking and found that in winter, warm (cold) conditions over south-western Greenland were related to high (low)

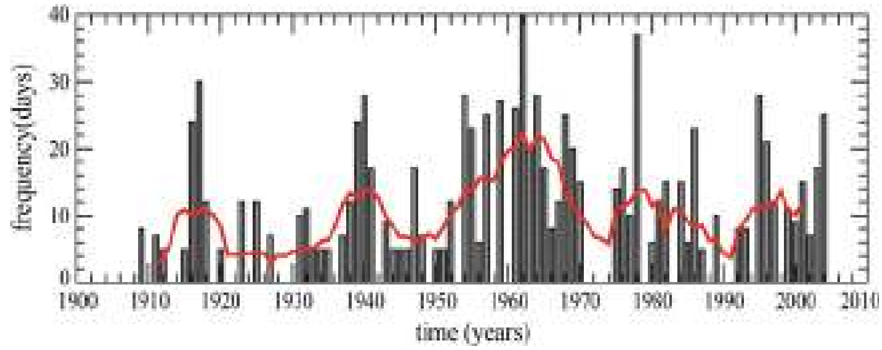


Fig. 4.7 Blocking index (bars) and its decadal variation (seven-year running mean; *red*) for boreal winter (DJF) 1908–2005. The blocking index takes into account spatial aspects and persistence, and it is defined as the number of blocked days per winter in the sector 80°–10°W. The blocking condition must be satisfied for an interval of at least 12.5° for at least five consecutive days (persistence criteria), Rimbu and Lohmann (2011)

blocking activity and a negative (positive) phase of the NAO. For summer, however, the authors found the opposite, that warm (cold) conditions over south-western Greenland were related to low (high) blocking activity and a positive (negative) phase of the NAO, even though a significant part of the North Atlantic blocking variability was not directly related to NAO variability, but rather to the exact position of the centre of blocking, which, in turn, did show dependence on the NAO phase. Furthermore, it is well known (e.g. Luo and Wan 2005; Barriopedro et al. 2006) that the frequency of blocking exhibits considerable inter-decadal variation. Rimbu and Lohmann (2011) constructed a North Atlantic blocking index (Fig. 4.7) which shows pronounced decadal variations with frequent blocking in the 1910s, 1940s and 1960s as well as after 1995, and low blocking particularly in the 1920s, 1950s, 1970s and early 1990s, in good agreement with the observed temperature anomalies in the Baltic Sea region during the twentieth century. The relationship, first discussed by van Loon and Rogers (1978), also holds further back in time; very mild south-western Greenland winter temperatures during the Late Maunder Minimum (late seventieth and early eightieth centuries, see Chap. 3, Sect. 3.5) coincides with above normal blocking frequency over Europe, cold winters and above (below) normal pressure over northern (southern) Europe (Luterbacher et al. 2001) and above normal sea ice (Koslowski and Glaser 1999). It has also been possible to simulate these changes in blocking frequency in reconstructed (Casty et al. 2005) and model data (Stendel et al. 2006).

4.2.4 Distant Controls of Circulation Changes

There are also indications that circulation changes in the Baltic Sea region are related to climate anomalies at further distances. Several authors have addressed the question whether the NAO is influenced by ENSO (El Niño/Southern Oscillation). Since there is no significant correlation between the two indices, the effects seem to be small (e.g. Rogers 1984; Pozo-Vázquez et al. 2001; Sutton and Hodson 2003). However, it can be expected that the influence of ENSO on the European climate is nonlinear and so should be analysed in terms of composites of strong anomalies of ENSO (Brönnimann et al. 2007). In this way, in periods of pronounced La Niña or El Niño, the European climate can be indirectly influenced by ENSO through teleconnections via a downstream propagation of tropical disturbance from the Pacific to the North Atlantic (Fraedrich 1994) to the stationary Rossby waves. Another indirect link exists via the effect of ENSO on tropical North Atlantic temperatures (e.g. Chiang et al. 2002). Jevrejeva et al. (2003) discussed the influence of the Arctic Oscillation (of which the NAO can be regarded as the European/North Atlantic part) and of ENSO on ice conditions in the Baltic Sea and found a weak, but non-negligible contribution from the latter. García-Serrano et al. (2011) applied a principal component analysis to the North Atlantic/European winter 200 hPa stream function and found a discernible El Niño signature. In contrast, von Storch (1987) did not find a robust ENSO signal in winter. Thus, many different mechanisms for controlling European climate have been suggested, but with little predictive skill.

Graf and Zanchettin (2012) discussed the effects of El Niño on North Atlantic/European climate. Distinguishing between ‘central Pacific’ and ‘east Pacific’ El Niños, they found a teleconnection via a ‘tropospheric bridge’ between the latter and cold European winters. Seager et al. (2010) related positive snowfall anomalies in the Arctic to excess moisture due to anomalously warm conditions in the preceding summer and autumn and stated that most models are unable to capture this wintertime cooling due to their poor representation of snow cover variability. Stroeve et al. (2011) and Jaiser et al. (2012) showed from ERA-Interim data that low ice concentrations over the Arctic Ocean lead to an increase in heat released into the atmosphere and, as a consequence, to a reduction in vertical static stability, leading to circulation anomalies over Europe in winter that resemble the negative phase of the NAO. In contrast, Ineson et al. (2011) related weaker winter westerlies and a negative NAO phase-like pattern to a minimum in solar ultraviolet (UV) irradiance. If this finding proves correct, it implies that low solar activity drives a cold winter in northern Europe and the United States.

Overland and Wang (2010) found a relationship between changes in atmospheric circulation in the Baltic Sea region and the loss of sea ice in the Arctic. Triggered by a reduction in Arctic summer sea ice caused by anomalous meridional flow, the resulting additional heat stored in the Arctic Ocean due to the increase in late summer open water area contributed to an increase in the lower tropospheric relative topography (500/1000 hPa), but not necessarily to changes in SLP. As a consequence, anomalous easterly winds were observed in the lower troposphere along 60°N in many regions, including northern Europe and the Baltic region. This is in contrast to early findings by Glowienka-Hense and Hense (1992), who concluded that Arctic sea-ice variability may have an effect on mid-latitude circulation through synoptic transient eddy forcing.

More specifically, Petoukhov and Semenov (2010) performed a series of experiments with the ECHAM5 model at low resolution (T42, i.e. approximately 300-km grid point spacing with 19 vertical levels) and found a dependence of central European winter temperatures from sea-ice cover in the Barents and Kara Seas. A gradual decrease in sea-ice cover from 100 % to ice-free conditions led to a strong temperature increase, and via a nonlinear relationship between convection over the ice-free parts and baroclinic effects triggered by changes in temperature gradients near the surface heat source, this resulted in a warming, then a cooling and at very low ice cover, again a warming over central Europe. Yang et al. (2011), using the EC-Earth model (Hazeleger et al. 2012) with considerably higher resolution (T159, i.e. approximately 80 km between grid points, with 31 vertical levels), confirmed a decrease in winter temperature with decreasing sea ice in the Barents and Kara Seas, but in a more linear way than by Petoukhov and Semenov (2010). In the light of the record, low Arctic ice cover and recent cold winters over Europe, these are interesting findings. As a consequence, transitions between different regimes of the atmospheric circulation in the subpolar and polar regions may be very likely. Mesquita et al. (2010) even found a connection to positive sea-ice anomalies in the Sea of Okhotsk via a westward shift in cyclosis and the build-up of a pattern resembling the negative phase of the NAO over the North Atlantic.

Many other authors have also discussed the low temperatures of the 2009/2010 and 2010/2011 winters over large parts of Europe (including the Baltic Sea region). Taws et al. (2011) observed a tripole pattern in sea-surface temperature (SST) anomalies related to a negative NAO phase. Guirguis et al. (2011) and Cattiaux et al. (2010) argued that only parts of Europe, Russia and the United States experienced cold anomalies, while extreme warm events were observed at several other locations in the Northern Hemisphere, thus providing a consistent picture of a regional cold event under global warming conditions. Cattiaux et al. (2010) and Ouzeau et al. (2011) highlighted the importance of an adequate representation of the stratosphere in the ARPEGE model for reproducing the cold anomalies over Europe. Cohen et al. (2010) related stratospheric temperature anomalies and their interaction with the troposphere. On the other hand, Jung et al. (2011) stated that internal atmospheric dynamic processes were responsible for the extended negative NAO phase in 2009/2010. Stroeve et al. (2011) argued that negative AO indices and corresponding low sea-ice volumes at the beginning of the melt season result in the summer melt of much of the multi-year sea ice due to ice transport into warmer southerly waters related to the atmospheric circulation anomalies. Some studies also suggest a link between autumn snow cover in Eurasia and Northern Hemisphere winter circulation (see Chap. 6).

4.2.5 Controls of the NAO

A spectral analysis of the NAO time series revealed little evidence for the NAO index to vary on any preferred timescales (Hurrell and Deser 2009). There were large changes from winter to winter and even within a season, but a decadal signal was also visible. For example, high NAO index values prevailed during the 1920s, while the 1960s were characterised by low values. Very high values were observed in the 1990s, together with a north-eastward displacement of the centres of action (Hurrell and van Loon 1997). Whether this is related to anthropogenic climate change or to what extent the NAO might change in the future due to global warming is still a matter of conjecture. The Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) (Meehl et al. 2007) stated that '… the most consistent results from the majority of the current generation of models show, for a future warmer climate, a poleward shift of storm tracks in both hemispheres…'. Ulbrich et al. (2009) concluded that most models agree that there are fewer, but more intense cyclones in many parts of the extra-tropics, including the North Atlantic/European region. They also noted that this conclusion can only be drawn when 'extreme' is defined as a function of core pressure, whereas no such increase (actually, a slight decrease in several models) is found when 'extreme' is instead defined from pressure gradients.

4.2.6 Circulation Changes in Contrast to Global Warming

In two recent articles, Bhend and von Storch (2008, 2009) presented a method to compare the consistency of observed trends with climate change projections, even if no estimates of natural variability exist. They found that anthropogenic forcing can explain a large part of the observed changes in temperature and precipitation over the Baltic Sea region and that this correlation is unlikely to be occurring by chance. However, it cannot fully explain the observed trends. Since, due to its stochastic nature, a relatively large part of the NAO could be unrelated to anthropogenic climate change, the NAO signal was removed and the analysis was repeated. The results indicate that the climate change signal in temperature and precipitation is robust with respect to the removal of the NAO for long-term means, whereas seasonal as well as spatial variability is underestimated. This may be due to additional forcing mechanisms not included in their model set-up (e.g. the indirect aerosol effect) or to a general underestimate of the model response to anthropogenic forcing (see Chap. 10 as well as Chaps. 23–25).

4.3 Surface Pressure and Winds

The wind climate, described through the statistics of near-surface wind speed and direction, has a strong impact on human activities and the Baltic Sea ecosystem. Extreme wind speeds are a direct threat to life and property and an indirect threat through wind waves, storm surges (Chap. 9) and coastal erosion (Chap. 20) leading to high economic loss. However, on the European scale at least, no trends were found for storm losses adjusted for inflation and changes in population and wealth in the period 1970–2008 (Barredo 2010). Nilsson et al. (2004) calculated a storm damage index for Sweden for the period 1901–2000 based on storms sufficient to cause forest damage. Although the 1980s suffered most extreme storm events in terms of windthrow, the authors noted several factors other than wind that increased or decreased storm damage. Widespread and severe damage usually relates to severe winter storms. A positive NAO index is generally associated with an increased number of extreme cyclones although they can also occur at negative phases of the NAO (Pinto et al. 2009). Among others, typical examples of severe winter storms causing widespread damage in the last decade have been Gudrun/Erwin on 8/9 January 2005 (Haanpää et al. 2006; Suursaar et al. 2006) and Kyrill on 18/19 January 2007 (Fink et al. 2009). Negative economic effects can also result from unusually calm conditions, especially for activities with an increasing dependence on wind energy.

Storms are also an essential factor for ventilation and mixing of the strongly stratified Baltic Sea. Inflow events from the North Sea importing salt and oxygen into the Baltic Sea basin are highly dependent on the wind climate and atmospheric pressure differences (Lass and Matthäus 1996; Gustafsson and Andersson 2001) and have a strong impact on the Baltic Sea ecosystem (see also Chap. 7). Warm water inflows into the Baltic Proper in summer indicate that pressure systems and wind conditions in summer also play a vital role (Feistel et al. 2004).

4.3.1 Wind Climate in Recent Decades

The temporal and spatial covariance of the wind climate is generally related to large-scale variations in atmospheric circulation over the North Atlantic and in winter to the NAO. Hence, changes in the synoptic-scale wind climate over the Baltic Sea region are closely related to variability in atmospheric circulation, baroclinic activity and changes in the North Atlantic storm tracks.

During the latter half of the twentieth century, the wind climate over the north-east Atlantic and northern Europe underwent large changes. Based on NCEP/NCAR reanalysis data (Kalnay et al. 1996; Kistler et al. 2001), the number of deep cyclones (core pressure < 980 hPa) in winter (DJFM) reached a minimum in the early 1970s and increased over the following decades peaking around the last decade of the twentieth century (Lehmann et al. 2011, Fig. 4.4). At the same time, a continuous north-eastward shift in the storm tracks regionally increased the impact and number of storms over northern Europe and thus the Baltic Sea in winter and spring, although there was a decrease in autumn (Fig. 4.8).

Consequently, a strong increase in storminess in the 1980s and 1990s across the North Sea (e.g. Carter and Draper 1988; Hogben 1994) raised public concern about the possible impact of increased greenhouse gas concentrations on the rougher wave and storm climate (Schmidt and von Storch 1993). Based on high-resolution meteorological data from SMHI for the period 1970–2007, regional changes were also found in the wind climate over the Baltic Sea region (Lehmann et al. 2011). Although wind speeds returned to average values by the last decade of the investigated period, there was a clear increase in mean geostrophic wind speed of 1.5 ms^{-1} for the period 1989–2007 compared to 1970–1988 in the southern and central Baltic Sea region in winter (DJF). This coincided with an increase in the number and spatial extent of deep lows (Fig. 4.8) over

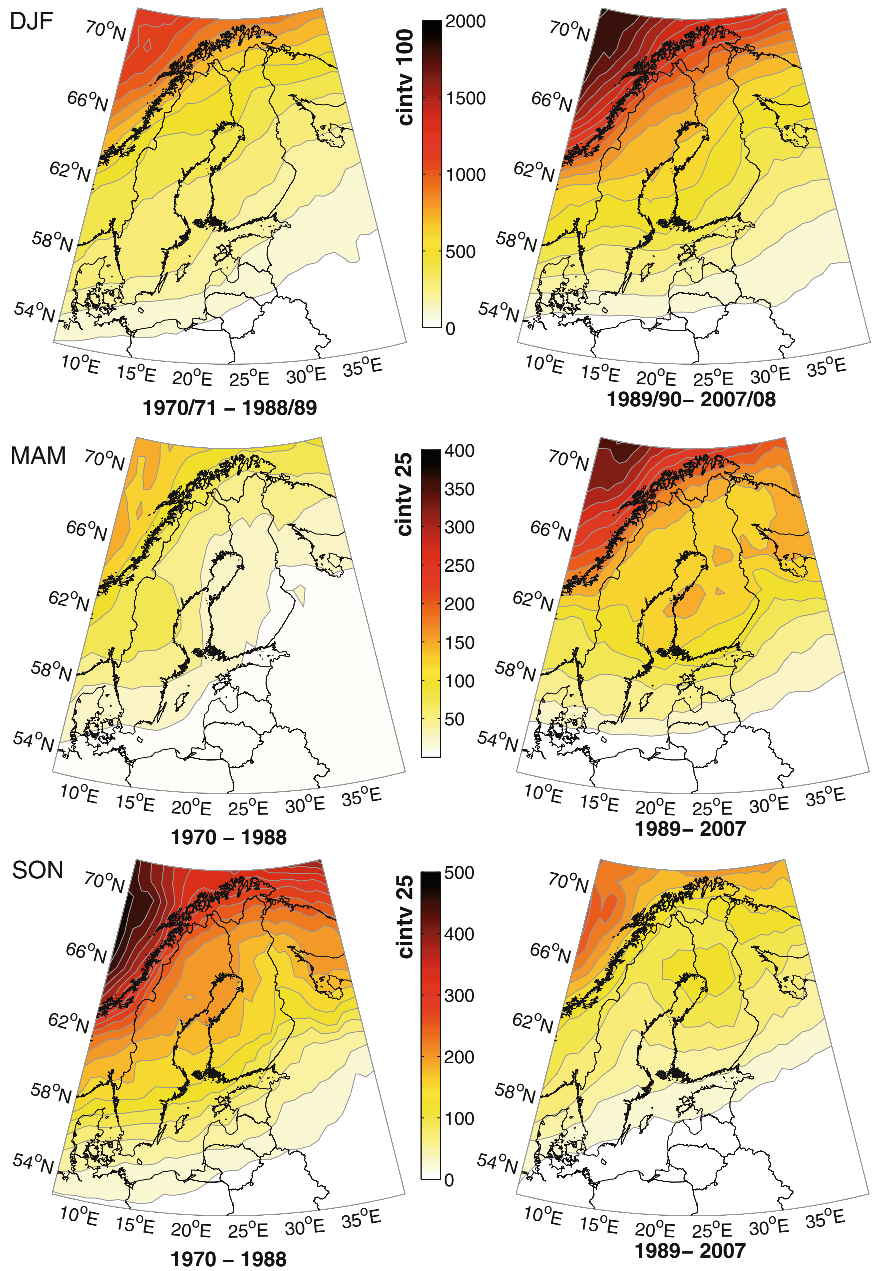


Fig. 4.8 Changes in the number of deep cyclones (core pressure < 980 hPa) between 1970–1988 and 1989–2008 over the Baltic Sea region for winter (DJF), spring (MAM) and autumn (SON) from SMHI data (Lehmann et al. 2011). Note different scales and contour intervals for the different seasons. Scale (contour interval, cintv) refers to the number of cases below 980 hPa

the Baltic Sea region. While the increase in mean geostrophic wind speeds in winter over the Bothnia Bay was only 0.5 ms^{-1} over this period, a general increase of 0.5–1 ms^{-1} took place over most areas in spring (MAM) together with a change to more westerly than south-westerly winds.

Comparable shifts for early spring were also reported for Finland by Keevallik and Soomere (2008) and Keevallik (2011) from the 1960s to 1990s with changes to more westerly than north-westerly winds. For the period 1966–2011, Jaagus and Kull (2011) also found a clear change in the main wind direction over Estonia in winter changing from south-east in the 1970s to south-west in the last decade. A general tendency towards more zonal and less meridional flow in winter is also confirmed for the easternmost Baltic Sea region for the period 1961–2003 accompanied by increasing variability (Khokhlova and Timofeev 2011).

In contrast, wind speeds in autumn (SON) decreased over the western and central Baltic Sea (by 1.5–2 ms^{-1}) and the Bothnia Bay (by 0.5 ms^{-1}) explained by a general decrease in the number and spatial extent of deep lows in 1989–2007 compared to 1970–1988 (Fig. 4.8). Over the Kiel Bight, this change in strong wind speeds (>13.9 ms^{-1}) is accompanied by a marked change in the frequency distribution of wind direction with a decrease in south-westerly and an increase in the easterly component of the winds (Lehmann et al. 2011).

The increase in mean wind speed since the 1960s and 1970s is accompanied by a relative increase in the frequency of storms over the southern North Sea and Baltic Sea ($\sim$1–2 % per year) in the period 1958–2001 based on numerically downscaled NCEP/NCAR reanalysis data (Weisse et al. 2005). Including also the Norwegian Sea, the positive trends in storm frequency were found to be statistically significant ($p < 0.05$). Although high annual geostrophic wind speeds (above the 99th percentile) returned to average or calm conditions over the north-east Atlantic, central Europe, the North Sea and the Baltic Sea at the end of the twentieth century (Matulla et al. 2008), there has been an upward trend in winter storminess for the past 50 years over northern Europe (Donat et al. 2011). Whether this trend is likely to persist over the longer term or is due to large (multi-)decadal variability is addressed in the following section.

4.3.2 Long-Term Wind Climate

Long data series of direct wind observations are sparse, and most measurements even in recent decades suffer from potential inhomogeneities due to changes in the environment (growing trees or new buildings in the vicinity, station relocation, etc.) or changes in methodology (different instruments, number of measurements per day, etc.) as discussed by the group '*Waves and Storms in the North Atlantic*' (WASA 1998; von Storch and Weisse 2008; Lindenberg et al. 2012). Also, cyclone detection and tracking algorithms to derive the frequency and intensity of deep lows as a proxy for storminess from historical pressure fields face the problem of lower data density and quality back in time (Smits et al. 2005), possibly leading to an apparent increase in high-latitude cyclone activity that is actually due to higher data density (see Sect. 4.3.4).

As synoptic-scale storms are generally linked to large-scale forcing over the pressure field, pressure gradients can be used to derive geostrophic wind speeds based on surface pressure readings (Krueger and von Storch 2011). A first study based on geostrophic wind speeds calculated from a triangle of station pressure over the German Bight by Schmidt and von Storch (1993) showed no long-term trend for the wind climate of 1876–1990. High annual wind speeds in the 1990s appeared to be comparable to high wind speeds in the 1880s as well as in the early and mid-twentieth century. Kaas et al. (1996) found no overall trends but considerable decadal variability. This was further confirmed by studies from the WASA group (WASA 1998), Alexandersson et al. (2000) and Matulla et al. (2008) using different pressure triangles over Europe, mainly the North Sea. For Finland, Suvilampi (2010) found a slight decrease in annual geostrophic wind speeds since 1884 and a weak but non-significant upward trend for the past 50 years. The tendency for a long-term decrease in annual wind speeds is also confirmed by Wern and Bärring (2009) for southern Sweden based on geostrophic wind speeds derived from pressure triangles. For the period 1901–2008, the authors found statistically significant negative trends in annual potential wind energy and mean and extreme (>25 ms^{-1}) geostrophic wind speeds. For the shorter period, 1951–2008, a tendency to negative trends in mean wind speeds was found for northern Sweden, while weak non-significant trends of both signs were found for central and southern Sweden. In general, the authors concluded that (multi-)decadal scale variations dominate rather than any long-term trends.

While the decrease in storminess from a peak around the 1880s happened quite suddenly in central Europe, there was a gradual slow-down over a period of decades in northern Europe until the 1960s (Figs. 4.8 and 4.9 in DJF). Matulla et al. (2008) found the increase in storminess starting in the late 1970s was most pronounced in NW Europe and more steady in central Europe. They also found general agreement between storminess over central Europe and NW Europe despite some difference in timing and/or magnitude. Bärring and Fortuniak (2009) also showed a correlation between inter-decadal variations over southern Scandinavia and similar variations over NW Europe.

Another way to estimate historical storminess is by using pressure-based single-station proxies such as different pressure tendencies per unit time, mean or low percentiles of surface pressure or, for example, the annual number of deep lows. Based on different storm indices derived from single-station pressure readings for Lund and Stockholm, Bärring and von Storch (2004) and Bärring and Fortuniak (2009) found no robust signs of any long-term trend in southern Sweden for the period 1780/1800 to 2005. Hanna et al. (2008) found similar results based on a daily pressure variability index calculated as absolute 24 h pressure differences, that is $\Delta p = |p_{t+24} - p_{t+0}|$, for the British Isles since 1830 and for Denmark since 1874 confirming increased storminess at the end of the nineteenth century and the 1980s to 1990s, with the 1880s being the stormiest decade. The informational value of five different pressure-based storminess indices including those used by Bärring and von Storch (2004) and Hanna et al. (2008) was evaluated by Krueger and von Storch (2012). The authors confirmed the general usefulness of the indices as storminess proxies, with absolute pressure tendencies per six or eight hours containing the highest informational value.

Schenk and Zorita (2011) released a new reconstruction of HIgh RESolution Atmospheric Forcing Fields (HiResAFF) for northern Europe for the period 1850–2009

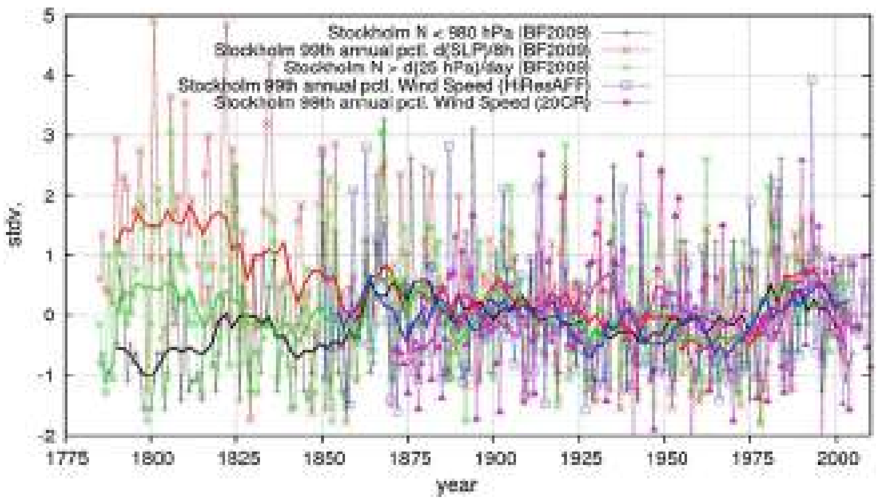


Fig. 4.9 Storminess indices of the annual number of deep lows (N < 980 hPa), the 99th percentile of pressure tendency per 8 h, the annual number of days exceeding a pressure tendency of 25 hPa for the Stockholm station 1785–2005 (Bärring and Fortuniak 2009) compared to the reconstructed annual 99th percentile of wind speeds in the vicinity of Stockholm 1850–2009 from HiResAFF (Schenk and Zorita 2011, 2012). Data normalised with respect to the period 1958–2005. *Bold lines* represent the 11-year running mean to highlight decadal variability

including wind. Based on the pattern similarity between daily SLP station data starting in 1850 and SLP observations since 1958, Schenk and Zorita (2012) reconstructed historical atmospheric fields by taking the daily atmospheric fields of regionally downscaled ERA reanalysis for any day for which the pattern similarity is maximised for an analogous day in 1958–2007. As shown in Fig. 4.9, the reconstructed 99th percentile of annual wind speeds from HiResAFF in the vicinity of Stockholm gives comparable results regarding long-term features of annual storminess derived from single-station proxies of Stockholm used by Bärring and Fortuniak (2009). The different storminess measures agree in showing increased annual wind speeds in the 1880s and 1990s and an unusually calm period around the 1960s to 1970s and a return to average conditions in recent years although the number of deep lows does not indicate calm conditions. Figure 4.9 also confirms the gradual decline in wind speeds since the end of the nineteenth century as reconstructed by Matulla et al. (2008) for northern Europe.

As discussed by Bärring and Fortuniak (2009), up to eight different proxies for storminess calculated from single-station pressure data represent different aspects related to storminess. Estimating the covariance over all indices, the derived first principal component (PC1) shows good agreement with the 99th percentiles of the geostrophic wind index from Trenberth et al. (2007) and the reconstructed NAO index from Luterbacher et al. (2002) with respect to long-term variability (see Fig. 4.1). In contrast to the number of deep lows (N < 980 hPa) in Fig. 4.9, the PC1 over all eight indices captures the calm period of the 1960s and 1970s indicating that it is better to use a number of different indices rather than relying on only one. The highest correlation between HiResAFF annual extreme wind speeds and single-station proxies was achieved for the pressure tendency over 8 h (r = 0.50) confirming the work of Krueger and von Storch (2012). Remarkably high values for the 8-h pressure tendencies on the one hand and very low values for the number of deep lows on the other hand indicate low confidence in the data before around 1850 probably due to irregular pressure readings. As irregular sub-daily observations hamper the detection of deep lows or pressure changes over 6 h, the estimate using annual numbers of days exceeding a pressure change of 25 hPa per 24 h in Fig. 4.9 (green line) is likely to be more reliable prior to around 1850 as only one observation per day is required. Also, the large differences between the Stockholm and Lund time series in the early historical period should be noted with care (Bärring and von Storch 2004).

While the previous studies analysed historical storminess on an annual basis only, Wang et al. (2009a) repeated and updated (1874–2007) previous studies based on the 99th percentiles of geostrophic wind speed over the NE Atlantic, and northern and central Europe and focused more on seasonal and regional differences. They found that the maxima in the 1990s were due to winter storminess, while the high annual storm values in the 1880s were mainly due to summer storminess. For the period 1878–2007, Wang et al. (2011) found weak negative trends in the 99th percentiles

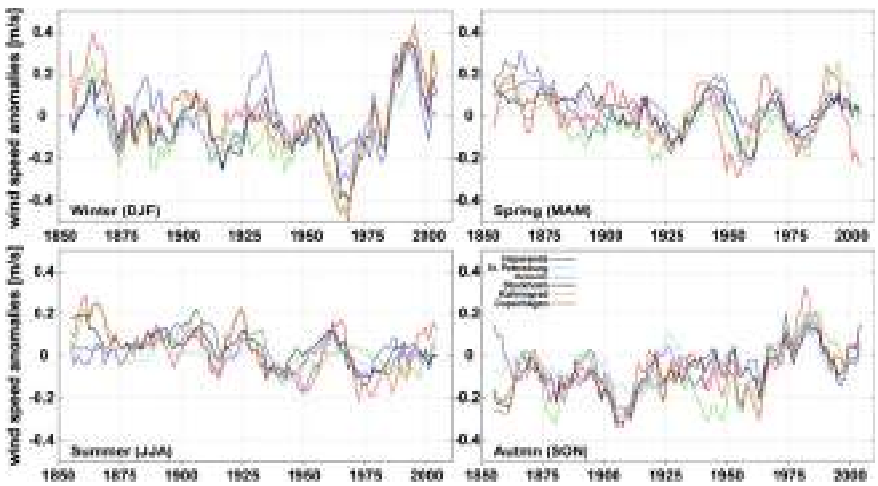


Fig. 4.10 Sliding decadal (11-year) mean seasonal wind speed anomalies for the Baltic Sea region for 1850–2009. Anomalies are calculated by subtracting the mean for 1958–2007. Time series are drawn from the gridded fields of HiResAFF (Schenk and Zorita 2011, 2012). Grid points are selected in the closest vicinity of Haparanda, St Petersburg, Helsinki, Stockholm, Kaliningrad and Copenhagen

over central Sweden and the south-western Baltic Sea in winter (DJF) and a significant ($p < 0.05$) negative trend over the south-western Baltic Sea in summer (JJA).

As shown in Fig. 4.10, HiResAFF confirms decreasing seasonal mean wind speeds in summer (Wang et al. 2011) and the peak in summer wind speeds in the 1880s (Wang et al. 2009a), that is over the southern Baltic Sea region. However, no increased summer winds are reconstructed for the 1880s over the northern and eastern Baltic Sea region highlighting regional differences in the wind climate. While Wang et al. (2009a) attributed high annual storminess in the 1880s mainly to higher storminess in summer, HiResAFF shows higher mean wind speeds in all seasons except autumn over the southern and central Baltic Sea region in the 1880s.

So far, all long-term reconstructions of the wind climate discussed here have been derived from (sub-)daily pressure observations relying on physical (triangle method) and empirical (analogue-upscaling) methods or from pressure tendencies and the number of deep lows as indirect storminess indices. While the reconstructions show good agreement in terms of a dominance of (multi-)decadal variability rather than robust long-term trends in wind speed, a recent study by Donat et al. (2011) differs in showing a significant long-term increase in winter storminess since 1871 for Europe based on the twentieth-century reanalysis (20CR) data (Compo et al. 2011). The model used for 20CR is very similar to those used for NCEP/NCAR reanalysis but uses a different data assimilation technique. Unlike NCEP/NCAR, 20CR uses only daily station SLP monthly SST and sea ice for data assimilation of historical observations since 1871. As the density of stations with daily SLP increases strongly over time, potential users of 20CR should be cautious about whether the 20CR trend is in fact an artefact caused by the lower station density in earlier times (e.g. Krueger et al. 2013) similar to other long-term trends found in reanalysis data subsequently identified as spurious (see Sect. 4.3.4).

4.3.3 Long-Term Trends Versus Decadal Variability

The findings of reconstructions based on geostrophic wind speeds derived from pressure triangles, different storminess proxies using single-station pressure indices and field reconstructions using analogue-upscaling, are in good agreement showing large decadal variability rather than robust trends in storminess over northern Europe since 1850. The 1880s and 1990s show maxima in annual mean and extreme wind speeds, while the 1970s were unusually calm. The past decade shows a return to average conditions, and only the summer wind climate over the southern Baltic Sea region shows a slight negative long-term trend. Studies analysing the wind climate of the past 40–60 years detect large changes in the recent past (Sect. 4.3.1) that are characterised by the rebound from very calm conditions in the 1960s at the beginning of many observational time series for wind, followed by the very stormy 1990s. Hence, while positive trends in this period indeed describe a dramatic change in wind state, the return to average conditions in the past decade and the long-term analysis of the wind climate over more than 150 years clearly commute the decadal trend into (multi-)decadal variability.

The physical explanation for these large changes from the 1970s to the 1990s relates to dynamical changes in the large-scale atmospheric circulation over the North Atlantic and the

NAO. Over this period, the NAO index switched from strongly negative to unprecedentedly high positive values highlighting the strong correlation of storminess with the NAO (Sect. 4.2). The NE shift of the NAO together with the increased pressure gradient over the North Atlantic extended the geographical influence and numbers of deep lows towards the Baltic Sea region (Fig. 4.8, Wang et al. 2006; Lehmann et al. 2011), explaining upward trends in annual and winter to spring storminess from the 1960s to 1990s. However, this relation depends on the region and time period (Matulla et al. 2008), where recent decades show a very high influence of the NAO (Alexander et al. 2005) with a weaker link in previous times (Alexandersson et al. 1998).

Regarding atmospheric circulation and weather type, there is a corresponding change from calm anticyclonic conditions towards more active cyclonic conditions at the end of the twentieth century for the winter season (Hurrell et al. 2003). In addition, the remarkably calm period during 1960s and 1970s coincides with a period of very high Euro-Atlantic atmospheric blocking frequency in winter (e.g. Rimbu and Lohmann 2011, Fig. 4.7) relative to the period 1908–2005, preventing or weakening zonal (westerly) flow and leading to low wind speeds and fewer storms over Scandinavia. In contrast, the 1990s show low blocking and high wind speeds.

The long-term negative trend in the wind climate for summer (Wang et al. 2009a, 2011) over the southern Baltic Sea region agrees with the findings of Kaszewski and Filipiuk (2003). Based on weather type classifications over central Europe for summer 1881–1998, they found a tendency towards less zonal and increased meridional flow which could explain the decreasing wind speeds in summer.

Whether external forcing over the past half century has influenced trends in atmospheric circulation and storminess is difficult to identify due to the large natural variability over the North Atlantic and northern Europe. According to Wang et al. (2009b), combined anthropogenic and natural forcing have had a detectable influence on the pattern of atmospheric circulation during boreal winter showing an upward trend in storminess and ocean wave heights in the high northern latitudes and a decreasing trend in the lower northern latitudes for 1955–2004. Further analysis for the first half of the twentieth century suggests that external forcing is less likely to have been an important factor for surface pressure and storminess (see Chaps. 23–25 for attribution). From the different long-term reconstructions of storminess based on surface pressure observations covering more than 150 years, the wind conditions of recent decades seem not to be unusual and to fall within the large range of natural variations which are to a large extent explained by the NAO.

4.3.4 Potential Inconsistencies in Long-Term Trends

As direct wind observations usually cover limited time periods and/or suffer from strong inhomogeneities in the data (Sect. 4.3.2), many studies rely either on reanalysis data or on different reconstructions derived from pressure observations. As previously discussed, the different pressure-based reconstructions show good overall agreement regarding long-term variations in storminess independent of the method used. The conclusion drawn from these studies—that northern European storminess is dominated by large multi-decadal variations rather than long-term trends—appears robust given that Krueger and von Storch (2011, 2012) also confirmed the informational value of most reconstruction methods used.

The analysis by Donat et al. (2011), however, does not agree with the previous reconstructions in suggesting a significant long-term increase in winter storminess since 1871 for Europe based on the 20CR data (Compo et al. 2011). Assimilating only daily station SLP monthly SST and sea ice from historical observations since 1871, the density of stations with daily SLP strongly increases over time in the 20CR model. As the discrepancy in 20CR compared to other reconstructions reduces in parallel to the increase in number of stations, increasing storminess with time could be an artefact due to the changing station density (Krueger et al. 2013) comparable to other spurious long-term trends found in reanalysis data (cf. Trenberth and Smith 2005 and Hines et al. 2000 in case of the of SLP, Bengtsson et al. 2004; Paltridge et al. 2009; Dessler and Davis 2010). At least average or higher wind speeds in the 1880s (in contrast to what is suggested by 20CR) are supported by direct observations for western Europe (Clarke and Rendall 2011) such as sand dune studies in southern Wales (Higgins 1933) and severe storm analysis by Lamb and Frydendahl (1991). Furthermore, Omstedt et al. (2004) found an unusually high frequency of cyclonic circulation at the end of the nineteenth century with a pronounced peak in cyclonic weather types in 1871–1885 relative to 1800–2000. According to historical weather records of gale days for Scotland, remarkably high values were recorded for 1884–1900 (Dawson et al. 2002) which contrasts with very low storm activity in the 1880s derived from the 20CR model data.

To what extent reanalysis products like ERA40 and NCEP/NCAR might also be compromised by similar problems regarding spurious long-term trends in pressure and wind needs further investigation. In general, variables derived from reanalysis data (wind speeds, pressure etc.) are assumed to be closely co-related to observations through data assimilation into state-of-the-art climate models. Even though reanalysis datasets are often referred to as 'observations', several studies highlight the possibility of detecting spurious long-term trends in reanalysis data caused, for example, by a regionally changing density of assimilated stations over time. As an example, Smits et al. (2005) found no trend in observed storminess over the Netherlands for 1962–2002, in contrast to the trend seen in NCEP/NCAR reanalysis data.

In addition to issues with data assimilation, there is also a resolution issue with the relatively coarse gridded NCEP/NCAR and ERA reanalysis. As might be expected (see Raible et al. 2008), this is more of a problem with NCEP/NCAR (triangular truncation of T62, approximately 1.9° in latitude and longitude) than ERA (for ERA40; T106, approximately 1.1°). Analysing storm tracks over the Euro-Atlantic sector (Trigo 2006) and the cyclone lifetime characteristics of the Northern Hemisphere (Löptien et al. 2008) in NCEP/NCAR and ERA, the results are comparable, or the summer season for the two different reanalysis products. Apart from spatial resolution issues, differences in dynamics, physical parameterisations of the models and assimilation of observations may also play a role in the quality of the data (see Ulbrich et al. 2009 and references therein).

4.4 Surface Air Temperature

4.4.1 Long-term Temperature Climate

Earlier studies detected a significant increase in surface air temperature in the Baltic Sea region during 1871–2004 (BACC Author Team 2008). Rather than showing a steady increase, however, temperature showed large (multi-)decadal variations dividing the twentieth century into three main phases: warming until the 1930s, followed by cooling until the 1960s and then another distinct period of warming during the final decades of the time series. Linear trends of the annual mean temperature anomalies during 1871–2011 were 0.11 °C per decade north of 60°N and 0.08 °C per decade south of 60°N in the Baltic Sea basin (Table 4.1). This is greater than for the trend in global mean temperature, which is about 0.06 °C per decade for 1861–2005 (IPCC 2007). All seasonal trends are positive and significant at the 95 % level, except winter temperature north of 60°N (lower significance due to the large variability). The largest trends are observed in spring (and winter south of 60°N) and the smallest in summer. The seasonal trends are also larger in the northern area, than the southern area. The annual and seasonal time series of surface mean air temperature for the Baltic Sea basin presented by the BACC Author Team (2008) have been updated and are shown in Fig. 4.11. The warming has continued over the past few years during spring and summer in the southern area and in autumn and spring in the northern area, and the winters of 2009/2010 and 2010/2011 were relatively cold.

Similar features are also evident in the long Stockholm temperature series (Fig. 4.12). Based on the same period (1871–2011), the trends and significance resemble those in the Baltic Sea basin north of 60°N.

Long-term variations and trends in the Baltic Sea basin are similar to those for European mean air temperature (Casty et al. 2007). A number of studies show similar warming trends for areas of the Baltic Sea basin and its vicinity: Finland (Tietäväinen et al. 2009), Sweden (Hellström and Lindström 2008, see Chap. 5, Fig. 5.18), Norway (Hanssen-Bauer et al. 2009), Czech Republic (Brázdil et al. 2009), Latvia (Lizuma et al. 2007; Klavinš and Rodinov 2010), Estonia (Kont et al. 2007, 2011; Russak 2009) and for the three Baltic countries together (Kriauciuniene et al. 2012). Long and homogeneous time series of spatial mean air temperature were created for Finland covering 1847–2008 (Tietäväinen et al. 2009). Trends were calculated for three periods: 1909–2008, 1959–2008 and 1979–2008. An increase in annual mean temperature was significant at $p < 0.05$ level for all three periods. Tietäväinen et al. (2009) found significant warming in winter (1959–2008 and 1979–2008), spring (1909–2008 and 1959–2008), summer (1909–2008 and 1979–2008) and autumn (1979–2008). The increase in annual mean air temperature for Finland during 1909–2008 was 0.09 °C per decade. The annual mean temperature time series for Finland agrees with the results in Fig. 4.11. The most significant warming in Finland was typical for spring where the trend value was 0.15 °C per decade during 1847–2004 (Linkosalo et al. 2009).

Temperature increased in south-eastern Norway during the twentieth century and the first decade of the twenty-first century by an average of 0.07 °C per decade (Hanssen-Bauer et al. 2009). However, the long-term temperature trend has varied throughout the century, starting with relatively low temperatures, followed by the well-known increase in the

Table 4.1 Linear surface air temperature trends (°C per decade) for 1871–2011 in the Baltic Sea basin

Datasets	Annual	Winter	Spring	Summer	Autumn
Northern area (north of 60°N)	**0.11**	0.10	**0.15**	**0.08**	**0.10**
Southern area (south of 60°N)	**0.08**	**0.10**	**0.10**	**0.04**	**0.07**

Trends shown in *bold* are significant at the $p < 0.05$ level. The trends were also tested by the nonparametric Mann–Kendall test. The results were consistent with the linear trend test. Data from the CRUTEM3v dataset (Brohan et al. 2006)

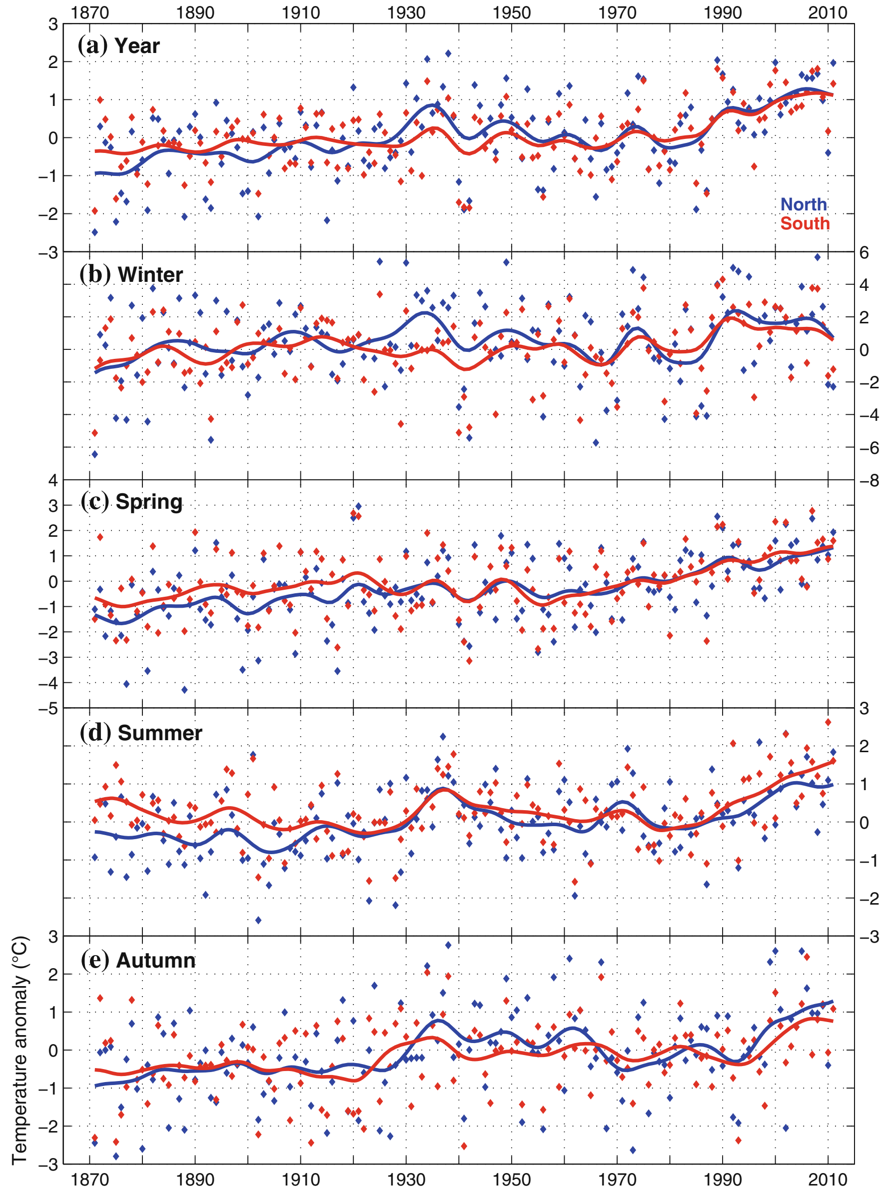


Fig. 4.11 Annual and seasonal mean surface air temperature anomalies (relative to 1960–1991) for the Baltic Sea basin 1871–2011, calculated from 5° by 5° latitude, longitude box average taken from the CRUTEM3v dataset (Brohan et al. 2006) based on land stations (from *top* to *bottom*: **a** annual, **b** winter (*DJF*), **c** spring (*MAM*), **d** summer (*JJA*), **e** autumn (*SON*). *Blue* comprises the Baltic Sea basin north of 60°N and *red* south of 60°N. The dots represent individual years and the smoothed curves (Gaussian filter, σ = 3) highlight variability on timescales longer than 10 years

1920s, referred to as 'the early twentieth-century warming', which ended in the 1930s. After this warm decade, temperature then decreased until the 1980s after which there has been a still ongoing rapid increase. The first decade of the twenty-first century has been the mildest of the series. In terms of seasonal trends, the trend for summer is weaker than for the other seasons. The strongest trend is for spring (Hanssen-Bauer et al. 2009).

The strong warming during spring is also supported by the work of Nordli et al. (2007). Using a composite series for February–April temperature based on ice break-up records on Lake Randsfjorden (1758–1874) and instrumental observation since 1875 (Fig. 4.13), these authors found that the high temperatures in spring and late winter between 1989 and 2011 were unprecedented since 1758.

Long-term series of annual mean air temperature in Rīga, Latvia, indicate a significant warming since 1850 of 1.4 °C (Lizuma et al. 2007). For the shorter period (1948–2006), statistically significant trends ($p < 0.05$) were found at all five stations studied in Latvia (Klavinš and Rodinov 2010). The highest temperature increase occurred in spring and winter. During the whole observation period (1795–2007), mean air temperature at Rīga-University Meteorological Station increased by 1.9 °C in winter, 1.3 °C in spring, 0.7 °C in autumn and 1.0 °C for the year as a whole (Klavinš and Rodinov 2010).

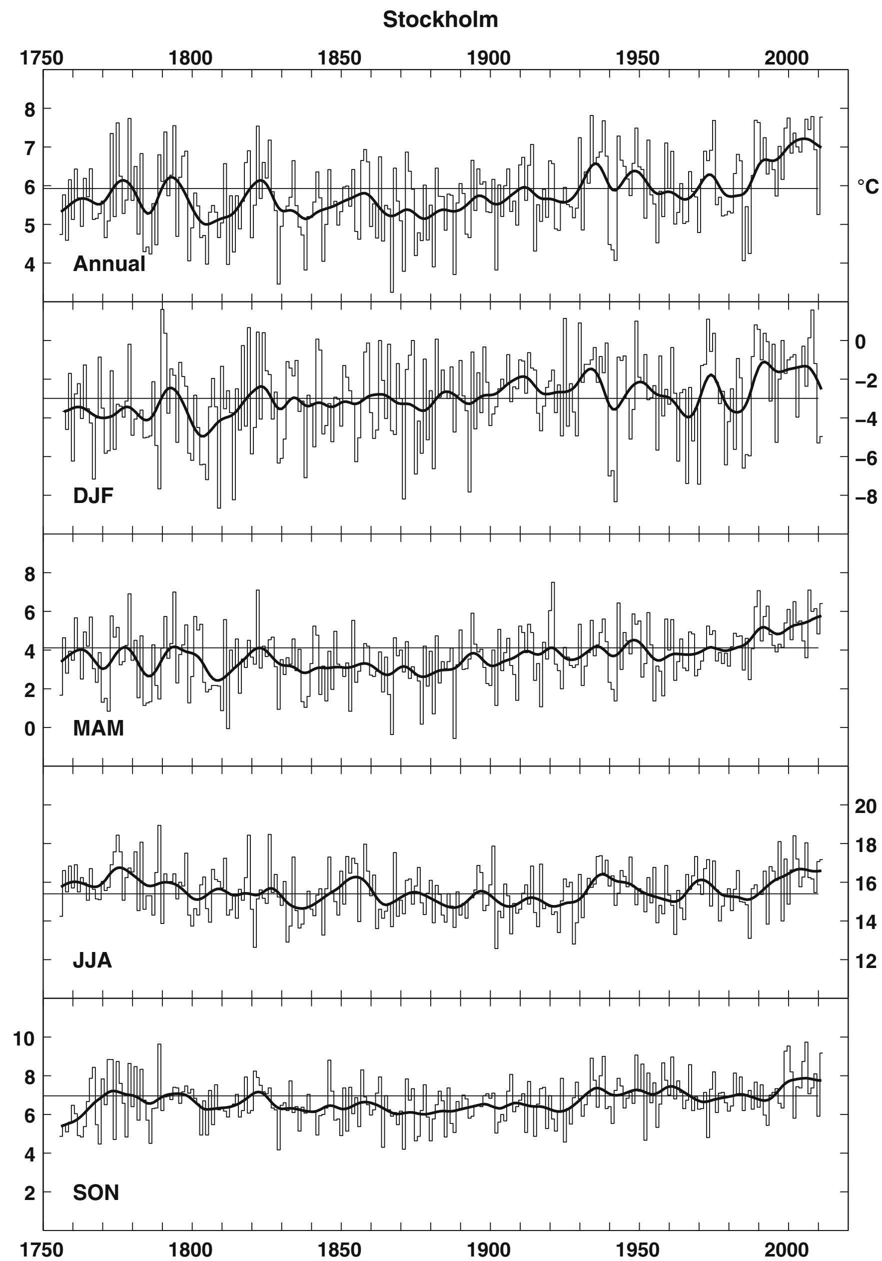


Fig. 4.12 Annual and seasonal mean surface temperatures (°C) in Stockholm 1756–2011, calculated from the homogenised daily mean temperature series by Moberg et al. (2002) after a correction for a suspected positive bias in summer temperature before 1859 (Moberg et al. 2003). The correction is the same as used by Moberg et al. (2005). Smoothed curves (Gaussian filter, σ = 3) highlight variability on timescales longer than 10 years

4.4.2 Temperature Trends in Recent Decades

The reanalysis data compiled from the mesoscale analysis system (MESAN) and ERA40 dataset indicated that during the period 1990–2004, all years except one, 1996, had a mean temperature above normal for most of Europe (Jansson et al. 2007). Lehmann et al. (2011) showed a warming trend in annual mean temperature of 0.4 °C per decade over the Baltic Sea basin with the strongest change in its northern part using the SMHI database with a 1° × 1° horizontal resolution and 3 h time step during 1970–2008 (Fig. 4.14). The strongest trend in the Gulf of Bothnia occurred in autumn and winter (0.5–0.6 °C per decade), while in the central and southern part of the Baltic Sea region, significant changes occurred in spring and summer (0.2–0.3 °C per decade). This is in agreement with annual mean surface air temperature at coastal stations in Estonia having increased by about 0.3 °C per decade during 1950–2009 (Russak 2009; Kont et al. 2011) and the Russian part of the Baltic Sea drainage basin, where temperatures increased by 0.4 °C per decade during 1976–2006 (Chap. 6). In terms of seasonal change, statistically significant warming occurred in winter, spring and summer, but not autumn, and in terms of months during the first half of the year (JFMAM) with the greatest change in March and April.

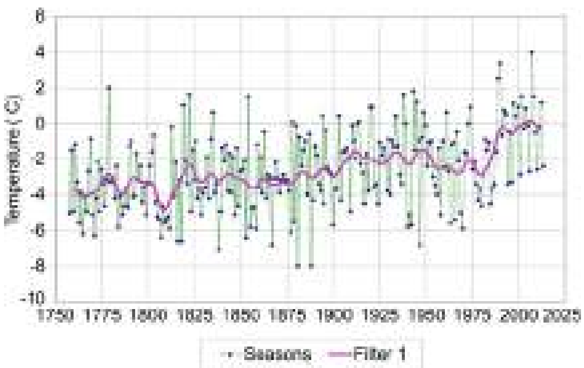


Fig. 4.13 Late winter/early spring (FMA) temperature for south-eastern Norway (Austlandet) based on ice break-up data from Lake Randsfjorden (1758–1874) and instrumental observations (1875–2011), updated from Nordli et al. (2007). Filter 1 means 10-year moving average series of original data

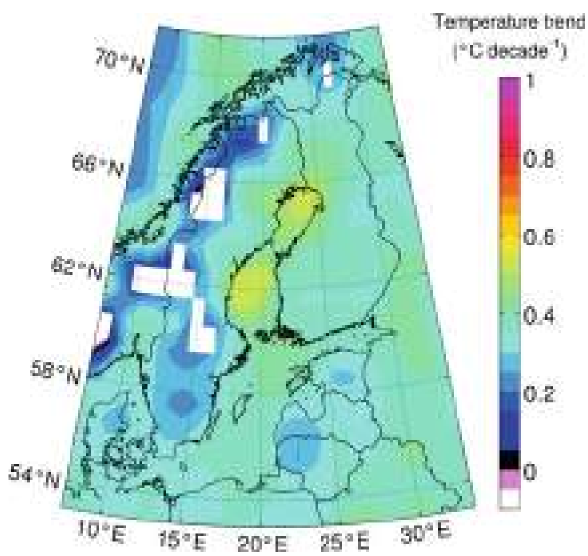


Fig. 4.14 Linear trend in annual mean surface air temperature during 1970–2007 based on the SMHI dataset. Trends in the light-shaded areas (as over parts of the Scandinavian mountains) are statistically non-significant at the $p < 0.05$ level (Lehmann et al. 2011)

4.4.3 Daily Cycle and Seasonality

Not only are the annual mean and seasonal mean temperatures changing, but the daily temperature cycle, which is reflected in the daily maximum and minimum temperatures, is also changing. Previous studies have shown that the daily minimum temperature has increased much more than the daily maximum causing a decreasing trend in diurnal temperature amplitude (BACC Author Team 2008). In Rīga, the mean minimum temperature increased by 1.9 °C during 1913–2006, while the mean maximum temperature increased by 1.7 °C. The mean maximum temperature increased most rapidly in the later part of spring (April–May), while the minimum temperature increased most in winter (Lizuma et al. 2007). Consequently, the daily temperature range has decreased (Avotniece et al. 2010).

Climate change is expressed not only in time series of climatic parameters but also in changes in seasonality. The length of the growing season and the sums of positive degree-days have previously been shown to increase, whereas the length of the cold season and the frost days have decreased (BACC Author Team 2008). Trends in start dates and duration of climatic seasons were analysed for Tartu, Estonia, during 1891–2003 by Kull et al. (2008). Some statistically significant changes ($p < 0.05$) were detected. The start of late autumn (i.e. the end of the growing season, indicated by a continuous drop in daily mean air temperature below 5 °C) became 8 days later and the start of winter (indicated by the formation of a permanent snow cover) 17 days later. The duration of summer increased by 11 days and of 'early winter' by 18 days, while the duration of winter proper has decreased by 29 days. Here, 'early winter' means the transition period with unstable weather conditions with the forming and melting of snow cover before the beginning of winter proper, which is defined by the presence of a permanent snow cover. The length of the growing season (defined by a daily mean air temperature permanently above 5 °C) increased by 13 days (Kull et al. 2008). In Poland, the number of ice days ($T_{max} \leq 0$ °C) and frost days ($T_{min} \leq 0$ °C) decreased by 2 and 3 days per decade, respectively, while the mean monthly minimum and maximum temperatures and the frequency of warm days ($T_{max} \geq 25$ °C) generally increased during 1951–2006. A warming trend occurred in winter, spring and mid- and late summer, but not in June or autumn (Wibig 2008a).

4.4.4 Temperature Extremes

Changes in temperature extremes may influence human activity much more than changes in average temperature. There have been a number of projects and studies on changes in temperature extremes over the past decade. Examples of projects investigating extremes of relevance for the Baltic Sea region include the following: European Climate Assessment (Klein Tank and Können 2003), Statistical and Regional dynamical Downscaling of Extremes for European regions (STARDEX; Haylock and Goodess 2004), European and North Atlantic daily to MULtidecadal climATE variability (EMULATE; Moberg et al. 2006) as well as several national projects. A large number of indices describing extremes have been elaborated (BACC Author Team 2008).

Using ensembles of simulations from a general circulation model (HadCM3), large changes in the frequency of 10th percentile temperature events over Europe in winter

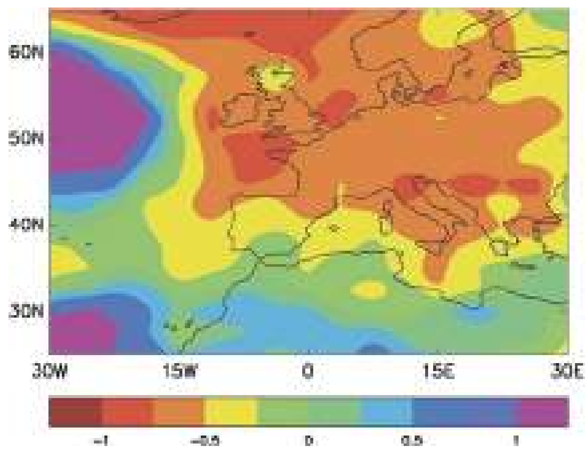


Fig. 4.15 Fractional change in the frequency of very low daily minima in winter surface temperature. The fractional change in occurrence of below-10th percentile daily minimum temperature events is plotted between the winters 1964/1965–1968/1969 and 1990/1991–1994/1995. The 10th percentile is defined over the former period for each ensemble and at each grid point. In order that the existing dataset could be used, percentiles are calculated over 1961–1990. All daily values for the model were pooled together over the December–February (DJF) period to calculate both percentile thresholds and changes in frequency (Scaife et al. 2008)

appear mainly explained by related changes in the NAO from the 1960s to 1990s (Scaife et al. 2008). A decreasing tendency in the frequency of very low daily minima in winter surface temperatures has been present across most of Europe including the Baltic Sea region over the past 50 years (Fig. 4.15). The number of frost days between the winters 1964/1965–1968/1969 and 1990/1991–1994/1995 decreased by 20–30 days in that region (Scaife et al. 2008).

Nine temperature indices were used for analysing weather extremes at 21 stations in Poland during 1951–2006 (Wibig 2008a). The indices are based on daily maximum and daily minimum temperatures. A statistically significant increase was detected for the annual numbers of days with daily maximum temperature above both 25 and 30 °C, while a statistically significant decrease was observed in the length of the frost season and in the annual number of frost days (daily minimum below 0 °C) and ice days (daily maximum below 0 °C). Positive trends in monthly temperature indices were observed from February to May and from July to September (Wibig 2008a).

Changes in temperature extremes were also observed in Latvia using daily data from two stations for the period 1924–2008 and three stations for the period 1946–2008 (Avotniece et al. 2010). The Mann–Kendall test indicates statistically significant positive trends in the number of tropical nights (T_{min} ≥ 20 °C) and of summer days (T_{max} ≥ 25 °C) and negative trends in the number of frost days (T_{min} ≤ 0 °C) and of ice days (T_{max} ≤ 0 °C). The number of hot days and nights increased in Lithuania during 1961–2010 particularly in July and August and in 1998–2010 (Kažys et al. 2011). Abnormally warm or cold weather conditions over several consecutive days can be used as an alternative to studying temperature extremes. An analysis of extremely warm and cold days in Łódź, Poland, during 1931–2006 was undertaken by Wibig (2007). A particular day was included in a warm (cold) period if its daily mean temperature was higher (lower) than 1.28 (−1.28) standard deviations for this particular calendar day, which corresponds to the 90 % (10 %) percentile in the case of a normal distribution. The duration of extremely mild periods has increased significantly in winter, while the number of heat waves has increased in summer as well as during the year as a whole. Accordingly, the length of cold waves has decreased significantly in winter. The annual number of cold days decreased by 0.87 days per decade (Wibig 2007). In spite of general warming, the frequency of cold periods in Poland has not decreased significantly (Wibig et al. 2009). Time series of the number of days with T_{min} ≤ 15 °C, T_{min} ≤ 20 °C and T_{max} ≤ 10 °C show a significant trend at only one of the nine stations (Zakopane) during 1951–2006. An increase in the frequency of heat waves occurred in the Czech Republic in 1961–2006 (Kyselý 2010). Owing to the increase in mean summer temperature, the probability of very long heat waves in the Czech Republic has risen by an order of magnitude over the past 25 years (Kyselý 2010).

Changes in surface air temperature are determined mostly by large-scale atmospheric circulation. As this influences large geographical areas, it is not surprising that temperature studies focusing on different areas show similar fluctuations and trends. Nevertheless, data quality is also very important in trend analysis, and thus, time series based on data from different sources and might contain significant inhomogeneities.

4.5 Precipitation

4.5.1 Long-Term Precipitation Climate

Change in precipitation during the twentieth century in the Baltic Sea basin has been more variable than for temperature. There have been regions and seasons of increasing precipitation as well as those of decreasing precipitation (BACC Author Team 2008). Time series of European mean precipitation is characterised by large inter-annual and inter-decadal variability and with no long-term trends apparent for 1766–2000 (Casty et al. 2007). There were no clear trends in Latvia during 1922–2003 (Briede and Lizuma 2007). Variations in annual precipitation with a periodicity of 26–30 years were detected for all three Baltic countries during 1922–2007 (Kriauciuniene et al. 2012). The periodicity varied by season: spring (24–32 years), summer (21–

33 years), autumn (26–29 years) but with no periodicity evident in winter. Summer precipitation in Finland during 1908–2008 showed a positive trend (Ylhäisi et al. 2010). A statistically significant trend in south-western Finland was detected in June and in north-eastern Finland in May, July and for the whole summer period (MJJAS). An increase in precipitation was detected in south-eastern Norway where annual precipitation increased by about 15–20 % during 1900–2010, with a higher increase in autumn and winter and a lower increase in spring and summer (Hanssen-Bauer et al. 2009). There has been no trend in autumn precipitation over the past 30 years in this area but a marked increase in winter precipitation (Hanssen-Bauer et al. 2009).

A comparison of annual mean precipitation for 1994–2008 with that for 1979–1993 showed less precipitation in the northern and central Baltic Sea region and more in the southern region (Lehmann et al. 2011). The pattern also varied by season (Fig. 4.16). Spatial distribution and trends in precipitation related to large-scale atmospheric circulation and local landscape factors were analysed for the three Baltic countries (Jaagus et al. 2010). A statistically significant positive trend for 1966–2005 was detected only for eastern Estonia, eastern Latvia and Lithuania as a whole in winter and for western Estonia in summer. A detailed study of summer precipitation data for Helsinki during 1951–2000 did not indicate any trends (Kilpeläinen et al. 2008).

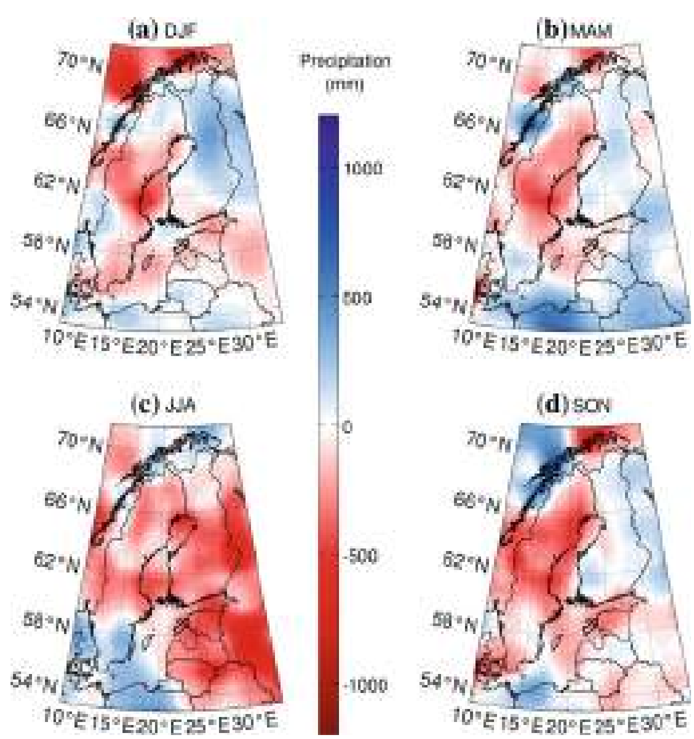
4.5.2 Precipitation Climate in Recent Decades

A general increase in precipitation during winter is typical for northern Europe over the past few decades. The greatest increase has been observed in Sweden and on the eastern Baltic Sea coast, while southern Poland has on average received less precipitation. A tendency of increasing precipitation in winter and spring was detected during the latter half of the twentieth century. Benestad et al. (2007) compared downscaled and modelled precipitation at 27 stations in Fennoscandia for 1957–1999. Only a few locations exhibited trends that were statistically significant at the 5 % error level.

4.5.3 Precipitation Extremes

According to the BACC Author Team (2008) and more recent studies, precipitation increase in northern Europe is associated with an increase in the frequency and intensity of extreme precipitation events. The climatology of extreme precipitation events is usually described by several indices of heavy precipitation. Zolina et al. (2009) showed mostly positive trends in daily precipitation and precipitation totals at 116 stations across Europe, including the Baltic Sea region, during 1950–2000. Although most of the trends are not statistically significant, some positive trends were identified for winter and spring, while negative trends mostly

Fig. 4.16 Change in total precipitation between 1994–2008 and 1979–1993 by season based on SMHI data (Lehmann et al. 2011)



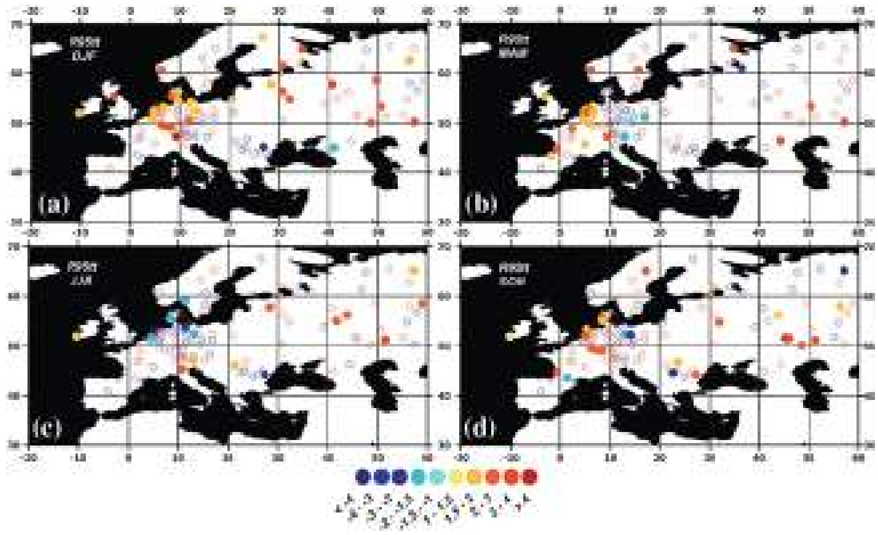


Fig. 4.17 Linear trends (% per decade) in the R95tt index (fraction of total precipitation above 95th percentile of rain-day amounts) for **a** winter, **b** spring, **c** summer and **d** autumn for the period 1950–2000. *Open circles* show all trend estimates and *closed circles* denote locations where the trends are significant at 95 %. *Blue* indicates negative trends and *red* indicates positive trends (Zolina et al. 2009)

occurred in summer and, in some cases, in autumn (Fig. 4.17).

Using ensembles of simulations from a general circulation model (HadCM3), Scaife et al. (2008) found large changes in the frequency of 90th percentile precipitation events over Europe in winter, which are related to changes in the NAO from the 1960s to 1990s. This relationship is likely to have led to an increased occurrence of heavy precipitation events over northern Europe and a decreased occurrence over southern Europe and extratropical North Africa during high NAO periods.

The duration of wet periods with daily precipitation exceeding 1 mm increased by 15–20 % across most of Europe during 1950–2008 (Zolina et al. 2010). The increased duration of wet periods was not caused by more wet days, but by short rain events having been regrouped into prolonged wet spells. Becoming longer, wet periods in Europe are now characterised by heavier precipitation. Heavy precipitation events during the past two decades have become more frequently associated with longer wet spells, and precipitation is now heavier than during 1950s and 1960s (Zolina 2011).

Wibig (2009) used daily data at five stations in Poland during the latter half of the twentieth century to analyse the number of days with precipitation exceeding given thresholds, the duration of wet and dry spells and precipitation amount in a single event. A positive trend was detected in the number of wet spells and days with precipitation and a negative trend in mean precipitation during any given spell.

Very few changes were detected at a number of stations in Poland during 1951–2006 (Wibig 2008a; Łupikasza 2010). Only the number of days with precipitation increased significantly. Positive as well as negative trends in indices of precipitation extremes were detected (Fig. 4.18). Decreasing trends dominated in summer, while increasing trends were more pronounced in spring and autumn. In summer and winter, decreasing trends were more spatially coherent in southern Poland (Łupikasza 2010). A similar result (i.e. lack of significant trends in extended precipitation time series) was found in Łódź for 1904–2000 (Podstawczyńska 2007); the only trends were an increase in dryness indices in February, April and August.

Trends in extreme precipitation in western Germany during 1950–2004 were analysed by Zolina et al. (2008). Only the northernmost part of the territory belongs to the Baltic Sea basin. Increasing trends in the 95th and 99th percentiles were detected in winter, spring and autumn and with a negative trend occurring in summer in northern Germany.

Kažys et al. (2009) and Rimkus et al. (2011) analysed long-term changes in heavy precipitation events in Lithuania during 1961–2008. They found increasing trends for the number of days with heavy precipitation (above 10 mm) and for heavy precipitation as a percentage of total annual precipitation (Fig. 4.19).

Mean values of daily heavy precipitation as a percentage of the annual total range from 33 to 44 % in Lithuania and in some years can be close to 60 %. In summer and autumn, the percentage of heavy precipitation is much higher than during the rest of the year. Analysis showed positive but mostly non-significant tendencies across large parts of Lithuania during the study period. This means that the temporal variability of precipitation has increased. This tendency is especially clear in summer when an increase in precipitation extremes can be observed in spite of neutral or negative tendencies in the total summer precipitation (Rimkus et al. 2011).

Valiukas (2011) analysed long-term fluctuations and trends in the occurrence of dry periods for Vilnius, Lithuania, during 1891–2010. Two indices—the Standardized

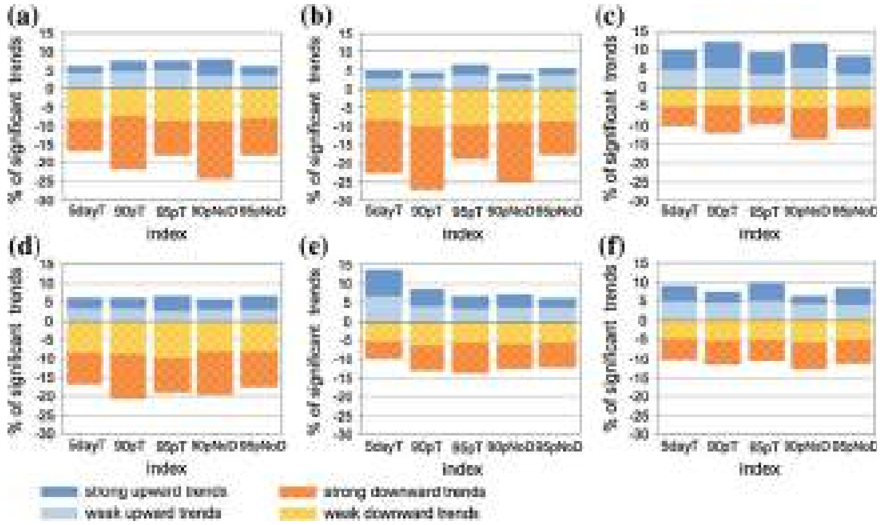


Fig. 4.18 Percentage of statistically significant trends in extreme precipitation indices calculated for each station and for each of the 30-year moving periods within 1951–2006 in Poland. **a** cool half-year, **b** summer, **c** spring, **d** warm half-year, **e** winter, **f** autumn. Weak trends—significant at $0.1 < p \leq 0.2$, and strong trends—significant at $p \leq 0.1$. The number of possible cases was counted by multiplying the number of stations by number of 30-year moving periods (26 stations with data covering the period 1951–2000 multiplied by 21 time series or 21 stations with data covering the period 1951–2006 multiplied by 27 time series) (Łupikasza 2010)

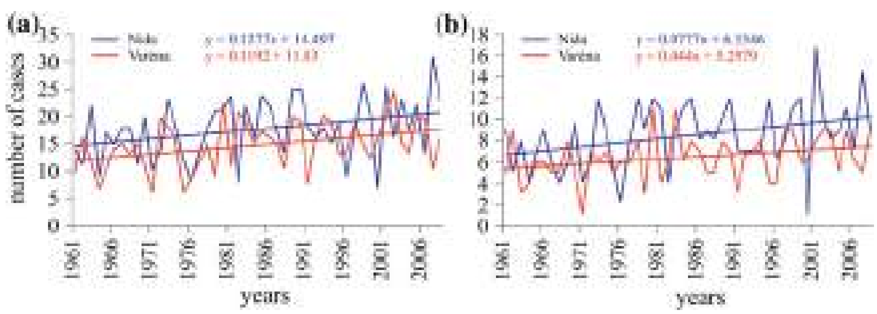


Fig. 4.19 Number of days with heavy precipitation **a** >10 mm per day and **b** >20 mm in three consecutive days in Nida (western Lithuania) and Varėna (south-eastern Lithuania) in 1961–2008. All trends are statistically significant according to a Mann–Kendall test (Rimkus et al. 2011)

Precipitation Index (SPI) for different time steps and the Selyaninov Hydrothermic Coefficient (HTC—ratio of precipitation and sum of daily mean temperatures divided by 10)—were used for identifying drought. Statistically significant trends were not detected. A small decrease in dryness was observed, however. A significant increase in the number of days with heavy precipitation (≥10 mm) was observed in Latvia during 1924–2008 (Avotniece et al. 2010). Analysis of extreme precipitation and drought events in Estonia using moving averages of daily precipitation revealed an increase in the occurrence of extreme events (Tammets 2007, 2010). During the period 1957–2006, the sum of wet and dry days has grown considerably in Estonia. The main cause was two extremely dry summers, 2002 and 2006. An increase in the number of wet and dry days and persistence of precipitation events could be due to the increase in persistence of weather patterns discussed in Sect. 4.2.

Przybylak et al. (2007) used the climate extreme index (CEI) according to Karl et al. (1996) to determine the variability of all climate extremes together for Poland over 1951–2005. Using gridded data from the NCEP/NCAR reanalysis, they found that CEI was at a maximum in the 1990s with a non-significant upward trend for the period as a whole. Increasing trends was seen for annual mean number of days with $T_{\max}$ and $T_{\min}$ above the 90th percentile and for percentage changes in areas of Poland where the precipitation minus potential evapotranspiration is considerably below normal. A decreasing trend was found for percentage changes in area of Poland with precipitation minus potential evapotranspiration considerably above normal, spatially averaged frequency of extreme 1-day precipitation totals above 15 mm and a considerably greater than normal mean number of days with precipitation (Przybylak et al. 2007).

4.6 Cloudiness and Solar Radiation

4.6.1 Cloudiness

Records of cloudiness and solar radiation are generally shorter than for temperature and precipitation. There were remarkable long-term fluctuations in mean cloudiness and sunshine duration over the Baltic Sea basin during the twentieth century. The trends were of almost opposite sign between the northern (Estonia) and southern (Poland) parts of the study region (BACC Author Team 2008). From the 1950s until the 1990s, total cloud cover decreased over Poland, while the amount of low-level clouds increased over Estonia. The trends were reversed in the 1990s. Fluctuations in sunshine duration were more or less of opposite sign compared to cloudiness. There is a trend of decreasing cloud cover over the Baltic Sea basin of 1 % per decade for 1970–2008 (Fig. 4.20), mostly in spring and autumn (Lehmann et al. 2011). However, cloud cover increased over parts of the mountainous regions of Scandinavia and the south-eastern Gulf of Finland, mainly during winter and summer.

Filipiak and Miętus (2009) undertook a detailed study of spatial and temporal variability in cloudiness at 41 stations in Poland for 1971–2000. Positive as well as negative changes in cloudiness were detected, mainly related to local variability; only a few were statistically significant. Time series of cloudiness in Łódź, Poland, for 1951–2000 indicate some trends in the frequency of specific cloud types (Wibig 2008b). Total cloud cover significantly decreased, while low-level clouds increased during the warm season. Straticform clouds became less frequent and convective clouds more frequent. An increasing trend in high-level clouds was also observed (Wibig 2008b). A clear decrease in cloudiness after the 1980s was recorded in Lund, Luleå, Sodankylä and Hamburg, that is in the Baltic Sea region (Stjern et al. 2009) mainly in March and September. An increase in annual mean low-level clouds was reported at Tartu-Tõravere, Estonia, from the 1960s to 1980s, and a rapid decrease since 1990s (Russak 2009).

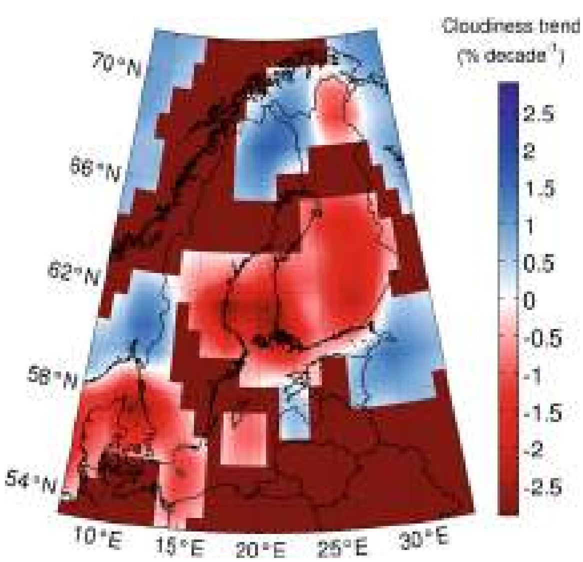


Fig. 4.20 Linear trend of total cloud cover during 1970–2008 based on meteorological data obtained from SMHI (Lehmann et al. 2011)

4.6.2 Sunshine Duration and Solar Radiation

Stjern et al. (2009) presented surface solar radiation data from 11 stations in north-western Europe and the European Arctic within the context of the ongoing discussion on global dimming and global brightening. They compared the records to records of cloud cover and to qualitative information on aerosol concentrations and atmospheric circulation patterns. The authors found a decrease in solar radiation of 18.3 % or 21.5 W m^2 during 1955–2003 (Fig. 4.21), while the 1983–2003 period showed a 4.4 % increase. The earlier part of the time series (before 1965) contains data from only two stations (Hamburg, Copenhagen) and so is less reliable. After 1965, no trend can be distinguished. Similar results on change in solar radiation were obtained for the Tartu-Tõravere Meteorological Station in Estonia during 1955–2007 (Russak 2009). A decrease in global and direct radiation occurred from the 1950s until the end of the 1980s (Fig. 4.22). After that an increasing trend is evident. The

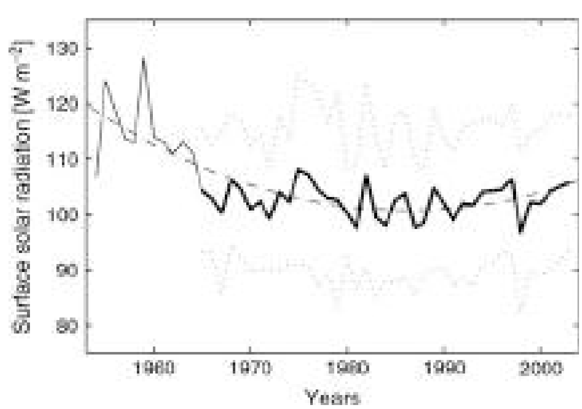


Fig. 4.21 Mean surface solar radiation at 11 stations. *Solid line* indicates where the mean comprises more than three stations and the *dotted line* an envelope of ±1 standard deviation. The *dashed line* represents a second-degree regression fit (Stjern et al. 2009)

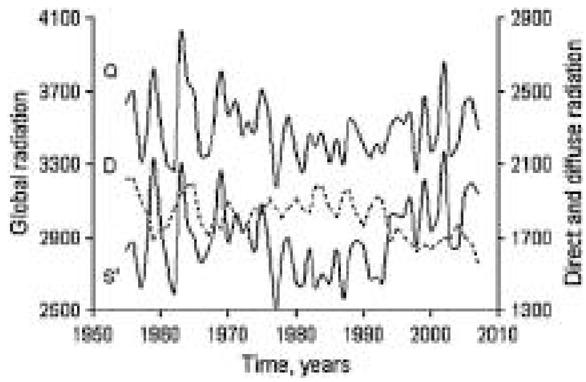


Fig. 4.22 Time series of global (Q), diffuse (D) and direct radiation incident on a horizontal surface (S') at Tartu-Tõravere meteorological station, Estonia. Unit is MJ m^{-2} (Russak 2009)

trend in solar radiation is opposite to that in low-level cloud cover.

Changes in global solar radiation based on the NCEP/NCAR dataset were analysed by Uscka-Kowalkowska et al. (2007) for central Europe during 1951–2005. The study area had 35 grid points and covered Poland, eastern Germany, Czech Republic, Slovakia, western Ukraine and Belarus, Lithuania and southern Sweden, areas which mostly belong to the Baltic Sea basin. A general increase in solar radiation was detected during the study period. The strongest trends were found in April, May, November and December, while little change was detected in March and October. Higher trend values are characteristic for the northern part of the study area (Uscka-Kowalkowska et al. 2007). Net radiation and net long wave radiation show a significant positive trend in Tartu-Tõravere, Estonia, during 1961–2002 (Fig. 4.22) (Russak 2009). Variability in the individual components of the radiation budget for the Baltic Sea region was discussed by the BACC Author Team (2008). No new evidence is available.

The studies mentioned here generally indicate negative trends in cloudiness and coincident positive trends in sunshine duration and solar radiation in the major parts of the Baltic Sea basin over recent decades. The number of investigations to date is not sufficient to enable more detailed and reliable conclusions.

4.7 Conclusion

Variations and trends of atmospheric parameters in the Baltic Sea region during the last 200–300 years can be summarised as follows. A northward shift in storm tracks and increased cyclonic activity have been observed in recent decades with an increased persistence of weather types. No long-term trend have been observed in annual wind statistics since the nineteenth century, but considerable variations on (multi-)decadal timescales have been observed. An anthropogenic influence cannot be excluded since the middle of the twentieth century. The pattern in wind and wave heights over the Northern Hemisphere with a NE shift of storm tracks appears to be consistent with combined natural and external forcing. Continued warming has been observed, particularly during spring and is stronger over northern regions (polar amplification) than southern. Bhend and von Storch (2009) detected that the significant warming trends over the Baltic Sea region are consistent with future climate projections under increased greenhouse gas concentrations. No long-term trend was observed for precipitation, but there is some indication of an increased duration of precipitation periods and possibly an increased risk of extreme precipitation events.

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5 Recent Change—River Run-off and Ice Cover

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Abstract

This chapter compiles and assesses information on run-off and discharge from rivers within the Baltic Sea drainage basin. Some information is also available on ice duration on inland waterways. Although decadal and regional variability is large, no significant long-term change has been detected in total river run-off to the Baltic Sea over the past 500 years. A change in the timing of the spring flood has been observed due to changes in the timing of snowmelt. Change in temperature seems to explain change in run-off better than does precipitation. Later start dates for ice formation on waterways, and earlier ice break-up dates have resulted in shorter periods of ice cover.

Keywords

Baltic sea • Climate change • River • Run-off • Discharge • Ice cover

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5.1 Introduction

This chapter compiles and assesses recent information on run-off and river discharge[1] within the Baltic Sea basin. Most of the information is based on peer-reviewed scientific publications, but a substantial share of the data has been provided by national authorities responsible for hydrological monitoring. Lake hydrology is not addressed in this chapter because widespread regulation for hydropower generation and the resulting complexity in lake level variation tend to hamper attempts to assess climatically induced change. Readers interested in the impact of climate change specifically on lakes are advised to refer to the European-scale review by Glen (2010).

[1] In this chapter, the terms *inflow*, *runoff* and *discharge* are used as follows. Runoff describes general, long-term and/or regional processes and is typically given as litres per second per square kilometre ($l\ s^{-1}\ km^{-2}$) (allowing comparisons between rivers of different sizes) or millimetres per year ($mm\ year^{-1}$) (allowing comparisons with precipitation and evaporation), whereas inflow and discharge typically refer to immediate channel flow and are given as cubic metres per second ($m^3\ s^{-1}$). Where possible, the terminology used in the original publications has been respected.

The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_5

The magnitude of water flow in a river is the result of various complex hydrological processes including precipitation, evapotranspiration, infiltration and storage (in the form of snow, soil moisture, and sub-surface and groundwater storage, etc.). Explaining changes in streamflow thus requires an understanding of these parameters of which precipitation is often pivotal. Meteorological parameters are addressed in Chap. 4, but some notes may be beneficial for general reference. Bordi et al. (2009) studied linear and nonlinear trends in drought and wetness within Europe in terms of the gridded standardised precipitation index (SPI) determined from monthly precipitation reanalyses by the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR). They concluded that time series of drought and wetness area coverage show a marked linear trend until about the end of the twentieth century at which point the trends reverse. The recent reversed nonlinear trend is particularly pronounced on the hydrological timescale. Substantial regional variation occurs in drought/wetness trends (Fig. 5.1).

Another European-wide example is provided by Stahl et al. (2010) who studied streamflow trends in 1962–2004 in a broad European study consisting of 411 near-natural small catchments in 15 countries. Positive trends with increasing streamflow were found in the winter months in most catchments, while in spring and summer months, strong negative trends (decreasing streamflow, shift towards drier conditions) were found specifically in the southern and eastern regions of the study area, as well as widespread across Europe. A marked shift towards negative trends was observed in April, gradually spreading across Europe to reach a maximum extent in August. Low flows have decreased in most regions where the lowest mean monthly flow occurs in summer, but vary for catchments which have flow minima in winter and secondary low flows in summer. The study by Stahl et al. (2010) largely confirmed findings from national- and regional-scale trend analyses.

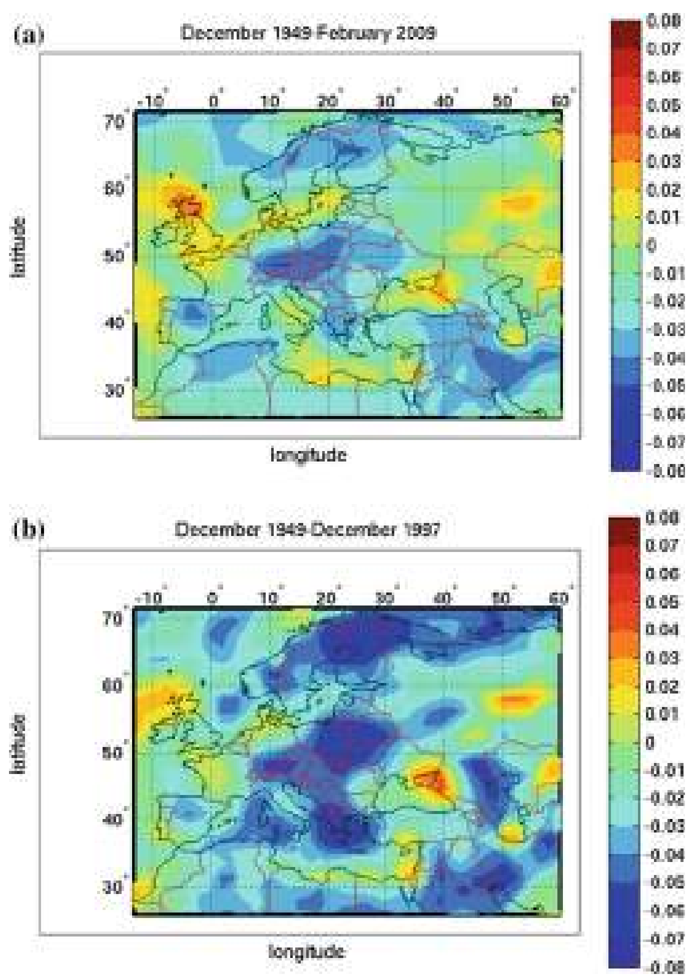


Fig. 5.1 Spatial distribution of the tendency towards wetter and drier conditions within Europe including the Baltic Sea basin. This is shown by plots of the angular coefficient $p1$ (in year^{-1}) of the linear trend fitting the SPI-24 time series for the two time sections: **a** December 1949 to February 2009 and **b** December 1949 to December 1997. Negative (*blue*) values of $p1$ denote a tendency towards drier periods, while positive (*red*) ones towards wetter periods (Bordi et al. 2009)

5.1.1 General Drainage Characteristics of the Baltic Sea Basin

The land surface of the Baltic Sea basin, in other words the drainage area of the Baltic Sea, covers 1.74 million km^2. It includes the territories of 14 countries. Three countries—Estonia, Latvia and Lithuania—are located entirely within the Baltic Sea basin, while only minor parts of the following countries drain towards the Baltic Sea: Czech Republic, Germany, Norway, Slovakia and Ukraine. The largest national areas of the Baltic Sea basin are those of Sweden (25.3 %), Russia (19.0 %), Poland (17.8 %) and Finland (17.4 %). These four countries together cover two-thirds of the basin.

The exact number of sub-basins within the Baltic Sea catchment area is a matter of definition, but Hannerz and Destouni (2006) have delineated 634 river basins of more than 6 km^2 in area (Fig. 5.2). The ten largest river basins draining into the Baltic Sea and some of their characteristics are summarised in Table 5.1. The ten rivers account for 58 % of the Baltic Sea catchment area and 55 % of inflow. The 11–20th ranked basins have a total area of 251,000 km^2 (14 % of the total) while the hundred largest basins cover about 86 % of the Baltic Sea catchment. The remaining 14 % (representing a quarter of a million square kilometres) are divided into numerous small catchments along the coastal regions and on the Baltic Sea islands. The total area of the Baltic Sea islands is almost 40,000 km^2, and there are around 200,000 islands in total.

Although there is some overlap, the ten largest river basins are not the top ten in terms of mean annual flow. The specific run-off is greatest in the north-western parts of the Baltic Sea catchment; three rivers from that region—Ångermanälven, Luleälven and Indalsälven—cover positions 8–10, displacing the Narva, Torne and Kymi rivers.

Fig. 5.2 Sub-basins of the Baltic Sea drainage basin greater than 6 km^2 in size (Hannerz and Destouni 2006)

The drainage area of Lule River is only 25,200 km^2 but the run-off, 19.0 l km^{-2} s^{-1}, leads to a mean annual discharge of 486 m^3 s^{-1}.

The first assessment of climate change in the Baltic Sea basin (BACC Author Team 2008) summarised the scientific

Table 5.1 Ten largest river basins draining into the Baltic Sea. The runoff values refer to the period 1950–90 (BACC Author Team 2008)

River	Country	Area (km^2)	Percentage of Baltic Sea drainage basin	Mean annual discharge (m^3 s^{-1})	Percentage of total river inflow to the Baltic Sea	Run-off (l km^{-2} s^{-1})
Neva	Russia/Finland	281,000	16.1	2460	17.6	8.8
Vistula	Poland/Ukraine/Belarus/Slovakia	194,400	11.2	1065	7.6	5.5
Odra	Poland/Germany/Czech Republic	118,900	6.8	573	4.1	4.8
Nemunas (Lithuanian)	Belarus/Lithuania/Russia	98,200	5.6	632	4.5	6.4
Daugava	Belarus/Latvia/Lithuania/Estonia/Russia	87,900	5.1	659	4.7	7.5
Narva	Estonia/Russia	56,200	3.2	403	2.9	7.2
Kemi	Finland	51,400	3.0	562	4.0	11.0
Göta	Sweden	50,100	2.9	574	4.1	11.5
Torne	Sweden/Finland	40,100	2.3	392	2.8	9.8
Kymi	Finland	37,200	2.1	338	2.4	9.1
Total		1,015,400	58	7658	55	8.2[a]

[a] Denotes average value

understanding of the Baltic Sea catchment hydrology based on publications up to 2006. The BALTEX database of river run-off was utilised for the period 1921–1998. The key findings of the 2008 assessment are summarised in Box 5.1.

Box 5.1 Key findings of the first assessment of climate change in the Baltic Sea basin

Annual and seasonal variation in total river inflow:

- The average annual inflow to the Baltic Sea for 1921–1998 was 14,119 $m^3 s^{-1}$ or 445 $km^3 year^{-1}$, excluding the Danish belts and sounds.
- The wettest year was 1924 with an inflow of 18,167 $m^3 s^{-1}$ (+28 % compared to the average annual inflow), and the driest year was 1976 with an inflow of 10,553 $m^3 s^{-1}$ (−25 %).
- The wettest decade was the 1990s with an average annual inflow of 14,582 $m^3 s^{-1}$ (+3.3 % compared to the long-term average), and the driest decade was the 1940s with an average annual inflow of 12,735 $m^3 s^{-1}$ (−9.8 %).
- No statistically significant linear trend could be found in the time series 1921–1998.

Regional variations and trends in run-off:

- There were positive trends in run-off for 1920–2002 at several stations in Denmark, southern Sweden and Lapland—although the trends were statistically significant at only two stations. Negative trends were rare and regionally scattered.
- Statistically significant positive trends in run-off were relatively common in winter (DJF) and spring (MAM) in the period 1941–2002 across the northern part of the Baltic Sea drainage basin.

Floods:

- A widespread pattern of snowmelt floods occurring earlier in the spring was likely to have been the result of higher temperatures.
- A positive phase in the North Atlantic Oscillation (NAO) often indicates high water levels earlier in spring and maximum levels above the long-term average with the opposite tendency in the southern part of the Baltic Sea drainage basin.

Lakes:

- Long-term variations in lake levels in the Baltic Sea basin were not analysed as widely as the variations in river discharge due to the regulation of many lakes: changes in their levels do not correctly reflect variation in climatic or physiographic factors.

Ice regime:

- In the Russian territory of the Baltic Sea basin, the duration of complete ice cover on river decreased during the latter half of the twentieth century: by 25–30 days on northern rivers and 35–40 days on southern rivers. The maximum ice-cover thickness decreased 15–20 % on all rivers studied.
- In Finland, in a data series extending to late seventeenth century, the ice break-up on lakes has moved 6–9 days earlier per hundred years. Since late nineteenth century, freezing has delayed by 0–8 days per century.

5.2 Basin-scale Change in Run-off Patterns

The Baltic Sea can be considered a large, semi-enclosed brackish water estuary draining into the North Sea via the Danish Straits. The inflow from rivers to the Baltic Sea is an important variable for both the physical and ecological processes of the sea. The form of precipitation falling on its drainage area has a major impact on the annual run-off regime. In winter, much of the precipitation is stored as snow, especially towards the northern part of the basin. Consequently, water levels and discharges in the northern area are at their lowest towards the end of the winter, before snowmelt. The highest water levels and discharges are recorded in spring or early summer owing to snowmelt. Water levels and discharges usually decrease during summer when evaporation/evapotranspiration is greatest and is normally larger than precipitation. Sometimes, if the summer is dry and warm, water levels can even drop below the winter minimum. Climate change is likely to have a clear influence on the seasonal flow regime as a direct response to changes in the form of the precipitation, as well as by altering the temperature–evapotranspiration regime.

Hansson et al. (2011) reported—based on temperature and atmospheric circulation indices from 1500 onwards—that run-off to the Baltic Sea appears strongly linked to temperature, wind and rotational circulation components in the northern region and Gulf of Finland. In contrast, run-off in the southern region is more associated with the strength and torque of the cyclonic or anti-cyclonic pressure systems. Although decadal and regional variability is large, no *statistically significant* long-term change has been detected in total river run-off to the Baltic Sea over the past 500 years (Fig. 5.3).

Analysis of run-off sensitivity to temperature suggests that the southern regions may become drier with rising air temperatures, whereas in the north and around the Gulf of Finland higher temperatures are associated with greater river run-off. As a whole, over the past 500 years, the total river run-off to the Baltic Sea has decreased slightly in response to the rise in temperature, at a rate of 3 %, or 450 $m^{-3} s^{-1}$, per 1 °C (Fig. 5.4).

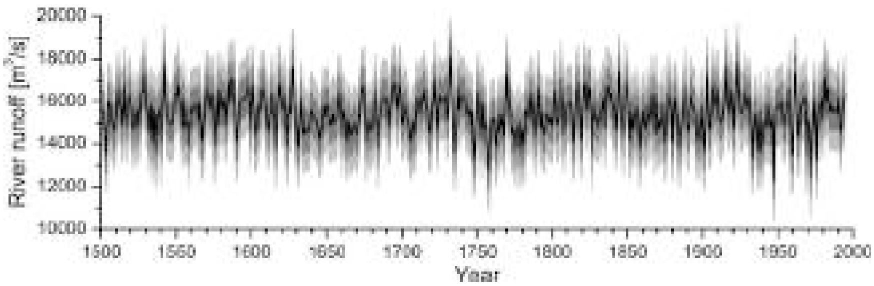


Fig. 5.3 Reconstructed annual river discharge to the Baltic Sea for the past 500 years. The *grey shading* indicates 1 and 2 standard errors of the reconstructed river discharge (Hansson et al. 2011)

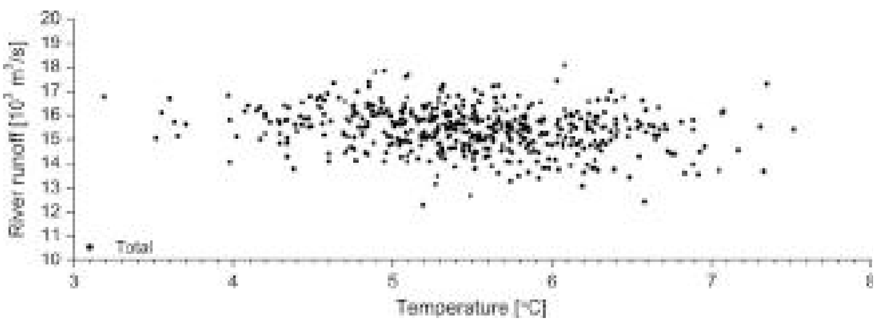


Fig. 5.4 Reconstructed total river run-off as a function of temperature in the Baltic Sea drainage basin for 1500–1995. A change of 1 °C results in a decrease in river run-off of 3 % (450 m^{-3} s^{-1}) (significant at the 95 % confidence level) (Hansson et al. 2011)

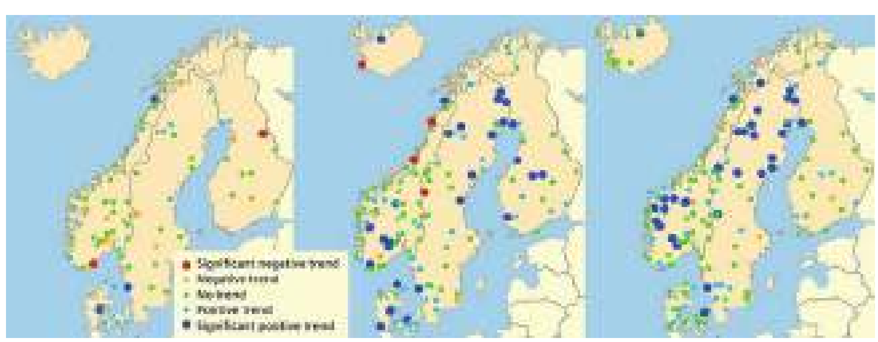
Regarding flow changes on a decadal scale, Hisdal et al. (2010) revised and extended their earlier analysis included in the first Baltic Sea assessment report (Hisdal et al. 2003; BACC Author Team 2008). Currently, the data comprise more than 160 streamflow records from the Nordic countries. The Mann–Kendall trend test was applied to study changes in annual and seasonal streamflow as well as floods and droughts for three periods: 1920–2002, 1941–2002 and 1961–2000.

Regional patterns detected by Hisdal et al. (2010) were influenced by the time period and the selection of stations. However, in general, trends towards increased streamflow dominated annual values (Fig. 5.5) and winter and spring seasons. Trends in summer flow were highly dependent on the period analysed while no trends were found for autumn. A signal towards earlier snowmelt floods was clear. Comparison of the findings to various streamflow scenarios showed the strongest trends detected to be coherent with changes expected during the scenario period, for example increased winter discharge and earlier snowmelt floods.

Hisdal et al. (2010) concluded by suggesting that the observed temperature increase has clearly affected streamflow in the Nordic countries. These changes correspond well to the projected consequences of a continued rise in global temperature, whereas the impacts of both the observed and projected changes in precipitation on streamflow are unclear.

The analyses by Hisdal et al. (2010) were further refined by Wilson et al. (2010) who considered the effect of spatial and temporal autocorrelation, investigated trend magnitude instead of statistical significance and extended the period

Fig. 5.5 Trends in annual streamflow within the Nordic countries for 1920–2002 (*left*), 1941–2002 (*middle*) and 1961–2000 (*right*) (Hisdal et al. 2010)



covered to 2005. They also applied the Mann–Kendall trend test to study changes in annual and seasonal streamflow as well as floods and droughts for three periods: 1920–2005 (68 stations), 1941–2005 (111 stations) and 1961–2000 (151 stations). The overall picture was that trends of increased streamflow dominate annual values and the winter and spring seasons. In all three periods, a signal towards earlier snowmelt floods was clear, as was the tendency towards more severe summer droughts in southern and eastern Norway. The trends in streamflow result from changes in both temperature and precipitation, but the temperature-induced signal is stronger than precipitation influences. This is a consequence of temperature affecting the timing of snowmelt and thus the seasonal distribution of flows rather than annual totals. A change in the timing of the spring flood has also been observed due to changes in the timing of snowmelt.

The tendency for a decrease in annual discharge in the southern catchments recognised by Hansson et al. (2011) was also detected by Gailiušis et al. (2011). They studied the variability of long-term monthly run-off in five rivers draining into the Baltic Sea: the Nemunas, Neva, Odra, Vistula and Luleälven. More than 100 years of data are available for all five rivers, and the data for the Nemunas at Smalininkai starting in 1812 are one of the longest data sets for river run-off in Europe. Variability within the long-term run-off series was analysed by comparing the annual, highest and lowest monthly river discharges against the long-term mean to derive 'anomalies', integrated curves of annual discharges and the results of the Mann–Kendall trend test. A decrease of about 10 % in annual discharge was observed in the rivers on the southern shore of the Baltic Sea (Fig. 5.6). Cyclical variation with an amplitude of 26–27 years in the annual discharge data series is characteristic of the Nemunas, Neva, Vistula and Odra rivers.

Variability in river discharge depends on natural environmental factors, especially on the cyclic variation in precipitation. However, drier and wetter phases in the studied rivers do not occur at the same time. The low phases in the annual discharges of the Nemunas and Neva have a lag of about 12 years compared to those of the Odra and Vistula. Discharge in the Luleälven is affected by high regulation. For a discussion of the impacts of damming and river regulation, see Chap. 17.

5.3 Regional and Seasonal Variations

5.3.1 Sub-basin-scale Changes

Centennial-scale change in historical sub-basin-scale river discharge has been modelled by Graham et al. (2009). They used the coupled atmosphere–ocean global climate model ECHO-G (Legutke and Voss 1999) downscaled with the Rossby Centre Regional Climate Model RCA3 (Samuelsson et al. 2010) for simulating temperature and precipitation. The results were used in the Hydrologiska Byråns Vattenbalansavdelning (HBV) hydrological model (Bergström 1976, 1992) to simulate river flows to four sub-basins of the Baltic Sea for the periods 1000–1199 and 1551–1929. Observations for the period 1921–2002 were used as reference. The basins studied were the Bothnian Bay, Bothnian Sea, Gulf of Finland and Gulf of Riga. For the southern basin (Gulf of Riga), the earlier periods have had higher river flow, while for the northernmost basin (Bothnian Bay) recent years show a slight increasing trend (Fig. 5.7).

In the Baltic States (Lithuania, Latvia and Estonia) in general, changes in streamflow over the twentieth century show a redistribution of run-off throughout the year: with a significant increase in winter river discharge and a tendency for decreasing spring floods (Reihan et al. 2007). Correlation between streamflow, air temperature and precipitation for the periods 1923–2003, 1941–2003 and 1961–2003 for a total number of 70 stations with a record length of 84 years of daily discharge data shows the strongest relations to occur in winter. Increasing winter discharge seems to be associated

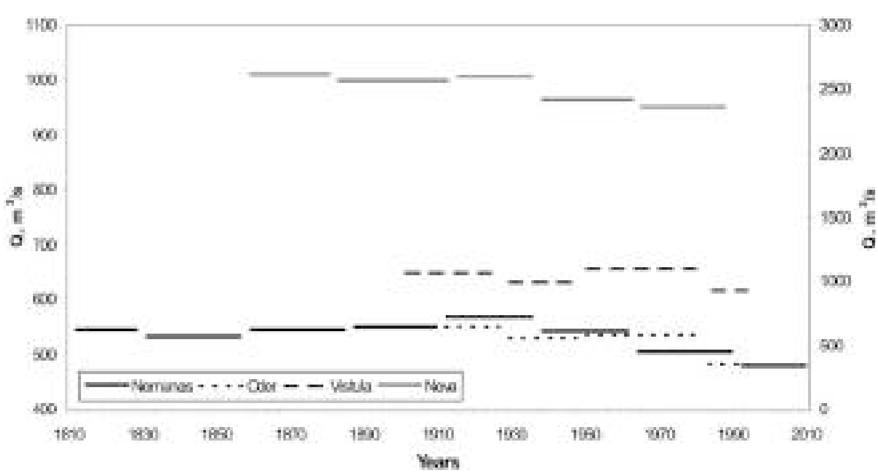


Fig. 5.6 Change in average annual discharge with the cyclicity of about 27 year amplitude for the Nemunas and Odra (Oder) rivers (*left axis*) and for the Vistula and Neva rivers (*right axis*) (Gailiušis et al. 2011)

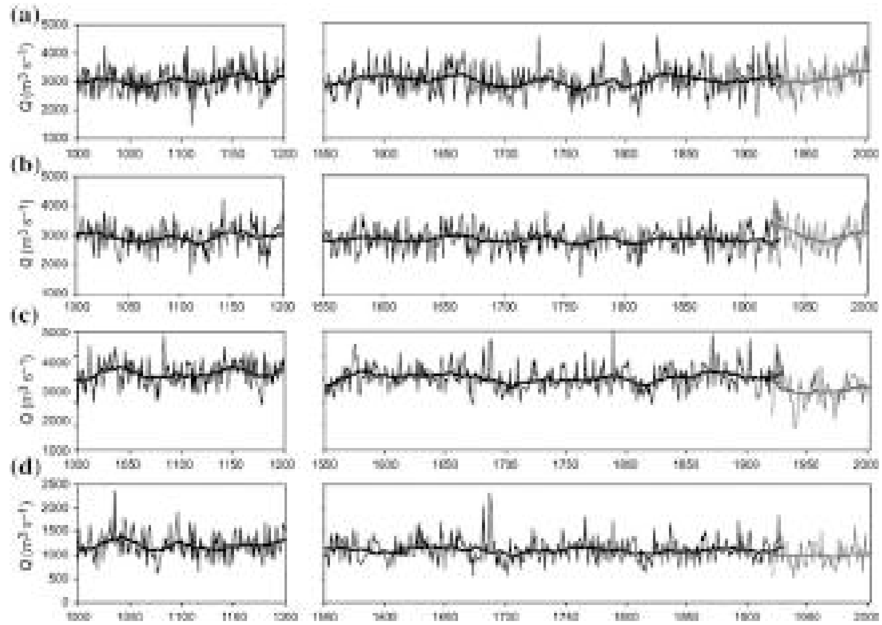


Fig. 5.7 Annual river flow to **a** Bothnian Bay, **b** Bothnian Sea, **c** Gulf of Finland, **d** Gulf of Riga for the periods AD 1000–1199 and 1551–1929 according to the HBV-Baltic simulation. The observed period 1921–2002 is shown in *grey* (Graham et al. 2009, modified)

with the significant rise in winter temperature. Average winter temperature in the Baltic States for 1961–2003 rose by 3 °C and precipitation by 43 mm, while run-off increased by 19 mm. A significant decrease in spring floods was also found.

There was no systematic pattern for the other seasons although differences in streamflow (no change, positive change and negative change) during summer and autumn in most cases reflect tendencies in precipitation and temperature. Although a clear tendency for low summer flow was lacking, a tendency for more years with low-water flow was found. In most cases, changes in river discharge were similar to the Nordic results reported by Hisdal et al. (2010). However, the decrease in spring floods in the Baltic rivers contrasts with the situation in the Nordic countries, where changes in winter snowmelt are not yet apparent in the river run-off data although they are expected in the future (Veijalainen et al. 2010). In general, the relation between the main meteorological and hydrological parameters and tendencies in river discharge trends is common across all three Baltic States and may reflect the regional impact of global climate change.

Kļaviņš et al. (2007, 2008) and Kļaviņš and Rodinov (2008) studied river run-off in the Baltic region with an emphasis on Latvia. The variability in long-term run-off for 1881–2006 was investigated in ten Latvian rivers and compared with rivers in neighbouring countries: Neva and Narva (Russia), Nemunas (Lithuania) and Pärnu (Estonia). The observation period of more than 150 years at the Riga University Meteorological Station shows that mean annual temperature has risen about 0.8 °C over the past century. A good coherence was found between changes in annual precipitation and run-off for the largest rivers in Latvia and other rivers in Baltic region (Fig. 5.8).

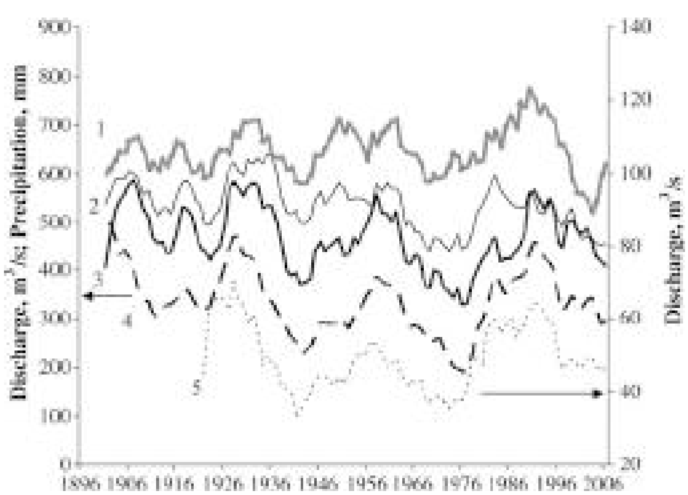


Fig. 5.8 Long-term change in precipitation and mean annual discharge for rivers in the Baltic Sea basin: *1* precipitation (Station Rīga University); *2* Nemunas; *3* Daugava; *4* Narva; *5* Pärnu. Curves 1–4: left axis, curve 5: right axis. Data were smoothed using a 6-year moving average

An increasing trend in annual mean discharge for the period 1961–2000 was found in all rivers although the trend was statistically significant in the Daugava, Gauja, Narva, Pärnu, Salaca, Dubna, Irbe and Abava rivers only. For all periods with regular hydrological observations, linear trend analysis showed a statistically significant decrease in discharge for River Nemunas (1896–2006), but for all other rivers the trends were not statistically significant. However, it was evident that discharge showed a stronger increase when the period analysed was reduced to the past 50 years. A particularly marked increase in winter discharge was detected over the past two decades. Discharge, precipitation and temperature trends were more pronounced in catchments further from the coast.

Trend analysis of the annual maximum and minimum discharges for the major rivers Daugava, Lielupe, Venta, Gauja and Salaca indicates a statistically significant decrease in maximum discharge (except for River Salaca) and a statistically significant increase in minimum discharge (except for River Gauja). The greatest number of years (3–6) with extremely high (probability ≤ 10 %) annual discharge for the major rivers in Latvia occurred during the high-flow periods of 1951–1962 and 1922–1936. Maximum discharge levels have been at their lowest over the past 50 years.

Periodic oscillations in discharge intensity and low- and high-water flow years are common for major rivers in Latvia as well as for those in the Baltic Sea region as a whole. Kļaviņš et al. (2008) found good coherence between changes in annual precipitation and river discharge. Spectral analysis revealed the following statistically significant periods of cyclicity in annual river discharge in the Baltic region: 38, 28, 14, 19, 5, 4 and 3 years. For the past 100–125 years in Latvia, low-discharge periods have been longer than high-discharge periods, lasting from 10 years up to a maximum of 21–27 years.

The winter increase and spring decrease in run-off in the Baltic States were also found by Kriaučiūnienė et al. (2012). They analysed the climate change impact on water resources in Estonia, Latvia and Lithuania (175,000 km^2) since the 1920s. Long-term temperature (40 stations), precipitation (59 stations) and river discharge (77 stations) data were used to generate ten regional data series for 1923–2007. Changes in the regional series between 1991–2007 and 1931–1960 were analysed against 1961–1990 as a reference period. The impact of changes in temperature and precipitation on river discharge was also assessed.

Anomalies in the regional discharge series depend on the climate type (marine or continental) and the sources feeding the rivers (precipitation, snowmelt, groundwater). Discharge in winter increased everywhere by 20–60 % compared with the reference period. A 10–20 % decrease in spring discharge occurred in the western regions of all Baltic States (marine climate zone), but there were no significant changes in spring discharge in the continental regions (south-eastern Lithuania and Latvia, eastern Estonia) (Fig. 5.9).

5.3.2 Regional Discharge Patterns by Country

River catchments are commonly shared by two or more countries, while hydrological offices and authorities operate on a national basis. Therefore, formulating a coherent picture of the hydrology of the Baltic Sea catchment area is complicated. However, international research projects and collaboration between national authorities allow the collection of discharge information by country as well as by catchment. The rest of this section reports the latest information on river flow to the Baltic Sea summarised by country. For Belarus, Denmark and Germany, no new information was available; see also BACC Author Team (2008).

5.3.2.1 Estonia

Reihan et al. (2007) analysed 14 daily streamflow records ranging in length from 45 to 98 years for the period 1903–2004. The tendency for a decrease in spring floods was notable in all periods for 80 % of stations over the continental regions. No trend was detected in spring, summer and autumn run-off, nor in summer droughts.

Long-term discharge trends were analysed by the Estonian Meteorological and Hydrological Institute (EMHI 2011) for the period 1957–2006 for 23 Estonian gauges. Trends were investigated by applying the Mann–Kendall test with a 5 % significance level. Annual mean run-off shows a significant positive trend over all Estonia except for the Narva river. The changes in annual run-off follow the changes in precipitation. Spring flood maximum discharge shows a significant negative trend over Estonia, while winter flood maximum discharge shows a significant positive trend over the country. Summer flood maximum discharge has increased in eastern Estonia, while a decrease has occurred in central and western parts of the country.

5.3.2.2 Finland

The published data on water level and flow regimes in the rivers of Finland extend until 2004 (Korhonen 2007; Korhonen and Kuusisto 2010). The monthly and annual mean outflow from the whole territory has been determined for the period 1912–2004, complemented by 25 discharge time series and 13 water level time series. Both unregulated and regulated rivers were examined. While most of the observation series examined started from the 1910s to 1930s, the longest continuous records date back to the mid-1800s.

The mean outflow from Finland in 1912–2004 was about 3300 m^{-3} s^{-1} (Fig. 5.10). The variation in annual mean discharge was 18 %. The lowest annual outflow from

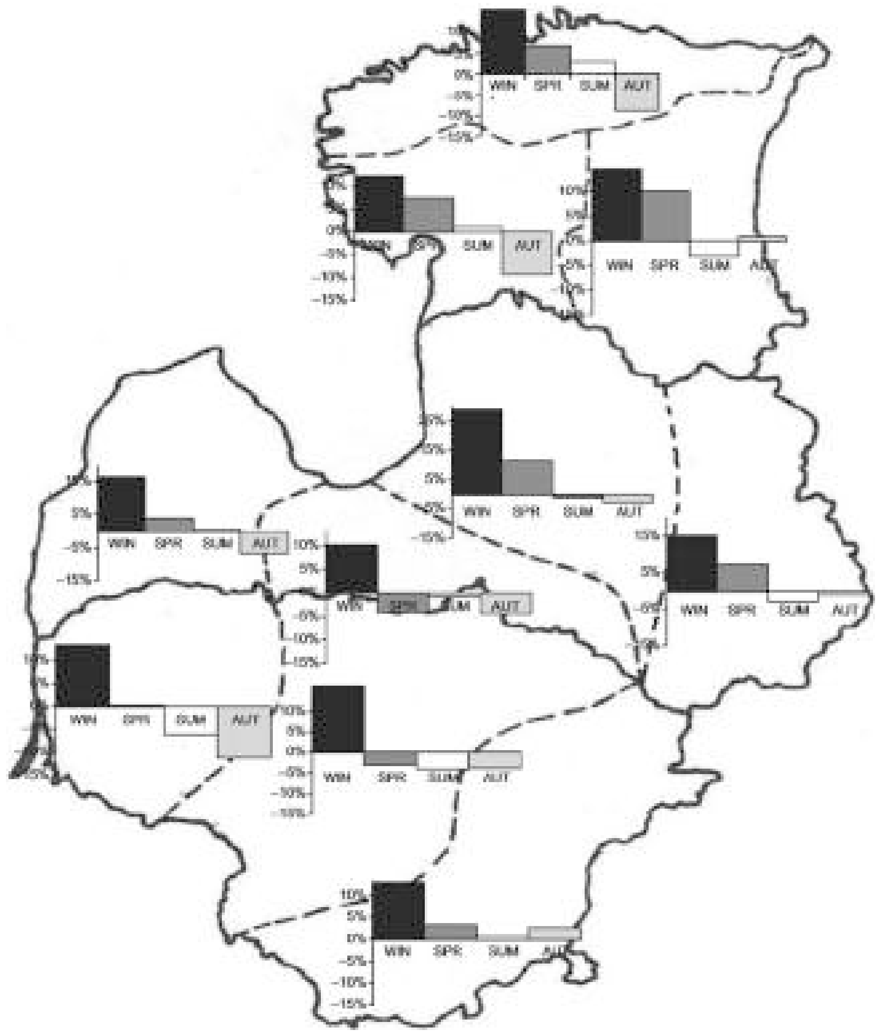


Fig. 5.9 Seasonal and regional anomalies in river discharge in the Baltic States in 1991–2007 relative to the reference period 1961–1990. *WIN* winter, *SPR* spring, *SUM* summer, *AUT* autumn (Kriaučiūnienė et al. 2012)

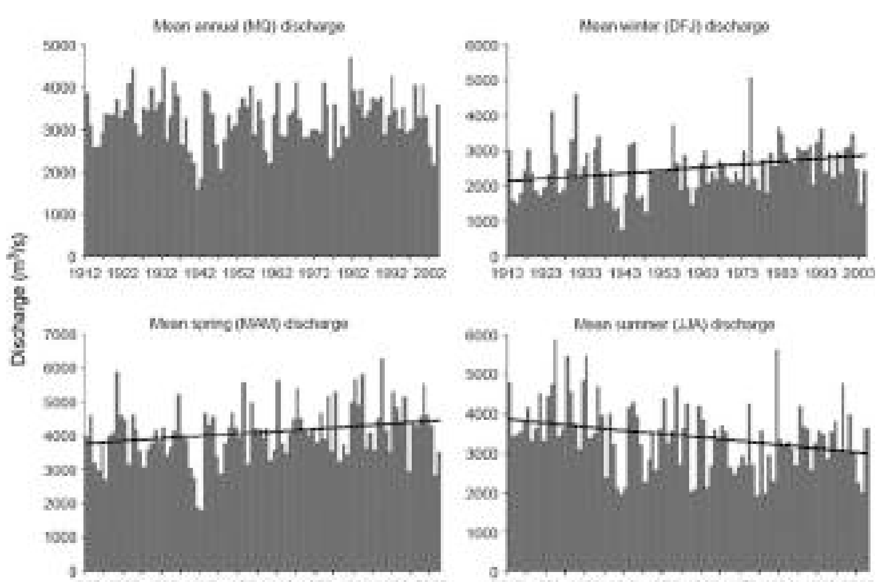


Fig. 5.10 Time series and trends in mean annual (MQ), winter (DJF), spring (MAM) and summer (JJA) outflow from Finland in 1912–2004. Only statistically significant trends are shown (Korhonen and Kuusisto 2010)

Finland was about 1600 m^{-3} s^{-1} in 1941 and the highest about 4700 m^{-3} s^{-1} in 1981, giving a ratio of about 3:1 between the highest and lowest annual outflow. The highest monthly outflows have normally been recorded in May during snowmelt floods, with a record value of about 10,350 m^{-3} s^{-1} in May 1920. The lowest monthly mean outflow was 640 m^{-3} s^{-1} in March 1942 giving a ratio of about 16:1 between the highest and lowest monthly outflow. Normally, outflow is lowest in winter and peaks in spring or early summer. The relative variation in monthly mean outflow was lowest in May (22 %) and highest in September (35 %).

The discharge regime has changed over the decades in response to both climatic fluctuations and human impacts, predominantly water regulation. Although there is no statistically significant change in the annual mean outflow from Finland for 1912–2004, climate change has affected the annual cycle of flow, particularly the seasonal distribution of flow. The most significant change has occurred in the hydrological regimes of winter and spring. Both seasons have become milder during the twentieth century (see also Chap. 4, Sect. 4.4), and consequently, late-winter and early-spring mean discharges have increased. However, the magnitudes of spring peak flow have not changed. Regulation has increased the winter and spring mean discharge in some places, while the summer flow has decreased. Winter and spring monthly mean discharges from Finland increased by 100–150 m^{-3} s^{-1} per decade during 1912–2004. June and July monthly mean discharges from Finland decreased by 85–195 m^{-3} s^{-1} per decade. Changes in seasonal discharge were different in different regions. Most drainage basins in Finland are affected by regulation, but especially in northern Finland and in Ostrobothnia. Winter and spring discharge increased mostly in the north, whereas summer discharge decreased specifically in southern Finland.

Long-term changes in the individual discharge time series were similar to the changes in the outflow from Finland. At most sites, the winter and spring mean discharges increased at both unregulated and regulated sites. However, in northern Lapland, it seems more likely that winter discharge decreased. The increase in winter discharge focused on late winter and the increase in spring discharge on early spring. The rise in winter and spring discharge can almost certainly be attributed to warming in winter and spring and earlier snowmelt. At some regulated sites, the release of water has been increased in winter and early spring in order to increase the storage capacity for snowmelt water. This explains the stronger winter and spring discharge trends at some regulated sites. Spring high flow occurred earlier within the year at about a third of the observation sites. In most cases, this shift in timing was 1–8 days per decade. There is no overall change in the magnitude of spring high flow. At a third of the unregulated sites, summer discharge has increased, whereas there has been a decrease in some monthly discharges at slightly less than half of the regulated sites. The decrease in summer discharge at regulated sites can be at least partly explained by higher water release in winter and spring. At about half of the unregulated observation sites, low flows have increased, at about half of the regulated observation sites they have decreased. Increase in low flow at unregulated sites can be explained by increased discharge in low-flow periods (winter and summer). Decrease in low flow at regulated sites is explained by zero flow when water gates are shut, but a similar situation is usually not possible in unregulated streams. Annual mean flow and annual high flow did not show statistically significant trends in general, apart from at a couple of sites. Changes in mean monthly or seasonal discharge were typically a small percentage of the period mean flow per decade, in most cases not higher than 10 %. Trends at regulated sites were stronger than trends at unregulated sites.

5.3.2.3 Latvia

Recent studies of run-off in Latvia include those by Apsīte et al. (2009, 2013). The authors studied a 1951–2009 data series of twenty-five river hydrological stations. The first 37-year period (1951–1987) showed 'no substantial' climate change impacts on river run-off, whereas the subsequent 22-year period (1988–2009) had a 'substantial' climate change signal in river run-off. In the period 1951–1987, a major part of the total annual river run-off (37–52 %) was generated during spring with a peak discharge of up to 30 % in April, followed by winter (17–30 %), autumn (17–25 %) and summer (9–14 %), with the lowest discharge (2–4 %) in July and August. In comparison with the study period of 1951–1987, the past two decades (1988–2009) showed a statistically significant mean increase of 11 % in river run-off during winter. In contrast, a statistically insignificant decrease took place in spring (8 %) and autumn (3 %). Summer showed little change.

Although run-off has increased significantly in January and February and decreased significantly in April and May over the past two decades, a major proportion of the total annual river run-off is still generated in spring (39 % on average) and discharge still peaks in April (17 % on average) (Fig. 5.11). Changes in seasonal and monthly river discharges were identified in all four hydrological districts studied (western, central, northern and eastern Latvia), but the greatest change occurred in the central district.

Trend analyses for both periods (1951–2009 and 1881–2009) suggested a statistically significant increasing trend for winter low flow (the 30-day minimum discharge) and significant decreasing trends for both the high discharge (the maximum discharge of the year) and the coefficient of uneven run-off distribution. Hence, the difference between the low-flow and high-flow discharges in spring has

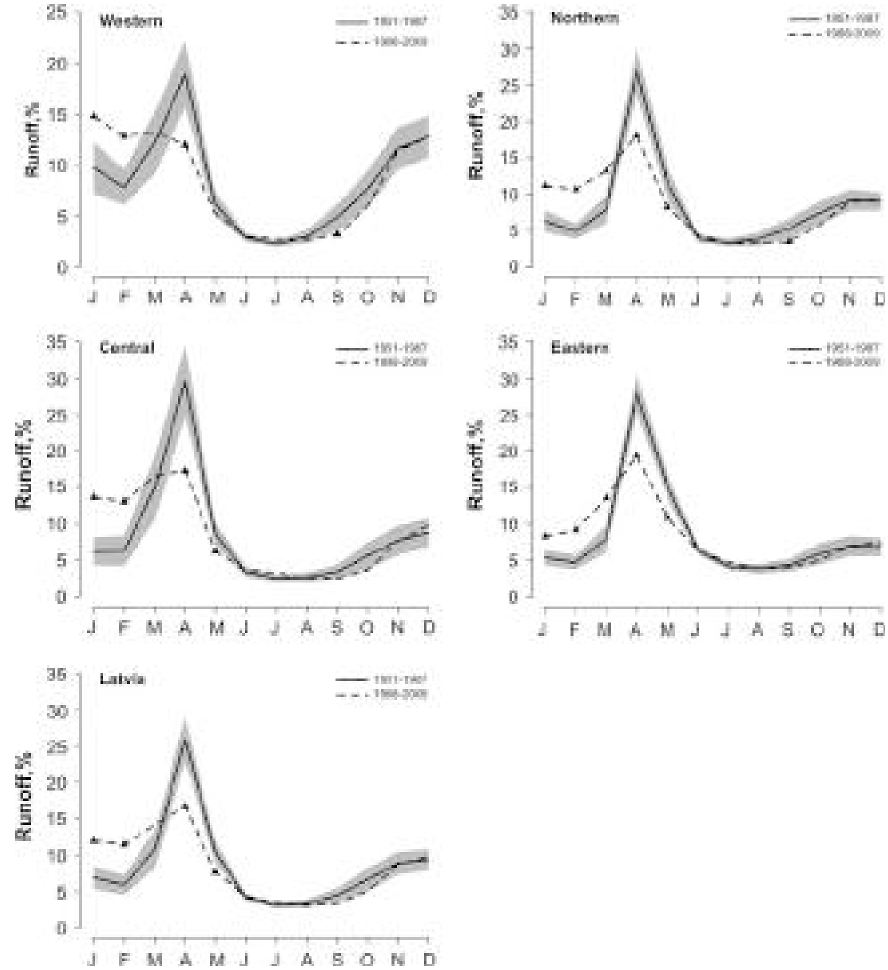


Fig. 5.11 Change in river hydrographs between two study periods in hydrological districts of Latvia and in Latvia as a whole. The *grey area* shows the 95 % confidence interval for the mean discharge values for 1951–1987. The *black triangles* represent statistically significant change in monthly mean discharge value for 1988–2009 (Apsīte et al. 2013)

decreased over recent decades, and the annual distribution of run-off has become more uniform. An example of long-term trends in the river Daugava annual discharge at the Daugavpils site is presented in Fig. 5.12, representing one of the longest observed data series in the Baltic countries.

5.3.2.4 Lithuania

The natural climate fluctuation in the Baltic Sea region, including Lithuania, depends partially on the processes of atmospheric circulation (see Chap. 4). Since the 1940s, deep cyclones have been more frequent across Lithuania during

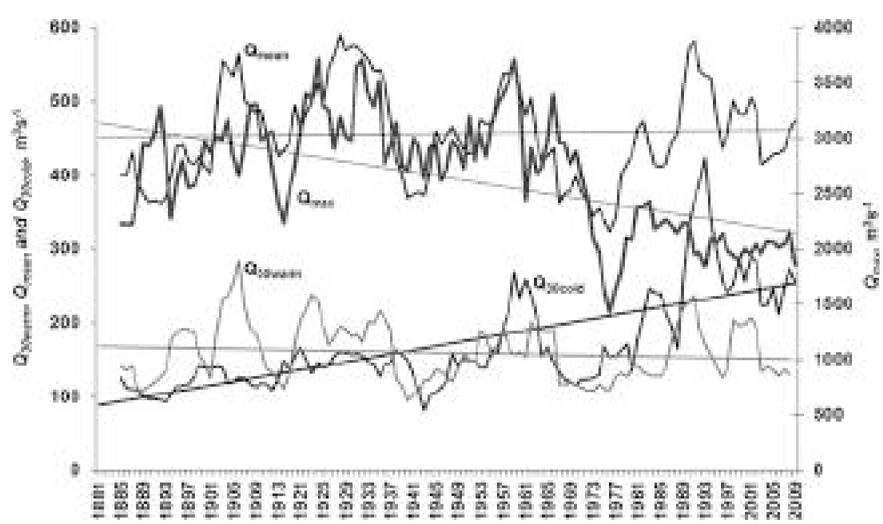


Fig. 5.12 Trends in the 30-day minimum discharge in low flow of cold ($Q_{30\ cold}$) and warm ($Q_{30\ warm}$) periods, annual mean discharge (Q_{mean}) and maximum of the year discharge (Q_{max}) at the hydrological station Daugava–Daugavpils for the period 1881–2009. Discharge curves are smoothed with a 5-year moving average (Apsīte et al. 2013)

winter (Bukantis et al. 2001). Consequently, winters have become warmer, long-lasting seasonal frost has decreased, and the contrast between seasons has diminished. Unusually, warm periods and the amount of winter precipitation increased in the final decades of the twentieth century in Lithuania.

Meilutytė-Barauskienė et al. (2010) studied Lithuanian rivers with a catchment area larger than 500 km^2 using data from 17 meteorological and 32 hydrological stations. Changes in air temperature from 1961–1990 to 1991–2007 in Lithuania can be summarised as follows (Meilutytė-Barauskienė et al. 2010): average annual temperature increased by 1.1 °C, average winter and spring temperatures increased by 0.57 and 0.7 °C, respectively, and the temperature contrast between seasons decreased. Annual and winter precipitation increased by 3 and 17 %, respectively. Spring, summer and autumn precipitation showed little change.

Winter run-off increased and spring run-off decreased during the final decades of the twentieth century with only small changes during summer and autumn. Meilutytė-Barauskienė et al. (2010) analysed the run-off time series for annual, seasonal and extreme events using the Mann–Kendall trend test. A summary of significant positive and negative trends is illustrated in Fig. 5.13. A significant negative trend was found in annual and spring, summer and autumn run-off over the period 1922–2003. The same tendency is characteristic for 1941–2003, while winter run-off increased in 54 % of stations. In the period 1961–2003, winter run-off again increased, whereas spring run-off and maximum discharge decreased in 85 % of stations. The smallest changes were found for autumn and summer, especially the driest month, July.

Findings by Kriaučiūnienė et al. (2008, 2012) demonstrated an uneven temporal and spatial run-off distribution in Lithuania. Annual river run-off varies from 4.2 to 14.0 l s^{-1} km^{-2}. Lithuania has been divided into three hydrological regions (western, central and south-eastern) according to different types of river feeding (precipitation, snowmelt, groundwater) and hydrological regime. Long-term regional series of temperature, precipitation and run-off compiled for these hydrological regions were analysed against the reference period 1961–1990. Annual temperature for the period 1991–2006 was about 15 % higher than during the reference period, whereas precipitation varied regionally from −3 % (western region) and −6 % (central region) to +4 % (south-eastern region). There was no significant change in run-off in 1991–2006 relative to the reference period. Climate change (increasing air temperature and decreasing summer precipitation) has had an impact on drought characteristics of river run-off (frequency, duration and discharge) in Lithuania. Change in low-flow conditions in Lithuanian rivers was analysed by Kriaučiūnienė et al. (2007). The drought data series were defined as a series of 30-day minimum discharge for the summer low-flow period. Representative historical daily data series from 30 hydrological stations were used to calculate trends. Three time periods (1922–2003, 1941–2003 and 1961–2003) were selected for analysis by the Mann–Kendall trend test. The cyclic variations in the time series of 30-day minimum discharge were typical for all Lithuanian rivers. The average periodicity of the cycles was found to be 27 years, including an average 'wet period' of 13 years and 'dry period' of 14 years. This is in line with the findings of Kriaučiūnienė et al. (2012) who found long-term variations in the regional time series of precipitation and discharge in all ten regions studied in Lithuania. The average periodicity of the wet and dry phases was 27–30 years, including an average wet period of 15 years and an average dry period of 14 years

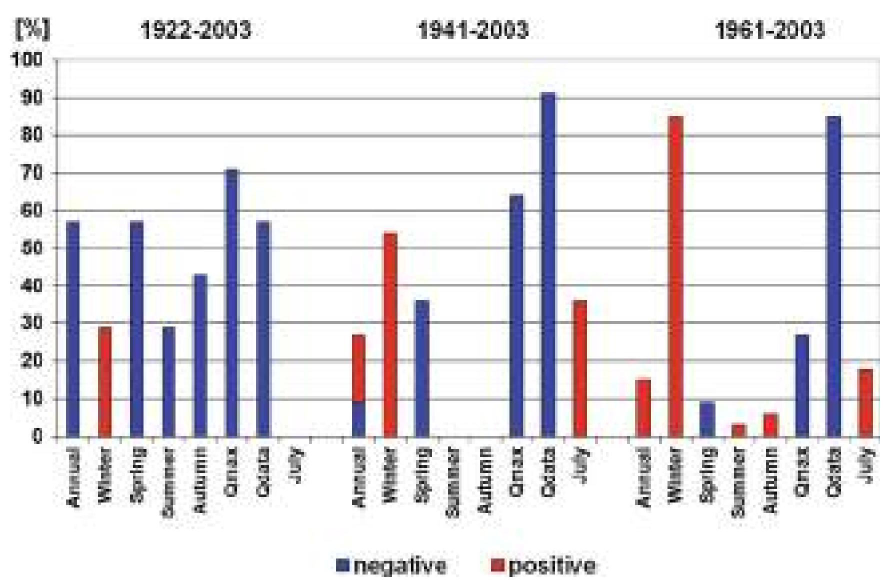


Fig. 5.13 Significant trends in annual, seasonal and extreme run-off from rivers in Lithuania for three periods over the past century. The *vertical axis* shows the percentage of stations encountering a positive (*red*) or negative (*blue*) significant trend (Meilutytė-Barauskienė et al. 2010)

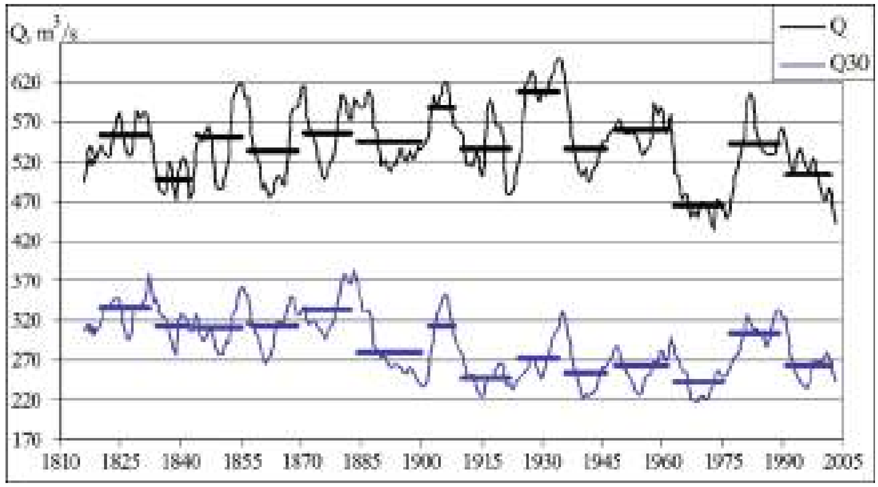


Fig. 5.14 Periodicity in annual (Q) and 30-day minimum discharge ($Q30$; 5-year moving average) in Nemunas, Lithuania (Kriaučiūnienė et al. 2007)

(see also Chap. 4, Sect. 4.5). The final period (1996–2007) of long-term variation in river discharge was considered ‘dry’, although was only 3 % lower than the multi-year average discharge. This small decrease in discharge is believed to stem from higher temperature despite a small (2 %) increase in precipitation in 1996–2007. The periodicity of variations in annual discharge and 30-day minimum discharge is similar (Fig. 5.14).

The analysis of minimum run-off using the Mann–Kendall trend test revealed no trends in annual variation in 30-day minimum discharge for the periods 1922–2003 and 1941–2003. Only a few rivers in western Lithuania had positive trends in the period 1961–2003. In Lithuanian rivers, droughts (30-day minimum discharges) occur in the period June to August. Since 1961, droughts have been observed less frequently in June and more frequently in August.

5.3.2.5 Poland

About 99.7 % (312,683 km^2) of Poland lies within the catchment of the Baltic Sea. Analysis of the total run-off from Poland during the period 1951–2010 shows a slight increasing tendency (Fig. 5.15).

Vistula (Wisla) is the longest river flowing into the Baltic Sea and has the second highest run-off after Neva river. The Vistula river catchment covers about 54 % of Poland. Analyses of Vistula discharge during 1921–2006 do not show any increasing tendency, but a periodicity of 13–18 years. Run-off is stable over the long term, ranging from a minimum of 18.76 km^3 in 1943 to a maximum of 51.09 km^3 in 1975. However, the past 20 years show an increasing tendency with the long-term maximum for 1921–2006 (51.09 km^3) exceeded in 2010 (54.58 km^3).

The Odra is the second longest river in Poland. Discharge at the Godzdovice gauging station has shown a slight increase over the past couple of decades although there is no trend over the full data series (1921–2006) (Institute of Meteorology and Water Management of Poland).

5.3.2.6 Russia

The Neva is the largest of the rivers flowing into the Baltic Sea, accounting for about 18 % of total run-off (Table 5.1). Figure 5.16 presents monthly and annual run-off for the Neva and Volkhov rivers and shows considerable interannual variability (Filatov et al. 2012). During the final decade, the annual means vary widely from a maximum of 2861 m^3 s^{-1} in

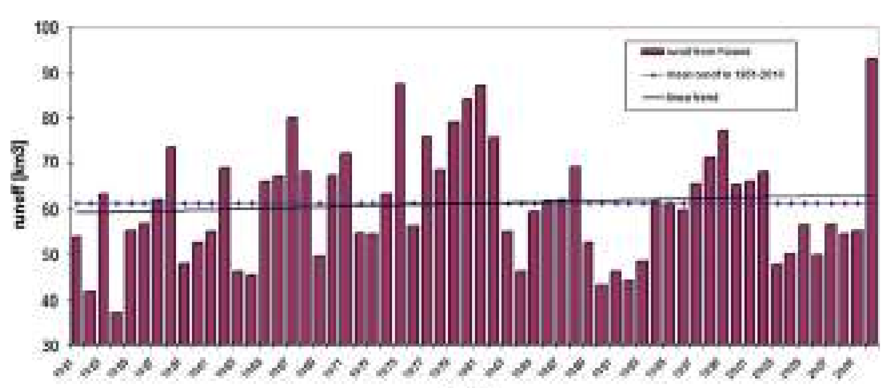


Fig. 5.15 Annual run-off from Poland in the period 1951–2010 (Institute of Meteorology and Water Management of Poland)

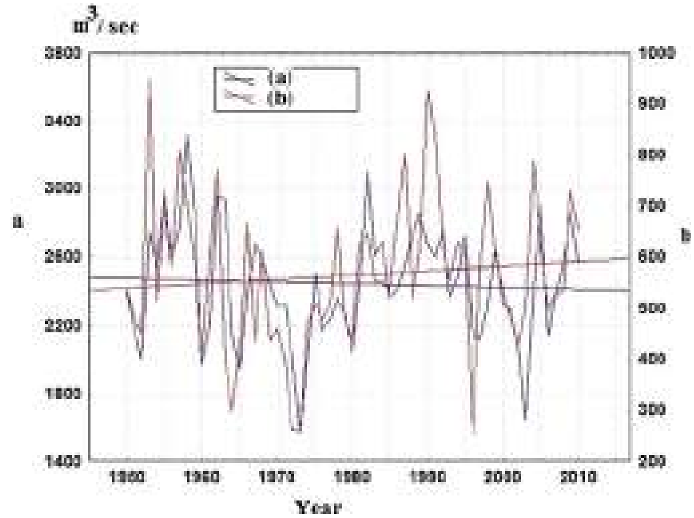


Fig. 5.16 Annual discharge for the Neva (a, left axis) and Volkhov (b, right axis) between 1950 and 2010 (Filatov et al. 2012)

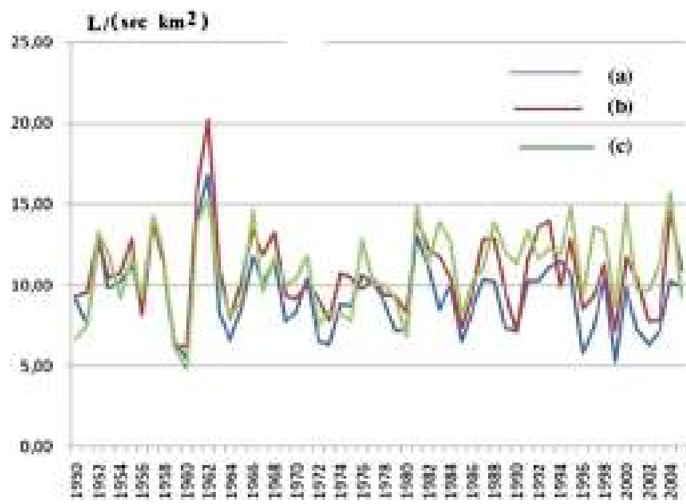


Fig. 5.17 Annual river run-off for the Shuja (a), Suna (b) and Vodla (c) between 1950 and 2006 (Filatov et al. 2012)

2000 to a minimum 1649 m^3 s^{-1} in 2003. No statistically significant trends were found in the run-off data, for the Neva or Volkhov rivers, or for the Shuja, Suna or Vodla rivers (Fig. 5.17). Recent decades have been characterised by a combination of high temperature and high run-off.

The Neva river is heavily regulated using the lakes of the catchment, and the share of winter low water accounts for about 25 % of annual run-off (Gronskaja 2008). The river run-off ratio between wet and dry years is roughly 2:1. The seasonal river run-off distribution has not changed but spring (AM) run-off has increased by 6–13 % while winter (DJF) run-off has decreased by 4–10 % (Shiklomanov 2008).

5.3.2.7 Sweden

Average long-term series of temperature and precipitation for Sweden were constructed by Alexandersson (2002). These data series were extended and supplemented by hydrological data by Lindström and Alexandersson (2004), Hellström and Lindström (2008) and Lindström (2011), for example. Figure 5.18 shows annual anomalies and long-term variations in precipitation, temperature, water resources and flood magnitude in Sweden for the period 1901–2010. The anomalies are relative to mean values for the reference period 1961–1990. Temperature has been unusually high in recent years, with a temperature anomaly of about 1 °C for Sweden as a whole. From 1988 onwards, all years except two have been warmer than the average for the reference period. Precipitation has increased much more than run-off. Hellström and Lindström (2008) were not able to establish the reason for this discrepancy, but suggested that the single most likely factor was a change in the precipitation measurement technique. Run-off was fairly stable over the study period. The 1970s was the driest decade (with a deficit of about 10 % compared to the reference mean), while the 1920s, 1980s and 1990s were the wettest (about +10 %). 2000 was the wettest year, with a precipitation anomaly of almost 40 %. The magnitude of the annual maximum floods in unregulated rivers was relatively stable over the study

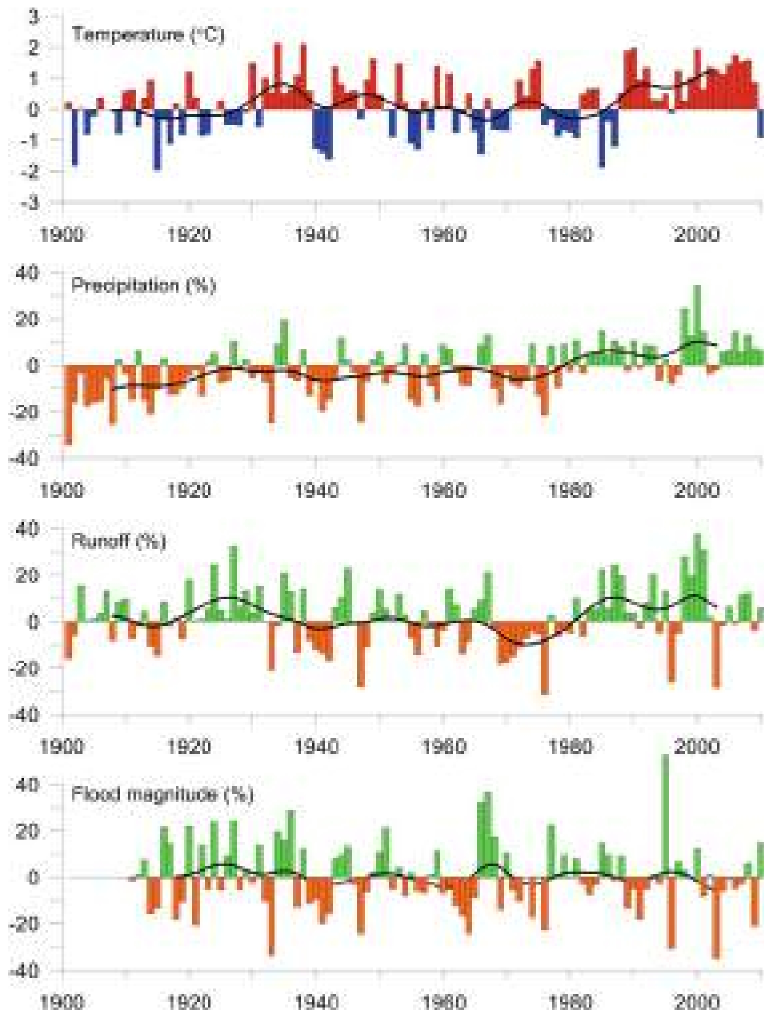


Fig. 5.18 Annual anomalies (relative to 1961–1990) and long-term variability in precipitation, temperature, runoff and flood magnitude in Sweden for 1901–2010. For flood magnitude, the years before 1911 were omitted due to data scarcity (Hellström and Lindström 2008)

period. 1995 was the year with highest floods on average, and 2003 was the year with lowest flood peaks.

5.4 River Ice Regime

The river ice regime is considered a sensitive indicator of climate change. Kļaviņš et al. (2007, 2009) studied long-term changes in ice break-up date and duration of ice cover for 17 river stations in the Baltic countries and Belarus. River Daugava has the longest data series in Europe starting in 1530. A pronounced decreasing trend is apparent for the past 150 years and is even more clearly evident over the past 30 years (Fig. 5.19). No decreasing trend was detected for the initial period (starting 1530), which includes the Little Ice Age. The records of historical observations show recurring episodes of mild and severe winters (see Chap. 4 for the past 200 years and Chap. 3 for the past 1000 years, including the Little Ice Age).

A decreasing linear trend indicates a reduction in ice-cover duration and earlier ice break-up. The ice-cover duration has declined by 2.8–6.3 days per decade during the past 30 years. In general, the shift in river ice break-up

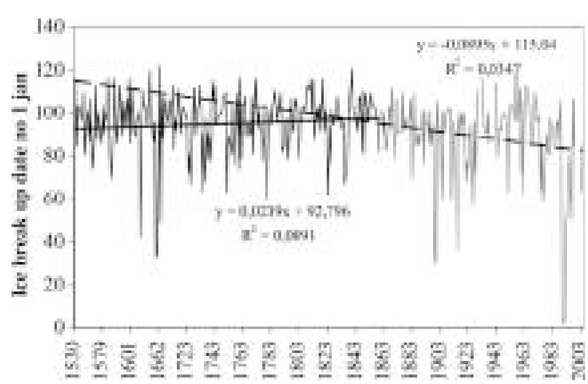


Fig. 5.19 Time series of ice break-up dates on River Daugava (*dashed line* shows trend from 1860 to 2003 and *continuous line* from 1530 to 1859) (Kļaviņš et al. 2009)

towards earlier dates, indicating an earlier start in river flooding, is in agreement with the increase in winter run-off for rivers in the Baltic Sea basin. The time of ice break-up depends not only on meteorological conditions in a given year and distance from the Baltic Sea, but also on global climate, making it a good integrating indicator of global climate change.

Both the ice regime and the seasonal river discharge are strongly influenced by large-scale atmospheric circulation processes over the North Atlantic, and this is manifested through a close correlation with the NAO (see Chap. 4). A strong negative correlation between the NAO index, Winter Baltic Climate Index (WIBIX, Hagen and Feistel 2005) and ice break-up events suggests that atmospheric processes over the North Atlantic were the main factor driving the river ice regime in the Baltic region for the period 1921–2000 (Kļaviņš et al. 2009). In winter, intense westerly circulation moves fronts and air masses through the mid-latitudes and across the Baltic Sea region, but during the warmer parts of the year it weakens considerably and precipitation events are largely due to different processes.

Stonevicius et al. (2008) reported a decrease in ice-cover duration in the lower reaches of the Nemunas River, Lithuania, over the past 150 years. They argued that variation in the river freeze-up and break-up dates is related to variation in climatic variables. The negative break-up trend exceeds that of the positive freeze-up trend. Low-frequency, large-scale atmospheric circulation patterns such as the NAO and the Arctic Oscillation (AO) (see Chap. 4) appear to have more influence on the break-up date than on the freeze-up date.

Šarauskienė and Jurgelėnaitė (2008) studied ice-cover data for 13 water measurement stations on eight rivers also in Lithuania. Variation in ice-cover data was investigated for three periods: 1931–1960, 1961–1990 and 1991–2005. The study suggests that warmer winters cause later freeze-up dates and shorter ice-cover duration on the rivers. Long-term observations of the Nemunas (1812-2006) at the Smalininkai station indicate that in the nineteenth century, ice cover formed on average 13 days earlier and remained unbroken on average for 30 days longer than in the twentieth century. Applying the Mann–Kendall trend test to the data showed very significant trends in the Nemunas ice data series of the last century: a negative trend in ice duration data and a positive trend in freeze-up date data. According to the ice-cover data, the Kaunas Hydro Power Plant has the largest anthropogenic impact on the Nemunas ice processes. After the dam was constructed in 1959, ice duration at Kaunas and Lampėdžiai WMS decreased on average by 15 and 5 days, respectively, compared to the period 1931–1960. Observational data from 1991 to 2005 indicate a strong decrease in ice duration compared to earlier periods. In urban areas, human activities cause rivers to freeze-up later and the period of ice cover to be shorter.

In Russia, the analysis of the interannual variation in ice phenomena dates and the duration of coverage for Lake Ladoga show very small but significant trends (1–6 days per 100 years). The trend is most marked for freeze-up dates (14 days per 100 years) but the coefficient of determination is no more than 3 % (Karetnikov and Naumenko 2008). Change in the thermal regime of the near surface water layer is revealed by the increase in duration of the ice-free period on Lake Onego (Onega) (Salo and Nazarova 2011). By the end of the twentieth century, the number of ice-free days had increased from 217 to 225 days on average (Fig. 5.20). The authors argued that the dates of ice-cover formation and decay on Lake Onego are determined not only by autumn and spring temperatures but also by large-scale processes characterised, for example, by the NAO index. Figure 5.21 illustrates the

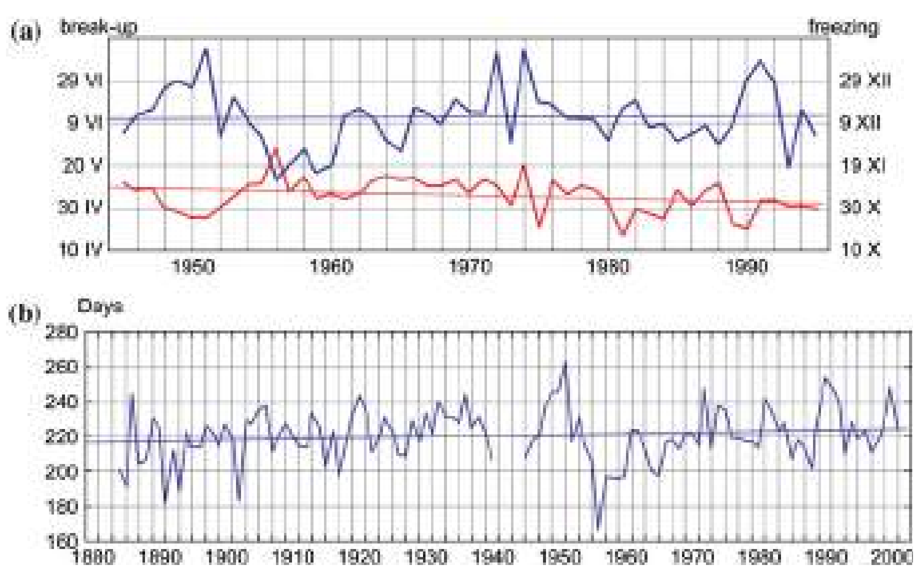


Fig. 5.20 Dates of **a** break-up (*red*) and freezing (*blue*), and **b** duration of ice-free period on Lake Onego (redrawn from Salo and Nazarova 2011)

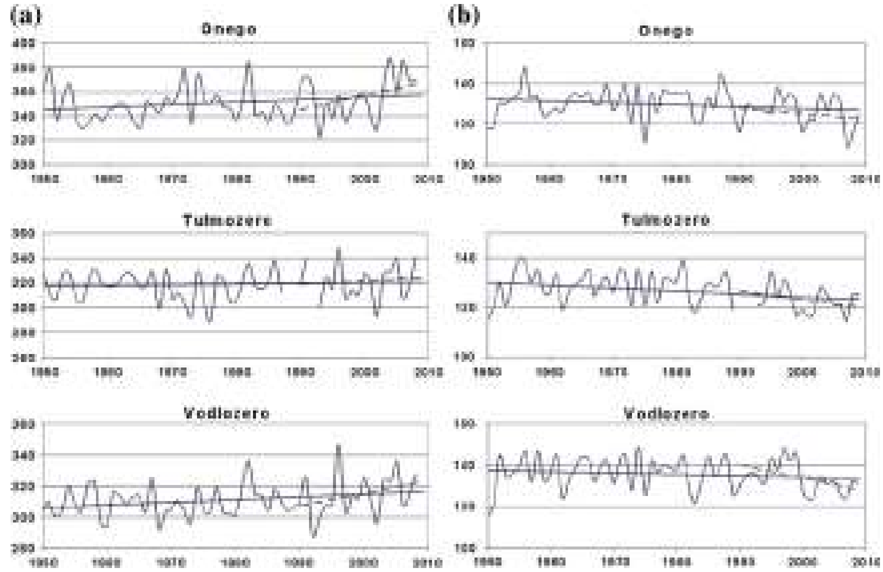


Fig. 5.21 Date of **a** freeze-up and **b** break-up on a large lake (Onego), a middle size lake (Vodlozero) and a small lake (Tulmozero) for 1950–2009. The linear trends for 1950–2009 and 1990–2009 are shown by the *solid* and *dashed line*, respectively (Efremova and Palshin 2011)

date of freeze-up and ice break-up on three lakes of different sizes in Russia (Efremova and Palshin 2011).

5.5 Conclusion

The recently published literature and data collected by national authorities on river hydrology of the Baltic Sea basin support and are generally in line with the findings reported in the first assessment of climate change in the Baltic Sea basin (BACC Author Team 2008). No statistically significant trends have been detected in annual river discharge in the Baltic Sea basin as a whole, and temperature change seems to explain change in run-off better than does precipitation. Winter discharge has increased due to higher temperatures and subsequent snowmelt, while spring discharge has decreased as less snow is present. Regional variations occur in response to climate (temperature, precipitation) and discharge cyclicity. Later start dates for ice formation on rivers and earlier ice break-up dates have resulted in shorter ice duration on rivers in the Baltic Sea basin.

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6 Recent Change—Terrestrial Cryosphere

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Abstract

This chapter compiles and assesses information on recent and current change within the terrestrial cryosphere of the Baltic Sea drainage basin. Findings are based on long-term observations. Snow cover extent (SCE), duration and amount have shown a widespread decrease although there is large interannual and regional variation. Few data are available on changes in snow structural properties. There is no evidence for a recent change in the frequency or severity of snow-related extreme events. There has been a decrease in glacier coverage in Sweden and glacier ice thickness in inland Scandinavia. The European permafrost is warming, and there has been a northward retreat of the southern boundary of near-surface permafrost in European Russia.

Keywords

seasonal snow cover • glaciers • seasonally frozen ground • permafrost

6.1 Introduction

The terrestrial cryosphere of the Baltic Sea drainage basin includes widespread seasonal snow cover and frozen ground as well as small glaciers in Sweden and in a few cases, Norway. In addition to long-term changes in climate, various components of the terrestrial cryosphere are affected by the seasonal changes in weather, especially winter air temperature and the form of the precipitation (liquid or solid).

This chapter updates information presented in the first Baltic Sea assessment (BACC Author Team 2008) on recent and current change in snow cover, and other components of the terrestrial cryosphere, based on the recent literature. Findings are based on long-term observations mostly from the European part of Russia, Estonia, Latvia, Lithuania, Poland, Finland and Sweden. Some information is available from Denmark, Norway and Germany, which all cover relatively small areas of the Baltic Sea drainage basin. Only small parts of Ukraine, Czech Republic and Slovakia are within the drainage basin, and no information on changes within the terrestrial cryosphere in these countries was available. The same is true for Belarus, although a significant proportion of the country is within the drainage basin. Some

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_6

findings are valid for the whole Baltic Sea drainage basin, while for others the perspective is restricted to limited areas only—glaciers contributing run-off to the Baltic Sea drainage basin found only in Sweden and a few remote regions of Norway—and permafrost is a marginal phenomenon.

Box 6.1 Summary of findings from the first assessment of climate change in the Baltic Sea basin

The first Baltic Sea assessment (BACC Author Team 2008) reported several climate-related changes in the snow cover of the Baltic Sea drainage basin. Study periods ranged from the period 1961–1990 to the whole of the twentieth century. Although winter air temperatures were observed to rise across all of northern Eurasia, impacts on snow cover varied across the Baltic Sea basin. A decrease in snow cover was observed in the south-western regions of the drainage basin, due to an increase in the proportion of precipitation in liquid form during winter, while an increase in snow storage and snow cover duration (SCD) was observed in the north-eastern regions. Most of the drainage basin has experienced earlier snow melt and a decrease in spring snow cover due to the rise in temperature. A recent decrease in SCD and snow water equivalent (SWE) was observed in the southern parts of all Fennoscandian countries. Despite this, total snow storage increased in the east and north. In the Scandinavian mountains, an increase in winter precipitation related to thicker snow cover was observed. In Estonia, a recent negative trend was observed in SCD, snow depthand SWE. A decrease in snow cover days was observed in Latvia, Lithuania and Poland. In the north-west of the East European Plain, snow storage increased in line with winter temperature and precipitation.

Since the first BACC assessment (BACC Author Team 2008), there have been three other significant assessments with at least some emphasis on northern European cryospheric conditions. The Global Outlook for Snow and Ice (UNEP 2007) reported that the Northern Hemisphere mean monthly SCE had declined at a rate of 1.3 % per decade during the past 40 years. It also reported a long-term increase in snow depth and SCD across most of northern Eurasia. A decreasing trend in SCE during winter in the Northern Hemisphere was also reported by Lemke and Ren (2007) in the Fourth Assessment report of the Intergovernmental Panel on Climate Change. The authors also concluded that lowlands in central Europe had seen recent reductions in annual SCD, whereas greater snow depth but a shorter snow season had been observed in Finland and the former Soviet Union. An assessment by Voigt et al. (2010), mostly concerning alpine and central European conditions, concluded that higher winter temperatures are the main reason for the decrease observed in snowfall and snow depth in most of the Europe.

Snow cover affects the winter and spring climate in many ways. For example, snow has a high albedo and so absorbs much less solar radiation than bare soil or vegetated surfaces; it also acts as a heat sink during the melt, keeping ground temperature near zero despite the high radiative fluxes (BACC Author Team 2008). When the snow cover is formed on tundra vegetation, snow depth determines whether the vegetation is still visible (Heino et al. 2006; Euskirchen et al. 2007). One example is from northern Scandinavia, where satellite data have been used to map the snowmelt date, and to evaluate the effect of low vegetation on snowmelt and through this, on surface albedo. During the study period of 1995–2011, more abundant low vegetation in northern Norway caused snowmelt to occur earlier than in northern Finland—causing the Finnish side of the border to have a greater surface albedo during the melt period. Reindeer grazing is one reason for differences in vegetation amount (Cohen 2011). A modelling study that eliminated snow cover from the climate system found that this resulted in higher mean annual surface air temperature; decreased soil temperature and increased permafrost extent; drying of upper-layer soils and changes in the annual cycle of run-off; and the disappearance of extreme cold air outbreaks (Vavrus 2007). Variability in SCE in Europe affects low-level atmospheric temperature, soil temperature, soil moisture, stream discharge and energy flow in the warming and melting of the snowpack (Henderson and Leathers 2010).

At a larger scale, Eurasian SCE affects the Northern Hemisphere winter circulation (Orsolini and Kvamstø 2009). The extent of autumn snow cover in Eurasia has been shown to influence the atmospheric circulation over the Northern Hemisphere during the following winter, and even the North American winter temperatures (see Chap. 4, Sect. 4.2.4). During winters 1967/1968–2007/2008, autumn snow cover from northern Scandinavia to the West Siberian Plain was associated with winter temperatures over the interior of North America (Mote and Kutney 2011). A relationship has also been observed between winter and spring Eurasian snow cover and spring and summer East Asian rainfall (Wu and Kirtman 2007).

Changes in seasonal snow cover (amount, extent and duration), glacier mass balance and frozen ground have various ecological and socio-economic consequences (see Chaps. 15–22). The terrestrial cryosphere has close connections to the hydrological regime (i.e. river run-off) described in Chap. 5.

Box 6.2 Observing the cryosphere

Observations of snow cover comprise (1) in situ measurements of snow fall, snow depth, SWE and/or snow structure; (2) airborne remote sensing observations; or (3) space-borne satellite observations. In situ observations are normally operated by hydrological and/or meteorological services. They are local in nature and not uniformly distributed. Data quality is affected by, for example, changes in station location and observation methodology. Snow observations have been made operationally in many countries for several decades.

The use of satellite observations for studying the cryosphere has been discussed by Sharkov (2003), Nosenko et al. (2005) and Sutyrina (2011). Satellites provide the opportunity to observe large-scale SCE by optical satellite imagery or by radar remote sensing. Snow depth and SWE are derived from data obtained from passive microwave sensors.

Cloud cover and highly variable illumination conditions (including the polar night) impede the use of monitoring methods reliant on reflected solar radiation. Dense forest cover and deep snow hinder the use of passive microwave sensors. Studies also reveal significant differences between remotely sensed and in situ observations of snow cover, for example owing to the algorithms that convert brightness temperatures observed from satellites to SWE. These problems were discussed by Boyarskii et al. (1994), Boyarskii and Tikhonov (2000), Luojus et al. (2009), Kitaev (2010), Metsämäki et al. (2010) and Kitaev and Titkova (2011). There are sometimes large discrepancies between the satellite data and the ground-based instrumental observations of the snow cover boundary during periods of snow formation and snowmelt. Radar-based remote sensing is problematic during the snowmelt season, and it is difficult to find a single method that functions well in mountainous, open and forested areas. The use of wide-swath synthetic aperture radar (SAR) was promising especially in the boreal zone (Khan et al. 2007; Lemke and Ren 2007; Luojus et al. 2007, 2009). Kärnä et al. (2007) and Takala et al. (2011) recently presented a snow mapping procedure that combines weather station measurements and microwave radiometer data.

Observations of structural snow parameters (stratigraphy, density, hardness, grain size and form, impurities) are far less frequent and regular than observations of the occurrence and amount of snow. An overall understanding of typical snow structures within the various parts of the Baltic Sea drainage basin is not available, although some fragmentary information on phenomena such as rain-on-snow and ice crusts is occasionally published.

The Swedish glaciers have been investigated since the early expeditions at the start of the twentieth century (e.g., Williams and Ferrigno 1993; Klingbjer and Neidhart 2006), and since 1946, there has been a systematic monitoring programme in northern Sweden (Holmlund et al. 1996; Jansson and Pettersson 2007). Within the scientific monitoring programme of the Tarfala Research Station, the yearly mass balance is calculated for five different glaciers in the region and the frontal positions of 20 glaciers are monitored.

Current understanding of the spatial patterns of frequency, intensity and duration of freeze/thaw cycles of the ground in the Baltic Sea region remains poor and has not been subject to systematic study. Observations include measurements of frozen ground depth and permafrost (e.g., active layer thickness). Permafrost temperatures are monitored either relatively close to the ground surface or in boreholes of 100 m or more deep (Lemke and Ren 2007).

Reanalysis products such as ERA40 (reanalysis for 1957–2002 made by the European Centre for Medium-Range Weather Forecasts) and NCEP/NCAR (continually updated data set from 1948, produced by the National Centers for Environmental Prediction and the National Center for Atmospheric Research) offer new possibilities for understanding the present state and recent change in the terrestrial cryosphere. According to studies by Khan et al. (2007, 2008) on reanalysed snow data from major Russian river basins for 1979–2000, the method reproduces the observed seasonal and interannual snow cover variability well, even though the absolute values may differ.

6.2 Recent and Present Change in Seasonal Snow Cover

6.2.1 Snow Cover Formation, Duration and Melt

According to Brown and Mote (2009), SCD has the highest sensitivity to climatic change of all snow cover parameters. Observations by NOAA polar orbiting satellites show a decrease in Northern Hemisphere terrestrial SCD during the period 1966–2007. The greatest decreases occurred in areas where the seasonal mean air temperature was between −5 and +5 °C (i.e., the mid-latitudinal coastal regions of the continents). Choi et al. (2010) found that the average Northern Hemisphere SCD decreased by 5.3 days per decade between the winters of 1972/1973 and 2007/2008. The most significant change occurred in the late 1980s. Takala et al. (2009) showed that the European regions of Eurasia experienced an increasing early melt onset between 1979 and 2007. At Sodankylä, northern Finland, melt onset advanced by 3.4 days per decade.

These findings are supported by an analysis in snow survey observations in northern Eurasia by Bulygina et al. (2011). They found a decrease in SCD since 1966. Also the spring snowmelt has become shorter and more intense in northern Eurasia, even though the maximum snow depth has increased across most of Russia.

SCD between 1976 and 2008 was shorter than the 1938–2008 average in the eastern part of the Baltic Sea region. This followed the sharp rise in air temperature and the increase in precipitation that began in 1976 (Fig. 6.1; Table 6.1). Furthermore, the INTAS-SCCONE project (International Association for the Promotion of Cooperation with Scientists from the New Independent States of the former Soviet Union—Snow Cover Changes Over Northern Eurasia) showed that SCD decreased within the Baltic Sea basin in western Scandinavia and in the south-west of the East European Plain over the past century. This corresponds to climatic conditions over the northern part of the East European Plain (ROSHYDROMET 2008; adapted from Kitaev et al. 2007, 2010). Conversely, an increasing trend in the number of snow cover days is seen in most of northern Eurasia (Heino et al. 2006).

Since the mid-twentieth century, SCD in Latvia has decreased by 3–27 days (Draveniece et al. 2007; Kļaviņš 2007; Kļaviņš et al. 2009). Mean SCD decreased by 17 days in Lithuania during 1961–2010. Only in the most eastern part of Lithuania was a positive tendency seen (Gečaitė and Rimkus 2010).

SCD in Poland increases from west to north-east and is greatest at high altitude in the mountains. A long time series (80 winters) showed no significant trends in SCD. However, in the latter half of the twentieth century, a slight negative trend occurred across most of the Poland, excluding the highest parts of the Sudety Mountains (Falarz 2010). In southern Poland, the foehn effect has a strong influence on snow cover (Falarz 2013). Northern Germany has very variable snow cover with the mean number of days with snow cover varying from less than 15 in the west to almost 40 in the east. The latter half of the twentieth century saw a decline in the number of days with snow cover in northern Germany (Bednorz 2007).

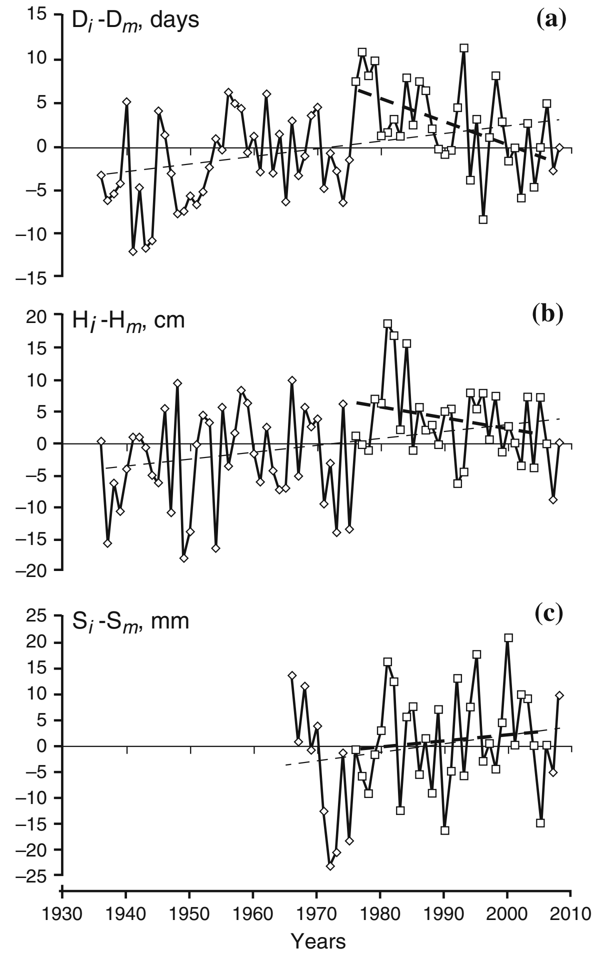

Fig. 6.1 Snow cover anomalies in the Russian part of Baltic Sea basin. **a** Snow cover days. **b** Snow depth in March. **c** Snow water equivalent in March. *Dots* show anomalies relative to the long-term average for 1938–2008. *Dotted lines* show linear trends (ROSHYDROMET 2008; Kitaev et al. 2007, 2010)

During the period 1905–2003, SCD did not change significantly in Sweden (northern part of the Baltic Sea basin). For the period 1961–2003, the number of days with snow cover decreased in southern Sweden by 20–40 %. In southern Sweden, seven of the ten years with the shortest SCD have occurred after 1974, and in mid-Sweden, five of the ten years with the shortest SCD have occurred since 1989 (Larsson 2004). In Norway, there has been a decrease in SCD at the majority of monitoring stations since 1914. This negative trend is more pronounced in the past few decades (Dyrrdal and Vikhamar-Schuler 2009; Dyrrdal et al. 2013).

The monthly key climatic reports from the period 1970–2009 collected by Cappelen (2000, 2003a, b, 2010) showed that Denmark (western part of the Baltic Sea drainage basin) has an ephemeral snow cover and that SCD varies greatly between years and between decades. Both the minimum (8.3) and maximum (54.2) number of snow cover days per

Table 6.1 Long-term (1936–2008) and recent (1979–2006) trends in monthly observations on anomalies of air temperature (*T*), precipitation (Pr), snow depth (*H*) and snow water equivalent (*S*) for the Russian part of the Baltic Sea drainage basin (ROSHYDROMET 2008; adopted from Kitaev et al. 2007, 2010)

Month	Period	*T*		Pr		*H*		*S*	
		b	R^2	*b*	R^2	*b*	R^2	*b*	R^2
Nov	1936–2008	0.10	1.9	3.87	30.0	1.3	20.9	−2.4	0.0
	1979–2006	0.12	3.2	2.05	3.5	8.4	1.9	3.1	5.1
Dec	1936–2008	0.10	1.4	7.52	30.5	1.3	20.1	1.2	0.0
	1979–2006	0.11	2.4	5.8	6.4	4.6	0.0	2.7	1.1
Jan	1936–2008	0.09	0.9	10.7	31.5	2.8	30.9	−1.6	0.0
	1979–2006	0.11	0.9	3.0	3.3	−2.9	6.2	0.9	0.4
Feb	1936–2008	0.13	1.3	14.6	33.2	4.2	36.3	−0.8	0.0
	1979–2006	0.19	1.8	6.2	5.6	−1.5	1.3	0.0	1.1
Mar	1936–2008	0.15	2.0	14.3	31.8	1.1	9.3	1.6	3.7
	1979–2006	0.18	2.2	6.2	1.8	−1.7	6.8	1.2	1.3

b coefficients of liner trend: *T* °C/10 years; Pr mm/10 years; *H* cm/10 years; *S* mm/10 years; R^2 coefficient of determination (explained variance), expressed as percentages (%)

year was observed during the period 2000–2009. Nevertheless, a weak but statistically significant trend towards shorter SCD is seen.

Snow is the origin of a significant fraction of run-off in the Baltic Sea basin. The water volume held by the snow cover and the spring melt rate are significant factors affecting the volume and timing of the spring floods (BACC Author Team 2008), and changes in SCD can be detected indirectly using hydrological observations. Germany has been divided into three regions, each with a different seasonality of flooding. A slight extension to the region situated in western and central Germany towards the southeast has been detected, indicating a spatial increase in winter flooding due to changes in snow conditions (Beurton and Thieken 2009).

Analysis of 19 river basins in Latvia during 1951–2006 has shown a tendency towards a decrease in spring floods and an increase in winter flow, due to changes in SCD and snow amount (Apsīte et al. 2009). Trends in spring flood volume, peak and timing observed in Lithuanian rivers during 1922–2003 indicate warmer winters and changes in snow cover (Meilutyte-Barauskiene and Kovalenkoviene 2007).

Snow is a significant recreational attraction in the Baltic Sea area. In Estonia, various forms of winter recreation have become increasingly popular, and this has led to an increase in snow observations, made not only by scientists. A combination of questionnaire-based analysis and scientific observations shows that variability in snow conditions has increased recently: with more significant variation experienced in lowlands than uplands and forest edges which retain more stable conditions (Vassiljev et al. 2010).

6.2.2 Snow Depth and Snow Water Equivalent

Large interannual variation is seen in the snow depth and SWE time series of the Baltic Sea drainage basin. Snow amounts do not show any significant trends over the period 1936–2008 in the Russian part of the Baltic Sea drainage basin (30–40°N; 60–65°E) (Fig. 6.1; Table 6.2) (ROSHYDROMET 2008; adapted from Kitaev et al. 2007, 2010). Using interpolation of measurements on a network of snow transects, the changes in SWE during 1966–2005 were studied for the northern part of the East European Plain (Baltic Sea watershed is included) and no significant trends were seen (Kitaev et al. 2007, 2010; Khan et al. 2008; Holko et al. 2009).

Nevertheless, some long-term trends have recently been reported, with some regional variation. According to INTAS-SCCONE, snow depth is still increasing despite the recent rise in global temperature, for the majority of northern Eurasia (Heino et al. 2006; Kitaev et al. 2010). Drozdov et al. (2010) reported a slight positive trend in snow accumulation in European Russia (eastern part of the Baltic Sea drainage basin), and Bulygina et al. (2011) reported an increase in mean winter and maximum snow depth across most of the Russian land area over the past four decades. In most areas, the number of days when snow depth is greater than 20 cm also increased. However, maximum winter snow depth decreased in western European Russia. An increase in SWE was seen in many areas, but decreased in western European Russia (Bulygina et al. 2011).

Table 6.2 Recent trends (1976–2006) in surface air temperature (*T*), precipitation (Pr), snow cover duration (*D*), snow depth (*H*), snow water equivalent (*S*) for the Russian part of the Baltic Sea drainage basin and East European Plain. (ROSHYDROMET 2008; adapted from Kitaev et al. 2007, 2010. See also Chap. 4, Sects. 4.4.2 and 4.5.2)

	T		Pr		*D*		*H*		*S*	
	b	R^2	*b*	R^2	*b*	R^2	*b*	R^2	*b*	R^2
Baltic Sea drainage basin	0.38	2	6.2	1.8	−2.6[a]	24.2	−1.7	6.8	1.2	1.3
East European Plain	0.68	7	0.61	4.2	−4.7	14	0.88	4.9	−1.5	2.7

b coefficients of liner trend: *T* °C/10 years; Pr mm/10 years; *D* days/10 years; *H* cm/10 years; *S* mm/10 years; R^2 coefficient of determination (explained variance), expressed as percentages (%)
[a]0.89 during the period 1936–2008 (R^2 = 12.6 %)

In Sweden (northern part of the Baltic Sea drainage basin), maximum snow depth did not change significantly during 1905–2003 (Larsson 2004). A slight increase was observed in the most southern and northern parts of the country. During the period 1961–2003, mid-Sweden experienced an approximate 30 % decrease in maximum snow depth (Larsson 2004). A snow depth record from the Swedish sub-Arctic (1913–2004; see also Fig. 6.2) shows an increase in winter mean snow depth of 2 cm (5 %) per decade since 1913, and 10 % per decade since the 1930–1940s (Kohler et al. 2006). Relatively shallow snow cover has been seen since the late 1990s, however (Åkerman and Johansson 2008; Callaghan et al. 2010).

There has been a general decrease in snow depth at the majority of monitoring stations since 1914 in Norway. A negative trend is more pronounced for the past few decades. In mountain regions, the variation in snow depth is driven by changes in precipitation and a temperature increase can even increase snow depth (Dyrrdal and Vikhamar-Schuler 2009; Dyrrdal et al. 2013).

Maximum SWEs decreased in southern and western parts of Finland during 1946–2001, but increased in the eastern and northern parts of the country (Venäläinen et al. 2009). Large decadal variability was seen (Venäläinen et al. 2009). This was also the case in southern parts of the Baltic Sea region in Poland; the 1960s were characterised by heavy snow loads, while the first half of the 1970s and the end of 1980s had thin snow cover (Bartoszek 2007). Nevertheless, maximum snow depth was observed to decrease in Poland and Estonia (Bednorz 2007). Maximum snow depth decreased by 3.5 cm in Lithuania during 1961–2010 (Gečaitė and Rimkus 2010).

6.2.3 Snow Cover Extent

The Fourth Assessment report of the Intergovernmental Panel on Climate Change reported that the Northern Hemisphere snow cover area had decreased in most regions, especially during spring and autumn in the period 1966–2005, due to the rise in air temperature. In areas with an increase in SCE, the reason was an increase in solid precipitation (Lemke and Ren 2007).

Brown (2000) reconstructed a long time series (1922–1997) of western Eurasian SCE anomalies for October, March and April (by producing areal snow cover indexes using station data from Canada, the United States, the former Soviet Union, and the People's Republic of China as well as continental SCE from NOAA satellite snow cover data) and reported a small long-term change in autumn SCE, but a rapid reduction in spring SCE, particularly in April. More recently, a fast decrease in spring SCE (1972–2013) has been observed in Europe, especially in Scandinavia, in spite of large decadal fluctuations (van Oldenborgh et al. 2009, see Fig. 6.3). Henderson and Leathers (2010) also reported a decrease in European SCE.

In Fennoscandia, there has been a decreasing trend in SCE especially since the 1970s, but with some regional exceptions (Venäläinen et al. 2009). SCE decreased in the Russian part of the Baltic Sea basin during the 1970–1990s; this decrease has since ceased (Bulygina et al. 2011).

6.2.4 Snow Structure and Properties

In the western half of the Eurasian continent, days with thaw have become more frequent since 1881. For example, in Fennoscandia, in the latter half of the twentieth century, the number of days with winter thaw increased by 6 in 50 years. The duration and maximum thickness of the basal ice layer in the European part of Russia have also decreased since 1966. Changes in open areas are more marked than in forested areas (Bulygina et al. 2010). Reindeer herders in northern Finland report slightly different experiences, however (ACIA 2004). Their experiential knowledge is that ground ice formation at the lichen layer has become more common.

The formation of ice crusts after rain-on-snow events, or surface thawing with subsequent refreezing, has been observed by satellite monitoring (Bartsch et al. 2010).

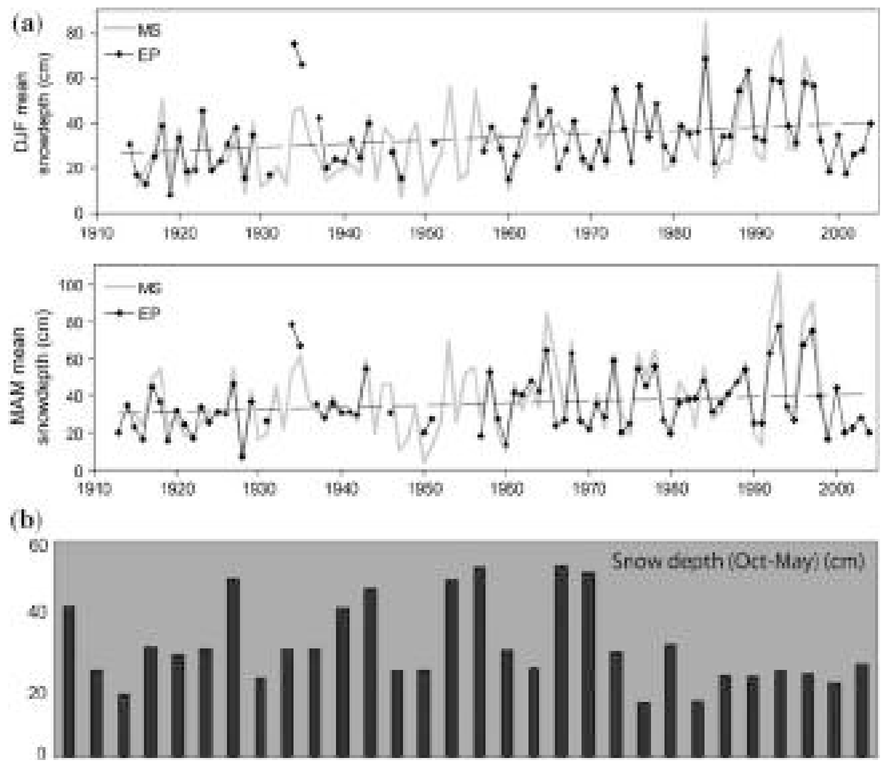

Fig. 6.2 Mean snow depth for Abisko, northern Sweden, during December–February and March–May for 1913–2004 **a** (Kohler et al. 2006) and 1978–2006 **b** (Åkerman and Johansson 2008; updated from Kohler et al. 2006)

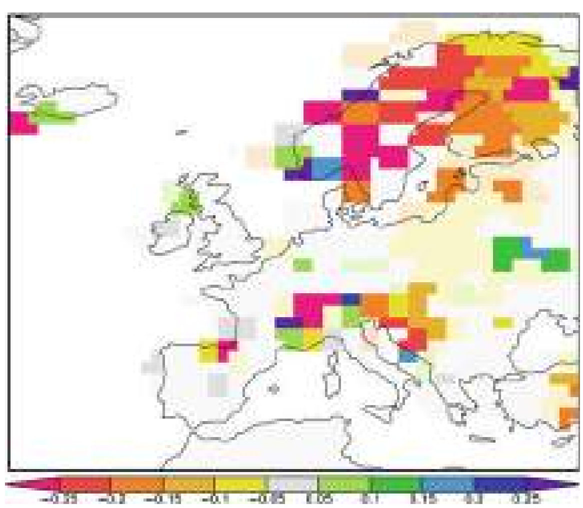

Fig. 6.3 Trend in observed March–May snow cover extent during the period 1972–2013 (°C per decade). Linear regression was calculated between the spring snow cover area and the globally averaged temperature anomalies—negative values mean decreasing snow cover extent with the rise in temperature. Only *grid boxes* with $p < 0.2$ are shown. (updated from van Oldenborgh et al. 2009)

Winter rain-on-snow events are associated with changes in air temperature in northern Eurasia. Such events are therefore sensitive to small changes in winter climate. The occurrence of winter month rain-on-snow events has been observed to increase with increased air temperature; the rate of increase ranged from 0.5 to 2.5 events per winter per °C and was greater at locations with low air temperature (Ye et al. 2008).

6.2.5 Extreme Events

Europe experienced several exceptional winters during the period 2000–2010. Winter 2005/2006 was notable because of the great snow accumulation at the end of the winter season, long SCD and heavy snowfall events in the low-lying areas. Climatologically, this winter was not particularly cold or wet, but it did have exceptionally few thawing episodes (Pinto et al. 2007). Winter 2006/2007 was exceptionally warm and was extremely likely to have been the warmest for more than 500 years (Luterbacher et al. 2007). The winter of 2009/2010 had large snowfall, which was associated with a negative North Atlantic Oscillation (NAO, see Chap. 4) and El Niño event (Seager et al. 2010).

The occurrence of a positive NAO phase has been shown to contribute to rapid snowmelt events in Polish–German lowlands (southern part of the Baltic Sea drainage basin) (Bednorz 2009), and the location of low-pressure systems over Europe has been shown to be responsible for heavy snowfalls in this region (Bednorz 2008). Extreme SCDs and

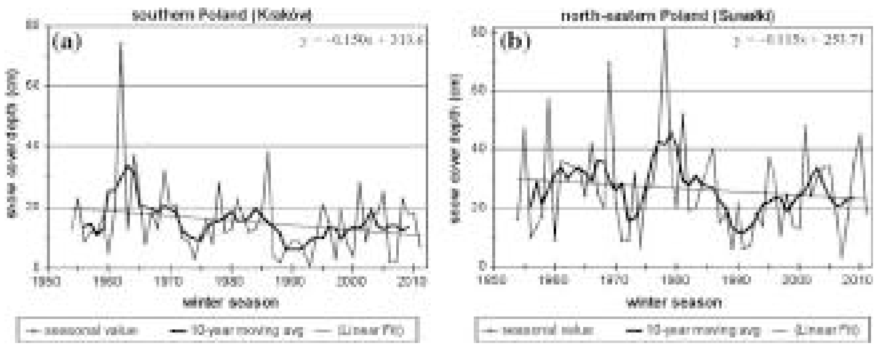

Fig. 6.4 Trend in the 90th percentile of daily snow cover depth for the winter season (1 December to 28 February 1954/1955–2011/2012). **a** Southern Poland. **b** North-eastern Poland (updated from Falarz 2008)

maximum seasonal snow depth values in Poland during the latter half of the twentieth century were analysed by Falarz (2008) (Fig. 6.4). Abundant snow cover has become rarer, and since the 1970s, a sparse snow cover has been observed more frequently than before. The changes are not statistically significant, however. Łupikasza et al. (2009) found no significant trends in extreme snow cover in Poland during the latter part of the twentieth century, but reported that since winter 1987/1988 the area of extremely thin snow cover has remained relatively large.

Extreme snow conditions are connected to, for example, snow-induced forest damage. In Finland, this damage is assumed when snow accumulation exceeds 20 kg m^{-2} during a 3-hour period or precipitation exceeds 20 mm during a 5-day period. According these criteria, snow-induced forest damage was expected in Finland on average 65 times a year during the period 1961–2000, but as often as 150 times a year during the mild 1990s. During 1961–2000, the maximum number of heavy snow-load events occurred in 1994 in northern Finland (Gregow et al. 2008; Kilpeläinen et al. 2010).

6.3 Recent and Present Change in Glacier Extent and Mass Balance

Perennial snow and ice extent in Scandinavia during 2000–2008 was studied by Fontana et al. (2010) using a Moderate Resolution Imaging Spectroradiometer (MODIS) data set with a 250 m spatial resolution. Large interannual variation was seen, and a strong negative relationship was found between snow and ice extent and positive degree-days during summer months. Snow and ice extent was significantly correlated with annual net glacier mass balances.

The mass balance record for Storglaciären in the Kebnekaise massif of northern Sweden is the longest continuous glacier mass balance record in the world. The Storglaciären record shows a fluctuating pattern in net mass balance; since 1992, the net mass balance trend has been largely negative (Jansson and Pettersson 2007; Fig. 6.5). In a study by Evans et al. (2008), the net mass balance of Storglaciären was shown to be related to the changing snowpack volume and the resulting winter mass balance during the years 1990–2006. A negative trend in the winter mass balance combined with the increasing trend in mass lost due to ablation has resulted in a decrease in glacier net mass balance and a rise in the snowline.

Other glaciers in the Tarfala Research Station mass balance programme exhibit similar trends. In inland Scandinavia, a cumulative loss in glacier ice thickness has been reported by the World Glacier Monitoring Service for 1967–2008 (WGMS 2008; Voigt et al. 2010). Recent thinning of 1 meter year^{-1} has been observed at the equilibrium line of a Norwegian ice cap, partly draining into the Baltic Sea drainage basin (Brown 2012).

The frontal positions of glaciers measured in Tarfala show retreat rates of −1 to −14 meters year^{-1} between 1915 and 1994 (Holmlund et al. 1996). More recently, the status of Swedish glaciers has been monitored using remote sensing and classification of satellite images for areal estimation of the glaciated areas (e.g., Klingbjer et al. 2005). In 1973, Østrem et al. (1973) compiled a glacier atlas over northern Scandinavia using aerial photographs and map data. In this report, the glaciers covered 321.8 km^2. In 2001, a new inventory was conducted using high-resolution satellite imagery as part of the Global Land Ice Measurements from Space (GLIMS) project (Armstrong et al. 2011) and the Swedish glaciers were reported to cover 264.5 km^2. However, the error margins associated with the satellite data limits the accuracy of these results. Nevertheless, the glacier data show a decrease in glacier coverage in Sweden with many glaciers retreating into protected niches. Fealy and Sweeney (2005) attributed the behaviour of Scandinavian glaciers since the 1970s to large-scale changes in atmospheric circulation. The strong correlation between the Arctic Oscillation(AO, see Chap. 4) and the winter mass balance of Swedish glaciers (Jansson and Linderholm 2005) suggests that Swedish glaciers are particularly sensitive to change in winter surface temperature. Regional downscaling of general circulation models(GCMs) using ERA-40 reanalysis data has suggested that Storglaciären exhibits a mass balance sensitivity to temperature change of −0.48 meters year^{-1} per °C (Radíc and Hock 2006).

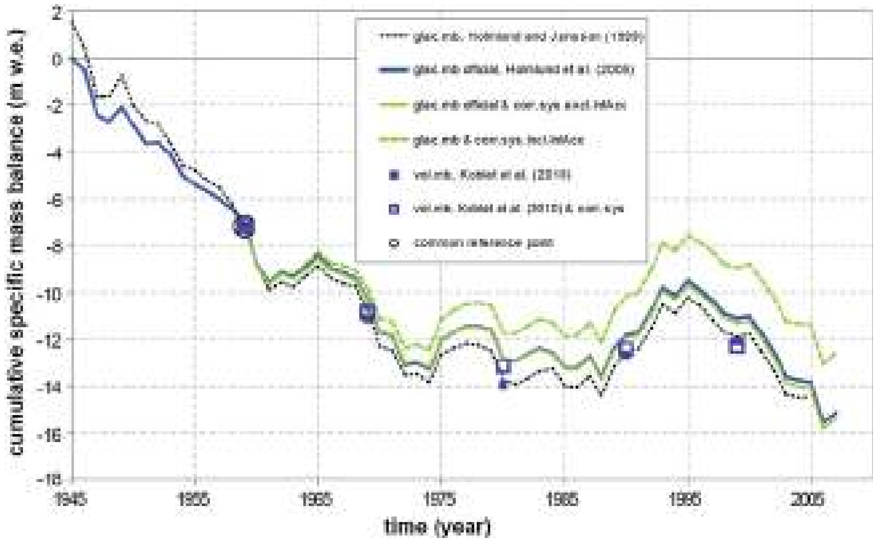

Fig. 6.5 Cumulative glaciological and volumetric mass balance series of Storglaciären, northern Sweden. *Different lines* represent different methods and corrections that have been used to estimate the net mass balance (Zemp et al. 2010)

6.4 Recent and Present Change in Frozen Ground

6.4.1 Seasonally Frozen Ground

In the northern part of the Baltic Sea basin in Fennoscandia, surface soil freezes and thaws during winter with large variability in frozen ground depth. Mellberg (2008) described the recent (1991–2007) seasonally frozen ground characteristics observed in Sweden. Maximum and minimum temperatures per winter season, the number of freeze days and the temperature trend per winter were studied at several sites around the country. A small warming trend during winter was observed in ground temperature at 10 cm depth.

Recent warming trends at Abisko (Sweden) are known to be consistent with those observed for the Scandinavian sub-Arctic and the rest of the Sweden. Schmidt (2011) studied the annual and seasonal warming trends at soil depths of 20–100 cm in Abisko (1985–2010) and found a decrease in the length of time that the ground remained frozen seasonally, with later freeze-up and earlier spring thaw. In contrast, short-term freeze/thaw cycles of the ground in the upper 20 cm appeared to be more frequent. Despite earlier studies indicating snow cover as the most important parameter influencing ground temperature, Johansson et al. (2008) and Schmidt (2011) failed to find such a correlation. In contrast, mean monthly air temperature is highly correlated with ground temperature in all seasons to 100 cm depth. The increase in regional mean annual air temperature over the period 1979–2002 (see Chap. 4) is positively correlated with slow soil surface movements due to freezing and thawing, but this is subject to large local and regional variability (Ridefelt et al. 2009).

Over the final decades of the twentieth century, the length of time over which the ground remained frozen reduced by two weeks in Lithuania (south-eastern part of the Baltic Sea drainage basin). In the period 1960–1979, the ground remained frozen throughout the entire winter season and the probability of thaw/freeze was only 35 %. In 1980–2000, the probability that thaw/freeze would occur during the winter was 100 %. In some regions, seven thaw/freeze events were recorded in one season (Taminskas et al. 2005). The depth of seasonally frozen ground has decreased in Lithuania since 1923. The greatest reduction occurred at the end of the twentieth and the start of the twenty-first centuries (Taminskas et al. 2006).

It has been suggested that atmospheric warming can lead to more frequent and stronger freeze/thaw events due to reduced snow insulation (Isard and Schaetzl 1998). In the southern Baltic Sea region, the increase in freeze/thaw event frequency was not observed with reduced snow cover. Kreyling and Henry (2011) reported a 50-year trend analysis of snow cover and frozen ground characteristics at 177 stations in Germany. SCD decreased by 0.5 d $year^{-1}$ with an increase in minimum soil temperature and a uniform decrease in freeze/thaw cycles at 5 cm depth. Henry (2008) suggested that changes in air temperature may have a greater impact on frozen ground characteristics than precipitation, but this has not yet been investigated for the Baltic Sea area.

6.4.2 Permafrost

Data collected from permafrost boreholes over the past decade indicate recent warming in the European permafrost, with the greatest warming at higher latitudes. Shorter-term extreme climatic events are also reflected in changes in

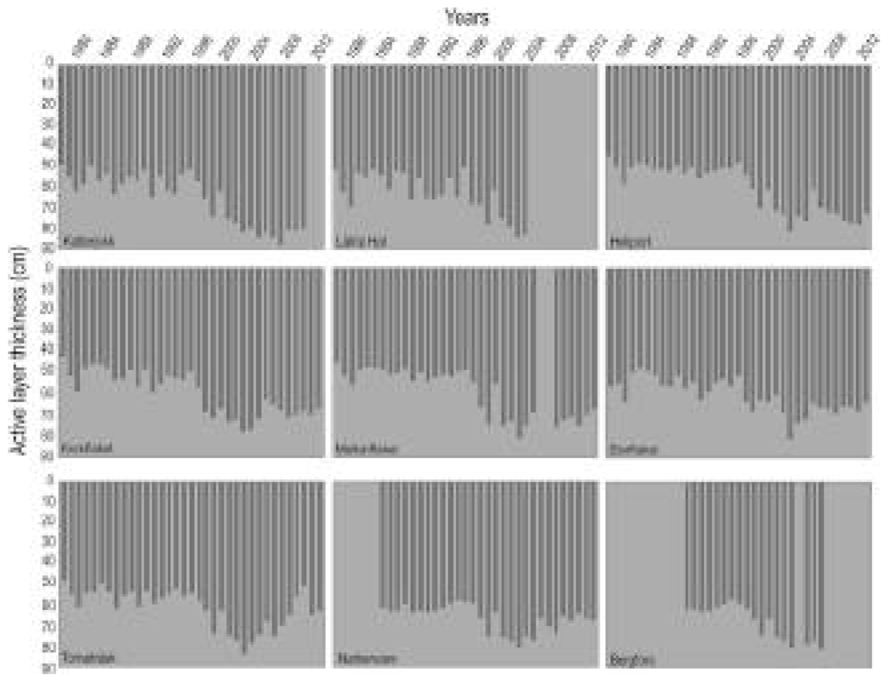

Fig. 6.6 Active-layer thickness from 1978 to 2012 at nine sites in sub-Arctic Sweden. The active layer has become deeper over the monitoring period, especially over the past decade (updated from Åkerman and Johansson 2008)

active layer thickness (Harris et al. 2009). A rise in ground temperature of 0.1–0.7 °C at the depth of zero annual amplitude in European Russia was observed during the monitoring period. The southern limit of patchy near-surface permafrost retreated northward by 20–50 km in European Russia between 1974 and 2008 (Drozdov et al. 2010).

Thawing permafrost and thicker active layers are also reported for sub-Arctic Sweden over the period 1978–2012 (Åkerman and Johansson 2008; Callaghan et al. 2010; Fig. 6.6). Permafrost degradation is correlated with increases in air temperature and is sensitive to changes in snow depth. The relationship between snow and permafrost is not straightforward, however, and snow structure also has an effect (Johansson 2009). New borehole data in the lowland peat mires of the Abisko area show ground temperature increased by 0.4–1 °C between 1980 and 2002 with mean annual ground temperatures close to 0 °C. Thus, permafrost in this region appears very vulnerable to the projected climate warming (Johansson et al. 2011).

6.5 Conclusion

There has been a 0.1 °C per decade increase in winter temperature for the period 1871–2011 in the northern part of the Baltic Sea area (Chap. 4, Sect. 4.4). In relation to this warming, the recent literature reinforces the findings of the first Baltic Sea assessment on changes observed in the terrestrial cryosphere.

SCE has shown mostly decreasing trends within the Baltic Sea basin. SCD has also decreased in several regions, especially owing to earlier snowmelt. Large interannual variation is seen in snow depth and SWE time series for the area. Nevertheless, a decreasing trend is seen in snow depth and SWE in several regions, especially in lowlands and coastal regions, where snow cover variability is dominated by air temperature (winter temperatures relatively close to 0 °C). In northern and eastern parts of the drainage basin, and in mountain regions where both precipitation and temperature control snow amount, colder average winters have led to an increase in annual snow depth and SWE. Few data are available on changes in snow structural properties. There is no evidence for a recent change in the frequency or severity of snow-related extreme events. The past decade saw an exceptionally warm winter in 2006/2007 and two winters with high snow accumulation (2005/2006 and 2009/2010).

A decrease in glacier coverage in Sweden has been observed, although variability is seen in the long-term mass balance record of an actively monitored glacier. A cumulative decline in glacier ice thickness has been reported by the World Glacier Monitoring Service in inland Scandinavia.

Current understanding of the spatial patterns of frequency–intensity–duration characteristics of freeze/thaw cycles of the ground in the Baltic Sea region remains poor and has not been subject to systematic study. Some warming trends, and some decreases in the duration and depth of seasonally frozen ground have been seen. Warming trends have been observed in the European permafrost, as well as a northward shift in the southern boundary of near-surface permafrost in European Russia.

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Recent Change—Marine Circulation and Stratification

7

Jüri Elken, Andreas Lehmann, and Kai Myrberg

Abstract

This chapter describes recent change in the circulation and stratification of the Baltic Sea. A recent warming trend in sea-surface waters has been clearly demonstrated by in situ measurements, remote sensing data and numerical models. Trends in sea-surface temperature (SST) for the past three to four decades based on remote sensing data generally agree with trends determined from in situ observations. Models suggest the current warming within the Baltic Sea lies within the range experienced during the past 500 years. The salinity and stratification of the deep waters are strongly linked to the major inflows of North Sea water that occur sporadically and bring high-saline water into the deep layers of the Baltic Sea. The major inflows normally occur during winter and spring and bring cold oxygen-rich waters into the deep basins. Since 1996, large inflows have also occurred during summer, bringing in warm low-oxygen water.

7.1 Introduction

Changes in thermohaline characteristics and stratification of the water column are usually analysed on the basis of data from repeat hydrographic observations, originating from shipborne monitoring programmes and/or permanent coastal and offshore oceanographic stations. Such datasets are described in recent books by Feistel et al. (2008) and Leppäranta and Myrberg (2009). In situ observations are often irregular in space and time, especially for offshore areas; therefore, trends and variability in the ocean state can only be determined by careful pre-processing of the data to address issues such as data homogenisation and the suppression of aliasing errors. In recent decades, horizontal undersampling by in situ measurements has been partially offset by the remote sensing of sea-surface properties from satellites; with these data free of spatial aliasing errors due to their high horizontal resolution. Satellite-derived cloud-free products for sea-surface temperature (SST) have been reasonably accurate since the 1980s. Many of the papers published over the past decade that address trends and variability in the circulation and stratification of the Baltic Sea are based on remote sensing data, often in combination with in situ observations and the results of numerical modelling. Improvements in the physical and numerical features of the models, accompanied by model validation studies, have increased confidence in the realism of the model results. Models provide dynamically balanced gridded datasets over decades or longer, including information for 'noisy' variables such as currents, transport and mixing fluxes. This chapter analyses the recent peer-reviewed literature, extending the findings from earlier review (Box 7.1).

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_7

Box 7.1 Key findings from the first Baltic Sea assessment (BACC Author Team 2008)

Temperature

- The late 1990s were characterised by warmer summers and colder winters compared to the years before and after this period.
- Heat content maxima were identified around 1975 and 1990. A heat content minimum was found in 1987, when the winter heat content and summer heat content both showed minimum values.
- There were indications of seawater warming, but it was unclear whether this was partly or even entirely due to a change in sampling frequency and/or to changing seasonal representation in the dataset.
- Since 1988, the seasonal temperature minimum of the intermediate winter water has increased by 1.5 °C.

Salinity and saltwater inflows

- The mean salinity of the Baltic Sea decreased during the early twentieth century and during the 1980s and 1990s. No long-term trend was found during the twentieth century.
- Since the mid-1970s, the frequency and intensity of major saltwater inflows from the North Sea have decreased.
- Major inflows were absent between February 1983 and January 1993. During this low-salinity phase, the deep water of the eastern Gotland Basin was poorly ventilated, with oxygen depletion as a consequence.
- Two low-salinity phases were found: one during the 1920s/1930s and one during the 1980s/1990s. They are explained by stronger than normal freshwater inflow and zonal wind velocity.

This chapter refers to the Baltic Sea gulfs and basins as shown in Fig. 7.1. For detailed descriptions of Baltic Sea geography and topography, see the books by Feistel et al. (2008) and Leppäranta and Myrberg (2009).

7.2 Trends and Variations in Water Temperature

The Baltic Sea is well stratified, with a seasonal cycle of temperature superimposed on the more or less permanent two-layer salinity stratification. While temperature and sea ice respond rapidly to the changes in atmospheric heat fluxes, variations in salinity are governed mainly by lateral transport processes, resulting together with diapycnal mixing (i.e. mixing across the surfaces of constant water density) in response times of many decades (e.g. Stigebrandt and Gustafsson 2003; Omstedt and Hansson 2006).

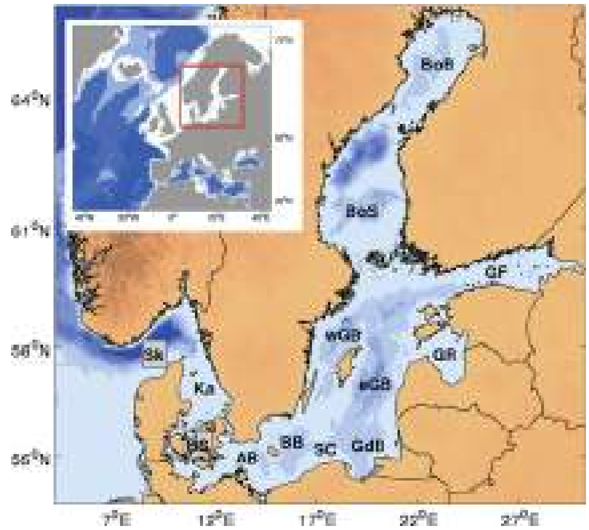

Fig. 7.1 Baltic Sea showing topography (greater depths presented by *darker blue colour*) and sub-regional notations: *Sk* Skagerrak, *Ka* Kattegat, *BS* Belt Sea, *AB* Arkona Basin, *BB* Bornholm Basin, *SC* (underwater) Stolpe Channel, *GdB* Gdansk Basin, *eGB* eastern Gotland Basin, *wGB* western Gotland Basin, *GR* Gulf of Riga, *GF* Gulf of Finland, *BoS* Bothnian Sea, *BoB* Bothnian Bay, Baltic Proper includes AB, BB, SC, GdB, eGB and wGB, Gulf of Bothnia includes BoS and BoB

Cold waters, formed during winter, extend down to the halocline which has a typical depth of 60–80 m in the Baltic Proper and somewhat less in the southern basins. During summer, when a seasonal thermocline develops in the surface waters at depths of about 15–20 m, the underlying cold intermediate layer generally retains a 'memory' of the severity of the previous winter. This dicothermal layer is often referred to as 'old winter water'. It is no surprise, therefore, that Hinrichsen et al. (2007) found the summer (July–August) temperature of the intermediate cold layer to be well correlated with the surface (down to the halocline) temperature of the preceding March. Deeper waters, below the halocline, are formed mainly by lateral advection of saline waters of North Sea origin that entrain and mix with ambient waters during their passage into and through the Baltic Sea. Below 100 m depth, the range in temperature variation within the Gotland Deep is only 5 °C (range 3–8 °C), compared to a range in surface temperature of up to 25 °C. According to the Baltic Sea hypsographic curves (the relationship between the surface area bounded by given depth contours and their depths), the volume of deep layers below 100 m represents only about 12 % (Leppäranta and Myrberg 2009) of the whole sea volume (21,205 km^3) and the contribution of lateral heat advection to the overall Baltic Sea heat content is relatively small.

The rise in air temperature in the Baltic Sea region (see Chap. 4, Sect. 4.4) is expected to drive a corresponding rise in water temperature. Indeed, MacKenzie and Schiedek (2007a) stated that since the 1860s a record warming of the

Baltic Sea and North Sea has occurred, especially during recent decades. They used data from daily monitoring at a few long-term stations, such as Christiansø, located close to the island of Bornholm, as well as data from irregular open water sampling, and applied advanced data homogeneity and spatial synchrony matching procedures to ensure sufficient data quality for climate analysis (MacKenzie and Schiedek 2007b). Their results showed little evidence of a gradual linear increase or decrease in SST since the mid-late 1800s, however; in contrast, modelling results by Gustafsson et al. (2012) suggest a rise in SST temperature of 0.8 °C over the past 150 years. There have been earlier warm periods in the mid-late 1800s and in the mid-1900s, especially in the 1930s. However, since about 1985, a warming of surface waters is evident in all datasets and in all seasons. The probability of extremely warm surface waters in winter and summer has increased since the 1990s by two- to fourfold. MacKenzie and Schiedek (2007a) argued that summer warming rates have almost tripled compared to those that could be expected from the observed increase in air temperature. In contrast to this and to the results of many other studies, Håkanson and Lindgren (2008) concluded, from a simple treatment of raw irregular HELCOM data 1974–2005, that '*there is no increase in surface-water temperatures in the Baltic Proper, but rather a weak opposite trend*'. It could be that this dataset contains more observations from the cold season over recent decades and so the rising trend in SST is not visible. Careful statistical pre-processing of irregular sampling data is vital when attempting to estimate multi-decadal change for a variable with high seasonal amplitude. Madsen and Højerslev (2009) have shown, on the basis of daily routine lightship observations in Danish waters during 1900–1998, that in Drodgen Sill, the mean SST by the end of the period was 0.7 °C higher than any previous observations. Since the 1990s, surface layers are also observed to have warmed near the Lithuanian coast (Dailidienė et al. 2011), recently at 0.3–0.9 °C per decade. The changes are more complex in the Gulf of Finland (Liblik and Lips 2011) and in the Gulf of Riga (Kotta et al. 2009); these studies stress the relation of temperature variations to the atmospheric circulation patterns expressed as the North Atlantic Oscillation (NAO, see Chap. 4, Box 4.1) and/or the Baltic Sea Index (BSI, difference of sea-level air pressure anomalies between Oslo and Szczecin, Lehmann et al. 2002) indices (e.g. Lehmann et al. 2011; Dippner et al. 2012). In the northern sea areas, the recent decrease in the extent and duration of ice cover (see Chap. 8) also has a strong influence on the trends in seawater temperature (see Sect. 7.5).

Variations in SST can be well resolved by remote sensing from satellites, using infrared AVHRR (Advanced Very High Resolution Radiometer) and MODIS (Moderate Resolution Imaging Spectroradiometer) sensors. Regular satellite coverage has been available in the Baltic Sea region since the mid-1980s. Although single remote sensing images are frequently disturbed by cloud coverage, skin layer uncertainties and other factors, the monthly mean SST from the remote sensing data agree well with those from in situ measurements in the offshore sea areas (Siegel et al. 2006; Bradtke et al. 2010). Lehmann et al. (2011) used remote sensing data for 1990–2008 to derive a linear trend of annual mean SST of up to 1 °C per decade in the northern part of the Bothnian Bay, but a high increase was also found in the Gulf of Finland and Gulf of Riga and in the northern Baltic Proper (Fig. 7.2). Warming of surface waters is lowest (0.3–0.5 °C per decade), north-east from Bornholm Island up to and along the Swedish coast, probably due to an increase in the frequency of coastal upwelling (Lehmann et al. 2012). Bradtke et al. (2010) also considered trends in monthly and seasonal mean values during 1986–2006 and found the highest positive trend (more than 2 °C per decade in August) in the Bothnian Sea and the northern Baltic Proper. At the same time, mean SST in March decreased. Siegel et al. (2006) studied the period 1990–2004 and found the highest rate of increase in the Bothnian Sea in July (more than 3 °C per decade), and in the Arkona and Gotland Sea in August and September (about 1.5 °C per decade). Trends in SST for the past three to four decades based on data from remote sensing generally agree well with trends determined from independent in situ observations of SST.

Deep-water temperature in the Baltic Proper is determined mainly by the lateral spread of submerged saline water of North Sea origin, reflecting surface thermal conditions during deep-water formation. Mohrholz et al. (2006) found that in Bornholm Basin, the mean temperature in the halocline increased during the period 1989–2004 by about

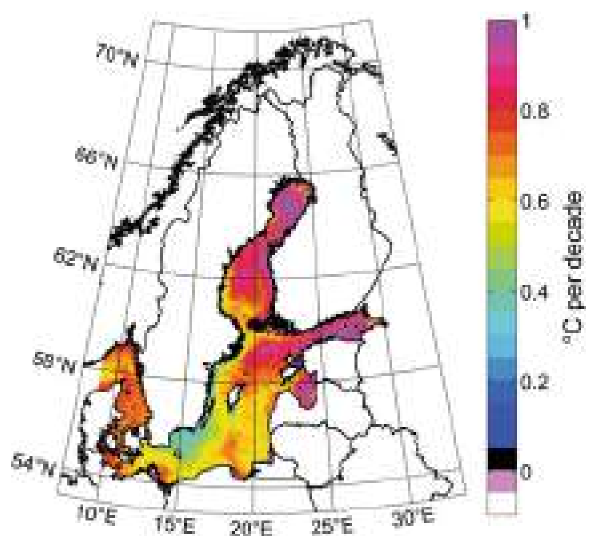

Fig. 7.2 Linear trend in annual mean sea surface temperature based on infrared satellite data (1990–2008) provided by the Federal Maritime and Hydrographic Agency (BSH), Hamburg (Lehmann et al. 2011)

1 °C compared to the longer period 1950–2004. They argued that this halocline temperature increase was caused by more frequent warm summer inflows since the 1990s.

An increase was also found in the annual minimum temperature of the intermediate layer, lying between the seasonal thermocline and the halocline. In the Bornholm Basin, Mohrholz et al. (2006) found a positive correlation with the NAO winter index for the period 1952–2004 (R^2 = 0.61, January and February). In light of the decadal change in the NAO index since the 1990s, when intensified cyclonic circulation and stronger westerly winds became evident in the Baltic Sea region relative to the 1970s and 1980s (see Chap. 4, Box 4.1), the intermediate layer temperature variations were interpreted as a 'regime shift' increase of about 1 °C (Mohrholz et al. 2006; Hinrichsen et al. 2007). In the Gulf of Finland, Liblik and Lips (2011) found a positive correlation (R^2 = 0.81) between the intermediate layer temperature and the winter BSI for the period 1987–2008.

It is an intriguing task to reconstruct past changes in climate elements and compare them with ongoing change. Based on a climate reconstruction since 1500 (Luterbacher et al. 2004), Hansson and Omstedt (2008) reconstructed the Baltic Sea water temperature and ice conditions for the period 1500–2001 using the PROBE-Baltic model. For the period since 1893, they used more detailed forcing data from the NORDKLIM database (see Hansson and Omstedt 2008 for details). Annual mean water temperature anomalies (Fig. 7.3), averaged over the whole sea domain (both by area and depth), reveal cold anomalies of about −0.7 °C in the decadal moving average during the 1690s and 1780s, and warm anomalies of up to 0.5 °C in the 1730s, 1930s and 1990s. The seawater warming during the present period is comparable in magnitude to that of the 1930s and the first half of the eighteenth century. The results of the modelling study suggest that the current warming within the Baltic Sea lies within the range experienced during the past 500 years. Nevertheless, the twentieth century is clearly the warmest, with the exception of the warm anomaly around the 1730s.

7.3 Changes in Salinity, Stratification and Water Exchange

While the thermal response of the Baltic Sea to the change in air temperature is similar to that of a large lake, freshwater discharge from land and restricted water exchange with the North Sea create strong salinity stratification, accompanied by along-basin gradients such as are seen in estuaries and fjords. The overall salt content of the Baltic Sea depends to a large extent on net precipitation and river discharge; with higher salinity during dry periods and lower salinity during wet periods. The salinity level is also governed by variability in the water exchange between the North Sea and Baltic Sea (Winsor et al. 2001, 2003; Meier and Kauker 2003; Gustafsson and Omstedt 2009). Termed Major Baltic Inflows (MBIs, Matthäus and Frank 1992) these events occur sporadically and bring in large volumes of highly saline water causing a sudden increase in bottom salinity.

The Gotland Deep (the deepest part of the eastern Gotland Basin, Fig. 7.1) is a representative location for describing salinity and stratification development within the Baltic Sea as a whole. Indeed, changes in mean salinity, calculated from Gotland Deep data only, are only 2 % different to changes calculated based on data from all sub-basins (Winsor et al. 2001). Observations (Fig. 7.4) reveal a low-salinity period above the halocline starting in the 1980s. Fresher periods also occurred in the 1900s and 1930s and to a less extent in the 1960s.

Salinity and stratification of the deep layers are highly affected by the occurrence of MBIs of North Sea water. These occur when high pressure over the Baltic Sea region with easterly winds is followed by several weeks of strong

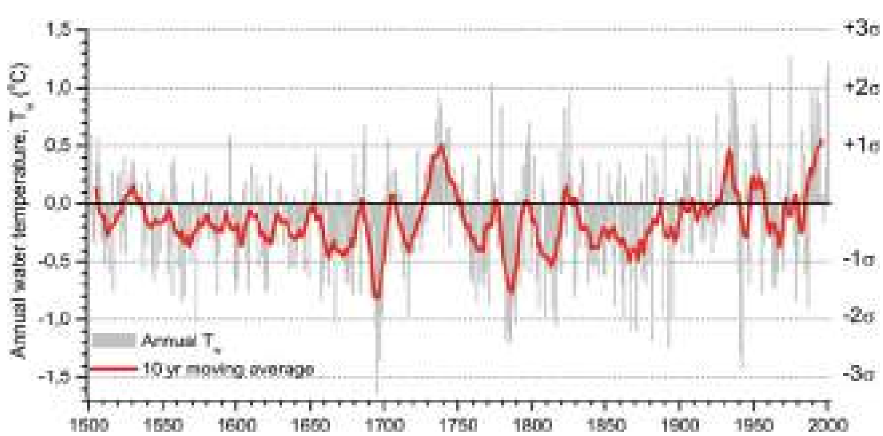

Fig. 7.3 Anomalies of the annual and decadal moving average of the modelled Baltic Sea spatial mean water temperature over the period 1500–2001. The *dotted horizontal lines* are the standard deviations of water temperature during the reference period 1900–1999 (Hansson and Omstedt 2008)

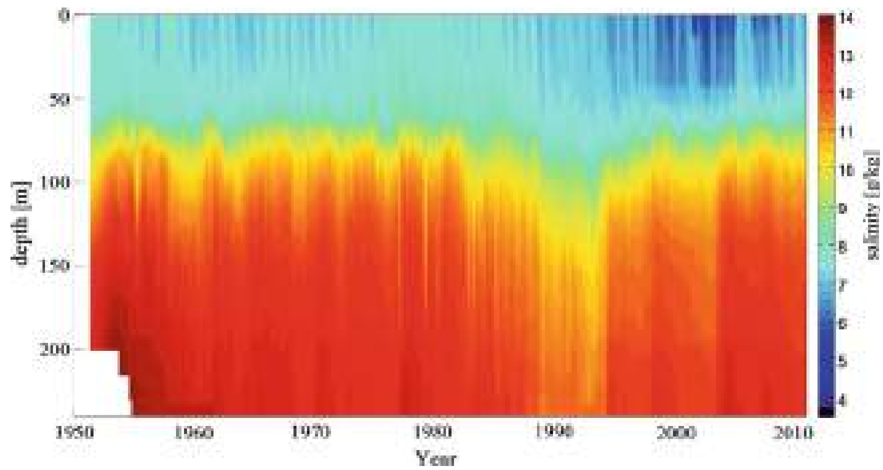

Fig. 7.4 Observed salinity in the Gotland Deep around 57°19.20′ N, 20°03.00′E as a function of depth and time. Data from the ICES Marine Data Centre (www.ices.dk)

westerly winds (e.g. Lehmann et al. 2002; Matthäus et al. 2008; Leppäranta and Myrberg. 2009). During the history of observations since the 1900s, the strongest inflow took place in November–December 1951 (e.g. Madsen and Højerslev 2009; see the high bottom salinities in Fig. 7.4). During the peak inflow, the difference in sea level between Gedser and Hornbaek (i.e. between the northern and southern ends of the Danish straits) was up to 1.5 m and the normal saline stratification in the Kattegat and the strait area broke down for several weeks. A MBI of comparable magnitude occurred in December 2014. On average, new high-saline water reaches the deep layers of the Gotland Basin with a delay of up to a year (e.g. Kõuts and Omstedt 1993; Matthäus et al. 2008), as can be seen also in Fig. 7.4. The inflow during winter 1976/1977 was followed by an exceptionally long stagnation period, when the strength of the saline stratification (bottom to surface salinity difference) decreased by about one and a half times, before the next inflow in 1993. In some areas, such as the Gulf of Finland, the halocline effectively disappeared. An extensive stagnation period also occurred in the 1920s and 1930s, after the very strong inflow in winter 1921/1922, coinciding with the shift from a wet period to a dry period over the Baltic Sea basin. Based on water age calculation, Meier (2005) identified a stagnation period of more than eight years also in the 1950s/1960s.

Since 1994, when stratification strength returned to the near-normal levels of the 1960s and 1970s, stagnation in terms of oxygen deficiency of the near-bottom waters continued (Conley et al. 2009, see also Chap. 18). In addition to smaller inflows, a series of larger inflows has also occurred since then. In contrast to the usual barotropic inflows (vertically uniform transport over the entrance sills) that occur in winter and spring and advect relatively cold water with high oxygen content into the Baltic Sea, the recent large inflows in the summers of 1997, 2002, 2003 (Feistel et al. 2006) and 2006 were of a two-layer (baroclinic) type that transported high-saline, but warm and low-oxygen water to the deep layers of the Baltic Sea. Thus, warm water inflows, whether baroclinic or barotropic, transport less oxygen to the Baltic Sea than cold water inflows, and higher temperatures increase the rate of oxygen consumption (through organic matter mineralisation) in the deep water and increase production of hydrogen sulphide (Matthäus 2006). Inflow activity is very clear in the daily temperature records for the deep layers in the Gotland Deep (Fig. 7.5). The low temperatures apparent during 2003 reflect the normal barotropic inflow in winter 2002/2003, described in many papers (e.g. Matthäus et al. 2008; Leppäranta and Myrberg 2009). Changing stratification strength also has a feedback to mixing processes. For example, Osiński et al. (2010) found that the major inflow in winter 2002/2003 increased the value of the first baroclinic Rossby radius of deformation (which determines the size of mesoscale eddies) in the southern Baltic Sea from about 4 km (in the pre-inflow period) to more than 9 km.

At the sub-regional scale, many aspects of the change in salinity and stratification are important in the context of ecological status and environmental and climatic impacts. When saline waters enter the Baltic Sea, the halocline is lifted up and this signal is dynamically transferred to the downstream basins (Meier 2007). Upstream from the Gotland Deep, in the south-western Baltic Sea, the variations in deep-water properties are generally of higher amplitude; downstream along pathways of deep-water advection they are damped due to a wide range of mixing processes (e.g. Reissmann et al. 2009). In the Bornholm Basin, deep temperature observations reveal waters of warm and cold inflows (Mohrholz et al. 2006) that can be later traced in the

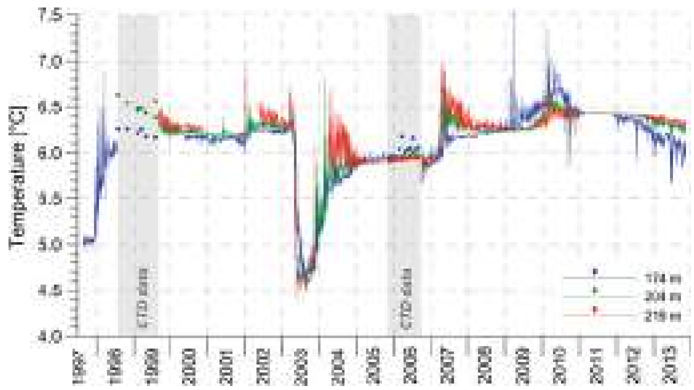

Fig. 7.5 Temperature time series from August 1997 to October 2013 recorded at the eastern Gotland Basin mooring (57°23′N, 20°20′E) near Gotland Deep at 174, 204 and 219 m depth. This 'Hagen curve' is redrawn from Feistel et al. (2008) by the authors using recent data

Gotland Deep. Bottom-salinity anomalies in the Bornholm Basin during 1961–2000 (Neumann and Schernewski 2008) range from −1.8 g kg^{-1} (1982) to 2.0 g kg^{-1} (1994), with no significant trend although a recent slight salinity increase could be seen. In the Lithuanian part of the Baltic Proper deep-water area, Dailidienė and Davuliene (2008) reported a strengthening of stratification for the period 1984–2005: decreased surface salinity and increased deep-water salinity.

In the Gulf of Finland, a sub-region with a free connection to the Baltic Proper and the highest freshwater discharge per unit sea volume, changes in salinity and stratification generally follow those of the Baltic Proper, but are not fully synchronous (e.g. Zorita and Laine 2000). On the basis of monitoring data for the period 1965–2000, Laine et al. (2007) found a continuous decrease in salinity and density stratification until the early 1990s, after which there was a slight increase. Based on independent data for 1987–2008, Liblik and Lips (2011) found that the deep salinities in summer increased after the 1993 major inflow by about 2 g kg^{-1}. Despite the increased mean stratification strength and more frequent occurrence of hypoxia events (see Chap. 18), at the annual scale ventilation of deep waters is still effective and the annual mean oxygen concentrations remain higher than during the 1960s and 1970s (Laine et al. 2007). This can be explained by decreased sea-ice cover (see Chap. 8), which favours wind mixing (Vermaat and Bouwer 2009) and by stronger south-westerly winds in winter that cause stratification collapse events, due to wind straining effects on estuarine gradients (Elken et al. 2014).

Reconstructing annual mean salinities since 1500 (Hansson and Gustafsson 2011) indicates that salinity has slowly increased by 0.5 g kg^{-1} since 1500, peaking in the mid-eighteenth century. Present salinity values are nearly as high as reconstructed for the earlier maximum salinity period. Historically, there have been several fresher periods when the mean salinity of the Baltic Sea decreased from a maximum of about 7.8 g kg^{-1} to about 6.5 g kg^{-1}. Hansson and Gustafsson (2011) also found a negative correlation between oxygen content and salinity, indicating that the major, upper, part of the water column was more efficiently ventilated when the Baltic Sea was in a fresher state.

7.4 Circulation and Transport Patterns and Processes

There are four mechanisms for inducing currents in the Baltic Sea: wind stress at the sea surface, a surface pressure gradient, a thermohaline horizontal density gradient and tidal forces. The currents are also steered by Coriolis-acceleration, topography and friction. Voluminous river inlets can produce local changes in sea-level height and thus also in currents. Due to the small size of the Baltic Sea basins, friction caused by the bottom and shores damp the currents considerably. The general circulation is typical of a stratified system with a positive freshwater balance. Inflowing waters settle in the receiving basin at the depth where the ambient water is of equal density; fresher water moves to the upper layer and saltier and denser water masses move to lower layers.

Over short timescales (1–10 days), the currents are mostly caused by wind stress and gradients in sea level, the latter particularly in straits. Due to the large variability in the winds over the Baltic Sea basin, the long-term wind-driven mean circulation is weak: transient currents are an order of magnitude greater than average currents. In coastal areas, drift currents generate upwelling and downwelling features that are affected by Kelvin-type waves. The Baltic Sea is laterally mixed by mesoscale eddies and deep-water circulation (e.g. Elken and Matthäus 2008). At the timescale of an hour to a day, there are several periodic dynamical processes in operation. The most important are inertial oscillations (13.2–14.5 h) and seiches (less than 40 h). For details, see Leppäranta and Myrberg (2009).

Thus, the long-term mean surface circulation observed in the Baltic Sea is principally due to the nonlinear combination of the wind-independent baroclinic mean circulation and the mean wind-driven circulation. Which of the two is more important is difficult to say in the case of a nonlinear system. The answer will depend on the particular location and on the timescale under study.

7.4.1 Surface Circulation and Related Processes—Recent Findings

Mean circulation within the Baltic Sea as a whole was recently modelled by Meier (2007). The results (Fig. 7.6) agree with the main characteristics of the early findings by Palmén (1930) and the outcome of earlier numerical modelling (Lehmann and Hinrichsen 2000; Lehmann et al. 2002) but also provide new fine-scale details. Mean transport above and below the halocline is in agreement with observational results and implies the existence of strong high-persistency cyclonic gyres both within the Baltic Proper and in the Bothnian Sea. In the eastern Gotland Basin, the model results reveal high transport around the Gotland Deep, especially below the halocline, reproducing the observed deep rim current (Hagen and Feistel 2007). Furthermore, modelling reveals that the strength and persistency of currents are lower in the Gulf of Riga, Gulf of Finland and Bothnian Bay in comparison with the Baltic Proper. This might be due to the impact of sea ice during winter. Close to the Swedish coast an intense southward-directed flow becomes visible, being a part of the cyclonic gyre of the Baltic Proper. This flow is directed into Bornholm Basin and the Arkona Basin. The main flow crosses the central Arkona Basin and bifurcates north of Rűgen Island. One branch leaves the Baltic Sea at the Darss Sill and the flow continues through the Belt Sea and the Great Belt. The other branch recirculates and forms a cyclonic gyre in the Arkona Basin. A flow also follows the Swedish coast into the Öresund and Kattegat. In the lower layer, the flow follows the topography, from the Darss Sill into the Arkona Basin and further towards the Bornholm Channel passing Rűgen Island. The deep waters flow further to the Bornholm Deep and into the Stolpe Channel with a high persistency. East of the Stolpe Channel, the main flow is directed along the south-western slope of the Gdańsk Basin. In the Gotland Deep, the flow is characterised by cyclonic gyres. The water masses to the north also have a cyclonic gyre which finally leads part of the water to flow into the western Gotland Basin.

The mean circulation is likely to vary over the long term, with changes induced by wind forcing, heat fluxes and ice extent, as well as freshwater balance and inflow activity. On the basis of hindcast modelling for the period 1958–2001, Jedrasik et al. (2008) showed that annual average surface

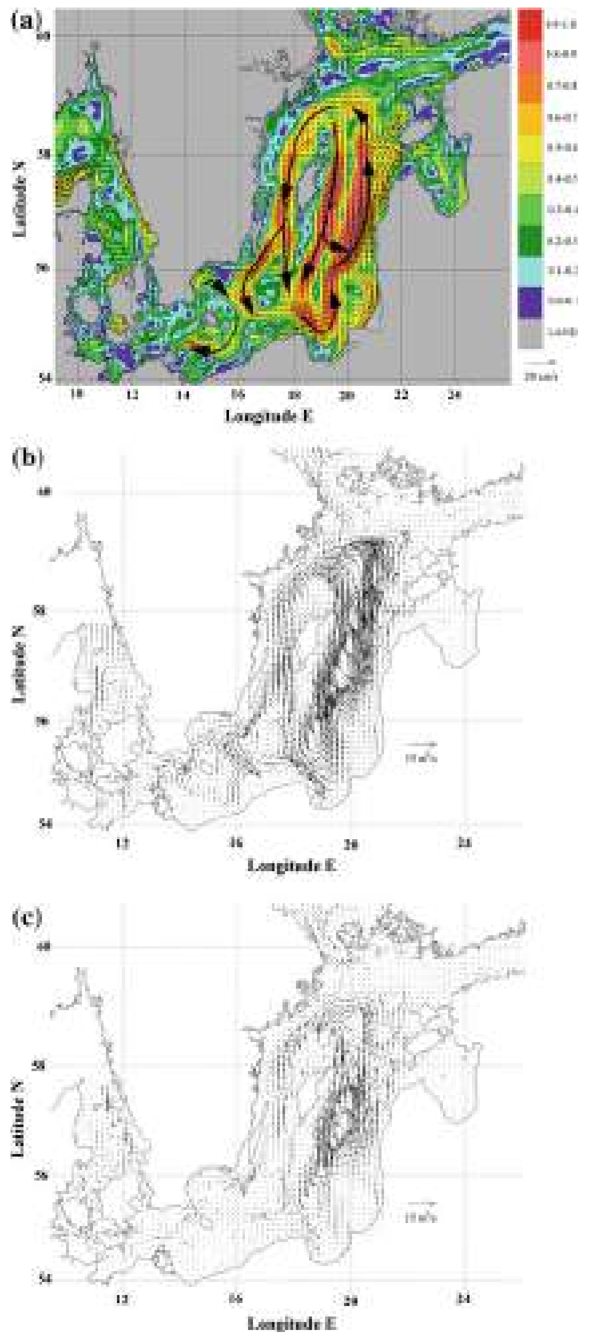

Fig. 7.6 Baltic Sea circulation in the Gotland Basin and adjacent sea areas from modelling results. Average barotropic currents (**a**) for 1992–1995 with the flow stability contours (redrawn from Lehmann and Hinrichsen 2000), and average transport per unit length for 1981–2004 above (**b**) and below (**c**) the halocline (redrawn from Meier 2007)

velocities (mean over the whole sea area, annual values of 16.5–19 cm s^{-1}) increased by 0.21 cm s^{-1} per decade. Over shorter periods (months), there is evidence from other researchers that water mass movements may take alternative paths. For instance, after the MBI in January 2003, saline water passing the Stolpe Channel flowed directly north-eastward along the eastern slope of the Hoburg Channel (the passage from the Stolpe Channel towards the eastern Gotland Basin) but after the baroclinic summer inflow in August/September 2002 the deep-water flow took the loop along the south-western slope of the Gdansk Basin (e.g. Meier 2007). Another example is from the Gulf of Finland, where less saline waters originating from the large rivers lying in the eastern part of the gulf usually spread along the northern coast. These less saline waters can occasionally detach from the coast and move for several months near the central axis of the gulf (Lips et al. 2011), as observed recently.

Finer-scale circulation in the Gulf of Finland was modelled by Andrejev et al. (2004a, b). Although the typical cyclonic mean circulation in the Gulf of Finland is apparent, the patterns and persistency of the currents according to Andrejev et al. (2004a) differ to some extent from the classical analyses by Witting (1912) and Palmén (1930). Both the mean and instantaneous circulation patterns in the Gulf of Finland contain many loops with a typical size that clearly exceeds the internal Rossby radius. The modelled circulation patterns reveal some non-trivial and temporally and spatially varying vertical structures.

Recent circulation studies in the Gulf of Finland have identified some new features of the system. Elken et al. (2011) carried out an EOF (Empirical Orthogonal Functions) analysis of hourly forecasts from the Baltic Sea operational High Resolution Operational Model for the Baltic, a version by the Swedish Meteorological and Hydrological Institute (HIROMB-SMHI) model for the period 2006–2008. It is possible to distinguish two regions with a specific regime of circulation variability. The western region behaves like a wide channel. Dominant EOF modes at different sections have similar patterns and their time-dependent amplitudes are well correlated. A prevailing mode of currents (23–42 % of the variance) is barotropic (unidirectional over the whole section) and its oscillation (spectral peak at 24 h) is related to the water storage variation (change in water volume) of the gulf. A two-layer flow pattern (surface Ekman transport with deeper compensation flow, 19–22 %) reveals both inertial and lower frequencies. The highest outflow of surface waters occurs during north-easterly winds. The eastern wider region has more complex flow dynamics and only patterns that are nearly uniform over the whole gulf were detected there. On the sea surface, quasi-uniform drift currents are deflected on average by 40° to the right of the wind direction (based on the best correlation between wind and rotated current deviations from their mean values; the model does not take into account wave effects on surface currents) and they cover 60 % of the circulation variance. Sea-level variability is heavily (98 %) dominated by nearly uniform changes which are caused by the water storage variation of the gulf. Sea-level gradients contain the main axis (23 %) and transverse (17 %) components, forced by winds of the same direction. The flows below the surface are decomposed into the main axis (24–40 %) and transverse (13–16 %) components that are correlated with the sea-level gradients according to geostrophic relations.

New detailed observational information for the Gulf of Finland was obtained by Lilover et al. (2011). They performed current velocity observations on Naissaar Bank in northern Tallinn Bay for 5 weeks in late autumn 2008 using a bottom-mounted ADCP (Acoustic Doppler Current Profiler) deployed at 8 m depth. Strong and variable, mainly southerly winds with speeds exceeding 10 m s^{-1} dominated in the area during 60 % of the study period. Bursts of seiche-driven currents with periods of 31, 24, 19.5, 16 and 11 h were observed after the passage of wind fronts. Inertial oscillations and diurnal tidal currents were relatively weak. The low-frequency current velocities gradually decreased towards the bottom at 3 cm s^{-1} over 4-m distance. The magnitude of the complex correlation coefficient between the current and wind for the whole series was 0.69, but was much higher (up to 0.90) within the shorter steady wind periods. The current was rotated $\sim$35° to the right of the wind. As an exception, during one period an anticlockwise surface-to-bottom veering of the current vector was observed. A topographically steered flow was seen either along isobaths of the bank during strong winds or along the ‘channel’ at the entrance to Tallinn Bay.

Soomere et al. (2011a) studied modelled circulation patterns and Lagrangian transport in the uppermost layer of the Gulf of Finland. Using the RCO (Rossby Centre Ocean) model data for 1987–1991, they revealed several normally concealed features of surface circulation. For certain years, a slow anticyclonic gyre may exist in the surface layer in the wide eastern and central part of the Gulf of Finland, reflecting a relatively weak coupling of the mostly Ekman-drift-driven surface-layer dynamics with that in the deeper layers. Semi-persistent (timescales of about a week, Viikmäe et al. 2010) patterns of rapid Lagrangian surface transport mostly follow the usual location of coastal currents but may stretch across the gulf during certain months and seasons (Soomere et al. 2011a).

The statistical analysis of Lagrangian surface transport has been employed to identify those areas of the Gulf of Finland from which the drift of passive tracers to the coast is unlikely (Soomere et al. 2011b). Transport is generally anisotropic and tracers in the surface layer generally have a greater chance of reaching the southern coast (Soomere et al. 2010), whereas in the eastern, wide part of the gulf, there is an extensive area of closed circulation from which transport to either of the coasts in unlikely (Soomere et al. 2011c). The results, however, are highly sensitive to the resolution of the underlying ocean model: the statistics of transport change substantially when the grid step is reduced from 2 miles to 1 mile; with a 1-mile grid step, all the significant eddies are resolved and a further decrease in grid step has little effect on results (Andrejev et al. 2011).

A numerical simulation of the circulation of the whole Baltic Sea was performed for the period 1991–2000 with a special focus on the Gulf of Bothnia (Myrberg and Andrejev 2006). Their results supported the traditional view of the cyclonic mean circulation in this basin. Persistency of currents ranges from 20 to 60 % and is greatest close to coasts, as observed many years ago by Witting (1912) and Palmén (1930). The simulation was performed using a barotropic model (Myrberg and Andrejev 2006), with a relatively high horizontal resolution (3.4 × 3.4 km), supporting the idea that the main features of the circulation can be reproduced with a barotropic, wind-driven model. However, the mean current velocities simulated by the model were clearly greater than those observed by Witting (1912) and Palmén (1930). The difference is apparently due to the insufficient resolution of the early measurements; these did not resolve mesoscale features, such as the pronounced differences in speed and direction of the coastal and open sea currents. In particular in the Bothnian Bay, the persistency pattern of the mean circulation was close to the results of Lehmann and Hinrichsen (2000), who presented depth-mean currents from the baroclinic model for a different period.

7.4.2 Dynamics in the Bottom Layer

Interaction between the upper and lower water layers is limited in the Baltic Sea due to the strong stratification. In the Kattegat, the dense North Sea originating waters form a deep-water pool, whereas the fresher Baltic waters are located in the surface layer. The deep-water circulation is characterised by dense bottom currents in the inflowing high-saline water at the mouth of the Baltic Sea. Convection and mechanical mixing, entrainment and vertical advection of water masses lead to interaction between the upper and lower water layers in other parts of the Baltic Sea.

Water effectively recirculates within the Baltic Sea even with the low-permeable halocline. This overturning circulation was termed the Baltic Sea 'haline conveyor belt' by Döös et al. (2004) as an analogy to the deep-water conveyor belt of the World Ocean. This vertical overturning circulation involves many important components: the gravity-driven dense bottom currents of the inflowing waters from the North Sea, the entrainment of ambient surface waters, mixing due to diffusion, interleaving of the inflowing water masses into the deep at the level of neutral buoyancy, and vertical advection due to the conservation and upward entrainment of deep water into moving surface water in the northern Baltic Proper.

Elken et al. (2003, 2006) investigated the large halocline variation and related mesoscale and basin-scale processes in the northern Gotland Basin—Gulf of Finland system. The authors suggested that long-lasting pulses of south-westerly winds cause an increase in the water volume of the Gulf of Finland. The resulting increase in hydrostatic pressure in the gulf leads to an outflow of deep water. Such counter-estuarine transport weakens the stratification of water masses at the entrance of the Gulf of Finland. As a consequence, the same energy input leads to an intensified diapycnal mixing as compared to the classical situation at the entrance (strong upward vertical advection). Owing to the variable topography both in the northern Gotland Basin and in the Gulf of Finland, the basin-scale barotropic flows are converted into baroclinic mesoscale motions with a large isopycnal displacement (more than 20 m within a distance of 10–20 km), which causes intra-halocline current speeds of more than 20 cm s^{-1}. So, Elken et al. (2006) concluded that the near-bottom layers of the Gulf of Finland actively react to the wind forcing, a reasoning which considerably modifies the traditional concept of the partially decoupled lower layer dynamics of the Baltic Sea. The multitude of processes at the entrance of the Gulf of Finland makes modelling the deep-water inflow extremely difficult. The internal wave activity is high, the production of strong eddies and topographically controlled local currents are frequent, and thus the diapycnal mixing is intense.

7.4.3 Mixing

There is a long-term approximate advective–diffusive balance in the deep water of the Baltic Sea (Stigebrandt 2001). Advective supplies of new deep water tend to increase the salinity while diffusive fluxes tend to decrease the salinity. However, this is not in balance over short timescales due to the discontinuous character of the advective supply of deep water. Since tides are usually weak in the Baltic Sea, most of the energy sustaining turbulence in the deep-water pools must be provided by the wind.

Based on long-term modelling of the large-scale vertical circulation in the Baltic Proper, Stigebrandt (1987, 2001) concluded that under contemporary conditions the basin-wide vertical diapycnal diffusivity (or diapycnal mixing coefficient) in the deep-water pools can be reasonably well described by

$$\kappa = \min(\alpha/N, \kappa_{\max})$$

where α and $\kappa_{\max}$ are constants and N is the Brunt-Väisälä frequency. In the horizontally integrated model for the Baltic Proper, Stigebrandt (1987) tuned α to equal 2×10^{-7} m^2 s^{-2}.

According to Meier et al. (2006), α depends on energy fluxes from local sources, such as wind-driven inertial currents, Kelvin waves and other coastally trapped waves. This means that mixing near the coasts and near topographic slopes is more thorough than in the open sea. Axell (1998) found, based on measurements, that $\alpha = 1.5 \times 10^{-7}$ m^2 s^{-2} and that there is also seasonal variability. For $N = 10^{-2}$ s^{-1} (typical for the halocline), the values of α/N are about 1.5×10^{-5} m^2 s^{-1}, while the normal level in the mixed layer is 10^{-3}–10^{-2} m^2 s^{-1}, which serves as a reference for κ_{max}.

The processes interlinked to diapycnal mixing are not yet fully understood. A key issue is to find the sources and paths for the energy sustaining the turbulence. It has been anticipated that internal waves and their dissipation plays a key role in the transfer of energy down into the deep water. Several mechanisms may generate internal waves.

In the DIAMIX (DIApycnal MIXing) project (Stigebrandt et al. 2002), Lass et al. (2003) measured dissipation rates and stratification between 10 and 120 m depths during a nine-day experiment in the eastern Gotland Basin. Their main finding was that there are two well-separated turbulent regimes. The turbulence in the surface layer, as expected, was closely connected to the wind. However, in the strongly stratified deeper water, turbulence was independent of the meteorological forcing at the sea surface. The integrated production of the turbulent kinetic energy exceeded the energy loss of inertial oscillations in the surface layer suggesting that additional energy sinks might have been inertial wave radiation during geostrophic adjustment of coastal jets and mesoscale eddies. The diapycnal mixing coefficient (α) of Stigebrandt (1987) was estimated to be 0.87×10^{-7} m^2 s^{-2}, several times less than earlier estimates based on the bulk methods.

A recent review paper by Reissmann et al. (2009) summarised the different mechanisms through which mixing takes place within the Baltic Sea. One involves the episodic overflow of high-saline water over the sills into the Baltic Sea, leading to entrainment and interleaving of the incoming water masses to the level of neutral buoyancy (e.g. Lass and Mohrholtz 2003). It is this mechanism that ventilates the Baltic deep waters (e.g. Meier et al. 2006). Owing to volume conservation, these episodic deep-water inflows drive vertical advection in the central Baltic Sea. Mixing due to inertial waves and the breaking of internal waves (Van der Lee and Umlauf 2011) also enhances vertical turbulent transport, as do Baltic Sea eddies (e.g. Lass et al. 2003) and coastal upwelling (Lehmann and Myrberg 2008). Winter convection and wind-induced mixing are also important mixing processes, but these only affect the layer above the halocline (e.g. Leppäranta and Myrberg 2009). Surface waves cause vertical mixing directly through wave breaking and indirectly through Langmuir circulation (e.g. Smith 1998). The effect of surface wave breaking is usually thought to penetrate to depths of only a few metres in the surface layer and it is often considered through the wind-speed-dependent (not the wave-dependent) friction velocity. Kantha and Clayson (2004) showed that the Stokes production of turbulent kinetic energy in the surface mixed layer is of the same order of magnitude as the shear production and must therefore be included in mixed layer models (see also the Baltic Sea case study by Kantha et al. 2010). The Stokes drift together with mean shear generates Langmuir cells. Taking Langmuir circulation into account in the vertical turbulence schemes affects the deepening of the mixed layer (e.g. Ming and Garrett 1997; Kukulka et al. 2010). Even though the small size of the Baltic Sea limits the growth of surface waves, the waves are high enough to be of significance even in the small sub-basins (e.g. Soomere and Räämet 2011; Tuomi et al. 2011). Summer is typically the season with the smallest mean and maximum significant wave height and winter the highest (excluding the seasonally ice-covered areas, see Chap. 9, Sect. 9.5). The importance of including the parameterisation of internal waves and Langmuir circulation in vertical turbulence schemes in multi-year simulations for the Baltic Sea was shown by Axell (2002).

7.5 Sensitivity to Changes in Forcing

Owing to the transient nature of the atmospheric conditions over the Baltic Sea, the flow field is highly variable, and thus changes in the resulting circulation and upwelling are difficult to observe. However, three-dimensional models, forced by realistic atmospheric forcing conditions and river run-off, have reached a state of accuracy such that the highly fluctuating current field and associated evolution of the temperature and salinity fields can be realistically simulated. Changes in the characteristics of the large-scale atmospheric wind field over the central and eastern North Atlantic can be described by the North Atlantic Oscillation (NAO, see Chap. 4, Box 4.1). Weakened westerlies across northern Europe, which is characteristic of the negative phase of the NAO, are a precondition for outflow from the Baltic Sea. Thus, negative phases of the NAO have the potential to indirectly affect circulation in the Baltic Sea and water mass exchange with the North Sea (Lehmann et al. 2002). The linear correlation between the volume exchange of the Baltic Sea and the NAO index is only $r = 0.28$ ($r = 0.49$ for the NAO winter index DJFM, December through March). A better relation of the wind field over the Baltic Sea to the large-scale atmospheric circulation is given by the (BSI), which is significantly related to the NAO. Furthermore, the BSI is highly correlated with the variation in water storage in the Baltic Sea and volume exchange with the Danish Sounds. For northern Europe, the NAO accounts for about 50 % of the dominant climate winter regimes, the 'blocking' and 'Atlantic Ridge' regimes account for another 27 and 23 %, respectively (Hurrell and Deser 2009; Lehmann et al. 2011). The local BSI includes all four regimes and thus better describes sea-level pressure variability over the Baltic Sea rather than the NAO alone.

Changes in the general wind conditions over the Baltic Sea lead to changes in upwelling. In a statistical study of upwelling based on satellite data for the period 1990–2009, Lehmann et al. (2012) analysed location and upwelling frequencies along the Baltic Sea coast during the thermally stratified period of the year. The most frequent upwelling occurred along the Swedish east coast (up to 40 % of the time) and the Finnish coast of the Gulf of Finland (15–20 %).

Generally, there was a positive trend in upwelling frequency along the Swedish coast of the Baltic Sea and the Finnish coast of the Gulf of Finland and a negative trend along the Polish, Latvian and Estonian coasts (Fig. 7.7). This is in line with the warming trend of annual mean SST derived from infrared satellite images (1990–2008) presented by Lehmann et al. (2011). The smallest trends occurred along the east coast of Sweden 0.3–0.5 °C per decade compared to 0.5–0.9 °C per decade in the central part of the Baltic Proper. The authors suggested that the decrease in the warming trend along the coast was due to increased upwelling connected with a shift in the dominant wind directions. The trend analysis of favourable wind conditions derived from wind station data for May to September over the period 1990–2009 supports this (Lehmann et al. 2012). There is a positive trend of south-westerly and westerly wind conditions along the Swedish coast and the Finnish coast of the Gulf of Finland and a corresponding negative trend of north-easterly and easterly winds along the east coast of the Baltic Proper, the Estonian coast of the Gulf of Finland and the Finnish coast of the Gulf of Bothnia. September contributes most to this trend, whereas in June and August a partial reverse of the trends occurs (see also Chap. 4, Sect. 4.3).

In a three-dimensional modelling study, Meier (2005) investigated the sensitivities of modelled salinity and water age on freshwater supply, wind-speed and sea-level amplitude in the Kattegat. Under steady-state conditions, the average salinity of the Baltic Sea is most sensitive to perturbations of freshwater inflow. Increased freshwater inflow and wind speed both resulted in decreased salinity, whereas increased amplitude of the Kattegat sea level resulted in increased salinity. The average water age was most sensitive to perturbations of the wind speed. In particular, decreased wind speed causes a significant increase in the age of the deep water. Long-term changes in freshwater and saltwater inflows and of low-frequency wind anomalies cause the Baltic Sea to adapt to a new steady state with a new salinity; stability and ventilation remain largely unchanged. Thus, for a change in state, the timescale of perturbations needs to be long compared to the turnover time of the freshwater content. By contrast, long-term changes in the high-frequency wind characteristics affect deep-water ventilation significantly.

Omstedt and Hansson (2006) analysed the Baltic Sea climate memory and response to change using both observations and modelling. Their findings can be summarised as follows. The averaged salinity of the Baltic Sea is nonlinearly dependent on and strongly sensitive to changes in freshwater inflow. The annual maximum ice extent is strongly sensitive to changes in the mean winter (DJF) air temperature over the Baltic Sea: at a mean air temperature of −6 °C the sea will become completely ice covered, whereas at 2 °C ice cover will not appear. In the Baltic Sea climate system at least two important timescales need to be considered: the first is associated with the water balance (salinity) and the e-folding time (i.e. the time interval for a change by the factor *e*) is approximately 33 years, while the other is associated with the heat balance and is approximately one year. Change in Baltic Sea annual mean water

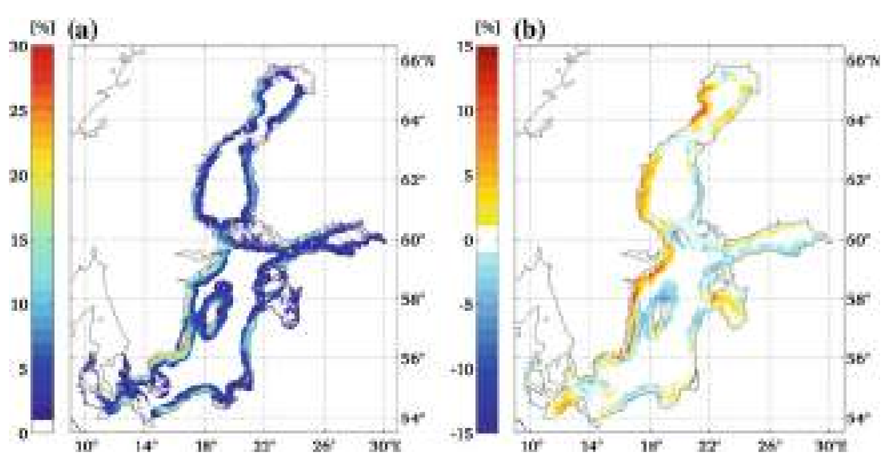

Fig. 7.7 Upwelling frequencies for May–September during the period 1990–2009 based on the automatic detection method for upwelling based on 443 sea-surface temperature maps (**a**) and the trend in upwelling frequencies (**b**), adopted from Lehmann et al. (2012)

temperature is closely related to the change in air temperature above the sea. However, in climate warming experiments, the water and air temperatures may differ due to changes in the surface heat balance components.

A study of climate change effects on the Baltic Sea ecosystem was presented by Neumann (2010). Two regional datasets for greenhouse gas emission scenarios (A1B and B1) for the period 1960–2100 were used to force transient simulations with a three-dimensional ecosystem model of the Baltic Sea. The outcome was a projected warming of the Baltic Sea of 1–4 °C, with a decrease in salinity and a much reduced sea-ice cover in winter. Most of the findings were consistent with an earlier study presented by Meier (2006). From a comparison of both studies using the same scenarios, it is clear that the magnitude of the warming is similar, thus demonstrating that warming is a robust feature. At the same time, the decrease in salinity differed, indicating far greater uncertainty with salinity-change simulations, albeit with a tendency towards reduced salinity. In addition, the season favouring cyanobacteria blooms is prolonged due to the change in temperature and mixing, with the spring bloom in the northern Baltic Sea beginning earlier in the season, while the oxygen conditions in deep water are expected to improve slightly.

Hordoir and Meier (2011) examined changes in future stratification in the upper water column using a three-dimensional ocean circulation model. They found a switch between processes controlling the seasonal cycle of stratification in the Baltic Sea at the end of twenty-first century. The air temperature increase was solely responsible for increased stratification at the bottom of the surface mixed layer. As in the present-day climate, winter temperatures in the Baltic Sea are often below the temperature of maximum seawater density at a given salinity, which means early warming causes thermal convection. Re-stratification at the beginning of spring is then triggered by the spread of freshwater. This process is believed to be important for the onset of the phytoplankton spring bloom. In a future climate, temperatures in the surface layers are likely to be higher than the temperature of maximum density for large parts of the year and thermally induced stratification would then start without prior thermal convection. Thus, freshwater-controlled re-stratification during spring would become considerably less important, and so this change in stratification might have an important impact on vertical nutrient fluxes and the intensity of the phytoplankton spring bloom in future. On the other hand, the freshening of waters will increase the temperature of maximum density and this would oppose the effect of increasing surface temperature. Changes in processes controlling the seasonal cycle of stratification were also investigated by Demchenko et al. (2011). The authors found differences in the formation and evolution of seasonal structural thermal fronts after winters of different severity. Structural fronts are related to the temperature of maximum water density. In spring, the front advances northwards at a rate of about 11–16 km d^{-1} traversing the breadth of the Baltic Sea within 8–10 weeks. After severe winters, the horizontal temperature gradient is more pronounced and the rate at which the Baltic Sea is traversed decreases. For a full overview of the potential impacts of these processes on biogeochemistry and ecosystems, see Chaps. 18 and 19, respectively.

7.6 Conclusion

A recent warming trend in sea-surface waters has been clearly demonstrated by in situ measurements, remote sensing data and modelling results. In particular, remote sensing data for the period 1990–2008 indicate that the annual mean SST has increased by up to 1 °C per decade, with the greatest increase in the northern Bothnian Bay and large increases in the Gulf of Finland, the Gulf of Riga and the northern Baltic Proper. Although the increase in the northern areas is affected by the recent decline in the extent and duration of sea ice, warming is still evident during all seasons and with the greatest increase occurring in summer. The least warming of surface waters (0.3–0.5 °C per decade) occurred north-east of Bornholm Island up to and along the Swedish coast, probably owing to an increase in the frequency of coastal upwelling. The latter is explained by the change in atmospheric circulation. Comparing observations with the results of centennial-scale modelling, recent changes in seawater temperature, appear to be within the range of the variability observed during the past 500 years.

The salinity and stratification of the deep waters, governed by interacting circulation, transport and mixing processes, are strongly linked to the Major Baltic Inflows of North Sea water that occur sporadically and bring high-saline water into the deep layers of the Baltic Sea. These major inflows are often followed by a period of stagnation during which saline stratification decreases and oxygen deficiency develops in the bottom waters. Major inflows normally occur during winter and spring; they bring relatively cold and oxygen-rich waters to the deep basins. Since 1996, several large inflows have occurred during summer. These baroclinic inflows have transported high-saline, but warm and low-oxygen water into the deep layers of the Baltic Sea.

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8 Recent Change—Sea Ice

Jari J. Haapala, Iina Ronkainen, Natalija Schmelzer, and Marzenna Sztobryn

Abstract

Sea ice conditions in the Baltic Sea have been systematically monitored for more than 100 years. All sea ice-related parameters display large interannual variability, but a change towards milder ice winters has been observed over the past 100 years: in particular, the annual maximum ice extent has decreased and the length of the ice season has become shorter. There is no correlation between consecutive ice seasons because the thermal memory of the Baltic Sea is only 2–3 months. Interannual variability in sea ice conditions is principally driven by the large-scale atmospheric circulation, described by the North Atlantic Oscillation. In addition to a tendency towards milder winters, the occurrence of severe ice winters has also decreased considerably over the past 25 years.

Keywords

Sea ice • Extent • Duration • Thickness • Climate variations and change • Baltic Sea

8.1 Introduction

The importance of understanding the variability and changes in sea ice has been recognised for centuries. The first written accounts of ice conditions in the Baltic Sea were documented in harbour logbooks of the Fourteenth century (Tarand and Nordli 2001; Schmelzer and Holfort 2011). A regular observational network was established in the nineteenth century among the Baltic Sea countries, mainly to provide guidance for shipping (Jevrejeva et al. 2004), but later for scientific analyses of ice conditions (Speerschneider 1927; Jurva 1937; Palosuo 1953). In the first Baltic Sea assessment (BACC Author Team 2008), Schmelzer et al. (2008) and Vihma and Haapala (2009) gave extensive summaries of the existing literature. This chapter updates those reviews but does not repeat the extensive discussions already published.

The state of sea ice is determined by its extent, thickness and drift. Among these variables, ice extent is the most reliable measured quantity. In the past, ice extent was determined by visual observations from the coast, ships and aircraft, but since the start of the satellite era in the 1970s, accurate daily measurements are possible. In the Baltic Sea, regular ice thickness measurements are limited to the landfast ice sites. These measurements, conducted by drilling, although accurate represent variations in ice thickness for coastal areas only. Sea ice drift can be determined by position logging buoys (Leppäranta et al. 2001) or by detecting sea ice displacement using pairs of satellite images (Leppäranta et al. 1998; Karvonen 2012), but so far, no studies have been conducted focusing on long-term variability and

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_8

changes in sea ice drift. Apart from these basic physical variables, the length of the ice season and dates of freezing and break-up of the ice cover have also been widely used as climatological indicators of ice season severity.

The Baltic Sea ice season usually lasts for six to eight months. Typically, coastal shallow areas in the Bothnian Bay become ice-covered in October–November. During the next phase, the ice cover advances towards the central parts of the Bothnian Sea, and coastal areas of the other Baltic Sea sub-basins also become ice-covered. On average, the ice-covered area is greatest in March, when the maximum ice extent in the Baltic Sea (MIB) is about 165,000 km^2, or 40 % of the total area of the Baltic Sea. The limit of the ice is typically located in the northern Baltic Proper, which means the Gulf of Bothnia, Gulf of Finland and Gulf of Riga are all completely ice-covered. In extremely severe winters, the entire Baltic Sea has been ice-covered (420,000 km^2). During the mildest winters, sea ice is formed only in the Bothnian Bay and the eastern Gulf of Finland. The minimum MIB to date is 49,000 km^2. The length of the ice season is 130–200 days in the Bothnian Bay, 80–100 days in the Gulf of Finland and 0–60 days in the southern Baltic Sea.

Near the shore, sea ice remains stationary, and its thickness is controlled by thermodynamic factors only. That ice is termed ‘fast ice’. In those regions, horizontal ice thickness variations are small, typically 10–20 cm within a scale of kilometres. Most of the monitoring data, in particular for ice thickness, have been collected from the fast-ice regions. The record for measured fast-ice thickness is 1.22 m from Tornio, Finland.

Ice thickness in the drift ice regions is much more variable. Drift ice is a mixture of different ice types, and within a scale of kilometres, ice thickness can vary from a few centimetres of thin new ice up to thick pressure ridges. Ice ridges, formed by compressive ice motion, are typically 3–5 m thick, but freely floating ridges 25 m thick have been observed in the Baltic Sea. Ice thickness measurements in drift ice areas have been made during several field campaigns, but the observations are too sparse in time and space to examine long-term change.

The state of sea ice depends on the surface energy balance, momentum flux and water flux. Some of the climate-related state variables, such as ice extent or freezing date, are mainly driven by the energy balance, and the observed variation in these variables very closely reflects large-scale variation in air temperature. However, many variables, such as ice thickness, ice type and duration of the ice-covered period, are also very dependent on winds and ocean currents and may reflect changes occurring at a local scale: this highlights the importance of a unified analysis of ice parameters for quantifying change in sea ice on a climatological scale. An alternative approach for analysing directly measured geophysical parameters is to use a particular ‘sea ice index’ as a general indicator of ice conditions. For the Baltic Sea, two methods have been used. Koslowski and Glaser (1995) used an index that describes the mass of sea ice by integrating ice thickness over the sea area in question, while Sztobryn et al. (2009) developed an index based on the normalised length of the ice-covered season of several coastal stations. In the Baltic Sea, all these variables display a large interannual variability and are very closely related to the variability in the large-scale atmospheric circulation (Omstedt and Chen 2001; Jaagus 2006; Vihma and Haapala 2009, see also Chap. 4, Sect. 4.2).

Recent years have highlighted the magnitude of the interannual variability of ice conditions in the Baltic. The winters of 2008 and 2009 were very mild; sea ice formed only in the Bothnian Bay and coastal areas of the Bothnian Sea and Gulf of Finland. The MIB was only 49,000 km^2 in 2008 and 110,000 km^2 in 2009. The winters of 2010 and 2011, on the other hand, are classed as severe ice winters, and the MIBs were correspondingly larger: 244,000 km^2 (2010) and 309,000 km^2 (2011). More importantly, the ice was thick, and severe storms induced pack ice compression events that caused major difficulties for winter navigation. In the past, there have been similar relatively short periods with both extremes of ice conditions, such as 1873–1876, 1938–1941 and 1985–1990. It is also evident that there is no correlation between consecutive ice seasons, simply because the thermal memory of the Baltic Sea is only 2–3 months (Leppäranta and Myrberg 2009).

The first Baltic Sea assessment (BACC Author Team 2008) concluded that climate warming can be detected from the time series of MIB and ice season length. This conclusion was based on studies utilising observations that extended only until 2000. The present review is an updated analysis and includes observations up to 2011. Observed trends in sea ice parameters are summarised in Table 8.1.

8.2 Ice Extent

Annual MIB is the most widely used indicator of sea ice changes because it integrates the winter period weather over the entire Baltic Sea basin (Fig. 8.1). The time series is based on various observation methods. The early part of the time series has considerable uncertainty. It is mainly based on correlation between air temperature recordings, observations from lighthouses, information from newspapers and records of travels on ice (Jurva 1937; Palosuo 1953; Seinä and Palosuo 1996). According to Jurva (1952), MIB has been fairly accurately determined from 1880 onwards from notes made onboard ships navigating in the middle parts of the Baltic Sea during winter (Vihma and Haapala 2009).

Table 8.1 Trends in the annual maximum sea ice extent (MIB), annual maximum sea ice thickness (MIT) and duration of the ice season (DIS), updated summary table based on Vihma and Haapala (2009)

	Region	Period	Trend	Statistical significance	Source
MIB	Baltic Sea	1901–1995	–	No (<90 %)	Haapala and Leppäranta (1997)
MIB	Baltic Sea	1720–1995	–	Yes (97 %)	Haapala and Leppäranta (1997)
DIS	Polish coast	1896–1993	–	Yes	Sztobryn (1994)
DIS	Szczecin Lagoon	1888–1995	–	Yes	Girjatowicz and Kozuchowski (1999)
DIS	Finnish coast	1889–1995	–	Yes (99 %)	Haapala and Leppäranta (1997)
DIS	Finnish coast	1950–2010	–	NR	Ronkainen (2013)
DIS	Port of Tallinn	1500–2000	–	NR but evident since mid-1800s	Tarand and Nordli (2001)
DIS	West Estonian archipelago	1949–2004	–	Yes	Jaagus (2006)
DIS	Southern coast of the Gulf of Finland	1949–2004	–	No	Jaagus (2006)
DIS	Gulf of Finland and Gulf of Riga	1900–1990	–	NR	Jevrejeva (2000)
DIS	Port of Riga	1529–1990	–	Yes (99.9 %) for severe winters; no for mild and average winters	Jevrejeva (2001)
DIS	Baltic Sea coasts	1900–2000	− and +	Depends on station	Jevrejeva et al. (2004)
MIT	Baltic Sea coasts	1900–2000	− and +	No	Jevrejeva et al. (2004)
MIT	Gulf of Bothnia	1899–1995	+ (Kemi)	Yes (Kemi)	Haapala and Leppäranta (1997), Launiainen et al. (2002)
			− (other)	No (other)	
MIT	Finnish coast	1950–2010	+ (Kemi)	NR	Ronkainen (2013)
			− (Loviisa)		
MIT	Gulf of Bothnia	1980–2000	–	NR but evident	Launiainen et al. (2002)
MIT	Northern Gulf of Finland	early 1900–1990s	–	NR	Alenius et al. (2003)

− decreasing trend; + increasing trend; *NR* statistical significance of the trend not reported

The MIB displays large year-to-year natural variability due to the large-scale variation in atmospheric circulation, commonly described by the North Atlantic Oscillation (NAO, see Chap. 4, Box 4.1). Vihma and Haapala (2009) used the mean January–March NAO index in their analysis and showed that during a negative NAO phase (NAO < −0.5), the mean MIB is 259,000 km^2 (range: 150,000–405,000 km^2), while during a positive NAO phase (NAO > +0.5), the mean MIB is 121,000 km^2 (range: 45,000–337,000 km^2) (Fig. 8.2).

In spite of the obvious correlation between the NAO and MIB, in some NAO-positive winters, sea ice extent has been greater than average. In the middle of winter, the heat content of the Baltic Sea is already low, and a long-lasting blocking situation could cause an anomalously cold period and extensive ice growth. Interestingly, the relationship between the NAO index and MIB has not remained constant over time (Omstedt and Chen 2001; Janssen 2002; Schrum and Janssen 2002; Chen and Li 2004). Meier and Kauker (2002) found that during two periods, around 1926 and 1966, the correlation increased simultaneously with improvements in the observation methods for MIB. Changes in the NAO–MIB relationship can, however, also be due to changes in the location of the atmospheric pressure patterns (Koslowski and Loewe 1994; Chen and Li 2004).

All previous studies have reported a significant decreasing trend in MIB (BACC Author Team 2008; Vihma and Haapala 2009). Including observations until 2011, the trend in MIB for the past 100 years is −3400 km^2 10 year^{-1} or ~2 % 10 year^{-1} (Fig. 8.3). Another apparent change is a fall in the frequency of severe winters over the past 20 years. Figure 8.4 shows that the modal probability of the MIB has remained the same whether the period considered is the past 100, 30 or 20 years, but mild ice seasons have become more common, and years when the Baltic Sea is almost completely ice-covered have not occurred over the past 25 years.

Schmelzer et al. (2012) examined changes in the probability of sea ice occurrence in the southern Baltic Sea. Figure 8.5 compares results for two 30-year periods, 1961–1990 and 1981–2010, and shows a general tendency

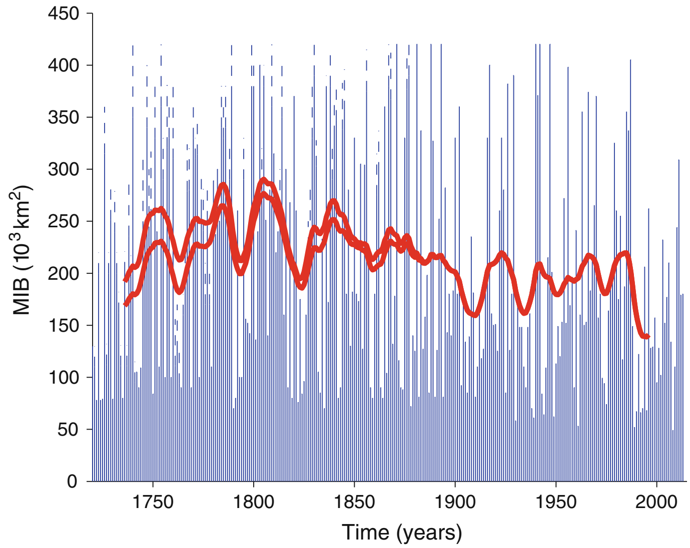

Fig. 8.1 Maximum ice-cover extent in the Baltic Sea (MIB), 1720–2012. The *dashed bars* represent the error range of the early estimates. The 30-year moving average is indicated by *two lines* representing the error range early in the series, converging into one line when high-quality data became available

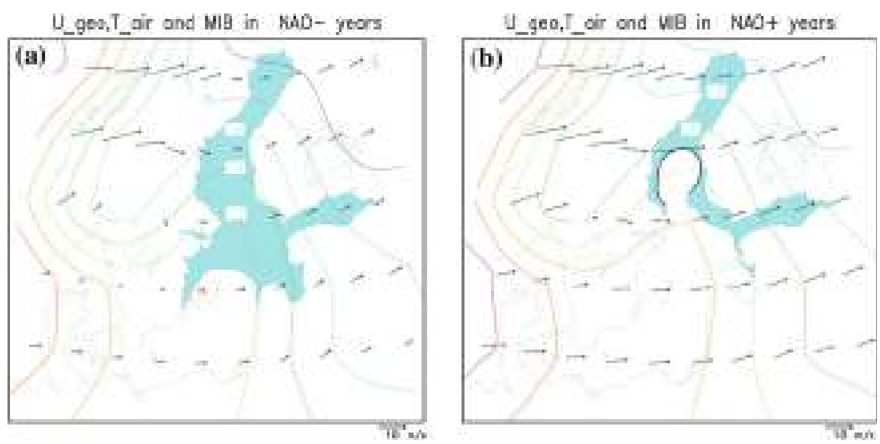

Fig. 8.2 Mean annual maximum sea ice extent (*blue*), 2-m air temperature (*coloured isolines*) and 10-m wind vectors over the Baltic Sea region during two phases of the North Atlantic Oscillation: a negative phase (NAO < −0.5) and a positive phase (NAO > +0.5) (redrawn from Vihma and Haapala 2009)

towards ice-free winters. Major changes have occurred in the offshore area of the western Baltic Sea and east of Bornholm Island, with the lowest frequency of ice occurrence observed in the past 30 years. There has been a concomitant decrease in the frequency of 'difficult' ice conditions (coverage > 7/10 and thickness > 10 cm).

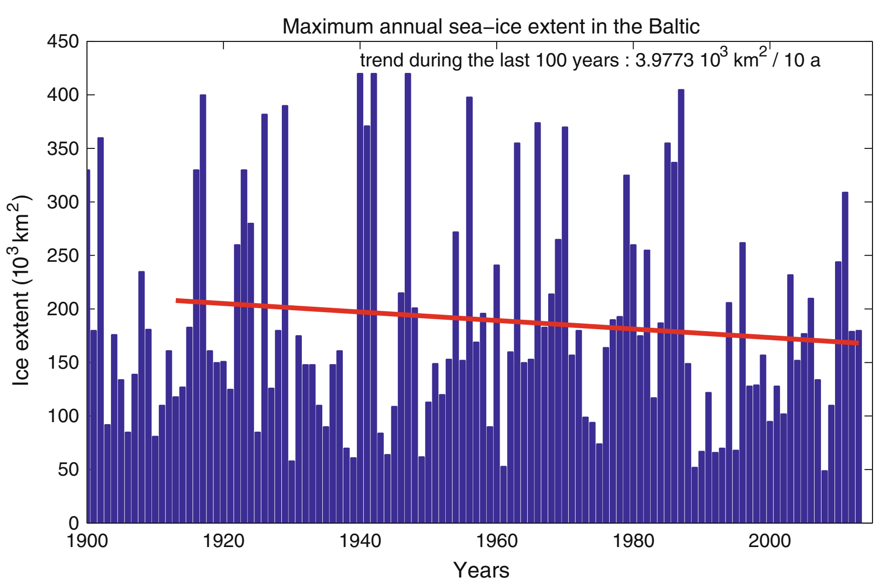

Fig. 8.3 The maximum extent of sea ice cover in the Baltic Sea, 1900–2012. The *red line* shows a long-term declining trend of ~2 % per decade

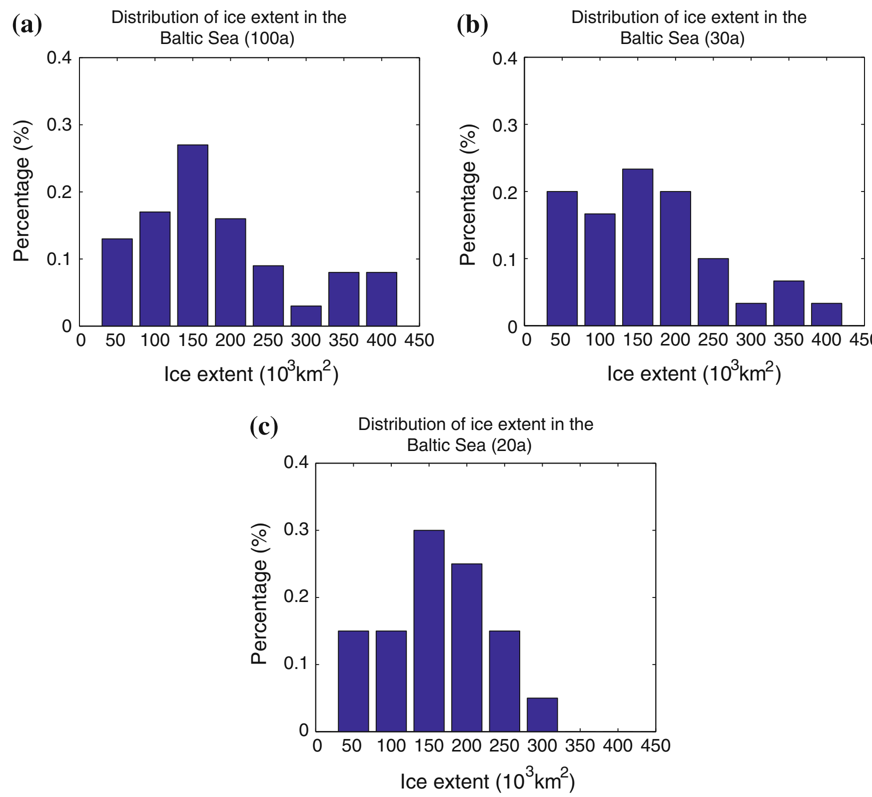

Fig. 8.4 Probability distributions of maximum sea ice extent in the Baltic Sea during 1911–2011, 1981–2011 and 1991–2011

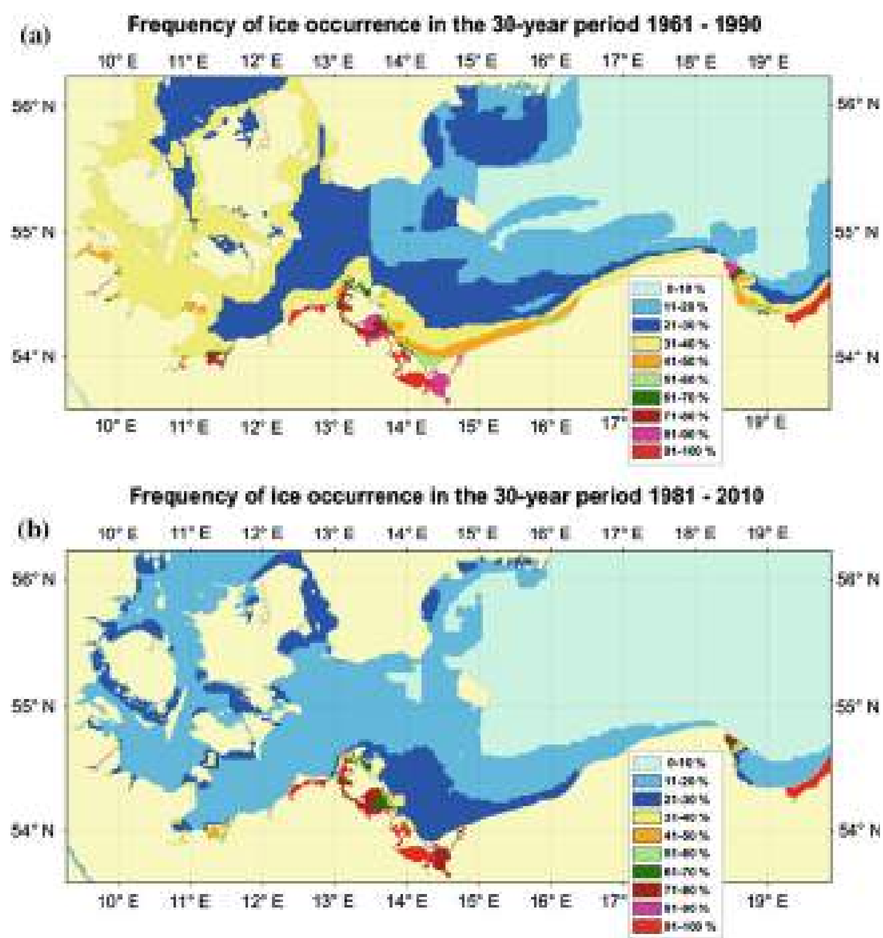

Fig. 8.5 Probability of sea ice occurrence in the southern Baltic Sea (Schmelzer et al. 2012)

8.3 Ice Duration

Jevrejeva et al. (2004) has compiled the most extensive analysis of ice season changes in the Baltic Sea to date, using time series at 37 coastal stations from the Bothnian Bay to the Arkona Basin. In general, observations showed a tendency towards milder ice conditions. For the variables studied, the greatest change occurred in ice season length, which has decreased by 14–44 days over the past 100 years, mostly due to earlier ice break-up.

A recent study by Ronkainen (2013) confirmed the trend towards a shorter ice-covered period in the Finnish coastal zone (Fig. 8.6 and Table 8.2). In the Bothnian Bay (Kemi), where the ice season is clearly longer than in the eastern Gulf of Finland (Loviisa), the trend is −18 d 100 year^{-1}. In the eastern Gulf of Finland, where sea ice is also formed every winter, larger changes have been observed. The length of the ice season has decreased by 41 d 100 year^{-1}, while the trend based on the last 50 observations is −62 d 100 year^{-1}.

In the southern Baltic Sea, the first sea ice is formed in the coastal lagoons between 10 and 20 December, near the coast around 1 January, in the sea area of the western Baltic and off the Polish coast in mid-January, and in the German coastal areas at the beginning of January (Schmelzer et al. 2008, 2012; Girjatowicz 2011). Comparing the different periods analysed, there has been no major change in the average date at which ice formation begins (Fig. 8.7). In terms of ice break-up, on average, the last ice in the inner coastal waters has disappeared by late February or early March, while residual ice has been observed in offshore waters until mid-March. The longest duration of ice in the coastal waters was observed in the 1961–1990 winters. In the past 30-year period, ice disappeared from the eastern coastal waters up to ten days earlier. In the Baltic Sea offshore waters, no change has been observed in this parameter, and in the western Baltic Sea, there has even been a weak trend towards a later break-up date.

Ice duration decreases from east to west, and from the inner waters towards the sea areas. A weak trend towards shorter ice duration was found for the 1981–2010 period. The maximum ice duration occurs in particularly cold and long winter periods. There has been at least one cold winter with strong ice formation in each of the 30-year periods analysed, for example 1962/63, 1978/79 and 1995/96. A significant decrease (30 days) has only been observed in the sea area west and east of Bornholm Island in the period 1971–2000. For the period 1981–2010, ice duration during very cold winters is nearly identical to that in 1971–2000.

8.4 Ice Thickness

For climate studies, the sea ice thickness or, preferably, the large-scale sea ice thickness distribution should be the main indicator of change in sea ice, since it is essentially the same as the mass of the ice pack. However, interpreting the causes of change in sea ice thickness is not straightforward, since in addition to the atmosphere–ocean energy balance, sea ice thickness depends on snow thickness and ice dynamics.

Monitoring activity in the Baltic Sea is limited to the landfast ice regions, where the sea ice could be thinner than in the drift ice regions. According to Jevrejeva et al. (2004), the ice thickness time series around the Baltic Sea coast does not show any consistent trend: both decreasing and increasing trends were reported. A recent study by Ronkainen (2013) supported these conclusions. In the Bothnian Bay (station

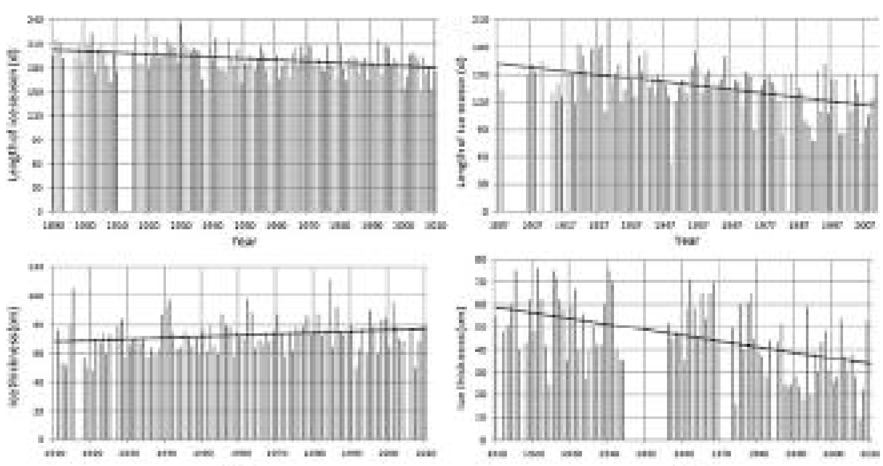

Fig. 8.6 Observed changes in **a** length of ice season in Kemi and **b** in Loviisa, and **c** ice thickness in Kemi and **d** in Loviisa (Ronkainen 2013)

Table 8.2 Change in freezing date, break-up date, length of ice season and ice thickness in the Finnish coastal zone of the Baltic Sea

		Kemi, Bothnian Bay	Utö, Archipelago Sea	Loviisa, Gulf of Finland
	Probability of ice appearance	1.00	0.81	1.00
	Trend 100 year^{-1}	0	−0.19	0
Freezing date	Number of observations	112	97	106
	Mean date	10 Nov	27 Jan	7 Dec
	Trend (days 100 year^{-1})	7	(24)	20
	Trend pre-1950 (days 100 year^{-1})	12	(65)	25
	Trend post-1950 (days 100 year^{-1})	0	(−15)	24
Break-up date	Number of observations	113	97	10
	Mean date	20 May	9 Apr	24 Apr
	Trend (days 100 year^{-1})	−11	(−16)	−20
	Trend pre-1950 (days 100 year^{-1})	−5	(−1)	−8
	Trend post-1950 (days 100 year^{-1})	−17	(−34)	−38
Length of ice season	Number of observations	113	121	102
	Mean (days)	190	50	137
	Trend (days 100 year^{-1})	−18	−46	−41
	Trend pre-1950 (days 100 year^{-1})	−17	−84	−32
	Trend post-1950 (days 100 year^{-1})	−16	−36	−62
Ice thickness	Number of observations	94		81
	Mean (cm)	73		46
	Trend (cm 100 year^{-1})	9		−25
	Trend pre-1950 (cm 100 year^{-1})	13		−29
	Trend post-1950 (cm 100 year^{-1})	4		−52

Data for three sites: Kemi (1890–2010), Utö (1889–2010) and Loviisa (1894–2010) (Ronkainen 2013). The parentheses indicate values that are not based on all data because the probability of ice occurrence is less than one

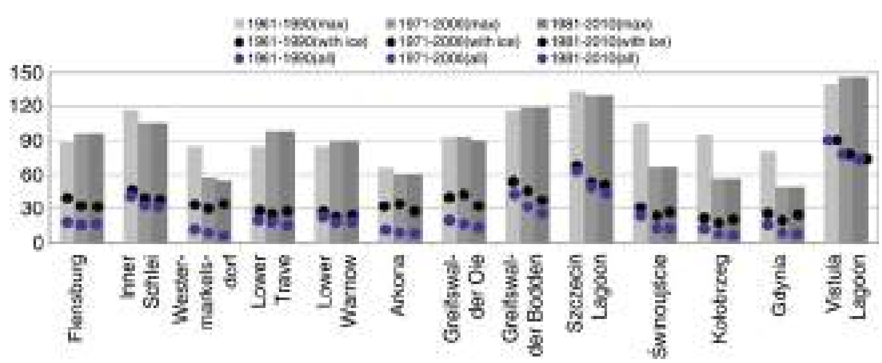

Fig. 8.7 Mean number of days with ice (all winters), mean number of days with ice (only winters with ice) and maximum number of days with ice on the southern Baltic Sea coast in the 30-year periods 1961–1990, 1971–2000 and 1981–2010 (Schmelzer et al. 2012)

Kemi), ice thickness shows a slight increasing trend (+9 cm) over the past 100 years, while in the Gulf of Finland, a thinning trend is observed (−25 cm).

In the southern Baltic Sea, ice thickness in the sheltered coastal waters reached 10–20 cm on average, and as much as 40–62 cm in the extremely hard ice winter of 1962/63 (Fig. 8.8). In most parts of the Baltic Sea offshore waters, ice thickness reaches 5–20 cm, but in particularly cold and long winters, the ice grows to 50 cm thickness in some areas. A maximum ice thickness of 70 cm was reached in the Kiel and Lübeck bays and in the waters of Rügen in the winter of 1962/63. In Szczecin Lagoon, in the Bay of Puck and in Vistula Lagoon, the ice reached 62 cm thickness. Since 1971, the maximum ice thickness has been about 50 cm in the southern Baltic Sea coast and up to 30 cm in the sea area north of Rügen.

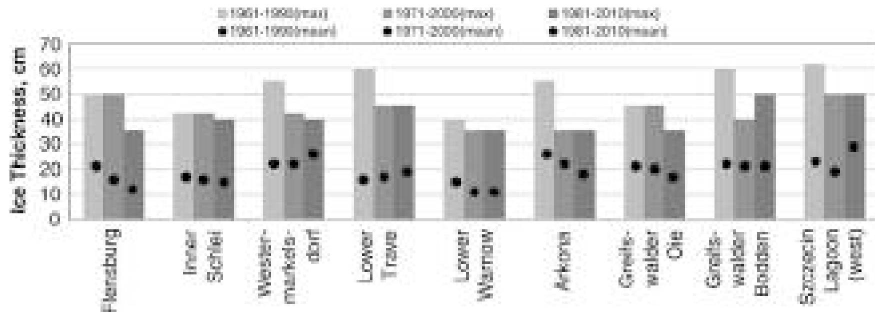

Fig. 8.8 Mean and extreme annual maximum ice thickness (for winters with ice only) on the German Baltic Sea coast within the 30-year periods 1961–1990, 1971–2000 and 1981–2010 (Schmelzer et al. 2012)

8.5 Conclusion

Sea ice conditions in the Baltic Sea have been systematically monitored for more than a century. All sea ice-related parameters display large interannual variability, but a change towards milder ice winters has been observed over the past 100 years: in particular, the annual maximum ice extent has decreased, and the length of the ice season has become shorter. Interannual variability in sea ice conditions is principally driven by the large-scale atmospheric circulation. In particular, the winters of strong westerly circulation, that is during positive phases of the NAO, have manifested as a minimum ice cover in the Baltic Sea. However, a winter without any ice formation in the Baltic Sea is far from the present climatology. According to some 300 years of records of annual maximum sea ice extent, the northernmost sub-basin, the Bothnian Bay, has been entirely ice-covered even during the mildest winters, and the length of the ice season near the coast has been 150 days at a minimum.

In addition to a tendency towards milder winters, the occurrence of severe ice winters, when the southern Baltic Sea is ice-covered, has also decreased considerably over the past 25 years. Although these observations are consistent with the changes in global climate, ice season length declined to a similar extent during the first half of the twentieth century, when anthropogenically derived greenhouse gas emissions are likely to have had almost no impact on climate. Long-term changes in the Baltic Sea regions have been interpreted as the result of a recovery from the 'Little Ice Age' (Omstedt et al. 2004), but changes in sea ice could also be due to shipping. Motivation for sea ice measurements has been the provision of data for shipping, and consequently, monitoring sites have typically been located near harbours. Ship-induced waves are known to prevent the formation of a permanent ice cover in autumn and also to enhance break-up of the ice cover in spring, and so an increase in the size of vessels and the intensity of shipping activity could also affect local ice conditions.

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9 Recent Change—Sea Level and Wind Waves

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Abstract
This chapter describes observed changes in sea level and wind waves in the Baltic Sea basin over the past 200 years and the main climate drivers of this change. The datasets available for studying these are described in detail. Recent climate change and land uplift are causing changes in sea level. Relative sea level is falling by 8.2 mm year^{-1} in the Gulf of Bothnia and slightly rising in parts of the southern Baltic Sea. Absolute sea level (ASL) is rising by 1.3–1.8 mm year^{-1}, which is within the range of recent global estimates. The 30-year trends of Baltic Sea tide gauge records tend to increase, but similar or even slightly higher rates were observed around 1900 and 1950. Sea level in the Baltic Sea shows higher values during winter and lower values during spring and this seasonal amplitude increased between 1800 and 2000. The intensity of storm surges (extreme sea levels) may have increased in recent decades in some parts of the Baltic Sea. This may be linked to a long-term shift in storm tracks.

9.1 Introduction

Sea surface height (SSH) is an important indicator of variability and long-term change in climate. An understanding of the processes driving future trends in SSH on global to regional scales requires an understanding of the interannual to decadal (long-term) variability in the observational period. This requires an accurate assessment of past and recent change in global and regional SSH, including changes in mean sea level (MSL), extreme sea level (storm surges) and wind-generated waves. This chapter reviews what is known about change in SSH in the Baltic Sea region over the observational period (effectively the past 200 years) and the main drivers of this change. This includes a description of the datasets available for studying sea level and wind waves and a review of major findings for the Baltic Sea region.

Box 9.1 Definition of key sea level terms
Change in the mean Baltic Sea sea level is the sum of global, regional and local effects. There are several definitions of the key terms necessary for understanding these effects in the literature. Thus, a clear definition of such terms is essential.

Mean sea level (MSL) is defined as the average height of the sea surface at a specific location, neglecting short-term effects of tides and storm surges. Some studies, for instance Woodworth et al. (2011), consider annual means to define MSL, although other studies focus on monthly or seasonal means. The MSL can in turn be absolute or relative.

Absolute sea level (ASL) is the height of the sea surface at a given location relative to a geocentric

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_9

reference such as the reference ellipsoid and is measured by satellite altimetry.

Relative sea level (RSL) is the height of the sea surface relative to the sea floor, and thus to land, at a given location, and is estimated using tide gauges or sea level reconstructions using information from the geological record. Relative sea level is more important than ASL for impact studies (see Chaps. 15–22). The land surface may undergo vertical movements that may occur at millennial timescales (e.g. the glacial isostatic adjustment, GIA) or at shorter timescales (e.g. due to water extraction or oil mining), or sudden changes (e.g. during earthquakes). In the Baltic Sea basin, GIA has a very important influence on RSL, stronger in the northern Baltic Sea and weaker in the southern Baltic Sea. The GIA is caused by the adjustment of the Earth's crust and mantle to the disappearance of the ice sheets of the last glacial age. Thus, RSL trends may differ strongly from ASL trends in the Baltic Sea region.

Global mean sea level (GMSL) is the average of sea level over the global ocean and over a period longer than the timescales for tides and surges.

Local sea level (LSL) reflects changes owing to tides and surges down to timescales of a few minutes, including *extreme sea level* variations, at a specific locality.

9.2 Sea Level Observations

9.2.1 Tide Gauges

The most direct measurement of RSL is from tide gauges. Globally, the coverage of sea level data from tide gauges is limited in time and space. Most data are from the Northern Hemisphere and are for the latter half of the twentieth century. The Baltic Sea has one of the longest running and most densely spaced tide gauge networks in the world; many stations have been in continuous operation since the late nineteenth century. For example, the Stockholm time series is one of the oldest sea level records (Ekman 2003; see also Woodworth 1999 for comparison with other similar records). A remarkable number of long and high-quality sea level records are available for the Baltic Sea basin and, to date, more than 45 stations provide at least 60 years of data.

Many of the sea level time series are freely available online through the Permanent Service for Mean Sea Level (PSMSL), which provides monthly and annual MSL measurements from around the world (Woodworth and Player 2003, www.psmsl.org), processed from high-frequency datasets. In these archives, stations with a full history of reference levels (benchmark datum history) have had their data adjusted to a fixed, revised local reference (RLR). Although the RLR time series are constructed to be continuous and relative to landmarks on the nearby land, corrections for the vertical movement of the reference (e.g. due to GIA) are not applied, because this requires a careful estimation of these movements usually involving a geodynamic model, which is not always available. Users of RLR data are aware of these limitations. The PSMSL RLR records are the most commonly used data bank for global and regional studies of historic sea level rise. However, as is the case for many other climate datasets (e.g. near-surface air temperature) for some locations significant differences exist relative to other datasets. The differences may stem from the original data source or from different error correction methods (e.g. Dimke and Fröhle 2009).

Historic sea level time series are available from several other sources, for example, for Stockholm (Sweden) (1774–2000; Ekman 2003), Kronstadt (Russia) (1816–1999; Bogdanov et al. 2000), Travemünde (Germany) (since 1826; Jensen and Töppe 1986) and various national institutions.

To illustrate possible data limitations, Richter et al. (2007) analysed RSL data from archives of six different national state authorities for the German southern Baltic Sea coast. The records in these archives were often incomplete and lacked important additional information and a systematic cataloguing. To produce homogeneous datasets from such documents (written on paper) requires considerable effort. This is not only due to obvious sources of uncertainty, such as different measurement techniques and sampling frequencies (e.g. Richter et al. 2007; Ekman 2009), but also to change in the reference points. Such change could arise from the relocation of landmarks and to vertical landmark movements caused by geodynamic processes or to man-made modification of the tide gauge surroundings (e.g. coastal or port construction activities) (e.g. Bogdanov et al. 2000). Long Baltic Sea sea level records from different sources (not only PSMSL) have been used in a wide range of research papers for global sea level studies (e.g. Douglas 1992; Nakada and Inoue 2005; Jevrejeva et al. 2006, 2008; Merrifield et al. 2009; Woodworth et al. 2009; Church and White 2011; Houston and Dean 2011) and regional Baltic Sea studies (e.g. Andersson 2002; Omstedt et al. 2004; Chen and Omstedt 2005; Jevrejeva et al. 2006; Hünicke et al. 2008) (Table 9.1). Figure 9.1 shows sea level records for the Baltic Sea with at least 60 years of data through to recent times from the PSMSL together with other long Baltic Sea sea level datasets used in the published literature.

Tide gauge operation, and the processing, distribution and archiving of the data are the responsibility of a single national agency per country, for instance the Swedish Meteorological and Hydrographical Institute (SMHI), the Danish Meteorological Institute (DMI), the Estonian Meteorological and Hydrographical Institute (EMHI), the Finnish

Table 9.1 Sources of sea level information used in the published literature organised according to the geographic regions of interest

Region	Source
North Atlantic and Europe, including Baltic Sea	Jevrejeva et al. (2005), Barbosa et al. (2008)
Baltic Sea basin wide	Omstedt and Nyberg (1991), Heyen et al. (1996), Carlsson (1997, 1998a, 1998b), Liebsch (1997), Janssen (2002), Baerens et al. (2003), Meier et al. (2004), Novotny et al. (2006), Hünicke and Zorita (2006, 2007, 2008, 2011), Barbosa (2008), Hünicke et al. (2008), Ekman (2009) and references therein; Hünicke (2010), Donner et al. (2012)
Southern Baltic Sea coast	Richter et al. (2007, 2011)
Lithuania	Dailidienė et al. (2004, 2005, 2006, 2011, 2012), Jarmalavicius et al. (2007)
Russia	Bogdanov et al. (2000), Averkiev and Klevannyy (2010), Navrotskaya and Chubarenko (2012)
Estonia	Suursaar et al. (2002, 2006a, 2010), Suursaar and Kullas (2006, 2009b), Suursaar and Sooäär (2007), Suursaar (2010), Tõnisson et al. (2011)
Poland	Pruszak and Zawadzka (2005, 2008), Richter et al. (2007, 2011)
Germany	Jensen and Töppe (1986), Liebsch (1997), Dietrich and Liebsch (2000), Liebsch et al. (2000, 2002), Jensen and Mudersbach (2004), Richter et al. (2007, 2011), Lampe et al. (2010), Dailidienė et al. (2011)
Denmark	Duun-Christensen (1990), Hansen (2007), Knudsen and Vognsen (2010), Barbosa and Madsen (2011), Hansen et al. (2011), Madsen (2011)
Sweden	Gustafsson and Andersson (2001), Kauker and Meier (2003), Omstedt et al. (2004), Chen and Omstedt (2005), Hagen and Feistel (2005), Madsen et al. (2007), Hammarklint (2009), Ekman (2009) and references therein
Finland	Johansson et al. (2001, 2003, 2004)
Gulf of Bothnia	Lisitzin (1973)

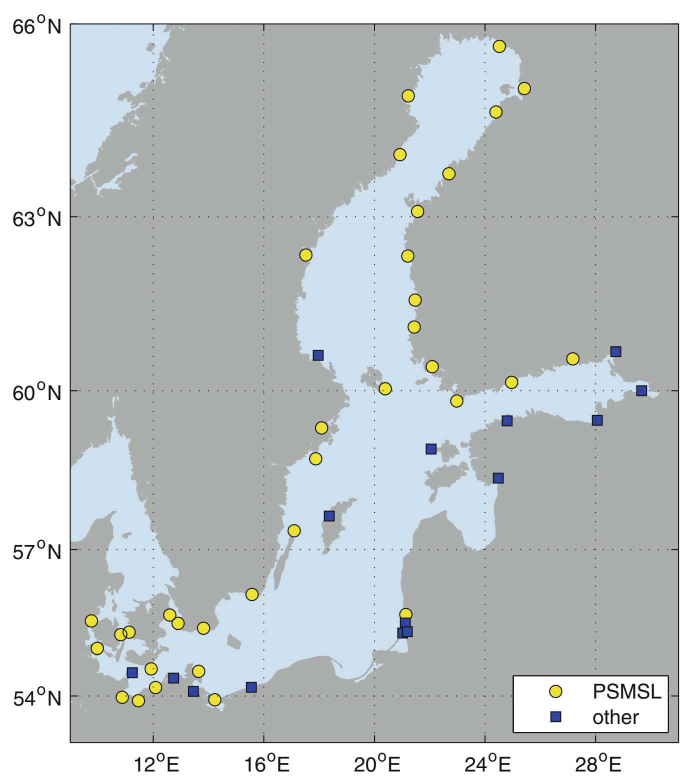


Fig. 9.1 Location of sea level records for the Baltic Sea containing at least 60 years of data through to recent times, from the Permanent Service for Mean Sea Level (PSMSL) and other long sea level datasets for this region used in the published literature (see also Table 9.1)

Meteorological Institute (FMI) and the Wasser-und Schifffahrtsamt (WSA) in Germany. Because some agencies do not supply their data to the PSMSL database, there is no basinwide freely accessible network that combines all long sea level records from Baltic Sea tide gauges.

For some applications, such as the study of extreme sea levels and storm surges, higher frequency data (more frequent than monthly average data) are needed. Some high-frequency data are available from the PSMSL/British Oceanographic Data Centre (BODC) and the University of Hawaii Sea Level Center (UHSLC, http://uhslc.soest.hawaii.edu/). However, most of the historic data are generally only available through the national agencies responsible for their collection. Real-time observations are available through the Baltic Operational Oceanographic System (BOOS, www.boos.org).

Tide gauge-derived sea level records are based on local observations of RSL presenting the position of sea level with respect to land. Thus, these data include changes in ASL as well as vertical crustal movements, which can be of different origin and take various forms. In the Baltic Sea region, the most obvious phenomenon is the long-term and more or less constant change in the Earth's crust caused by the GIA (e.g. Milne et al. 2001).

The GIA evolves on millennial timescales and so can be assumed to be approximately linear on timescales of a few centuries. However, other short-term geodynamic phenomena can lead to a contamination of tide gauge-derived sea level records. For example, sinking of piers due to unstable foundations or land sinking due to groundwater extraction, etc., which occur on timescales of years and decades, must be taken into account in the analysis of tide gauge data.

Owing to uncertainty in estimating the strong GIA signal in tide gauge records (measured relative to land), the dense network of Baltic Sea sea level observations is not generally used in the estimation of linear trends in global average sea level. For example, Baltic Sea tide gauge data are excluded from some well-known reconstructions of GMSL (e.g. Jevrejeva et al. 2006). The authors justified this exclusion by arguing that the trend in Baltic Sea sea level is strongly influenced by limited water exchange with the North Sea which leads to sea level rate curves being dissimilar from those for other regions. The strong GIA signal in the Baltic Sea region means that the RLR records must be carefully corrected before they can be used to estimate climate-driven trends in sea level.

9.2.2 Satellite Altimetry and GPS Measurements

Satellite altimetry enables continuous and spatially resolved measurements over the global ocean of SSH relative to a reference geopotential ellipsoid. The use of satellite radar altimeters to measure global SSH began in 1978 with a measurement accuracy of tens of metres. More recent high-quality satellite altimeter missions started with TOPEX/Poseidon, launched in 1992, and were continued with Jason-1 and Jason-2. These satellites were specifically designed to measure SSH with a space-time point accuracy of a few centimetres. Since then, satellite altimetry has been an independent source of SSH measurements of the open ocean, allowing for more accurate estimates of globally averaged and large-scale sea level change (Cazenave et al. 2008) than point tide gauge data. The development of satellite radar altimetry techniques, complemented in 2002 by satellite measurements of the temporal development of the gravity field, made precise quasi-global and near-continuous measurements of SSH available for the study of sea level variability and change (Woodworth et al. 2011). These data have been used extensively to map global sea level change in recent years, verifying the suggestion (based on tide gauge records) that sea level change is not spatially uniform (e.g. Cazenave and Lovell 2010, see also Fig. 9.2).

Satellite altimeter holdings are usually available as 'along track' datasets (with a time series of heights at a number of latitude grid points for each satellite pass) or as 'gridded' datasets (mapped and gridded, for example on a 1 × 1° resolution). The spatial coverage of these monthly datasets is limited to the latitude band between 65°S and 65°N due to satellite orbit constraints. Several satellite datasets combining subsequent satellite missions, thus providing time-continuous records, are freely available from different sources (e.g. www.aviso.oceanobs.com, http://sealevel.colerado.edu, www.cmar.csiro.au/sealevel). These datasets may include several corrections, such as the barometer correction (due to the air pressure at the sea surface), seasonal cycles (annual + semi-annual) or GIA correction (due to the effect of the crust and mantle adjustment on the geopotential ellipsoid). While interest in global sea level studies often concerns ASL variability, satellite data must be treated with great care for regional purposes. For instance, the comparison of tide gauge data with data derived from satellite altimetry in the Baltic Sea region requires several corrections, such as for the GIA-related strong vertical movement of land and for the GIA-related change in the geoid. Furthermore, atmospheric pressure also influences sea level. Whether or not this latter effect should be included in the corrections may for instance depend on whether the objective of study is to determine long-term trends due to increased ocean temperature (correction required), or comparison to tide gauge data (correction not required) (Barbosa 2011). For example, Fernandes et al. (2006) showed that a specific atmospheric correction applied to altimetry observations can significantly affect the resulting sea level information, as the atmospheric pressure varies geographically and covers a wide range of spatial and temporal scales. Thus,

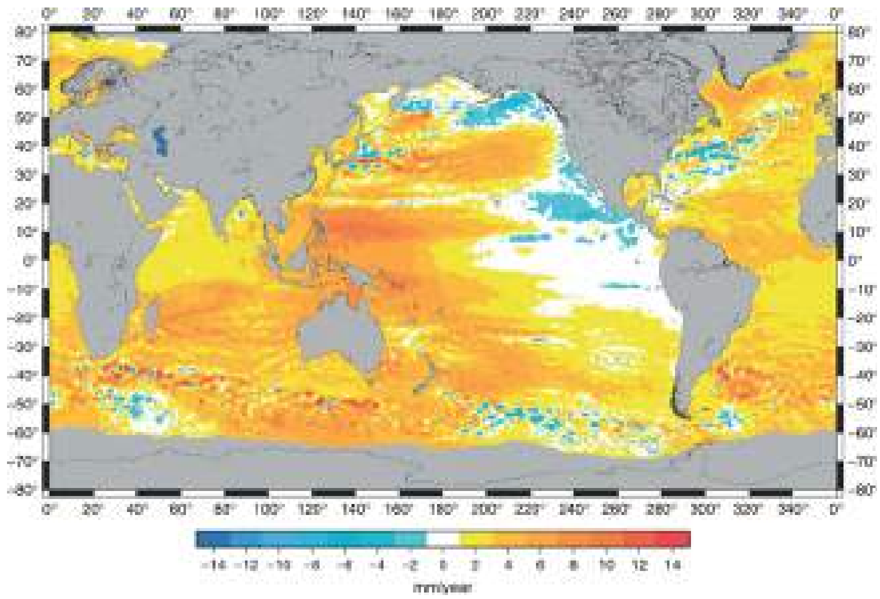


Fig. 9.2 Spatial trends in sea level rise between January 1993 and October 2010 computed from the multi-mission (Topex/Poseidon, Jason-1, Jason-2, ERS, and Envisat satellites) gridded sea level products available from the CLS/AVISO Website at weekly intervals (Cazenave and Remy 2011)

the application of these corrections strongly depends on the research question.

In contrast to global studies, only a few studies for the Baltic Sea region have so far (to 2012) used satellite datasets together with tide gauge readings to address different research questions (e.g. Liebsch et al. 2002; Novotny et al. 2005, 2006; Madsen et al. 2007; Hünicke and Zorita 2011; Madsen 2011).

There are two other factors to consider when using satellite altimetry products in the near-coastal zone, in general and specifically for comparison with tide gauge observations. First, some of the applied corrections to the raw satellite data are not valid in the near-coastal zone, typically 50 km from any coast or island (Madsen et al. 2007). The most important is the wet tropospheric correction (Obligis et al. 2011): the altimetry estimation from the satellite radar travel-time is routinely recalibrated for the effect of changing tropospheric humidity, which is simultaneously retrieved by an on-board microwave sensor. This retrieval is erroneous in coastal regions and must be corrected. Second, the gridded altimetry products are interpolated in space and time, and across missing data. Interpolation should be undertaken with great care in the coastal zone, where the variability in space and time is much greater than in the open ocean. Some products include many interpolated data, while others permanently mask out areas which may have periods with invalid measurements, for instance in the case of seasonal ice cover. Along-track data with customisable processing are available from the Radar Altimeter Database System (RADS) database (Naeije et al. 2008; http://rads.tudelft.nl/).

Another important feature of the satellite era is the application of the Global Positioning System (GPS) to measure continuously the vertical land movements (stated hereafter also as station velocities). Before the satellite era, change in sea level could only be assessed relative to points on land. Permanent GPS observations since the 1990s have accumulated sufficiently long time series to allow determination of the isostatic uplift (that is, the GIA). Uncertainty in the determination of the station vertical velocity may be strongly dependent on the station and on the model assumed for the measurement noise. Vestøl (2006) indicated an accuracy of between ±0.4 mm year^{-1} and ±0.5 mm year^{-1} for Fennoscandia, Lidberg et al. (2007, 2010) between 0.05 and 1.96 mm year^{-1} for Fennoscandia, and Richter et al. (2011) between ±1.4 and ±2.2 mm year^{-1} for the southern Baltic Sea. In the latter study, it became evident that the determination of reliable station velocities in the southern Baltic Sea poses a challenge (as the uncertainties exceed the small rates of land movement themselves; see Sect. 9.3.1.2 for more information).

GPS measurements of vertical velocities within the Baltic Sea area are collected through different networks and projects, such as the BIFROST (Baseline Inferences for Fennoscandian Rebound Observations, Sea level and Tectonics) and the European Terrestrial Reference System (EUREF) Permanent Network (EPN) (EUREF 2011) networks as well as by the Satellite Positioning Service of the German State Survey (SAPOS). Good coverage exists for Germany, Denmark, Sweden and Finland, while coverage in other countries around the Baltic Sea is more limited (Lidberg et al. 2007; Knudsen and Vognsen 2010; Richter et al. 2011). The BIFROST network started in 1993 and comprises the permanent GPS network SWEPOS (SWEPOS™, SWEPOS 2011) in Sweden and FinnRef (FinnRef®, FGI 2011) in Finland. There is also a GPS station network in Norway (SATREF®, SATREF 2011). A combined study of

GPS stations relevant for the GIA process in Fennoscandia was reported by Lidberg et al. (2010).

The advantage of GPS measurements is that they measure rates of vertical land movement, whether or not they are due to GIA or to other geological processes. Their disadvantage lies in their relatively short time span (since the 1990s). The determination of vertical crustal deformation rates finds application for the calculation of trends in ASL (see Sect. 9.3.2.1). Analysis of the GPS data in the Baltic Sea area yields a spatial pattern that demonstrates that ongoing crustal deformation in Fennoscandia is dominated by GIA.

9.3 Change in Mean Sea Level

9.3.1 Main Factors Driving Sea Level Change

The overall change in MSL at the Baltic Sea coast is the combined result of, on the one hand, factors acting at large spatial scales, such as the postglacial rebound and the change in global ocean mass, and on the other, the contribution of regional and local oceanographic and atmospheric factors. Sea level within the Baltic Sea can deviate markedly from that of the North Sea immediately outside the Danish Straits (Madsen 2011).

9.3.1.1 Large-Scale Factors

Long-term trends in GMSL for the past century have been mainly determined by the thermosteric contribution, that is, the thermal expansion of the sea water with rising ocean temperatures. Although the global mean trend is positive, there is strong regional variability due to the large-scale ocean circulation that modulates heat uptake by the ocean. Another large-scale contribution to the rise in GMSL has been the melting of land ice (glaciers and the polar ice sheets), although there is considerable uncertainty about the net contribution from the Antarctic Ice Sheet over recent decades (Cazenave and Remy 2011). This uncertainty originates in determining the net mass balance between melting, calving and precipitation over the ice sheet. Climate model simulations indicate that rising temperatures may have caused an increase in precipitation over polar ice sheets. Observations, however, do not show a trend in precipitation, but in these areas, they are sparse and do not have good spatial coverage. A third large-scale contribution to the rise in GMSL is land water storage and the use of subsurface water. The former limits sea level rise because some of the water that would normally flow into the ocean remains stored on land; while the latter increases sea level because it mobilises water into the ocean that would otherwise remain on continental areas.

Long tide gauge observations over the twentieth century indicate that GMSL has risen on average by 1.7 mm year^{-1} (Bindoff et al. 2007). Tide gauges are located at the coast and are not fully representative of the ocean regions. Satellite altimetry data for the global ocean within the latitude band between 65°N and 65°S (see Sect. 9.2.2) indicate an average sea level rise for 1993–2007 of 3.1 mm year^{-1} (±0.4 mm year^{-1}) with smaller values since 2003 of 2.5 mm year^{-1} (Cazenave et al. 2009). For more detailed information about the global sea level budget, see Chap. 14.

9.3.1.2 Regional and Local Factors

Land Movement

The Baltic Sea is a region strongly influenced by GIA, which results even today in a maximum uplift of the Earth's crust in the Gulf of Bothnia of roughly 10 mm year^{-1} (resulting in a negative trend in RSL) (e.g. Hammarklint 2009) and subsidence in parts of the southern Baltic Sea coast of about 1 mm year^{-1} (e.g. Richter et al. 2011). The local pattern of land movement may be modified by local natural or man-made land uplift or subsidence (see Sect. 9.2).

Nowadays, different methods exist to study land movement effects. Traditionally, land uplift rates in the Baltic Sea region have been determined from sea level measurements (e.g. Vermeer et al. 1988). They are based on local (e.g. tide gauge) observations of long-term trends in RSL, present the position of sea level with respect to land and, therefore, include both the signal related to vertical crustal movements and to changes in ASL. The first consistent map of the postglacial uplift of Fennoscandia was constructed by Ekman (1996), mainly on the basis of RSL records reduced to the common 100-year period 1892–1991, but also taking into account lake-level records and repeated high precision levelling. Two years later, Ekman (1998) presented an updated map of absolute vertical velocities (based on the results of Ekman 1996) by applying a GMSL rise of 1.2 mm year^{-1}, deriving the rise of the geoid (relative to the ellipsoid) from computations of Ekman and Mäkinen (1996). In 2007, Rosentau et al. (2007) augmented the map of Ekman (1996) with data from tide gauge measurements from the southern Baltic Sea (e.g. from Dietrich and Liebsch 2000), and Ågren and Svensson (2007) presented an apparent land uplift map based on a combination of sea level results, GPS results and repeated levelling.

Hansen et al. (2011) obtained an alternative estimate of the land uplift in the south-western Baltic Sea region based on tide gauge data and their determination of an ASL curve at the island of Læsø in Kattegat. After grouping the tide gauge stations, this study indicates a jump in the land rise coefficients across southern Denmark (not seen in other studies including precise levelling campaigns).

In Estonian studies, the applied land uplift rates are still based on a map compiled on the basis of national high precision levelling (1933–1943, 1956–1970, 1977–1985) by Vallner et al. (1988). Along the Estonian coast, uplift rates vary between 0.5 and 2.8 mm year^{-1}. According to Suursaar et al. (2006a), the accuracy of these rates, compared to the map of Ekman (1996) of the postglacial uplift of Fennoscandia, lies in the range ±0.4 mm year^{-1}.

Lithuanian sea level studies do not consider any land movement effects in their analysis of sea level variability because 'researchers recommend different estimates of land sinking in the Lithuanian Region' (0–2 mm year^{-1}) (Dailidienė et al. 2006).

Land movement due to the GIA effect can be calculated with an ice load model (e.g. Peltier 1998). This estimation does not consider other land motions (e.g. sinking piers or short-term motion due to earthquakes). Milne et al. (2001) considered the GIA correction for the Baltic tide gauge records as critical because of the large GIA amplitude ($\sim$10 mm year^{-1}) relative to the global trend ($\sim$1–3 mm year^{-1}).

However, recently, new geodetic techniques such as the application of GPS (see Sect. 9.2.2) allow for measurements of more precise absolute rates of vertical land movements and lead to considerable progress by comparison with those derived from tide gauges, ice load models or corrections from geological data. Vertical land movements (station velocities) based on observations at permanent GPS stations in Sweden and Finland (1993–2000) were presented by Johansson et al. (2002), Scherneck et al. (2002) and used by Milne et al. (2001, 2004), for example for the estimation of regional sea level rise, together with Fennoscandian tide gauge records. An update of station velocities was presented by Lidberg et al. (2007) for the period 1996 to mid-2004, including some additional sites in Norway, Denmark and northern Europe, and showing much smaller uncertainties in station velocity compared to Johansson et al. (2002). An overview of published station velocities was provided by Lidberg et al. (2010). Figure 9.3 shows the vertical velocity rates for Scandinavia from the analysis of a regional GPS network by Richter et al. (2011). The application of these rates to the presented map of RSL changes in Fig. 9.7 (left) can be used to estimate ASL changes in the Baltic Sea region.

Hill et al. (2010) introduced a new technique to combine geodetic observations and models for GIA fields in

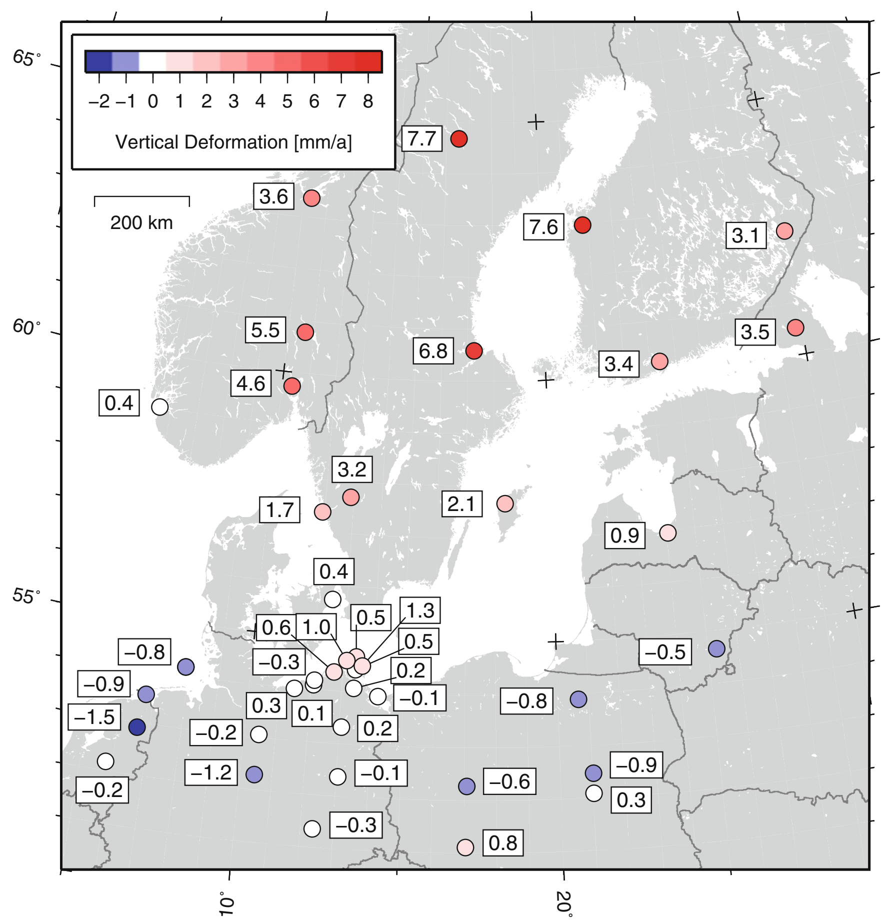


Fig. 9.3 Estimation of vertical velocities (crustal deformation rates) in the Baltic Sea region (for tide gauge correction) (Richter et al. 2011)

Fennoscandia by assimilating GPS-derived and tide gauge data together with gravity rates from the Gravity Recovery and Climate Experiment (GRACE) into their model approach (see also Chap. 14). The updated model shows a spatial pattern and magnitude of peak uplift (9.5 ±0.4 mm $year^{-1}$) that is relatively consistent with previous findings (e.g. Johansson et al. 2002; Milne et al. 2004), but with a different—more easterly—peak uplift location in the middle of the northern Gulf of Bothnia. Previous studies (e.g. Milne et al. 2001; Johansson et al. 2002) indicated the peak uplift slightly to the west of the northern Gulf of Bothnia.

Meteorological Influence

In addition to the long-term global sea level rise and land movements, sea level in the semi-enclosed Baltic Sea is modulated by meteorological factors, especially wind forcing. Winds play a key role, as persistent winds from the south-west or north-east transport water into or out of the Baltic Sea, respectively, thereby raising or lowering Baltic Sea sea level as a whole. According to Ekman (2007), temporary winds redistribute the water within the Baltic Sea, producing high or low sea levels at the ends of the basin depending on wind direction.

Sea level variations in the Baltic Sea at interannual to decadal timescales are strongly influenced by the strength of westerly winds, closely related to the North Atlantic Oscillation (NAO). The NAO represents the large-scale circulation over the north-west Atlantic (see Chap. 4, Box 4.1 for detailed information). A positive NAO (strong Azores High and Icelandic Low) is associated with warm and humid winters and strong westerly winds in the Baltic Sea area, which causes sea level to rise; a negative NAO (weak Azores High and Icelandic Low) is associated with cold and dry winters which causes sea level to fall. The correlation between the NAO and Baltic Sea sea level is especially strong in winter (e.g. Andersson 2002; Dailidienė et al. 2006; Hünicke and Zorita 2006; Suursaar and Sooäär 2007), and in northern and eastern parts of the Baltic Sea (Johansson et al. 2004; Suursaar et al. 2006a) (see Fig. 9.4, right panel). However, in the southern Baltic Sea, the correlation between the winter NAO index and winter sea level is low (Hünicke and Zorita 2006; Jevrejeva et al. 2005). In addition, in the Baltic Sea basin as a whole, the strength of the correlation between the winter NAO index and winter MSL changes significantly over time (Fig. 9.4, left panel; also shown by Jevrejeva et al. 2005).

Although the NAO is the dominant sea level pressure (SLP) large-scale pattern of the North Atlantic in wintertime, it explains (on average) only 32 % of the total variability of sea level at interannual timescales (e.g. Kauker and Meier 2003; Jevrejeva et al. 2006). Thus, estimation of the full amount of variability that can be explained by the atmospheric circulation (and not only by the NAO) must consider the whole SLP field of the North Atlantic–West European region. Such an analysis was presented for the first time for the Baltic Sea by Heyen et al. (1996).

In a more recent study, Hünicke et al. (2008) statistically analysed the influence of different atmospheric forcing factors (temperature, SLP and precipitation) and found that decadal sea level change in the Baltic Sea basin varies geographically. In the central and eastern parts of the Baltic Sea,

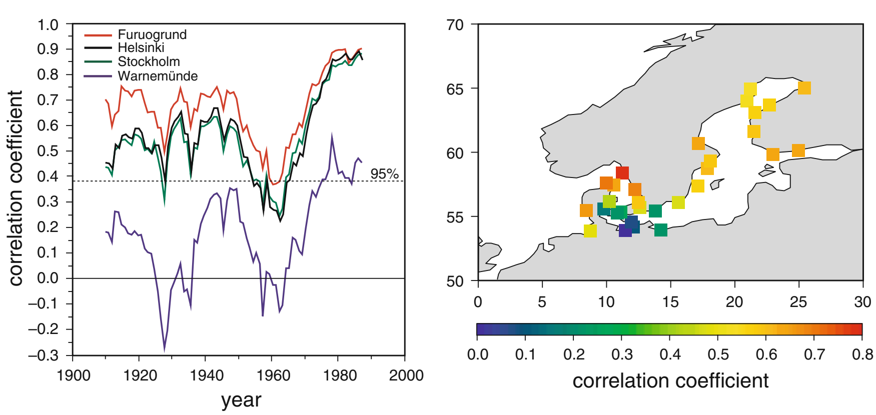


Fig. 9.4 *Left* Sliding correlation (21-year window) between four sea level stations that should be representative of the behaviour of sea level in the Baltic Sea and the winter NAO index (DJF); *Right* Correlation between winter means of the **NAO** index and winter mean (linearly detrended) Baltic Sea sea level 1900–1998 (redrawn from Hünicke and Zorita 2006)

winter sea level variations at decadal timescales are explained well by SLP variations (and not only the NAO). For the southern part, area-averaged precipitation seems to better explain statistically the decadal variations. Thus, depending on future climate trends (see Chaps. 11–14), a different behaviour of future sea level trends in southern and eastern and northern parts of the Baltic Sea might be possible.

On the other hand, Ekman (2009) offered a different, purely dynamic, explanation for the observed correlation between SLP and the southern Baltic (low correlation) and northern Baltic (high correlation). This explanation involves the basinwide response of SSH to sustained westerly winds (long-term response). The variations should be greater away from the connection to the North Sea and smaller closer to it. However, questions remain, such as the instability of the correlation with the NAO over time and the long-term correlation between southern Baltic Sea sea level and precipitation, which is evident in low-pass filtered decadal time series.

Omstedt et al. (2004) investigated variability and trends in observed time series for the Baltic Sea region from the past two centuries and found pronounced positive trends in Stockholm sea level variations and anti-cyclonic circulations. For the period 1985–2000, trends in air temperature and sea level anomalies were found to be positive, lying outside the range of normal variation for the past 200 years (see also Sect. 9.3.2.1).

Getzlaff et al. (2011) observed an offset in detrended monthly MSL anomalies (taken from PSMSL) for the periods 1970–1988 and 1989–2008 and associated it with a shift in strong wind events and with the change in prevailing wind directions (as a result of an increased positive NAO) (Lehmann et al. 2011). The shift was found to occur uni-directionally over the whole Baltic Sea area, which suggests a change in the prevailing westerly wind situation controlling MSL variations. The results agree with findings of a numerical model simulation and confirm findings of recent studies of historic sea level time series (e.g. Johansson et al. 2004).

Dailidienė et al. (2011) studied water level and surface temperatures of coastal lagoons along the southern and south-eastern shores of the Baltic Sea (Curonian and Vistula Lagoons and Darß-Zingst Bodden chain), relating the observed increase in water level (greatest in the east) and water temperature since the 1980s to changes in the NAO.

9.3.2 Variations Within the Observational Period (Past 200 Years)

Within the Baltic Sea, the trends in coastal sea level display a strong influence of the isostatic dynamics following the last deglaciation. Thus, RSL is falling in the northern Baltic Sea and rising in the southern Baltic Sea, due to the combined effect of crustal deformation and volumetric sea level changes. Superimposed on the long-term trends that affect RSL, many climate factors, such as changes in water density, changes in the total volume of the Baltic Sea and currents, can modulate ASL within the whole or parts of the Baltic Sea basin. Sea level may thus display considerable variability on timescales ranging from minutes through the annual cycle, to decades.

9.3.2.1 Long-Term Trends and Decadal Variations

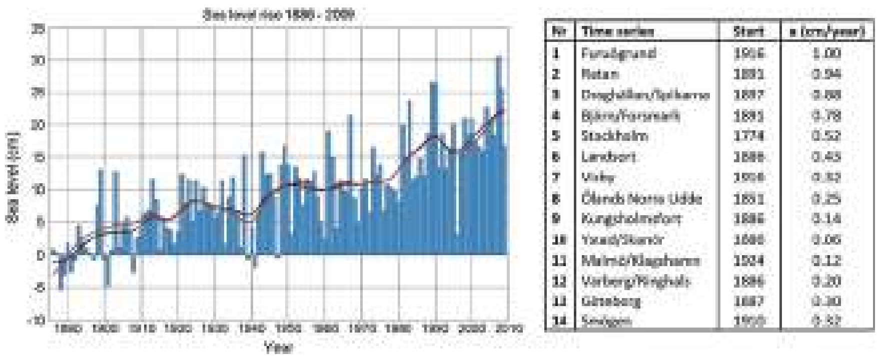
Absolute Sea Level

Some studies have analysed observations from permanent GPS stations to determine crustal deformation rates and used these and repeated levelling data to derive ASL changes from tide gauges. For instance, Milne et al. (2001) estimated a regional ASL rise of 2.1 ± 0.3 mm $year^{-1}$ using annual tide gauge means from records (provided by PSMSL) at 20 sites in Fennoscandia spanning 35 years or more, and using only data acquired during or after 1930. Hill et al. (2010) combined geodetic observations and models and estimated a spatially averaged rate of ASL change of around 1.5 mm $year^{-1}$, based on monthly RLR tide gauge means (from PSMSL) from records at 40 sites in Fennoscandia that provide more than 40 years of data. Richter et al. (2011) derived a mean ASL value of around 1.3 mm $year^{-1}$ for 13 tide gauges for which a GPS station is located nearby. Nine of these gauges were sites on the southern Baltic Sea coast, spanning almost 200 years, and complemented by monthly tide gauge means from 51 (basinwide) PSMSL sites that report more than 60 years of data in the period 1908–2007.

Additionally, regional focused studies that combined geodetic and tide gauge data information found ASL values of around 1.3 mm $year^{-1}$ for the Scandinavian coast (1891–1990, Vestøl 2006), 1.8 mm $year^{-1}$ for Danish sea level stations (1900–2000, Knudsen and Vognsen 2010) and 1.5 mm $year^{-1}$ for Swedish tide gauge stations (1886–2009, Hammarklint 2009, see Fig. 9.5). Ekman (2009) indicated a rate of climate-driven sea level rise in Stockholm of about 1 mm $year^{-1}$. For more detailed information, see the original references.

When the reported uncertainty is considered, the results of all the previously cited studies lie within the error bars of the rate of GMSL rise of 1.7 ± 0.5 mm $year^{-1}$ reported in the Fourth Assessment of the Intergovernmental Panel on Climate Change (Bindoff et al. 2007). However, it should be remembered that regional long-term trends in sea level can deviate substantially from the global mean.

This was also demonstrated by Madsen (2011), who estimated linear sea level trends from a gridded satellite altimetry dataset (1992–2008, multi-satellite open ocean



Nr	Time series	Start	a (cm/year)
1	Furuögrund	1916	1.00
2	Ratan	1891	0.94
3	Draghällan/Spikarna	1897	0.88
4	Björn/Forsmark	1891	0.78
5	Stockholm	1774	0.52
6	Landsort	1886	0.43
7	Visby	1916	0.32
8	Ölands Norra Udde	1851	0.25
9	Kungsholmsfort	1886	0.14
10	Ystad/Skanör	1886	0.06
11	Malmö/Klagshamn	1924	0.12
12	Varberg/Ringhals	1886	0.20
13	Göteborg	1887	0.30
14	Smögen	1910	0.32

Fig. 9.5 Annual sea level means averaged for 14 Swedish sea level records corrected for land uplift (shown in the right table for each location) and compared to the 1886 level. *Black line* time-filtered version together with the filtered Stockholm sea level time series (*red line*) (Hammarklint 2009)

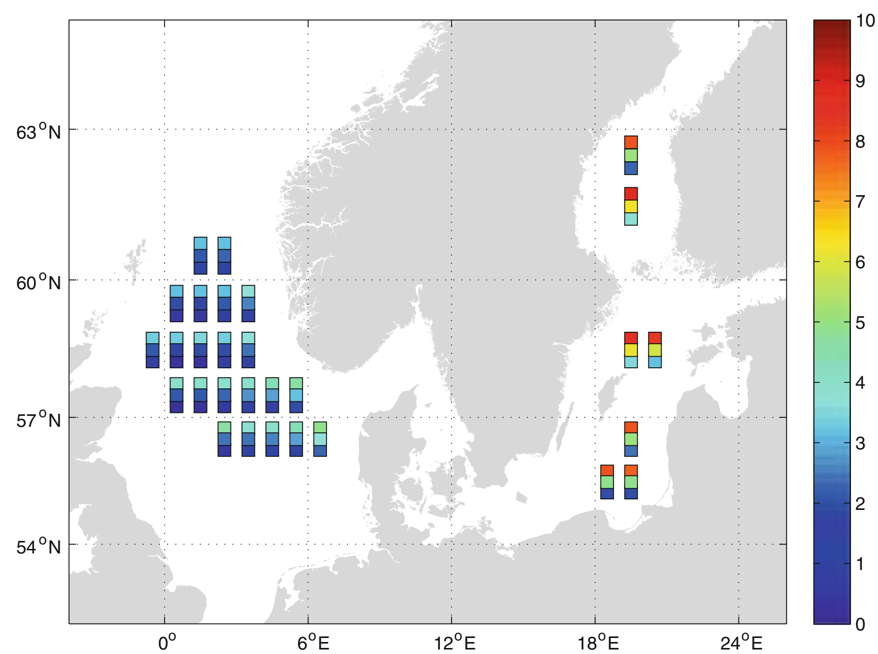
product with 1-degree, 10-day resolution). This resulted in a best estimate mean trend of 5.8 mm year^{-1} for the Baltic Sea area, but with a high uncertainty in the trends of a few millimetres per year (Fig. 9.6). For comparison, the global mean value (3.23 ± 0.04 mm year^{-1}) and the Atlantic mean (3.3 ± 0.1 mm year^{-1}) were also calculated for the same dataset and time period. Madsen (2011) concluded that the uncertainty in the estimation of regional trends is much greater, both because of the smaller number of samples and, especially for the Baltic Sea, because of the large natural variability. This also means that the trends are sensitive to the choice of time period. With these reservations, it can be seen that the North Sea trend is likely to be lower than the global mean and that the Baltic Sea trend is very likely to be higher than the global mean and perhaps just as interesting for the Danish area, higher than the North Sea trend.

Relative Sea Level

In recent years, changes in Baltic Sea RSL, as the most important factor in impact studies, have been the subject of a wider range of publications, mainly on the regional (e.g. national) to local (e.g. only one tide gauge) level. However, a comparison of the different results is hampered by different observation periods (and the treatment of data gaps therein) and analysis techniques. Thus, the cited RSL trend values that follow must be treated with great care.

Navrotskaya and Chubarenko (2012) studied annual (and extreme) sea levels in the Russian sector of the Vistula Lagoon situated in the south-eastern part of the Baltic Sea. They found annual mean linear RSL trends in the range 1.7–1.9 mm year^{-1} (Baltiisk, 1860–2006; Kaliningrad, 1901–2006) to 3.6–3.7 mm year^{-1} (Kaliningrad, Baltiisk, Krasnoflotskoe, 1959–2006). On the sea coast (Pionerskii,

Fig. 9.6 Linear trends in sea surface height (mm year^{-1}, *centre boxes*) and 90 % confidence intervals (*upper and lower boxes*) as calculated from satellite altimetry (Madsen 2011)



1959–2006), the trend in the latter half of the twentieth century amounted to 2.6 mm year^{-1} only.

To the authors' knowledge, no published sea level studies are available for the Latvian coasts, but linear trend estimations at two sites (Liepaja 1865–1936 and Daugavgriva 1872–1938) are available through PSMSL.

For the Estonian coastline, Suursaar and Kullas (2009b) estimated linear RSL trends of 1.5 mm year^{-1} (Pärnu 1924–2008), 1.8 mm year^{-1} (Tallinn, 1899–2005) and 0.5 mm year^{-1} (Narva-J. 1899–2008).

Dailidienė et al. (2006, 2012) estimated linear annual mean RSL long-term trends for Lithuanian tide gauges, ranging from 1.3 mm year^{-1} for the centennial trend (Klaipeda Strait 1998–2002; situated on the open Baltic Sea coast) to 2.4–2.9 mm year^{-1} for sea level stations reporting between 1961 and 2002 (Juodkranté, Vida, Venté; situated in the Curonian Lagoon).

Figure 9.7 displays maps of the secular (100 years) RSL changes compiled by Richter et al. (2011) based on long tide gauge measurements of the Baltic Sea region based on data from PSMSL (including Polish, German, Danish, Swedish and Finnish tide gauges) and data compiled on the basis of historic documents (at German and Polish sites).

The pattern in RSL trends (Fig. 9.7, left panel) shows a clear north–south gradient within the Baltic Sea region, reflecting the crustal deformation due to the GIA effect. In the northern part, stations indicate large negative RSL trends. A maximum rate of 8.2 mm year^{-1} is found in the Gulf of Bothnia, in the area of predicted maximum GIA-induced crustal uplift (e.g. Peltier 2004). Tide gauge measurements along the southern Baltic Sea coast yield positive rates of around 1 mm year^{-1}, which implies a rising sea level relative to the Earth's crust. However, the pattern of trends over the southern region is not uniform and displays a clear gradient in the north-easterly direction (Fig. 9.7, lower right panel).

The spatial pattern in long-term trends in sea level should, in general, also reflect the spatially varying fingerprints of sea level change due to recent mass changes in the polar ice sheets and, accordingly, the change in the gravity field of the Earth's surface (Tamisiea et al. 2003; Milne et al. 2009). As demonstrated by Mitrovica et al. (2001), the pattern caused by the melting of the Greenland Ice Sheet is negligible for the Baltic Sea region, as its zero-line intersects this region, and the remaining variations are below the accuracy of the derived RSL rates, ranging between 0.1 and 0.3 mm year^{-1}. On the other hand, the pattern caused by the melting of the Antarctic Ice Sheet does affect the Baltic Sea region, but can be expected to be nearly constant over the entire region (see also Chap. 14). There exists, however, considerable uncertainty about the sign of the mass balance of the Antarctic Ice Sheet in the twentieth century, due to the competing effects of melting, calving and precipitation.

Richter et al. (2011) also analysed the variation in the annual mean RSL at long tide gauge records, such as the Polish tide gauges Swinoujscie and Kolobrzeg (Fig. 9.7, upper right panel). Both time series show consistent behaviour with a slight negative trend throughout the first decades of the time series up until 1860, followed by an increasing trend of around 1 mm year^{-1}. The authors suggested, as a possible explanation for this trend, climatic effects related to the Little Ice Age. This is consistent with Ekman (2009 and references therein) for the GIA-corrected ASL trend of 1.01 mm year^{-1} for the Stockholm time series. (Figure 9.7 includes, for illustration, the RSL Stockholm series that displays a strong negative trend due to the GIA). However, it should be remembered that due to decadal

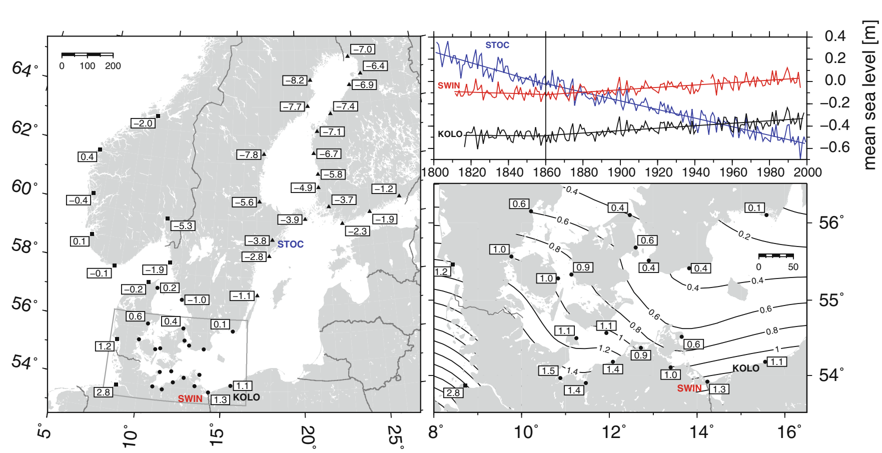


Fig. 9.7 Maps of secular (100 years) RSL changes, based on tide gauge measurements of the entire Baltic Sea region (*left panel*) and, in more detail, the southern Baltic Sea coast (*right panel below*) together with changes in the linear trend of the (arbitrarily shifted) annual RSLs at Stockholm, Swinoujscie (SWIN) and Kolobrzeg (KOL) between the period before and after 1860. The *symbols* represent the affiliation to different reference stations (*dots* warnemünde, *triangles* Stockholm, *squares* Smögen) (redrawn from Richter et al. 2011)

variations in the RSL trend, a comparable determination of secular RSL changes at different stations requires the application of identical observation periods (Richter et al. 2011) and analyses techniques.

The long-term trends of Baltic Sea RSL in the period with direct observations have been analysed by Barbosa (2008) and Donner et al. (2012). As well as estimating trends in the median sea level, they also estimated trends in the quantile of the probability distributions of monthly MSL at different gauges along the Baltic Sea coast (Fig. 9.8). The trends clearly exhibit the effect of the GIA. Interestingly, the trends in the median sea level do not always coincide with the trends in the extreme high and low quantiles.

Whereas the low quantiles of the distributions show basically the same trend as the median, the upper quantiles tend to display a more positive trend. Therefore, the higher values of RSL are increasing more rapidly, or decreasing more slowly in the regions with isostatic uplift. This happens more markedly in the northern Baltic Sea and has also been confirmed by more locally focused studies on Estonian sea level (Suursaar and Kullas 2006).

The reasons for this different behaviour are not clear, and many factors such as the atmospheric circulation could contribute. In the Estonian case, due to the form of the coastlines, the fingerprint of forcing by the atmospheric circulation at interannual timescales is clearly detectable. Therefore, a reasonable conjecture is that the atmospheric circulation, in particular the NAO, may also be involved in the long-term trends of the upper sea level quantiles in wintertime. However, the NAO exhibits a positive trend only in the last decades of the instrumental record, and not over the whole twentieth century. Also, the NAO trend in the very last two decades has been negative (Pinto and Raible 2012). Other atmospheric circulation patterns different from the NAO may also play a role (see Sect. 9.3.1.2 *Meteorological influence*). For example, Omstedt et al. (2004) found that the frequency of anti-cyclonic situations over the Baltic Sea increased in the late twentieth century imprinting its influence on several climate variables, also on the Stockholm sea level data. The causes of linear trends in sea level, therefore, may depend on the timescale over which these trends are calculated. At decadal and multi-decadal timescales, the internal variations of the atmospheric circulation may exert a strong influence. The NAO, as previously indicated, undergoes multi-decadal phases of increase and decrease. Over longer timescales, such as the centennial scale, the influence of the atmospheric circulation is strongly averaged out and is likely to become weaker relative to other external forcing. In this respect, Ekman (2009) found that the probability distribution of monthly sea level in Stockholm

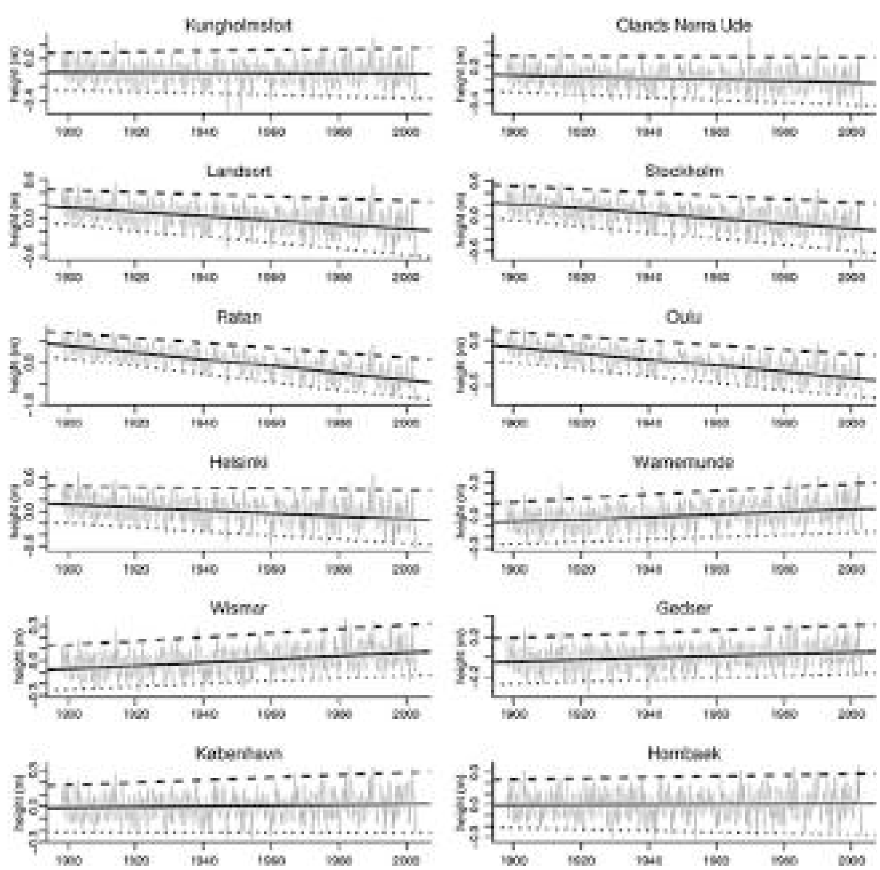


Fig. 9.8 Time series of three quantiles (median, and the 1 and 99 % quantiles) of the distribution of de-seasonalised monthly sea level at several stations of the Baltic Sea in the period 1890–2010 (Barbosa 2008)

has also undergone large changes over time, with a much narrower spread prior to 1950 than in later decades.

Baltic Sea sea level has undergone decadal variations around the quasi-linear long-term trend dominated by the GIA. These decadal variations were found to depend on the season. For example, Chen and Omstedt (2005) found for the Stockholm sea level record that all months except June and August display positive sea level trends. The authors connected the negative trends in June and August to a negative long-term trend in Baltic Sea precipitation during these months. However, they stated that these findings have still to be justified by a detailed analysis of the hydrological budget.

An illustration of decadal long-term wintertime variations in sea level for the past 200 years was presented by Hünicke et al. (2008) for four different sites (Fig. 9.9).

Ignoring the GIA-related trend, winter Baltic Sea sea level generally displays higher values around 1820, 1910 and in the most recent decade, and lower values around 1875, 1940 and 1970. However, the homogeneity of the data may be compromised at the beginning of the record. Since the decadal variations are not totally coherent through time, the precise mechanisms responsible for them have not been ascertained. The decadal variations may have been caused mainly by atmospheric circulation, but also by precipitation and variations in ocean currents.

Scotto et al. (2009) analysed the spatial coherence of long (>30 years) monthly records of RSL from 14 tide gauges in the Baltic Sea by applying cluster analysis. The coherent regions were defined in terms of the statistical properties regarding their serial auto-correlation. At timescales of up to three months, the usual Baltic Sea sub-basins emerged as individual clusters. For longer timescales, up to six months, all regions with the exception of the areas close to the North Sea displayed high coherency.

9.3.2.2 Changes in Seasonal Variability

The long-term changes in Baltic Sea sea level have been also found to depend on the season. Sea level displays an annual cycle and is generally higher during winter months and lower during spring (as demonstrated by Hünicke and Zorita 2008, see Fig. 9.10).

Changes in the annual sea level cycle were studied by Ekman and Stigebrandt (1990), Ekman (1999) for the Stockholm sea level record (covering the ninetieth and twentieth centuries). The authors associated the increase in the amplitude of the dominant annual component to changes in winter wind condition. Hünicke and Zorita (2008) statistically analysed 30 sea level records from PSMSL (covering the twentieth century) and four long historic sea level records (covering the ninetieth and twentieth centuries, including Stockholm), together with climatic datasets, to investigate centennial trends in the amplitude of the annual cycle. They detected a statistically significant increase in amplitude (winter-spring sea level) in the annual cycle in almost all sea level records investigated, but of a weaker sign than the decadal variations in the annual cycle. As the magnitude of the trends appeared almost uniform, with the exception of the Skagerrak area, they concluded that the driving mechanism for the trends in the annual cycle is very likely to be of non-regional origin. Testing several hypothesised factors to explain these centennial trends (wind through the SLP field, the barometric effect, temperature and precipitation), only precipitation remained a plausible candidate in their analysis. The physical mechanisms responsible for the long-term trend due to changes in precipitation could be related to salinity changes and their effect on changes on water density. This remains to be resolved.

Plag and Tsimplis (1999) investigated the spatial and temporal variability in the seasonal cycle of sea level for PSMSL tide gauge records in the North Sea and Baltic Sea.

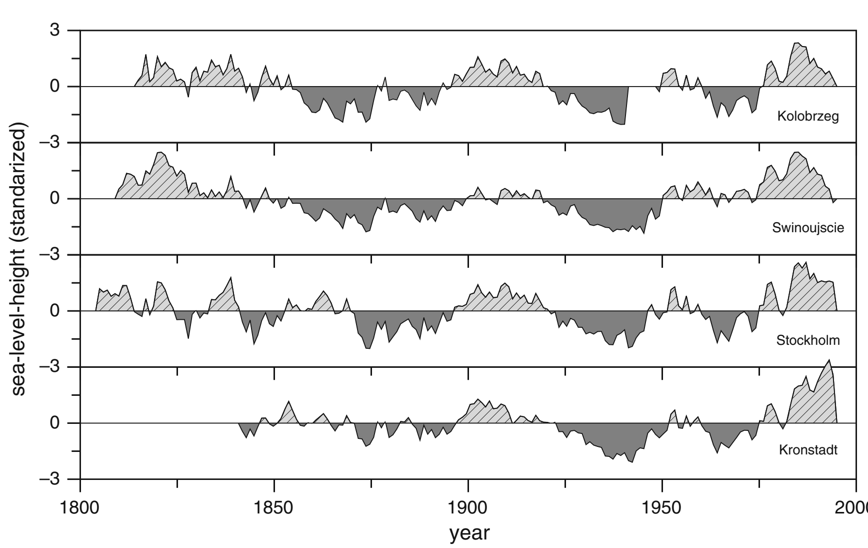


Fig. 9.9 Relative winter mean (DJF) sea level heights in the Baltic Sea. The series are smoothed by an 11-year running mean to highlight the decadal variations and are standardised by unit standard deviation. The long-term trend was eliminated (redrawn from Hünicke et al. 2008)

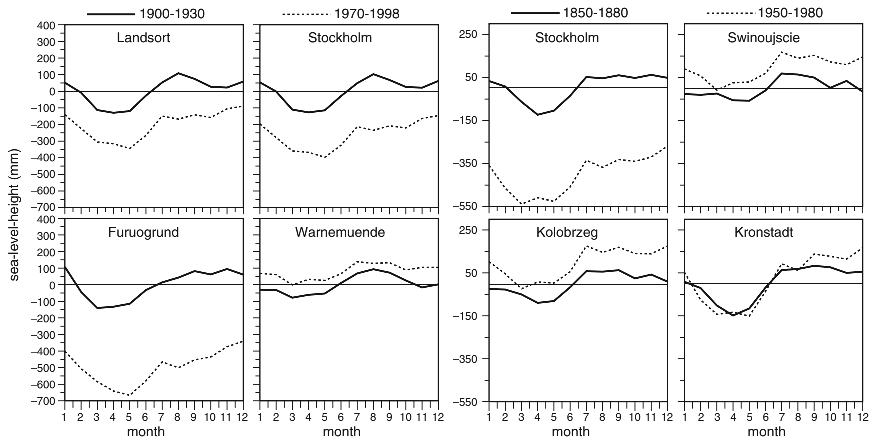


Fig. 9.10 Annual sea level cycle at various Baltic Sea sites derived from monthly means. *Left panel*: early and late twentieth century (1900–1930 *solid line*, 1970–1998 *dashed line*). *Right panel*: mid-nineteenth and mid-twentieth century (1850–1880 *solid line*, 1950–1980 *dashed line*) (redrawn from Hünicke and Zorita 2008)

They related the variability detected to changes in the position and extent of the separation zone between maritime and continental climate regions.

However, the decadal variations in the amplitude of the annual sea level cycle can be due to slow variation in the wind forcing, especially at regional scales and at locations where the geometry of the coastline is favourable for wind-driven change to become evident. For instance, in the Gulf of Riga and Väinameri area, results from a two-dimensional hydrodynamic model indicate that a relatively modest increase in wind speed (2 m s^{-1}) could be responsible for a mean increase in sea level of up to 5 cm, in addition to an analogous change in the Baltic Sea MSL (Suursaar and Kullas 2006). Consequently, a total wind-induced average sea level rise of 7–10 cm could occur at locations like Pärnu and Matsalu, Estonia. In these areas, changes of a similar magnitude probably occurred between 1950 and 1990 (Suursaar and Kullas 2006).

9.3.2.3 Is Sea Level Rise Within the Baltic Sea Accelerating?

A relevant question in the context of climate change is, whether the rise in global sea level is accelerating? Projections of the future magnitude of global sea level rise are still very uncertain, mainly due to the difficulty in estimating the effects of warming on the dynamics of the polar ice sheets. Also, there is a wide spread in the projections of the future thermosteric effect simulated by the suite of climate models for 2100, between about 10 and 40 cm in the global mean (see also Chap. 14). Global sea level is currently rising at about 3.2 mm $year^{-1}$ (Church and White 2011). If this trend continues through the twenty-first century, sea level rise by 2100 would be about 30 cm. The projections from global climate models, augmented by the more uncertain estimates of the dynamics of ice sheets and coastal glaciers, imply therefore an acceleration from the present rate of global sea level rise (see also Chap. 14). This is not necessarily the case for regional sea level rise, however, especially for the Baltic Sea with its complex coastline and the many different processes modulating its response to global warming. Nevertheless, the question as to whether the rate of sea level rise in the Baltic Sea is accelerating is relevant for adaptation to climate change and other planning purposes over the coming decades.

A preliminary indication of acceleration may be conveyed by a change in the long-term linear trends in two different periods of the recent past. Ekman (2009) suggested that the rise identified in the Stockholm sea level record is accelerating because the trends before and after 1950 look different (Fig. 9.7, upper right panel). However, this simplistic view should be complemented, if possible, by studies that analyse jointly the observed evolution of sea level and the theoretical evolution that sea level should display as a response to anthropogenic climate change. A change in the rate of sea level rise may not necessarily be projected by numerical climate models.

Most studies implicitly define the concept of 'acceleration' of sea level rise, possibly influenced by anthropogenic climate change, in one of two ways. The first approach is to look for a long-term increase in the *rate of* sea level rise, while the other compares the recent rate of sea level rise against rates observed in the past. The difference between both approaches can be illustrated by the following example. Given a sea level record, the linear trend of sea level change in sliding windows of fixed length (such as 30 or 50 years) can be calculated (Richter et al. 2011). If the linear rate in the most recent window is the highest, it could be claimed, following the second approach, that the present rate of increase is unprecedented, and thus possibly related to anthropogenic climate forcing. However, due to decadal variability in the rate of change, the linear rate in the most recent window may not be the highest over the whole record length. However, if the linear trends over the sliding windows itself display a linear trend, it could be claimed, following the first approach, that sea level rise is accelerating in that particular record. In a variant of the first approach, the

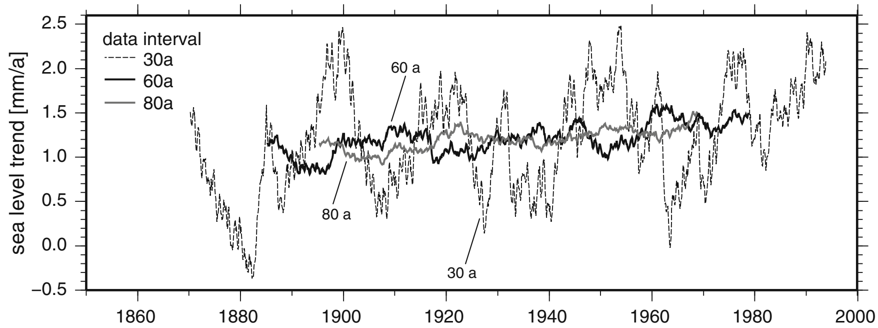


Fig. 9.11 Linear trends calculated in sliding windows of fixed length for the annual sea level record in Warnemünde (Germany), a station in the southern Baltic Sea. The three series show the results for different window lengths (redrawn from Richter et al. 2011)

long-term trends in sea level rise is determined by fitting a second-order polynomial in time to a sea level record (Church and White 2006; Woodworth et al. 2009). If the quadratic coefficient is found to be significantly different from zero, acceleration could claim to have been identified. More loosely, if abrupt changes are identified in the linear trends—denoted inflexion points—these may be attributed to acceleration or deceleration. Sea level acceleration has been looked for using both approaches in global studies. For the Baltic Sea, these approaches have yielded contradictory results. Woodworth et al. (2009) found a positive quadratic term for the rate of rise in mean Baltic Sea sea level of the order of 0.005 mm year^{-2}, derived from the global sea level reconstructions of Church and White (2006) and consistent with values found specifically for the Stockholm sea level record (Woodworth et al. 2011). However, the Baltic Sea series provided by Jevrejeva et al. (2008) displayed a *deceleration* inflexion point around 1920.

The different perspectives on acceleration may be illustrated by the results of Richter et al. (2011) (Fig. 9.11) for the Warnemünde sea level record, a station in the southern Baltic Sea. The 30-year trends indicate that the present rate of rise is not unprecedented in the record. Although the recent linear trends are high, similar or even slightly higher rates were observed around 1900 and 1950. These findings generally agree with the (non-peer-reviewed) results of Zorita and Hünicke (2010), who presented sliding linear trends of long sea level records, estimated by linear fit in moving 31-year windows for Stockholm (Sweden), Kronstadt (Russia), Kolobrzeg and Swinoujscie (Poland).

The time series of sliding 30-year trends (Fig. 9.11) illustrates the very high variability in sea level rise in the southern Baltic Sea over relatively short periods. Although the 30-year trend time series displays a slight positive upward trend, its statistical significance is low due to the large decadal variability. For trends over wider sliding windows (such as 60 or 80 years—see Fig. 9.11), the variability is reduced and a long-term increase in the linear rates of rise is easier to see. However, wider windows reduce the number of independent degrees of freedom which also compromises any estimate of statistical significance.

It is clear therefore that, depending on which approach is used, the same long-term sea level record may appear to exhibit both the presence and absence of a significant acceleration in sea level rise.

In cases where acceleration in sea level rise has been identified, this should ideally be accompanied by an estimate of its level of statistical significance. As in any statistical test, a general algorithm to attach a certain level of statistical significance of an estimated value cannot exist, since the significance level depends on the formulation of the null-hypothesis, against which the statistical test is conducted. This formulation is subjective, and different authors may consider different null-hypotheses to describe the statistical properties of the natural fluctuations in the rate of sea level change. Donner et al. (2012) reported different levels of statistical significance for the estimated acceleration depending on whether the data were assumed to be serially correlated or serially independent. The authors also attempted to estimate the acceleration in the rates of change of the quantiles of the distribution of monthly sea level for a set of Baltic Sea coastal stations. The record length across the dataset is not uniform and so the spatial patterns of the acceleration estimate and its level of statistical significance may be strongly influenced by data availability. When the analysis is applied to the period covered by all sea level records (1951–2000), none of the quantiles of the probability distribution display statistically significant accelerations if the de-seasonalised monthly sea level data are assumed to be serially correlated. However, they do underline that statistical detection of acceleration in only 50 years of data may not be possible due to the short record length.

9.4 Extreme Sea Levels

Extremes in coastal sea level oscillations pose a serious threat to coastal populations as their impacts can be catastrophic. At many coasts around the world, extreme sea levels are essentially determined by tides; however, this effect is negligible in the Baltic Sea due to the semi-enclosed nature of the basin. Lowe et al. (2010) reviewed change in extreme sea levels in recent decades and concluded that—globally—there is little evidence of change in extreme sea levels over an extended period, a behaviour which differs significantly from the change in GMSL. In their quasi-global investigation, Menendez and Woodworth (2010) showed that recent change in extreme sea levels is mostly due to the change in MSL (by a shift in the frequency distribution), while meteorological contributions vary on timescales of years and decades but mostly show no clear long-term trend. Thus, the question arises as to whether the amplitude and frequency of extreme sea level events in the Baltic Sea is changing and if so, how they compare to changes in Baltic Sea MSL.

9.4.1 Main Factors Affecting Extreme Sea Levels in the Baltic Sea

Because of its elongated shape, semi-enclosed configuration and presence of shallow bays exposed to the direction of strong winds, storm surges occur frequently on exposed coasts of the Baltic Sea. In contrast to tide-dominated basins, extreme sea levels in the Baltic Sea are mainly due to wind (wind set-up). Other contributors include the inverse barometer effect, standing waves (seiches) and the propagation or amplification of remotely generated long waves. Winds affect sea level in two main ways (e.g. Samuelsson and Stigebrandt 1996; Ekman 2007). First, wind may build up a sea level slope within the Baltic Sea, resulting in the strongest deviations at the ‘ends’ of the Baltic Sea (in the Belt Sea and the gulfs of Finland, Bothnia and Riga). Second, when a strong persistent wind from the south-west or north-east is blowing over the Baltic Sea and its entrance, water is transported into or out of the Baltic Sea, thereby raising or lowering sea level in the basin as a whole. Although the amplitude of such events is normally less than 50 cm, they can provide preconditions for much larger local-scale storm surges when combined with short-term storm winds during cyclones (Hupfer et al. 2003). All over the sea, both extreme high and low sea level events tend to occur in the meteorologically more variable winter months (e.g. Sztobryn et al. 2009; Suursaar 2011).

Owing to the large meridional extent of the Baltic Sea, the required forcing conditions, as well as storm surge risks, may vary strongly in different parts of the basin. As a general rule, in the eastern section near the coasts of Lithuania, Latvia, Estonia, Russia and Finland, strong easterly winds tend to lower sea level, and westerly winds to raise it (Suursaar et al. 2002, 2006a). This applies for short-time variations, as well as for low-frequency variations through the corresponding changes in the Baltic Sea sea water volume. The lowest recorded sea level in the Gulf of Riga (−130 cm at Riga, −125 at Pärnu) occurred in December 1959 after a month with strong (20 m s^{-1}) easterly winds during an anti-cyclonic blocking pattern (Suursaar et al. 2002).

The highest surges usually occur in the eastern part of the Gulf of Finland (Fig. 9.12). Such events have been analysed by Klevannyy et al. (2001), Nekrasov et al. (2007), Averkiev and Klevannyy (2010). As most of the cyclones over Scandinavia and the Baltic Sea travel from south-west or west to the east, storm surges are usually generated in the eastern or north-eastern sections of the basin. The surges are particularly pronounced in bays away of the nodal (central) part of the sea, such as the Neva Bay of St Petersburg (up to 421 cm). At the same time, negative surges or sea level lowering occur along the opposite, Swedish, Danish or German coasts. Those high and low sea level areas evolve in time, as the cyclones travel through a particular area.

For extreme surges to occur in the coastal waters of Estonia and St Petersburg (and for negative surges on the Swedish and Danish coasts), the centre of a powerful cyclone should bypass Estonia to the north over the Scandinavian Peninsula and Bothnian Sea to make the local wind direction veer from south-west to north-west (Suursaar et al. 2006b; Averkiev and Klevannyy 2010). As the strongest winds occur a few hundred kilometres to the right of the cyclone track, reduced friction on the sea surface and the elongated shape of the Baltic Sea together with the Pärnu Bay (in case of Pärnu tide gauge) or Gulf of Finland (for Narva-Jõesuu and St Petersburg) provides a span for surge waves to increase towards the east. The diminishing water depth and narrowing gulf width reinforce this effect. Since 1703, there have been about 300 flooding events in St Petersburg due to a sea level rise above the critical value of 160 cm (Bogdanov et al. 2000). In 1970, a decision was taken to build a 25-km-long dam with huge closable gates to protect the city. This structure was finished in 2011.

A map of historical water-level maxima around the Baltic Sea coast was recently compiled by Averkiev and Klevannyy (2010) (Fig. 9.12). The authors compiled information from various sources and reported the data according to the relevant national water-level reference system. Data from stations of short duration using tide gauges are also included. Thus, the map is not homogeneous and must be considered a rough picture of extreme water-level maxima in the Baltic Sea (see also Fig. 9.13 for comparison). More detailed information is available from the original references (e.g. Hupfer et al. 2003; Sztobryn et al. 2005 and references therein; Dailidienė et al. 2006; Suursaar et al. 2006a; Kowalewska-Kalkowska and Wisniewski 2009).

Cyclone Gudrun on 9 January 2005, which attained a hurricane-like intensity according to the mean wind speed measurements (up to 34 m s^{-1}) in Denmark, produced its highest sea level at most of the Estonian tide gauges (275 cm at Pärnu; the previous highest was 253 cm on 19 October 1967) and some Finnish tide gauges (Helsinki 151 cm,

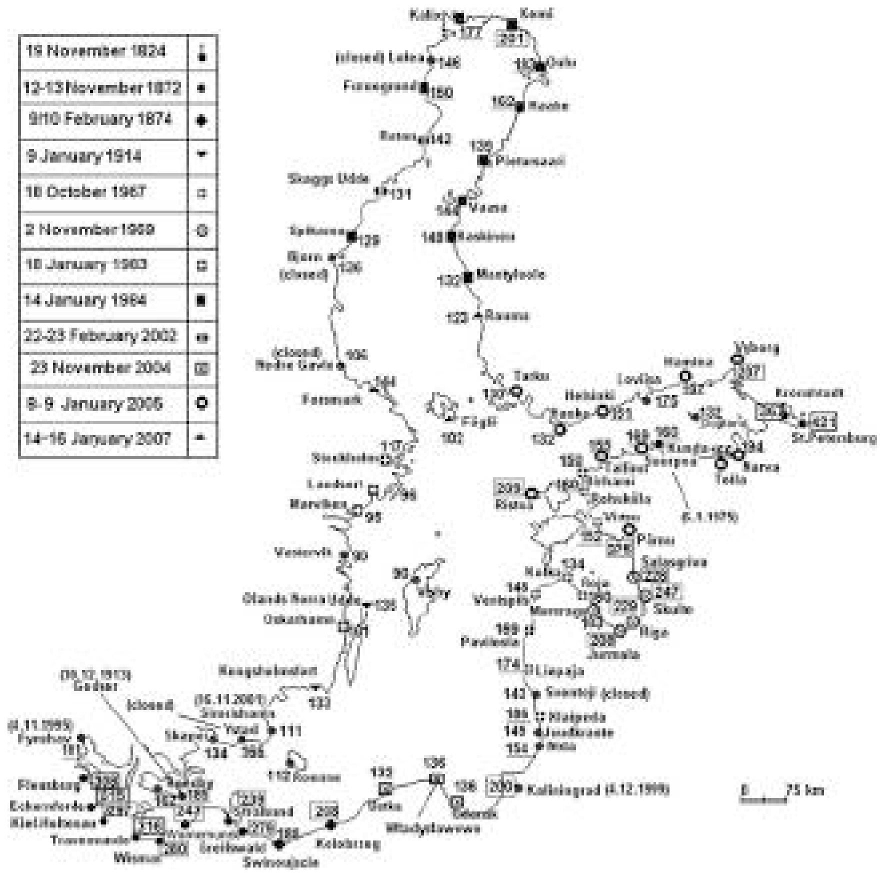


Fig. 9.12 Historical water level maxima (cm) in the Baltic Sea (Averkiev and Klevannyy 2010)

Hamina 197 cm, Hanko 132 cm, Turku 130 cm). However, previous records were not surpassed in the Gulf of Bothnia (Kemi 201 cm in 1982) or in St Petersburg, where sea level reached a relatively modest 230 cm. Gudrun's eye passed 300 km north of Estonia, heading north-east and creating strong south-westerly winds (and later west–north-westerly winds once it had passed) reaching an average speed of 28 m s^{-1} over one hour. Another factor contributing to Gudrun's storm surge was the relatively high background sea level (70 cm) in the Baltic Sea at the time (Suursaar et al. 2006b). The high background sea level was due to strong cyclonic activity during the preceding month, which had funnelled additional water through the Danish Straits into the Baltic Sea.

Narrow bays in the Belt Sea and in the south-western Baltic Sea have experienced the second highest storm surges (see Fig. 9.13), but also the strongest negative surges. Sztobryn et al. (2009) described the characteristics of negative surges in the southern Baltic Sea using data (1955–2005) from three German gauge stations (Wismar, Warnemünde,

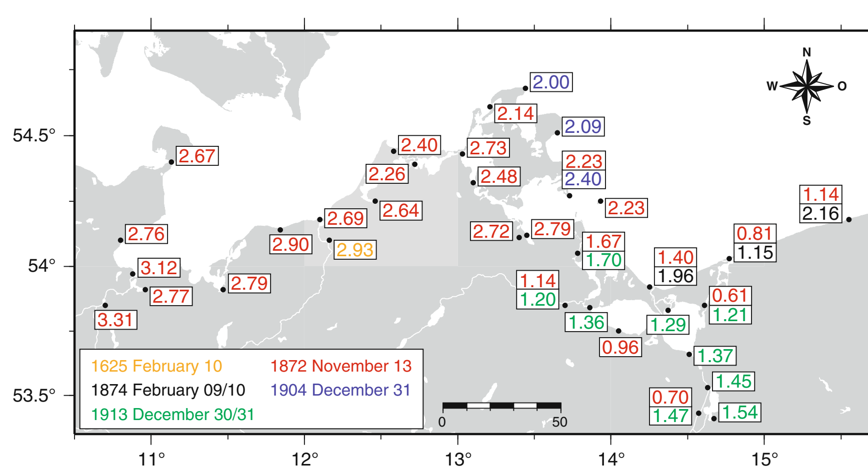


Fig. 9.13 Absolute water levels (in m relative to the NN height system) along the southern Baltic Sea coast according to data from tide gauge observations and storm surge marks (redrawn from Richter et al. 2011)

Sassnitz) and two Polish gauges (Swinoujscie, Kolobrzeg). Their results confirmed that strong offshore winds (usually accompanied by low-pressure systems crossing the Baltic Sea) or storms driving the water away from the coast are the most important factor for the development of negative surges. The frequency and severity of negative surge events on the south-western Baltic Sea coast was found to decrease from west to east due to the baylike shape of the coast with an eastward opening (Sztobryn et al. 2009).

The maxima in many tide gauge records of the area were established by one spectacular north-easterly storm in November 1872 (see Fig. 9.13). A negative surge of about 1 m occurred in the Gulf of Finland (Ekman 2009). Colding (1881) analysed the event in Denmark. The author concluded that a storm raises sea level at the coast proportional to the square of the wind velocity and that the effect is also proportional to the fetch and inversely proportional to the sea depth.

Other major storm surge prone areas include the Gulf of Riga, and particularly Pärnu Bay (up to 275 cm) and the northern part of the Bothnian Bay (Kemi, 201 cm). Located close to the nodal line of the sea, the line of no vertical displacement in standing, basinwide oscillations—the Lithuanian and Latvian coasts in the Baltic Proper, the islands of Gotland, Åland and the Swedish coasts located leeward—will not experience high sea level events even during extreme storms. For a discussion of impacts on the coastlines of the Baltic Sea due to extreme sea level events, see Chap. 20.

9.4.2 Statistics and Long-Term Trends in Extreme Sea Levels

Richter et al. (2011) analysed long time series of extreme sea levels inferred from tide gauge records and historical documents of flood marks at the south-western Baltic Sea coast (Fig. 9.13). Changes in the magnitude of climate-driven extremes over the past 200 years could not be detected. This result contrasts with the findings of Sztobryn et al. (2005), who analysed extreme sea levels at the southern Baltic Sea coast (western and central parts) between 1950 and 2000 and found an increase in storm surges in the decades to 1990.

Woodworth and Blackman (2004) analysed the Stockholm time series (Fig. 9.14) and showed that the spread of higher percentiles is greater than that of the lower percentiles. They explained this result primarily by asymmetries in air pressure data and the wind fields. Also, the higher percentiles have similar time dependence on MSL which means that sea level extremes have a greater response to changes in the winter NAO index than MSL.

To summarise, Donner et al. (2012), Barbosa 2008 (see Sect. 9.3.2.1 and Fig. 9.8) found that trends in median sea level do not always coincide with the trends in extreme high and low quantiles. Upper quantiles tend to display a more positive trend than the median and the lower quantiles. This is consistent with the findings of Johansson et al. (2001), who investigated temporal changes in extreme sea level in Finland. They used homogenised Finnish series (13 stations, mostly covering the period 1923–1999) and found a significant 2–4 mm year^{-1} rise in annual sea level maxima. At the same time, the rise in annual sea level minima was only around 1 mm year^{-1}. Along the Lithuanian coast, the average rise in maxima was around 2–3 mm year^{-1} (Dailidienė et al. 2006).

Studies at the Estonian coastline showed a remarkable rise in annual sea level maxima of 3.5–11 mm year^{-1} after correcting for local uplift rates (Vallner et al. 1988; Suursaar and Sooäär 2007). Even excluding the prominent event of 2005, the rise in annual sea level maxima was significantly higher than the rise in MSL (1–2.6 mm year^{-1}) and the rise in annual sea level minima (0.8–3.1 mm year^{-1}). The hydrodynamic mechanism responsible for these differences appears linked to the location of Estonian and Finnish tide gauges with respect to the predominant strong winds

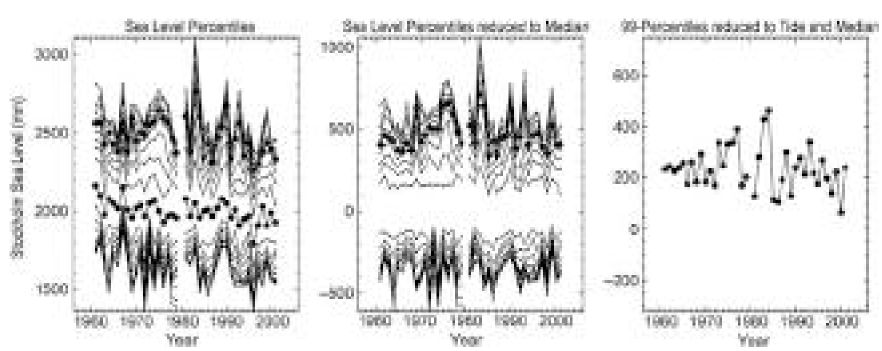


Fig. 9.14 An example of percentile time series from Stockholm, Sweden, showing the 19 separate percentile series. The 50th percentile or median (no. 10 of 19) and 99th percentile (no. 15 of 19) series are shown by *large dots*; (*middle*) as in (*left*) but with each percentile time series reduced to the median values; (*right*) the difference between the observed 99th percentile series and that from tidal analysis of each year of data (redrawn from Woodworth and Blackman 2004)

(Suursaar et al. 2006a, Suursaar and Kullas 2006). Analysis of local wind data from Estonian coastal stations has shown that, although the mean wind speed is likely to have decreased over the past 50 years, the westerly wind component and the extreme wind events have increased (e.g. Suursaar and Kullas 2009a, b; Jaagus 2009, Jaagus and Kull 2011, see also Chap. 4, Sect. 4.3). This is in good agreement with findings on changes in cyclone trajectories above the Baltic Sea (Sepp et al. 2005; Jaagus et al. 2008). The magnitude of an expected extreme event at a given location is usually described by the magnitude of the expected maximum extreme event over a certain period, such as 100 years or 1000 years. Alternatively, an observed extreme event may be described as the one expected to occur once in a certain period. Normally, a storm surge prone area (city, port) has established critical sea level values, usually based on statistical analyses of long tide gauge records. However, these projections are only truly valid in an unchanging climate. To incorporate the effects of future climate change requires additional information, climate simulations from numerical models and/or extrapolation of recently observed trends. However, as extreme events are by definition rare, empirical extrapolation carries large uncertainties.

There are many examples worldwide of ‘storms of the century’ or ‘one in a thousand years’ surge events that occur more frequently than expected from statistical analysis. This may be due to the limited number of observations available for estimating the magnitude of extreme events. For example, a comparison of empirical return periods against the corresponding theoretical distributions shows a reasonable fit for three Estonian tide gauges, but no fit in the case of Pärnu maxima (Suursaar and Sooäär 2007). The same situation exists for the Travemünde tide gauge in Germany, where sea level in 1872 was extraordinarily high and appeared to be an outlier in relation to the observed distribution of extreme events. However, when the tide gauge data were supplemented by historic storm surge water levels and simulations from numerical models, the 1872 value fitted well with the distribution based on this larger dataset (see Mudersbach and Jensen 2009). This is in general agreement with the findings of the MUSTOK project (model-based investigation of extreme water levels at the German Baltic Sea coast, see Jensen et al. 2009 and references therein). One of the assumptions in extreme value analysis is that there is no physical limit that an extreme event can attain, which is not a totally valid assumption in the case of storm surges. MUSTOK aimed at combining model-based estimates of the maximum attainable storm surge, high-quality data from instrumental measurements, and the more uncertain information from historical sources, to provide better estimates of storm surge return periods at several sites in the Baltic Sea. Including physically based limits for the possible extremes can substantially modify the probability of storm surge occurrence. For instance, using recent instrumental data only would have suggested that the storm surge in Travemünde was a once-in-10,000 year event, whereas including all three datasets in the projection would shorten the return period to 3400 years.

9.5 Wind Waves

The properties of wind waves depend primarily on wind speed and duration, and effective fetch length. In semi-enclosed shallow basins, wave properties are also modified through wave–seabed interaction (via refraction, shoaling, breaking or reflection) and diffraction behind obstacles. The Baltic Sea is a challenging location for wave scientists; the area has a very complex geometry and the associated high variability in wind fields gives rise to extensive spatio-temporal variability in the wave fields. Wave simulations therefore require a high spatial resolution and a careful choice for the location of wave measurements. The presence of sea ice (see Chap. 8) may substantially affect wave patterns and complicates wave measurements in winter. As floating devices are normally removed before the ice season (Kahma et al. 2003; Tuomi et al. 2011), measured wave data (especially in the northern Baltic Sea) display extensive time gaps and estimates of commonly used variables (e.g. annual mean wave height or period) become biased for seasonally ice-covered seas (Tuomi et al. 2011). Relatively shallow areas and convergent wind patterns may lead to unexpectedly high waves, formed through occasional wave energy concentration in some areas (Soomere 2003, 2005; Soomere et al. 2008a). Also, specific wave generation conditions for offshore winds over irregular coastlines (Kahma 1981; Kahma and Calkoen 1992) or under so-called slanting fetch (when wind blows obliquely across the coastline) frequently occur in some sub-basins (Pettersson et al. 2010).

9.5.1 Instrumental Measurements

Long-term (>10 years) instrumentally measured wave data are available only at three sites (Fig. 9.15). In terms of climate change, only data from upward-looking echo sounders at Almagrundet (1977–2003, Broman et al. 2006) and from a directional wave-rider at Darss Sill have been analysed (Soomere and Kurkina 2011; Soomere et al. 2012). The data from the wave-rider in the northern Baltic Proper were discussed by Kahma et al. (2003), Tuomi et al. (2011). Numerous relatively short-term wave measurement sites around Finland in the 1980s and 1990s were briefly described by Soomere (2008). Satellite altimeter data for wave properties have been used in a very small number of studies (Cieślikiewicz and Paplińska-Swerpel 2008; Tuomi et al. 2011).

9.5.2 Visual Observations

Historic wave observations from ships in offshore regions combined with the results of hindcast simulations have been used to generate wave atlases for the Baltic Sea (Rzheplinsky 1965; Russian Shipping Registry 1974; DWD 2006; Lopatukhin et al. 2006) and its sub-basins (Rzheplinsky and Brekhovskikh 1967; Druet et al. 1972; Schmager 1979; Sparre 1982).

Visual observations from anchored lightships and permanent coastal stations (Fig. 9.15, Table 9.2) cover a much longer time interval than instrumental measurements. Visually observed wave heights generally agree well with significant wave heights, whereas the visually estimated wave period is, on average, a few tenths of a second shorter than the peak period from hindcast simulations (Gulev and Hasse 1998, 1999). In contrast to offshore ship-based wave observations for estimates of wave climate (Gulev and Hasse 1999; Gulev et al. 2003), use of observations from coastal sites is more problematic. On the one hand, such data pose intrinsic quality and interpretation problems, have a poor temporal resolution, contain a large element of subjectivity and a substantial amount of noise (Zaitseva-Pärnaste et al. 2009), and only partially characterise the open sea wave fields (Soomere 2005). On the other hand, they have exceptional temporal coverage: regular observations have been carried out using a unified procedure at several locations for almost seven decades (Zaitseva-Pärnaste et al. 2011; Pindsoo et al. 2012).

Visual wave observations from Danish lightships have been made at multiple locations including Gedser Rev and Halsskov Rev and a permanent station at Drogden with six to eight observations per day (Sparre 1982; Fig. 9.15). The data are digitised from 1931 until the withdrawal of the ships in the 1970s and 1980s. Observations from Drogden continued until 1994. The ships were located at shallow sites of particular danger to ship traffic. This may affect the wave field statistics. Also, the lightships were relocated short distances on several occasions and there are some temporal gaps. Nevertheless, the dataset constitutes a unique source of information on the western Baltic Sea historic wave climate.

Wave observations on the eastern coast of the Baltic Sea have been made in many locations since the mid-1940s or 1950s (Table 9.2) usually three times per day using perspectometers (binoculars with specific scaling) and/or buoys or bottom-fixed structures to better characterise wave properties. The observers scanned areas at least 4 m deep about 200–400 m from the waterline. The observation conditions vary considerably between locations; for example at Pakri (Estonia), the observer was located on top of a 20-m-high cliff and the water depth of the area observed was 8–11 m. All sites of visual observations presented in Fig. 9.15 only

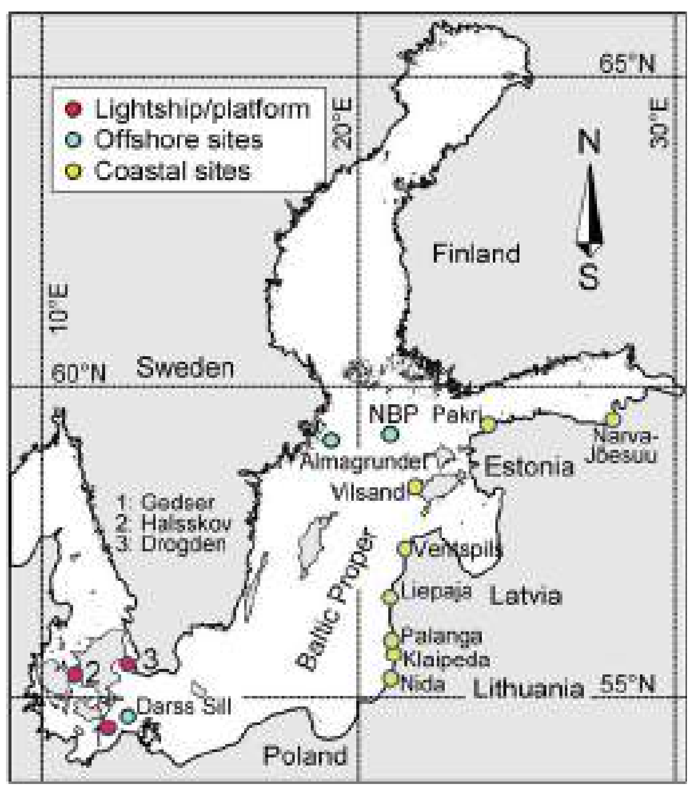


Fig. 9.15 Location of long-term (>15 years) instrumental wave measurements in the Baltic Sea (including information from Soomere and Räämet 2011; Sparre 1982)

Table 9.2 Sources of climatic wave information in the published literature

Location	Time series	Analysis of temporal change?	Source
Instrumental measurements			
Almagrundet, upward-looking echo sounders	1977–2003	Yes	Broman et al. (2006)
Darss Sill, directional wave-rider	Since 1991, excl. ice seasons	Yes	Soomere et al. (2012)
Northern Baltic Proper	Since 1996, excl. ice seasons	No	Tuomi et al. (2011)
Visual observations			
Global, including Baltic Sea (observations from ships)	1958–1997	No	Gulev et al. (2003)
	1970–2011	No	Gulev et al. (2005)
Danish waters (Drogden, multiple locations of lightships)	1931–1994	–	Sparre (1982)
Swedish waters (multiple locations and lightships)	1965–1972	No	Wahl (1974)
Lithuanian waters (Nida, Palanga, Klaipeda)	1954–2009	Yes	Kelpšaitė et al. (2008, 2011), Zaitseva-Pärnaste et al. (2009, 2011)
Latvian waters (Ventspils, Liepaja)	1946–2011	Yes	Pindsoo et al. (2012)
Estonian waters, NE Baltic Proper (Vilsandi) and the Gulf of Finland (Pakri, Narva-Jõesuu)	1946–2009	Yes	Soomere and Zaitseva (2007), Zaitseva-Pärnaste et al. (2009), Soomere et al. (2011)
Hindcasts			
Global, including Baltic Sea	KNMI/ERA-40 Wave Atlas	Yes	Sterl and Caires 2005; low spatial resolution (1.5° × 1.5°)
Baltic Sea basinwide	1999	No	Jönsson et al. (2003)
	2001–2007	No	Tuomi et al. (2011)
	1947–1988	No	Mietus and von Storch (1997)
	1958–2001	No	Cieślikiewicz and Paplińska-Swerpel (2008)
	1992–2002	No	Schmager et al. (2008)
	1958–2002	No	Augustin (2005), Schmager et al. (2008)
	1970–2007	Yes	Räämet and Soomere (2010, 2011)
Southern Baltic Sea	1978–1996	No	Blomgren et al. (2001)
Entrance to Warnemünde harbour	1956–1993 (single storms) 1988–1993	No	Gayer et al. (1995)
Pomeranian Bay	1997–1998	No	Paplińska (1999)
Tallinn Bay, Gulf of Finland	1981–2000	No	Soomere (2005)
	1981–2007, single-point winds	Yes	Kelpšaitė et al. (2009)
Hindcasts for single points in north-eastern Baltic Proper and Gulf of Finland		Yes	Suursaar and Kullas (2009a, b), Suursaar (2010, 2013), Suursaar et al. (2010, 2012)

partially represent the open sea wave conditions. Although the potential distortions due to the sheltering effect of the shoreline and the relatively shallow water depth may affect the results of individual observations, it is likely that long-term variations and trends in offshore wave properties would be evident through the analysis of large pools of visual observations. Owing to short daylight duration, only one to two observations per day are possible in northern areas in autumn and winter. To avoid the bias caused by a varying number of observations per day, the analysis of wave data is mostly based on the set of daily mean wave heights (Soomere and Zaitseva 2007; Zaitseva-Pärnaste et al. 2009, 2011).

9.5.3 Hindcast Simulations

The spatial resolution of numerically reconstructed global wave datasets such as the KNMI/ERA-40 Wave Atlas (1.5° × 1.5°) (Sterl and Caires 2005) is insufficient for the Baltic Sea. Several attempts to numerically reconstruct the wave climate in the Baltic Sea using a better resolution (down to about 3 nautical miles or about 5 km) have been undertaken for many areas and single locations (Table 9.2). Schmager et al. (2008), Soomere (2008) reviewed the relevant literature until 2007 and described the basic features of the wave climate.

Long-term reconstructions of the Baltic Sea wave fields are a complicated task and usually contain high uncertainties (Cieślikiewicz and Paplińska-Swerpel 2008; Kriezi and Broman 2008). The largest source of uncertainty is the wind information. The quality and spatial resolution of the wind data (down to about 10 km, Tuomi et al. 2011) have increased over the past decade. Simulations of long-term change in wave fields are, however, hampered by substantial temporal inhomogeneity (Tuomi et al. 2011), large spatial differences in data quality (Räämet et al. 2009; Soomere and Räämet 2011), and by the inability of even the most advanced atmospheric models to reproduce air flow accurately in several Baltic Sea sub-basins (Keevallik and Soomere 2010). Nevertheless, long-term numerical reconstructions of change in the Baltic Sea wave climate are available for 1958–2002 based on output from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCAR/NCEP) wind reconstructions (Augustin 2005; Schmager et al. 2008; Weisse and von Storch 2010) and for 1970–2007 based on adjusted geostrophic winds (Räämet and Soomere 2010; Soomere and Räämet 2011).

The relatively small size of the Baltic Sea, frequent large-scale homogeneity in the wind fields and the short saturation time and memory of wave fields make it possible to use simplified wave hindcast schemes (Soomere 2005), high-quality wind data from a few points (Blomgren et al. 2001) and/or properly calibrated simple fetch-based wave models (Suursaar and Kullas 2009a, b; Suursaar 2010) to reproduce local wave statistics with an acceptable accuracy. The use of such models for identifying spatial change in wave statistics is limited, however, as they can basically only reproduce change in the local wind field.

9.5.3.1 Long-Term and Extreme Wave Properties

Existing sources of Baltic Sea wave information make it possible to identify basic long-term wave properties (average and extreme height, occurrence distribution and height–period combinations) and their spatial variations. The Baltic Sea wave climate is, on average, very mild. The typical long-term significant wave heights are about 1 m offshore in the Baltic Proper (Kahma et al. 2003; Broman et al. 2006; Schmager et al. 2008; Tuomi et al. 2011), 0.6–0.8 m in the open parts of larger sub-basins such as the Gulf of Finland (Soomere et al. 2010) or Arkona Basin (Soomere et al. 2012) according to measurements and numerical simulations, and well below 0.5 m in semi-sheltered bays such as Tallinn Bay (Soomere 2005; Kelpšaitė et al. 2009) according to simulations (Fig. 9.16). These values are 10–20 % lower in the nearshore regions (Suursaar and Kullas 2009a, b; Suursaar 2010). The most frequent wave heights are also about 20 % lower than the long-term average wave height (Kahma et al. 2003; Soomere 2008; Soomere et al. 2012).

The spatial pattern of hindcast average wave heights (Fig. 9.16, left panel) contains either an elongated maximum

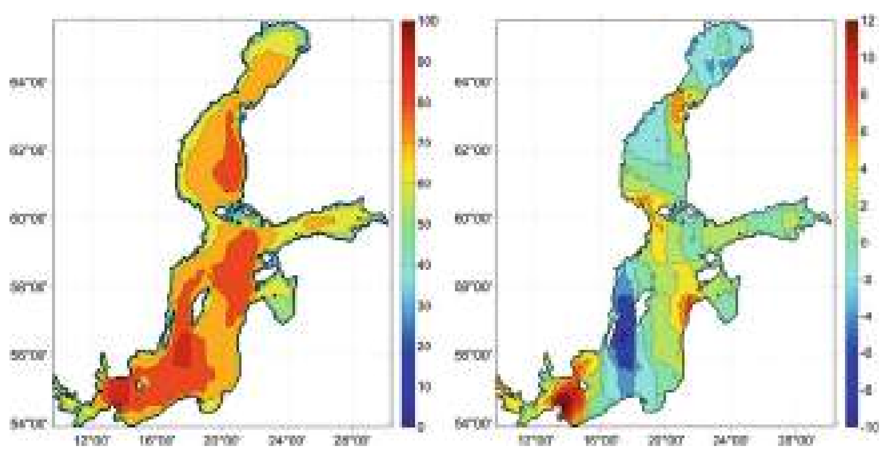


Fig. 9.16 *Left* Numerically simulated average significant wave height (colour bar, cm; isolines plotted after each 10 cm) in the Baltic Sea in 1970–2007 based on adjusted geostrophic winds from the Swedish Meteorological and Hydrological Institute (Räämet and Soomere 2010); *Right* Long-term change in the annual average significant wave height (cm, based on the linear trend, isolines plotted after each 2 cm) for 1970–2007 (Soomere and Räämet 2011). Note that a local maximum in the Arkona Basin is evidently caused by overestimation of the 10-m wind speeds from the geostrophic wind data

(Augustin 2005; Tuomi et al. 2011) or several local maxima in the eastern Baltic Proper (Jönsson et al. 2003). A hindcast using geostrophic winds indicates another maximum to the south of Gotland (Räämet and Soomere 2010). This may be explained by large interannual and decadal variability in wind patterns over the area that naturally causes extensive variations in both average and extreme wave patterns. A very similar spatial pattern of heights to the one shown in Fig. 9.16 (left panel) is found for extreme waves (threshold of the 1 % or the 99th percentile of significant wave height for each year) based on geostrophic winds (Soomere and Räämet 2011).

The empirical distributions of the occurrence of different wave heights at offshore measurement sites (such as Almagrundet, Bogskär, the wave-rider in the northern Baltic Proper and at the Darss Sill) resemble a Rayleigh distribution, with typical values of the shape parameter of 1.5–1.8 (Soomere 2008; Soomere et al. 2011). The most frequent wave heights are usually in the range 0.5–0.75 m. Significant wave heights greater than 4 m occur offshore with a probability of about 1 %, that is, during about 60 h per year. Significant wave heights over 7 m have been measured, on average, about twice per decade in the northern Baltic Proper (Soomere 2008; Tuomi et al. 2011) and so can be considered extreme conditions.

The wave climate is, however, highly intermittent and the sea occasionally experiences very high waves. The largest instrumentally measured values of significant wave heights (H_s) are H_s = 8.2 m in the northern Baltic Proper on 22 December 2004 (Directional Waverider, Tuomi et al. 2011) and H_s = 7.82 m on 13/14 January 1984 at Almagrundet [upward-looking echo sounder; estimated from the 10th highest waves based on Rayleigh distribution; an alternative estimate from the wave spectrum is H_s = 7.28 m (Broman et al. 2006)]. The highest individual instrumentally measured waves were 14 and 12.75 m, respectively. Numerical simulations indicate that H_s may reach 9.5–10 m in the north-eastern Baltic Proper at the entrance to the Gulf of Finland, to the north-west of the Latvian coast and in the south-eastern part of the Gulf of Gdansk (Schmager et al. 2008; Soomere et al. 2008a; Tuomi et al. 2011). The properties of waves in a particular region and storm event depend on the match of the geometry of the sea area and the wind pattern of the storm (Augustin 2005; Soomere et al. 2008a; Schmager et al. 2008).

Typical and extreme wave heights are much lower in sub-basins of the Baltic Sea. A large proportion of the low waves (H_s < 0.25 m) for most of the coastal visual observation sites (except for Vilsandi) (Zaitseva-Pärnaste et al. 2009, 2011) resembles an analogous property of simulated distribution in semi-sheltered bays (Soomere 2005). The frequency of occurrence of waves with H_s > 4 m is very low for all sub-basins except the Bothnian Sea (Tuomi et al. 2011) and waves with H_s > 2 m may be considered extreme for semi-sheltered areas such as the Darss Sill (Soomere and Kurkina 2011; Soomere et al. 2012). However, even sheltered bays may experience very strong waves during violent storms from unfavourable directions; for instance, H_s > 4 m apparently occurred in the interior of Tallinn Bay on 15 November 2001 (Soomere 2005). The maximum measured significant and single wave height in the Gulf of Finland was 5.2 and 9 m, respectively, on 15 November 2001 (Tuomi et al. 2011). Other sub-basins are not covered by regular wave measurements. The largest instrumentally measured wave height in the southern Sea of Bothnia (6.5 m) was recorded on 26 December 2011 (Pettersson et al. 2012). The maximum H_s recorded at the Darss Sill is 4.47 m (Soomere et al. 2012). Numerically simulated maxima of H_s are 7.6 m in the Sea of Bothnia, about 5 m in the Gulf of Finland and Gulf of Riga, and 6.7 m in the Arkona Basin (6.23 m according to another simulation).

The most frequent periods are 3–5 s offshore and 2–4 s in coastal areas (Soomere 2008). Most of the combinations of wave heights and periods roughly correspond to wave fields with a Pierson-Moskowitz (PM) spectrum (Soomere 2008; Räämet et al. 2010; Soomere et al. 2012). This indicates a large proportion of fully saturated seas in this region at wind speeds up to about 8 m s^{-1} (Schmager et al. 2008). The properties of the roughest seas, however, match better a JONSWAP spectrum (corresponding to fetch-limited seas and characterised by shorter periods than wave fields with a PM spectrum), especially in areas with limited fetch such as the Darss Sill (Soomere et al. 2012). Swells are very limited in all parts of the Baltic Sea.

9.5.3.2 Spatio-Temporal Variations

Several observation sites reveal short-term (weekly) features in the wave activity that reappear regularly, for example, a relatively calm period at the end of December and beginning of January in the northern Baltic Sea (Soomere et al. 2011). These features are probably site-specific and only persist for a few decades (Soomere et al. 2012). The extensive seasonal variation in wind speed (Mietus 1998) causes substantial variability (by a factor of two in coastal areas and up to three in offshore regions) in wave height at monthly scales (Schmager et al. 2008; Soomere and Räämet 2011) (Fig. 9.17). The calmest months are April to July and the windiest October to January (see also Chap. 4).

The most extensive interannual and decadal variations in wave properties exist in the visually observed wave data (Fig. 9.18). The appearance and spatial coherence of such variations has undergone major change. Short-term (interannual and up to 3 years) variability in annual mean wave height displays a consistent pattern (with a typical spatial scale of more than 500 km) from the southern Baltic Proper to the eastern Gulf of Finland from the mid-1950s to the mid-1980s. The coherence is lost in the mid-1980s (Soomere et al. 2011): since then, years with relatively large wave heights at the coasts of the Baltic Proper correspond to calm years in the eastern Gulf of Finland and vice versa.

Studies of variations at longer timescales reflect greater spatial variability of wave properties at different sites. Augustin (2005) identified an increase of 0.3 m in the simulated annual 99th percentile of H_s in the Baltic Proper (58° N, 20° E) that was mostly due to an increase in the frequency of severe wave events (Sect. 2.3.5 in BACC Author Team 2008). Simulations over the period 1970–2007 (Soomere and Räämet 2011) confirmed the presence of an increasing trend at this location but also indicated a complex spatio-temporal pattern of change (with scales down to about 100 km) in the Baltic Sea wave fields (Fig. 9.16). Moreover, the change in wave height over time in the northern Baltic Sea does not follow the gradual increase in annual mean wind speed in the northern Baltic Proper (Island of Utö, Soomere and Räämet 2011).

Analogous changes of even greater magnitude were identified from visual observations for the eastern Baltic Sea coast and from instrumental measurements at Almagrundet (Fig. 9.18). There is an overall decrease in wave height from about 1960 until the end of the 1970s (Soomere et al. 2011). Wave height increases substantially in the northern Baltic Proper from the mid-1980s until the mid-1990s (Broman et al. 2006; Soomere and Zaitseva 2007), similar to what happens in the south-western Baltic Sea and North Atlantic Ocean (Gulev and Hasse 1999; Weisse and Günther 2007). The increase was followed by a considerable decrease since 1997 (Broman et al. 2006; Soomere and Zaitseva 2007). Wave height at the south-eastern Baltic Sea (Lithuanian) coast showed the opposite: a rapid decrease until about 1996 followed by a rapid increase (Zaitseva-Pärnaste et al. 2011). The variations were weaker in the simulations of Soomere and Räämet (2011) and in reconstructions using a fetch-based model and local wind data (Suursaar and Kullas 2009a, b; Suursaar 2010).

Such extensive variations raise the issue of the significance of factors such as instrument error, observers' error or

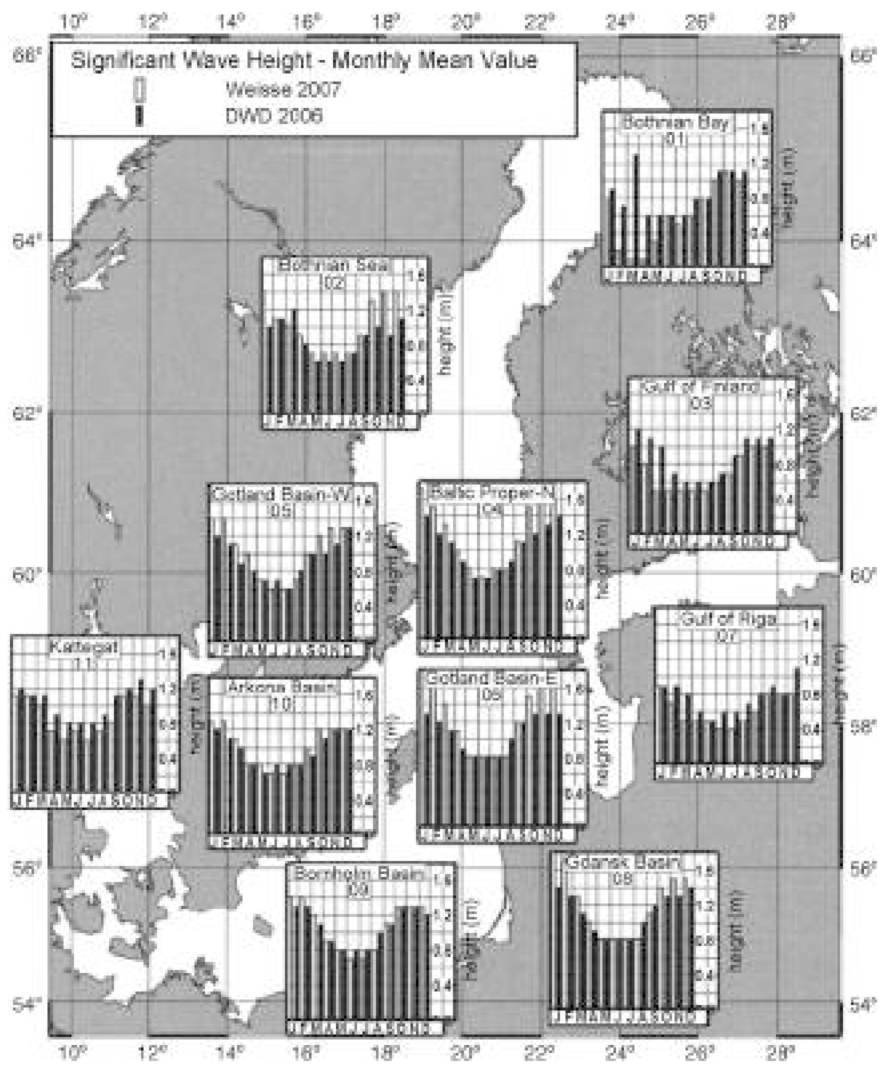


Fig. 9.17 Monthly mean significant wave height at selected sites in the Baltic Sea based on numerical simulations and DWD (German Weather Service) observations (Schmager et al. 2008)

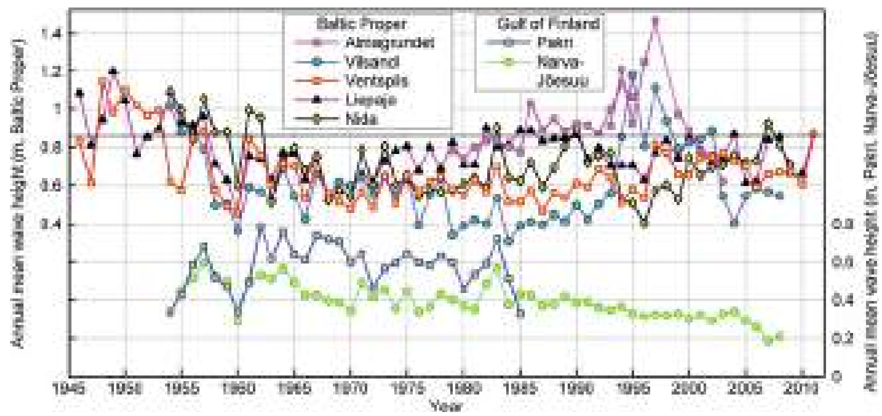


Fig. 9.18 Annual mean observed (Nida, Vilsandi, Pakri, Narva-Jõesuu, Liepaja, Ventspils) and instrumentally measured (Almagrundet) wave heights in the Gulf of Finland and Baltic Proper (redrawn from Soomere et al. 2011, including information from Soomere 2013)

noise in the data. The quality of data showing a decrease in wave intensity at Almagrundet for 1997–2003 was questioned by Broman et al. (2006), but the similarity of the changes at Almagrundet and Vilsandi still suggests that both datasets probably reflect real change (even if magnitude is possibly overestimated). The timing of the variations matches an almost twofold increase in the number of atmospheric low-pressure observations at Härnosand in the 1990s (Bärring and von Storch 2004) and so may simply reflect a change in the trajectories of cyclones.

For some locations along the Estonian coast, simplified fetch-based models using one-point coastal wind have demonstrated that wave intensity changes quasi-periodically and reveals no statistically significant trend (Suursaar and Kullas 2009a, b; Suursaar 2010). Quite large variations in the average wave periods (from about 2.3 s in the mid-1970s to 2.65 s around 1990) were found for selected sites (Suursaar and Kullas 2009b). This is apparently a local effect, because in most of the Baltic Sea the most frequent wave periods and the distribution of the different wave periods are effectively unchanged (Soomere et al. 2012).

Substantial changes (up to 90° from NW to SW) in the most frequently observed wave directions occurred at Narva-Jõesuu during the latter half of the twentieth century (Räämet et al. 2010) probably as a reflection of substantial change in local wind direction (Jaagus 2009; Jaagus and Kull 2011). Similar changes (of much smaller amplitude) were observed at the Lithuanian coast and were interpreted as a possible reason for change in the distribution of erosion and accumulation areas (Kelpšaitė et al. 2011). In general, it is likely that change in the simulated directions of wave propagation may result in substantial changes in the patterns of wave-driven sediment transport (Viška and Soomere 2012).

The changes discussed here are not necessarily evident in all sub-basins of the Baltic Sea. For instance, there has been effectively no change in the annual mean H_s at the Darss Sill (Soomere et al. 2012). The overall course in the wave activity in several other parts of the Baltic Sea also reveals no clear long-term trend (Soomere and Zaitseva 2007; Soomere 2008; Soomere et al. 2012). Only at Narva-Jõesuu in the Gulf of Finland is the wave intensity gradually decreasing (Soomere et al. 2011).

As previously mentioned, numerical simulations have revealed a complicated pattern of trends in wave intensity (Fig. 9.16, right panel). The spatial pattern of changes in the extreme (99th percentile) wave heights largely follows the pattern of changes in average wave height (Soomere and Räämet 2011). The decrease in wave intensity has been greatest between Öland and Gotland, and to the south of these islands to the Polish coast. In agreement with Augustin (2005) and the BACC Author Team (2008), a considerable increase in wave activity was indicated by the model in most of the northern Baltic Proper. The local maximum in the Arkona Basin and a rapid increase in simulated wave height in this region may stem from the overestimation of geostrophic wind speed in this part of the basin (see Pryor and Barthelmie 2003). There is a very slow decrease (about 0.01 m s^{-1} $year^{-1}$) in the annual mean wind speed at Kalbådagrund, Gulf of Finland (Soomere et al. 2010) where numerical simulations indicate very minor change in wave intensity (Suursaar and Kullas 2009b; Suursaar 2010; Soomere et al. 2010).

In some areas, change in average wave height and extreme wave height have opposite signs (Augustin 2005; Soomere and Healy 2008). For example, no long-term trend in average wave height exists at this site but the 99th percentile considerably decreased over 1991–2010. Hindcast simulations suggest a sawtooth-like behaviour for the 99th percentile, with a gradual increase for 1958–1990 from about 4 to about 5 m, a sudden decrease in 1991–1992 and a subsequent increase (Soomere and Kurkina 2011). Suursaar and Kullas (2009b) noted a negative trend in the 99th percentile near the north Estonian coast and a weak, opposite, gradually positive trend in average wave height. Simulations using the WAM wave model (Komen et al. 1994) showed that, unlike for average wave height, there has been a substantial decrease (of about 10 %) in maximum wave height near the southern coast of the Gulf of Finland and an almost equal increase to the north of the axis of the Gulf of Finland (Soomere and Räämet 2011). This feature may potentially cause enhanced coastal erosion in the affected areas in the north-eastern Gulf of Finland (Ryabchuk et al. 2011). It is apparently related to the major change in wind direction over the Estonian mainland: the frequency of south-westerly winds has increased considerably over the past 40 years (Jaagus 2009; Jaagus and Kull 2011).

9.6 Conclusion

The GIA—a readjustment of the Earth's crust to the off-loading of the ice sheets in the northern hemisphere that formed during the last glaciation—exerts a strong influence on sea level relative to land. Land uplift is stronger in the northern Baltic Sea, attaining rates close to 10 mm $year^{-1}$, whereas in the southern Baltic Sea, it is close to equilibrium with some areas sinking by about 1 mm $year^{-1}$. Analysis of individual records of land-locked tide gauge measurements corrected for the vertical land movements indicate that Baltic Sea sea level may have risen during the twentieth century at rates of around 1.5 mm $year^{-1}$, which are close to the rate of global sea level rise. However, uncertainty is large because the margin of error of the estimates of land uplift—based on GPS data—may be locally of the order of 1 mm $year^{-1}$, and the time overlap between tide gauges and GPS measurements is short. An accurate estimate of absolute, climate-induced, Baltic Sea sea level rise over the twentieth century, is still not available, but is unlikely to deviate much from the global average. In more recent decades, as satellite altimetry data have become available, the basinwide rate of sea level rise may be around 5 mm $year^{-1}$ (with an uncertainty of roughly ±3 mm $year^{-1}$) with the central estimate thus higher than the recent global mean of 3.2 mm $year^{-1}$.

Sea level records display long-term developments that are still not fully explained. These include (1) the long-term widening of the annual cycle of sea level (winter maxima minus spring minima), (2) the more positive long-term trends of upper-range sea level measurements as opposed to lower-range sea levels—thus implying a long-term widening in the range of sea level observations, and (3) the time-varying correlation between sea level records and the main patterns of atmospheric climate variability, such as the NAO.

Extreme sea level is caused by storm surges driven by the passage of atmospheric cyclones. There is some evidence that the intensity of storm surges may have increased in recent decades in some parts of the Baltic Sea, and this has been attributed to long-term shifts in the tracks of some types of cyclone rather than to long-term change in the intensity of storminess. Analyses of storm surges have focused on local records, however, and there is no systematic basinwide analysis of change in storm surges yet available (see Chap. 4).

Since the first Baltic Sea Assessment (BACC Author Team 2008), new long-term series of instrumental wave data (longer than 20 years) and visually observed wave data (longer than 60 years) as well as several new high-resolution (∼3 miles) long-term wave hindcast simulations have become available. New estimates of the basic properties of the wave climate in the Baltic Sea as a whole and its sub-basins have been derived, such as distributions of various wave properties (periods, height), spatial patterns of long-term change and decadal variations on timescales of 20–30 years. Analyses show no significant change in average wave activity in the Baltic Sea basin. However, extensive spatial patterns of changes within the basin exist, possibly leading to long-term variations in areas with the greatest wave intensity. Regional studies have even revealed different trends in average and extreme wave conditions that are probably due to systematic change in wind direction.

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Part III

Future Climate Change

Introduction

Regional climate models have been used extensively since the first assessment of climate change in the Baltic Sea region published in 2008, not least for studies of Europe (which includes the Baltic Sea catchment area). Therefore, conclusions regarding climate model results have a better foundation than was the case for the first Baltic Sea assessment.

The general issue of model skill and validation is addressed in Chap. 10, both for dynamical and statistical downscaling methods. It presents the main sources of uncertainty for climate projections: parameterisation of sub-grid processes both in general circulation models (GCMs) and regional climate models (RCMs), limited amount and accuracy of input data, limited information on other climate relevant changes, and natural climate variability. For the Baltic Sea area the lack of an oceanic component in RCMs is a major restriction.

Chapters 11–14 report model-based projections of anthropogenic climate change in the Baltic Sea basin. The range of subjects includes results from numerical climate models as well as statistical downscaling studies regarding changes in the atmosphere (Chap. 11), and goes on to cover resulting consequences for the hydrological situation in the Baltic Sea basin (Chap. 12). Model-based projections of the state of the ocean are addressed in Chap. 13. In Chap. 14, the consequences of climate change on sea level in the Baltic Sea area are discussed. The timescale of the projections is normally around 100 years; usually a comparison of projections for the middle or end of the current century compared to conditions at the end of the twentieth century. This time period was chosen mainly because anthropogenic climate change on this time horizon is expected to be beyond the uncertainty range of natural variability.

10 Projected Change—Models and Methodology

Joanna Wibig, Douglas Maraun, Rasmus Benestad, Erik Kjellström, Philip Lorenz, and Ole Bøssing Christensen

Abstract

General (global) circulation models (GCMs) are a useful tool for studying how climate may change in the future. Although GCMs have high temporal resolution, their spatial resolution is low. To simulate the future climate of the Baltic Sea region, it is necessary to downscale GCM data. This chapter describes the two conceptually different ways of downscaling: regional climate models (RCMs) nested in GCMs and using empirical and/or statistical relations between large-scale variables from GCMs and small-scale variables. There are many uncertainties in climate models, including uncertainty related to future land use and atmospheric greenhouse gas concentrations, limits on the amount of input data and their accuracy, and the chaotic nature of weather. The skill of methods for describing regional climate futures is also limited by natural climate variability. For the Baltic Sea area, the lack of an oceanic component in RCMs and poor representation of forcing by aerosols and changes in land use are major limitations.

10.1 Introduction

The development of general circulation models (GCMs) has created a useful tool for projecting how climate may change in the future. Such models describe the climate at a set of grid points, regularly distributed in space and time and with the same density over land and ocean. Their temporal resolution is relatively high, but their spatial resolution is limited by computing power. Many important processes, such as cloud formation, convection, and precipitation, occur at spatial scales much smaller than the distance between grid points. This means that these so-called sub-grid processes are not explicitly simulated by the models, but must be approximated with simplifying algorithms referred to as parameterisations. The low spatial resolution also means that the topography, coastline, and processes at the land–air, ocean–air, and land–ocean boundaries are coarsely represented in GCMs.

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_10

The resolution of present-day GCMs, defined as a distance between two neighbouring points, is of the order of 100–300 km. However, the skilful scale (i.e. the scale at which the climate models are able to capture climate features) is larger, at about 8 grid point distances (Grotch and MacCracken 1991; von Storch et al. 1993), so about 1000–2500 km. This means that GCMs are well able to simulate the atmospheric state at scales greater than the skilful scale in spite of providing values within a grid scale (von Storch et al. 1993). However, the work of Grotch and MacCracken (1991) was based on old models with a small number of vertical levels and simple ocean–atmosphere coupling. A more comprehensive discussion about the skilful scale issue was given by Benestad et al. (2008).

To generate estimates of regional climate, that is, at a scale smaller than the skilful scale, it is necessary to downscale GCM results. Downscaling is understood as a process linking large-scale variables with small-scale variables. There are two conceptually different ways of downscaling. The first uses regional climate models (RCMs) nested in GCMs. RCMs have much higher resolution and can describe local features better, but are still able to simulate the atmospheric state in a realistic manner in their skilful scales. The second uses empirical and/or statistical relations between the large-scale results from GCMs and small-scale variables that describe regional and/or local climate conditions.

Climate projections differ significantly from weather forecasting. Forecasts cannot predict weather with high accuracy beyond a few days. Numerical weather forecasts take observations as a starting point. The number of observations is limited as is the accuracy with which they are made. Small disturbances in the data can cause a large effect on weather after some time. Lorenz (1963) referred to this as the ‘butterfly effect’. Climate models are not concerned with weather on a particular day or month or even year but with the statistical features of states of the atmosphere over long periods.

There are also other differences between weather and climate. Weather is forecast for a relatively short time—a few days, generally less than two weeks. This is because changes in weather are caused mainly by changes in the atmosphere. Even changes in oceanic processes have only a very limited influence on the weather because of the longer timescales of typical processes occurring in the oceans. In the case of climate, however, other factors must be taken into account. Climate variations are also caused by changes in the environment: ocean, vegetation, ice, sun, and the composition of the atmosphere. Some of these can be predicted with high accuracy, while others cannot. Among those that cannot are land-use change and the composition of the atmosphere, especially in relation to greenhouse gases (GHGs) and aerosols. As future climate change is to a high degree related to the extent of change in these environmental variables, predicting the future climate requires reliable estimates of the future composition of the atmosphere and land use. As the concentration of GHGs and aerosols in the future atmosphere is so difficult to predict because of the many influencing factors, scenarios are developed based on projections of the future evolution of the world population and economy (see Chap. 11, Sect. 11.2) and it is these scenarios that are used as the basis for projections of future climate.

Beside the uncertainty related to the limited information on land use, and the atmospheric concentrations of GHGs and aerosols, there are also other sources of uncertainty in models. These include limited amounts of input data and their limited accuracy. Due to the chaotic nature of the climate system, a very small difference in initial conditions can generate different climate features, as each simulation generates a different set of realisations. If this were the only source of uncertainty, the differences between simulations should remain within the range of typical climate variability. However, this is not the case. Many sub-grid-scale processes must be simulated in models in a more or less complex form and are not well described by the models. For example, simulations of cloud formation, their optical and radiative features, and the creation of precipitation still carry considerable model error.

For climate models to be useful, they need to be evaluated. As future climate predictions cannot be evaluated by direct comparison with observations, models are evaluated by comparing simulations with observations of the past climate. In theory, this should make it possible to select the best model, but this is not the case in practice. One model can usually describe a particular parameter better than another model, while the second model better describes a different variable or even the same variable, but in another part of the world. There are no objective ways to choose the best model, because none are able to exactly reproduce the observed mean climate and its variability. Differences between simulations and real climate data can be estimated on the basis of a so-called reference period (in the past) for which observational data are available. The differences, usually referred to as ‘biases’, vary in space and typically also in daily and annual cycles.

The models describe climate at a set of grid points. Because of numerical constraints in GCMs and RCMs, model results at neighbouring grid points are more correlated than actual measurements from two observation points at the same distance (Déqué 2007). This is one reason why the distributions of simulated variables are usually smoothed in comparison with measured station data. Simulations tend to underestimate the highest values and overestimate the lowest (Déqué 2007). This means that the bias is different in different parts of the distribution.

There are a number of sources of uncertainty in climate projections, and thus, preparing scenarios for future change in climate variables is a big challenge. No single method can be used for all variables and all regions.

Natural variability is an important source of uncertainty in climate projections (Deser et al. 2012a). The term 'natural climate variability' refers to variations in climate unrelated to human influences (BACC Author Team 2008). Deser et al. (2012a) analysed how the amplitude of natural variability varies with location in North America. There has been no similar study for the Baltic Sea basin, but these results are also relevant for this region. The analysis showed that natural variability is generally smaller in summer than that in winter and at lower latitudes rather than at higher latitudes. Also, that regional averaging does not always reduce the uncertainty in climate projections (Deser et al. 2012a). Natural climate variability cannot be reduced by better models, downscaling techniques or improved GHG emission scenarios and as a result limits climate predictability. However, natural climate variability can be described and to some extent quantified in an ensemble approach (see Deser et al. 2012a).

10.2 Dynamical Downscaling

The methodology used to achieve climate simulations in high resolution for a specific region by applying RCMs is referred to as 'dynamical downscaling'. RCMs are based on atmospheric limited-area models used in numerical weather prediction. The first application of RCMs for long-term simulations goes back to the work of Dickinson et al. (1989) and Giorgi and Bates (1989). Today, RCMs are used by many institutions and have been applied for a large number of studies, and RCM climate change projections have been undertaken for regions on all continents. There are several recent reviews of RCM methodology and their application (e.g. Giorgi 2006; Foley 2010; Rummukainen 2010).

10.2.1 Methodology for Dynamical Downscaling

Owing to limitations in computational power, the spatial and temporal resolution of GCMs covering the whole globe cannot be refined arbitrarily. For long-term climate change simulations, state-of-the-art GCMs can go down to nominal horizontal resolutions of about 100 km on current supercomputing systems. As atmospheric systems can be resolved only within several grid boxes, their effective resolution is much coarser, however. Therefore, GCMs can simulate large-scale climate features (i.e. synoptic lows), but not mesoscale atmospheric features (e.g. regional winds generated by mountains), which are necessary for a realistic simulation of regional climate.

Consequently, the principal concept of RCMs is to perform long-term climate change projections with an increased spatial resolution (down to about 50–10 km) for a specific region of interest only. RCMs are limited-area versions of three-dimensional atmospheric circulation models, which in principle use the same set of dynamic equations and physical parameterisations as GCMs. Like GCMs, they include for land grid points a model describing the thermodynamic properties of the upper soil levels. The main difference between RCMs and GCMs (apart from sometimes different parameterisation schemes) is their lateral boundary, as they do not work globally. Because the RCM does not have any information outside its modelling domain, it needs to be provided with information about the atmospheric state at its lateral boundaries, the so-called lateral boundary conditions (LBC). In contrast to GCMs, the solution of an RCM consequently transforms from an initial-value problem into a lateral boundary value problem for longer integration times. The information at the lateral boundaries is taken from the output of the 'driving model', which can be a GCM, a global (re-) analysis, or—when using a 'double nesting' technique—from RCM output simulated on a larger domain in coarser resolution. In order to provide a smooth transition and to avoid numerical problems, a careful LBC treatment is essential for RCM integrations. In the early 1970s, Davies (1976) invented the 'sponge zone', a zone of around 5–10 grid boxes at all lateral boundaries, in which the LBC and the internal solution of the RCM are merged with decreasing weight of the LBC from the boundary towards the centre of the domain. This kind of treatment of the lateral boundaries is still used in most RCM simulations. Additionally, at the lower boundary over sea areas, values for sea-surface temperature (SST) and ice coverage have to be prescribed during the integration. This information is mostly extracted from the driving model like the LBC, as most RCMs are still pure atmospheric models without a coupled ocean component.

10.2.2 Performance of RCMs in Reproducing Recent Climate

A benchmark test for RCMs is that they can reproduce the main features of the climate of the past few decades when forced with realistic boundary conditions. In this respect, it is common to evaluate simulations in which RCMs have been downscaling reanalysis data. Extensive model evaluation has been undertaken for single RCMs (e.g. Samuelsson et al. 2011) or for a large number of models (e.g. Christensen et al. 2010). However, studies on RCM performance focusing on the Baltic Sea region remain few (e.g. Lind and Kjellström 2009). This section therefore presents results from a range of

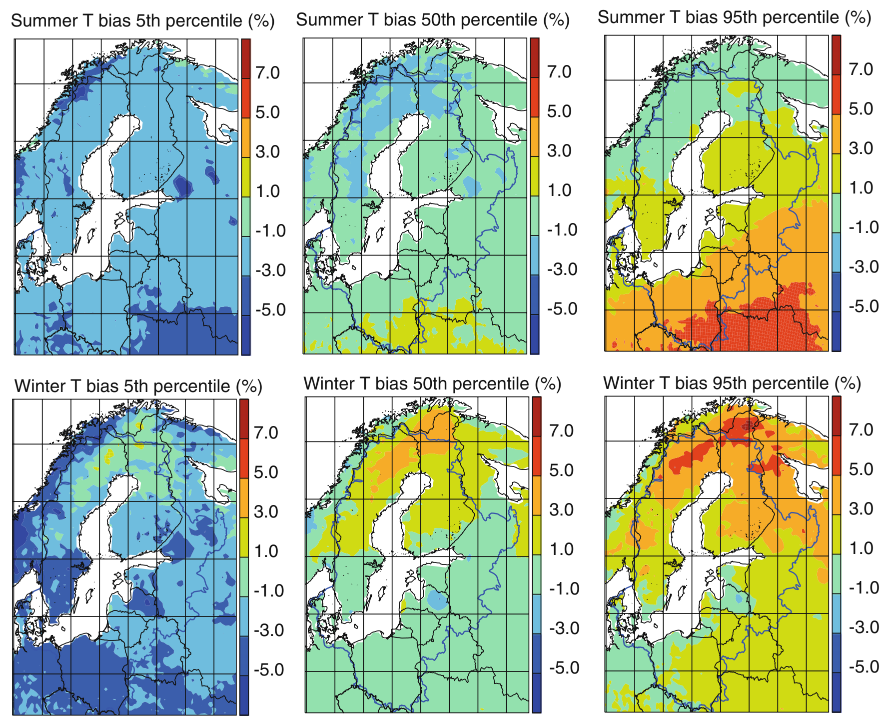


Fig. 10.1 Simulated mean temperature bias with respect to the daily gridded observational data set based on European Climate Assessment & Dataset information (E-OBS) for 1961–2000. The maps show the pointwise smallest (*left*), median (*middle*), and largest (*right*) biases from an ensemble of nine RCMs with lateral boundary conditions from ERA-40 reanalysis data. *Upper row* shows summer (JJA) biases and *lower row* winter (DJF)

RCMs from the ENSEMBLES project (Christensen et al. 2010) to illustrate the degree to which current RCMs can reproduce the recent past climate. The nine RCMs are listed in Chap. 11 and Table 11.1, and this section presents the results for their forcing by ERA-40 reanalysis data (Uppala et al. 2005) rather than GCM output at their lateral boundaries. As example of model performance, this section shows comparisons of how the ensemble of ERA-40-driven simulations reproduces seasonal mean temperature (Fig. 10.1) and precipitation (Fig. 10.2) for the Baltic Sea region with respect to the daily gridded observational data set based on European Climate Assessment & Dataset information (E-OBS) (Haylock et al. 2008). Nine RCMs were used: C4IRCA3, KNMI-RACMO2, DMI-HIRHAM5, ETHZ-CLM, HadRM3Q0, HadRM3Q16, MPI-REMO, HadRM3Q3, and SMHIRCA (for documentation on the individual models, see Christensen et al. 2010; data are available from http://ensemblesrt3.dmi.dk/). The maps show grid-point-wise model performance, and as an estimate of the spread, the nine sets of results for each grid point are sorted resulting in an approximate 5th percentile corresponding to the lowest value, a median, and an approximate 95th percentile corresponding to the largest value.

In summer, the temperature climate is reproduced to within ±3 °C in all models in most of the region (Fig. 10.1). An exception is the southernmost part where maximum errors are greater than 5 °C in the warmest (95th percentile) model. Another exception is the relatively large local negative biases found over the big Russian lakes: Lake Ladoga and Lake Onega. These biases are unlikely to be real but probably reflect that the E-OBS data build on land-based observations, while some of the RCMs include lake models. Such a lake model has the effect of delaying the summertime maximum temperature by about one month, implying that the June–July–August average is lower compared to the surrounding land areas (Samuelsson et al. 2010). In the north, on the other hand, no models overestimate the temperatures indicating a systematic cold bias in most models in that area. In winter, most models tend to be too warm in parts of the northern basin indicating a too weak annual cycle, while in the south, there are both models over- and underestimating temperature. An interesting feature is the local cold bias in eastern Latvia. As Christensen et al. (2010) pointed out, there is no reason why the RCMs should have a local bias like this and it may therefore indicate that it is in fact the E-OBS data that are biased.

Precipitation seems to be overestimated in the Baltic Sea region in most RCMs, both in winter and summer. An exception is again the southern part of the basin where there are models with a dry bias in summer. The driest model is also the model with the largest positive bias in temperature, indicating a possible feedback between precipitation, soil moisture, and temperature. The biases in wintertime precipitation are apparently large, in the wettest models more than 50 % in most of the region. However, it should be noted that the wintertime observations of precipitation in this area may be biased due to undercatch related to snow and wind (e.g. Rubel and Hantel 2001). A local overestimation

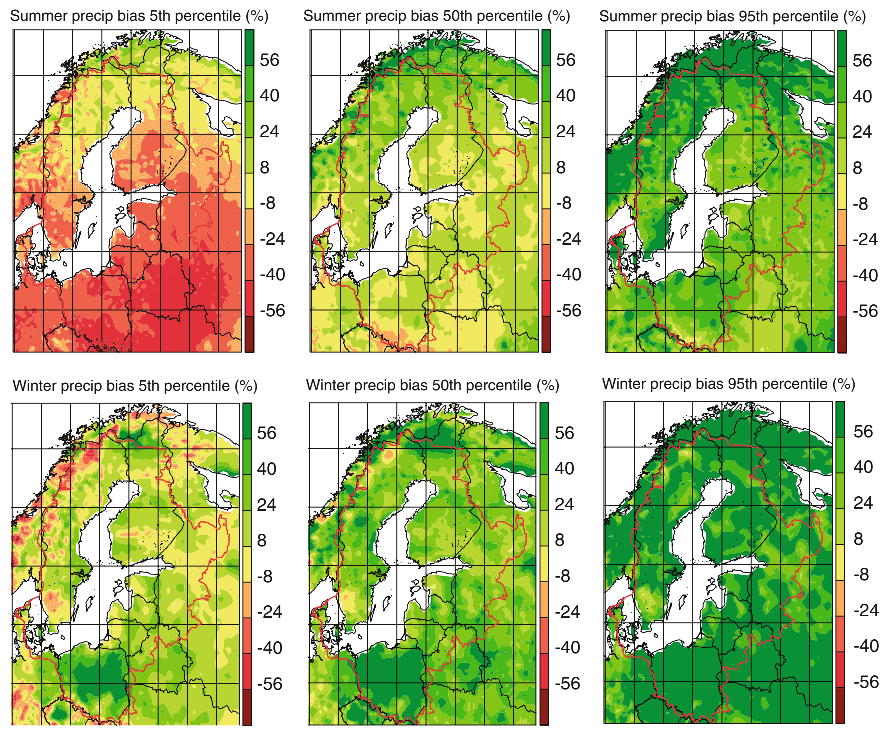


Fig. 10.2 Simulated precipitation bias with respect to the daily gridded observational data set based on European Climate Assessment & Dataset information (E-OBS) for 1961–2000. The maps show the pointwise smallest (*left*), median (*middle*), and largest (*right*) biases from an ensemble of nine RCMs with lateral boundary conditions from ERA-40 reanalysis data. *Upper row* shows summer (JJA) biases and *lower row* winter (DJF)

appears over large parts of Poland (Fig. 10.2). But, as discussed by Christensen et al. (2010), this could also be the result of a bias in the observations as there is no reason why the RCMs should show such a strong local deviation from the surrounding areas (see Chap. 11).

10.2.3 Developing and Extending RCMs

In global climate projections, coupled atmosphere–ocean models are state of the art (Meehl et al. 2007). However, RCM climate change projections are in general still carried out for the atmosphere only, prescribing SST data taken from the driving model (Christensen et al. 2007; Kjellström et al. 2013). Consequently, the quality of the prescribed SST/sea ice data depends on the quality of the global modelling system. In particular, for a relatively small and semi-enclosed sea like the Baltic Sea, data quality might be limited by the coarse resolution of the global ocean component. Figure 10.3 shows the land–sea mask of the global ocean model MPI-OM in grid resolution 1.5° (GR15), as used as one of the main coupled atmosphere–ocean GCMs (AOGCMs) for driving the RCM model suite within the EU project ENSEMBLES (www.ensembles-eu.org). A better representation of the water body of such oceans can be generated by the use of high-resolution regional ocean components, which can be coupled to the atmospheric RCM (analogue to global coupled model systems).

Pioneering work in this area has been done for the Baltic Sea region. This includes establishing atmosphere–ocean–sea ice models (e.g. RCAO, Döscher et al. 2002), some including additional river routing schemes allowing the

Fig. 10.3 Land–sea mask of the global ocean model MPIOM in resolution GR15 (courtesy of M. Böttinger, DKRZ)

modelling of the complete hydrological cycle (e.g. BALTIMOS, Lehmann et al. 2004). Recently, Meier et al. (2011) showed that a coupled RCM of this type (RCAO) has the potential to improve the results in downscaling experiments driven by GCMs considerably because SSTs and sea ice concentrations are more realistic than those taken directly from the driving GCM. This adds a major caveat to the utility of other downscaling methods relying on SSTs from GCMs in the area, for example the RCM simulations from the ENSEMBLES project.

In conventional RCM simulations, the driving model data are used only in the lateral boundary zone, while in the inner model domain, the RCM is not forced to the driving model. This causes an ill-posed boundary value problem and can lead to a different large-scale flow in the RCM simulation with respect to the driving model. In a perfect boundary setting, this 'freedom' of the RCM may lead to considerable deviations between the real and simulated local climate state (e.g. Winterfeldt and Weisse 2009). With a nudging technique, the solution of the driving model can be prescribed for the whole RCM domain. However, by a scale-independent nudging method, the desired small-scale circulation features generated by the RCM would also be suppressed. In order to circumvent this clear disadvantage, the method of 'spectral nudging' was introduced (von Storch et al. 2000; Feser et al. 2001), in which just the large-scale circulation is relaxed towards the driving model in the inner RCM domain, while the small-scale circulation remains untouched (large-scale constrained RCM simulation). This method leads to a system where empirical data (i.e. large-scale flow and surface details) are systematically combined with the theoretical understanding (i.e. the RCM). While the spectral nudging technique is now becoming more popular (e.g. Miguez-Macho et al. 2004; Castro et al. 2005), a debate on this technique is still ongoing; however, improvements through the application of spectral nudging are evident when the driving model represents a realistic large-scale flow, such as using reanalysis data as the driving model (Winterfeldt and Weisse 2009). In contrast, in the situation with a coarse GCM having an unrealistic large-scale circulation (caused by poorly represented topography due to the coarse resolution) as the driving model, even an RCM using spectral nudging could not alter the prescribed large-scale flow. In regions with complex terrain, the simulated flow in the reanalysis data and therefore the nudging constraints might themselves be biased. In such situations, spectral nudging might lead to unrealistic local-scale flow (Radu et al. 2008). The ability of a non-nudged RCM to improve the large-scale climate inside its domain can be evaluated using the 'Big Brother' approach (Denis et al. 2002). In this approach, a reference climate is established by performing a large-domain high-resolution RCM simulation termed 'the Big Brother'. Here, the short scales are filtered out and this filtered reference is used to drive the same nested RCM (the 'Little Brother') integrated in the smaller domain but with the same resolution. Differences between climate statistics of both can be attributed to errors associated with the nesting and downscaling technique, allowing them to be distinguished from model errors. This ability is model and region dependent.

At present, most RCMs still use the hydrostatic approximation, assuming the vertical structure to be in hydrostatic equilibrium, and consequently neglecting vertical acceleration. This assumption is valid for nominal horizontal resolutions roughly above ~10 km. Most current RCM climate change projections still use coarser nominal horizontal resolutions, between 50 and 20 km (e.g. PRUDENCE and ENSEMBLES), but due to increasing computer power, the resolution of some RCM climate change simulations is increasing to about 10 km. The expected further increase in computational resources will presumably mean a further increase in RCM resolution, leading to the use of non-hydrostatic RCMs. Kendon et al. (2012) reported on a recent study showing results from dynamical downscaling with a non-hydrostatic RCM at 1.5 km grid spacing.

To date, climate change projections have been carried out in a one-way nesting mode, meaning that the RCM does not give information back to the driving model. The first studies of two-way nesting, allowing feedback from the RCM to the GCM (Lorenz and Jacob 2005; Inatsu and Kimoto 2009), indicate the potential for improving the driving global simulation, even in regions far from the two-way nested RCM domain (there are no examples demonstrating the two-way nesting approach for the Baltic Sea region available yet).

In addition to ocean models, lake models have also been coupled to RCMs. This is an important development for the Baltic Sea basin where a large number of lakes exist and a large fraction of the land area is covered by lakes. In a study with an RCM coupled to a lake model, Samuelsson et al. (2010) found that including lakes warmed the climate and that the largest warming occurred in autumn and winter in southern Finland and western Russia where differences of more than 1 °C were obtained.

In recent years, RCMs have begun to incorporate more processes. One example is the work of Wramneby et al. (2010) where a process-based model of vegetation dynamics and biogeochemistry has been coupled to an RCM. They showed that including dynamic vegetation that responds to climate change has an impact on the climate simulated. For the Baltic Sea region in particular, they found reduced albedo resulting from the snow-masking effect of forest expansion when dynamic vegetation is included. This leads to an enhancement of the winter warming trend.

10.3 Statistical Downscaling

Statistical downscaling is an approach that bridges the gap between model output (GCMs or RCMs) and regional or local-scale climate. Rummukainen (1997) distinguished between the model output statistics (MOS) methods and the perfect prognosis (PP) methods. The MOS methods find relationships between model simulations and observations in the historical (reference) period and then use them in future climate simulations (Wilby and Wigley 1997; Maraun et al. 2010). The PP methods identify empirical relationships linking large-scale atmospheric predictors and local/regional predictands. This relationship is assessed on the base of observations in the historical (reference) period and is then used in simulations of future climate.

10.3.1 Model Output Statistics

The biggest disadvantage of the RCM methodology is probably the occurrence of systematic biases in the present climate simulations. These systematic biases, also seen in GCMs, are because dynamic climate simulations carried out with GCMs and RCMs are bound only to changing atmospheric GHG concentrations. Due to their coarse resolution and parameterisations, GCMs and RCMs are not perfect; so even the mean climatological values produced by these models deviate from the corresponding observations. In RCM climate projections, the systematic biases are nonlinear combinations of the systematic errors of the driving GCM and the systematic errors of the RCM itself. Another limitation is that there is still the need to downscale area averages given as grid values in model output to point values necessary for impact studies (Xu et al. 2005). Given the discrepancies between observations and model results for present-day climate, a method is needed to cope with the biases. Given that good observation data sets exist, more realistic data sets of forcing fields incorporating the projected changes can be created and used for impact studies (Piani et al. 2010). This can be achieved through the methods known as MOS. These are statistical models linking simulated variables to observations. There are generally two groups of MOS methods: one is known as the bias correction method (Déqué et al. 2007; Piani et al. 2010), while the other is known as the perturbation of observed data (POD) or the delta change (DC) method (Hay et al. 2000; Lenderink et al. 2007a; van Roosmalen et al. 2011). A review of MOS methods was reported by Maraun et al. (2010).

10.3.1.1 Bias Correction Method

Validating models by comparison with observations makes it possible to quantify model biases, defined as differences in the mean as well as higher order statistical moments. An assessment of bias is the first step before using the model output to force impact models. Unfortunately, model bias is not uniform in space or time and so its identification needs long and homogenised data sets with high spatial resolution. A bias has a seasonal cycle, so its correction often means applying it to individual months or seasons separately. Because GCM and RCM outputs are given in a set of grid points, they are usually volume averages and cannot be directly compared with observations as these are point values. Volume averaging is a type of smoothing that makes high values lower and low values higher, so the range of volume averages is usually much lower than the range of point values. It means that the bias can vary also within different parts of the distribution.

Bias correction or scaling is based on the assumption that the statistical relationship between observations and RCM simulations for the present-day climate is the same as that between the future climate and RCM simulations of the future climate, which may not be true (Christensen et al. 2007; Boberg and Christensen 2012). The bias correction values are calculated by comparing observations with RCM simulations for the same period. Since two climates are compared—the real one and a simulated one—the study period should be relatively long, covering at least 30 years. The corrections can be additive or multiplicative, depending on the variable.

In some cases, the impact models need only seasonal or monthly mean values. Then, it is enough to compare long-term means of observations and RCM simulations for the present-day climate (Schmidli et al. 2006; Graham et al. 2007b). The corrections calculated for RCM simulations under the present-day climate are then applied to the RCM simulations for the future climate to generate more accurate future scenarios. In many cases, the bias correction factors are considered individually for different intensities (i.e. parts of the variable distribution). This is sometimes referred to as distribution-based scaling (DBS; Yang et al. 2010; van Roosmalen et al. 2011). Déqué (2007) and Piani et al. (2010) gave a detailed description of the method. It is generally a quantile mapping approach, where quantiles are empirical cumulative distribution functions or statistical distributions fitted to simulated and observed data.

10.3.1.2 Perturbation of Observed Data

The second MOS method is the DC or POD method (Hay et al. 2000; van Roosmalen et al. 2011). In this approach, the long-term mean additive or multiplicative change factor is calculated on the basis of an RCM projection of the future and present-day (reference) climate and applied to the observation record (Yang et al. 2010; van Roosmalen et al. 2011). These factors can differ seasonally and for different part of the frequency distribution (Olsson et al. 2009). In the DC method, there is no need to identify the bias. Instead, the absolute or relative delta change factors (DCF) are assessed by comparing the climate model outputs representing present-day and future climate (Semadeni-Davies et al. 2008; Olsson et al. 2009). The observed variable is then rescaled and used as input for impact models.

One of the differences between MOS and other statistical downscaling methods is that MOS calibration is specific to the numerical model for which it has been developed and cannot be used with other numerical models (Maraun et al. 2010). Calibrations can be based on RCM driven by reanalysis or GCM climate simulation forced by external factors. In the first case, there is a direct correspondence between simulated and observed variables; in the second, only statistics of simulated and observed variable distributions can be compared. Because, in this second case, the simulated climate is one randomly selected from many possible choices it is always a risk that the bias determined is an artefact generated by this random choice. On the other hand, calibrating with a reanalysis-driven RCM is of little use, because no reanalysis of the future is available.

10.3.2 The 'Perfect Prognosis' Approach

The PP approaches establish the statistical relationship between large-scale predictors and regional or local-scale predictands. The local variable of interest, denoted by y, depends not only on the large-scale predictors $\boldsymbol{X}$, but also on the local geographic parameters denoted by g. Mathematically, this can be expressed as follows:

$$y = f(\boldsymbol{X}, g) + \eta,$$

where η means a residual noise term.

Figure 10.4 illustrates how the local conditions depend on both the geography and the large-scale situation. In this case, the snow only stays where the temperature is below freezing, which is only above a certain altitude. Furthermore, the large coherent extent of the snow shows that the local temperature is part of a larger pattern. Although the exact value of y may vary from location to location (small-scale noise η), it is possible to say from this photograph that the temperature in the snow-covered region shown is mainly below freezing. In this example, the large-scale condition $\boldsymbol{X}$ is the snow cover, but it is better to use a predictor with a more direct physical relationship to the predictand. $\boldsymbol{X}$ can often be the mean sea-level pressure (SLP) or the large-scale temperature pattern.

Two steps can be distinguished in the downscaling procedure: the identification of large-scale predictors and the development of a statistical model linking the local predictand with the large-scale predictors.

There are four requirements that the predictors should fulfil. Most important is the existence of a strong statistical relationship between predictors and predictand, typically manifest by high-correlation coefficients. The relationship between predictors and predictand needs to be stable over time. Suitable predictors should also be reasonably well

Fig. 10.4 The Rondane mountain range in Norway during autumn, illustrating how local conditions such as snow cover depend on both geography and large-scale weather (*photograph* R.E. Benestad)

simulated by GCMs. For climate change analysis, it is important that predictors capture the global warming signal (Wilby et al. 2000).

The predictor, being a large-scale variable, is defined at a huge number of grid points. It is therefore convenient to reduce this dimensionality, because there is usually a high correlation between values at neighbouring grid points. One way of doing this is to decompose the field variable into a smaller number of modes of variability. The large-scale variability can be described in terms of orthogonal empirical functions (EOFs) (Lorenz 1956; North et al. 1982; Benestad 2001). The spatial structures of EOFs describe a set of spatially coherent 'modes' that describe the variations in the gridded data. The leading modes describe the structures that are most pronounced and have the greatest spatial scales, and the higher order modes are associated with less variance and smaller spatial scales.

Often, only a small number of leading EOFs describe the major part of field variability (Wilks 1995). It is therefore possible to describe the main features of the gridded data in terms of a relatively small number of EOFs. Each spatial EOF pattern is associated with a vector of weights, describing how strongly this pattern is present at any time of the record. This time series is often referred to as a 'principal component' (PC). The PCs are the basis for the downscaling model calibration, for instance a multiple regression against the predictand. The benefit of using EOFs is that they are orthogonal and make the model calibration easier and more robust (no co-linearity).

The reduction in dimensionality can also be obtained by a transformation of field values into other indices. In the case of SLP fields, these can be the indices of zonal and meridional flow, vorticity, or other indices, such as the North Atlantic Oscillation index (Conway and Jones 1998; Wilby and Wigley 2000). Weather types represent another type of transformation. Here, the large-scale field, usually SLP or geopotential height, is mapped into a set of categories—weather types—by a clustering algorithm like k-means.

The transformation procedure should generate a predictor that has high predictive power, that is, explains a high percentage of the variability of the predictand. Some methods, such as canonical correlation analysis (CCA) or the singular value decomposition (SVD) method, directly seek the modes having the highest correlation or covariance with the predictand field, while others do not.

10.3.2.1 A Brand of Calibration Strategies

The brand 'PP methods' describe a class of empirical–statistical downscaling models that involve a specific strategy for model calibration (Wilks 1995). These use observations [raw and gridded data, or re-analyses (Kalnay et al. 1996; Simmons and Gibson 2000)] to calibrate against an observed predictand. First, a predictor is taken from historical data, and then, a relation is found with the predictand (downscaling model calibration). Then, the climate model results are compared with the predictors used to calibrate the downscaling model, and steps are taken to ensure that the model results correspond with the calibration data (e.g. through a regression analysis). The PP method may involve linear and nonlinear methods.

10.3.2.2 Regression Methods

Regression models include linear and nonlinear relationships between predictors and the predictand (Benestad et al. 2008). Among them are the multiple regression (Murphy 1999), the CCA method (Busuioc et al. 1999), and the SVD method (Bretherton et al. 1992). The difference between these approaches is that the multiple regression minimises the root-mean-square errors (distance between predictions and observations), the CCA maximises the correlation, and the SVD maximises the covariance between two fields. Artificial neural networks also represent nonlinear regression models (Crane and Hewitson 1998).

10.3.2.3 Weather Classification Methods

The weather classification methods involve various strategies, such as analogues (Zorita and von Storch 1999; Timbal et al. 2008), circulation classification schemes (Bárdossy and Caspary 1990; Jones et al. 1993), cluster analysis (Corte-Real et al. 1999; Huth 2000), and neural nets. The analogue model involves searching the record of past events and taking the day that most closely matches the situation wanted to predict. Cluster analysis bases the predictions on a number of closest states (Wilks 1995), either by taking the mean of the days with close matches or by using the observed values for all days that match the predicted state, and constructing a statistical distribution (histogram). From this sample, or a fitted probability density distribution, a random value may be drawn. Neural nets involve various adaptive learning algorithms, such as 'artificial intelligence' (Wilby et al. 1998; Hewitson and Crane 2002). The analogue model, circulation classification schemes, and cluster analysis all involve a re-sampling of past measurements. These re-sampling techniques suffer from one caveat that the tails of the distributions will be distorted because the sampling cannot produce new record-breaking values (Benestad 2008). Even stationary series are expected to produce new record-breaking events, given sufficiently long intervals for observations. Theory of independent and identically distributed (iid) series shows that the expected occurrence of new record-breaking events will converge towards zero, but never actually become zero. Nevertheless, this implies that the upper and lower tails of the distribution of the results from the re-sampling methods may be distorted and that the results may have to be re-calibrated. A re-calibration can be performed once the theoretical probability distribution function is known through local quantile mapping.

10.3.2.4 Weather Generators

Stochastic weather generators are statistical models producing high-resolution local-scale time series of a suite of elements such as temperature and precipitation among others, whose large-scale statistics follow the required criteria (Richardson 1981; Wilks and Wilby 1999; Olsson et al. 2009; Willems and Vrac 2011). Among many applications, they can serve as a computationally effective tool to produce site-specific data sets at the required time resolution (Semenov et al. 1998).

The distribution used is usually different for different climate variables. For temperature, the normal distribution is the most popular (Semenov et al. 1998). More complicated is the generation of precipitation data, and different functions are used. Among the most popular are the Markov chain, the semi-empirical, and the Neyman–Scott rectangular pulse (NSRP) weather generator. In the Markov chain generator, precipitation occurrence and totals are produced separately (Sunyer et al. 2012). Two states are possible: wet or dry days. The amount of precipitation on a rainy (wet) day is most often generated using a gamma or exponential distribution (Benestad 2007). In the semi-empirical generator, a few distributions can be defined, for instance for wet and dry spell lengths and precipitation amount. In the NSRP weather generator, Kilsby et al. (2007) proposed four different steps. A storm origin is described by the Poisson process. Separate rain cells within a storm are separated by time intervals taken from exponential distribution. The duration and intensity of each rain cell are also described by exponential distributions, and their sum gives a rainfall total.

Weather generators can be used when the observation records are relatively short. They can also supply many weather 'realisations' having the same overall statistics. A wide suite of statistics can be used to fit the model: mean, variance, skewness, autocorrelation, and many others. Weather generators can also serve to produce data for locations where there is information about the statistical distribution and time structure. For places with only short records of high-temporal-resolution data but longer series with data of low resolution, it is possible to use information from the longer records to make inferences about the distributions, and it is in principle possible to produce projections for temporal scales higher than those usually produced by RCMs (6 h).

10.3.2.5 Randomisation

Models generally underestimate the local-scale variance. To resolve this, Karl et al. (1990) proposed the use of a scaling factor to ensure that the variance of the projected surface values will match the observed variance. But this could increase the error of the estimates, a phenomenon called 'inflation'. Von Storch (1999) argued that this was not a good method because of the need to relate the variance of the predictor to the variance of the predictand. Instead, this author proposed a randomisation method that relied on adding a noise (not necessarily a white one, a random signal with a constant power spectral density was adequate). Another method of resolving the issue of underestimating local-scale variance was developed by Bürger (1996) and called the 'variance-optimised' version of expanded downscaling. Bürger and Chen (2005) compared all these methods. They found that inflation for multi-site downscaling did not describe spatial correlation. Randomisation has a problem with simulating variance in a future climate. The Bürger (1996) method is very sensitive to the quality of normalisation.

10.4 Ensembles, How to Use Them and How to Assess an Error of Projection

All techniques developed to derive regional-scale climate information are associated with uncertainties. This is true both for the direct use of global climate model output and for information emanating from dynamic or statistical downscaling techniques. Uncertainties related to forcing, climate sensitivity, and natural variability can, at least to some degree, be treated by utilising climate change information from ensembles including a large number of climate change experiments (Benestad 2011).

10.4.1 Different Types of Ensembles

Ensembles of climate change simulations can be constructed such that they sample different GCMs with different climate sensitivity under different GHG emission scenarios starting from different initial conditions. Such climate change experiments could be performed by the use of multi-model ensembles (e.g. van der Linden and Mitchell 2009). Under a given forcing scenario, the spread between the different model results can then be taken as an indicator of uncertainty related to structural differences between models, differences in parameterisations, and different initial conditions. In total, there are around 20–30 different coupled AOGCMs worldwide that can constitute such a multi-model ensemble (status as of 2012).

A problem in the context of uncertainty is that different climate models are not totally independent of each other but rather share parts of the code. This means that any multi-model ensemble will contain members that are related to each other. Furthermore, the degree of freedom in a GCM is very large, implying that even if all different GCMs are used, the full range of model uncertainty will not be sampled by a multi-model ensemble. As an alternative, perturbed physics ensembles with a much larger number of ensemble members have been developed (e.g. Murphy et al. 2007). In these ensembles, one model is used as a reference. In addition to the reference simulation, a large number of simulations with the same model are performed where one or more of the model parameters have been altered within their uncertainty bounds. In this way, the parametric uncertainty can be addressed along with the uncertainty related to initial conditions.

Even if the number of simulations is much larger in a perturbed physics ensemble compared to that in multi-model ensembles, this type of experiment will not sample the structural differences between different GCMs, and therefore, the full model uncertainty is not sampled by perturbed physics ensembles, either. Recently, comparisons have been performed between perturbed physics ensembles based on different GCMs (Yokohata et al. 2010) and between perturbed physics ensembles based on one GCM and multi-model ensembles (Collins et al. 2011). In the ENSEMBLES project (van der Linden and Mitchell 2009), uncertainties due to structural effects as determined from the multi-model CMIP3 GCM ensemble were added to the parametric uncertainties from the HadCM3 perturbed physics ensemble to yield a total uncertainty that could be used in the production of probabilistic climate change projections. In both multi-model ensembles and perturbed physics ensembles, it is not possible to distinguish between uncertainty related to model formulation and that related to initial conditions unless several ensemble members sampling also initial conditions are performed for each multi-model or perturbed physics ensemble member.

10.4.2 Are Ensemble Projections Better Than Those Based on Single Climate Projections?

The multi-model ensemble means have been shown to outperform the single model simulations. This has been shown to result from the fact that models are overconfident, that is, they have a too small spread in the ensemble, centred at the wrong value (Weigel et al. 2008). The good performance of the multi-model ensemble means holds true in a general sense, although for individual variables, seasons, and regions, it is possible to find single models that are better than the ensemble mean. This has been shown in a number of studies at the European scale based on RCMs downscaling reanalysis data in the ENSEMBLES project (e.g. Kjellström et al. 2010; Lenderink 2010; Lorenz and Jacob 2010). This is also illustrated for the Baltic Sea region in Figs. 10.1 and 10.2. There is no reason why the ensemble mean (or median in this case) should systematically show the smallest biases. For instance, the warmest model is better at reproducing the temperature in the far north in summer and the coldest model is better in the north in winter. Similarly, the driest model appears to outperform the ensemble average in summertime precipitation in the far north. A practical problem here is that different models perform best for different aspects; no one model performs best for everything (e.g. Christensen et al. 2010). This makes it difficult to know which model to choose and favours the use of the multi-model ensemble mean over the results of any single model.

10.4.3 Performance-Based Weighting of Ensembles

Climate models differ in their agreement with observations. The idea of performance-based weighting of ensembles is to utilise these differences to derive weights that can be applied when results from different models are to be combined in a common climate change signal. The rationale would be to give models with a better agreement to observations greater weight than those with less good agreement. However, there are a number of issues. For example, a model can have a good agreement for one variable but not for others, for one season but not for others, and the agreement can be due to compensating errors, etc. Furthermore, any performance-based weights will need to be calculated based on agreement in past decades and so are not necessarily applicable to future climate conditions. Also, regardless of how objective the methods used to derive weights are, there is a high degree of subjectivity as to which metrics to use and what observational data should be used in the analysis (e.g. Christensen et al. 2010).

In the ENSEMBLES project, a weighting system was designed and tested. It consists of a combination of a series of weights derived from evaluating different aspects of RCM performance. These aspects include reproduction of large-scale atmospheric circulation patterns, mesoscale patterns, daily temperature and precipitation distributions and extremes, trends, and the annual cycle (Christensen et al. 2010). Christensen and co-workers found no compelling evidence of an improved description of mean climate states when the weights were used. Furthermore, they concluded that using model weights added another level of uncertainty to the generation of ensemble-based projections. A particular problem related to RCM ensembles was that the underlying GCM simulation largely governed the results. Application of weights that are determined for RCMs in reanalysis-driven simulations (Christensen et al. 2010) on GCM-driven simulations with the same RCMs may therefore not lead to an improvement in the overall ensemble skill (Déqué and Somot 2010).

10.4.4 Design and Use of GCM-RCM Ensemble Regional Climate Projections

Traditionally, climate change ensembles are 'ensembles of opportunity', that is, they are the result of a compilation of more or less coordinated climate change experiments. This means that there have not been any deliberate attempts to design the ensemble so as to sample uncertainty in any specific way. Recently, however, there have been some attempts to design GCM-RCM ensembles in order to sample various kinds of uncertainty in a more systematic way. The PRUDENCE project mainly addressed uncertainty related to RCM formulation with 11 RCMs downscaling one and the same GCM under the same GHG emission scenario, but there were also other GCMs and emission scenarios included in that project (Christensen and Christensen 2007). Based on these results, Déqué et al. (2007) concluded that uncertainty in future European climate change is generally more associated with the choice of GCM than with which RCM is used, particularly for temperature. Consequently, in the ENSEMBLES project, there was an emphasis on having a larger ensemble with more GCMs involved (van der Linden and Mitchell 2009). In a recent study, Déqué et al. (2012) investigated sources of uncertainty in the ENSEMBLES GCM-RCM ensemble. This new study confirmed the results of Déqué et al. (2007) in that the choice of GCM is the dominant source of uncertainty. But there are exceptions, such as for summertime precipitation, when it is RCM formulation that may be the dominant source of uncertainty. Other examples of GCM-RCM ensembles involve ensembles with the Norwegian RCM sampling several GCMs (Haugen and Iversen 2008) or the Swedish RCM sampling a range of different GCMs under different GHG emission scenarios and in some cases with different initial conditions (Kjellström et al. 2011). Based on the results from the ENSEMBLES simulations and the Swedish model, Kendon et al. (2010) also concluded that sampling GCM uncertainty is most important, but RCM uncertainty also needs to be sampled, at least for some regions and seasons.

10.5 Validation Techniques

Any downscaled simulation of present-day climate or a future climate scenario is a more or less simplified representation of reality. A validation against observational data is therefore crucial to assess the quality of the simulation, in particular for a further use in impact studies. To this end, a set of indices is usually derived to describe the properties of interest from the reference data set and the model simulation to be validated. Agreement between the reference and the model is quantified by suitably chosen measures. As discussed in Sect. 10.4, the errors and uncertainties of downscaled climate simulations arise from an imperfect model formulation, uncertain future concentrations of GHGs, and internally generated climate variability. In a downscaling context, the uncertainty due to imperfect model formulation originates from three parts: errors of the driving GCM, errors inherent in the downscaling approach, and errors in observations themselves. The first two types of error are of interest in the validation.

When validating a downscaling system with boundaries from a GCM against observational data, the combined GCM/downscaling error can be evaluated. The influence of the driving GCM on the downscaled simulation can be assessed by combining a single downscaling method with different GCMs and then comparing the different results (e.g. Nikulin et al. 2011). In such a control run setting, care must be taken not to mix the model error and internal climate variability on long timescales. In particular, the estimation of extreme properties requires long time series and the typical 30-year period might not be long enough to gain robust estimates (Kendon et al. 2008). The downscaling error can be separated from the GCM error by driving the downscaling method with 'perfect boundary conditions' (Frei et al. 2003), that is, observational data or—as a proxy—reanalysis data. In a perfect boundary setting, the simulated and reference weather sequences are more or less synchronised, allowing for relatively short validation periods (although care should be taken not to be dominated by individual events). The nesting procedure for RCMs into large-scale low-resolution data at the lateral boundaries of the RCM domain is often supported by a spectral nudging technique that poses additional large-scale constraints onto the largest waves in the interior of the RCM domain (von Storch et al. 2000; Feser et al. 2001). To isolate the error due to nesting in dynamical downscaling in both control run and perfect boundary setting, an approach to separate different error sources in an RCM pseudo-reality, the Big Brother approach (see Sect. 10.2.3) can be used (Denis et al. 2002).

Before using a regional climate projection for follow-up studies, the assessment of not only the downscaling error but also the GCM error is essential, as misrepresentation of large-scale patterns (e.g. the position of the storm tracks) or temporal structure (e.g. blocking frequency and duration) is important practical limitations.

10.5.1 Validation Data

Ultimately, the reliability of any validation depends on the observational data used, either as a reference data set or to provide the forcing in a perfect boundary setting. The typical problems with reference data are inhomogeneities, outliers, and biases (e.g. Jones 1995). Inhomogeneities are systematic changes in the observational data such as slow creeping trends or jumps in the time series (its mean or other moments) due to changes in the measurement system or the surrounding environment; they might increase uncertainties and induce spurious trends (e.g. Yang et al. 2006; BACC Author Team 2008, Annex 5). Outliers are erroneously high (or low) values, such as caused by multiple-day counts of precipitation measurements; they are particularly detrimental for the estimation of extreme properties but may also affect the validation of other quantities. Biases are caused by systematic peculiarities that lead to a misrepresentation of the local climate by the measurement device, such as wind shadows due to buildings or wind-induced precipitation undercatch. Depending on the property of interest, addressing these issues might be essential for a reliable validation. Another common issue is the availability of long reference data sets, which are needed for robust estimates of the indices of interest, especially for extremes and long-term variability. In particular processes with strong small-scale variability such as precipitation, station data cannot directly be compared with regional climate data, which are considered to represent areal averages instead of point measurements (Chen and Knutson 2008). To overcome this spatial mismatch, gridded data sets have been derived by interpolation and averaging from dense station networks. Prominent examples are the UK Met Office gridded daily precipitation data set (Perry et al. 2009) and the E-OBS daily data set of temperature and precipitation (Haylock et al. 2008) derived from the European Climate Assessment & Dataset database (http://eca.knmi.nl; Klok and Klein Tank 2009) as part of the ENSEMBLES project. Crucial for the usefulness of gridded precipitation data sets is the density of the underlying rain gauge network. For instance, it has been shown that the first version of the E-OBS data set has incorporated too few rain gauges to represent extreme precipitation in some mountain regions (Maraun et al. 2011).

To validate large-scale features, reanalysis data are often taken as reference such as the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) (Kalnay et al. 1996) or the ERA-40 (Uppala et al. 2005) and ERA-Interim (Dee et al. 2011) reanalysis. These are numerical model hindcasts into which observational data have been assimilated. As the output from numerical models, these data are globally complete at the given resolution and provide a sequence of climate states (usually provided every 6 h) consistent within the numerical model. However, due to model biases, the reanalysis data can substantially deviate from reality. Also, reanalysis data do not resolve small-scale features and are therefore not suitable for validation on scales typically relevant for impact studies. Furthermore, it is necessary to be aware whether the observations representing the variable of interest have been assimilated into the model. For instance, precipitation is generally not assimilated into the reanalysis model but fully generated by the model parameterisations; such data are obviously not suitable as reference for validation. Recent projects such as the North American Regional Reanalysis (Mesinger et al. 2006) therefore assimilate further variables such as precipitation. Their completeness and consistency make reanalysis data an ideal candidate to provide boundary conditions for a perfect boundary validation.

10.5.2 Validation Indices

To validate climate simulations, several indices have been proposed, depending on the application of the downscaled product. Comprehensive lists of indices are available from the 'Expert Team on Climate Change Detection and Indices' (Peterson et al. 2001), the STARDEX project (Goodess et al. 2005), and the ENSEMBLES project (van der Linden and Mitchell 2009). Typical validated indices characterise statistics of the variable of interest such as mean, variance, or even the spatial and temporal structure.

The indices to validate the distribution of the variable of interest are statistics such as mean and variance or specific quantiles. For instance, a widely used index for strong but not yet extreme events is the 90th percentile. More generally, the indices can be the parameters of a parameterised formulation of the distribution such as the shape parameter describing the tail behaviour. To obtain results as robust as possible, the representation of extreme events should, if possible, be based on parametric distributions motivated by the extreme value theory, that is, the generalised extreme value (GEV) distribution to validate maxima of long blocks and the generalised Pareto distribution (GPD) to validate excesses of high thresholds (Coles 2001). The spatial indices are, for example, spatial correlations, cluster sizes, and indices describing spatial patterns. The temporal indices are autocorrelation functions, the annual cycle, variability on interannual to decadal timescales, and trends. Other temporal indices describe the length of events such as droughts or wet spells, and the transition probability between different states (e.g. from dry to wet). The corresponding extremal indices (which do not necessarily follow the extreme value theory) would be the maximum length of an event in a defined period, such as a season. To increase confidence in future projections, it is also important to assess the representation of relevant physical processes (e.g. Schär et al. 1999; Lenderink and van Meijgaard 2008; Kendon et al. 2010; Maraun et al. 2011). Of course for every validation procedure, particularly if hypothesis tests and statistical models are involved, the assumptions to be made should be clearly laid out.

An ongoing debate concerns whether the validation should use the data directly with grid box resolution, or whether the data should be smoothed in advance. On the one hand, it is argued that regional climate simulations are not meant to be interpreted on a grid box level and so the former choice would be too rigid. While on the other, it is a matter of fact that RCM simulations are often used on the grid box level, and a validation should not influence the corresponding performance. Furthermore, in impact studies, the simulated unsmoothed fields are often required even when they are not interpreted on a grid box level. Smoothing might then hide important spatial properties such as the spatial correlation structure.

The validation indices need to be carefully selected. In particular, they need to be independent of calibration or tuning. That is, for PP statistical downscaling and MOS, calibration and validation need to be carried out as a cross-validation on different data sets (e.g. different time periods). Even in cross-validation, the significance of apparently good performance needs to be critically assessed. If the indices are the predictands explicitly modelled in the PP approach or corrected using MOS, they will probably closely resemble the reference indices even in the validation period. Here, good agreement does not necessarily imply a high skill to represent future climate. A similar argument holds for RCMs, as these are in general tuned to properly simulate the observed climate of a specific region. This is discussed in more detail in Sect. 10.5.6.

10.5.3 Validation Measures

To quantify the discrepancy between the modelled and reference validation indices, a range of validation measures has been defined. In some validation studies, the discrepancies have not been quantified at all, but have only been visually inspected. On the other end of the range are statistical tests which explicitly address the significance of the discrepancies. In all cases, deviations should be interpreted carefully. Whereas a visual inspection might overlook important misspecifications, a significance test might as well be misleading (BACC Author Team 2008, Annex 8). Apart from false-positive results, the power of a test might simply be too low to detect model errors due to a lack of data or, in contrast, a significant deviation might simply be completely irrelevant.

The validation in a control run setting is fundamentally different from that in a perfect boundary setting. In a control run setting, the weather sequences between the model simulation and the validation data are independent. The validation can therefore only be based on long-term (climatological) statistics or, more general, distributions. In a perfect boundary setting, the modelled and observed weather sequences are more or less synchronous, given spectral nudging or a small domain and strong lateral forcing. Therefore, in addition to a distribution-wise validation, measures developed for the validation of weather forecasts can be applied for an eventwise validation.

10.5.4 Measures for Distribution-Wise Validation

Simple measures that can be applied to either spatial fields or time series are absolute and relative biases, for example, in mean and standard deviation. Spatial fields can furthermore be validated by their pattern correlation and (root)-mean-squared error relative to the reference pattern, which can be visualised in Taylor diagrams (Taylor 2001). It should be noted, however, that Taylor diagrams do not address the overall biases and provide no confidence intervals. They have been introduced in the AMIP project to synthesise the results of a large number of models in a single diagram. Further insight can be gained by calculating corresponding measures for quantiles or parameters of distributions. For the comparison of the overall distribution, the chi-square test or the Kolmogorov–Smirnov test might be applied (e.g. Semenov et al. 1998; Bachner et al. 2008). The graphical tools for the comparison of distributions are probability (PP) plots and, in particular for extremes, quantile (QQ) plots (e.g. Déqué 2007; Coles 2001). For a list of measures to validate distributions, see Ferro et al. (2005).

10.5.5 Measures for Eventwise Validation

In a perfect boundary setting, a broad range of additional validation measures can be applied. If the modelled and observed time series are synchronous and their phases are expected to match, measures can be applied that have been applied to validate weather forecasts. The same measures that are only applicable to spatial fields in a distribution-wise validation can in this context be applied to validate individual time series. These are, for example, cross-correlations and (centred) root-mean-squared errors, which then, for variables close to normally distributed, can also be visualised by Taylor diagrams. The measures to validate the occurrence of events are the hit rate and the false alarm rate, which are summarised in contingency tables (e.g. Wilks 1995). From these, it is possible to derive frequency biases and odds ratios. Also, continuous variables can be compared using these measures by defining suitable thresholds. Several downscaling approaches predict local-scale probability density distributions rather than specific values; their performance can be validated by probability scores. The classic measure to validate the occurrence of events is the Brier score (Brier 1950). Continuous events (i.e. intensities) can be validated by the continuous ranked probability score (e.g. Jolliffe and Stephenson 2003) and the quantile verification score (e.g. Friederichs and Hense 2007). Absolute score values are often difficult to interpret; therefore, they are usually compared with a reference forecast such as the climatology or the best-performing method. Such relative measures are skill scores, which can be derived from the aforementioned scores. A comprehensive list of further scores is available from, for example, Wilks (1995) and Jolliffe and Stephenson (2003). As an alternative to simple cross-correlations, one can assess the performance on different timescales using the squared coherence (Brockwell and Davis 1991); for an example, see Maraun et al. (2011). In essence, in this setting, a more rigorous validation is possible, as the capability of a model to simulate the occurrence and magnitude of individual events can be assessed. Of course, this setting does not in general allow for the assessment of GCM errors.

10.5.6 Validation in a Climate Change Context

A high skill of a downscaling method in the current climate does not necessarily imply a high skill in a future climate (e.g. Christensen and Christensen 2007). In PP statistical downscaling, the predictor–predictand relationships might be non-stationary in time, for example, because not all relevant factors controlling the local-scale variable have been included in the model. Also, it is not a priori clear whether the parameterisations of RCMs might capture the changing climate conditions. Finally, biases are not stationary under climate change (e.g. Christensen et al. 2008; Maraun 2012).

To at least partly address these shortcomings, it has been suggested to choose time periods climatically, as different as possible, to calibrate and validate statistical downscaling models (Maraun et al. 2010). This approach is of course limited by the availability of long time series of high quality. For dynamical downscaling, a similar approach is to check whether a RCM performs well in different present-day climates (Christensen et al. 2007). Consensus between different simulations is often seen as a measure of skill. Similarly, a comparison of statistical and dynamical downscaling might provide some insight into the reliability of future simulations. For instance, relationships within statistical downscaling models have been used to validate dynamical climate models (e.g. Busuioc et al. 2001; Maraun et al. 2011). Closely related is the use of RCMs as pseudo-realities to assess the stationarity of predictor–predictand relationships and model biases (e.g. Frias et al. 2006; Vrac et al. 2007; Maraun 2012). The value of model consensus and related concepts is, however, limited as deficiencies might be common to all models. Therefore, understanding the relevant underlying processes and the quality of their representation by the models used is essential to assess the reliability of future climate simulations (Maraun et al. 2010).

10.6 Skill of Downscaling Methods

This section gives a brief overview of the advantages and disadvantages of different downscaling methods. A more detailed discussion can be found in Benestad et al. (2008) and Maraun et al. (2010).

The quality of a downscaling product stands and falls with the ability of the forcing GCM to provide meaningful large-scale boundary conditions. As downscaling aims to correct local-scale misrepresentations due mainly to topographic and small-scale circulation effects, it cannot correct the misrepresentation of the large-scale atmospheric flow. For northern Europe and the Baltic Sea, the most obvious shortcoming of many GCMs is the position, strength, and variability of the main westerly flow. The circulation in many GCMs is too zonal (van Ulden et al. 2007). The large-scale circulation plays a dominant role in the European winter climate (Hurrel and van Loon 1997; Wibig 1999), but also strongly influences summer precipitation in northern Europe (Wibig 1999; Boé et al. 2009). A major shortcoming of the current generation of GCMs is the representation of blocking events (e.g. Palmer et al. 2008; Hinton et al. 2009). Consequently, a large part of the uncertainty in northern and central European temperature and precipitation projections stems from the driving GCM (Déqué et al. 2007).

The main rationale for using dynamical downscaling is that RCMs are based on physical laws. As a consequence, RCMs are in general expected to adequately describe climate change on regional scales. Although the related stationarity issues are more severe for statistical downscaling, it should be noted that parameterisations are developed and tuned for specific climates and might be at least slightly misspecified under future climate conditions. As RCMs calculate the state of the atmosphere regularly in three-dimensional space and in time, output can be generated for a large number of variables at or close to the surface as well as for levels above at temporal frequencies down to the internal computational time step of the respective RCM on a regular grid.

A practical advantage of dynamical downscaling approaches is that they are in principle applicable to any region of the world, whereas statistical downscaling approaches rely on high-quality data for the calibration. As parameterisations must be tuned for different climatic regions, however, RCM simulations for regions without proper validation data should not be taken face value. Although station coverage in the Baltic Sea region is generally very dense (e.g. van Engelen et al. 2008), this problem is not negligible here. Lind and Kjellström (2009) have shown that the observational estimates of precipitation differ to such a high degree that RCM evaluation was affected.

RCMs have been shown to adequately simulate European daily temperature and precipitation intensities, although considerable biases must be expected (e.g. Fig. 10.5; Jacob

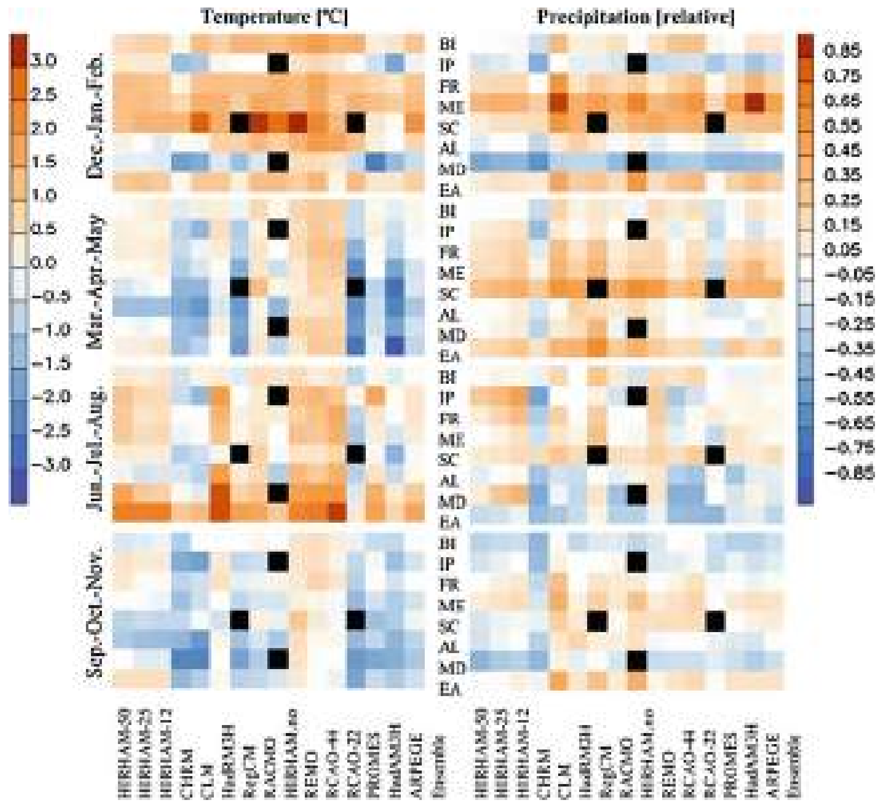


Fig. 10.5 A schematic overview of seasonal bias in the PRUDENCE regional models. In each panel, *rows* are the analysis areas, and *columns* correspond to models. *Rows* of panels signify the four seasons, the *left column* of panels is temperature biases (*left colour bar*, °C), while the *right column* of panels signifies precipitation (*right colour bar*, relative change). Areas not covered by a particular model are indicated by *black squares* (after Jacob et al. 2007)

et al. 2007). For instance, in winter, model results tend to be too wet in northern Europe, too warm in summer and winter, and too cold in spring and autumn (Jacob et al. 2007). On the one hand, RCMs generally overestimate the number of wet days; this 'drizzle effect' is partly because RCMs simulate area averages rather than point values. While on the other, RCMs underestimate heavy precipitation events (e.g. Fowler et al. 2007b). Generally, bias is different in different part of the distribution (e.g. Jeong et al. 2011; Fig. 10.6).

A major advantage of RCMs is the simulation of spatially coherent fields. In general, RCMs with a typical resolution of 25 km overestimate the spatial coherence of precipitation events, in particular for convective precipitation. It should be noted that RCMs provide meaningful information only on the scale of a few grid cells (e.g. Fowler and Ekström 2009). In particular, local precipitation is dominated by internal climate variability (Maraun 2012).

As RCMs integrate the equations governing the atmospheric circulation, they in principle provide a coherent picture. However, biases in one variable may propagate into strong biases in dependant variables (e.g. Fig. 10.7); for example, Yang et al. (2010) have shown for Sweden that small temperature biases may, via the nonlinear interaction with precipitation around the melting point, lead to large biases in spring river run-off. Inconsistencies arise in particular where parameterisations come into play. For example, Graham et al. (2007b) have shown for the drainage areas to the total Baltic Sea basin and Bothnian Sea basin that the partition of precipitation into run-off and evapotranspiration is in general biased towards the latter.

Only a few RCM validation studies consider sub-daily scales, which are particularly relevant for heavy precipitation events. Jeong et al. (2011) have shown that the spatial pattern of the diurnal precipitation cycle in Sweden is reasonably captured by the RCA3 model (Uppala et al. 2005), but the afternoon peak occurs too early and is spatially too uniform. The RACMO2 model accurately simulates the intensity scaling of heavy hourly precipitation with temperature for very intense precipitation, but fails to represent the temperature influence on moderate precipitation intensities beyond 20 °C (Lenderink and van Meijgaard 2008).

In general, increasing model resolution improves model simulations, in particular for precipitation in complex terrain (e.g. Salathé 2003; Hohenegger et al. 2009).

MOS (see Sect. 10.3.1) aims to improve misspecification of dynamical downscaling, although recent work has demonstrated the ability to directly apply MOS to GCMs (Eden et al. 2012). An underlying assumption of MOS is stationarity of the bias. Yet Christensen et al. (2008) inferred a dependence of biases on temperature, indicating potential non-stationarities. In a pseudo-reality, Maraun (2012) found biases in seasonal temperature and precipitation to be relatively stable across Europe, but identified non-stationarities for some regions and seasons; for the Baltic Sea region,

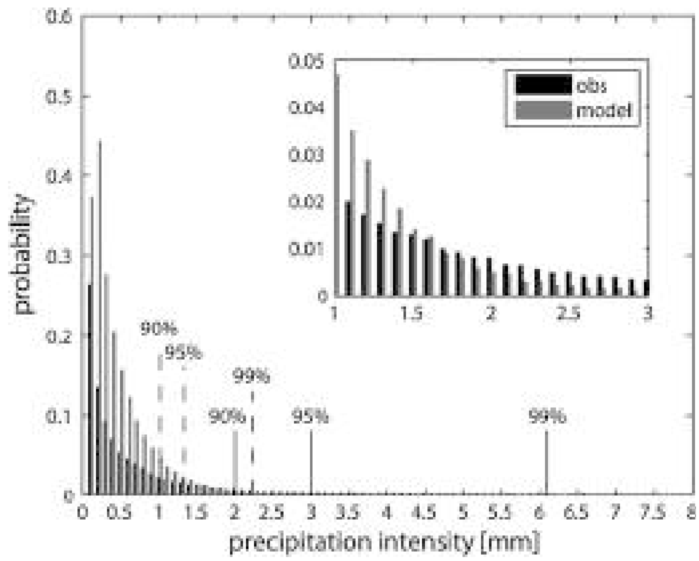


Fig. 10.6 Probability density of precipitation intensity on the eastern coast of Sweden in the warmest months of the year (April–September). *Inner box* highlights the probability of precipitation intensity from 1–3 mm h^{-1}. The 90th, 95th, and 99th percentiles are marked for the observations (*solid vertical lines*) and the model simulation (*dashed vertical lines*) (Jeong et al. 2011)

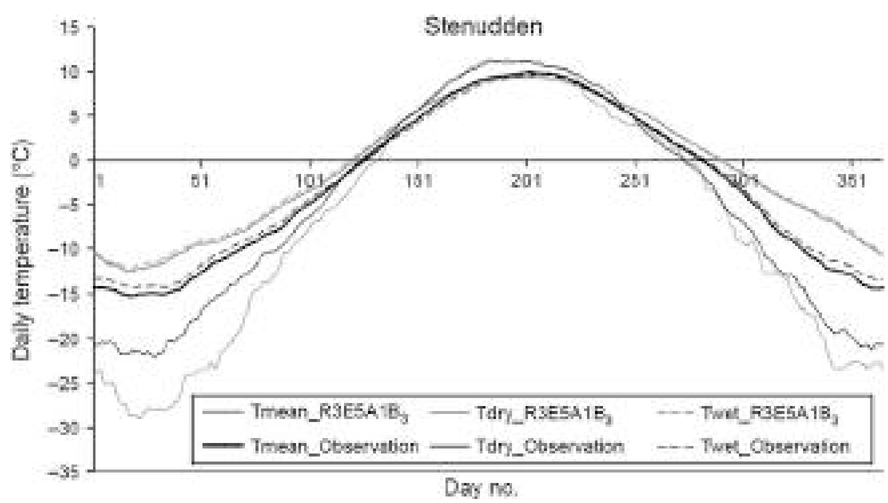


Fig. 10.7 Impact of precipitation on temperature bias for Stenudden, northern Sweden. Annual cycle of daily temperature from observations (1961–1990) and a model (R3E5A1B3) simulation for the same period. *Tmean* mean daily temperature; *Twet* mean daily temperature for days with precipitation; *Tdry* mean daily temperature for dry days (Yang et al. 2010)

temperature biases appear to be non-stationary because of uncertainties in sea ice parameterisations.

MOS has been shown to successfully correct temperature biases as well as biases in precipitation intensities and the number of wet days (e.g. Hay and Clark 2003; Lenderink et al. 2007b; Piani et al. 2010). MOS is particularly suitable to correct orographic effects on precipitation intensity in regions where the topography is misrepresented by the coarse model grid. Furthermore, Widmann et al. (2003) developed a non-local MOS that corrects systematic spatial displacements of precipitation. Yang et al. (2010) applied MOS to improve the correlation between simulated temperature and precipitation.

The good example of the skill of the DBS methodology was presented by Piani et al. (2010). Bias corrections were assessed for the 10-year period 1961–1970 and then applied to the simulated data for the period 1991–2000 and compared with observations from this period. The periods were chosen to maximise the time lag between them and test whether the bias correction estimated in one period can be applied in the other period with different climatic conditions. The results were surprisingly good (Fig. 10.8). Not only did the mean and higher moments of the scenario data fit well with the observed data, but also indices depending on autocorrelation spectra, such as drought and heavy precipitation, were well projected.

MOS is not capable of correcting the misrepresentation of the temporal structure of a simulated variable; for example, MOS cannot correct errors in the length of dry, hot, or cold spells inherited from GCMs. In particular, the POD method should be considered carefully. Because it only scales observed time series, owing to its construction, it ignores any changes in the atmospheric dynamics which might change the temporal structure of future weather. Nevertheless, applying MOS separately to seasons, individual months or even shorter parts of the year might improve the

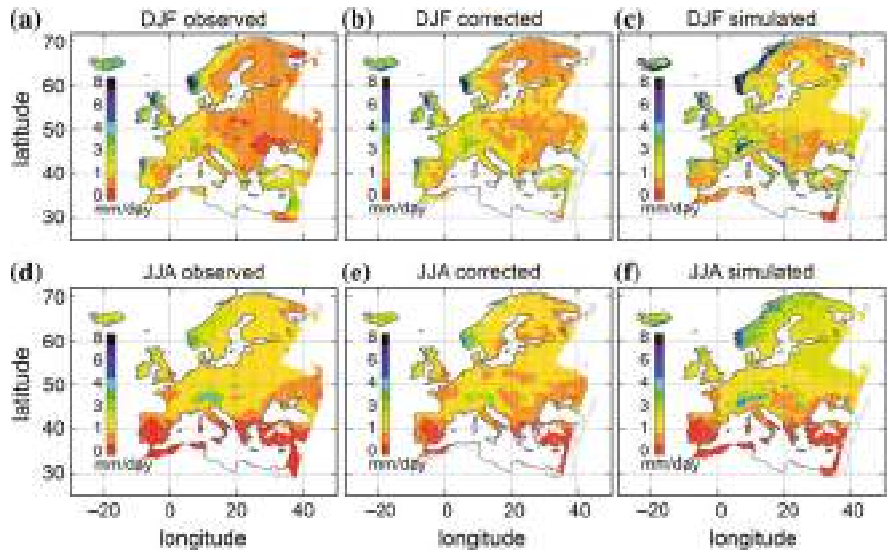


Fig. 10.8 Validation of methodology: seasonal mean daily precipitation. Application of bias correction, derived from simulated and observed data for 1961–1970, to model data for 1991–2000. **a** Mean observed daily precipitation for winter (DJF) 1991–2000, **b** as '**a**' but for corrected simulated data, **c** as '**a**' but for uncorrected simulated data, and **d**–**f** as '**a**–**c**' but for summer (JJA) (Piani et al. 2010)

representation of the annual cycle (e.g. Boé et al. 2007; Leander and Buishand 2007).

PP statistical downscaling (see Sect. 10.3.2) is often a computationally cheap alternative to dynamical downscaling. Its main rationale is the explicit use of empirical knowledge by including observational data in statistical models. By construction, the properties which are directly modelled as predictands should be simulated bias free over the calibration period, a characteristic often required by impact modellers. In particular, over complex terrain, PP intrinsically accounts for local effects which might not be captured by the coarse topography and imperfect parameterisations of RCMs. The predictor selection is crucial for the performance of PP approaches; non-stationarities may arise if the predictors do not capture the climate change signal. Predictors should therefore be physically motivated and as close to the underlying processes as possible, see Benestad et al. (2008) for details.

In general, PP methods perform better during winter than summer; Wetterhall et al. (2007) demonstrated this tendency for Sweden.

A shortcoming of many traditional PP methods is the underrepresentation of temporal variability. Most PP methods can be interpreted as some kind of linear or nonlinear, continuous, or categorical regression models. Such models are in general intended to predict the mean of a distribution, disregarding the variability around the mean. As previously discussed, inflation (Karl et al. 1990) or similar approaches do not resolve this problem, but rather create time series with an incorrect temporal structure (von Storch et al. 2000). Instead, many randomisation procedures have been suggested, ranging from generalised linear models (e.g. Chandler 2005) via mixture models (e.g. Vrac and Naveau 2007) to full weather generators. These models provide a realistic temporal structure, which might be explicitly modelled by Markov processes on short temporal scales, and imposed on longer timescales by large-scale predictors (Wilks and Wilby 1999). Weather generators have also been constructed to simulate sub-daily precipitation (e.g. Cowpertwait et al. 1996; Jones et al. 2009).

A main disadvantage of PP approaches is the handling of spatial coherence. Many PP approaches are used for single locations. Here, the downscaling to local scales is a major advantage over RCMs, which operate on scales of tens of kilometres. The large-scale predictors might impose a coherent spatial structure, which however is often too smooth and can be improved by the addition of a stochastic factor—a weather generator. Randomisation leads to improved local temporal variability, but at the same time might destroy spatial coherence. Therefore, spatial dependence needs to be modelled explicitly by complex multi-site models (e.g. Yang et al. 2005), which provide output at discrete points in space. The development of downscaling methods to full spatial fields for climate change studies is still in its early stages; an example is the Gaussian process-based disaggregation of areal rainfall by Onibon et al. (2004).

A practical advantage of the PP approach is its computational cost; for single sites, it can easily be applied to large ensembles of GCMs, and conditional on one GCM, numerous realisations can be carried out by randomisation.

In general, whether dynamical downscaling or PP is preferable depends on the problem addressed. In many situations, both methods are complementary and should be used in combination. With the availability of large RCM ensembles from the ENSEMBLES project (van der Linden and Mitchell 2009), MOS corrections have become increasingly attractive, attempting to combine the best of both worlds.

10.7 Added Value of Dynamical Downscaling

It is assumed that GCMs are able to provide a reliable description of large-scale weather phenomena and their dynamics. RCMs can resolve mesoscale atmospheric features explicitly and they add small-scale structures to the large-scale circulation provided by the driving model (Feser 2006). Local climate is influenced by large-scale dynamics, regional physiographic features such as local orography, land-sea contrasts land use, and soil type, as well as by small-scale atmospheric features such as frontal systems or convective cells (Lenderink et al. 2007b; Feser et al. 2011). This is particularly the case for the simulation of precipitation. Consequently, the simulated mean precipitation patterns as well as the extreme values are enhanced, especially for complex terrain (e.g. Christensen and Christensen 2001; Feldmann et al. 2008; Suklitsch et al. 2008). For many variables, the explicit treatment of small-scale atmospheric features leads to added value (AV) with respect to the driving model. Assessment of the AV of large-scale constrained versus unconstrained simulations was discussed by Castro et al. (2005) and Rockel et al. (2008). For different realisations of an RCM simulation, generated by small changes in the set-up of the RCM (e.g. domain size/location, initialisation time), substantial variability between the individual realisations is well known (e.g. Ji and Vernekar 1997; Rinke and Dethloff 2000; Weisse et al. 2000), demonstrating the need for ensemble RCM simulations with a large number of realisations. Many publications demonstrate that RCMs are able to realistically simulate climate in comparison with raw or gridded observations or reanalysis. Most of these state the superiority of RCM simulations compared to those from GCMs, but without giving proof.

One of the most important purposes of regional climate modelling is increasing knowledge of the real world (Laprise 2005). This additional knowledge is commonly termed 'added value' (Feser 2006). Identification of AV is not an easy task. Small-scale atmospheric fields are usually less energetic than large-scale fields (Laprise 2005), so scale decomposition is sometimes necessary to separate the finer scales.

Di Luca et al. (2012) used a diagram, adapted from Orlanski (1975) and von Storch (2005), to illustrate the concept of AV for the range of scales represented by global and regional models, relative to the characteristic temporal and spatial scales of atmospheric processes (Fig. 10.9). Regional climate modelling is mainly expected to add value at regional dimensions below 300 km and temporal scales less than 30 min, which are absent in GCMs.

Evaluation of a hypothesis of AV implies a comparison of the performance of the RCM with that of the driving GCM (Feser et al. 2011). To date, the number of studies in which the AV of RCMs is directly analysed is limited and only a few concerns the Baltic Sea region.

RCMs could provide AV by adding variability at scales not well resolved by GCMs (at Fig. 10.9 referred to as AV1). However, RCMs can also improve climate simulation at scales resolved by both RCMs and GCMs. This component of AV is referred as AV2 in Fig. 10.9. Because separation of scales is usually made while assessing AV, it is convenient to analyse both types of AV separately (Di Luca et al. 2012). RCMs operate in a limited domain, and two-way interactions between the regional domain and the rest of the globe do not usually occur. In many simulations, the spectral nudging technique ensures that the large scales are not altered too much by the regional model. For all these reasons, AV2 has not been clearly identified and existing analyses are usually limited to AV1.

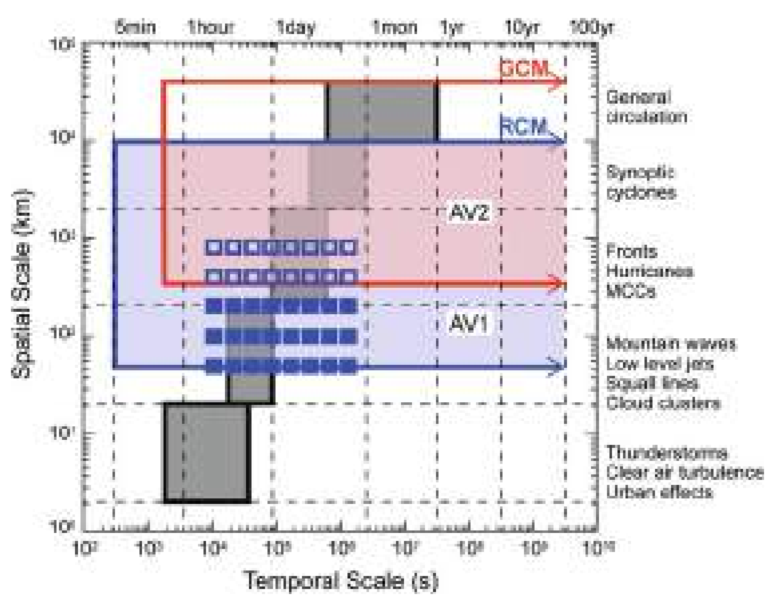


Fig. 10.9 Characteristic temporal and horizontal spatial scales of atmospheric processes (in *black*) and the range of scales represented in RCMs (*blue line*) and GCMs (*red line*). *Red* and *blue* shaded regions represent the added value of type 1 (AV1) and 2 (AV2), respectively (redrawn from Di Luca et al. 2012)

As RCMs can resolve mesoscale atmospheric features explicitly, they do add small-scale structures to the large-scale circulation provided by the driving model (Feser 2006). This explicit treatment of small-scale atmospheric features leads, for many variables, to an AV with respect to the driving model. This is particularly the case for the simulation of precipitation, which also depends strongly on topography and land–sea contrast, which are better represented at the increased resolution of the RCM. Consequently, the simulated mean precipitation patterns as well as the extreme values are enhanced, especially for complex terrain (e.g. Christensen and Christensen 2001; Feldmann et al. 2008; Suklitsch et al. 2008). For the Baltic Sea region, Walther et al. (2013) demonstrated the improved simulation of the daily precipitation cycle for spring and summer with increasing RCM resolution; an example for a station in central southern Sweden is displayed in Fig. 10.10.

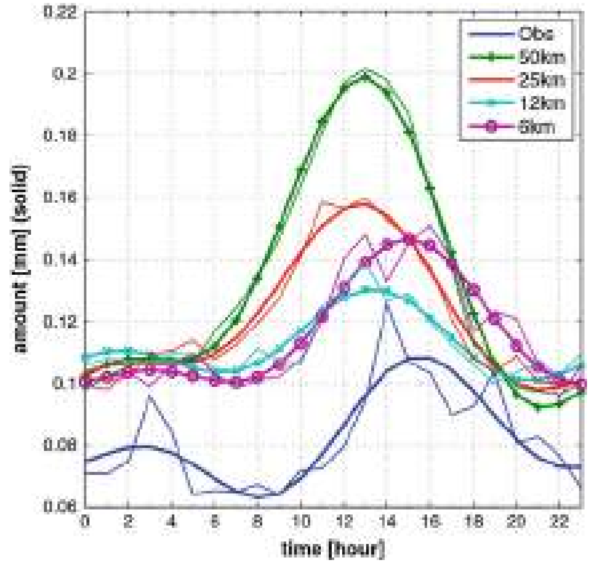


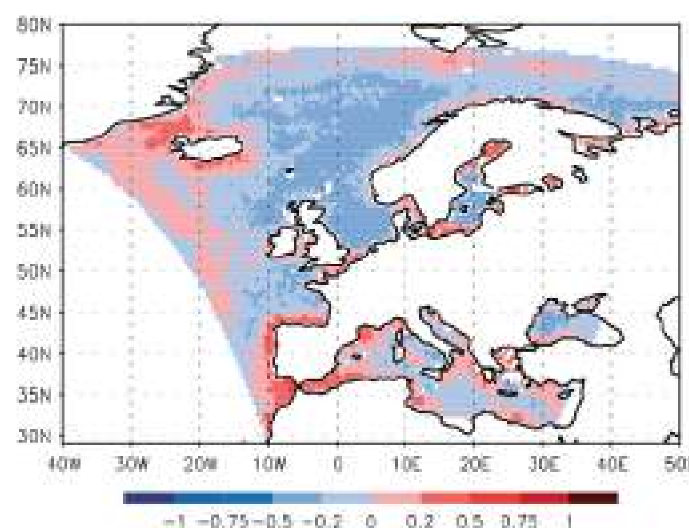
Fig. 10.10 Estimated diurnal cycle of precipitation amount from observation and the RCA3 regional climate model (RCM developed by the Rossby Centre of SMHI) simulations for four different resolutions for the station 'Malexander' in central southern Sweden (Walther et al. 2013)

Winterfeldt et al. (2011) analysed AV in dynamically downscaled wind speed fields. They used the Brier skill score (BSS) to detect the AV of the regionally modelled (with spectrally nudged-REMO) wind in comparison with the global reanalysis (NCEP). As seen in Fig. 10.11, the RCM provides AV along the coasts and in narrow bays and straits, in places with complex coastlines or topography. Over open seas and oceans, as well as the interior of Baltic Sea, the BSS is negative, indicating that in these regions, dynamical downscaling does not add value.

Feser (2006) analysed AV in the case of SLP and 2 m air temperature provided by the REMO RCM in comparison with NCEP reanalysis data. Spatial filters were used to separate the data into two domains: that represented best at the large-scale and that represented well by the REMO model (Fig. 10.12). The effect of spectral nudging was also analysed. For SLP, no AV is provided by RCM simulation without nudging. The small improvement was obtained when spectral nudging was applied. For 2 m air temperature, significant AV was obtained at both scales when spectral nudging was applied. Without nudging, only the improvement in the scale represented well by RCM was provided (Feser 2006). AV is small in the case of SLP and only for the scale well resolved by RCM, because it is the driving fields (from the GCM) that are the most relevant for SLP. For 2 m air temperature, regional and local factors exert a strong impact on its spatial distribution and the AV of RCMs can be significant (Feser et al. 2011).

Zahn et al. (2008) have shown that RCMs can provide AV in describing mesoscale phenomena such as polar lows. Figure 10.13 presents the SLP and 10 m wind speed fields filtered with a digital bandpass filter that allows better

Fig. 10.11 Brier skill score using QuikSCAT level 2B12 as the source of ground-truth data, global reanalysis (NCEP reanalysis) as the reference forecast, and a regional model (spectrally nudged-REMO) as the forecast, after Winterfeldt et al. (2011)



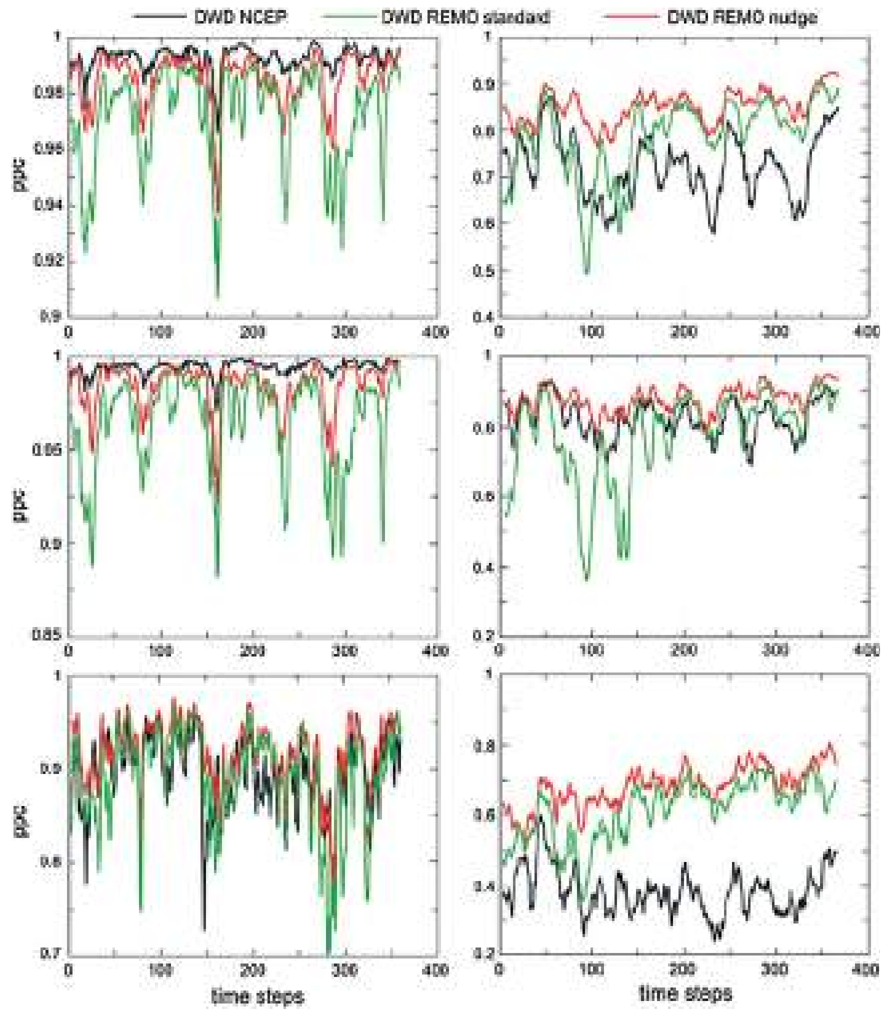


Fig. 10.12 *Left* Six-hourly time series of sea-level pressure pattern correlation coefficients (pcc) between DWD (Deutscher Wetterdienst—German weather service) analyses and reanalyses or RCM data after Feser (2006) for winter 1998/99. *Right* Time series of 2 m temperature anomaly (pcc) for summer 1998 for (*top*) full fields, (*middle*) low-pass-filtered, and (*bottom*) medium-pass filtered fields

presentation of phenomena at the 200–600 km spatial scale. The community land model (CLM) was able to identify the polar low along the Norwegian coast, although it was still too coarse to describe fine detail.

Di Luca et al. (2012) also introduced the concept of potential AV, as a type of necessary condition for AV. It is a fine spatial scale variability that would be present in regional climate statistics but absent on a coarser grid. The presence of *potential* AV in RCM simulations indicates the possible existence of AV but does not prove it. Di Luca et al. (2012) investigated the existence of potential AV at the temporal scale for different regions and seasons. They showed that for precipitation, potential AV increases for short temporal scales, the summer season and in regions with complex

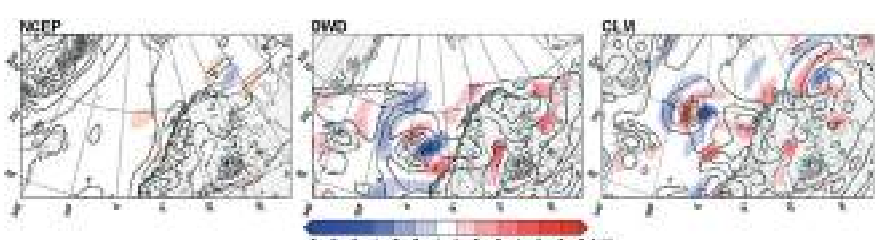


Fig. 10.13 Bandpass-filtered mean sea-level pressure (*isolines* hPa) and 10 m wind speed anomalies (*shaded*) at 0600 UTC 15 October 1993: NCEP analysis, DWD (Deutscher Wetterdienst—German weather service) analysis data, and a simulation by the regional climate model CLM (after Zahn et al. 2008). The position of the pressure minimum of the polar low in the CLM simulation is indicated (*yellow dot*), after Feser et al. (2011)

orography, and decreases when statistics are averaged over long time periods or over a wide spatial domain.

10.8 Downscaling in the Context of Climate Change Impact Studies

The GCMs were not designed for direct application in impact models. Prudhomme et al. (2002) stated that the quality of their output did not allow for direct use in hydrological impact studies, because the spatial and temporal scales were too coarse. Wilby et al. (1999) recommended the use of downscaling techniques before the GCM output data could be used in impact studies. There are many possibilities for downscaling GCM output: direct use of RCM output (Wood et al. 2004), use of bias-corrected RCM output (Wood et al. 2004; Fowler et al. 2007a), statistical downscaling (Wilby et al. 2000; Müller-Wohlfeil et al. 2000), stochastic weather generators (Evans and Schreider 2002) or weather typologies, and/or indices (Pilling and Jones 2002). The skills of different downscaling methods differ considerably between variables and regions.

Hydrologic simulation was found to be sensitive to biases in the spatial distribution of temperature and precipitation at the monthly level, especially where the seasonal snow pack transfers run-off from one season to the next (Fowler et al. 2007c).

Using the example of the Lule River in northern Sweden and two GCMs used to force the same RCM, Graham et al. (2007a) have shown that the choice of driving GCM has a greater impact on results than the choice of GHG emission scenario. The strong impact of the choice of GCM was also emphasised by Widmann et al. (2003), Jasper et al. (2004), Salathé (2005), and Wilby et al. (2006).

Fowler et al. (2007a) stated that at least two variables—temperature and precipitation—had to be downscaled for impact studies in hydrology. In impact models, the physical consistency between variables is very important. To obey this requirement for physical consistency, multi-variate methods should be applied which yield simultaneous correction of relevant variables. This is possible when RCMs are used (Fowler and Kilsby 2007; Fowler et al. 2007a; Graham et al. 2007a, b), but is generally not in statistical downscaling. A multi-site approach should be used when spatial consistency is needed.

10.9 Conclusion

GCMs are a useful tool for studying how climate may change in the future. Such models describe the climate on a set of grid points, regularly distributed in space and time using the same density over land and ocean. Their temporal resolution is relatively high; however, their spatial resolution is low. To simulate regional climate, that is, at a scale smaller than the skilful scale, it is necessary to downscale the GCM results. Downscaling is understood as a process that links large-scale variables with small-scale variables. There are two conceptually different ways of downscaling. One uses RCMs nested in GCMs; RCMs have much higher resolution and can describe local features better and are still able to simulate the atmospheric state in a realistic manner in their skilful scales. The other group of downscaling methods uses empirical and/or statistical relations between the large-scale variables simulated by GCMs and small-scale variables describing regional and/or local climate conditions.

There are many sources of uncertainty in climate model results. These include uncertainty related to limited information on future land use and atmospheric GHG concentrations, limits on the amount of input data and their accuracy, and the chaotic nature of weather. Many sub-grid processes must be represented in models in a simplified form and are not well described by the models. For example, the modelling of cloud formation, the optical and radiative features of clouds, and the creation of atmospheric precipitation still carry considerable model error. The skill of methods for describing regional climate futures is also limited by natural climate variability.

The quality of a downscaling product rests with the ability of the forcing GCM to provide meaningful large-scale boundary conditions, because a large part of the uncertainty in northern and central European temperature and precipitation stems from the driving GCM (Déqué et al. 2007). The main shortcoming of GCMs in Europe is that in many, circulation is too zonal in winter (van Ulden et al. 2007).

RCMs are able to simulate spatially coherent fields, but the parameterisations are developed and tuned for specific climates and might be slightly misspecified under future climate conditions. RCMs have been shown to adequately simulate European daily temperature and precipitation intensities, although considerable biases must be expected (e.g. Jacob et al. 2007). The biases in one variable may propagate into strong biases in dependant variables (e.g. Yang et al. 2010).

Using an ensemble of RCMs is one way of filtering a random error and assessing uncertainty. However, the models within an ensemble are not fully independent because of using shared codes. There is still debate about ensemble design. There have been some recent attempts to design GCM-RCM ensembles in order to sample various kinds of uncertainty in a more systematic way. The uncertainty in future European climate change is generally more associated with the choice of GCM than RCM (Déqué et al. 2007), although for summer precipitation, the RCM formulation may be the dominant source of uncertainty.

Only a few RCM validation studies consider sub-daily scales. Jeong et al. (2011) showed that the diurnal precipitation cycle in Sweden is reasonably well captured by the RCM at SMHI, but that the afternoon peak in precipitation occurs too early and is spatially too uniform. Increasing model resolution will in general improve model simulations, particularly for precipitation in complex terrain (Salathé 2003).

Using coupled AOGCMs is state of the art for global climate projections (Meehl et al. 2007). RCM climate change projections are in general still carried out for the atmosphere only, prescribing SST data from the driving GCM (Christensen et al. 2007). The quality of the prescribed SST/sea ice data depends on the quality of the global modelling system. For a relatively small and semi-enclosed water body like the Baltic Sea, data quality might be limited by the coarse resolution of the global ocean component.

Non-GHG forcings, such as aerosols and land-use change, are not fully represented in RCMs. This can be a source of major uncertainty in projections of future climate as a large part of the simulated multi-decadal variance in North Atlantic SSTs depends on the atmospheric levels of aerosols.

Natural climate variability limits the skill of future climate predictability in many regions (Deser et al. 2012a). In locations where the amplitude of natural variability is high, predictability is low. Conversely, in locations with low natural variability, predictability is higher. The uncertainty of future climate projections is largely a consequence of the chaotic nature of large-scale atmospheric circulation patterns, and improving models or GHG scenarios cannot eliminate this uncertainty (Deser et al. 2012b).

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Projected Change—Atmosphere

11

Ole Bøssing Christensen, Erik Kjellström, and Eduardo Zorita

Abstract

This chapter builds on the comprehensive summary of climate change scenarios in the first BACC assessment published in 2008. This chapter first addresses the dynamical downscaling of general circulation model (GCM) results to the regional scale, focussing on results from 13 regional climate model (RCM) simulations in the ENSEMBLES project as this European-scale ensemble simulation is also relevant for the Baltic Sea region and many studies on temperature, precipitation, wind speed and snow amounts have been performed. This chapter then reviews statistical downscaling studies that use large-scale atmospheric variables (predictors) to estimate possible future change in several smaller scale fields (predictands), with the greatest emphasis given to hydrological variables (such as precipitation and run-off). For the Baltic Sea basin, the findings of the statistical downscaling studies are generally in line with studies employing dynamical downscaling.

11.1 Introduction

A comprehensive summary of existing scenarios for the Baltic Sea region up to 2006 was published in the first assessment of climate change for the Baltic Sea basin (BACC Author Team 2008). Since then several large systematic efforts have been made to perform numerical model simulations and extract knowledge about anthropogenic climate change in this region. At the international level, a large number of global climate change scenarios have been produced over recent decades in climate model intercomparison projects (CMIP), often in connection with work on the latest Intergovernmental Panel on Climate Change (IPCC) assessment reports (IPCC 2001, 2007). The fourth IPCC assessment report (IPCC 2007) built on the World Climate Research Programme's (WCRP) Coupled Model Intercomparison Project phase 3 (CMIP3) multi-model data set (Meehl et al. 2007) with the participation of many general circulation models (GCMs) and use of several IPCC SRES scenarios (Nakićenović et al. 2000).

At the European level, some of these scenarios have been dynamically or statistically downscaled to a higher horizontal resolution allowing for detailed analysis of climate change on a local to regional scale. Regional climate model (RCM) simulations in the PRUDENCE project (Christensen and Christensen 2007) have been analysed, and simulations from the ENSEMBLES project (van der Linden and Mitchell 2009) have been made publicly available and analysed (e.g. Hanel and Buishand 2011; Kyselý et al. 2011; Räisänen and Eklund 2011; Déqué et al. 2012; Kjellström et al. 2013) and used for impact studies (e.g. Wetterhall et al. 2011). Finally, at the Baltic Sea level, several national initiatives have

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_11

resulted in extensive analyses of possible climatic futures for areas including the Baltic Sea basin (e.g. Lind and Kjellström 2008; Kjellström and Lind 2009; Benestad 2011; Kjellström et al. 2011a; Nikulin et al. 2011).

Probabilistic climate change information has been derived from the GCM scenarios (Räisänen 2010) and RCM scenarios (Buser et al. 2010; Donat et al. 2011). In addition, the wider range of GCM scenarios has been used to set regional scenarios in a broader context (Lind and Kjellström 2008).

This chapter relies on existing literature on climate change scenarios with a focus on northern Europe and in particular the Baltic Sea region. Some original summarising plots of public data from the ENSEMBLES project have been added because not all of the expected analyses of the ENSEMBLES archive have yet been published; at the same time, this collection of RCM data is an essential improvement on the state of the science since the first BACC assessment. For a more detailed description of GCMs and downscaling techniques, see Chap. 10 as well as IPCC (2007) and BACC Author Team (2008) and references therein. Much literature exists related to past and present climate change in the Baltic Sea, and conversely about future climate at the European scale, but there is little targeted research on future climate change in the Baltic Sea area.

Most RCMs do not include a dynamic ocean model, implying that the surface properties for the Baltic Sea (sea surface temperatures—SSTs—and sea ice) are taken directly from the driving GCM. As the GCMs have only a very crude representation of the Baltic Sea, this constitutes an additional source of uncertainty for the regional scenarios. Based on experiments with the Rossby Centre RCM RCAO in both coupled and uncoupled mode, Meier et al. (2011) concluded that the coupled model version has the potential to improve the results of downscaling considerably, as SSTs and sea-ice conditions are more realistic than in the corresponding GCMs. However, their results show that differences between the two model versions are different between different seasons with a greater improvement in summer compared to winter, since winter climate in the Baltic Sea region is more strongly governed by conditions over the North Atlantic. They also concluded that the parameterisation of air-sea fluxes needs improving in RCAO. The rest of this chapter focuses on results from atmosphere-only RCMs available at the time of writing (November 2012).

11.2 Emission Scenarios

The SRES emission scenarios—based on different storylines for the future development of world population and economy (Nakićenović et al. 2000)—were used in simulations for the IPCC's Fourth Assessment (IPCC 2007). Hence, most existing climate change scenarios build on these emission scenarios. All SRES scenarios show rising amounts of greenhouse gases (GHGs) in the atmosphere leading to rising global temperature. Depending on the scenario selected, the amplitude of the projected climate change will differ. Even though mitigation effects are not explicitly incorporated in the SRES scenarios, the range in potential futures depends on GHG emission amounts which do reflect the expected effects of mitigation action. The difference in emissions between high and low emission scenarios lead to different climate change trajectories, most notably from the mid-twenty-first century onwards. For the next few decades, much of the expected warming will be accounted for by GHG concentration increases caused by historical and current emissions.

Most downscaling experiments build on the SRES scenarios A2, A1B and B2, implying that the more extreme scenarios (A1FI on the high side and B1 on the low side) have not been studied as extensively. Pattern-scaling techniques have been used in order to 'translate' information between scenarios including the more extreme scenarios (e.g. Ruosteenoja et al. 2007; Kendon et al. 2010). Regional simulations for Europe with low-emission stabilisation scenarios exist with the ENSEMBLES E1 stabilisation scenario (see van Vuuren et al. (2007)). A very high emission scenario downscaling has also been performed (RCP 8.5, see Christensen 2011). However, none of these simulations have yet been analysed in the literature and no studies with specific focus on the Baltic Sea basin have been performed.

The IPCC Fifth Assessment has used data from the CMIP5 project, which contains output from many more models, together with the IPCC's new family of representative concentration pathway (RCP) scenarios (van Vuuren et al. 2011). Simulations for this archive had only just started to appear at the time of writing (end of 2012), and nothing had been published.

11.3 Global Climate Models

Most regional climate change information from global models in the last few years originates from the CMIP3 project underlying the IPCC fourth assessment (IPCC 2007). In that project, about 20 different coupled atmosphere–ocean general circulation models (AOGCM) were used in a number of different experiments including simulations of the twentieth century with observed forcing and a number of SRES scenarios for the twenty-first century. In addition to GCM and scenario uncertainty, uncertainty due to natural variability was also considered in CMIP3. This was achieved by multiple simulations with individual GCMs that differed only in terms of their initial conditions. Extensive documentation of model performance and climate change projections are available (IPCC 2007).

11.4 Regional Climate Models

Downscaling of GCM results to the regional scale has been pursued with a number of RCMs in the context of EU-projects (e.g. PRUDENCE and ENSEMBLES), other international projects (Climate and Energy Systems in the Nordic region; Kjellström et al. 2011b) and national efforts (e.g. Iversen 2008; Kjellström et al. 2011a). Most of the existing scenarios are at a horizontal resolution of 50 or 25 km, but attempts have been made at even higher resolution, around 12 km by Larsen et al. (2009) and 10 km by Jacob et al. (2008). In addition to downscaling of climate change scenarios, observation-based reanalysis data sets have also been extensively downscaled, particularly in recent years (e.g. Feser et al. 2001; Hagemann et al. 2004; Christensen et al. 2010; Samuelsson et al. 2011). These experiments allow a comparison of RCM results and observational data for the most recent decades and thereby an evaluation of the RCMs (see also Chap. 10, Sect. 10.2.2).

The international WCRP-sponsored CORDEX project (Giorgi et al. 2006, http://cordex.dmi.dk/) coordinates regional simulations for areas covering the whole Earth. European simulations at 12- and 50-km resolution have been performed by several institutions in Europe and publications based on these data are starting to appear (Vautard et al. 2013).

11.5 Temperature

Air temperatures in the Baltic Sea area are projected to increase with time, with the increase generally greater than the corresponding increase in global mean temperature. This is usually the case for land areas, which warm more quickly than sea areas but is also the case for the Baltic Sea region, largely due to the strong winter increase (Figs. 11.1, 11.2 and 11.3). This winter increase is the result of a positive feedback mechanism involving declining snow and sea-ice cover, leading to even higher temperatures—reduced snow and ice cover will enhance the absorption of sunlight, and so enable greater amounts of heat to be stored in the soil and water.

The strong increase in winter daily mean temperature is most pronounced for the coldest episodes (Kjellström 2004). This is also the case for the most extreme daily maximum and minimum temperatures (Kjellström et al. 2007; Nikulin et al. 2011) with a significant decrease in probabilities of cold temperatures (Benestad 2011). In summer, warm extremes are projected to become more pronounced. For example, Nikulin et al. (2011) showed that warm extremes in today's climate (1961–1990) with a 20-year return value (defined as the temperature that will be exceeded once every 20 years as a statistical average) will occur around once every 5 years in Scandinavia by 2071–2100 according to an ensemble of six RCM simulations, all downscaling GCMs under the SRES A1B scenario.

Figure 11.1 shows the annual cycle of temperature change for northern Sweden according to 23 different CMIP3 GCM simulations as well as for the ENSEMBLES RCM simulations (see Table 11.1). An increase in temperature is evident in all seasons, despite a large spread between different GCMs in their response to the change in forcing. This large spread is directly reflected in downscaling studies. Figure 11.2 shows example results from 16 regional climate change simulations with the Rossby Centre RCM (Kjellström et al. 2011a). These simulations include different emission scenarios, different forcing GCMs and different ensemble members allowing an illustration of the uncertainties related to climate change, discussed in Chap. 10, Sect. 10.4.4. It is clear that the spread due to different GCMs contributes strongly to the overall uncertainty in northern Sweden. It is also apparent that overall uncertainty increases with time, as the spread between the 16 scenarios is greater for 2071–2100 (data points with '3' in Fig. 11.2) than 2011–2040 (data points with '1' in Fig. 11.2). Furthermore, the impact of different emission scenarios increases over time as can be seen by comparing the SRES scenarios A2, A1B and B1 simulated by the same ensemble member of ECHAM5 (denoted M, A and P in Fig. 11.2). Finally, the three ECHAM5 A1B simulations with ensemble members differing only in initial conditions (A, B and C in Fig. 11.2) show a large spread in temperature and precipitation change, illustrating that natural variability also contributes to uncertainty on longer time scales, such as the 30-year period used here. The relation between change in temperature and precipitation seems robust, since the results are near the straight line; however, amplitude as a function of emission scenario shows considerable uncertainty.

An ensemble of 13 RCM simulations from the ENSEMBLES project was analysed in this chapter. The RCMs are listed in Table 11.1. This represents the complete set of ENSEMBLES regional simulations that extend to the end of the twenty-first century, except for the ICTP, UCLM and CNRM models which have been excluded from the extremes analysis as they use a different grid projection to the others and so cannot be directly compared. To ensure that the ensemble maps are directly comparable, they have also been excluded from the analysis of mean fields. For each grid point and each of the 13 RCM simulations, there is a value of the quantity in question, such as the projected change in summer temperature between 1961–1990 and 2070–2099. As an estimate of the spread, the 13 results were

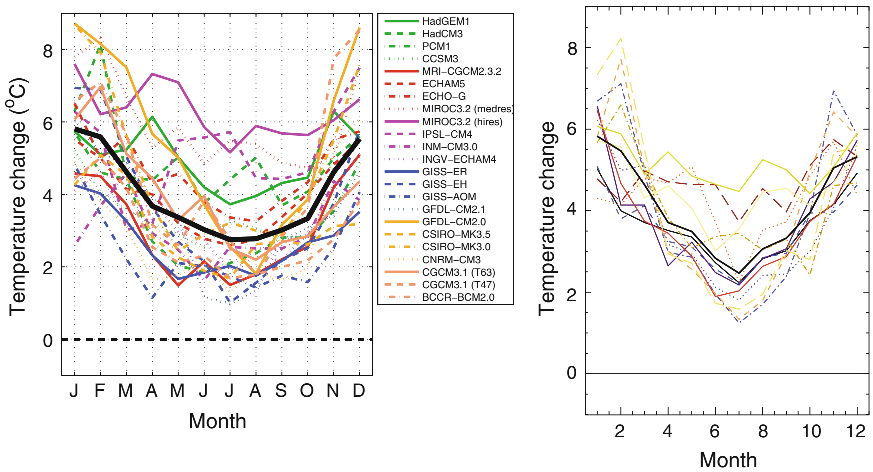

Fig. 11.1 Projected change in average monthly temperature in northern Sweden for 2071–2100 relative to 1961–1990. *First panel* results for 23 CMIP3 AOGCM-simulations under the SRES A1B scenario (Lind and Kjellström 2008). *Second panel* results for the 13 RCMs listed in Table 11.1. The *thick black lines* show the average of the individual model results, and the *dashed line* indicates no change

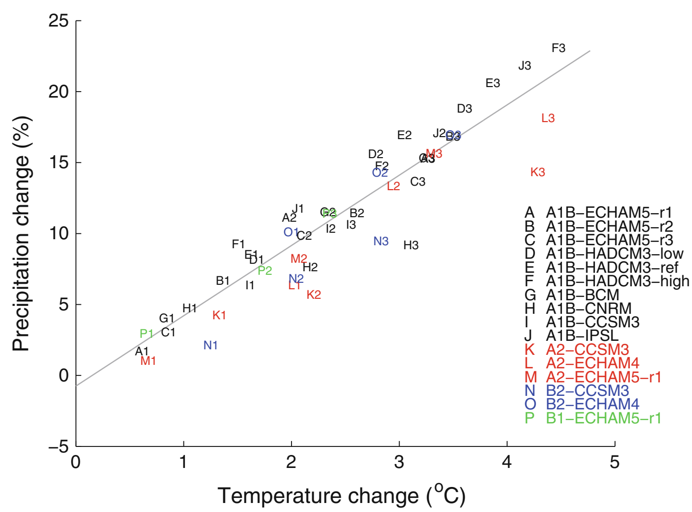

Fig. 11.2 Comparison of projected change in annual mean precipitation and near-surface air temperature over all land grid points in the Baltic Sea basin for several RCM simulations with the Rossby Centre regional climate model RCA3. Emission scenarios and forcing AOGCM for each RCA3 simulation are given in the legend (*A*–*P*). Projected change is shown for three time periods relative to 1961–1990: *1* 2011–2040, *2* 2041–2070 and *3* 2071–2100. *Colour* indicates emission scenario. The *grey line* is a least-square fit to the data (slope k = 5.0 %/°C; correlation coefficient r = 0.93). For further details see Kjellström et al. (2011a)

sorted producing an approximate 5th percentile corresponding to the lowest value, a median, and an approximate 95th percentile corresponding to the highest value. This is illustrated for surface air temperature in Fig. 11.3.

An inter-model spread similar to that in Figs. 11.1 and 11.2 is also seen in Fig. 11.3. The north–south gradient of greatest warming in the north in winter is general, but there is a spread in the magnitude of the change. This spread is somewhat smaller than in the GCM results, as shown in the two panels of Fig. 11.1: the range for the GCM results is roughly 3–9 °C in winter and 1.5–6 °C in summer for northern Sweden, whereas the range for the RCM results is around 4–8 °C in winter and 1.5–4.5 °C in summer. An RCM ensemble sampling more of the GCM uncertainty than in the ENSEMBLES data and more emission scenarios than just SRES A1B is likely to result in a greater spread. Summer warming in the Baltic Sea basin is less than the winter warming and is relatively homogeneous across the area. The above-average warming over the Baltic Sea basin may be an artefact due to the lack of a coupled ocean model in the

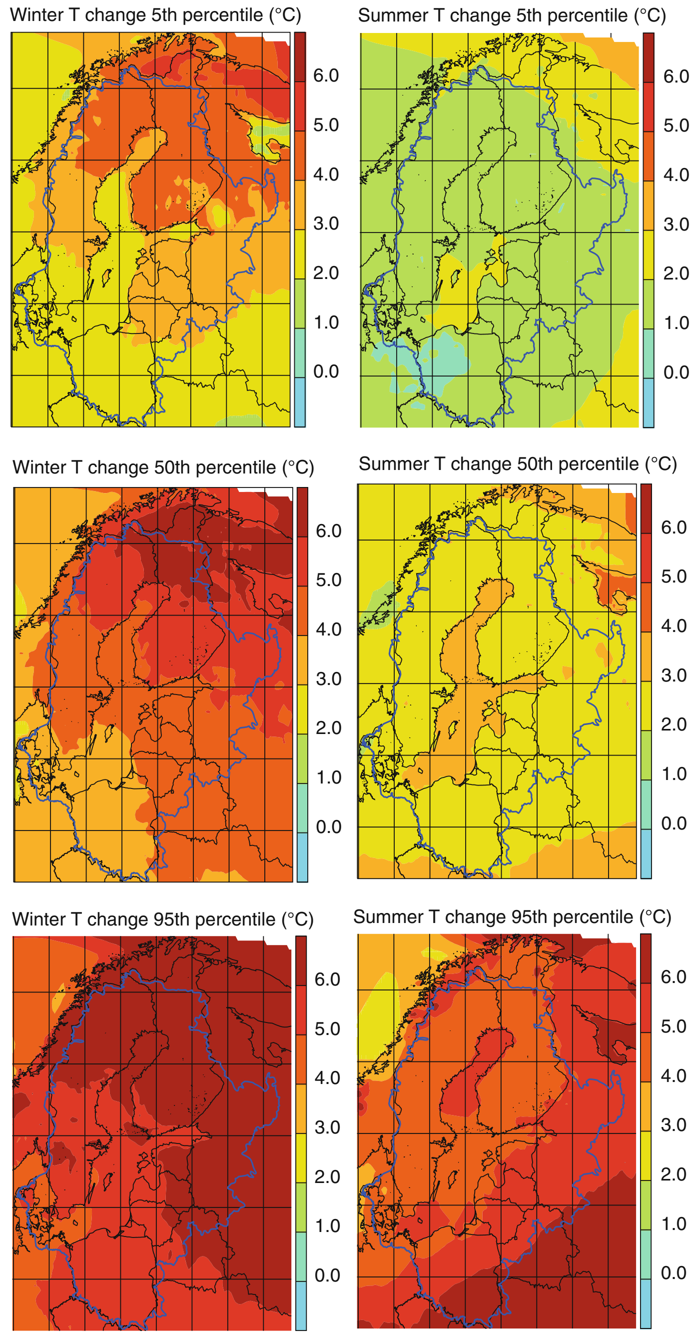

Fig. 11.3 Projected change in surface air temperature for 2070–2099 relative to 1961–1990 using the SRES A1B scenario as simulated by 13 RCM models from the ENSEMBLES project. *Left column* winter (DJF), *right column* summer (JJA). *Upper row* 5th percentile (corresponding to the lowest model result), *middle row* 50th percentile (corresponding to the median model result) and *lower row* 95th percentile (corresponding to the highest model result). The *blue line* indicates the Baltic Sea catchment area

models used here (Meier et al. 2011). The highest percentile summer warming is large in the south-east of the region. This is related to the large-scale pattern of warming in Europe with the strongest summer warming in southern Europe. Also, in the very north-east of the region, there is a large warming, probably connected to ice-albedo feedback. Similar results also exist for other GCM/RCM combinations (Christensen and Christensen 2007; Kjellström et al. 2011a).

The results shown in Fig. 11.3 are consistent with the results for an earlier period (2021–2050) based on a larger ensemble of RCM-GCMs as presented by Déqué et al. (2012). They found that even though the total uncertainty related to the choice of model combination (GCM/RCM) and sampling (natural variability) is large, it is still not enough to mask the temperature response, not even for the relatively short-term 2021–2050 time frame.

Table 11.1 Selection of ENSEMBLES RCM simulation used in this chapter

Acronym	GCM	RCM
MPI-M-REMO_ECHAM5	ECHAM5	REMO
SMHIRCA_ECHAM5-r3	ECHAM5	RCA
KNMI-RACMO2_ECHAM5	ECHAM5	RACMO2
DMI-HIRHAM5_ECHAM5	ECHAM5	HIRHAM5
SMHIRCA_BCM	BCM	RCA
DMI-HIRHAM5_BCM	BCM	HIRHAM5
DMI-HIRHAM5_ARPEGE	ARPEGE	HIRHAM5
ETHZ-CLM_HadCM3Q0	HadCM3Q0	CLM
METO-HC_HadRM3Q0_HadCM3Q0	HadCM3Q0	HadRM3
SMHIRCA_HadCM3Q3	HadCM3Q3	RCA
METO-HC_HadRM3Q3_HadCM3Q3	HadCM3Q3	HadRM3
C4IRCA3_HadCM3Q16	HadCM3Q16	RCA
METO-HC_HadRM3Q16_HadCM3Q16	HadCM3Q16	HadRM3

All simulations follow the SRES A1B scenario. For details of individual models, see Christensen et al. (2010); data are available from http://ensemblesrt3.dmi.dk/

11.6 Precipitation

A warmer atmosphere can hold more moisture, so in response to rising global temperatures, climate models also project an intensification of the global hydrological cycle (e.g. Held and Soden 2006). On a European scale, this implies more precipitation in northern Europe and less in southern Europe, both in winter and summer (Christensen et al. 2007). Between these areas of projected increase and projected decrease, there is a broad zone of 500–1000 km or more where only small changes are projected (e.g. Kjellström et al. 2011a). This transition zone shifts with the seasons and is located more to the south in winter and more to the north in summer. As a consequence, precipitation is projected to increase over the entire Baltic Sea run-off region in winter, while in summer, increased precipitation is mostly projected for the northern half of the basin only. In the south, precipitation is projected to change very little, although with a large spread between different models including both increases and decreases. There is a strong correlation between precipitation and temperature increase on an annual basis, as seen in Fig. 11.2.

Figure 11.4 shows the annual cycle of precipitation change for northern Sweden according to the same CMIP3 simulations as in Fig. 11.1. Change in the seasonal cycle reveals greater projected increases in winter than in summer when some models show only small changes and in two cases even a decrease in precipitation. Projected changes in the seasonal cycle for individual models differ more than the corresponding changes in temperature implying that the overall uncertainty is greater for precipitation.

Figure 11.5 shows the projected change in precipitation by the end of the twenty-first century for the same 13 ENSEMBLES simulations as in Fig. 11.3. For both winter and summer, there is a clear north–south gradient: the further north the more positive the change. An exception is the Norwegian west coast where relatively small increases, or even decreases, are projected in winter and in some case also in summer. This relatively small increase is related to change in the large-scale atmospheric circulation as the wind direction relative to the Scandinavian mountains largely determines the sign of change (e.g. Räisänen et al. 2004). Over the rest of the domain there is a clear increase in winter and changes of both signs in summer with an indication of a positive signal in the southernmost parts of the area. With a larger set of ENSEMBLES simulations, but a shorter-term future period, Déqué et al. (2012) found significant positive

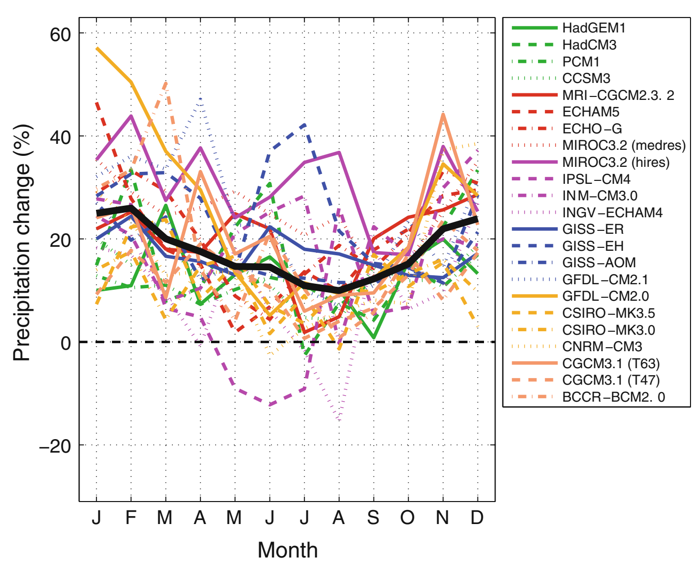

Fig. 11.4 Projected change in average monthly precipitation in northern Sweden for 2071–2100 relative to 1961–1990. Results for 23 CMIP3 AOGCM-simulations under the SRES A1B scenario (Lind and Kjellström 2008). The *thick black line* shows the average of the individual model results and the *dashed line* indicates no change

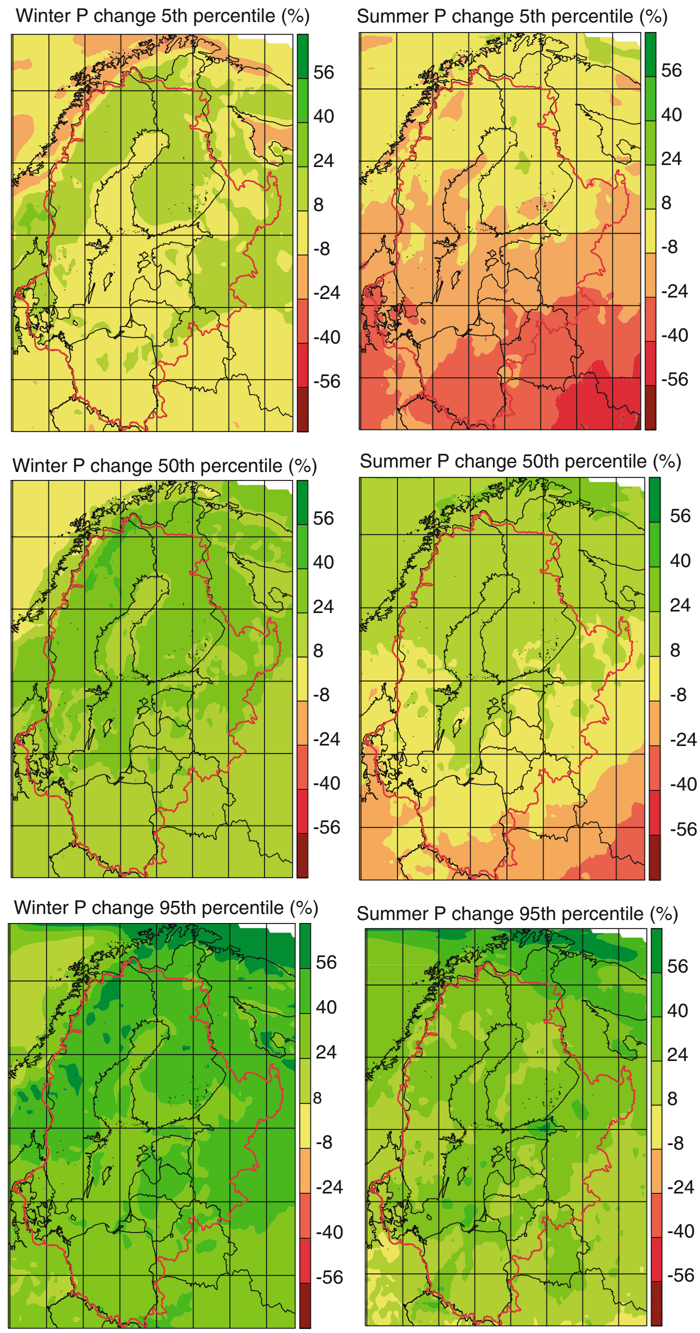

Fig. 11.5 Projected change in average precipitation for 2070–2099 relative to 1961–1990 using the SRES A1B scenario as simulated by 13 RCM models from the ENSEMBLES project. *Left column* winter (DJF), *right column* summer (JJA). *Upper row* 5th percentile (corresponding to the lowest model result), *middle row* 50th percentile (corresponding to the median model result) and *lower row* 95th percentile (corresponding to the highest model result). The *red line* indicates the Baltic Sea catchment area

summer precipitation signals for almost all land points in the Baltic Sea catchment. As discussed in Sect. 11.5 for temperature, the spread in Fig. 11.5 for precipitation may also differ given a different set of GCM/RCM combinations as indicated by the spread in the larger range of GCMs in Fig. 11.4.

Extreme weather events are very important for many aspects of society. Extreme precipitation is responsible for urban flooding, and this aspect of anthropogenic climate change has received considerable attention (see Chap. 22). As the water-holding capacity of the atmosphere increases under a warmer climate, precipitation extremes are also projected to increase (e.g. Lenderink and van Meijgaard 2010).

Building on the PRUDENCE project, Christensen and Christensen (2003) reported that projections showing a considerable decrease in average summer precipitation also showed an increased likelihood of very extreme precipitation. More intense precipitation can be expected on time scales ranging from single rain showers to long-lasting synoptic-scale precipitation.

It is likely that changes in precipitation extremes of a shorter duration may exceed those for longer time scales, as indicated by Lenderink and van Meijgaard (2010). They showed that the change in hourly precipitation extremes in one RCM considerably exceeded the prediction from the theoretical Clausius–Clapeyron relation that sets an upper bound on the water vapour content of the atmosphere.

As an example of changes in daily precipitation, Nikulin et al. (2011) investigated an ensemble of RCM simulations with the RCA model and showed that the 20-year return value of precipitation extremes in the 1961–1990 period was projected to decrease to 6–10 years in 2071–2100 for summer over northern Europe and to 2–4 years in winter in Scandinavia. Similarly, Larsen et al. (2009) reported that the return period for 20-year rainfall events on a 1-hour basis decreased to 4 years over Sweden based on a high-resolution RCM integration.

For the Rhine catchment, Hanel and Buishand (2011) investigated annual maxima of daily precipitation in 15 RCM simulations from the ENSEMBLES project and found an overestimation of the amount of these extremes, particularly in summer, when compared to a gridded observation

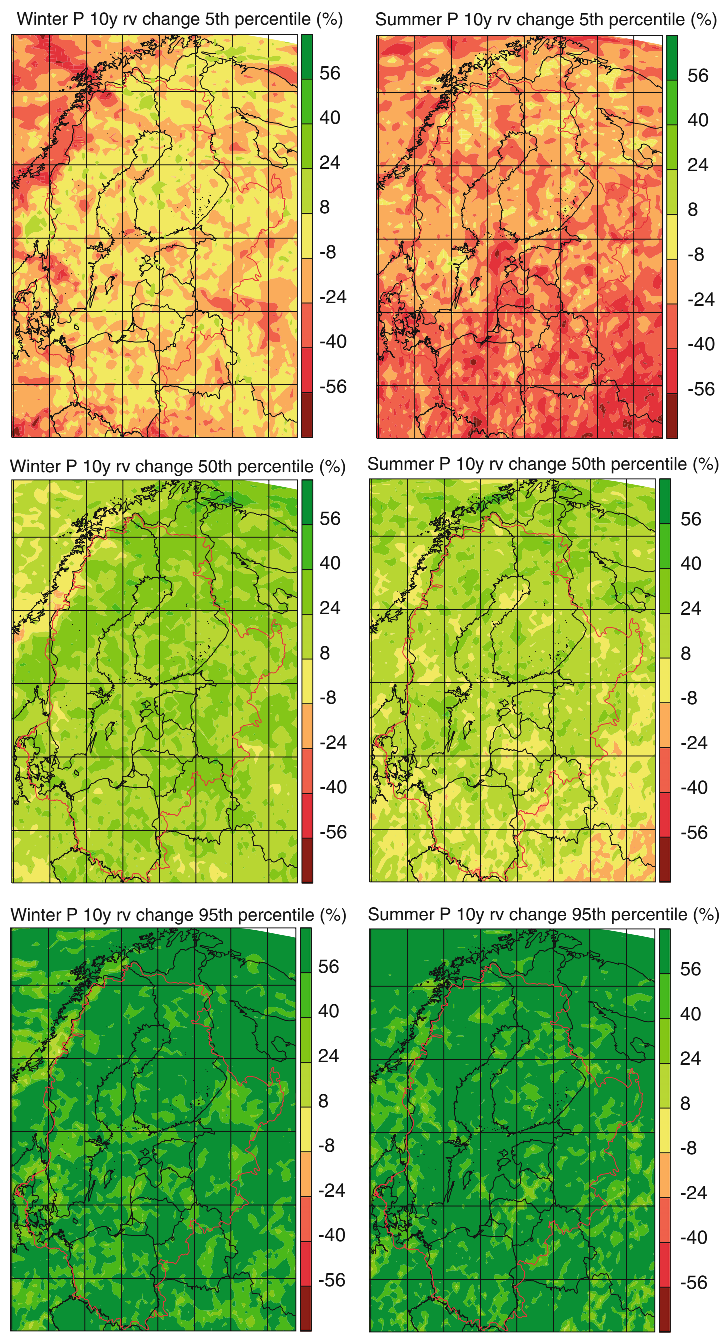

Fig. 11.6 Projected change in extreme precipitation calculated as 10-year return values for 2070–2099 relative to 1961–1990 using the SRES A1B scenario as simulated by 13 RCM models from the ENSEMBLES project. *Left column* winter (DJF), *right column* summer (JJA). *Upper row* 5th percentile (corresponding to the lowest model result), *middle row* 50th percentile (corresponding to the median model result) and *lower row* 95th percentile (corresponding to the highest model result). The *red line* indicates the Baltic Sea catchment area

set; this was partly attributed to a low density of observations used in constructing the gridded data set. The RCM models all projected increases of extreme precipitation with long return periods.

An analysis covering the 13 models from the ENSEMBLES project listed in Table 11.1 is illustrated in Fig. 11.6. The change in extreme precipitation is shown calculated as 10-year return values (the daily precipitation amount so large that it will be exceeded only once every 10 years on average). The median signal is consistently positive across the domain. The increase in the Baltic Sea basin is of the same order for both summer and winter, but the inter-model spread is larger in summer, corresponding to the greater influence of local processes in this season. It is apparent that the relative change of the extreme precipitation in winter (Fig. 11.6) looks very much like the relative change in average precipitation (Fig. 11.5), indicating no change in the shape of the intensity distribution function. The situation is very different for summer, where the projected change in extreme precipitation is considerably more positive than the change in average precipitation.

11.7 Wind

Changes in the wind climate are even more uncertain than is the case for the precipitation climate, both for seasonal mean conditions and for extremes on shorter time scales (e.g. Kjellström et al. 2011a; Nikulin et al. 2011). Figure 11.7 shows average changes over the Baltic Sea in 16 RCA3 simulations at three different time horizons in relation to the temperature change. It is evident that the correlation is much weaker than in the case of precipitation (cf. Fig. 11.1) and also that the changes are relatively small including mostly increasing, but in a few cases for one GCM decreasing wind speed. In many of the integrations, increasing wind speed is seen over ocean areas that are ice-covered in today's climate but not in the future climate, probably due to reduced static stability in the lower atmosphere as the surface gets warmer (e.g. Kjellström et al. 2011a).

The 13 ENSEMBLES RCMs of Table 11.1 are plotted in Fig. 11.8 for the projected change in average wind speed on a seasonal basis. The span of projected change is smaller than seen in Fig. 11.7 due to the small number of GCMs and the use of only one emission scenario (SRES A1B). For both winter and summer, the sign of the change varies but with a slight tendency towards an increase, particularly over sea areas (as is also indicated in Fig. 11.7).

Extremes of wind speed are relevant for projections of change in storm frequency, although it should be noted that wind speeds in RCMs are grid point averages as well as averages over model time steps of a few minutes. The potential for wind power is proportional to the third power of the wind speed, and so it is relevant for wind speed to investigate extremes of wind power.

Donat et al. (2011) investigated mid-century as well as end-of-century changes in the annual 98th percentile daily maximum wind in 14 ENSEMBLES RCM simulations for 2021–2050 and 11 models for 2070–2099 of which nine are part of the 13-member ensemble employed for the present

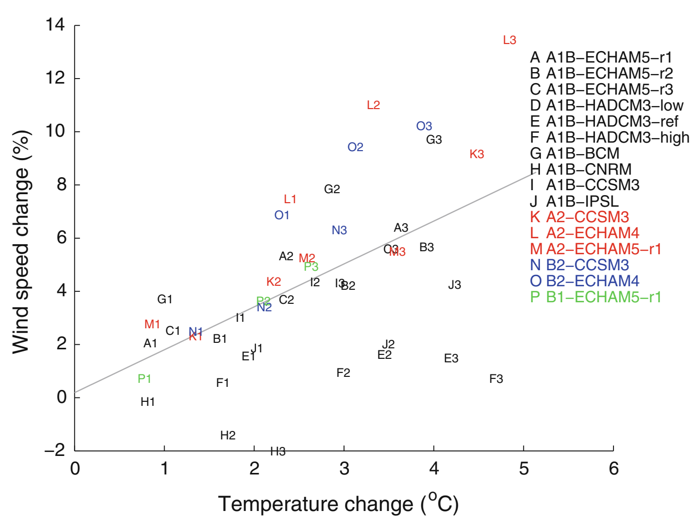

Fig. 11.7 Comparison of projected change in annual mean wind speed and near-surface air temperature over all ocean grid points in the Baltic Sea basin for several RCM simulations with the Rossby Centre regional climate model RCA3. Emission scenarios and forcing AOGCM for each RCA3 simulation are given in the legend (*A*–*P*). Projected change is shown for three periods relative to 1961–1990: *1* 2011–2040, *2* 2041–2070 and *3* 2071–2100. *Colour* indicates emission scenario. The *grey line* is a least-square fit to the data (slope $k = 1.6$ %/°C; correlation coefficient $r = 0.53$). For further details see Kjellström et al. (2011a)

analyses. The ensemble average, such as the driving GCMs, showed a tendency to increase in a belt from the British Isles to the Baltic Sea and a tendency to reduce in the Mediterranean area. Nikulin et al. (2011), based on an ensemble of one RCM downscaling six different GCMs under the A1B scenario, found increasing wind speed expressed as 20-year return periods of annual maximum wind speed over the Baltic Sea in five out of six simulations.

Figure 11.9 shows the projected change in extreme wind speed, with the 10-year return value of daily maximum wind speed plotted as an example. It should be remembered that a maximum calculated by an RCM is an average over a model time step of several minutes as well as over the grid square in question. The models are the same 13 ENSEMBLES models as in Fig. 11.6. It is clear that the spread is greater than is the case for average change (see Fig. 11.8). There is a very slight and insignificant median increase in the lower part of the area, consistent with findings by Donat et al. (2011). Generally, there are changes of both signs over the entire area, with small median decreases in winter and small increases in summer over land grid points; but there is a large inter-model spread.

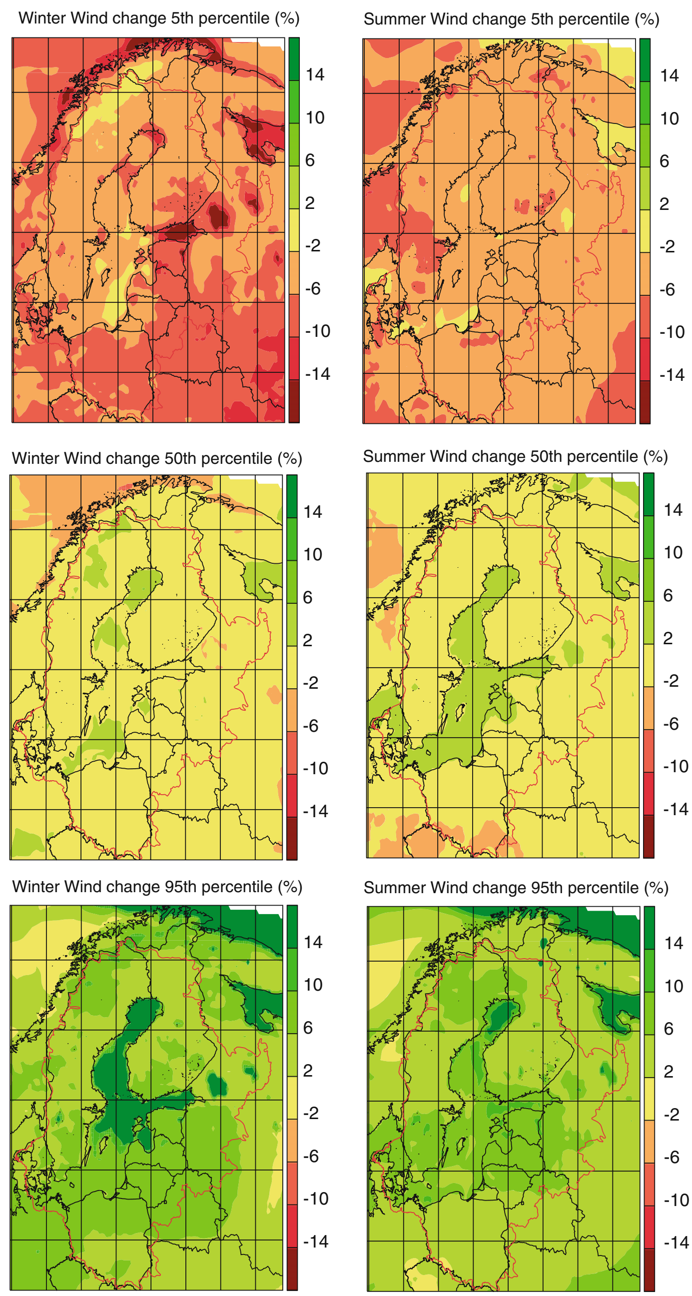

Fig. 11.8 Projected change in average wind speed for 2070–2099 relative to 1961–1990 using the SRES A1B scenario as simulated by 13 RCM models from the ENSEMBLES project. *Left column* winter (DJF), *right column* summer (JJA). *Upper row* 5th percentile (corresponding to the lowest model result), *middle row* 50th percentile (corresponding to the median model result) and *lower row* 95th percentile (corresponding to the highest model result). The *red line* indicates the Baltic Sea catchment area

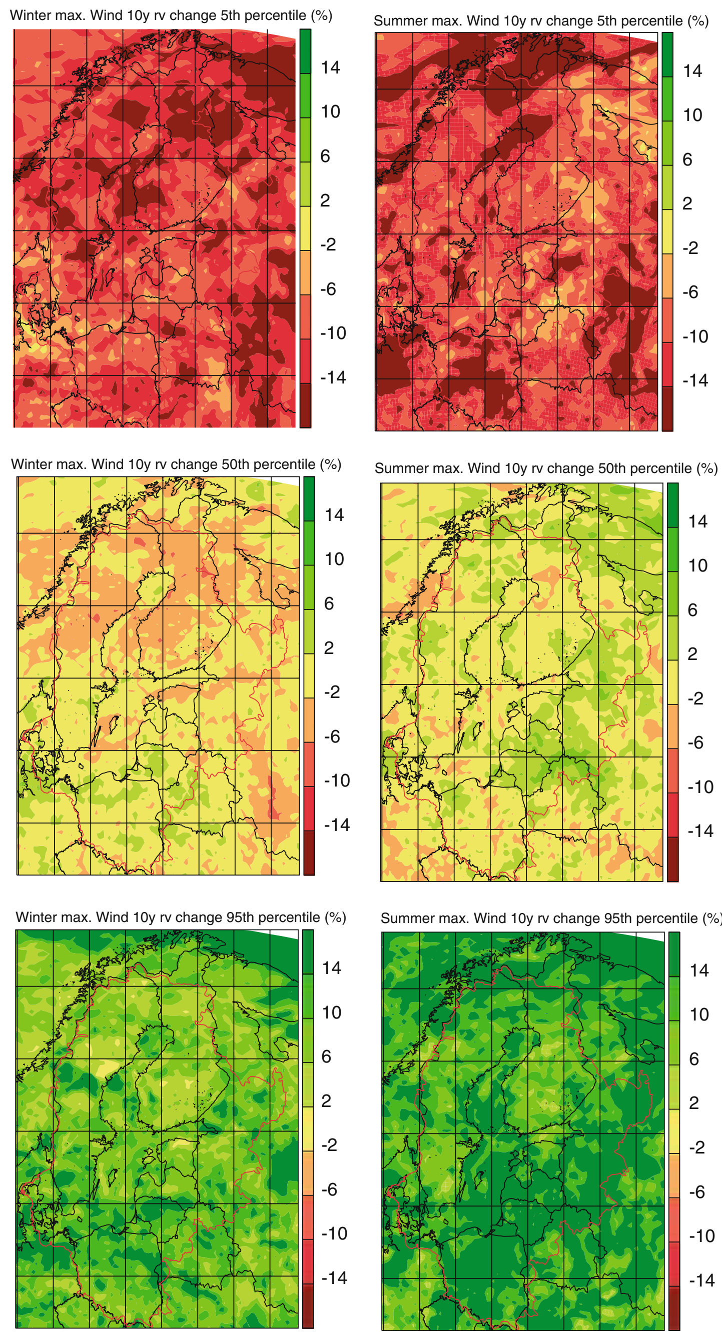

Fig. 11.9 Projected change in the 10-year return value of daily maximum wind speed for 2070–2099 relative to 1961–1990 using the SRES A1B scenario as simulated by 13 RCM models from the ENSEMBLES project. *Left column* winter (DJF), *right column* summer (JJA). *Upper row* 5th percentile (corresponding to the lowest model result), *middle row* 50th percentile (corresponding to the median model result) and *lower row* 95th percentile (corresponding to the highest model result). The *red line* indicates the Baltic Sea catchment area

11.8 Snow

Although rising temperatures are expected to lead to decreased snow cover, as more precipitation falls as rain and snow melt accelerates, they are also expected to lead to increased winter precipitation in Scandinavia, possibly compensating for the former effect. Changes in snow cover from climate models need to be analysed quantitatively in order to estimate the relative importance of these two counteracting effects.

Data from the ENSEMBLES project were analysed by Räisänen and Eklund (2011) who concluded that snow volume will decrease across Europe in the future, even though the Scandinavian mountain areas may experience a slight and statistically insignificant increase. Such an increase was also proposed by Schuler et al. (2006) in a detailed study for Norway based on two RCM scenarios

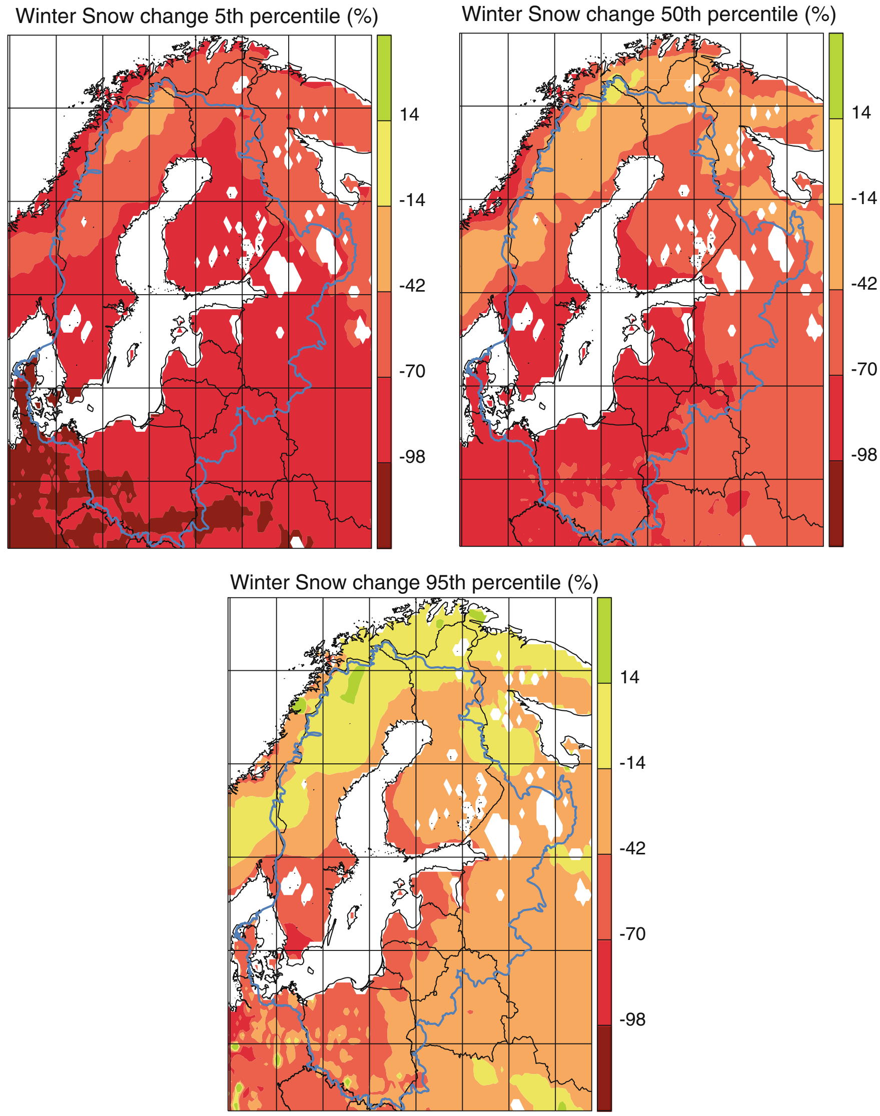

Fig. 11.10 Projected change in average winter snow amount for 2070–2099 relative to 1961–1990 using the SRES A1B scenario as simulated by 12 RCM models from the ENSEMBLES project. *Left* 5th percentile (corresponding to the lowest model result), *middle* 50th percentile (corresponding to the median model result), *right* 95th percentile (corresponding to the highest model result). The *blue line* indicates the Baltic Sea catchment

forced with different GCMs. The authors also pointed out that in extreme years, the maximum amount of snow could be greater than in extreme years of the recent past, even if snow amount is reduced on average. Figure 11.10 shows the projected change in average winter snow amount for 12 ENSEMBLES RCM models. One simulation (DMI-HIRHAM5_ARPEGE) with a known bug in the surface description was excluded because the error was judged to lead to unrealistic changes in snow amount. However, the error was not judged to have a decisive influence on the development of temperature and precipitation and so this model was included in the analyses reported in Sects. 11.5 and 11.6.

Only very small high-altitude mountain areas in a few simulations are projected to experience an increase in snow amount. The southern half of the Baltic Sea region is projected to experience significant reductions in snow amount with median reductions of around 75 %.

11.9 Statistical Downscaling

This section summarises the literature on the application of statistical downscaling methods for the Baltic Sea area that have become available since the first BACC assessment (BACC Author Team 2008). Statistical downscaling employs statistical relationships between large-scale variables (predictors) and smaller scale fields (predictands) such that, for example, robust changes in predictors as simulated by climate models can be used to estimate change in the predictands, assuming that the observation-based statistical relations themselves are stable under climate change. Predictors are frequently chosen as atmospheric fields, whereas predictands can be atmospheric, oceanic or related to ice. In this respect, a decision was made to include a review of publications on statistical downscaling even though the predictands may not be atmospheric.

As discussed in Chap. 10, Sect. 10.1, the skill of GCMs to simulate change at regional and local scales is still lower than at the continental scale. Dynamical (RCM-based) and statistical downscaling techniques are alternative approaches, which may be used to try to bridge this gap. Chap. 10, Sect. 10.3, includes a more detailed description of statistical downscaling. The validity of each of the downscaling methods depends on certain assumptions. For statistical downscaling, the strongest assumption is that the observed relationship between large-scale climate anomalies and regional climate variables does not change in the future. This assumption is quite restrictive and may not always be fulfilled. More importantly, it cannot easily be verified. The choice of large-scale predictors, their physical relationship to the regional predictands, the plausibility of the stationarity of their mutual relationship and the uncertainties arising from the statistical model itself must all be carefully weighted before the results obtained by statistical downscaling can be incorporated in a decision-making process. On the other hand, if conducted properly, statistical downscaling may provide a more explicit analysis of the sources of variability of regional climates and thus may advance understanding of the possible drivers of regional climate change in the future. For variables that are unlikely to be incorporated in RCMs in the foreseeable future, such as ecosystem variables, statistical downscaling is the only method that provides an estimate of climate change impact, although always within the caveats mentioned before. On these grounds, statistical downscaling and dynamical downscaling should be viewed as complementary.

Statistical downscaling methods have mostly been applied to climate variables that strongly depend on uncertain physical parameterisations within the models that may lack general validity, such as precipitation, cloudiness and extreme winds. The predictors used in statistical downscaling are normally chosen to be large-scale variables that are regarded as well simulated by climate models (see also Chap. 10). These variables tend to be fields with large spatial coherence, such as sea-level pressure (SLP) or geopotential height. However, it is not assured that the predictors that best describe the variability of a predictand in the twentieth century—which is the usual period for calibrating statistical downscaling models—will also be the ones that best estimate changes in the future. For instance, SLP and geopotential height are good predictors for observed seasonal mean winter precipitation in many areas of western Europe. However, in future climates, changes in precipitation may result not only from change in atmospheric circulation but also from changes in the water content of the atmosphere. Some analyses of statistical downscaling methods in the virtual world provided by climate models appear to indicate that for the Baltic Sea area this latter contribution cannot be neglected (Frías et al. 2006).

For the Baltic Sea area, statistical downscaling methods have mostly been applied to estimate future changes in hydrological variables, such as precipitation and run-off, and storm-related variables such as wind. The usual large-scale predictors are SLP and geopotential height. One particular aspect of the applications of statistical downscaling to the Baltic Sea so far is the frequent use of nonlinear statistical methods, such as weather typing, fuzzy networks and clustering algorithms, whereas for other areas linear regression methods, such as principal component regression, have generally been more frequently found in the literature.

Rogutov et al. (2008) gave an example of how a standard statistical downscaling method for precipitation should look. The authors considered the whole of western Europe, but the results are also relevant for the Baltic Sea area. The method used principal components regression, which has been applied not only for downscaling purposes but also for climate reconstructions based on proxy data since mathematically the problem is very similar. Both the predictor (SLP) and the predictand (precipitation) are decomposed by a previous principal component analysis, and the leading components are retained for further analysis. This ensures that the covariance matrices that result in the regression analysis are not singular, avoiding over-fitting of the statistical model. There is no clear rule to determine the optimal number of retained principal components, but the number can be approximately estimated by sensitivity calculations until the skill of the reconstructed predictand, when compared to observed data, does not grow.

Linear regression methods produce predictands with the same probability distribution as the predictors. Since atmospheric circulation variables tend to be approximately normally distributed, linear statistical downscaling methods may work well to estimate changes in monthly, seasonal or annual precipitation, which also tend to be approximately normally distributed. However, this is not the case for daily precipitation. In this case, more sophisticated nonlinear methods are needed, for instance those based on classification of weather types (see Chap. 10, Sect. 10.3). Wetterhall et al. (2009) employed a classification scheme constructed on a fuzzy logic algorithm to estimate changes in daily precipitation over Sweden based on the output of the GCM HadAM3H driven by the SRES scenarios A2 and B2. They also employed a weather generator that takes into account the weather type and is able to replicate the serial autocorrelation of daily precipitation. The advantage of this approach is that it is in theory able to estimate changes not only in daily precipitation amount and occurrence but also in block maxima, that is, maximum 3- or 5-day precipitation. Under these two scenarios, and conditional on the GCM used, Wetterhall et al. (2009) found that precipitation in Sweden tended to increase in the twenty-first century and that the maximum 5-day precipitation also became larger

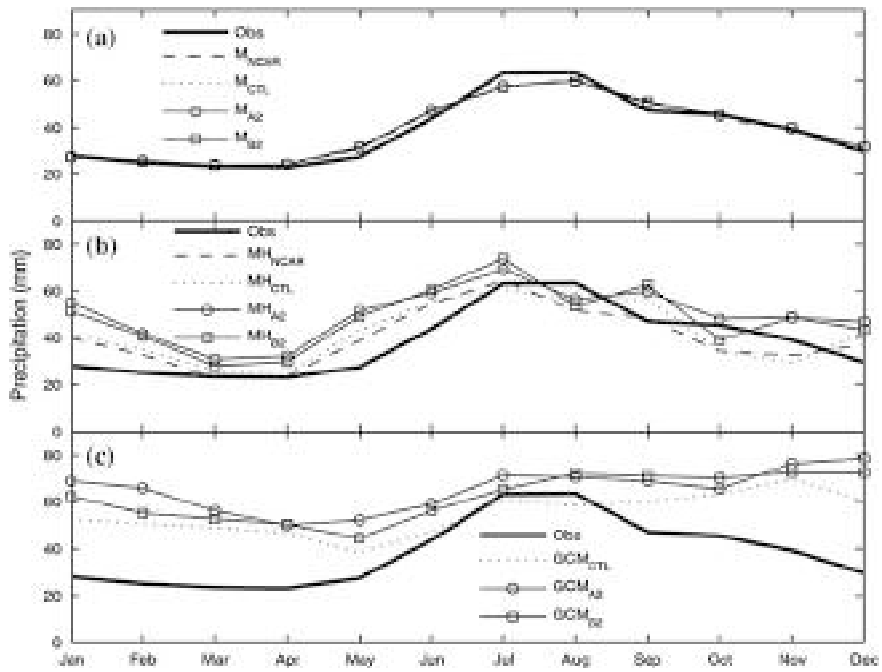

Fig. 11.11 Observed, simulated and downscaled annual cycle of precipitation over Sweden. *Bold line* represents the station observations; *dashed line*, output of the downscaling model driven by NCEP/NCAR reanalysis; *thin lines*, output of the downscaling model driven by a control climate simulation (M_{CTL}) and by SRES scenarios A2 and B2 ($M_{A2,B2}$) with the model HadAM3. The statistical downscaling model did not include specific humidity as a predictor in panel (**a**), but this was included in panel (**b**). Panel (**c**) shows the direct output from the global climate model HadAM3H for this region. (Wetterhall et al. 2009)

(Fig. 11.11, panels b and c). This increase was not due to changes in the frequency of weather patterns in the future but rather to the increase in specific humidity in the atmosphere, roughly in accordance with the tests of statistical downscaling methods in simulated climates conducted by Frías et al. (2006). These conclusions were reached by driving the statistical downscaling method by different combinations of the meteorological forcing from the GCM, as illustrated in Fig. 11.11. This figure shows that the direct output of the GCM is not able to reproduce well the annual cycle of precipitation over Sweden and that it underestimates its amplitude and is biased high (panel c). The statistical downscaling method improves the GCM simulation and reproduces well the observed annual cycle when driven by the meteorological reanalysis or by a control simulation of the present-day climate (panel a). However, the statistical downscaling method does not indicate large changes in precipitation when driven by scenario simulations excluding changes in the atmospheric-specific humidity (panel a). Only when this variable is explicitly used to drive the statistical downscaling method are future precipitation changes evident (panel b).

Downscaled precipitation can be used in conjunction with temperature to drive hydrological models and estimate changes in future run-off (see Chap. 12). However, there are important caveats to be borne in mind when applying climate model data to drive models of climate impacts, since climate impacts may be sensitive not only to the simulated relative changes in climate from the present state but also to the absolute level of temperatures and precipitation simulated by the climate model. Climate models are seldom bias free, that is, the simulated mean present climate may deviate from the observations, and sometimes by non-negligible amounts (IPCC 2007); therefore, the biases make the direct application of simulated or downscaled precipitation or temperature problematic. Often it is necessary to apply an empirical bias correction method.

Sennikovs and Bethers (2009) proposed a bias correction method for precipitation and temperature that may be subsequently applied to drive a hydrological model in the eastern Baltic Sea area. The bias correction method is based on a spatially explicit comparison between the probability distribution function of the climate variables simulated by an RCM and those derived from observations. A correction function is applied to the simulated data that aligns the simulated and observed quantiles of the probability distribution. This correction function may in general be nonlinear, although in some cases a simple realignment of the mean and a re-scaling to obtain the same variance as the observations may be adequate. Sennikovs and Bethers (2009) applied their methodology to several RCMs participating in the European project PRUDENCE (Christensen and

Christensen 2007), selecting the eastern Baltic Sea area for further analysis. They found that RCMs tend to produce a reasonable annual cycle of temperature but clearly overestimate precipitation in winter and underestimate precipitation in summer. By applying their bias correction method to daily precipitation, they were able to bring the model results much closer to observations. The same correction function would then be applied to the output of scenario simulations, under the assumption that the causes that produce the mean biases in the models remain unchanged under a future climate.

The corrected values of precipitation and temperature were used by Apsīte et al. (2010) to drive a hydrological model to estimate changes in run-off in the eastern Baltic Sea catchment area. Future run-off will be modulated by changes in two factors. On the one hand, evaporation will tend to increase due to higher air temperatures, while on the other, precipitation is expected to increase, as simulated by most RCMs participating in the PRUDENCE project (Christensen and Christensen 2007). A surprising result of the study by Apsīte et al. (2010) was that the first factor seemed to be more important in the future, and river run-off would tend to decrease according to the RCM simulations analysed. Also, important is that the annual cycle of run-off tended to change considerably, with the late spring maximum observed in the present climate shifting to earlier seasons even into the months of January and February. This is a consequence of the rising temperatures and an earlier onset of the melt season, as well as changes in the annual cycle of precipitation and increased evaporation. This represents a major shift in the annual run-off cycle that may have profound consequences on many economic sectors should it remain unmanaged by reservoirs. However, the study by Apsīte et al. (2010) is based on the mean of data simulated by 21 models and does not indicate the spread of individual simulations.

Previous projections of river run-off into the Baltic Sea indicated that the uncertainty was large enough to encompass a broad range of projections, from slight reductions to substantial increase (Graham 2004). In a further study aimed at reconstructing run-off in the past 500 years, Hansson et al. (2011) also applied a statistical downscaling method using predictors from climate reconstructions of atmospheric circulation and temperature. Although the study is not focused on future projections but rather on past evolution of run-off, their findings about past variability in river run-off were also interpreted in the context of future climate change. Hansson et al. (2011) briefly indicated that if their statistical downscaling model is correct, run-off would tend to increase in the northern Baltic Sea and decrease in the southern Baltic Sea. This result is mainly driven by the signal of increasing temperature in the northern Baltic Sea catchment area and by a decrease in precipitation in Central Europe. Chap. 12 deals explicitly with run-off projections.

The estimates of changes in extreme wind events over a few days show similar characteristics to the estimates of changes in daily precipitation, and thus, similar downscaling methods have also been applied, again in classification algorithms. Leckebusch et al. (2008) presented a cluster analysis based on the k-means method to identify the weather situations that give rise to extreme winds in western Europe. The k-means method is a standard clustering algorithm that in this case was applied after a pre-filtering by Principal Component Analysis. The same algorithm was then applied to a climate change simulation with the model ECHAM4/OPYC3 driven by an older future scenario (IS92a) of changing GHG concentrations in the atmosphere. However, the basic results of this simulation are not qualitatively different from the more modern simulations based on SRES scenarios. The authors found that the frequency of extreme winds increases over the whole of western Europe and in the southern Baltic Sea, which is consistent with the simulated increase in the intensity of the North Atlantic Oscillation (NAO) in this model. It should be noted that many GCMs project an increase in the intensity of the NAO in the future.

Finally, statistical downscaling methods have also been applied to the estimation of future changes in sea ice and snow in the Baltic Sea area (Jylhä et al. 2008). This study analyses the results of simulations by GCMs and RCMs participating in PRUDENCE over the Baltic Sea area driven by future scenarios of atmospheric GHG concentrations. However, to estimate changes in sea-ice cover the authors applied a statistical downscaling method, since the resolution of the RCM is too coarse to represent sea ice at the coastlines. The predictor in this regression model is air temperature and the regression model is calibrated using observations between 1902 and 2001. The calibrated regression model is then applied to the projected temperature change. In addition, a bias correction must be applied to account for the mean temperature bias in the model, since the formation of sea ice is a strongly nonlinear process that depends on absolute values of temperature and not only on the change in temperature. This study adopts the so-called delta change correction, which amounts to realigning the temperature simulation in the present climate simulation with the observed mean temperature, conserving the temperature change signal as given by the differences between scenario and present-day simulation. The main conclusion of the study is that coastal sea-ice cover will be dramatically reduced in the coming decades regardless of the future GHG emission scenario, even though SRES scenario A2 foresees the largest future GHG emissions and yields the largest reductions in sea-ice cover in the Baltic Sea (see also Chap. 13, Sect. 13.4 for potential changes in Baltic Sea ice).

11.10 Conclusion

Several numerical climate change simulations have been undertaken since the first BACC assessment (BACC Author Team 2008). Models now operate at higher horizontal resolution. Furthermore, the simulations cover a larger degree of the uncertainty range including: a wider range of emission scenarios (sampling the uncertainty in forcing), more climate models (addressing model uncertainty) and ensemble members (addressing natural variability). The picture emerging from these simulations confirms the findings of previous studies (e.g. BACC Author Team 2008) in terms of climate change in the Baltic Sea region. Climate model studies suggest that

- The future climate will get warmer, especially in winter. Changes increase with time and/or rising emissions of GHGs. There is a large spread between the different models, but they all project warming.
- Cold extremes in winter and warm extremes in summer are expected to change more than the average conditions, implying a narrower (broader) temperature distribution in winter (summer).
- Future precipitation will be higher than today. The increase is projected to be greatest in winter. In summer, models project an increase in the far north and a decrease in the south. For the transition zone between these two regions, the sign of change is uncertain.
- Precipitation extremes are expected to increase although with a higher degree of uncertainty compared to the projected change in temperature extremes.
- Future changes in wind speed are highly dependent on changes in the large-scale atmospheric circulation simulated by the GCMs. The results diverge and it is not possible to estimate whether there will be a general increase or decrease in wind speed in the future. A common feature of many model simulations, however, is an increase in wind speed over oceans that are ice-covered in the present climate but not in the future. Future changes in extreme wind speed are uncertain.
- The increased number of high-resolution regional model studies driven by many different GCMs has enabled a tighter connection with hydrological models, even though various forms of bias correction are necessary as an interface between the two types of model. Furthermore, the large number of available simulations enables some estimation of uncertainty of impacts.
- Statistical downscaling studies using atmospheric predictors have addressed several predictands, with the greatest emphasis given to hydrological variables. The findings of detailed studies have been in line with those from studies employing dynamical downscaling.

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12 Projected Change—Hydrology

Torben Obel Sonnenborg

Abstract

This chapter reviews studies on projected hydrological changes within the Baltic Sea catchment area published since the first assessment of climate change in the Baltic Sea region in 2008. Hydrological impact studies have been carried out in almost all countries in the area. The large differences in hydrological conditions (present and projected) from northern Scandinavia to the southern Baltic Sea area are addressed. The chapter considers the impacts of snow accumulation and melt, river discharge and flooding. Water resources with studies on dry periods and groundwater resources are also covered. In contrast to the first assessment, uncertainty has received significant attention. In contrast to traditional hydrological studies, projections of climate impacts on hydrology are associated with uncertainties related to the models and greenhouse gas (GHG) emission scenarios used. In several studies, individual uncertainty sources are quantified and compared.

Keywords

Hydrological change • Hydrological modeling • Climate model projection uncertainty • Impact model uncertainty

12.1 Introduction

This chapter reviews studies on projected hydrological changes within the Baltic Sea catchment area published since the first assessment of climate change in the Baltic Sea region (BACC Author Team 2008). The majority of the studies have been performed at a national level. This chapter is therefore structured by country within the Baltic Sea catchment area. Impact assessments associated with the changes in climate and hydrology are found in Chaps. 15–22.

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12.2 Country-Specific Projections

12.2.1 Belarus

No separate studies on catchments in Belarus were found. However, the river Neman/Nemunas covers part of Belarus and is addressed under Lithuania (Sect. 12.2.7).

12.2.2 Denmark

Thodsen et al. (2008) analysed the impacts of direct and indirect effects of climate change on suspended sediment transport in Danish rivers. Two lowland catchments were examined, an alluvial catchment dominated by sandy soils and a moraine catchment dominated by clay soils. Both catchments were modelled using a rainfall-run-off model for discharge and a regression model for the relation between discharge, precipitation and suspended sediment transport. Climate input from the Danish regional climate model (RCM)

The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_12

HIRHAM representing the SRES A2 scenario for 2071–2100 was used. The effect of climate change appeared to be more important than the change in length of the growing season. While changes in mean annual precipitation of 6–7 % were projected, the mean annual river discharge was projected to increase by 11–14 %, and the sediment transport by 24–27 % in the moraine catchment and 9–17 % in the alluvial catchment, depending on the growing season scenario used. Hence, the study indicates that the climate change effect increases down the chain—from precipitation to discharge to sediment transport—presumably because of strong seasonal changes.

Jeppesen et al. (2009) used the NAM rainfall-run-off model at ten small catchments in Denmark to quantify the effects of the climate projection provided by the ECHAM4/OPYC general circulation model (GCM) (SRES A2 scenario) downscaled by the Danish HIRHAM RCM (25-km grid). Annual discharge for the period 2071–2100 was projected to increase by 9–34 % compared to the control period 1961–1990. Strong seasonal changes were observed in response to seasonal changes in precipitation, with monthly run-off generally increasing in winter and decreasing in late summer. The projected discharge was used with a statistical phosphorus loading model. This resulted in an increase in phosphorus loading of 3.3–16.5 % and a decrease in phosphorus concentration of −2.2 to −13.4 %.

van Roosmalen et al. (2007) used the distributed, physically based hydrological model MIKE SHE to evaluate the impact of climate change on two large-scale catchments, a sandy area in the western part of Denmark and a clay area in the eastern part. The two models were forced by projections from the PRUDENCE project (Christensen and Christensen 2007) as simulated by the HIRHAM RCM nested in the HadAM3H GCM. The SRES A2 and B2 scenarios for the period 2071–2100 were considered, and the delta change approach was used for bias correction (see Chap. 10, Sect. 10.3.1.2). The authors found the magnitude of the hydrological response to be highly dependent on the geological settings of the model area. In the sandy catchment, groundwater recharge increased as a result of higher winter precipitation. Groundwater levels increased significantly, up to 3 m, resulting in elevated base flow and drain flow to the rivers. Significant increases in winter river discharge occurred, while summer discharge was only slightly affected. In contrast, only small changes in groundwater level occurred in the clay area since the infiltration capacity of the less permeable shallow geology was exceeded. Here, the increase in precipitation resulted in significant changes in drain flow and overland flow to rivers, with changes in mean monthly river discharge of up to +50 % in winter and −50 % in summer.

In a subsequent study, van Roosmalen et al. (2009) used the same model set-up to quantify the combined effects of climate change, sea level rise and change in land use on irrigation demand and water resources. The study showed that climate change had the most substantial effect on the hydrological system. However, indirect effects were also found to be significant. Irrigation demands were found to increase by up to 90 % for the SRES A2 scenario. Although groundwater levels generally increased with increasing mean precipitation, the water content of the root zone was found to decrease during summer in response to decreasing summer precipitation and increasing evapotranspiration. Changes in land cover from grass to forest and changes in growing season length resulted in minor effects on groundwater recharge. More significant effects were found when assuming that plant respiration became more water efficient with increasing carbon dioxide (CO_2) concentrations. A simple approach was used where future potential evapotranspiration (PET) was assumed to equal present PET and not to increase in response to the projected rise in future temperature. The resulting reduction in actual evapotranspiration was found to have a relatively large impact on groundwater recharge, groundwater levels and stream discharge. Sea level rise of 0.5 and 1.0 m was shown to have a pronounced effect on groundwater levels in coastal areas, where groundwater was affected up to 10 km inland from the coast.

van Roosmalen et al. (2011) studied the impact of the bias correction method on the response of a hydrological model. The delta change method (Chap. 10, Sect. 10.3.1.2), where the observed database of meteorological variables are perturbed according to the changes projected by the climate model, was compared to a distribution-based scaling (Chap. 10, Sect. 10.3.1.1; Piani et al. 2010), where the projected meteorological variables are corrected. In contrast to the delta change method, the distribution-based scaling method reproduced the dynamics of the climate model, for example prolonged periods of drought and possible changes in the number of days with precipitation. Comparing the hydrological simulations using both methods, only small differences in the hydrological variables were found. Only average quantities were analysed, however, such as annual groundwater recharge, mean change in groundwater level or mean monthly river discharge. The authors recommended that additional analyses are needed, addressing extremes and sensitivity to catchment characteristics, such as sandy versus clay catchments.

12.2.3 Estonia

No recent studies on the impact of climate change on basins in Estonia were found.

12.2.4 Finland

The effect of climate change on design floods was evaluated for 34 dams in Finland by Veijalainen and Vehviläinen (2008).

Design floods are hypothetical floods defined by their probability of occurrence and are used for planning and floodplain management. A 14-day design precipitation with a 1000-year return period was generated for both present-day and future climate conditions. The design precipitation, which depends on location and time of year, is defined such that the largest precipitation occurs on the 9th day of the period and corresponds to the 1000-year maximum of 1-day precipitation, while the precipitation sum of days 7–11 corresponds to the 1000-year maximum 5-day precipitation. The design storm is moved through the 40-year observation period and the 30-year future period. Results representing 2071–2100 from three GCMs, three emission scenarios and two estimates of design precipitation were applied in order to evaluate the uncertainty of the resulting change in design flood. The rainfall-run-off model WSFS was used. In northern Finland, the timing and magnitude of the floods were on average found to remain unchanged. Warmer winters with less snow accumulation were partly compensated for by increases in winter and spring precipitation. On dams in western and central Finland, the design floods increased, while the timing remained unchanged (summer and autumn). In eastern Finland, the time of the design floods changed from spring to summer, while both increasing and decreasing magnitudes of the design floods were found. The range of changes in the simulated design floods was large at most sites resulting in projections of both increasing and decreasing design floods at several dams. However, the relative contribution from the individual uncertainty sources was not quantified. The authors drew attention to the fact that additional uncertainty sources such as the impact model type or the parameters of the impact model would result in even larger uncertainty ranges.

Veijalainen et al. (2010a) assessed the impact of climate change on the regulation of three lakes in eastern Finland. The rainfall-run-off model WSFS was forced by results from 14 projections of climate change generated by combining four GCMs, where one was found as an average of 19 GCMs (IPCC 2007) and three SRES scenarios. Two scenarios were downscaled using the RCA3 RCM. The delta change approach was used for bias correction in all cases. Clear changes occurred in the seasonality of run-off and water levels, with decreases in late spring and summer and increases in late autumn and winter. The changes were primarily due to changes in snow accumulation and melt with changes in precipitation and evaporation less important. Current regulation permits were found to be unsuited for the projected future hydrological conditions in many lakes.

The effects of climate change on discharge and fluvial erosion potential were studied by Lotsari et al. (2010) for a sub-Arctic catchment in the border area between Finland and Norway in northern Fennoscandia. Impact modelling was carried out by combining results from the rainfall-run-off model WSFS with the two-dimensional hydraulic model TUFLOW. Future scenarios using three different emission scenarios and two different GCMs were applied. Additionally, two GCM projections were downscaled using the RCA3 RCM. The period 2070–2099 was considered, and the delta change approach was applied for bias correction. Based on annual maximum discharges projected by WSFS, floods with return periods of 2 years and 250 years were estimated using an extreme value-type I distribution. The floods were used as input to the hydraulic model. For all eight scenarios, the flood with a return period of 2 years was found to decrease in a future climate, while for the 250-year flood a decrease was found in seven out of eight cases. The reason for decreasing flood discharges was found to be a decline in snowfall together with a shorter snow accumulation period and warm spells with snow melt during winter. As a result, future erosion power was reduced, that is diminishing flow velocity, bed shear stress and stream power.

Veijalainen et al. (2010b) used the same methodology as Lotsari et al. (2010) to assess the impacts of climate change on flooding in Finland. However, the number of climate change projections was increased to 20, including three emission scenarios (SRES A2, A1B and B1), five GCMs and four RCMs. The delta change approach was used for transferring the climate model results to the hydrological model. Changes in flooding were evaluated at 67 study sites covering Finland. The 100-year floods were on average found to decrease by 8–22 % in 2070–2099 compared to the reference period 1971–2000. However, considerable variation between regions was observed. In areas currently dominated by spring snowmelt floods, the 100-year flood generally decreased due to less snow accumulation. In areas where autumn and winter flooding currently occur frequently, the projected increases in temperature and precipitation result in increasing floods. For the central lakes characterised by long-lasting volume floods, a clear increase was found. The changes in discharge were not linearly reflected in flood area extent. The characteristics of the river channels and floodplains were found to greatly influence the spatial extent of flood inundation. Flat floodplains showed a larger change in inundation than the projected change in discharge, whereas floodplains with greater variation in topography experienced less change in inundation. Generalisations based on a few case studies or a few climate scenarios in countries with variable hydrological conditions should be avoided, however.

Based on a statistical relation between the elevation of the groundwater table and snowmelt, precipitation and evapotranspiration, Okkonen and Kløve (2010) projected the fluctuations in the groundwater table for the period 2010–2039. The analysis was based on a lumped conceptual water balance model and climate projections using the SRES A2 scenario (climate model unspecified). The elevation of the groundwater table for a catchment in central Finland was found to increase in winter and decrease in summer. However, the changes were small with differences in mean monthly values of up to about 20 cm.

12.2.5 Germany

No studies on German catchments areas in the Baltic Sea basin were found.

12.2.6 Latvia

Apsīte et al. (2011) evaluated the impact of climate change on river run-off at eight river basins in Latvia for 2071–2100. Based on the SRES A2 and B2 scenarios, results from the GCM–RCM combination HadAM3H-RCAO were used to force a conceptual rainfall-run-off model. While mean annual temperature, precipitation and evapotranspiration were projected to increase, the mean annual river run-off was projected to decrease by 2–24 %. However, strong seasonal changes were found especially in winter, spring and autumn. A shift in the timing of maximum run-off from spring to winter was projected, and winter run-off was shown to increase significantly. Summer run-off was not projected to change much. A higher frequency of days with heavy rainfall was also projected. It was concluded that the river run-off regime will become similar to that of present-day western European rivers with two principal periods, one with high flow during winter and one with low flow during summer.

12.2.7 Lithuania

Kriaučiūnienė et al. (2008) carried out an impact assessment of climate change on the river Nemunas located in eastern Lithuania and western Belarus. Based on projections from two GCMs (ECHAM5 and HadCM3), each forced by three emission scenarios (SRES A2, A1B, and B1), the delta change method was used to transfer the climate change signal to the conceptual rainfall-run-off model HBV. The changes in river run-off were forecast for five 10-year periods covering 2011–2100. The impact of climate change increased over time and generally showed the same tendency during the scenario period. River discharge decreased significantly, and the spring flood became earlier and decreased greatly. The largest changes in average river discharge were found for the SRES A1B scenario, with a decrease of up to 41 %, whereas the effects for A2 and B1 were comparable and less significant. The reason is that temperature increased the most for the A1B scenario, whereas precipitation was almost unchanged. Hence, locally, the good scaling between temperature increase and change in precipitation shown in Chap. 11, Fig. 11.2 may not be valid. Owing to the rise in temperature, the probability of snowfall decreased resulting in a reduced spring flood with the effect that the maximum flood discharge also decreased significantly.

The study by Kriaučiūnienė et al. (2008) was extended by Kriaučiūnienė et al. (2009) to include the impact of hydrological model uncertainty. Using a generalised likelihood uncertainty estimation (GLUE; Beven and Binley 1992) methodology, the impact of model parameter uncertainty was compared to the uncertainty from the choice of GCM and emission scenario. One thousand parameter sets were initially chosen using a Monte Carlo method and subsequently ranked according to a likelihood function that expresses the match to the observed data. A threshold value was defined in order to select the parameter sets resulting in the best match to the observed discharge. The choice of emission scenario was found to be responsible for 75 % of the uncertainty, while the choice of GCM was responsible for 18 %. Hence, the emission scenario has a greater influence on forecasting run-off than the choice of GCM. However, it should be noted that only two GCMs and three emission scenarios were analysed. The impact of parameter uncertainty was only 7 % of the total uncertainty. However, the significance of the parameter uncertainty depends to a high degree on the choice of threshold value used in the GLUE methodology, which is subjective. If all 1000 parameter sets are considered, the parameter uncertainty accounts for 23 % of the total uncertainty.

12.2.8 Norway

Beldring et al. (2008) presented projections of climate change impacts on four catchments in Norway using the rainfall-run-off model HBV based on scenarios from two GCMs (HadAM3H and ECHAM4/OPYC3), one RCM (HIRHAM), two SRES scenarios (A2 and B2) and two methods for bias correction of climate model results. The choice of bias correction method was found to have an especially important effect on the projections of river discharge. The delta change method was compared to an empirical adjustment procedure developed by Engen-Skaugen (2007) that not only adjusts the mean but also the variance of the RCM data. The delta change approach was found to overestimate temperatures around 0 °C because the same changes are applied to all temperature intervals resulting in an overestimation of snowmelt intensity. The empirical adjustment procedure of Engen-Skaugen (2007) was found to be more accurate than the delta change approach as changes in the frequency of temperature and precipitation events and trends in the climate scenarios are preserved. The most important impacts of climate change were found to be earlier snowmelt and reduced snow storage, with the result that snowmelt floods occur earlier and winter and autumn discharge decrease. The changes were found to be caused more by change in temperature than precipitation.

12.2.9 Poland

Szwed et al. (2010) analysed the impact of climate change on water resources in Poland with a focus on agricultural effects. Based on results from the GCM–RCM combination, ECHAM5-MPI-M-REMO a water balance given by precipitation minus evapotranspiration was calculated for 2061–2090. The water balance for the country as a whole is projected to become more negative, indicating increasing water deficit. A cumulative probability curve for water deficit shows a shift to lower values, with a decrease in the median value from −32 to −50 mm. In addition, the maximum number of consecutive dry days is projected to increase across most of Poland. The authors concluded that the water budget is likely to become increasingly stressed, meaning greater additional water supplies would be needed to exploit the agro-potential of the environment. However, the already limited water resources of Poland do not allow large-scale irrigation and so the situation is likely to become increasingly severe in the future.

12.2.10 Russia

No specific modelling studies on future catchment discharges to the Baltic Sea are available. However, part of Russia is included in the study by Veijalainen et al. (2010a) reported in Sect. 12.2.4 for Finland.

12.2.11 Sweden

Yang et al. (2010) used results from the GCM/RCM ECHAM5/RCA3 forced by the SRES A1B scenario as input to the rainfall-run-off model HBV (Bergström 1976) to quantify the effects of projected climate change on three catchments located in the northern, middle and southern part of Sweden. Two methods for adjusting the output from the RCMs were tested: the delta change method (Chap. 10, Sect. 10.3.1.2) and an approach referred to as distribution-based scaling (DBS; Chap. 10, Sect. 10.3.1.1). While the delta change method uses observed data as a baseline and only the mean is adjusted, the DBS method uses the RCM results as baseline and adjusts the entire frequency distribution. The DBS approach was found to better preserve the future variability of the RCM output. Based on comparison of future discharge from the HBV model, greater variability in discharge was found using the DBS-adjusted data resulting in, for example, larger extreme discharges than the delta change approach. DBS was found to be more sensitive to the projections used and preserved the annual variability from the corresponding climate model projection.

Olsson et al. (2011) used an ensemble of climate projections to investigate uncertainties in hydrological changes. Twelve climate model projections were used to simulate the inflow to Lake Vänern by the HBV rainfall-run-off model. Results from three GCMs were downscaled using four RCMs operated at different resolutions. In all cases, the climate signal from the RCM was adjusted using DBS, which was established for the reference period 1961–1990. The impact of emission scenario, GCM, initial conditions for the GCM, RCM, and resolution of the RCM were examined. All projections were found to accurately reproduce the observed discharge to the lake during the reference period. Subsequently, the changes calculated by the 12 models for the period 1991–2008 were evaluated against observed changes. The performance of the different projections varied widely with respect to simulating changes in monthly mean discharge. All projections underestimated the observed increase in January–February, and most overestimated the discharge in March. During the rest of the year, most scenarios were out of phase with the changes simulated using the observed precipitation and temperature. Projections for 2009–2030 suggested that winter discharge would increase and that summer and autumn discharge would decrease. However, the projections disagree on both the sign and magnitude of changes for all months of the year. The greatest sources of uncertainty were identified as the GCM used and its initialisation, which is in contrast to the study of Kriaučiūnienė et al. (2009) for Lithuania who concluded that the emission scenario was the most important source of uncertainty. The RCM and its resolution were found to have a smaller influence. It was found that the difference in hydrological change using the SRES A2 scenario compared to the A1B scenario is larger than the difference between scenarios B1 and A1B. This indicates that the effect of the emission scenario may depend on the specific scenario chosen. The effect of GCM initialisation shows the importance of natural variability within the models on the resulting climate change projections. The authors indicated the importance of further development of climate models that are in phase with the historical climate and thus potentially able to generate decadal forecasts rather than projections. However, several uncertainty sources were not covered in the study, including the choice of bias correction method and the choice of hydrological model.

Wetterhall et al. (2011) used a response surface approach to handle the uncertainties in impact scenario modelling.

A framework consisting of four steps was proposed, including the definition of scenarios of changes in key climatic variables (e.g. precipitation or temperature), performing a sensitivity analysis of the climate change using the impact model, identifying critical thresholds (e.g. extreme flows) and evaluating the probability exceedance of the thresholds. The response surfaces are generated by perturbing the observed time series incrementally and using the resulting data as input to the impact model, in this case the rainfall-run-off model HBV. The probability of exceeding the defined threshold is plotted on diagrams as a function of the key climatic variables. This information can subsequently be used as an easy way to estimate the risk of reaching a predetermined threshold given information on projected changes in the key climatic variables from one or more climate models. Based on three test cases, the Lule River, Lake Vänern and Lake Mälaren, the method was found to provide a visualisation tool for expressing probabilistic hydrological change that is able to assess the uncertainties caused by climate models. Results suggested that low water levels in Lake Mälaren are likely to become more common in future, while the run-off at Lule River is very likely to increase in winter and spring and decrease in summer. Wetterhall et al. concluded that the thresholds should be selected carefully based on a good understanding of local conditions. The method also requires that the climate change can be expressed by a few variables, for example change in mean temperature or precipitation.

12.3 Conclusion

The studies cited in this chapter generally confirm the conclusions of the first assessment of climate change in the Baltic Sea basin (BACC Author Team 2008). For areas presently characterised by spring floods due to snow melt, the floods are likely to occur earlier in the year and their magnitude is likely to decrease owing to less snowfall and a shorter snow accumulation period. As a consequence, sediment transport and the risk of inundation are likely to decrease. In the southern part of the Baltic Sea area, increasing winter precipitation is projected to result in increased river discharge during winter. In addition, groundwater recharge is projected to increase in areas where the infiltration capacity is not currently exceeded, resulting in higher groundwater levels. Decreasing precipitation combined with rising temperature and evapotranspiration during summer is projected to result in a drying of the root zone which would drive increasing irrigation demands in the southern part of the Baltic Sea area.

The issue of uncertainty in the model projections has received increasing attention. Many studies have been carried out using a range of emission scenarios, GCMs and RCMs. The results indicate that the choice of GCM is especially important for the projected changes in climate (see also Chap. 11). A single study, by Olsson et al. (2011), suggested that the initialisation of the GCM is also important, at least for near-future projections. This indicates that natural variability in the GCM has a large effect on the projected changes, especially if short time spans are considered, or if the time spans investigated are periods for which the effect of increasing greenhouse gas (GHG) emissions is low, such as the near-future or emission scenarios with low GHG emissions. Olsson et al. (2011) proposed that climate models that are able to reproduce historical variability and so provide forecasts (as opposed to projections) of the future climate should be developed. A few studies have evaluated the impact of how climate model results are transferred to the hydrological model, also referred to as bias correction. It is indicated (Beldring et al. 2008; Yang et al. 2010) that the commonly used delta change approach may not produce satisfactory results, for example when dealing with snowmelt or extreme discharges. However, the quantity of work undertaken on this topic is low and more research is needed to quantify the accuracy and uncertainty associated with other bias correction methods.

The effect of impact model uncertainty has only been investigated in one study (Kriaučiūnienė et al. 2009). Several uncertainties are associated with impact modelling, including parameter uncertainty and model structure uncertainty. The values of the parameters of a hydrological model are normally found through calibration against historical data and are always associated with uncertainty. This uncertainty will translate into uncertainty in the projected changes. Model structure may also affect the response to climate change. In most of the studies cited in this chapter, conceptual rainfall-run-off models have been used to quantify the impact of climate change on river discharge. However, different types of hydrological model including physically based and/or distributed models may respond differently. Hence, studies are needed that not only consider the impact of climate projection uncertainty but also consider the hydrological model uncertainty.

The many uncertainties involved in projections of climate change impacts pose a challenge in presenting the results to stakeholders and decision-makers. The approach of Wetterhall et al. (2011) where response surfaces for exceeding a predetermined threshold are generated as a result of changes in climate variables (e.g. temperature and precipitation) is interesting as it effectively visualises the impact of climate change. Although associated with limitations on the ability to capture all changes in magnitude and especially the dynamics of future climate, the method provides a useful screening tool for assessing the impact of climate change on key impact variables and thresholds.

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13 Projected Change—Marine Physics

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Abstract

This chapter assesses recent results of changes in water temperature, salinity, sea ice, storm surges and wind waves during the twenty-first century in scenario simulations for the Baltic Sea. There have been several improvements since the first Baltic Sea assessment of climate change: the number of relevant scenario simulations has increased, ensembles of transient simulations with improved models based upon the scenarios and global models of IPCC's Fourth Assessment Report (AR4) have been analysed, and changes in biogeochemical cycles are now considered. The scenario simulations project that water temperatures will increase in the future, with the greatest changes in the northern Baltic Sea during summer. In agreement with earlier studies, sea-ice cover is projected to decrease drastically. Salinity is projected to decrease due to increased river run-off, whereas the impact of wind changes on salinity is negligible because the latter is relatively small. However, uncertainty in salinity projections is large owing to considerable bias in the simulated water balance. According to one study, salt transport into the Baltic Sea is unchanged. Sea-level rise has greater potential to increase surge levels in the Baltic Sea than increased wind speed, and changes in wind waves are projected to be small.

Keywords

Baltic sea • Climate change • Assessment • Scenario simulations • Dynamical downscaling • Climate models

13.1 Introduction

Model-based projections of climate change for the marine parts of the Baltic Sea region are sparse compared to studies for the atmosphere. The first assessment of climate change in the Baltic Sea area (BACC Author Team 2008) concluded that 'the mean annual sea surface temperatures could increase by some 2–4 °C by the end of the twenty-first century. Ice extent in the sea would then decrease by some 50–80 %. The average salinity of the Baltic Sea is projected to decrease between 8 and 50 %. However, it should be noted that these oceanographic findings are based upon only four regional scenario simulations using two emissions scenarios and two global models'. At that time, only one study addressing uncertainties was available and this was based on a larger ensemble of 16 simulations to project Baltic Sea salinity by the end of this century (Meier et al. 2006c).

The reason why only a few Baltic Sea scenario simulations have been performed compared to scenario simulations for the atmosphere might be that the former are computationally demanding due to the high horizontal and vertical grid resolutions required. Furthermore, from the ocean perspective, the dynamical downscaling technique used to investigate changes in climate at regional scales requires coupled atmosphere–ice–ocean models (Döscher et al. 2002; Räisänen et al. 2004) instead of uncoupled regional climate model (RCM) simulations (Kjellström et al. 2011; Nikulin et al. 2011) because sea-surface temperature (SST) and sea-ice concentration fields from general circulation models

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_13

(GCMs) used as surface boundary conditions in RCMs might be biased due to the coarse resolution of GCMs. Such biases would affect model sensitivity and hence the results in climate projections (Meier et al. 2011d). To reduce computational demands, scenario simulations were performed for selected time slices only (e.g. Madsen 2009) which requires the application of the 'delta' or 'delta change' approach (e.g. Meier 2002).

Since the first assessment of climate change in the Baltic Sea basin (BACC Author Team 2008), the number of relevant scenario simulations has increased considerably. In particular, a large number of scenario simulations were carried out during 2009–2011 within the ECOSUPPORT project (advanced modelling tool for scenarios of the Baltic Sea ECOsystem to SUPPORT decision making, www.baltex-research.eu/ecosupport). The following items characterise the new simulations compared to the first BACC assessment:

- The horizontal resolution of atmosphere and ocean model components increased to typically less than 25 and 3.6 km, respectively (e.g. Meier et al. 2011b).
- New model versions of GCMs and RCMs were used.
- The results and assumptions of the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC 2007) were used instead of those from the IPCC's Third Assessment Report (IPCC 2001).
- Multi-model ensemble modelling was introduced to estimate uncertainties due to biases in the Baltic Sea models (e.g. Meier et al. 2011b).
- Instead of time slices often combined with the delta approach (e.g. Meier 2006), transient simulations (1960–2100) were performed (e.g. Neumann 2010).
- Coupled physical–biogeochemical models were used (Neumann 2010; Meier et al. 2011a, b, c, 2012a, b, c; Eilola et al. 2012; Neumann et al. 2012).

This chapter relies on literature concerning climate change scenario simulations for the Baltic Sea between 2007 and (March) 2012. The focus is on projected changes in water temperature, salinity, sea ice, storm surges, and wind waves. Changes in marine biogeochemical variables such as oxygen, nutrients, and phytoplankton (chlorophyll) concentrations are addressed in Chaps. 18 and 19. The graphics included in this chapter (with the exception of Fig. 13.6) show ensemble mean changes projected for 2069–2098 relative to a baseline/reference of 1978–2007 in transient scenario simulations calculated with three coupled physical–biogeochemical models for the Baltic Sea following Meier et al. (2011b, 2012c). The Baltic Sea models are forced with RCM data driven by two GCMs at the lateral boundaries and two greenhouse gas (GHG) emission scenarios (SRES A1B and A2, see Nakićenović et al. 2000). The atmospheric forcing fields of the RCM were analysed by Meier et al. (2011d). Results at six monitoring stations representing

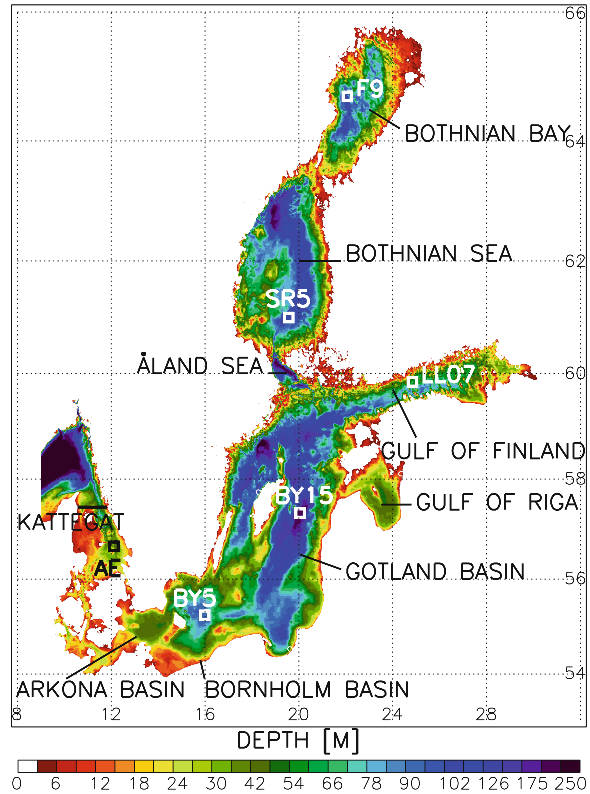

Fig. 13.1 Bottom topography in the Baltic Sea and selected Swedish and Finnish monitoring stations at Anholt East (AE), Bornholm Deep (BY5), Gotland Deep (BY15), LL07 in the Gulf of Finland, SR5 in the Bothnian Sea and F9 in the Bothnian Bay

different sub-basins of the Baltic Sea are reported in this chapter. The locations of the monitoring stations are depicted in Fig. 13.1.

13.2 Water Temperature

Projected change in seasonal and annual mean ensemble average SST is shown in Fig. 13.2. The projected change is greatest in the Bothnian Bay and Bothnian Sea during summer and in the Gulf of Finland during spring. According to the mini-ensemble, which is forced by the SRES A1B and A2 scenarios, summer SST could increase by about 2–4 °C in the southern and northern Baltic Sea, respectively. The north–south gradient in SST change is caused at least partly by the ice-albedo feedback, which increases SST sensitivity in the northern Baltic Sea (Meier et al. 2011d).

The projected change in annual mean temperature at the six monitoring stations (Fig. 13.1) is 2–3 °C (surface water layers) and 0–2 °C (deep-water layers) (Fig. 13.3). At Anholt East in the Kattegat, temperatures in the deep-water remain unchanged due to the assumption in the scenario simulations that the vertical temperature and salinity profiles at the lateral boundaries in the Kattegat or Skagerrak (depending on the

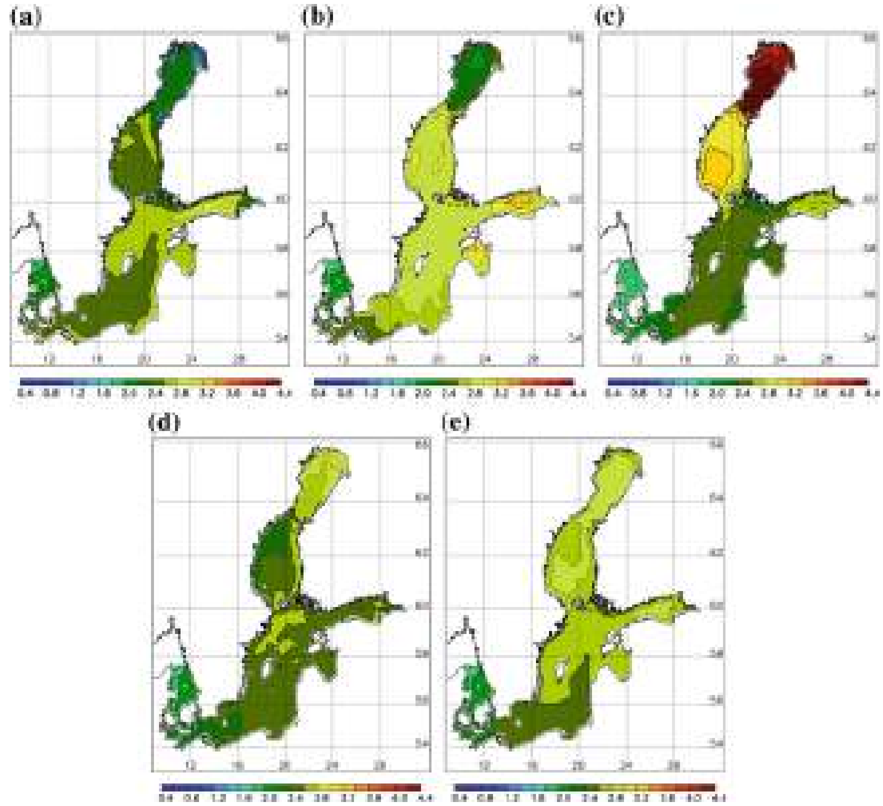

Fig. 13.2 Projected change in seasonal (**a** DJF, **b** MAM, **c** JJA, **d** SON) and **e** annual mean ensemble average SSTs for 2069–2098 relative to a baseline of 1978–2007. See Meier et al. (2012a)

ocean model used) do not change over time. Sensitivity experiments showed that artificially fixing the boundary data does not affect the Baltic Sea interior (Meier 2002). This is because in the model even during major Baltic inflow events the inflowing water volume is smaller than that of the Kattegat surface layer. Because at the time scale of inflows, no water from outside the model domain is advected through the Danish straits into the Baltic Sea and because at this time scale, the heat budget of the Kattegat surface layer is controlled by vertical heat fluxes, water masses that flow into the Baltic Sea are usually realistically simulated.

In all sub-basins, the surface layer is projected to warm more than the deep-water. Although this increasing temperature differential would cause an increase in vertical stratification, the projected salinity change is more significant in terms of water density than the projected temperature change. Hence, scenario simulations suggest that the vertical stratification between surface and deep-water layers would actually decrease (see Sect. 13.3).

The magnitude and spatial patterns of change in water temperature are relatively similar when different Baltic Sea models use the same atmospheric forcing (Meier et al. 2011a, b, 2012a, c). Even with atmospheric forcing from different RCMs or different GHG emission scenarios (SRES A2, B2, A1B, B1), temperature changes are similar (Meier 2006; Neumann 2010; Neumann and Friedland 2011; Gräwe et al. 2013) although temperature changes in the B1 scenario appear smaller than those in the other scenarios and a stabilising tendency at the end of the scenario simulation is observed (Neumann 2010; Neumann and Friedland 2011).

Most of the scenario simulations for the Baltic Sea were performed with regionally limited ocean models with lateral boundaries in the Kattegat or Skagerrak. Due to the proximity of the lateral boundary, results for the Kattegat are not reliable. Instead, projections for the North Sea should be employed to study changes in the Kattegat and Skagerrak (Ådlandsvik 2008; Holt et al. 2010). Only Madsen (2009) investigated both shallow seas—the Baltic Sea and the North Sea—simultaneously and found that warming is greater in the Baltic Sea than that in the North Sea.

Holt et al. (2010) found in the A1B scenario that the shelf sea regions warm substantially more than the open ocean, by 1.5–4 °C depending on location. These results agree with observed warming trends in the Baltic Sea, North Sea and other shelf seas (Belkin 2009).

13.3 Salinity

Results of the multi-model ensemble simulations by Meier et al. (2011b, 2012c) indicate that projected changes in sea-surface salinity are small in the northern and eastern Baltic Sea (smallest in the Bothnian Bay) and greatest in the Danish straits region, especially in the Belt Sea (Fig. 13.4).

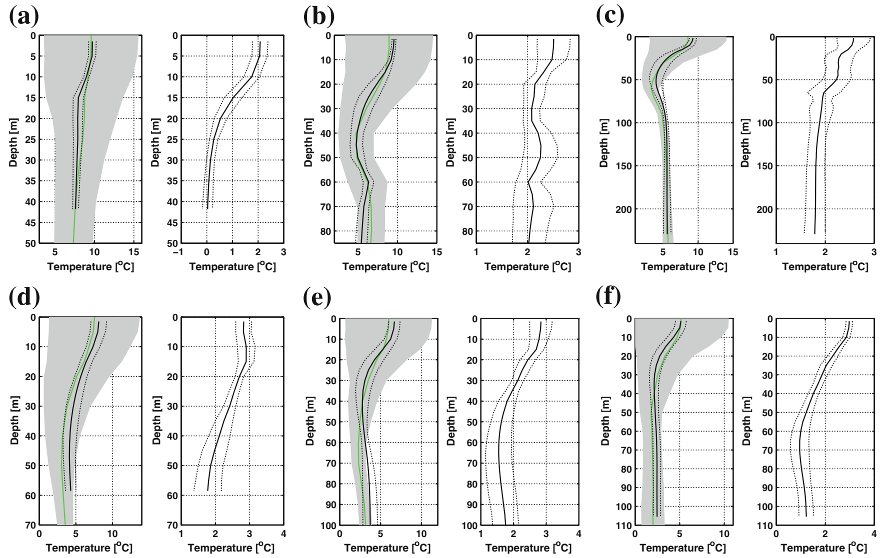

Fig. 13.3 Ensemble average vertical profiles and projected change in temperature for 2069–2098 relative to a baseline of 1978–2007 at the monitoring stations at **a** Anholt East (AE), **b** Bornholm Deep (BY5) and **c** Gotland Deep (BY15), and in the **d** Gulf of Finland (LL07), **e** Bothnian Sea (SR5) and **f** Bothnian Bay (F9) (for locations, see Fig. 13.1): observations (*green*), baseline 1978–2007 (*black*). The range in variability is indicated by the ± one standard deviation band around the ensemble average of model results (*dotted lines*) or observations (*grey shaded area*), see Meier et al. (2012b)

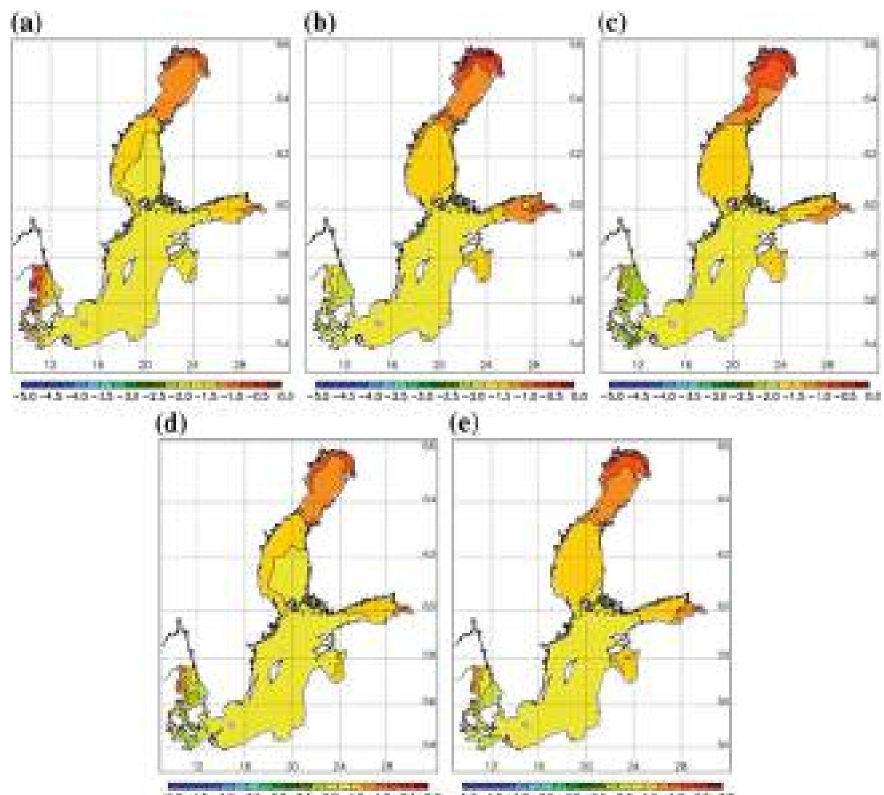

Fig. 13.4 Projected change in seasonal (**a** DJF, **b** MAM, **c** JJA, **d** SON) and **e** annual mean ensemble average sea-surface salinity for 2069–2098 relative to a baseline of 1978–2007. See Meier et al. (2012a)

The latter are due to a shift in the fronts within the transition zone. Seasonal change in sea-surface salinity is projected to be minimal.

The projected reductions in salinity in the Bornholm Basin and Gotland Basin are almost constant with depth and amount to 1.5–2 g kg^{-1} in the ensemble mean (Fig. 13.5). However, change in the deep-water is greater than that in the

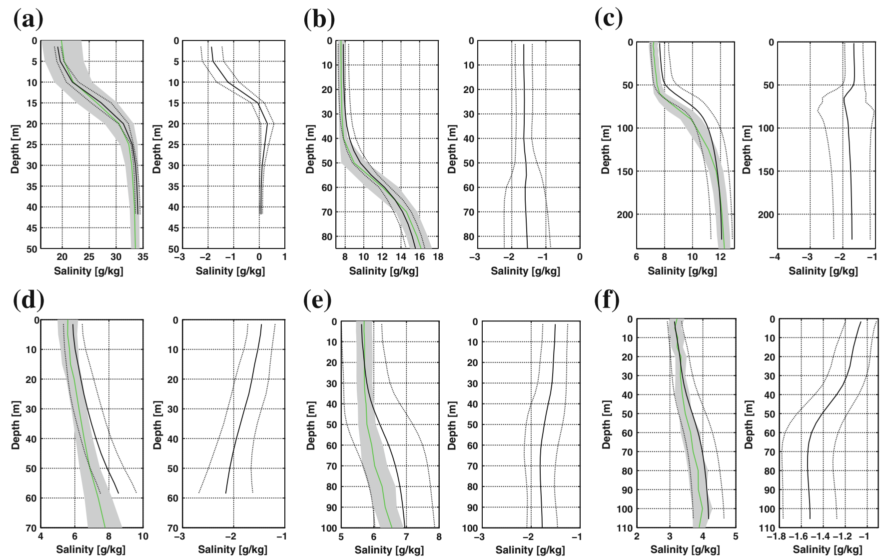

Fig. 13.5 Ensemble average vertical profiles and projected change in salinity for 2069–2098 relative to a baseline of 1978–2007 at the monitoring stations at **a** Anholt East (AE), **b** Bornholm Deep (BY5) and **c** Gotland Deep (BY15), and in the **d** Gulf of Finland (LL07), **e** Bothnian Sea (SR5) and **e** Bothnian Bay (F9) (for locations, see Fig. 13.1): observations (*green*), baseline 1978–2007 (*black*). The range in variability is indicated by the ± one standard deviation band around the ensemble average of model results (*dotted lines*) or observations (*grey shaded area*), see Meier et al. (2012b)

surface layer in these sub-basins. In more weakly stratified regions such as the Gulf of Finland or Bothnian Bay, the difference in the projected change for salinity in surface and bottom layers is even greater causing a reduction in vertical stability. These results are consistent across the three Baltic Sea models. The projections of halocline depth are most uncertain in the Gotland Basin, as indicated by the standard deviation among the ensemble members (Fig. 13.5).

In the ensemble presented, the salinity changes projected are driven by changes in run-off which is projected to increase by 15–22 % (Meier et al. 2012b). The latter figures are estimated from the difference between precipitation and evaporation over land calculated from the RCM output directly (Meier et al. 2012b). If a hydrological model is used to calculate run-off from the same atmospheric forcing, run-off changes are smaller and increase by 4–13 % (Arheimer et al. 2012). These discrepancies illustrate the uncertainty in hydrological modelling.

In the ensemble presented by Meier et al. (2012b), changes in wind speed play only a minor role in the salinity changes in contrast to earlier findings by Meier (2006). In the latter scenario simulations driven by ECHAM4/OPYC3, monthly mean increases in wind speed of almost 30 % were found during February (Räisänen et al. 2004). Hence, salinity projections remain uncertain in accordance with earlier results by Meier et al. (2006c) owing to the uncertainty in wind speed projections over the Baltic Sea region (Kjellström et al. 2011; Nikulin et al. 2011). However, there is generally a tendency for both increased mean (Kjellström et al. 2011) and extreme (Nikulin et al. 2011) wind speeds over the sea.

Although all studies based on dynamical modelling suggest that in a future climate, Baltic Sea salinity will decrease or remain unchanged compared to the present-day climate (Meier 2006; Meier et al. 2006c, 2011b, 2012b; Neumann 2010), Hansson et al. (2011) claimed that run-off from the total Baltic Sea catchment would decrease if air temperature rises. Hansson et al. (2011) reconstructed river run-off to the Baltic Sea for the period 1500–1995 using a statistical approach with air temperature and atmospheric circulation indices as predictors. They found that over the past 500 years, the total river run-off to the Baltic Sea showed no significant long-term trend but decreased slightly in response to the observed rise in temperature, at a rate of 3 % per 1 °C rise (see also Chap. 5, Fig. 5.3). However, their approach considerably underestimates interannual variability. In addition, changes in future climate that by the end of the century are projected to be outside the range of decadal variability are very likely not comparable with changes in past climate because past anthropogenic warming is small and other drivers may have controlled the air temperature–run-off relationship.

Studies of both past and future climates suggest that increased total run-off would increase the ventilation of the upper halocline due to weakened stratification causing improved oxygen conditions in the upper deep-water (e.g. Gustafsson and Omstedt 2009; Meier et al. 2011b). Despite changing halocline depth, stratification changes in the Baltic proper due to increased freshwater supply are expected to be minor (Meier 2005). Increased freshwater supply would drive an increased recirculation of brackish surface waters and consequently reduced saltwater fluxes into the Baltic Sea. For the hydrography of the north-western European continental shelf, Holt et al. (2010) found that the strength of seasonal stratification may increase by about 20 % on the shelf, compared with 20–50 % in the open ocean. The former being controlled by temperature and the latter by salinity.

Hordoir and Meier (2011) found increased spring and summer stratification under a warmer climate, with the changes greater in the northern Baltic Sea than the southern. The north–south gradient might be partly explained by the salinity-dependent temperature of maximum density. In the present climate, winter temperatures in the Baltic Sea are often below the temperature of maximum density such that warming during spring causes thermal convection (see also Chap. 7, Sect. 7.1). In a future climate, temperatures are expected to be typically higher than the temperature of maximum density and thermally induced stratification may start without prior thermal convection (Hordoir and Meier 2011). In particular, in the northern Baltic Sea, these stratification changes might affect vertical nutrient fluxes and thus the intensity of the spring bloom in future climate.

Gräwe and Burchard (2011, 2012) and Gräwe et al. (2013) studied local changes in the western Baltic Sea with a high-resolution model. They found no significant trend in potential energy measuring the competition between stratification and mixing. They also found no clear tendencies in the projected change in saltwater transport for either medium or major inflow events.

13.4 Sea Ice

Whether there is a reduction in sea-ice cover in the future depends mainly on the projected change in air temperature over the Baltic Sea in winter; other drivers such as wind are considered less important (e.g. Tinz 1996; Meier et al. 2004a, 2011d; Meier 2006; Jylhä et al. 2008; Neumann 2010). Hence, the projected changes in sea-ice cover depend on the GHG emission scenario, the GCM and the Baltic Sea model used. Both dynamical modelling and statistical modelling suggest that the relationship between annual maximum ice extent and winter mean air temperature changes is nonlinear (Meier et al. 2004a; Jylhä et al. 2008), that is, even under a warmer future climate, sea ice is likely in the northern Baltic Sea (Fig. 13.6).

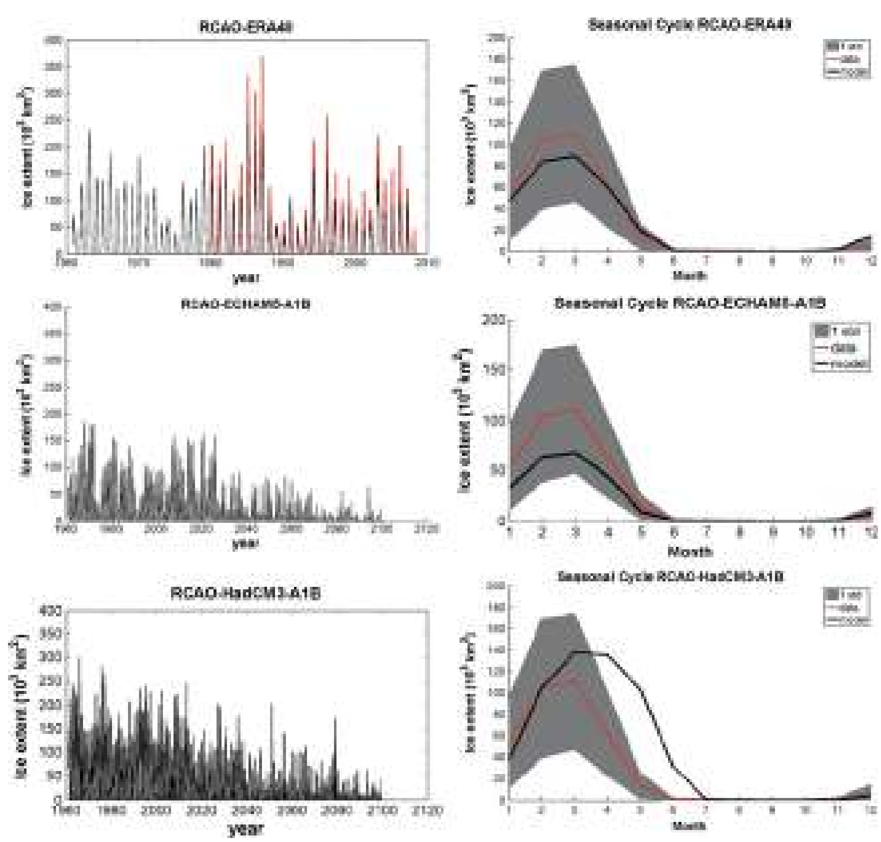

Fig. 13.6 Sea-ice extent as function of time for 1961–2007 and 1961–2100 in hindcast and scenario simulations, respectively (*left panels*): observations (*red*), model results (*black*). The mean seasonal cycles for 1980–2007 are also shown (*right panels*). The three rows show results from RCAO-ERA40, RCAO-ECHAM5-r3-A1B and RCAO-HadCM3-ref-A1B using a horizontal resolution of 50 km for the atmosphere model (Meier et al. 2011d)

For instance, Jylhä et al. (2008) examined the Baltic Sea ice cover using a statistical model that related annual maximum ice extent to wintertime coastal temperatures. They found that all model simulations, irrespective of the forcing scenario (SRES A2, B2 scenarios) and driving RCM (seven RCMs of the PRUDENCE project), produce considerably milder sea-ice conditions. However, they used only one driving GCM and concluded that a larger number of GCMs as drivers of the RCMs is likely to have resulted in wider ranges in sea-ice estimates than those in their study (see Fig. 13.6). This could explain why Jylhä et al. (2008) found more frequent unprecedentedly mild years for 2070–2100 than in the scenarios by Meier et al. (2004a). Nevertheless, all new scenario simulations indicate a strong decrease in sea-ice extent in agreement with earlier studies summarised in the first BACC assessment (BACC Author Team 2008).

Although new sea-ice models with explicitly resolved sea-ice categories for ridged and rafted ice (Haapala et al. 2005) have been applied in scenario simulations for the Baltic Sea, results for ice categories are not yet published.

13.5 Storm Surges

Past sea-level variability, trends and possible drivers are addressed in Chap. 9. Future changes in the mean sea level of the Baltic Sea caused by large-scale drivers such as steric expansion, geoid changes, melting of mountain glaciers and ice caps and changes in the mass balance of the Greenland and Antarctic ice sheets are discussed in Chap. 14. This chapter focuses on changes projected in regional sea level caused by changes in the regional wind field. A few scenario simulations for the Baltic Sea that consider other regional drivers such as changes in sea-ice cover and regional thermosteric and halosteric changes are also available (e.g. Madsen 2009; Hünicke 2010; Gräwe and Burchard 2012).

Tidal amplitude in the Baltic Sea is small. However, tidal amplitudes in the Skagerrak, particularly for the semi-diurnal constituents, are significant. On the European Shelf, future sea-level rise is projected to cause non-negligible increases and decreases in the amplitude of the principal lunar, semi-diurnal tidal constituent (M_2) depending on the assumed large-scale sea-level rise (Pickering et al. 2012; Müller et al. 2013).

For the Baltic Sea, results of the transient (1960–2100), multi-model ensemble simulations by Meier et al. (2011b, 2012c) indicate that at the end of the century, changes in mean sea-surface height are projected to be greatest during spring and up to 20 cm in coastal areas of Bothnian Bay (Fig. 13.7). The annual mean sea-surface height is projected to reach a maximum of 10 cm. The large spring signal is caused by one of the three Baltic Sea models and is related to the earlier melt of sea ice in a future climate. In model results by Meier et al. (2012a), only mean sea-level changes caused by regional wind changes and in one model even by steric effects are considered and neither the large-scale sea-level rise nor land uplift is included and has to be added to compile plausible future sea-level scenarios (Meier et al. 2004b). The changes projected in sea-level extremes (or storm surges) from the scenario simulations by Meier et al. (2011b) have not yet been analysed.

An evaluation of the earlier scenario simulations by Meier et al. (2004b) or Meier (2006) indicates that during the control period, both mean seasonal cycles and 100-year surge events in the Baltic Sea are simulated well relative to observations. However, extreme sea-level events in the western Baltic Sea and Danish straits were considerably underestimated, probably because the regional topography was not sufficiently resolved using a Baltic Sea model with a relatively coarse horizontal resolution of about 11 km only (Meier 2006). To address this shortcoming, Gräwe and Burchard (2012) used a high-resolution local model for the western Baltic Sea (with a spatial resolution of about 1 km), nested into a RCM in a dynamical downscaling approach, which also took into account baroclinic effects. They found that the quality of surge heights in their model was significantly better than in the driving model. Taking available global sea-level rise scenarios and simulated regional wind speed changes, Gräwe and Burchard (2012) found that sea-level rise has greater potential to increase surge levels in the Baltic Sea than does increased wind speed. Using the SRES A1B scenario and assuming a sea-level rise of 50 cm and a change in mean wind speed of approximately 4 %, resulted in projected surges with a 100-year return period for 2071–2100 that were greatest at the stations Lübeck, Koserow and Gedser in the western Baltic Sea and up to 2.7 m compared to simulated surge heights of less than 2.1 m for 1961–2000. However, the relative impact of changing wind speed on sea-level extremes might be greater for stations in the eastern Baltic Sea, for example in St Petersburg, depending on the driving GCM used (Meier 2006). Furthermore, Gräwe and Burchard (2012) found that changes in storm surge height in the scenarios can be consistently explained by the increase in mean sea level and variation in wind speed, supporting earlier approaches by Meier et al. (2004b) and Meier (2006).

In addition to the impact of changing winds on sea levels, Madsen (2009) analysed steric effects using a coupled North Sea and Baltic Sea model. The author found maximum halosteric changes in annual mean sea level of about 6.5 cm in the Baltic Sea. Hence, locally generated steric effects will be small compared to those of the NE Atlantic Ocean, basically because of the relatively small water depths. Changes in sea level from dynamical adjustment of global and regional changes might be greater but have not been studied in detail.

Hünicke and Zorita (2008) found an increase in sea-level amplitude (i.e. the difference between mean winter and spring sea level) and suggested that this could be due to the

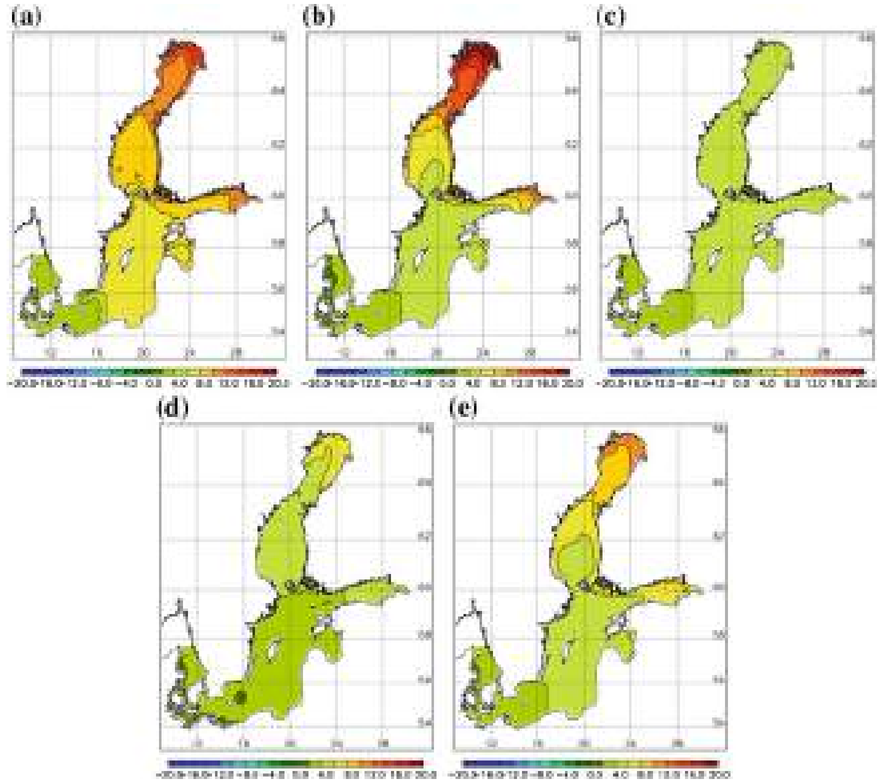

Fig. 13.7 Projected change in seasonal (**a** DJF, **b** MAM, **c** JJA, **d** SON) and **e** annual mean ensemble average sea-surface height for 2069–2098 relative to a baseline of 1978–2007, see Meier et al. (2012a)

long-term trend in seasonal precipitation. The results fit well with the projections by Hünicke (2010) who applied a statistical downscaling approach to the output of five GCMs. The author found that using both sea-level pressure and precipitation as the predictor resulted in significant future change in the sea level of Baltic Sea. Although Hünicke's estimates are lower, they are of the same order of magnitude as the projected rise in global sea level.

The implications of sea-level rise for various coasts have been analysed by several authors. For instance, impacts on the Estonian, Polish and Danish coasts were discussed by Kont et al. (2008), Pruszak and Zawadzka (2005, 2009) and Fenger et al. (2008), respectively. For further details, see Chap. 20.

13.6 Wind Waves

Since the first BACC assessment (BACC Author Team 2008), when scenario simulations with simplified wave models only were available (Meier et al. 2006a, b), advanced wave models have been applied for the Baltic Sea (e.g. Weisse and Günther 2007) and projections of the future wave climate have been made both at the Swedish Meteorological and Hydrological Institute (Kriezi and Broman 2008) and at the Helmholtz-Zentrum Geesthacht (Groll and Hünicke 2011).

In a recent study by Dreier et al. (2011), a statistical model based on wind–wave correlations and wind data derived from RCM simulations forced by ECHAM5/MPI-OM and two GHG emission scenarios (SRES A1B, B1) was used to perform scenario simulations of the future wave climate along the German Baltic Sea coast. The results indicate only a small increase in overall wave energy input in a future climate. However, the frequency of wave directions from the west could increase by 3.5 % compared to present-day conditions.

13.7 Conclusion

Recent studies confirm the findings of the first assessment of climate change in the Baltic Sea basin (BACC Author Team 2008), namely that under all GHG emission scenarios used (covering the range between SRES B1 and A2), water temperature is projected to increase significantly and sea-ice cover to decrease significantly. Warming would enhance the stability across the seasonal thermocline.

Although one study claimed an increase in salinity in the Baltic Sea under a warmer climate, most studies project decreased salinity and reduced stability across the permanent halocline due to higher, spatially integrated run-off from land. No clear tendencies in saltwater transport were found. However, the uncertainty in salinity projections is likely to be large due to biases in atmospheric and hydrological models.

Although wind speed is projected to increase over sea, especially over areas with diminishing ice cover, no significant trend was found in potential energy (measuring the competition between stratification and mixing). In accordance with earlier results, it was found that sea-level rise has greater potential to increase surge levels in the Baltic Sea than does increased wind speed.

In contrast to the first BACC assessment (BACC Author Team 2008), the findings reported in this chapter are based on multi-model ensemble scenario simulations using several GHG emissions scenarios and Baltic Sea models. However, it is very likely that estimates of uncertainty caused by biases in GCMs are still underestimated in most studies.

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Projected Change—Sea Level

14

Aslak Grinsted

Abstract

Global warming is causing sea levels to rise, primarily due to loss of land-based ice masses and thermal (steric) expansion of the world oceans. Sea level does not rise in a globally uniform manner, but varies in complex spatial patterns. This chapter reviews projections of the individual contributions to sea-level rise. These are used to assemble a mid-range scenario of a 0.70 ± 0.30-m sea-level rise over the twenty-first century (based on the SRES A1B scenario) and a high-end scenario of 1.10 m. The sea-level projection was regionalised to the Baltic Sea area by taking into account local dynamic sea-level rise and weighting the components of the sea-level budget by their static equilibrium fingerprint. This yields a mid-range Baltic Sea sea-level rise that is ~80 % of the global mean. Ongoing glacial isostatic adjustment (GIA) partly compensates for local sea-level rise in much of the region. For the mid-range scenario, this equates to a twenty-first century relative sea-level rise of 0.60 m near Hamburg and a relative sea-level fall of 0.35 m in the Bothnian Bay. The high-end scenario is characterised by an additional 0.5 m.

Keywords

Regional • Sea level rise • Climate change

14.1 Introduction

Global warming is causing sea levels to rise, primarily due to loss of land-based ice masses and thermal (steric) expansion of the world oceans (Meehl et al. 2007). The rise in global mean sea level (GMSL) is projected to accelerate over the twenty-first century (Meehl et al. 2007; Rahmstorf 2007a; Grinsted et al. 2010; Jevrejeva et al. 2010, 2012b). Sea level does not rise in a globally uniform manner, but has been observed to vary in complex spatial patterns (e.g. Douglas 2001; Bindoff et al. 2007). The projected changes in regional relative sea level (RSL, see definition of key terms in Chap. 9, Box 9.1) will deviate markedly from the global mean for a variety of reasons. In the Baltic Sea region, there is a large ongoing land uplift caused by glacial isostatic adjustment(GIA) due to the loss of the Fennoscandian ice sheet at the end of the last glacial period. In the Bothnian Bay, the RSL changes due to this adjustment are of the order of 1 m per century (Hill et al. 2010). The dynamic sea surface topography response to climate change will be far from uniform, and similarly, mass loss from ice sheets will not distribute evenly across the world oceans (Mitrovica et al. 2001; see Chap. 9). A practical and common approach to projecting regional sea-level rise is to project the individual major contributions to GMSL rise and combine this budget with the corresponding spatial fingerprints of each contributor (e.g. Slangen et al. 2011; Perrette et al. 2013; Spada et al. 2013). Published regional sea-level rise projections have generally focused on the global scale and have had insufficient detail to resolve the Baltic Sea. It has therefore been necessary to construct new sea-level projections specifically for this chapter using published fingerprints combined with a review of the projected sea-level budget.

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_14

There will be changes in sea-ice cover, salinity and atmospheric forcing of the Baltic Sea (see Chap. 13). These changes in local boundary conditions may influence decadal sea-level variability (e.g. Hünicke and Zorita 2006) and the statistics of extreme storm surges and wave heights (Chap. 13), but may not be major contributors to century-scale changes in mean sea level: some studies show that stronger winds and increased run-off may contribute in the order of 5-cm local sea level (LSL) rise for some locations in the Baltic Sea (Hünicke 2010; Meier et al. 2004, 2006, 2011). This contribution is not considered in this chapter as this effect is included separately in models of changing storm surge statistics (Chap. 13).

14.2 Sea-level Budget

There are different approaches to making sea-level rise projections. The traditional approach has been to build models of the individual major contributors and from these project the evolving budget (e.g. Meehl et al. 2007). As an alternative, semi-empirical models have been constructed where the rate of global sea-level rise is statistically related to global temperature (Rahmstorf 2007a; Horton et al. 2008; Vermeer and Rahmstorf 2009; Grinsted et al. 2010) or radiative forcing (Jevrejeva et al. 2009, 2010, 2012b; Moore et al. 2010). Semi-empirical models generally project greater twenty-first century sea-level rise than the budget from process-based models. For example, the semi-empirical model by Grinsted et al. (2010) resulted in projections of twenty-first century sea-level rise that are about a factor of three greater than those from the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4). The gap is partly explained by dynamical ice sheet discharge, which was not modelled by AR4 generation ice sheet models. Since then, there has been a large ongoing effort to improve the projections of land-based ice loss. The improvements have increased the process-based projections since AR4 (e.g. Spada et al. 2013), and the gap between process-based and semi-empirical projections has narrowed in recent years to the point where confidence intervals overlap. There is, however, still considerable uncertainty in the evolution of the sea-level budget. There is disagreement over the level of confidence that should be assigned to current semi-empirical versus process-based projections (Holgate et al. 2007; Rahmstorf 2007a, b, 2010; Schmith et al. 2007; von Storch et al. 2008; Vermeer and Rahmstorf 2009, 2010; Taboada and Anadón 2010; Rahmstorf et al. 2012b). Process-based models have not been able to fully account for the rate of twentieth-century sea-level rise (Rahmstorf 2007a, 2010; Rahmstorf et al. 2012a), except when combinations of process models with exceptionally large twentieth-century contributions are picked (Gregory et al. 2013). Furthermore, there is still a high degree of structural uncertainty in models of ice sheet loss. For example, modelling of dynamical ice sheet discharge and ice–ocean interaction is relatively immature, and projections of dynamical ice loss are often not driven by a specific emission scenario (e.g. Bindschadler et al. 2012). This uncertainty is also reflected in the large spread of estimates given in an expert elicitation of ice sheet experts (Bamber and Aspinall 2013). For these reasons, there is a lack of confidence in process-based model projections of sea-level rise. On the other hand, the statistical methods used to calibrate some semi-empirical models have been criticised in the literature (Holgate et al. 2007; Rahmstorf 2007b; Schmith et al. 2007; Taboada and Anadón 2010; Vermeer and Rahmstorf 2010; Rahmstorf et al. 2012b; Grassi et al. 2013). In addition, the physical justification for the semi-empirical model formulations has been questioned (von Storch et al. 2008; Vermeer and Rahmstorf 2010; Jevrejeva et al. 2012a). Ice sheet mass loss has a nonlinear equilibrium response (e.g. Levermann et al. 2013), which cannot be captured by current semi-empirical models. However, recent modelling studies have found that ice sheet volume response on century timescales is remarkably linear in imposed forcing (Bindschadler et al. 2012; Winkelmann and Levermann 2012) which suggests that semi-empirical models can approximate ice sheet mass loss over a few centuries.

For regional sea-level projections, it is necessary to know the detailed budget as the different contributors will have very different spatial 'fingerprints' in the Baltic Sea. Semi-empirical models of GMSL rise do not inform on the partitioning between contributors and cannot be used directly in regional sea-level rise projections. Thus, this chapter reviews process-based estimates and combines these with published regional fingerprints to construct mid-range and high-end scenarios of the projected rise in regional sea level. This exclusive focus on process-based estimates over semi-empirical models should be kept in mind when assessing the likelihood of the resulting regional sea-level rise scenarios.

The mid-range regional sea-level rise scenario is based on an assessment of projections using the SRES A1B scenario for 2090–2099 with respect to 1980–1999. On century timescales, the major sea-level contributors (land ice loss and thermosteric expansion) will respond to the integrated climate forcing history. A practical approach to estimating the LSL response for scenarios other than A1B can be estimated by scaling the A1B projections with the relative steric response between the two scenarios (see Table 14.1). This makes it possible to estimate LSL for the new generation of representative concentration pathways (RCPs) in the CMIP5 model inter-comparison project from the A1B projections presented in this chapter. The steric scaling ratio of

Table 14.1 The 5–95 % ranges in the projected steric contribution to global mean sea-level rise by 2090–2099 relative to 1980–1999 (Meehl et al. 2007; Yin 2012)

Greenhouse gas scenario	Steric sea-level rise (m)	Percentage contribution to A1B rise
B1	0.10–0.24	76
B2	0.12–0.28	89
A1B	0.13–0.32	100
A1T	0.12–0.30	93
A2	0.14–0.35	109
A1FI	0.17–0.41	129
RCP2.6	0.09–0.22	69
RCP4.5	0.13–0.27	90
RCP6.0	0.14–0.29	95
RCP8.5	0.20–0.40	135

The projected sea-level rise for each scenario is also expressed as a percentage relative to the sea-level rise projected under the A1B scenario (see Sect. 14.2)

133 % between the RCP8.5 and SRES A1B scenarios is in close agreement with Nick et al. (2013) for the Greenland ice sheet (GrIS) dynamical response. The rate of glacier wastage will depend on the surface area exposed, and the glaciers will thus have a muted response to warming as the global glaciated area shrinks over time. This results in an important nonlinear response which this simple steric scaling does not account for. Applying the simple scaling coefficients in Table 14.1 to the Marzeion et al. (2012) glacier mass loss projections results in a 3 cm positive bias for RCP8.5. The error introduced by this convenient scaling approach will be smaller than the uncertainties in the A1B LSL projections themselves.

14.3 Steric Expansion

Increasing greenhouse gas (GHG) concentrations are causing a radiative imbalance of the Earth, which will cause the Earth to heat until thermal radiation once more balances incoming solar radiation. The majority of the extra heat retained due to the energy imbalance imposed by the emission scenario selected will be absorbed by the ocean, and as a consequence, the volume of the oceans would on average increase due to thermosteric expansion (Table 14.1). Steric expansion will be greatest in the open ocean where the water column is deepest (Landerer et al. 2007; Yin et al. 2010). This would lead to a differential increase in the steric sea surface heights (SSH), which would drive a redistribution of ocean mass from the ocean interior to shallower regions (Landerer et al. 2007; Yin et al. 2010). Changes in ocean circulation and in the hydrological cycle will also induce thermosteric and halosteric changes. The combined sea surface topography response is a complex spatial pattern of sea-level rise. The sea surface topography has been simulated by the ensemble of models in the CMIP3 archive (Meehl et al. 2007; Yin et al. 2010; Slangen et al. 2011), but the Baltic Sea region is very poorly resolved by the majority of the models. The dynamic sea surface topography has therefore been separated into a global average steric response, and the deviation from the mean referred to as the dynamic sea level (DSL) (following Landerer et al. 2007; Meehl et al. 2007; Yin et al. 2009, 2010) depicted in Fig. 14.1. The full CMIP3 ensemble of projections will be used for the A1B global steric contribution, but for the DSL contribution in the Baltic Sea region, the calculation is restricted to the mean of three models (GFDL CM2.1; MIROC 3.2 HiRES; MPI ECHAM5) (see Fig. 14.1). The

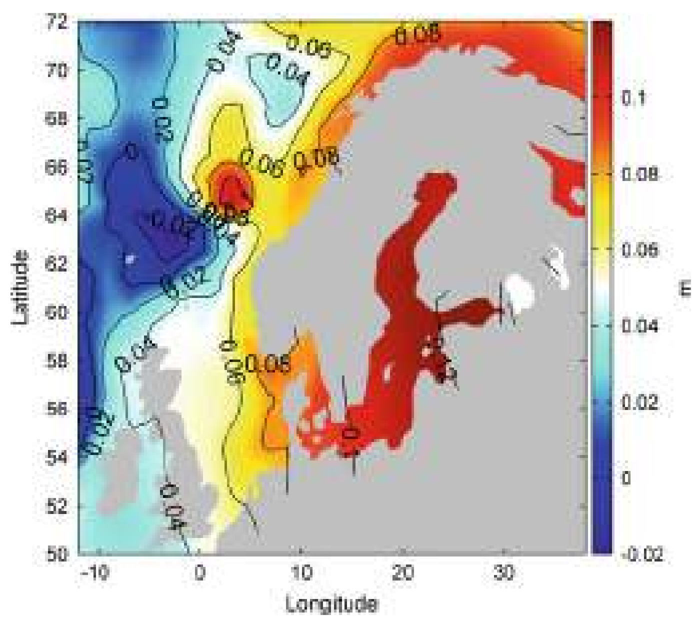

Fig. 14.1 Spatial pattern of the dynamic sea level (DSL) response (m) projected for the SRES A1B scenario for 2095 relative to the 1990 baseline. The DSL response is calculated as the average response of GFDL CM2.1, MIROC 3.2 HiRES, MPI ECHAM5 (Meehl et al. 2007). The three models are from the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (provided by J. Gregory) and selected on the basis that they should resolve the Baltic Sea

uncertainty in the DSL response near the Baltic Sea is estimated to be ±20 cm by considering the spread in the ensemble (Yin et al. 2010; Pardaens et al. 2011). This uncertainty is greater than the projected DSL changes in the region.

Freshwater fluxes arising from a negative GrIS mass balance have been shown to perturb the North Atlantic circulation and thus induce changes in SSH (Stammer 2008; Hu et al. 2009, 2011; Stammer et al. 2011). The model runs in the CMIP3 archive have not been forced with the additional freshwater flux coming as a consequence of decay of land-based ice masses. Hu et al. (2011) modelled the combined effect of forcing a model with both the SRES A1B scenario and a freshwater hosing from the GrIS. Using an unrealistically large hosing flux equivalent to 60 cm of global sea-level rise results in an additional global average steric contribution of about 2 cm by the end of the twenty-first century and that the hosing has little detectable influence on DSL near Scandinavia and in the Baltic Sea. Stammer et al. (2011), however, found that atmospheric feedbacks increase the hosing response significantly in the North Atlantic.

14.4 Geoid Changes

Large masses of land ice such as that contained in the GrIS gravitationally attract the oceans around them. This gravitational pull will be reduced if the ice mass is reduced through a negative mass balance. Reducing the ice load will also cause an immediate elastic rebound of the solid Earth, as well as perturb the Earth's rotation. These effects will combine to form a new static equilibrium in the sea-level configuration (Mitrovica et al. 2001, 2009; Bamber and Riva 2010; Kopp et al. 2010). The net static equilibrium sea-level response is that the sea-level contribution from melting land ice will not be distributed evenly across the Earth but will be characterised by a spatial fingerprint. Sea level will even drop up to ~2000 km away from the source (Mitrovica et al. 2001). As a consequence, the Baltic Sea region will only feel a small fraction of the global average sea-level contribution from the GrIS mass loss, but a slightly greater than average response from the Antarctic Ice Sheet (see Fig. 14.2). The steric response of the oceans leads to a mass redistribution towards shelf areas. This also induces self-attraction and shelf loading effects that lead to an additional minor LSL rise

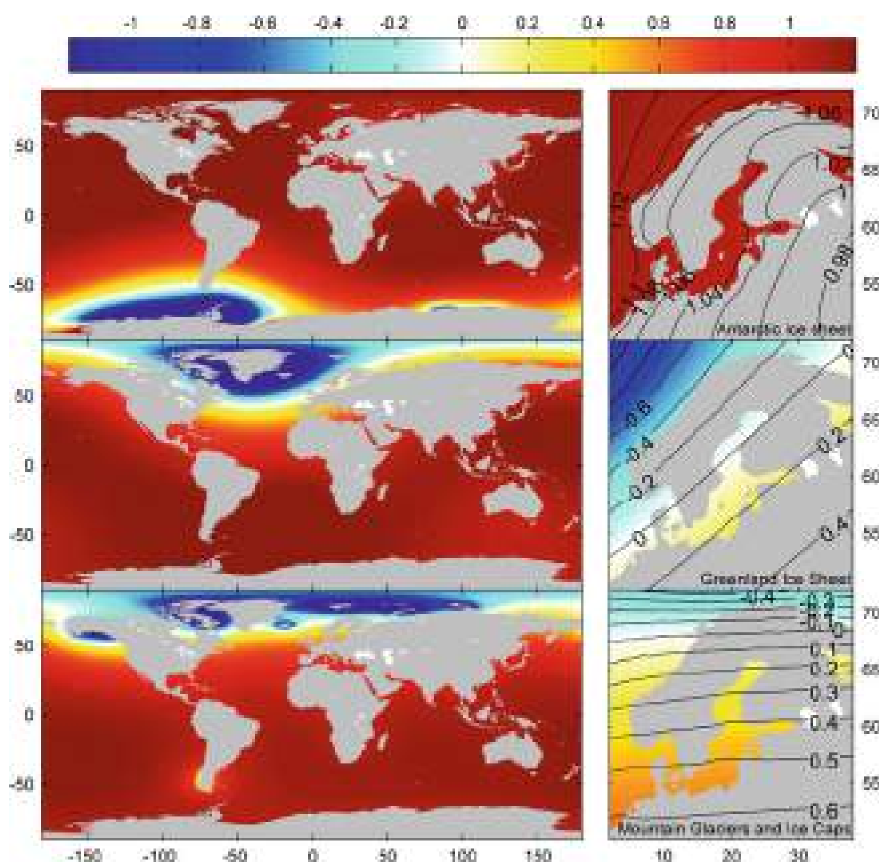

Fig. 14.2 The spatial fingerprint of sea-level rise expressed as a ratio to the global mean sea level equivalent loss from **a** the Antarctic ice sheet (Bamber and Riva 2010); **b** the Greenland ice sheet (Bamber and Riva 2010); **c** mountain glaciers and ice caps (Slangen et al. 2011)

of roughly 8 % of the steric and DSL contributions (Richter et al. 2013).

It is important to use realistic estimates of the patterns of ice loss when calculating the spatial fingerprints (Mitrovica et al. 2011). This chapter uses the present-day spatial patterns of ice loss for the two large ice sheets (Bamber and Riva 2010) and the projected pattern of mass loss from mountain glaciers and ice caps (Slangen et al. 2011). The total static equilibrium response is then calculated by scaling these fingerprints with the projected mass loss of land ice from these three sources: the GrIS, the Antarctic Ice Sheet, and mountain glaciers and ice caps.

14.5 Mountain Glaciers and Ice Caps

Glacier inventories are incomplete, but there are an estimated 300,000–400,000 glaciers and small ice caps in the world (Dyurgerov and Meier 2005). Detailed observations of these glaciers are sparse, and this leads to substantial uncertainty in their present-day volume and present-day rates of mass loss. The uncertainty in total volume (Grinsted 2013) will propagate to the projections of the contribution from mountain glaciers and ice caps (Slangen and van de Wal 2011). Some projections of the glacier contribution to global sea-level rise are based on a semi-empirical approach, where mass loss is related to global temperature change (e.g. Meehl et al. 2007). Marzeion et al. (2012) modelled the global glacier response of the globally complete Randolph Glacier Inventory (Arendt et al. 2012) to the CMIP5 ensemble of general circulation models (GCMs). Bahr et al. (2009) estimated how far present accumulation area ratios (AARs) are from being in equilibrium, and from that estimate a lower bound of glacier mass loss. The AAR-derived lower bounds are substantially greater than other projections. The contributions have been summarised in Table 14.2.

14.6 Greenland Ice Sheet

The surface mass balance (SMB) of the GrIS has been projected to contribute to global sea-level rise as SMB becomes increasingly negative. Several studies have modelled the GrIS SMB response to the projected climates in the CMIP3 simulations. These studies in general project a greater range than that reported in the IPCC AR4 (Graversen et al. 2011; Yoshimori and Abe-Ouchi 2012) and also project a generally greater mass loss. Regional climate projections of GrIS SMB are typically not coupled to ice sheet wastage and thus do not include the elevation mass balance feedback. This feedback is small over a single century, but it has been estimated that including it would increase the SMB contribution by ∼4 % over the twenty-first century (Edwards et al. 2013). Models of dynamical ice loss are lacking, and no models include a prognostic model of all key dynamical processes such as calving, grounding line migration and the impact of changing basal hydrology on ice dynamics. Observations suggest that dynamic ice loss scales with the SMB for the GrIS (Rignot et al. 2008). Dynamic mass loss projections are mostly based on heuristic estimates and statistical extrapolations of present-day trends and accelerations, which are not directly related to any GHG emission scenario (see Table 14.3). To conclude, there are large uncertainties in projections of GrIS mass loss. This uncertainty is much reduced for sea-level projections in the Baltic Sea due to the spatial fingerprint of GrIS mass loss.

Table 14.2 Review of projected contributions of mountain glaciers and ice caps to global mean sea-level rise for 2090–2099 relative to a baseline of 1990–1999

Source	Contribution (m sea-level equivalent)	Present-day volume (m sea-level equivalent)	Method
Radić and Hock (2011)	0.08–0.16		Surface mass balance model (A1B)
Radić and Hock (2010)		0.50–0.65	
Meehl et al. (2007)	0.08–0.15	0.15–0.72	Semi-empirical (A1B)
Meier et al. (2007)	0.10–0.24		Statistical extrapolation
Pfeffer et al. (2008)	0.17–0.55		Statistical extrapolation + heuristic dynamic
Bahr et al. (2009)	0.18–0.38	0.86	Accumulation area ratio conservative estimate of equilibrium (may not be reached by 2100)
Slangen et al. (2011)	0.10–0.20	0.50	A1B uncertainty study
Marzeion et al. (2012)	0.13–0.22		RCP6.0
Katsman et al. (2008)	0.07–0.19	0.10–0.40	A1B
Katsman et al. (2011)	0.07–0.20	0.15–0.60	High-end estimate

Table 14.3 Estimated contributions from Greenland ice sheet mass loss to global mean sea-level rise for 2090–2099 relative to a baseline of 1990–1999

Source	Contribution (m)	Method
Surface mass balance		
Meehl et al. (2007)	0.0–0.08	Positive degree day (A1B)
Fettweis et al. (2008)	0.03–0.05	Temperature index from the energy balance model (A1B)
Fettweis et al. (2013)	0.03–0.10	Regional climate model
Mernild et al. (2010)	0.12	Energy balance model (A1B)
Yoshimori and Abe-Ouchi (2012)	0.03–0.17	Positive degree day, systematic examination of uncertainties (A1B)
Franco et al. (2011)	0.05	GCM output mapped through a regression of Greenland climate anomalies on RCM output (A1B)
van Angelen et al. (2013)	0.07	RCM + snow model (RCP4.5)
Dynamical		
Pfeffer et al. (2008)	0.09–0.47	Heuristic
Price et al. (2011)	>0.06	Estimate of committed dynamical loss
Graversen et al. (2011)	0.03	Stationary tuned outlet-glacier sliding (shallow ice)
Rignot et al. (2011)	0.07	Statistical extrapolation of acceleration
Katsman et al. (2011)	0.10	Heuristic
Nick et al. (2013)	0.04–0.09	RCP4.5 scaled-up response from four major outlets
Total		
Katsman et al. (2008)	0.02–0.17	Semi-empirical
Meier et al. (2007)	0.05–0.25	Extrapolation
Graversen et al. (2011)	0.00–0.17	Shallow ice, tuned sliding in outlet glaciers (A1B)
Seddik et al. (2012)	0.10–0.17	Full stokes, PDD, range from: tuned sliding and doubled sliding
Bamber and Aspinall (2013)	0.08–0.31	Expert elicitation. Converted to total loss assuming constant acceleration from the present day (Shepherd et al. 2012)

14.7 Antarctic Ice Sheet

The Antarctic Ice Sheet SMB is projected to increase in a warmer climate. The increased moisture-holding capacity of warmer air would bring increased precipitation to the continent. This effect is modelled to dominate over increased ablation in the marginal areas of the ice sheet, and SMB modelling projects a negative contribution to sea-level rise. SMB models are typically not coupled to ice dynamics and so do not account for any mass balance induced increases in ice discharge which can offset 15–35 % of the mass gained from increased snowfall (Winkelmann et al. 2012). The dynamic mass loss of the Antarctic Ice Sheet is, as for the GrIS, primarily estimated using heuristic approaches and statistical extrapolation. Pollard and De Conto (2009) employed a novel approach to simulate grounding line mass flux over the past 5 million years and found the West Antarctic Ice Sheet to be particularly sensitive to warming with a 3 m contribution to sea-level rise for 2 °C warming. Estimates of current Antarctic Ice Sheet mass balance are negative and dominated by mass loss in regions of the West Antarctic Ice Sheet which are thought to be most dynamically sensitive (Rignot et al. 2011). Results are summarised in Table 14.4.

14.8 Glacial Isostatic Adjustment

During the last glacial period, ice sheets 1 km thick covered Fennoscandia, the British Isles and North America. The loss of these ice sheets is causing an ongoing viscoelastic response of the Earth. In particular, the unloading of the Fennoscandian ice sheet causes a present-day isostatic uplift of the order of ~1 m per century in the Bothnian Bay (e.g. Hill et al. 2010). Local land uplift will be perceived at the coast as a drop in RSL, and for the Baltic Sea region, the magnitude of this effect in many places exceeds present-day LSL rise. There are many methods for estimating the present-day GIA. The GIA can be modelled given a global ice sheet history and viscoelastic Earth structure. The present-day GIA can be detected in land movement (by levelling and GPS), in RSL trends (paleoshorelines, tide gauges) and in gravity data (by absolute gravimetry and satellite data from the GRACE mission).

Table 14.4 Estimated contributions from the Antarctic ice sheet mass loss to global mean sea-level rise for 2090–2099 relative to a baseline of 1990–1999

Source	Contribution (m)	Method
Surface mass balance		
Meehl et al. (2007)	−0.12 to −0.02	Positive degree day
Genthon et al. (2009)	−0.10	PDD CMIP3 corrections
Krinner et al. (2007)	−0.12	GCM
Bengtsson et al. (2011)	−0.04	High-res GCM
Ligtenberg et al. (2013)	−0.05 to −0.03	Regional climate model
Dynamical		
Pollard and de Conto (2009)	0.33	~3 m/2 °C/1000 years (WAIS volume)
Pfeffer et al. (2008)	0.12–0.55	Heuristic
Katsman et al. (2011)	0.07–0.49	Statistical extrapolation/ heuristic max
Rignot et al. (2011)	0.07	Statistical extrapolation
Spada et al. (2013)	0.12–0.38	Mid-range/high-end estimate
Total		
Rignot et al. (2011)	0.36	Statistical extrapolation
Katsman et al. (2011)	−0.01 to 0.41	Heuristic (moderate-severe)
Katsman et al. (2008)	−0.02 to 0.14	Semi-empirical
Meier et al. (2007)	0.05–0.06	Statistical extrapolation
Pfeffer et al. (2008)	0.14–0.62	Heuristic
Spada et al. (2013)	0.05–0.30	Mid-range/high-end estimate

This assessment uses the GIA estimate of Hill et al. (2010), which is designed to optimally reproduce present-day instrumental observations in the Baltic Sea region.

14.9 The Compiled Budget

This section constructs an overall sea-level budget based on the projections for the major individual contributors to global sea-level rise detailed in the previous sections. One potential issue with this approach is that all the components must be completely independent to avoid double accounting. In particular, the small glaciers near the ice sheet margin may in some studies be counted as part of the ice sheet contribution. Groundwater mining (Wada et al. 2010; Konikow 2011) and reservoir construction (Chao et al. 2008) add an additional non-climatic contribution to sea level. Reservoir building is currently slowing, while groundwater mining is expected to increase with the rise in world population (Shiklomanov and Rodda 2003), and it is expected that these sources will contribute to a net sea-level rise in the future (Rahmstorf et al. 2012b; Wada et al. 2012).

Table 14.5 Mid-range and high-end estimates of the sea-level rise budget for the twenty-first century

	Mid-range estimate (m) A1B	High-end estimate (m)
Global sea level contributor		
Greenland ice sheet (SMB)	0.07 ± 0.07	0.12
Greenland ice sheet (DYN)	0.07 ± 0.07	0.12
Antarctic ice sheet (SMB)	−0.07 ± 0.05	−0.05
Antarctic ice sheet (DYN)	0.21 ± 0.21	0.40
Mountain glaciers and ice caps	0.15 ± 0.05	0.18
Global mean steric	0.22 ± 0.09	0.25
Groundwater mining and reservoir storage	0.05 ± 0.07	0.08
Total GMSL	0.70 ± 0.3	1.10
Semi-empirical GMSL models	0.96 ± 0.4	
Local sea level contributor		
Baltic Sea hosing response	Excluded ± 0.10	0.20
Baltic Sea fingerprint error (Geoid and DSL)	Excluded ± 0.30	
Regional self-attraction and loading	Excluded ± 0.03	
Baltic Sea inverse barometer	Excluded ± 0.02	
GIA uncertainty (Hill et al. 2010)	Excluded ± 0.10	

The upper part of the table shows contributors to global mean sea level (GMSL), and the lower part lists terms that do not contribute to GMSL but which can have a significant impact locally in the Baltic Sea

The local inverse barometer contribution has only a minor impact on RSL rise in the Baltic Sea region (Stammer and Huttemann 2008) and is excluded. Based on the individual estimates given in Tables 14.1, 14.2, 14.3 and 14.4 and taking into account the uncertainties in the budget, a GMSL rise of 0.70 ± 0.3 m is projected under the SRES A1B scenario, for the period 1990–2095 (see Table 14.5). This tally of contributions overlaps with the range of published semi-empirical models for GMSL which project 0.96 ± 0.4 m rise (Rahmstorf 2007a; Horton et al. 2008; Grinsted et al. 2010; Jevrejeva et al. 2010; Rahmstorf et al. 2012b). The use of semi-empirical models has been discussed in the literature (Holgate et al. 2007; Rahmstorf 2007a, b; Schmith et al. 2007; von Storch et al. 2008; Vermeer and Rahmstorf 2009, 2010; Taboada and Anadón 2010; Rahmstorf et al. 2012b), but it is not understood why semi-empirical models consistently give greater rates of sea-level rise than the tally-based approach. This chapter relies on the tally-based approach.

The mid-range estimates in the budget are combined with their respective spatial fingerprints to provide LSL projections for the Baltic Sea region. The change in LSL is combined with the local GIA to yield local projections of RSL rise (Fig. 14.3). Locally, there may be additional sources of

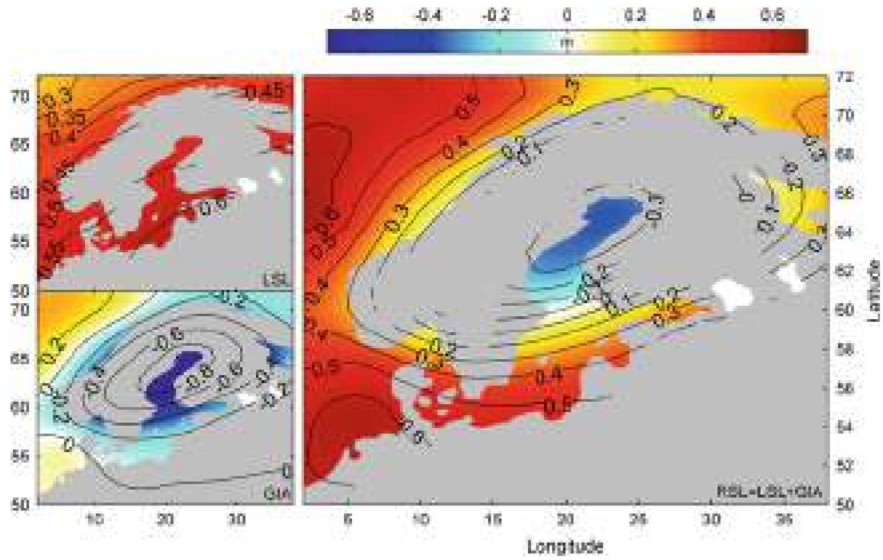

Fig. 14.3 *Right panel* shows the projected regional sea-level rise for 2090–2099 relative to the 1990–1999 baseline under the SRES A1B scenario, decomposed into local sea-level rise (*upper left*) and glacial isostatic adjustment (*lower left*; Hill et al. 2010). There may be additional local sources of vertical land movement that should be considered in adaptation

uplift/subsidence that should be taken into account in infrastructure planning. For example, the Frederikshavn tide gauge shows evidence of sinking due to gas leakage from an underground reservoir. The dominant uncertainty in the GMSL budget is the dynamic contribution of the Antarctic Ice Sheet. Furthermore, the uncertainty in RSL projections will be greater than for GMSL projections because of uncertainty in the spatial fingerprints and in the spatial distribution of change in DSL.

The projected sea-level budget is poorly constrained, and for some infrastructure decisions, a high-end scenario may be warranted (e.g. Lowe et al. 2009; Katsman et al. 2011; Spada et al. 2013). Therefore, a heuristic high-end scenario (Fig. 14.4) is constructed from high estimates of the projected budget, incorporating the possibility of an error in the spatial fingerprints in the Baltic Sea. The high-end heuristic is targeted at the combined 95th percentile based on the estimated uncertainty ranges in Table 14.5 while allowing for a more intense forcing scenario than the SRES A1B scenario (see Table 14.1). Of particular importance in the Baltic Sea is the long-tailed uncertainty of the Antarctic Ice Sheet, dynamic contribution for which the high-end estimate of 40 cm used is similar to that of Spada et al. (2013). However, this should not be interpreted in terms of a strict confidence bound given how the uncertainty ranges have been derived. At present, this high-end estimate is not considered likely from a process model perspective; however, there are several other lines of evidence that point to even greater sea-level rise being plausible. The total ice sheet contribution adopted here is 25 cm lower than the very likely upper range derived from an expert elicitation (Bamber and Aspinall 2013). Furthermore, the high-end scenario is more

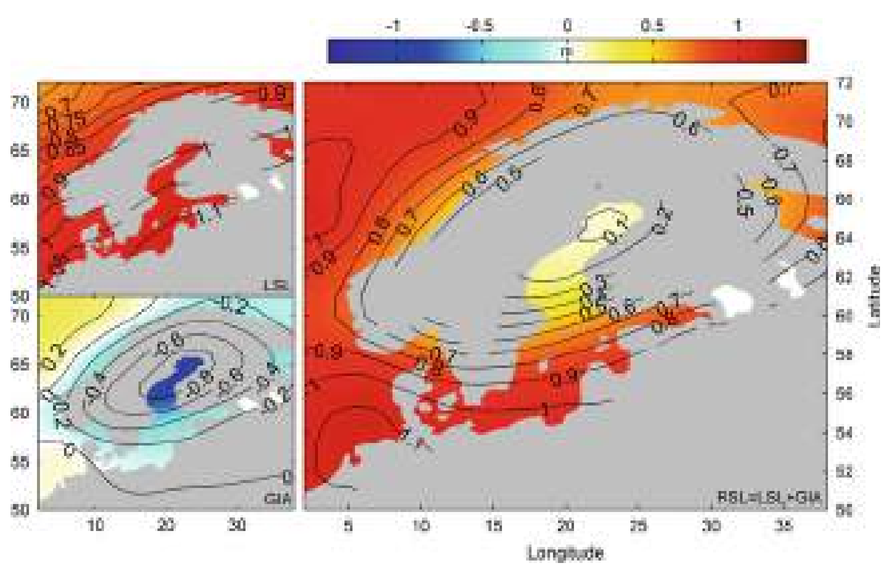

Fig. 14.4 A high-end estimate of projected sea-level rise in the Baltic Sea. *Right panel* shows the projected regional sea-level rise for 2090–2099 relative to the 1990–1999 baseline, decomposed into local sea-level rise (*upper left*) and glacial isostatic adjustment (*lower left*; Hill et al. 2010)

in line with central semi-empirical projections than the mid-range estimate (see Table 14.5; and Perrette et al. 2013). The individual contributions are expected to co-vary with global mean warming and thus climate sensitivity. Allowing for some uncertainty, covariance will further increase the likelihood of the high-end scenario.

14.10 Conclusion

By reviewing recent projections of the individual major contributions to global mean sea-level rise, it has been possible to assemble estimates of global mean sea-level rise over the twenty-first century: a mid-range scenario of 0.70 ± 0.30 m (based on the SRES A1B scenario) and a high-end scenario of 1.10 m (Table 14.5). This sea-level projection was regionalised by taking into account local dynamic sea-level rise (Fig. 14.1) and weighting individual components of the sea-level budget by their static equilibrium fingerprint (Fig. 14.2). This reveals a local sea-level projection that is ~80 % of the global mean for the mid-range scenario (Fig. 14.3). Ongoing GIA partly compensates for local sea-level rise in much of the Baltic Sea region. For the mid-range scenario, this equates to a twenty-first century relative sea-level rise of 0.60 m near Hamburg and a relative sea-level fall of 0.35 m in the Bothnian Bay (Fig. 14.3). The high-end scenario is characterised by an additional 0.5 m (Fig. 14.4). The dominant sources of uncertainty in sea-level projections for the Baltic Sea are the future rate of mass loss from the Antarctic Ice Sheet, uncertainties in the spatial fingerprints of each contributor to GMSL rise and the uncertainty in the spatial expression of DSL (Table 14.5). To better constrain sea-level projections, it is necessary to validate models of the most uncertain contributors to GMSL rise and LSL rise against their observed contributions in the coming decades.

Several studies have investigated the impact of LSL rise scenarios on the Baltic Sea coastline. These scenarios have usually been based on global sea-level models (Johansson et al. 2004; Staudt et al. 2004; Meier et al. 2006; BACC Author Team 2008) or adopted idealised sea-level rise scenarios such as 30 cm per century or 1 m per century (Kont et al. 2008; Pruszak and Zawadzka 2009). The LSL rise scenarios in these studies did not consider the spatial fingerprinting of land ice loss and so miss important regional effects. Nonetheless, the sea-level rise scenarios considered in these studies are still relevant as they fall inside the uncertainty envelope (Table 14.5).

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Part IV

Environmental Impacts of Climate Change

Introduction

The six chapters of this section describe the environmental impacts of climate change on the coastal and marine environments of the Baltic Sea basin; specifically, impacts on the atmosphere, biogeochemistry, ecosystems and coastal processes of the catchment and the Baltic Sea itself. Chapter 15 demonstrates that atmospheric chemistry is currently affected more by changes in human emissions than by direct climate change effects. Chapter 16 addresses the special vulnerability of coastal ecosystems to climate change, with a particular emphasis on birds and natural forests. Chapter 17 represents an in-depth look at changes in the aquatic biogeochemistry of the catchment area, showing that it is still very difficult to separate climate change impacts from the effects of other human-induced drivers. Chapters 18 and 19 deal, respectively, with the biogeochemistry and ecosystems of the Baltic Sea itself. The authors show that changes are clear but that it is difficult to identify any single factor, such as warming, as being responsible for the changes observed. Finally, Chap. 20 documents the clearly visible changes in the coastline over the past two decades.

The chapters demonstrate that most of the observed environmental changes are due to several interrelated factors of which climate change is but one. There is a need for attribution research to establish the respective impacts and interactions of the different anthropogenic pressures, such as climate change, eutrophication, and changes in land use.

Environmental Impacts—Atmospheric Chemistry

15

David Simpson, Jerzy Bartnicki, Jukka-Pekka Jalkanen, Hans-Christen Hansson, Ole Hertel, Joakim Langner, and Sara C. Pryor

Abstract

This chapter addresses sources and trends of atmospheric pollutants and deposition in relation to the Baltic Sea region. Air pollution is shown to have important effects, including significant contributions to nitrogen loading of the Baltic Sea area, ecosystem impacts due to acidifying and eutrophying pollutants and ozone, and human health impacts. Compounds such as sulphate and ozone also have climate impacts. Emission changes have been very significant over the past 100 years, although very different for land- and sea-based sources. Land-based emissions generally peaked around 1980–1990 and have since reduced due to emissions control measures. Emissions from shipping have been steadily increasing for decades, but recent measures have reduced sulphur and particulate emissions. Future developments depend strongly on policy developments. Changes in concentration and deposition of the acidifying components generally follow emission changes within the European area. Mean ozone levels roughly doubled during the twentieth century across the northern hemisphere, but peak levels have reduced in many regions in the past 20 years. The main changes in air pollution in the Baltic Sea region are due to changes in emissions rather than to climate change.

15.1 Introduction

In relation to the Baltic Sea region, this chapter attempts to answer the overriding questions: What are the main atmospheric pollutants and where do they originate? How has deposition changed over time and how good are the estimates? How will climate change affect future trends?

The main pollutants addressed in this chapter are the acidifying compounds (sulphur and nitrogen oxides, as well as ammonia/ammonium) and ozone. Historically, much of the international activity concerning air pollution in Europe stemmed from the recognition that increasing emissions and deposition of sulphur in Europe after the Second World War led to the acidification of freshwaters and sensitive soils in northern Europe and parts of the Baltic Sea region (Odén 1968). Following a coordinated research effort by the Organisation for Economic Co-operation and Development (OECD 1977), transboundary fluxes of acidifying pollutants were proven to be significant sources of acid deposition across Europe (Eliassen 1978; Tørseth et al. 2012, and references therein). The Convention on Long-range Transboundary Air Pollution (CLRTAP) was established in 1979. The network of monitoring sites, and the modelling framework, established for the OECD project was later continued under the European Monitoring and Evaluation Programme (EMEP), which was established to provide scientific support

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_15

to the CLRTAP contracting parties. Sulphur deposition has decreased significantly since the beginning of the 1980s due to the successful implementation of emission reductions in Europe under the protocols of CLRTAP (Mylona 1996; Vestreng et al. 2007).

Emissions and deposition of reactive nitrogen (Nr) species, such as nitrogen dioxide (NO_2), nitric acid (HNO_3), ammonia (NH_3) and ammonium (NH_4^+), have proven harder to reduce. Indeed, deposition of Nr species is now of particular concern for the Baltic Sea and surrounding semi-natural ecosystems as here the atmospheric supply of nitrogen can form an appreciable part of the total nitrogen load; Bartnicki et al. (2011) calculated that about one-quarter to one-third of Nr input to the Baltic Sea originates from airborne nitrogen deposited directly onto the sea surface. In addition, part of the nitrogen deposition into the Baltic Sea drainage basin reaches the sea via runoff from land (Seitzinger et al. 2002).

Measurements are essential for understanding the state of the atmosphere. However, the chain of processes linking emissions, atmospheric dispersion, chemical transformation and loss from the atmosphere of polluting Nr compounds is extremely complex, and observations can typically address only a small part of this chain. The following sections make use of both measurements and modelling results, discussing in turn emissions, concentrations and deposition. The ecosystem effects of air pollution are largely outside the scope of this chapter (but see Sutton et al. 2011 for a comprehensive discussion). This chapter concludes with a short discussion on the human health effects of air pollutants—probably the main driver of EU policy with regard to air quality.

15.2 Emissions

15.2.1 Land-Based Sources

Land-based emissions of the main acidifying compounds can be grouped into sulphur compounds and Nr species, where the predominant Nr groups are NH_3 and NO_X (=NO + NO_2) but also include other inorganic and organic nitrogen compounds (Hertel et al. 2012). Anthropogenic sulphur emissions decreased by more than 80 % from 1980 to 2004 in the HELCOM countries and by more than 70 % over the same period for the whole of Europe (Table 15.1; Vestreng et al. 2007). Between 1980 and 1990, the relative reduction in emissions was about the same in the whole of Europe as in the HELCOM countries, while after 1990, the reduction was greater in the HELCOM countries. Energy production from fossil fuel combustion has been the most important source of sulphur emissions. The reduction in emissions has mainly been achieved through a combination of reduced sulphur content of fuels, switching fuels and flue-gas desulphurisation in coal-fired power plants. The downturn in the economy in eastern Europe after 1989 was also an important factor.

Table 15.2 presents national NO_X emission trends for the HELCOM countries and for Europe as a whole, and gives an overview of the emissions in the EMEP inventory between 1980 and 2005. The relative share of emissions from road transport is also shown. It should be noted that official EMEP emissions had to be supplemented by data from other sources for many countries. These data sources included data from the GAINS model (www.gains.iiasa.ac.at/gains) developed at the International Institute for Applied Systems Analysis

Table 15.1 Trends in sulphur emission (kt SO_2 $year^{-1}$) for the HELCOM Contracting Parties and for Europe as a whole, 1980–2004

Country	1980	1985	1990	1995	2000	2004
Denmark	450	333	176	133	27	23
Estonia	287	254	274	117	96	90
Finland	584	382	259	95	74	83
Germany	7514	7732	5289	1708	630	559
Latvia	96	97	97	47	10	4
Lithuania	311	304	263	92	43	40
Poland	4100	4300	3278	2381	1507	1286
Russian Federation	7323	6350	6113	3101	2263	1858
Sweden	491	266	117	79	52	47
Total Europe	55,340	48,448	42,896	26,282	18,263	15,162

Vestreng et al. (2007)

Table 15.2 Trends in nitrogen oxide emissions (kt NO_X as NO_2 year^{-1}) for the HELCOM Contracting Parties and for Europe as a whole, 1980–2005

Country	1980	1985	1990	1995	2000	2005
Denmark	273 (26)	291 (32)	274 (38)	264 (37)	207 (39)	186 (37)
Estonia	67 (43)	74 (41)	74 (41)	38 (42)	35 (38)	32 (34)
Finland	295 (36)	275 (44)	299 (53)	258 (51)	235 (45)	177 (32)
Germany	3334 (35)	3276 (38)	2861 (47)	2170 (53)	1817 (55)	1443 (45)
Latvia	61 (43)	67 (41)	67 (30)	40 (37)	38 (42)	41 (43)
Lithuania	152 (36)	166 (34)	158 (34)	65 (36)	49 (51)	58 (58)
Poland	1229 (38)	1500 (26)	1581 (25)	1121 (28)	838 (27)	811 (28)
Russian Federation	3280 (37)	3600 (33)	3600 (31)	2563 (36)	2357 (40)	2795 (43)
Sweden	404 (44)	426 (41)	314 (55)	280 (54)	231 (49)	205 (41)
Total Europe	23,944 (36)	24,550 (36)	25,256 (38)	20,507 (41)	17,809 (42)	17,059 (39)

Percentage contribution from road transport in brackets. Vestreng et al. (2009)

(IIASA) and for a few countries from EDGAR emission data (www.mnp.nl/edgar). The coverage of officially reported emissions is about 40 % in the 1980s, increasing to nearly 60 % after 1990. The level of confidence is considered to be higher for the reported and reviewed emission data, due to country-specific insight and the detailed input to the calculations. The emissions and their uncertainties are discussed further by Vestreng et al. (2009).

As also noted by Vestreng et al. (2009), a trend study by Konovalov et al. (2008) applying inversion techniques with satellite measurements between 1996 and 2004, broadly confirmed that the NO_X emission trends in Europe have been decreasing, and further indicated that the quality of the EMEP inventory has increased over the past few years. This study also suggested that the greatest uncertainties in emission inventories probably concern southern and eastern Europe.

Long-term trends in emissions are the result of two main factors: changes in fuel use and changes in emission factors. Older technology tends to be associated with higher emissions due, for example, to inefficiency or lack of control measures. Figure 15.1 shows the development of fuel use and NO_X emissions from 1880 to 2005, as calculated by Vestreng et al. (2009). Dramatic changes are seen after 1945, when liquid fuel use and road transport emissions increase significantly. Fuel use peaks around 1980 and emissions peak around 1990. The reduction in emissions in later years is stronger than the reduction in fossil fuel use and reflects the increasing use of improved emission control technologies, particularly with respect to road vehicles.

To illustrate the uncertainty associated with such inventories, Vestreng et al. (2009) compared estimates of long-term NO_X emission changes with three other major inventory efforts. In recent years, all studies gave similar estimates for both western and eastern Europe. For the earliest years (around 1920), the Vestreng et al. (2009) and van Aardenne et al. (2001) values were also similar, although this may reflect a lack of alternative data rather than inherent accuracy. However, between 1950 and 1990, significant differences were seen in western Europe and more so in eastern Europe. Lack of information on emission factors for older vehicles and combustion appliances is a major limitation in

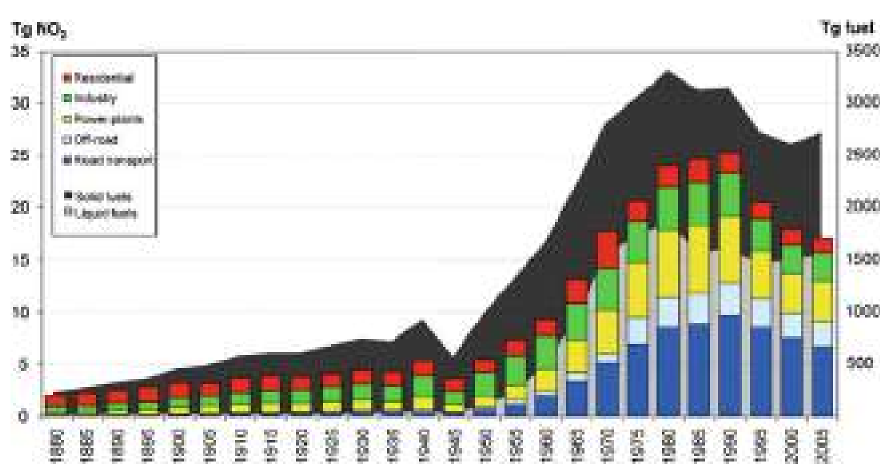

Fig. 15.1 Sector trends in European NO_X emissions (million t year^{-1}, as NO_2) and European solid and liquid fossil fuel consumption (million t year^{-1}) since 1880. Vestreng et al. (2009)

Table 15.3 Annual emissions of carbon monoxide (CO), non-methane volatile organic compounds (NMVOC), nitrogen oxides (NO_X), sulphur oxides (SO_X), fine particulate matter (less than 2.5 µm in diameter; $PM_{2.5}$) and coarse particulate matter (between 2.5 and 10 µm in diameter; PM_C) from Baltic Sea shipping as reported to EMEP

Year	CO (kt) (C)	NMVOC (kt)	NO_X (kt) (as NO_2)	SO_X (kt) (as SO_2)	$PM_{2.5}$ (kt)	PM_C (kt)
1980	29.0	8.0	1157	456	na	na
1985	29.0	8.0	1157	456	na	na
1990	24.2	8.2	775	336	na	na
1995	27.4	9.2	881	382	na	na
2000	31.1	10.5	907	376	21.9	1.20
2005	35.2	11.8	996	426	24.8	1.40
2009	38.8	13.1	1074	244	16.6	0.92

EMEP CEIP (www.ceip.at/ceip/). na: no data available

estimating such emissions. (Emission trends for sulphur are usually simpler to make than for Nr compounds, since uncontrolled emissions depend in a fairly straightforward way on the sulphur content of fuel, which is reasonably well documented.)

Ammonia emissions are dominated by agricultural sources, and in Central to northern Europe, these contribute 85–98 % of atmospheric emissions (www.emep.int). In the HELCOM countries, Denmark, Poland and Germany achieved considerable emission reductions over the period 1985–2010, while Sweden and Norway achieved marginal reductions only. The reasons for this difference are partly structural changes and increased efficiency in agricultural production methods as seen in Denmark (Skjøth et al. 2008) but also political restructuring affecting industry and energy production in eastern Europe as seen in Poland (Reis et al. 2009). One cause of uncertainty in the quantification of NH_3 emissions is the incomplete data sets—for example due to the lack of data on natural emissions (e.g. Reis et al. 2009) and limited access to high-quality emission factors. Uncertainty associated with the national totals may be down to 5–10 % in the most data-rich regions such as Denmark (Geels et al. 2012a), but the lack of data and poor emission factors may cause uncertainties of up to a factor of ten in data-poor regions (e.g. Nowak et al. 2012). Another cause of uncertainty is the use of fixed emission factors in the calculation of national emission inventories for NH_3, a practise that is used despite the fact that the emission process depends on meteorology. This point is addressed in Sect. 15.2.3.

15.2.2 Shipping

The Baltic Sea is a very busy shipping route with over 2000 vessels sailing at any given time. Until recently, emissions from shipping were based on national statistics only, with assumed activity statistics and emission factors. However, in the past few years, ship emission estimates have become more consistent through the inclusion of real ship activity data offered by the Automatic Identification System (AIS).

15.2.2.1 Historical Perspective

Compared to estimates of current ship emissions in the Baltic Sea (see Sect. 15.2.2.2), the accuracy of past and future emission estimates is much lower. Nevertheless, it is important to evaluate general trends in emissions over time in the Baltic Sea (see Table 15.3).

The reported shipping emissions of carbon monoxide (CO) and non-methane volatile organic compounds (NMVOC) in 2009 are higher than in 1980 (by 34 and 64 %, respectively), whereas NO_X and sulphur oxide (SO_X) emissions in 2009 are lower than in 1980 (7 and 46 %, respectively). Annual emissions of both particulate matter (PM) components are more than 20 % lower in 2009 than in 2000.

For all pollutants, annual emissions in 1985 are the same as in 1980. There is a clear increasing tendency for emissions of CO, NMVOC and NO_X in the period 1990–2009. Emissions of SO_X also show increasing trends until 2005 and then a significant drop between 2005 and 2009. Some of the differences between the consecutive values (e.g. 1985–1990) in Table 15.3 are hard to understand and suggest problems in the reporting of the inventories.

Estimates of ship emissions in the Baltic Sea before 1980 are difficult to make. However, historical data concerning global ship emissions are available based either on energy statistics (e.g. Endresen et al. 2003) or international shipping statistics (e.g. Corbett and Koehler 2003). Eyring et al. (2005b) used an activity-based approach to calculate total fuel consumption and global annual ship emissions back to 1950. Based on these data, Fig. 15.2 shows the relative global annual ship emissions of NO_X, CO, sulphur dioxide (SO_2) and PM, expressed as a fraction of the 2001 values. Emissions of all these pollutants rose significantly between 1960 and 1980, due to a rapid increase in the merchant fleet: the number of ships doubled in the period 1960–1980. Furthermore, the tanker business reached its peak in 1973–1975, and the

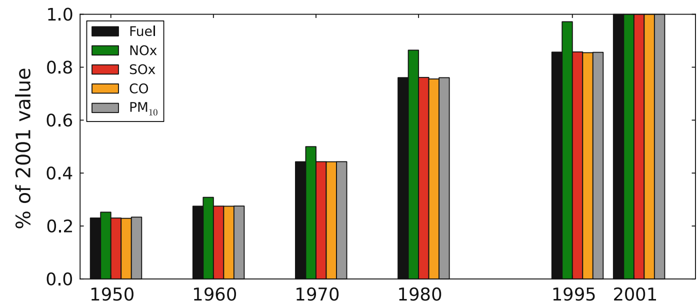

Fig. 15.2 Development of global shipping fuel consumption and emissions, relative to 2001. Based on Eyring et al. (2005b)

container vessel class was introduced (Eyring et al. 2005b). Assuming that ship traffic on the Baltic Sea developed in a similar manner to the global trends, these data would indicate that annual ship emissions on the Baltic Sea in 1950 were at least 60 % lower than ship emissions in 2000, for all pollutants considered. It should be noted, however, that applying global inventories to the Baltic Sea leads to an underestimate of ship emissions, mainly due to a lack of accounting for 'short-sea' traffic (i.e. the movement of cargo and passengers on routes in enclosed sea areas or near coastlines without crossing an ocean) (Jalkanen et al. 2009).

15.2.2.2 Recent Developments

Emission estimates for shipping have become more consistent in the past few years with the inclusion of real ship activity data offered by AIS. This device automatically reports the identity, location and speed of any vessel. The use of AIS data in ship emission modelling was demonstrated in the Baltic Sea area, as part of the so-called Ship Traffic Emissions Assessment Model (STEAM, STEAM2, Jalkanen et al. 2009, 2012). The amount of data provided by AIS is immense, for example the vessel movement data for the Baltic Sea over five years (2006–2010) comprises over 1100 million position reports. Using AIS data also removes the need to use average speeds or estimated travel distance between ports. AIS became obligatory for all large ships in 2005, which is the earliest year when such emission studies are possible.

Significant uncertainties remain, especially on the sulphur content of ship fuel for which compliance with the SO_X Emission Control Area (SECA) requirements of the International Maritime Organisation is assumed (IMO 1998). A recent study indicates, however, that this assumption is in reasonable agreement with experimental measurements (Berg et al. 2011). Currently, the sulphur content of marine fuels is restricted to 1 % (w/w) in SECA areas of the Baltic Sea, North Sea, English Channel and North America. In addition, the EU sulphur directive commits all vessels to using 0.1 % fuel inside EU port areas. This helps to mitigate the harmful emissions of PM from ships in areas that are close to human populations (e.g. ports).

Emissions from Baltic Sea shipping usually peak during summer months, because of increased passenger traffic at this time. During these months, there is a significant number of small craft (Fig. 15.3). Commercial marine traffic also peaks during summer months, but this effect is less pronounced than for small craft.

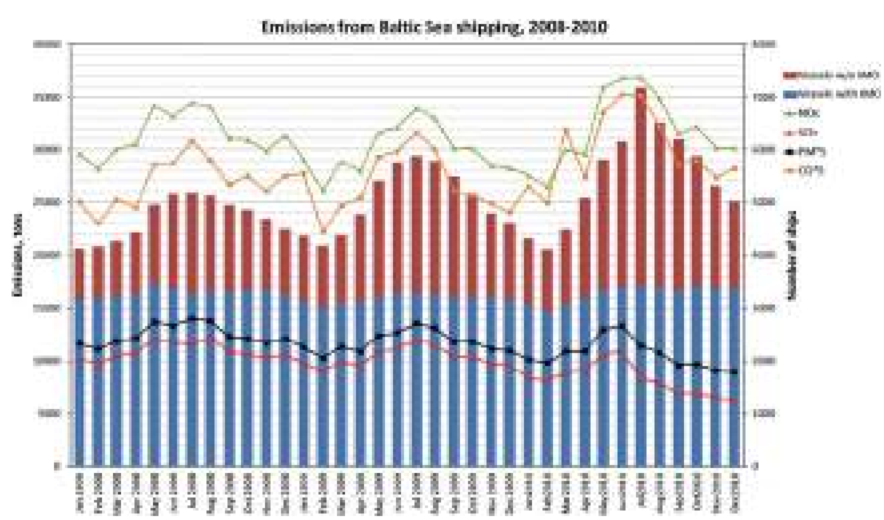

Fig. 15.3 Emissions from Baltic Sea shipping in 2008–2010. *Blue bars* indicate the number of large vessels (with an IMO registry number), and *red bars* illustrate the number of small vessels (without an IMO registry number). *Lines* represent pollutant emissions each month. Note that, emissions of particulates (PM) and carbon monoxide (CO) have been multiplied by five. STEAM2 model (Jalkanen et al. 2012)

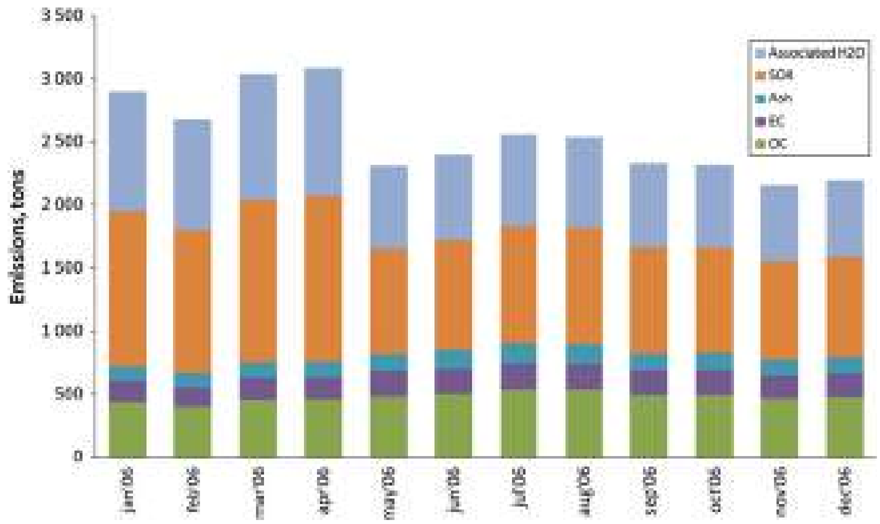

Fig. 15.4 Emissions of particulate matter from Baltic Sea shipping through 2006. The reduction in emissions from May 2006 was due to the required reduction in the sulphur content of marine fuels in the SO_X Emission Control Area (SECA). The pre-SECA sulphur content of 2.4 % was reduced to 1.5 %, but the emissions of PM components unrelated to fuel sulphur remain unaffected. STEAM2 model (Jalkanen et al. 2012)

In May 2006, the Baltic Sea became the first SECA and the North Sea/English Channel followed in November 2007. The sulphur content of marine fuels was reduced to 1.5 % from a global average of 2.4 %, reducing both gaseous and particulate sulphur emissions from shipping (Fig. 15.4). A further reduction in fuel sulphur from 1.5 to 1.0 % and the requirements of the EU sulphur directive in port areas decreased the SO_2 emissions by 20 % during 2010 (compared to an uncontrolled emission level). Particulate matter emissions from ships are not entirely associated with sulphur emissions, but nonetheless a decrease of 9 % was observed in PM emissions as a result of sulphur reductions in 2006 (Fig. 15.4).

Emissions of sulphate aerosols and associated water were also reduced by the fuel sulphur requirements, whereas the emissions of elemental carbon, organic carbon and ash increased because these components are not dependent on fuel sulphur content of marine fuels and reflect the increased amount of fuel used by Baltic Sea shipping. Sulphate aerosol emissions from ships in 2010 were almost half (47 %) those in 2009.

Emissions from the shipping sector are also affected by economic factors. The global recession of the late 2000s reduced emissions from Baltic Sea shipping by 0.5–5 % depending on the pollutant. Passenger traffic was relatively unaffected by the recession, whereas traffic by bulk, vehicle and container cargo carriers declined. However, by 2010, emission levels had already recovered relative to those before 2008.

The EMEP ship emission inventories for NO_X and SO_X (309 and 190 kt in 2006) show reasonable agreement with STEAM inventories (336 and 144 kt in 2006), but the PM and CO emissions are significantly larger in STEAM than EMEP. The methods used in the construction of these two emission inventories are very different but both include uncertainties arising from the fuel sulphur content. In addition, EMEP inventories are not based on real vessel movements and do not include the contribution from ships in port areas. The use of AIS improves geographical accuracy thus enabling local-scale studies in port areas.

At least two factors affect the reported emission trends for Baltic Sea shipping: the increase in ship traffic, which contributes to the increased number of vessels in the area, and the strong increase in small vessels with AIS equipment. AIS is voluntary for small vessels, whereas it is required for large ships. The popularity of AIS in small vessels may explain some of the increase in reported emissions, especially for CO, in the Baltic Sea area. Over the long term, this will help improve the quality of emission projections because less small vessel traffic will fall outside vessel tracking and emission calculations. However, no centralised database exists for the technical specifications of small vessels which are required for emission studies. To date, default specifications of tugboats are used for small vessels, which may overestimate their contribution.

15.2.2.3 Future Projections

Estimates of future shipping emissions vary widely. Although there is some variation in forecasts of ship movements, the main cause of variation in emission projections relates to the wide range of technologies that could be used to reduce pollutant emissions. Figure 15.5 illustrates the range in one such study (Eyring et al. 2005a) where projections started in 2002. As noted by Eyring et al. (2010),

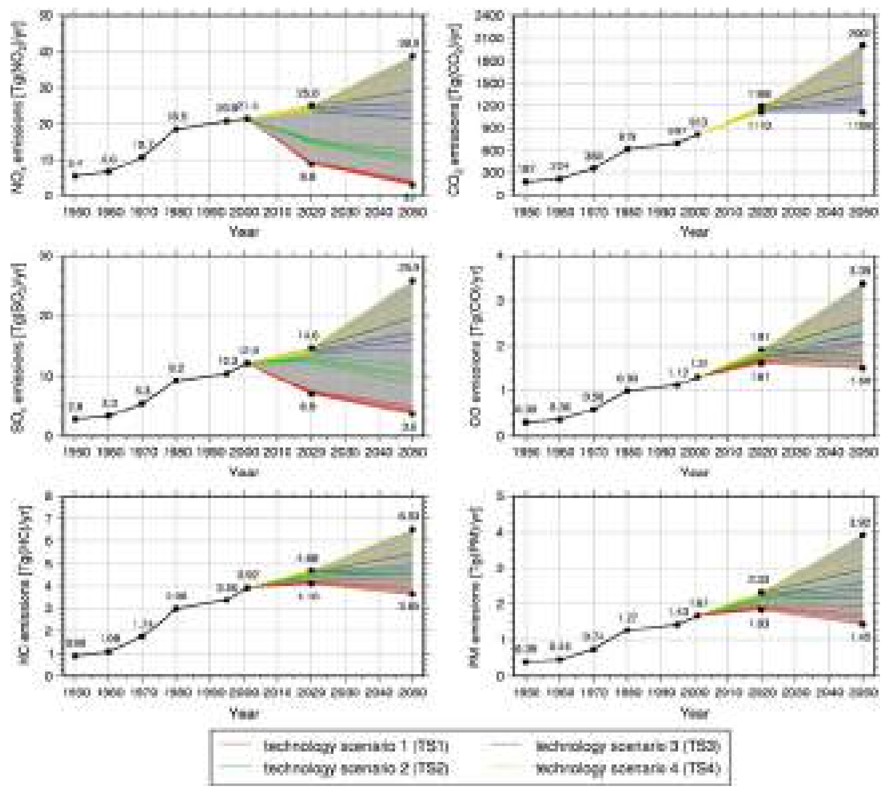

Fig. 15.5 Possible range of future emissions of nitrogen oxides (NO_X), carbon dioxide (CO_2), sulphur oxides (SO_X), carbon monoxide (CO), hydrocarbons (HC) and particulate matter (PM) according to four technology scenarios (TS1-4) and four ship traffic demand scenarios (giving four lines per technology scenario). Eyring et al. (2005a; which also provides details of the scenarios), figure redrawn by V. Eyring

even this wide range in projected emissions could not include the unexpectedly high growth rates in shipping traffic between 2002 and 2007.

The Baltic Sea nations have finalised an application to the IMO for an NO_X emission control area (NECA), but this has not yet been submitted to the IMO because of a lack of political consensus in the Baltic Sea countries. The decision taken by the IMO Marine Environmental Protection Committee (66th session, 2014) allows a date for NECA establishment beyond the previously agreed 2016. Declaration of the Baltic Sea as a NECA would make NO_X reduction techniques mandatory for all vessels built after the starting date defined by the countries applying the NECA status. Designation of the Baltic Sea as a NECA should cut NO_X emissions from ships by 60 %, but as the stricter emission standards will apply only to newly built ships, the gradual phase-out of existing more polluting vessels means that the full effects will not be seen until 30 years after the effective date (Acid News 2012).

15.2.3 Land and Sea Emissions—Impact of Climate Change

The wider impact of climate change on combustion emissions (e.g. NO_X, SO_X) involves complex policy and technology decisions and is beyond the scope of this chapter. In contrast, emissions of NH_3 are highly dependent on climatic variables, especially temperature (e.g. Hertel et al. 2012, and references therein). An important new understanding is that the projected rise in temperature may be enough to induce significant increases in NH_3 emissions owing to increased evaporation; Skjøth and Geels (2013) and Sutton et al. (2013) suggested possible increases of 20–50 % over the next century. Model studies to assess the potential impact of this important climate effect on nitrogen deposition and the exceedance of critical loads have begun, but as noted by Sutton et al. (2013), they will require a new paradigm of emission and deposition calculations.

A rise in temperature is also commonly assumed to cause an increase in biogenic VOC emissions and hence results in higher O_3 and PM levels (e.g. Langner et al. 2012b; Doherty et al. 2013; Hedegaard et al. 2013). However, some studies indicate that increased levels of atmospheric CO_2 may reduce some biogenic VOC emissions (e.g. Young et al. 2009); the net impact of climate change on such emissions is still unknown.

A potentially important impact of climate change on shipping emissions concerns the Arctic. The observed decline in Arctic sea ice is likely to result in a significant increase in shipping activities. According to Corbett et al. (2010), NO_X emissions from Arctic shipping could increase by a factor of 1.7 (by 2030) and 3.8 (by 2050), relative to 2004. If new shipping routes are introduced (such as the

Northern Sea Route and Northwest Passage), these increases will be even higher. The impacts of these changes are discussed in Sect. 15.4.

15.3 Observed Concentrations and Deposition

Concentrations of many pollutants over the Baltic Sea region have changed significantly over the past century, mainly due to changes in emissions, either within Europe or globally. However, systematic measurements (with a few exceptions) did not really begin until the 1970s and later. Many of the sulphur- and nitrogen-related pollutants have their greatest environmental impact after they are deposited, while concentrations of aerosol species in particular are of concern due to their health impacts (Sect. 15.6). Ozone is a toxic gas affecting human health and vegetation.

15.3.1 Sulphur and Nitrogen

Tørseth et al. (2012) used data from a selection of long-term monitoring sites within the EMEP network to analyse trends in sulphur, Nr compounds and O_3 across Europe. Although they reported concentration information for many compounds, this section focuses on a few key findings plus some complementary studies. Sulphur compounds have been measured for the longest period within EMEP. Figure 15.6 illustrates both the network development and trends in total SO_4^{2-} (including sea-salt sulphate) from 1974 to 2009.

There is a clear reduction in total SO_4^{2-} concentration across the entire region. Tørseth et al. (2012) also presented (more limited) data on non-sea-salt (nss) SO_4^{2-} and SO_2. The EMEP monitoring results for SO_4^{2-} in air and precipitation from 1980 to 2009 reflect the emission reductions achieved throughout Europe, whereas the observed reductions in SO_2 are even greater than implied by emission reductions. This apparent nonlinearity is probably due to a combination of factors, including changes in the oxidation capacity of the atmosphere, changes in cloud water pH, changes in surface deposition and changes in emission height (Lovblad et al. 2004; Fowler et al. 2007, 2009).

Earlier assessments (Bartnicki and Lovblad 2004; Lovblad et al. 2004) also found the widespread reduction in observed concentrations of SO_2 and nss-sulphate in air in the Baltic Sea countries in the period 1980–2000 to be in line with sulphur emission reductions. The reduction in concentrations of SO_2 observed in the Baltic Sea region were in the range 85–90 % for stations in Denmark, Finland, Sweden and Germany, while the reductions were smaller at 50–80 % for stations in Belarus, Estonia, Latvia, Lithuania and Poland (Lovblad et al. 2004). The nss-sulphate concentrations in air and precipitation have also decreased substantially, but the decrease has not been as large as the decrease in SO_2. Vuorenmaa (2004) found significant decreasing trends in SO_4^{2-} in precipitation all over Finland in the period 1986–2000.

Tørseth et al. (2012) compared the observed trends in sulphur (from 1980) and Nr (from 1990) to 2009. The analysis for the Nr compounds was especially limited because of the lack of complete and consistent data across all years, but the results illustrate the significant changes in air quality that have taken place in this period. As shown in Fig. 15.7 and discussed in more detail by Tørseth et al. (2012), the trends in SO_X, NO_X and NH_3 emissions are generally reflected in the measurements, with the exceptions of SO_4^{2-} (discussed above) and NO_3^-. Colette et al. (2011) also found robust trends in NO_2 over the past ten years, with the majority of monitoring stations in Europe showing reductions of about 50–60 % over this period.

The concentrations of total airborne NO_3^- decreased on average by only 8 %. These differences in trends can partly be explained by a shift in equilibrium towards more particulate ammonium nitrate relative to HNO_3 caused by a reduction in SO_2 emissions (Fagerli and Aas 2008). A more rapid oxidation of NO_X may also have contributed (Monks et al. 2009). The total reduction in observed concentrations of NO_X from 1980 to 2009 is larger than from 1990 to 2009. A similar trend is not seen in the emission data, but the discrepancy may partly be explained by significant changes in the number and location of sites when comparing the 1980s with more recent years (Tørseth et al. 2012).

Tørseth et al. (2012) reported that the total European NH_3 emissions decreased by 29 % from 1990 to 2009, although with large regional differences. A majority of the EMEP sites show a decreasing trend both in air and precipitation, on average 24–25 %. It should be noted that some EMEP sites are, due to their location in rural districts, partly affected by local NH_3 emissions (Tørseth et al. 2012). Pihl Karlsson et al. (2011) also investigated changes over about a decade in sulphur and nitrogen air concentrations, deposition and soil water concentrations in forest ecosystems in Sweden as well as in other Nordic countries. Their analysis of the time series 1996/97–2007/08 showed that SO_2 and NO_2 air concentrations have decreased substantially, whereas there was no trend for NH_3. This study found reductions in SO_4^{2-} deposition for the majority of monitoring sites across Sweden and other Nordic countries and that the reductions were of the same order of magnitude as the European emission reductions. Soil water SO_4^{2-} concentrations decreased at most, but not all, monitoring sites across Sweden in parallel with the SO_4^{2-} deposition reductions. The soil water acidification indicators—pH, acid neutralising capacity (ANC) and inorganic aluminium concentrations—indicated acidification recovery at some sites, but there were also many sites with no significant change. Despite the substantial decrease

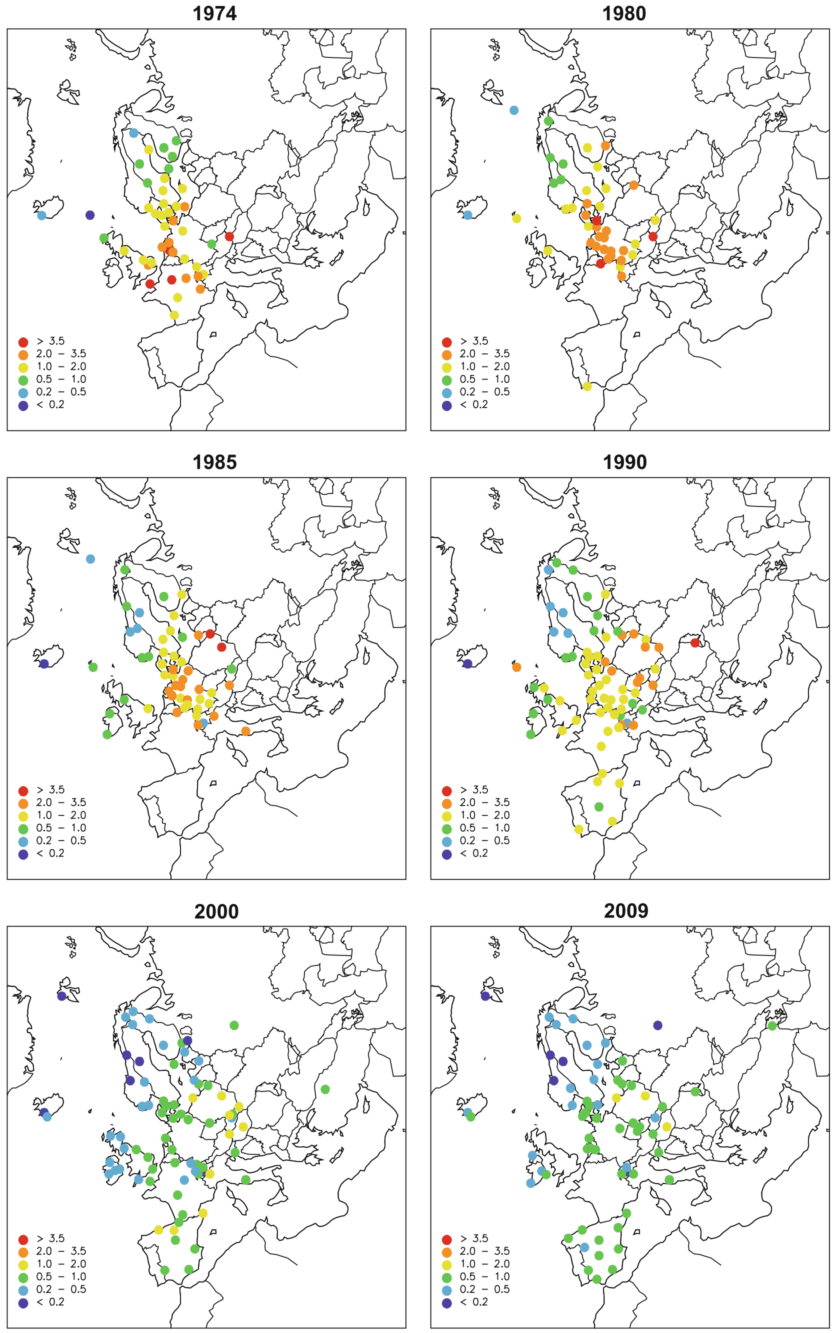

Fig. 15.6 Annual mean concentrations of sulphate (SO_4^{2-}) in aerosols across Europe from 1974 to 2009, as estimated from the EMEP network for selected stations. Unit: μg S m^{-3}, Tørseth et al. (2012)

in NO_2 air concentrations, no statistically significant decrease in the bulk deposition of inorganic nitrogen could be demonstrated. (They were not able to include dry nitrogen deposition in the trend analyses, however.) Elevated NO_3^- concentrations in the soil water occurred intermittently, but also after massive felling caused by severe storm events, particularly at some sites in southern Sweden. This indicated that nitrogen stocks in the forest soils of southern Sweden are increasing and may be approaching saturation.

15.3.2 Ozone

There has been a significant increase in O_3 since the start of the twentieth century (Fig. 15.8), which is largely attributed to changes in anthropogenic emissions of NO_X and other precursors (Monks et al. 2009; Parrish et al. 2009, 2012). Annual trends appear to have flattened out in Europe since about 2000. The reasons for this are not fully understood, but reductions in European emissions are certainly affecting

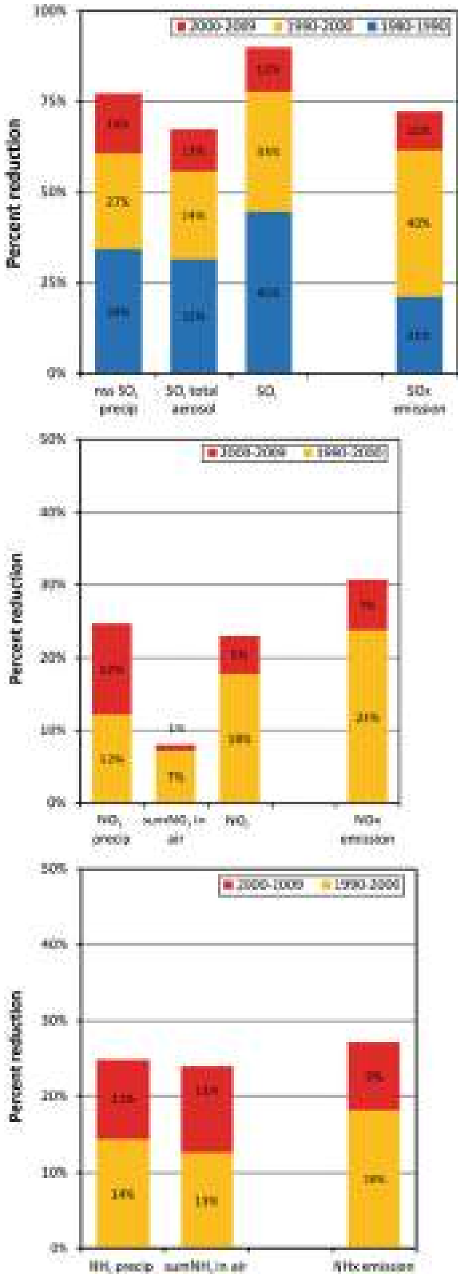

Fig. 15.7 Observed reduction in sulphur and nitrogen compounds relative to emission reductions in Europe for ten-year periods from 1980 (sulphur) and 1990 (nitrogen). Sulphur reductions are calculated using data from the 14 sites that measured all three compounds since 1980; although data from 14 sites were also used for nitrogen, it is not necessarily the same site used for all compounds, Tørseth et al. (2012)

O_3 trends. In general, mean O_3 levels in Europe are increasing in wintertime because of the reduction in NO_X emissions (the NO_X-titration effect, important in winter, in which NO_X acts as a sink rather than a source of O_3). In summer, the European emission reductions help to reduce O_3, although these are sometimes counteracted by increasing hemispheric background levels (e.g. Jenkin 2008).

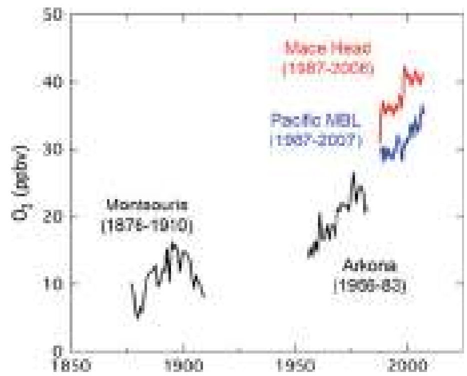

Fig. 15.8 Comparison of the 12-month running mean ozone (O_3) concentrations from three northern European sites and the Pacific marine boundary layer (MBL), Parrish et al. (2009)

Colette et al. (2011) presented trends in O_3 (as well as NO_2 and PM_{10}) over the past ten years. They found that (as expected) the trends in O_3 reflect trends in NO_2. On average, O_3 increases of 0.37, 0.27 and 0.05 $\mu g\ m^{-3}$ year^{-1} were found at urban, semi-urban and rural sites, respectively. For daily means, the trends were positive at about 31 % of stations, whereas for daily peaks, only 18.5 % of stations showed positive trends. This difference between mean and peak O_3 trends is reasonably well understood (e.g. Jonson et al. 2006; Vautard et al. 2006; Jenkin 2008; Wilson et al. 2012); in short, mean O_3 is increasing, while peaks in O_3 are being reduced. Figure 15.9 illustrates the decline in peak O_3 concentrations at some Nordic sites.

15.4 Modelled Concentrations and Deposition

The number of model studies specifically addressing sulphur deposition to the Baltic Sea region is limited. Many studies cover the region but do not report estimates specifically for the Baltic Sea or for the surrounding catchment. Hongisto et al. (2003) reported estimates for 1993–1998 for the Baltic Sea and some of the surrounding land areas, and in the work of EMEP, model estimates of sulphur deposition to the Baltic Sea and individual surrounding countries are available for different years starting from 1985. Based on the EMEP model results, total deposition of nss-sulphur to both the Baltic Sea and the countries bordering the Baltic Sea decreased by about 80 % between 1985 and 2009 to a level of about 7700 t S year^{-1} for the region as a whole region and about 1300 t S year^{-1} for the Baltic Sea itself in 2009 (Table 15.4). The EMEP estimate for the Baltic Sea is comparable to that by Hongisto (2003) for 1993–1998 of about 1500 t S year^{-1}.

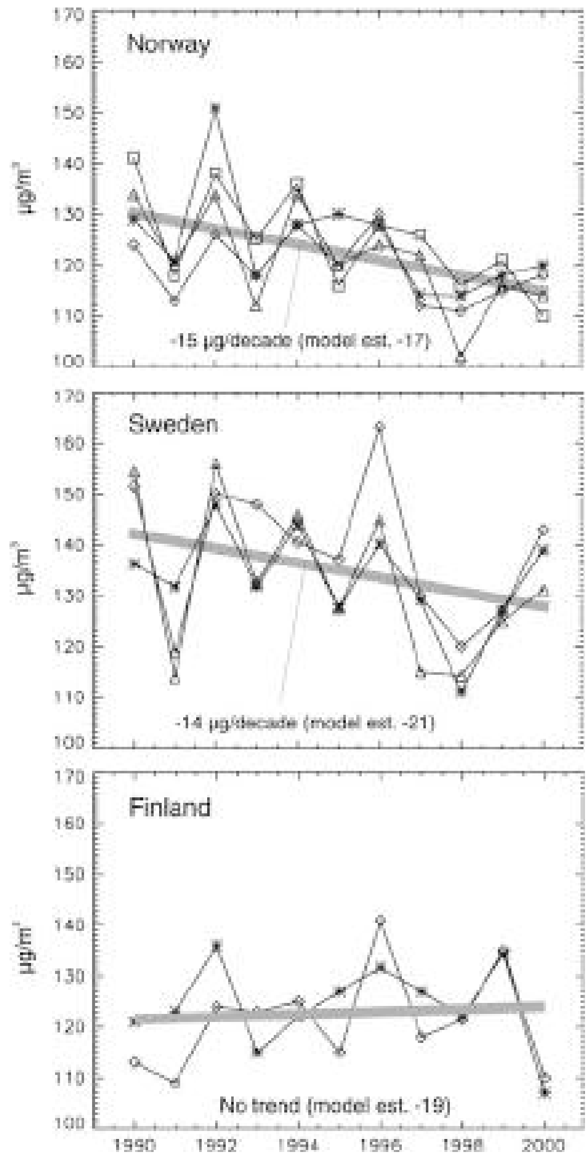

Fig. 15.9 Hourly observed ozone (O_3) 99th percentile concentrations for 1990–2000 at sites in the southern parts of Norway, Sweden and Finland. The *shaded grey line* indicates the regression of the average of the 99th percentiles. The trend estimated by this regression is given together with the estimated trend in modelled 99th percentiles (MATCH model), Solberg et al. (2005)

Several modelling studies have specifically addressed the deposition of Nr to the Baltic Sea and surrounding areas (de Leeuw et al. 2001, 2003; Hertel et al. 2002; Schlunzen and Meyer 2007; Bartnicki and Fagerli 2008; Langner et al. 2009; Bartnicki et al. 2011; Geels et al. 2012b). As illustrated in Figs. 15.10 and 15.11, wet deposition of nitrogen is greater than dry deposition of nitrogen, and oxidised nitrogen deposition is greater than reduced Nr deposition. All studies show that modelled dry deposition of both the oxidised and reduced forms of nitrogen exhibits strong south–north gradients across the Baltic Sea region, declining by well over an order of magnitude from Denmark to the northern part of Sweden (Langner et al. 2009). Indirect estimates of the atmosphere as an ‘external’ source of nitrogen to the Kattegat undertaken within the MEAD project suggest that the atmosphere may account for a substantial fraction (~40 %) of the total flux (the sum of land runoff, upwelling flux and atmospheric deposition) during the summer months, dropping to less than 20 % for the year as a whole (Spokes et al. 2006). The estimates are in broad agreement with data for the Baltic proper (Elmgren and Larsson 2001; Rolff et al. 2008; Langner et al. 2009) but the wet, and particularly the dry, deposition fluxes are comparatively poorly constrained. Deposition to ice in the northern Baltic Sea comprises 6 % of the annual nutrient supply and up to 40 % of the annual cadmium and lead flux into the Bothnian Bay, implying that sea ice may play a key role in determining the timing and magnitude of chemical fluxes to the water column (Granskog and Kaartokallio 2004).

In general, nitrogen deposition originating from emissions on land has a strong gradient towards the sea. Ammonia is efficiently dry deposited close to the source areas and most of the reduced nitrogen that reaches the open sea comes in the form of ammonium particles which are efficiently wet deposited. NO_X deposition has a weaker gradient, reflecting a longer residence time in the atmosphere (NO and NO_2 do not deposit efficiently, but are transformed to HNO_3 which is efficiently dry deposited or forms NO_3^- aerosols). Furthermore, slower deposition processes for aerosols over water surfaces are assumed in all models.

Some studies have assessed the contribution of different countries to Nr deposition in the Baltic Sea region. For example, Geels et al. (2012b) using the DEHM model, estimated that the nine countries bordering the Baltic Sea contributed about 50 % of the Nr deposited in both 2007 and a projected 2020 scenario, with Germany the largest single contributor (Fig. 15.12). Bartnicki et al. (2011), using the EMEP MSC-W model (Simpson et al. 2012), found greater contributions from some countries, with five contributing about 55 % of total Nr deposition, and emissions from international shipping on the Baltic Sea contributing 4–5 % (Fig. 15.13). They also found Germany to be the single largest contributor (almost a factor of two greater than Poland at 12 %), but that even the UK made a significant contribution (7 %). As to source types, Hertel et al. (2002) estimated that around 40 % of the nitrogen deposited to the North Sea originated from agricultural activities and that around 60 % was from combustion sources.

Table 15.5 compares nitrogen deposition estimates from a number of studies using several chemical transport models (CTMs). For current years, the CTMs gave similar estimates: 230–260 kt N for 1995 or near 200 kt N for 2006/2007. To estimate changes in nutrient loads to the Baltic Sea over a longer period (1850–2006), Ruoho-Airola et al. (2012)

Table 15.4 Sulphur deposition trends for the HELCOM Contracting Parties and Baltic Sea (t S year^{-1})

Country	1985 EMEP (1994)	1990 EMEP (1994)	1995 EMEP (1998)	1993–1998 Hongisto (2003)	2009 EMEP (2011)
Denmark	716	547	320		183
Estonia	530	430	311		133
Finland	2262	1659	968		431
Germany	17,080	12,428	5440		1966
Latvia	850	600	439		211
Lithuania	1160	890	565		276
Poland	12,517	11,456	5948		2524
Sweden	2737	2205	1359		641
Baltic Sea	5853	4218	2572	1526	1334
Total	43,705	34,433	17,922		7699

Note that, EMEP model versions and inventories have changed over the course of the reports cited, so the data are not fully consistent

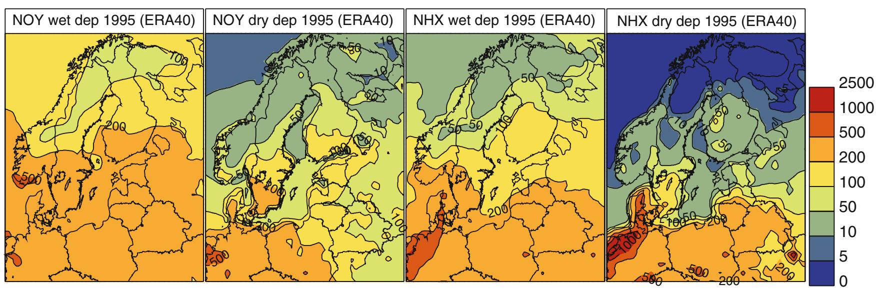

Fig. 15.10 Simulated wet and dry deposition of oxidised (NO_y) and reduced (NH_X) nitrogen in 1995 from the MATCH-ERA40 model. Units: mg N m^{-2} year^{-1}(Langner et al. 2009)

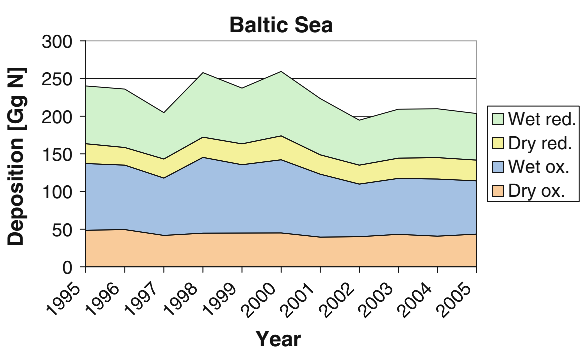

Fig. 15.11 Time series of annual nitrogen deposition to the Baltic Sea for 1995–2005 as calculated by the EMEP MSC-W model. Adapted from Bartnicki and Fagerli (2008)

combined time series of deposition data, published historical monitoring data and deposition estimates, and recent emission estimates. Figure 15.14 presents their estimates of annual deposition, with changes in oxidised Nr similar to the changes in emissions of Fig. 15.1.

15.4.1 'Climate' Meteorology, Impacts of Climate Change

Modelling historical and future changes in concentration and deposition shares many common features. In both cases, the models must be driven by 'estimates' of meteorology rather than by observed meteorological fields. Hindcasting recent decades is easier than forecasting the future, although the lack of satellite data and other observations in earlier years makes such data less reliable than current numerical weather prediction systems can deliver.

The most valuable historical data set for CTMs is ERA40, a set of meteorological data going back to 1957 that has been produced from global meteorological reanalysis by the European Centre for Medium-range Weather Forecasts (ECMWF) (Uppala et al. 2005). For future meteorology, many global climate models (i.e. general circulation models; GCMs) are available, but for estimating changes over the Baltic Sea region, it is better to run finer resolution models that are forced by GCMs and more capable of capturing the effects of local topography and land cover. An important

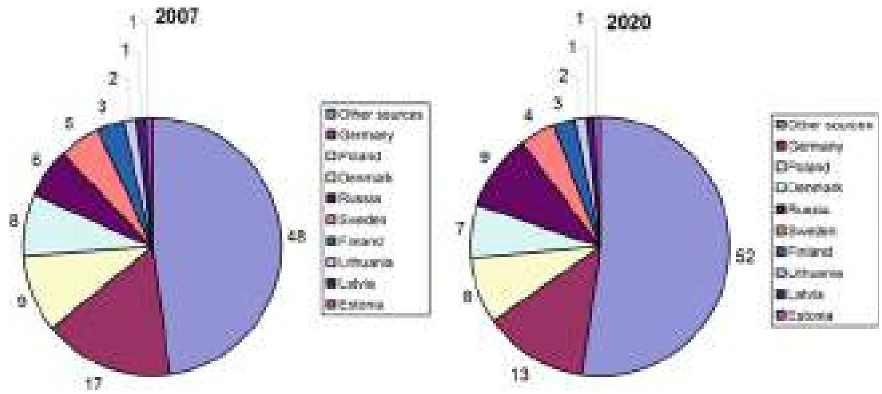

Fig. 15.12 Nitrogen deposition to the Baltic Sea as calculated using the DEHM model, subdivided according to the contribution from the nine bordering countries and other sources (i.e. remaining emissions in the model domain). The contributions are given as percentages for the present-day scenario and a projection for 2020, Geels et al. (2012b)

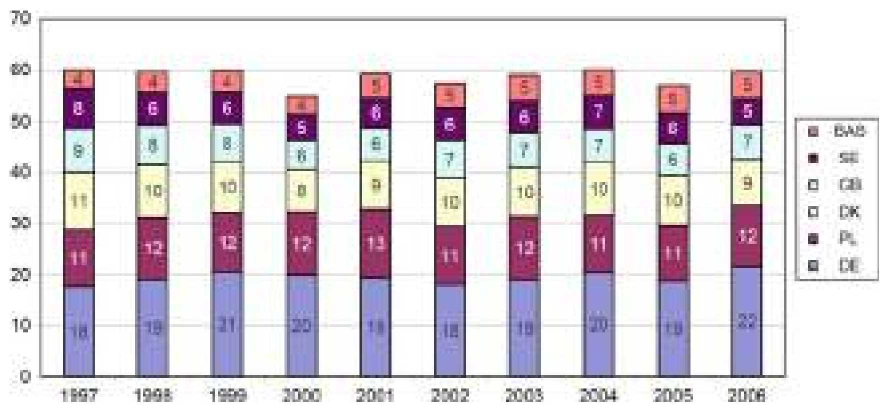

Fig. 15.13 Major contributions to the annual deposition of nitrogen to the Baltic Sea basin for 1997–2006. Contributions are shown as a percentage of total deposition. International ship traffic on the Baltic Sea (*BAS*), Sweden (*SE*), UK (*GB*), Denmark (*DK*), Poland (*PL*) and Germany (*DE*), Bartnicki et al. (2011)

data set in this respect is that generated by the Rossby Centre regional climate model, RCA3 (Samuelsson et al. 2011).

Estimates of future concentrations and deposition depend on forecasts of both emissions and meteorology. Many studies have shown that meteorological/climate factors are generally less important than emission changes for concentrations and deposition (Hole and Engardt 2008; Langner et al. 2012a, b; Doherty et al. 2013; Engardt and Langner 2013; Hedegaard et al. 2013). Such simulations suggest that specification of future emissions is almost certainly the major factor when attempting to project future deposition amounts. Figure 15.15 (from Langner et al. 2012a) presents a clear illustration of this for O_3 (see also Engardt et al. 2009), comparing daily maximum values calculated by the MATCH model for scenarios that examine the relative impact of changes in climate, European emissions and increasing hemispheric background levels of O_3. Emission changes are clearly of most importance in these model runs, although the background O_3 level is also significant. Wild et al. (2012) highlighted the large uncertainties in assumptions concerning changes in background O_3 levels.

The MATCH model has been run for both historical and future scenarios in a number of studies. Andersson et al. (2007) used ERA40 to run MATCH for 1958–2001. Hole and Engardt (2008) used 30-year periods of meteorological data produced by RCA3, with forcing by the ECHAM4/OPYC3 GCM (Roeckner et al. 1999). These simulations were for the SRES A2 scenario (Nakićenović 2000) in 'transient' mode from 1961 to 2100 with gradually changing climate forcing, that is changing atmospheric aerosol and greenhouse gas concentrations. MATCH was applied to data for three different time slices (1961–1990, 2021–2050 and 2071–2100), representing past and future climates. The two set-ups are denoted MATCH-ERA40 and MATCH-RCA3, respectively, and Langner et al.

Table 15.5 Comparison of model estimates of total, dry and wet deposition of nitrogen (kt N yr^{-1}) to the Baltic Sea

Year	Model	Dry	Wet	Total	Comments
1995	HILATAR			255	(a)
	EMEP rv2.5 EMEP			244	(a, b)
	rv3.1 MATCH-			230	(c)
	ERA40			260	(a)
1996–2000	EMEP			300	(a, b)
	MATCH-ERA40			271	(a)
2006	EMEP rv3.1			199	(c)
2007	DEHM			203	(d)
Historical					
1961–1990	MATCH-RCA3	41	207	248	(a), year 2000 emissions
Projections					
2020	DEHM			165	(d), projected emissions
2021–2050	MATCH-ERA40	42	206	248	(a), year 2000 emissions
2071–2100	MATCH-ERA40	44	218	262	(a), year 2000 emissions

Extended from Langner et al. (2009)
Notes (a) Langner et al. (2009); (b) EMEP model rv2.5 from ca. 2005–2006; (c) EMEP model rv3.1 from 2008, data from Bartnicki et al. (2011); (d) Geels et al. (2012b)

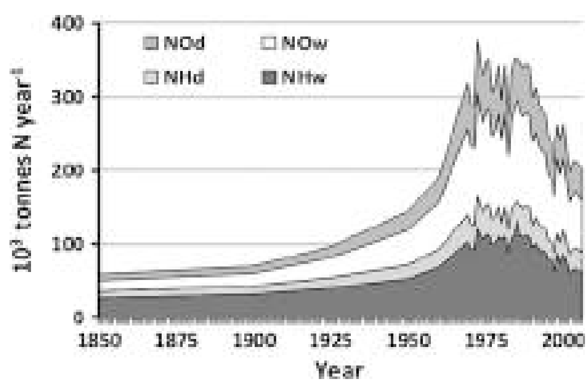

Fig. 15.14 Estimated annual integrals of inorganic nitrogen deposition to the Baltic Sea, 1850–2006. The graphic shows dry and wet deposition of oxidised nitrogen (NO_d and NO_w) and dry and wet deposition of reduced nitrogen (NH_d and NH_w), Ruoho-Airola et al. (2012)

(2009) investigated the results of these MATCH model runs for the Baltic Sea region in particular.

In a study to examine the potential effect of historical and possible future climate change (not emissions) and variability on atmospheric deposition of nitrogen to the Baltic Sea based on the MATCH model under the assumption of constant emissions, Langner et al. (2009) found modest projected increases when averaged over the entire Baltic Sea region (of 4–5 %), but generally increased deposition of oxidised nitrogen over the Baltic Sea. This tendency is of smaller magnitude than current interannual variability (Hongisto 2011).

For Europe, Geels et al. (2012b) used an inventory based on a combination of the EU Thematic Strategy on Air Pollution and scenarios for the 27 EU countries made by IIASA as part of the work towards a revised EU directive on national emission ceilings (NEC-II). For the remaining European countries and the western Asian countries, the projected emissions were based on estimates provided in the EU Clean Air for Europe (CAFE) programme. For the rest of the northern hemisphere, 2020 emissions were based on the RCP 3-PD projections (van Vuuren et al. 2007). Ship emissions from the area around Denmark were assumed to follow new regulations adopted by the IMO, and the same projections were used for the North Sea and Baltic Sea. For the nine countries bordering the Baltic Sea, nitrogen emissions were projected to decline by about 50 % between 2007 and 2020 (although for Russia the projected decrease was just 11 %).

Another potential impact of climate change is increased emissions from shipping due to reduced sea ice in the Arctic. Tuovinen et al. (2013) assessed the changes in nitrogen deposition and the phytotoxic ozone dose (POD) arising from both climate change alone, and from potential changes in ship emissions, including a possible new Northeastern Passage route (Corbett et al. 2010). The emissions changes resulted in significant changes in POD and nitrogen deposition, mainly along the Norwegian coastline. POD values increased in some areas by more than 10 %, but decreased close to sources; such nonlinearity is expected and common in O_3 scenarios, but highlights the need for high spatial resolution in such model simulations.

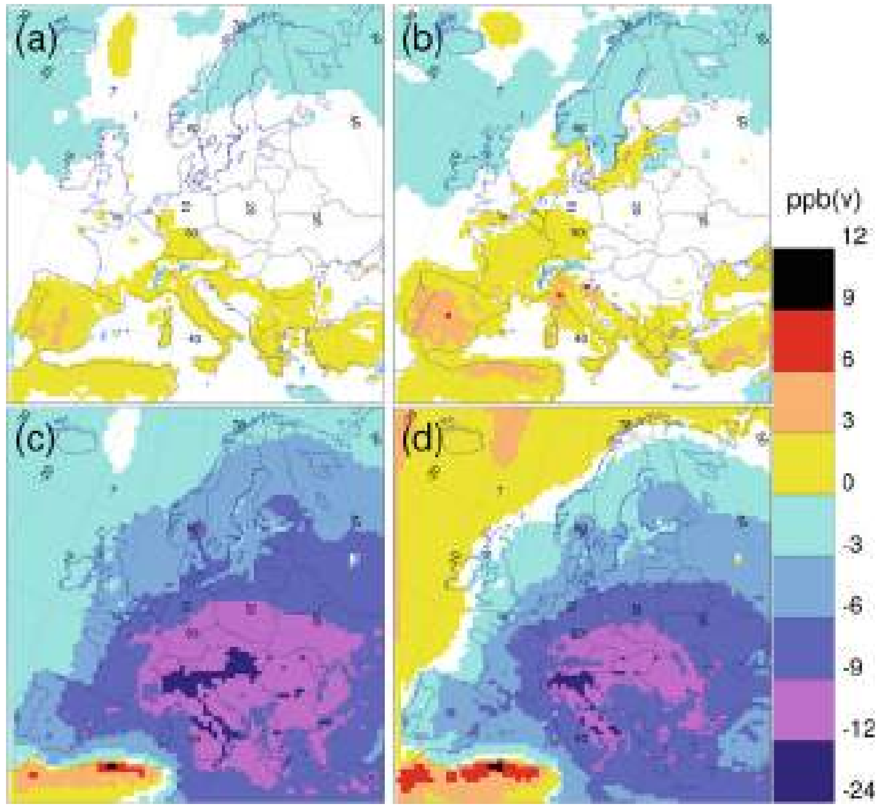

Fig. 15.15 Projected change in summer (April–September) daily maximum surface ozone (O_3) concentration for 2040–2059 relative to 1990–2009 **a** and **b** changes due to change in climate only; RCA3 downscaling of ECHAM5 and HadCM3, respectively. **c** ECHAM5 downscaling and change in European O_3 precursor emissions. **d** Increasing boundary case; change due to combined change in climate, change in O_3 precursor emissions and an increasing hemispheric background of O_3 level of 0.1 ppb(v) year^{-1}. Non-significant changes at the 95 % confidence level are masked *white*, Langner et al. (2012a)

15.5 Uncertainty of Estimates

Table 15.5 presented a comparison of nitrogen deposition estimates from a number of studies using CTMs. As was apparent, for any given year, the different CTMs seemed to give similar estimates. Much of the similarity may be ascribed to the use of similar emissions data and that much of the deposition is driven by precipitation events that are not that sensitive to model formulation. Consistency of emissions data, or deposition estimates, is not always guaranteed, however. For example, Winiwarter et al. (2011) found large discrepancies between the emissions and deposition over Europe as estimated by Schulze et al. (2010) and those from the EMEP inventory and EMEP model estimates. Model calculations clearly need to be thoroughly evaluated against observations.

15.5.1 Wet Deposition

Comparison of model results for wet deposition or concentrations in precipitation is in many ways more difficult than comparing gas concentrations. The most important issue concerning the wet removal of species in CTM models is probably the meteorological input; model performance for wet deposition fluxes or concentrations in precipitation is strongly limited by the quality of the numerical weather prediction models providing meteorological data. For example, models generally have problems with sub-grid precipitation, simulating precipitation more often, but in lower amounts, than reality. As precipitation scavenging is a complex and nonlinear process (e.g. Barrie 1992) such issues will cause errors in modelled wet deposition that are difficult to evaluate. There are also many uncertainties inherent in the deposition monitoring methods themselves (Draaijers and Erisman 1993; Erisman et al. 2005).

Large differences between models were found in the global models participating in the COSAM study, in which the wet deposition efficiency varied by a factor of four (Roelofs et al. 2001). A similar spread was also found for global models by Dentener et al. (2006) and Textor et al. (2006).

At the European scale, the EMEP MSC-W model (Simpson et al. 2012) has been subject to extensive evaluation against observed wet deposition estimates. Standard scatter plots showing model performance against observed concentrations of NO_3^- and NH_4^+ are available in the annual EMEP status reports (e.g. Fagerli and Hjellbrekke 2008; Berge and Hjellbrekke 2010). The EMEP MSC-W model results have also been compared to observed wet deposition for nitrogen from the ICP Forests monitoring network (Simpson et al. 2006). Differences in mean values between modelled and observed (ICP Forests) SO_4^{2-}, NO_3^- and

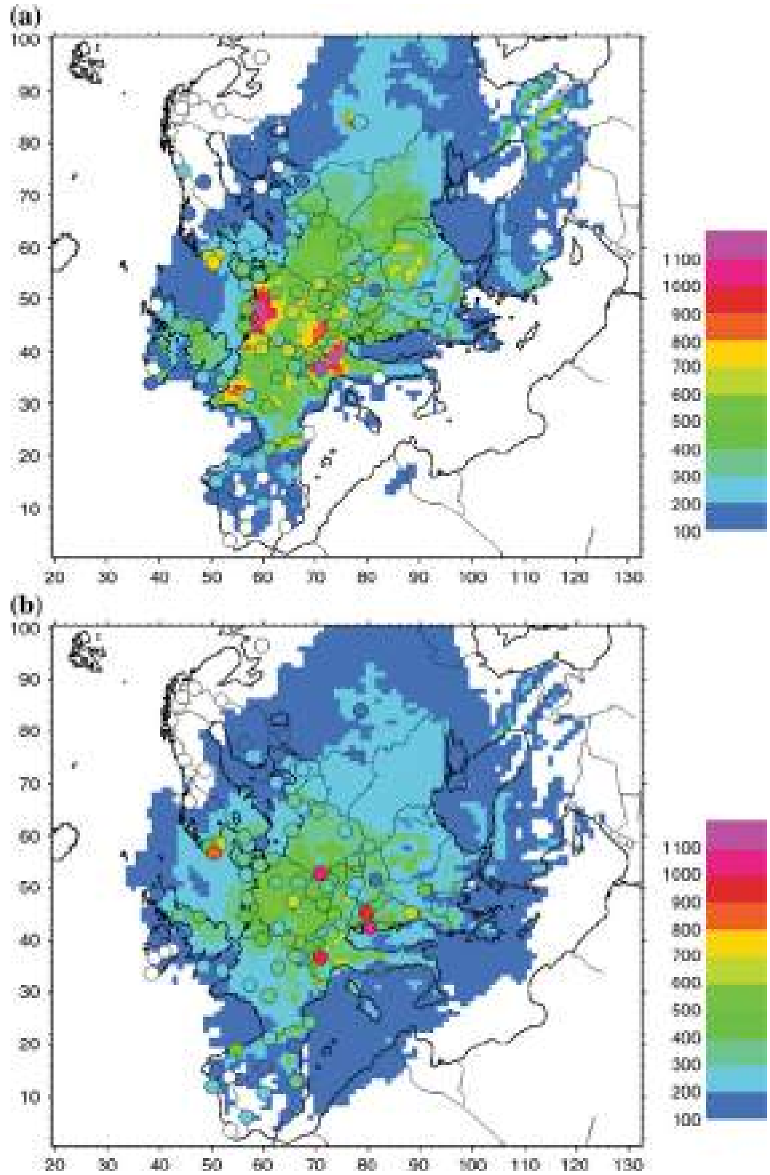

Fig. 15.16 Comparison of modelled and observed annual means of reduced nitrogen **a** and oxidised nitrogen **b** in wet deposition (mg N m^{-2} $year^{-1}$). Data are for 2001. The bullets depict observations with the same colour bar as the modelled field. Measured annual means are calculated by using the measured precipitation amount and the nitrate and ammonium concentration in precipitation, Simpson et al. (2011). The model used is the EMEP MSC-W model (Simpson et al. 2012)

NH_4^+ total and wet deposition were within 20 % (in 1997) and 30 % (in 2000), with the EMEP model showing slightly lower values than the observations (Simpson et al. 2006). Modelled and observed concentrations of SO_4^{2-}, NO_3^- and NH_4^+ in precipitation were on average very similar (differences of 0–14 %), and the correlation between modelled and observed data was high for this type of comparison (R^2 = 0.4–0.8 for most components and years).

Figure 15.16 compares measured wet deposition of oxidised and reduced nitrogen against results from the EMEP model. In these plots, the measured deposition is calculated using the measured precipitation amount and the NO_3^- and NH_4^+ concentration in precipitation. For reduced nitrogen, Fig. 15.16a reveals good agreement between modelled and measured values, across almost all of Europe. The high-modelled values near northern Italy are reflected in the measurements. Unfortunately, other regions with high estimated wet deposition have only a limited number of measurement sites (e.g. the Netherlands, Belgium), and so it is difficult to evaluate model performance here. The EMEP model has a tendency to underestimate wet deposition in Nordic sites.

For oxidised nitrogen (Fig. 15.16b), five sites stand out as having much higher measured wet deposition than modelled. The reason for this seems to be that the observed precipitation at the sites far exceeds the modelled precipitation (by a factor of two for the Norwegian site). However, there is good agreement between model results and measurements at almost all other sites, which provides confidence that the modelled wet deposition budget is within the uncertainty of the measured value.

15.5.2 Dry Deposition

Although wet deposition represents an important fraction of the nitrogen deposition over the Baltic Sea region, Figs. 15.10 and 15.11 show that dry deposition is also important. Many of the physical/chemical processes controlling dry deposition of Nr compounds were discussed by, for example, Flechard et al. (2011), Hertel et al. (2012) and Fowler et al. (2009). Deposition processes over land and sea involve different processes and challenges.

15.5.2.1 Land

Efforts to estimate aerosol particle dry deposition to terrestrial ecosystems face many of the same challenges as efforts to estimate deposition to water surfaces (see Sect. 15.5.2.2), but vertical velocities over terrestrial surfaces are typically greater (reducing the uncertainty associated with direct micrometeorological techniques) and platforms suitable for deployment of flux instrumentation are more readily available (and do not exhibit motion as would be experienced on a ship). Nevertheless, only limited direct measurements are available and are principally focused on size-resolved, rather than chemically resolved, fluxes (Pryor et al. 2008a; Fowler et al. 2009). Furthermore, there has recently been a greater focus on the aerosol particle diameters that dominate aerosol number concentrations (i.e. sub-micron) (e.g. Pryor et al. 2009) rather than aerosol particles in sizes that may dominate the chemical flux. Those recent studies have tended to indicate a very strong influence of vegetation canopy morphology and aerosol properties on deposition velocities, and thus, they explain—at least to some degree—the large variability in measurement data sets of particle number fluxes to vegetated surfaces taken under superficially similar atmospheric conditions (Petroff et al. 2009). Recent instrumentation innovations, such as the time-of-flight mass spectrometer (TOF-MS) and aerosol mass spectrometers (AMS), that are capable of measuring the size and chemically resolved aerosols with high time resolution has facilitated initial direct flux measurements (e.g. Nemitz et al. 2008; Thomas et al. 2009) over terrestrial surfaces. However, there remain comparatively large uncertainties on aerosol particle fluxes and the technical challenges, for example, artefacts associated with hygroscopicity or other non-stationarity in the aerosol size distribution (Kowalski 2001; Pryor and Binkowski 2004) exceed those associated with atmosphere–surface exchange of gases (Pryor et al. 2008a).

Recent studies within the EU NitroEurope (NEU) project (Sutton et al. 2007) also illustrate the level of uncertainty in dry deposition estimates. Flechard et al. (2011) conducted inferential modelling with deposition codes from three European dry deposition models at selected sites across Europe. This study suggested that NH_3 is the single highest atmospheric Nr dry input in many parts of Europe. At suburban sites of the NEU network, HNO_3 and particulate NO_3^- and NH_4^+ also contributed significant fractions of the total dry deposition. There were however substantial differences between models, with annual deposition rates varying as much as twofold between models at given monitoring sites. This highlights the variability in model parameterisations, stemming from the variability in measured deposition rates and canopy resistances. For NH_3, the stomatal compensation point and the external leaf surface (or non-stomatal) resistance were the largest sources of divergence between models.

The importance of bidirectional fluxes was also discussed by Geels et al. (2012b). They noted that although several parameterisations of bidirectional fluxes over land exist for NH_3, they have so far mainly been used in field-scale NH_3 exchange models (Massad et al. 2010). Geels et al. (2012b) also noted that bidirectional fluxes have been observed over marine surfaces (Hertel et al. 2006) and the inclusion of such fluxes in a CTM can lead to a redistribution of the deposition in the coastal areas and hence in the gradients of nitrogen deposition over the sea (Sorensen et al. 2003).

Model estimates of aerosol deposition velocity (Vd) differ greatly among the various modelling approaches and parameterisations (see Ruijgrok et al. 1997 for a review), but it is in the size range 0.1–1.0 μm that variability and uncertainty are greatest. Whereas mechanistic models predict very low deposition velocities for fine aerosols, typically of the order of 0.1 mm s^{-1}, field measurements suggest that Vd is 1–3 orders of magnitude higher (Zhang et al. 2001; Gallagher et al. 2002). Nevertheless, such field measurements are also subject to great uncertainty (Rannik et al. 2003; Pryor et al. 2008a, b). This is especially relevant for Nr in the aerosol phase, as NH_4^+ and NO_3^- are mostly (>90 %) present as sub-micron particles.

15.5.2.2 Sea

The difficulty in making in-site direct aerosol particle dry deposition observations over water derives principally from the following: (i) the bidirectionality of the flux (i.e. the surface acts as both a source and sink for particles), (ii) challenges in making direct size and composition resolved measurements with sufficient time resolution to allow application of micrometeorological techniques and (iii) the typically low turbulence intensity (which both suppresses vertical transport and can challenge flux detection) (Pryor et al. 2008a). For this reason, the overwhelming majority of studies focused on aerosol particle atmosphere–surface exchange (including those focused on nutrient supply) continue to take time-averaged measurements of aerosol particle size and composition and apply a parameterised model of the dry deposition rate to determine the deposition flux (e.g. Matsumoto et al. 2009; Buck et al. 2010; Uematsu et al. 2010). Such studies and recent numerical modelling have suggested a key role for atmospheric transport and deposition of aerosol particles in nutrient supply (Krishnamurthy et al. 2010), as well as toxin transport (Paytan et al. 2009) to aquatic ecosystems. Also, that in some environments and for some key micro- and macro-nutrients, dry deposition of aerosols to water surfaces may dominate over the wet deposition flux (Uno et al. 2007; Tian et al. 2008). However, aerosol deposition velocities used in such studies are poorly constrained and flux estimates derived thus exhibit large uncertainties, in part because aerosol particle dry deposition velocities exhibit multiple functional dependencies beyond the direct dependence on aerosol particle diameter. For similar reasons, many postulated functional dependencies remain essentially unverified. For example, it has been proposed that transfer across a thin laminar layer close to the surface is a major limiting factor for deposition rates (Slinn and Slinn 1980; Giorgi 1986; Hummelshøj et al. 1992; Pryor et al. 1999) and that the observed increase in particle dry deposition with wind speed may be linked to disruption of that layer by bubble burst activity. Indeed, one model study showed that bubble burst activity almost doubled the deposition velocity of aerosol particles in the diameter range of 0.1–1.0 µm (Pryor and Barthelmie 2000), however, in a wave tunnel experiment deposition velocities for magnesium oxide particles in the diameter range 0.1–1.0 µm showed an enhancement of ≤30 % (Larsen et al. 1995).

15.6 Human Health Effects of Air Pollution

In a major review of the health effects of fine particulate air pollution, Pope and Dockery (2006) stated that ‘Despite important gaps in scientific knowledge and continued reasons for some scepticism, a comprehensive evaluation of the research findings provides persuasive evidence that exposure to fine particulate air pollution has adverse effects on cardiopulmonary health’. In their extensive review, Pope and Dockery (2006) concluded ‘There has also been emerging evidence of PM-related cardiovascular health effects and growing knowledge regarding interconnected general pathophysiological pathways that link PM exposure with cardiopulmonary morbidity and mortality’. They pointed to substantial progress in research providing several interesting observations and viable hypotheses for how the human body reacts to inhaled particles. Most investigations to establish the relative importance of different air pollution components for health effects indicate particle mass to be of major importance. However, additive or synergistic effects of co-occurring pollutants may also important (Pope and Dockery 2006).

Estimates of current and future health impacts of air pollution were presented by Amann et al. (2011) using the GAINS model. The GAINS methodology is based on using pollutant concentration data from GAINS itself or EMEP, linked to population maps across Europe. For different emission scenarios, GAINS quantifies premature mortality that can be attributed to long-term exposure to $PM_{2.5}$. The link to health effects follows the outcomes of the American Cancer Society cohort study and its re-analysis (Pope et al. 2002, 2009).

For the health impact assessment of policy scenarios, GAINS calculates the loss in statistical life expectancy as the total amount of life years lost (YOLL) for the entire population over 30 years. Health impacts for people younger than 30 years, and in particular impacts on infant mortality, are presently not considered in GAINS.

Figure 15.17 shows the impact of PM across Europe for 2005 and a 2020 scenario. The GAINS model estimates that the average loss in statistical life expectancy that can be attributed to exposure to fine particulate matter ($PM_{2.5}$) could decline from 7.4 months (2005) to 4.4 months (2020)

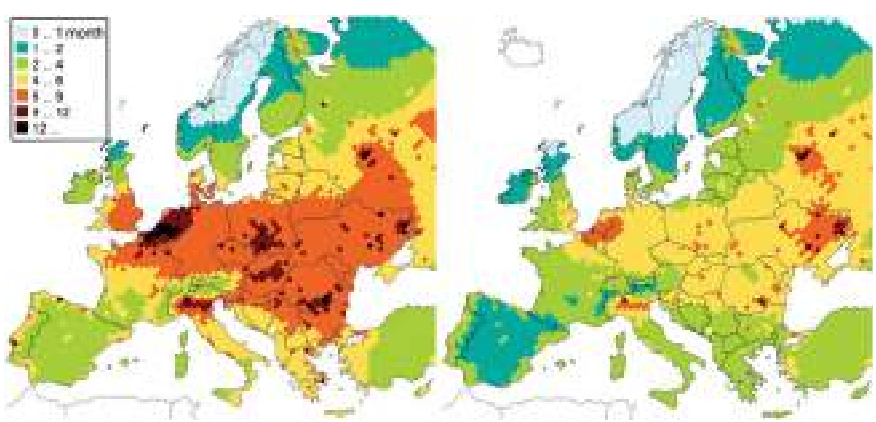

Fig. 15.17 Loss in statistical life expectancy attributable to exposure to fine particulate matter ($PM_{2.5}$) from human activities. *Left* 2005; *right* baseline projection for 2020, Amann et al. (2011)

in the EU-27 and to 6.1 months in non-EU countries. Particularly, high threats to human health occur in industrial areas, where air pollution is estimated to shorten life expectancy by more than a year (Amann et al. 2011).

15.7 Conclusion

This chapter has documented the changes in atmospheric chemical components over the Baltic Sea region, using a mixture of measurements, emissions estimates and modelling. The focus was on the period from around 1900 to 2050, since this period has received the most attention in the literature. The majority of the information concerns sulphur and Nr compounds, as well as O_3, since these pollutants have significant ecosystem and health impacts.

Emissions of sulphur and nitrogen species, and of other ozone precursors such as hydrocarbons, have changed significantly over the last 100 years, although very differently for land- and sea-based sources. Land-based emissions generally peaked around 1980–1990 and have since been reduced as a result of emissions control measures. Emissions from shipping have been steadily increasing for decades, but recent measures have reduced sulphur and particulate emissions. Future developments depend strongly on policy developments.

In general, the main driver of changes in atmospheric concentration and deposition with time is found to be changes in emissions rather than impacts of meteorological changes. The dramatic increase in emissions after the 1940s, in Europe and North America, resulted in substantial changes in Nr and sulphur deposition and in O_3 levels. Reductions in emissions in Europe starting around the 1980s have resulted in significant reductions in sulphur and oxidised Nr compounds in the European atmosphere. Emissions of reduced Nr compounds have not declined to the same extent, and indeed in some areas, NH_3 emissions are increasing.

For the future development of air pollution in Europe, some climate-induced changes are potentially important, however. For example, potential increased shipping activity and new shipping routes within the Arctic may lead to increased nitrogen deposition in environmentally sensitive areas and even to increased phytotoxic ozone uptake. A new understanding is also that higher temperatures may increase NH_3 emissions from evaporative sources over land by very substantial amounts (e.g. 20–50 %), a process that is not yet included in NH_3 emission inventories. In summary, while it seems likely that air pollution impacts from sulphur and oxidised nitrogen will be substantially reduced in future compared to recent years, the situation for ozone and reduced nitrogen is still unclear, and very dependent on future emission control measures, both at regional and (for ozone) hemispheric scale.

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Environmental Impacts—Coastal Ecosystems, Birds and Forests

16

Pekka Niemelä, Harri Tolvanen, Mia Rönkä, Seppo Kellomäki, Joachim Krug, Guy Schurgers, Esa Lehikoinen, and Risto Kalliola

Abstract
This chapter describes observed historical and projected future impacts of climate change on the coastal and terrestrial ecosystems of the Baltic Sea basin. Because terrestrial and aquatic ecosystems interact, this chapter gives particular emphasis to the coastal zone as a contact area for terrestrial, marine and atmospheric processes. Archipelagos and post-glacial land uplift are particular features of the Baltic Sea basin and so receive special consideration. This chapter comprises three main sections. The first describes coastal zone and archipelago ecosystems in the Baltic Sea region and evaluates the potential impacts of climate change. The second examines a case study for the effect of current and future climate change on coastal bird populations and communities. The third evaluates the effects of current and future climate change on forests and natural plant communities in the Baltic Sea basin and the ways in which terrestrial ecosystems may interact with aquatic ecosystems. Climate-related changes in carbon storage are also discussed.

16.1 Introduction

This chapter describes observed historical and projected future impacts of climate change on the coastal and terrestrial ecosystems of the Baltic Sea basin. This region contains many types of ecosystem, such as boreal and hemiboreal forests, peatlands, arable and other human modified lands, freshwater and riparian ecosystems and diverse coastal environments. Terrestrial and aquatic ecosystems interact and so this chapter focuses particularly on the coastal zone as a contact area for terrestrial, marine and atmospheric processes. Archipelagos are a particular feature of the Baltic Sea basin and so receive special emphasis. Thus, the focus of this chapter is on ecosystems that received little or no attention in previous assessment of climate change in the Baltic Sea basin (BACC Author Team 2008).

A special feature of the Baltic Sea region is land uplift. Currently, the rate of land uplift in the Baltic Sea region is −1 to 9 mm year^{-1} (Leppäranta and Myrberg 2009; see also Chap. 9, Sect. 9.2). Although marine regression has been the prevailing process on the coasts, periods of marine transgression (shoreline advance) have also occurred (Björck 1995). Thus, the transitional zone between the terrestrial and marine realms is considered particularly changeable, which highlights the necessity to address the aquatic dynamics of

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_16

the Baltic Sea within the contexts of sea-level rise (Chaps. 9 and 14) and drainage basin processes. It is important to emphasise the combined effect of climate change impacts and post-glacial land uplift on coastal ecosystems, especially in the northern parts of the Baltic Sea basin.

The sensitivity to human-induced change is particularly evident in terms of the water volume of the Baltic Sea. Water volume divided by the number of inhabitants in the drainage basin equates to about 250,000 m^3 per capita. This is only 0.13 % of the value calculated for the World Ocean and the total world population (186,000,000 m^3 per capita). A strong human influence in the Baltic Sea catchment area also results from the high population pressure (80–90 million inhabitants) and because around 20 % of the overall land area is arable land concentrated towards the south (Sweitzer et al. 1996). Thus, the marine ecosystem is under considerable pressure from the land: both in abiotic terms (such as temperature change and pollutant inputs) and biotic terms (such as from invasive species).

This chapter focuses on ecosystems and processes that are unique and vulnerable to environmental change, including climate change. Section 16.2 describes coastal zone and archipelago ecosystems in the Baltic Sea region and evaluates the potential impacts of climate change on the interaction between aquatic and terrestrial ecosystems. Section 16.3 examines the effect of current and future climate change on coastal birds and bird communities. Section 16.4 evaluates the effects of current and future climate change on forests, natural vegetation and carbon storage in the Baltic Sea basin.

16.2 The Coastal Zone and Shorelines

16.2.1 Heterogeneity Implies Regionally Different Effects

Coastal areas play key roles in the interaction between terrestrial and aquatic systems. The coastal zone comprises the marine areas under terrestrial influence, as well as the terrestrial environment under marine influence (Tolvanen and Kalliola 2008). The ecosystems of the coastal areas are unique in combining both realms. Human occupancy and its impacts are exceptionally diverse in the often densely populated coastal areas.

The total shoreline of the Baltic Sea measures 76,000 km (HELCOM 2012). The distribution of shorelines of different complexity (Bartley et al. 2001) shows remarkable variety, from soft-formed depositional shores to the complex fractal shorelines of the bedrock-dominated archipelagos. *Shore density* (length of shoreline per unit area) provides a good indicator of shoreline complexity, which is highly variable in the Baltic Sea region (Fig. 16.1).

Shore density is not the only parameter with significant regional variation on the Baltic Sea coasts: the amount of nearshore areas also varies, both on the littoral and terrestrial sides (Fig. 16.1). The importance of *edge effects* becomes apparent when high shore density areas are considered as ecosystem patches—the nearshore ecosystems prevail in high shore density areas compared to the less complex

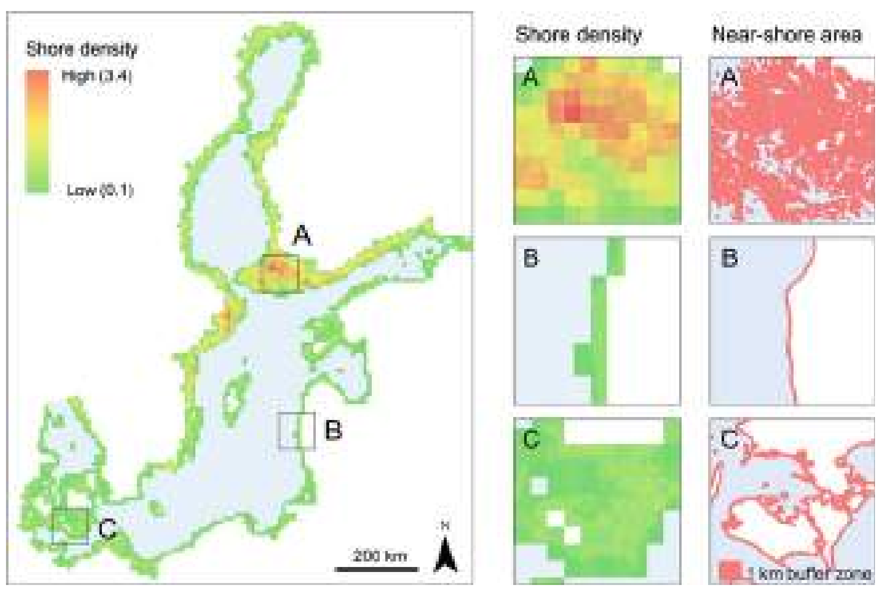

Fig. 16.1 Shore density in three 100 km^2 square cells on the Baltic Sea coast. The insets show detail concerning shore density and the nearshore zone as a buffer area extending 1 km from the shoreline. The nearshore zone covers 74 % in area *A*, 2 % in *B* and 19 % in *C*. Computed from HELCOM (2012)

coastal settings in other areas. Biodiversity of plant and animal communities is generally highest on the edges of different habitats because edges share species from both habitats. In addition, areas with different land–water transitions may respond differently to anticipated change, for example climate-related sea-level rise.

In areas with a wide coastal zone, from the mainland to the outermost islets, the transition from the terrestrial to the marine environment is often gradual. A wide coastal area may show distinct inner, middle and outer zones, with characteristic physical dimensions and ecological responses (e.g. Häyrén 1900; Granö 1981; Jaatinen 1984; Granö and Roto 1989b). This transition may be particularly sensitive to environmental change in the interface between the terrestrial and marine systems.

Post-glacial land uplift progresses at different rates in different Baltic Sea coastal areas. While uplift is almost 1 cm year^{-1} in the Bothnian Bay, the southernmost coasts of the Baltic Sea undergo slight submergence (Eronen 2005; Myrberg et al. 2006; see also Chap. 9). Figure 16.2 shows the proportion of coastal areas in each uplift category. These differences are crucially important in relation to projections of global sea-level rise (see Chap. 14). It is noteworthy that land uplift is also a process that creates particular coastal environments, such as closing bays (flad to glo-lake continuum (see Tolvanen et al. 2004)), and these processes work at different rates in different parts of the Baltic Sea.

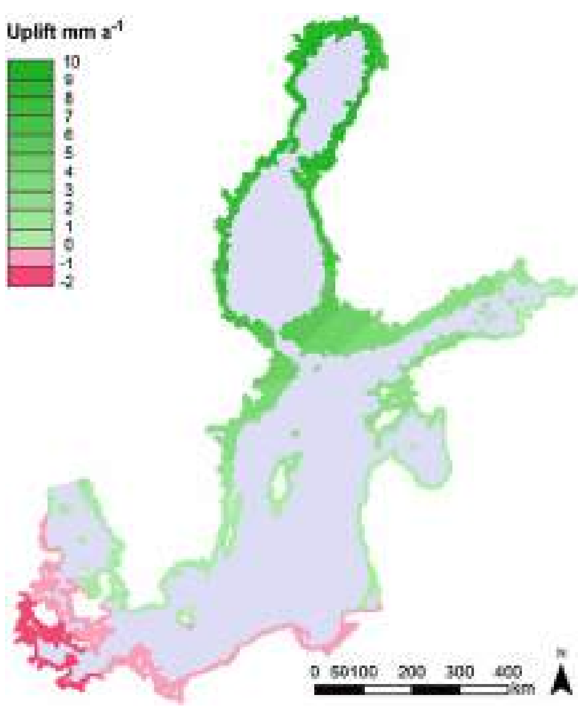

Fig. 16.2 Rate of post-glacial land uplift in 25 km^2 square cells along the coasts of the Baltic Sea (uplift rates from Voipio and Leinonen 1984)

There is a wide variety of *shore types* on the Baltic Sea coasts: chalk cliffs, barrier islands with coastal lagoons, sandy beaches, flat clay shores, rocky shores and esker shores (see also Chap. 20). All these environments are subject to particular erosional and depositional processes, which are mediated by different shore-forming forces such as wave energy. These drivers, in turn, vary in response to changing weather conditions making them sensitive to climate change. Furthermore, the prevailing processes in the shore areas are also affected by the post-glacial land uplift of the Fennoscandian crust. There are several ways to classify coastal areas (e.g. Bird 2000; Fairbridge 2004; Finkl 2004) and the Baltic Sea coasts have been classified by several authors (Furman et al. 1998; Frisén et al. 2005; HELCOM 2012; see also Chap. 20). HELCOM (2012) presented six coastal types based on the structural appearance of the coast:

- Archipelagos—the mainland shoreline does not face the open sea
- Bodden coasts with lagoons—low elevation coasts with dynamic sediment balances
- Cliffs—erosional shores of soft sedimentary rocks
- Fjords and rocky shores—hard rock environments with glacial erosion patterns
- Intended low coasts—low elevation coasts with marshes and skerries
- Open low coast—often sandy dynamic coasts.

Shore openness defines the shore's exposure to the open sea and so portrays the wave formation potential (Ekebom et al. 2003; Tolvanen and Suominen 2005; Murtojärvi et al. 2007). On coasts with low shore density, the shore openness is generally high, while on high shore density coasts, most of the shoreline is sheltered by surrounding islands.

Owing to its restricted connection to the World Ocean, the Baltic Sea is effectively non-tidal. There are, however, irregular water level fluctuations caused by changes in atmospheric pressure, precipitation, wind strength and standing wave oscillation (*seiche*) (Myrberg et al. 2006; Chap. 9). The combined effect of these factors may under some circumstances cause local sea-level fluctuations in large bays of up to 3.2 m in the northernmost part of the Bothnian Bay (Frisén et al. 2005). An increase in the magnitude and frequency of extreme weather events may increase the frequency of extreme water level events (i.e. surges) on the Baltic Sea coasts. During warmer than normal winters, storms can also cause local sea-level rise, which together with ice dams in rivers and seiche fluctuations may increase winter flooding of Baltic Sea shores.

The formation of sea ice influences Baltic Sea shores (see also Chap. 8). Ice usually forms first in shallow inland bays, extending gradually towards the open sea. Sea ice in contact with the shore can cause local erosion through tearing as well as deposition by ice push (e.g. Forbes and Taylor 1994). Annual sea ice dynamics vary from year to year depending

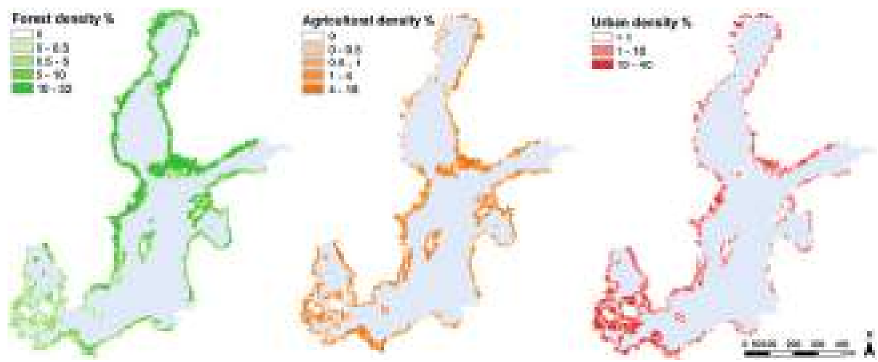

Fig. 16.3 Land use within 1 km of the shoreline (excluding the Russian Federation) in 100 km^2 square cells. Computed and generalised from the CORINE land use classification, EEA (2006)

on changes in winter temperature, storms and water level fluctuations and sometimes also on the weight of snow on ice (Chap. 8). These variations influence shore geomorphology, ecosystems and human livelihoods. If winters become warmer, it is likely that significant changes will occur in the ice conditions of the Baltic Sea (Omstedt and Nyberg 1997). As a consequence, sea ice interaction with the shores of the Baltic Sea may decrease in importance, and open water conditions with wave energy-induced dynamism may become more important (see also Chap. 20).

Human influences of terrestrial origin are diverse and induce various pressures on the coastal and marine systems. Eutrophication is especially sensitive to changes in atmospheric and hydrological processes, with consequent alterations in nutrient input and biological processes (HELCOM 2009). Maritime activities and shore protection also cause physical stress, which is particularly relevant for shallow coasts and archipelagos (e.g. Eriksson et al. 2004).

Land use varies greatly within the Baltic Sea basin (see also Chaps. 21 and 25). The most notable contrast is between the agricultural south and the forested north (Sweitzer et al. 1996). When only the immediate vicinity of the shoreline is considered, the north–south contrast is less striking (Fig. 16.3). The land-use patterns within 1 km of the shoreline reflect the high human occupancy of the coastal areas: urban areas are frequent, especially in the Scandinavian countries and Finland, where shoreline development and private land ownership on shores has been allowed to a greater degree than in the Baltic States and Poland.

16.2.2 Impacts of Climate Change on Coastal Areas

The Baltic Sea coasts show a large variety of environments on which climate change may have strong effects (Kont et al. 1997, 2003; Neumann and Friedland 2011; Störmer 2011) (Table 16.1). Some of the ecosystem responses are particular to the relatively simple-structured brackish water ecosystems of the Baltic Sea. Therefore, a comprehensive view of the regional geography and distribution of the different habitats within the drainage area is important when assessing the pattern of future change.

Table 16.1 Summary of potential climate change impacts on coastal areas of the Baltic Sea

Change	Consequences		Response
Atmospheric warming (see Chap. 4)	Warmer terrestrial ecosystems (e.g. Hickling et al. 2006)		Northward migration of terrestrial and aquatic species, longer productive season (e.g. Chapin et al. 2007; MacKenzie et al. 2007; Burrows et al. 2011)
	Warmer coastal sea water (e.g. Omstedt et al. 2004)		
	Decreased extent, thickness and duration of annual sea ice in coastal waters (e.g. Vihma and Haapala 2009)		Changes in marine ecosystems and physical features of the sea (see Chap. 19)
Potential increase in precipitation (e.g. Zolina et al. 2010)	Potentially increased terrestrial run-off	Decreased coastal water salinity	Changes in species composition and ecosystem function (see Chap. 19)
	Increased river-borne sediment, dissolved organic material and nutrient loads	Changes in water chemistry, decreased coastal water transparency	Eutrophication; shallower distribution of aquatic plants, decreased and fragmented benthic and littoral habitats (see Chap. 19)
Acceleration in global sea-level rise (e.g. Omstedt et al. 2004; Donner et al. 2012)	Decreased or reversed relative land uplift (e.g. Hammarklint 2009)	Shoreline retreat slowed or reversed (e.g. Johansson et al. 2004)	Changes in littoral ecosystems (see Chap. 19), need for coastal defence (see Chap. 20)
	Increased land submergence (e.g. Richter et al. 2012)	Accelerated shoreline advance (e.g. Johansson et al. 2004)	

16.2.3 The SW-Finnish Archipelago as an Example of a Particularly Sensitive Coastal Environment

The presence of extensive archipelago coasts is a particular feature of the Baltic Sea. In these areas, the interaction between the terrestrial and marine realms can be particularly complex. The Fennoscandian ice sheet of the Pleistocene glaciation retreated from SW Finland about 10,500 cal. year BP (Frisén et al. 2005; see also Chap. 2). The region is now characterised by a fragmented bedrock surface of the Fennoscandian Shield, with local relative elevations up to 100 m. As the bedrock base surface is slightly tilted to the west and is partially submerged, the result is an archipelago coast with an east–west transition from a land-dominated inner archipelago to a water-dominated skerry landscape at the edge of the open sea (Fig. 16.4). The width of this transition zone is up to 150 km of continuous archipelago area. On a 1:10,000 map, the SW-Finnish archipelago alone, excluding the Åland Islands, contains more than 56,000 islands. About 8500 of these are larger than 1 ha (Stock et al. 2010). Water depth is generally less than 20 m, but exceeds 100 m in areas of bedrock faults. The archipelago has about 15,000 km of shoreline within an area of roughly 10,000 km^2, giving an average shore density of 1.5 (Granö et al. 1999), ranging locally from 0 to 12.5 (Tolvanen and Suominen 2004).

The bedrock base is partially covered by till or fluvioglacial deposits and occasionally by marine sediments since the area was initially submerged after the deglaciation. Shore processes have affected these sediments, leaving the highest bedrock areas bare and accumulating sediments on the slopes and in nearby sea basins (Granö and Roto 1986). The current glacio-isostatic land uplift rate of 3–5 mm year^{-1} on the SW-Finnish coast (Suutarinen 1983; Kakkuri 1987; Vestøl 2006) has created a mosaic of bare bedrock and sedimentary surfaces along the shores, which are subject to primary succession.

The exceptionally high shore density is a result of abundant small-scale geomorphological details and gives rise to some special shore forms characteristic of the land uplift environment and habitats (Schwartz et al. 1989; Munsterhjelm 1997; Tolvanen et al. 2004). These land uplift habitats harbour specialised fauna and flora important for biodiversity (Vartiainen 1988). The physical conditions on the islands and shores are diverse and can differ notably between adjacent islands of similar size. The small-scale mosaic of islands and shallow water areas, with numerous small habitat patches, creates habitat edges and thus increases the biodiversity of the region (von Numers 1995; Boström et al. 2006). Shallow semi-enclosed bays often indicate intermediate characteristics between terrestrial and marine systems.

Salinity in the SW-Finnish archipelago ranges from 3.5 to 7.0 (Viitasalo et al. 1990). Annual ice cover can last for up to 100 days (Seinä and Peltola 1991), but ice duration is shortening (Haapala and Leppäranta 1997). Water temperature in the surface mixed layer reaches 20–25 °C in August. December is the windiest month with an average wind speed of 8.3 m s^{-1}, while May, June and July show the lowest average wind speed of 5.3 m s^{-1} (FMI 1991). The predominant wind directions during the ice-free season are southerly, south-westerly and westerly (Heino 1994).

Water currents in the SW-Finnish archipelago are typically slow (<15 cm s^{-1}) (Virtaustutkimuksen neuvottelukunta 1979), but faster currents occur occasionally in narrow straits. Flow direction and intensity vary depending on atmospheric pressure and wind patterns. Islands and underwater thresholds create small local basins, in which freshwater run-off mixes with offshore water, resulting in a dynamic mosaic of sea water of different origins and properties.

There are four-dimensional geographical patterns and seasonal dynamics in the water properties of the SW-Finnish archipelago. Importantly, many of these spatial or temporal patterns differ for the different sea water properties (turbidity, salinity, acidity and temperature) (Suominen et al. 2010a). For example, salinity shows considerable spatial, seasonal and interannual variability (Suominen et al. 2010b), challenging the physiological limits of many species living at the edge of their salinity tolerance. Eutrophication resulting from terrestrial run-off is a common environmental problem in the region (Lundberg et al. 2005; HELCOM 2009).

Penetration of solar radiation into the surface waters is strongly influenced by many optical constituents of terrestrial origin, such as suspended sediment and dissolved organic material, as well as by in situ phytoplankton populations (Kirk 2011). The inner archipelago is relatively turbid throughout the year due to suspended sediment from rivers and shallow clay bottoms. In the middle and outer

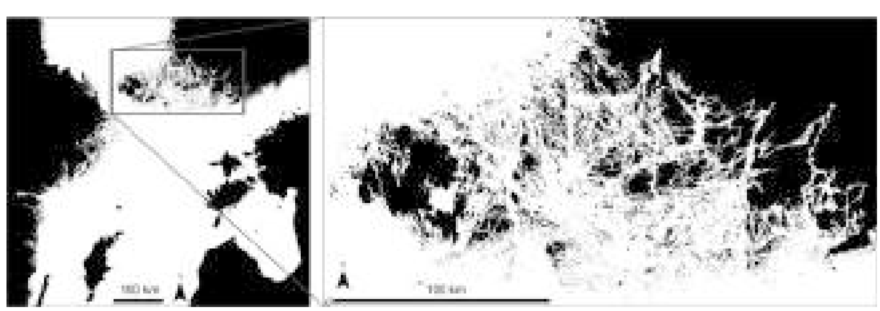

Fig. 16.4 The SW-Finnish archipelago, including the Åland Islands towards the west. Shoreline data from HELCOM (2012)

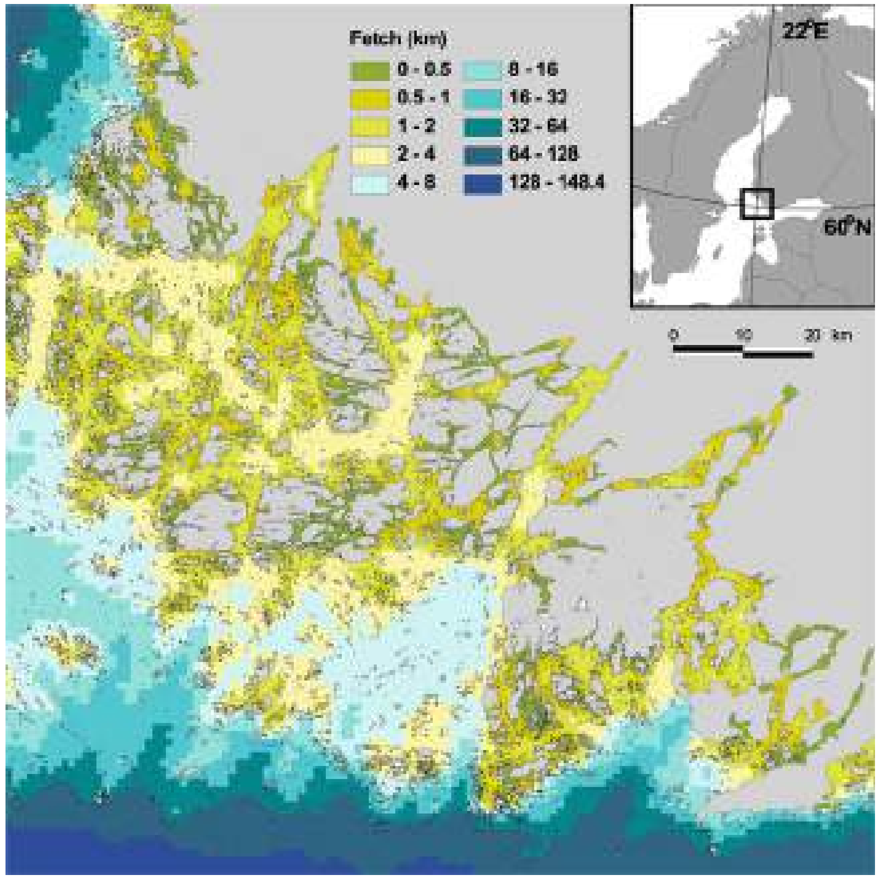

Fig. 16.5 Sea surface openness as average fetch in 1 km^2 square cells on the SW-Finnish archipelago coast, computed after Tolvanen and Suominen (2005)

archipelago areas, chlorophyll and humic compounds tend to regulate the underwater light field (Erkkilä and Kalliola 2004; Suominen et al. 2010a). Periods of high run-off and precipitation increase the freshwater input and local turbidity near river mouths.

In the complex archipelago areas, shore openness and the consequent shore sensitivity to storm damage create a mosaic of open and sheltered areas with a gradual transition towards the pelagic environment (Fig. 16.5). In addition to the zonal characteristics of the entire archipelago coast (Häyrén 1900; Granö 1981; Tolvanen and Suominen 2004), the abundance of littoral areas vulnerable to shore disturbance processes results in a variety of local patterns and characteristics. Due to continuous land uplift, the duration of exposure to shore processes for a particular site has been 400–800 years (Pyökäri 1986; Granö and Roto 1989a).

16.3 Climate Change Effects on Coastal Birds and Seabirds

Since the 1970s, the effects of climate change on birds have been analysed in a rapidly increasing body of literature (for a review, see Møller et al. 2010). According to the Millennium Ecosystem Assessment, climate change is likely to become one of the most significant drivers of biodiversity loss by the end of this century (Millennium Ecosystem Assessment 2005).

This section reviews the effects of climate change on coastal birds and seabirds in the Baltic Sea. Equally important are climate change effects on the distribution of land birds in the catchment area, but these effects are still to be consistently analysed. Discussion of land birds is therefore limited to a few general references and the chapter focuses on coastal birds and seabirds that rely on the habitats provided by the coasts, archipelagos and open sea areas of the Baltic Sea.

16.3.1 Coastal Birds and Seabirds as Components of Baltic Sea Ecosystems

Seabirds and coastal birds are an integral part of coastal and marine ecosystems and link into ecosystems at a number of trophic levels (Tasker and Reid 1997), usually as predators near the tops of food chains. Thus, gulls, ducks and waders can play important roles in the mass and energy fluxes of food webs, as well as in food web control (Moreira 1997; Eybert et al. 2003). Seabird nutrient transport between marine and terrestrial realms may shape entire ecosystems (Croll et al. 2005).

As an addition to their role in ecosystems, birds provide ecosystem services that benefit humans (Şekercioğlu et al. 2004; Millennium Ecosystem Assessment 2005; Şekercioğlu 2006). In Sweden, Rönnbäck et al. (2007) identified more than forty categories of goods and services provided by coastal ecosystems, including several services connected to coastal birds and seabirds.

16.3.2 Climate Change Effects

Climate change may impact coastal birds and seabirds of the Baltic Sea area by affecting water salinity, temperature and acidity, air temperature during breeding and non-breeding seasons, as well as rainfall and windiness. These environmental factors can affect birds directly by causing adult or juvenile mortality or indirectly by altering, for instance, the abundance and quality of food. Climate change also affects physical habitats through sea-level rise, storm events and erosion. Furthermore, climate change may interact with other environmental processes, such as eutrophication. These effects may lead to changes in bird population sizes and distributions during breeding and non-breeding seasons and thus alter the composition of bird communities.

16.3.2.1 Salinity

In the Baltic Sea, changes in salinity can have a cascading effect on food webs through the whole pelagic ecosystem (Hänninen et al. 2003; Wasmund et al. 2011). Some climate change scenarios project a decrease in Baltic Sea salinity, while others project an increase (BACC Author Team 2008; Chap. 13). A fall in salinity would lead to a decrease in marine species and an increase in freshwater species (Möllmann et al. 2005; MacKenzie et al. 2007), while an increase in salinity would cause the reverse. Changes in salinity would probably affect coastal birds and seabirds indirectly, mainly through changes in food availability. For instance, distribution and size of the blue mussel *Mytilus edulis* are highly dependent on salinity (Westerbom et al. 2008), whereas the Baltic clam *Macoma balthica* has a greater capacity to acclimatise to highly dilute brackish conditions (Jansen et al. 2009). As it is mediated by food availability, the relationship between salinity and bird populations may not become apparent immediately (Rönkä et al. 2005).

16.3.2.2 Water Temperature

The marine ecosystem is sensitive to changes in water temperature, which can affect the plankton (Dahlgren et al. 2011; Wasmund et al. 2011) as well as fish reproduction and survival (Margonski et al. 2010), and may thus have an impact on the food resources of coastal birds and seabirds. The rise in surface water temperature in the Baltic Sea over the past 100 years (Belkin 2009) may have had significant ecological consequences (MacKenzie and Schiedeck 2007; see also Chaps. 18 and 19).

16.3.2.3 Eutrophication and Oxygen Deficiency

Climatic factors can promote eutrophication by enhancing nutrient inputs through increased run-off and precipitation (Hänninen and Vuorinen 2011). Both eutrophication and the rise in water temperature have caused a reduction in the oxygen content of the Baltic Sea water, which has in turn resulted in negative impacts on some prey items, such as different life stages of the eastern Baltic cod *Gadus morhua* (Hinrichsen et al. 2011). While some coastal birds might initially benefit from eutrophication, at some point the effects can turn negative (Rönkä et al. 2005).

16.3.2.4 Food Web Structure

Changes in food webs may have severe effects on coastal birds and seabirds at the top of the food chain. Climate change has been shown to affect the populations of oceanic seabirds by diminishing their food resources (Montevecchi and Myers 1997; Barbraud and Weimerskirch 2003). In addition to food abundance, the quality of food items may change (Riou et al. 2011). The low-energy value of fish was considered the cause of a major breeding failure of common guillemots *Uria aalge* in the North Sea (Wanless et al. 2005). Similarly, Österblom et al. (2001) found a long-term decline in the individual mass of common guillemot *Uria aalge* chicks when leaving a colony in the Baltic Sea that coincided with a decline in the condition of the sprat *Sprattus sprattus*. However, on the Norwegian coast, milder winters have been found to increase the availability of fish prey for the lesser black-backed gull *Larus fuscus fuscus* (Bustnes et al. 2010).

Climate change may also alter predation pressure on birds by facilitating invasions of mammalian predators. In the Baltic Sea area, non-indigenous mammalian predators that may prey on coastal birds and seabirds are mainly the American mink *Mustela vison* and the raccoon dog *Nyctereutes procyonoides*. The raccoon dog probably benefits from climate warming (Melis et al. 2010), which may allow it to increase and expand its distribution northwards (Helle and Kauhala 1991; Melis et al. 2010). However, evidence of the raccoon dog's negative impacts on bird populations is still scarce (Kauhala and Kowalczyk 2011). The American mink in turn reduces the breeding densities of several birds nesting on small islands in the Finnish archipelago (Nordström et al. 2002, 2003), but there are few assessments of the possible effects of climate change on its abundance and distribution. In Iceland, climate change events may have contributed to a reduction in the mink population (Magnusdottir 2012) and an American study has shown its sensitivity to hydrological shifts resulting from climate change (Schooley et al. 2012).

16.3.2.5 Weather Conditions

Climate change may affect the breeding performance of seabirds by altering weather conditions during breeding or by affecting the condition of the birds after winter (Hildén 1964; Milne 1976; Lehikoinen et al. 2006). Weather, especially temperature, rainfall and wind, is important for the breeding success of common eider *Somateria mollissima*, velvet scoter *Melanitta fusca*, mute swan *Cygnus olor* and tufted duck *Aythya fuligula* (Koskimies 1955; Hildén 1964; Koskimies and Lahti 1964; Koskinen et al. 2003). If climate change increases the frequency of extreme weather events, there may be strong effects on reproductive success.

High temperatures during the breeding season may cause heat stress in adults, particularly in species that make long foraging trips, forcing the other parent to spend extensive periods in continuous nest attendance (Oswald et al. 2011). Climate change can also affect breeding success through complex predator–prey interactions, for instance by changing the availability of prey other than birds (Hario et al. 2009). However, even a severe crash in fledgling production need not affect the size of the future local breeding population (Rönkä et al. 2005).

Severe winters have been suggested to influence populations of the great crested grebe *Podiceps cristatus* (von Haartman 1945), mute swan (Koskinen et al. 2003; Rönkä et al. 2005), coot *Fulica atra* (von Haartman 1945, Rönkä et al. 2005), tufted duck (Hildén 1966, Hildén and Hario 1993), mallard *Anas platyrhynchos*, common eider and common goldeneye *Bucephala clangula* (Rönkä et al. 2005). Furthermore, high winter mortality has been assumed to be the main reason for the decline in razorbill *Alca torda* and greater scaup *Aythya marila* during the Second World War (Hildén and Hario 1993). The effects of cold winters can be additive, and it may take years for a species to recover (Rönkä et al. 2005). Winter severity may also affect the density-dependence of survival (Barbraud and Weimerskirch 2003).

16.3.2.6 Sea Ice

A climate-induced decrease in the ice coverage of the Baltic Sea would improve wintering conditions for Baltic Sea coastal birds and seabirds by alleviating competition for food. In addition, wintering birds would not congregate in areas of intense shipping. This would reduce their vulnerability to oil hazards, which are an important mortality factor for the long-tailed duck *Clangula hyemalis* (Hario et al. 2009). However, the net effect of climate change on the risk from oil pollution is unclear; while a decrease in ice cover might reduce the risk from oil pollution, an increase in windiness and extreme weather events might enhance it.

16.3.2.7 Migration

Many Baltic Sea coastal birds and seabirds migrate only as far as the western or southern Baltic Sea or the North Sea (Cramp and Simmons 1977; Pihl et al. 1995; Gilissen et al. 2002). A minor component of the Finnish and Swedish populations of some coastal birds remains in the northern Baltic Sea (Gilissen et al. 2002). Winter severity in the Baltic Sea reflects winter severity in the North Sea and further off the coast of western Europe, as well as in central Europe (Hurrell 1995). Winter severity in western Europe affects the non-breeding survival of several coastal birds and seabirds (Nilsson 1984; Koskinen et al. 2003). In addition to mortality from starvation and cold, severe winters may force birds to migrate further than normal and this incurs extra energy costs. Furthermore, when the Baltic Sea is largely ice-covered, seabirds are forced to feed in small open areas which increases competition for food and possibly the risk of disease (Grenquist 1965; Hario et al. 1995).

Climate change affects the arrival and departure times of migrants (Forchhammer et al. 2002; Jonzén et al. 2002; Hüppop and Hüppop 2003; Lehikoinen et al. 2004; Lehikoinen and Sparks 2010; Lehikoinen and Jaatinen 2011) and the timing of breeding (Forchhammer et al. 1998; Both and Visser 2001; Møller 2002; Sanz 2002). A mismatch between hatching time and resource availability has been observed in passerines (e.g. pied flycatcher *Ficedula hypoleuca*, Both et al. 2006) as well as in ducks (Oja and Pöysä 2007). Trophic mismatches may cause population declines particularly in long-distance migrants in seasonal habitats (Both et al. 2010). In addition to trophic constraints, long-distance migrants may suffer from increasing competitive pressure by residents and short-distance migrants along with climate change, contributing to population declines in long-distance migrants (Lemoine and Böhning-Gaese 2003).

Climate change may also drive changes in migratory routes, stopover sites and migratory tendencies within species and populations. For instance, delayed departure from breeding areas may be the cause of recently observed northward shifts in wintering ducks (Lehikoinen and Jaatinen 2011). Milder winters may allow birds to winter closer to breeding grounds and thus contribute to higher survival, which has been proposed for the mallard (Gunnarsson et al. 2012). However, temperature may be less important in shaping the wintering distributions of European dabbling ducks than factors such as feeding ecology (Dalby et al. 2013).

Species ranges are expected to move poleward with climate change (Thomas and Lennon 1999; Hickling et al. 2006), and this has already been shown for the breeding ranges of central European and Arctic birds (Brommer et al. 2012). More study species (41; 69 %) shifted their range margin northwards than southwards (18; 31 %) (X^2, p = 0.00275). Only a few range margin shifts were greater than three grid squares. An example of a species with a large shift in the southern border northward is the ruff *Philomachus pugnax*. Large shifts in the northern border occurred in coot, mute swan, common eider, razorbill and Arctic skua

Stercorarius parasiticus suggesting range extension as the southern border did not move. The latter three species have extended their range much further north on the Atlantic Ocean and North Sea coasts. It is still unclear if climate change is involved in the range shifts and changes in the area of occupancy of the study species, since at least some of the shifts seem to have other causes.

16.3.2.8 Sea Level

Climate change may also affect the physical and biological features of islands through sea-level rise, storm events and erosion. By 2100, many regions currently experiencing a fall in relative sea level can be expected instead to experience a rise (Chap. 14), reversing current land uplift in Baltic Sea archipelagos. Together with storm events and erosion, sea-level rise could affect the succession in coastal bird and seabird breeding habitats. In addition, climate factors affect the productivity of the vegetation cover, which in turn may influence species richness and distribution, as shown by Eronen et al. (2011) for Finnish land birds.

16.3.2.9 New Species

Climate change may facilitate the settlement of new species that are either spreading to the Baltic Sea area naturally or that have been deliberately or accidentally introduced by humans. The effects of climate change on salinity and water temperature can facilitate invasion by non-indigenous aquatic species, which can cause major changes in nearshore ecosystems (Zaiko et al. 2011) and thus affect the food resources of birds. Climate change may also increase the survival and reproductive prospects of vagrant birds and increase the numbers of, for example, vagrant Siberian birds in Europe (Jiguet and Barbet-Massin 2013).

16.3.2.10 Changes in Distribution

Changes in the occurrence and distribution of bird species can be assessed using bird atlas data, such as the Finnish Bird Atlas data (Väisänen et al. 1998; Valkama et al. 2011). Changes in distribution of Finnish coastal and seabird species were studied by comparing data in the first and second Finnish Bird Atlases (1974–1989) (Väisänen et al. 1998) with those in the third (2006–2010) (Valkama et al. 2011). The data presented here cover 59 species in archipelago and coastal 10 km^2 square cells, and to avoid bias from differences in research activity, only those cells where the research activity grade was the same in both data sets were compared.

According to a preliminary analysis (Lehikoinen E unpubl.), the area of occupancy (for a description of this concept see Rassi et al. 2010, p. 31) had contracted for 10 (17 %) study species and expanded for 12 (20 %), while 37 (63 %) showed no significant change. Species with the largest contractions were the ruff, common pochard *Aythya ferina*, greater scaup and northern pintail *Anas acuta*. Species with the largest expansions were the great cormorant *Phalacrocorax carbo*, barnacle goose *Branta leucopsis*, whooper swan *Cygnus cygnus*, and Canada goose *Branta canadensis*. Of the species with the largest expansions, the great cormorant settled on the Finnish coast in 1996, the barnacle goose in 1981 and the Canada goose in the 1970s (Valkama et al. 2011).

The observed changes in species distribution and density (e.g. Virkkala and Rajasärkkä 2011) can be contrasted with the projections of European species range shifts by Huntley et al. (2008) using a model based on climate-envelopes. The authors predicted that with the current rate of climate change, bird communities in the Baltic Sea area (including all species) may be among the richest in Europe by 2100. Baltic Sea countries may experience a loss of some species which would in turn be replaced by particularly southern species. The predictions infer major changes in coastal bird and seabird communities in particular.

16.3.3 Implications for Seabird Management and Conservation

The effects of climate change on the breeding and feeding ecology of coastal birds and seabirds may first become visible near the limits of their ranges (Barrett and Krasnov 1996; Montevecchi and Myers 1997). Climate change effects might become discernible relatively early in the ecosystems of the Baltic Sea that can be considered marginal because of its distinctive features.

As integral parts of marine ecosystems, seabirds and coastal birds may fill some of the gaps in knowledge about marine and coastal ecosystems under stress. They may even act as indicators of climate change (Furness and Camphuysen 1997; Rönkä et al. 2005), providing early warnings of unforeseen environmental impacts and means to monitor changes at lower trophic levels. Habitat changes, as well as density and distribution shifts should be taken into consideration in species and habitat protection, including the management of currently protected areas and the planning of future conservation efforts (Virkkala and Rajasärkkä 2011). As climate change is already rapidly reshaping species distributions, ignoring future dynamics could lead to misguided and potentially ineffective conservation decisions (Kujala 2012).

Effective measures for the management and conservation of coastal birds and seabirds and their habitats require insight into their population processes and the factors affecting their distribution and abundance. A coherent monitoring system is needed, addressing population size, reproductive success and mortality (Järvinen 1983; Kilpi 1985; O'Connor 1985; Tiainen 1985; Elmberg et al. 2006; Sutherland 2006; Rönkä et al. 2011). Background data on environmental factors should also be collected.

Owing to the uncertainties concerning climate change scenarios (see Chaps. 10–14), assessing possible effects on coastal birds and seabirds is challenging. The multiple and partly contradictory effects and possible thresholds in the relationship between environmental drivers and ecosystem change add to the complexity. In addition, differences in the ecology of species must be considered (Rönkä 2008). Due to the many environmental pressures on the Baltic Sea, more studies are needed on the relative effects and possible interactions of different environmental changes, such as climate change and eutrophication (Kotta et al. 2009; Põllumae et al. 2009). In addition, a flyway approach is needed as some effects of climate change may act in breeding or wintering areas far from the Baltic Sea (Hario et al. 2009).

16.4 Climate Change Effects on Forests and Natural Vegetation

This section describes the effects of climate change on trees and forests in general—impacts on agriculture and managed forestry are addressed in Chap. 21.

16.4.1 Effects on Forest Growth

The overriding impacts of climate change on forest growth within the Baltic Sea basin, after the greater risks of extreme weather events such as prolonged drought, storms and flood (Lindner et al. 2010), are considered to be changing patterns of precipitation, increased carbon dioxide (CO_2) concentrations and rising temperature. The latter may drive an increase in evapotranspiration and, potentially, increased microbial activity which could cause more rapid decomposition of soil organic matter (Magnani et al. 2007). Increased microbial activity would lead to mineralisation and related fertilising effects which would be especially important at previously nutrient-deficient forest sites. Nitrogen emissions from human activities are already having potential fertilising effects so it is difficult to isolate improved nitrogen availability due to higher temperatures. Nevertheless, the end result is expected to be increased tree growth in forests, with the impact greatest at high latitudes and on sites of historical nutrient deficiency (Burschel and Huss 1997; Dengler 1992). Although understanding of nutrient dynamics at high latitudes is limited (Gundale et al. 2011), it seems likely that rising temperature is the major driver (Karjalainen 1996a).

Schulze et al. (2009) stated that at any one moment, the rate of carbon uptake at a particular site would depend on the age composition and density of the stand. They also stated that forest management influences growth rates by controlling species composition and stand density and that management practices have changed over recent decades. Additional factors that might influence forest growth rates are increased temperature and CO_2 concentration or nitrogen deposition from the atmosphere. According to Körner et al. (2005), increased CO_2 concentration alone would not cause increased growth, which assumes that other limiting factors such as nutrient deficiency may dominate. Similarly, Magnani et al. (2007) concluded that forest net carbon sequestration is overwhelmingly driven by nitrogen deposition, largely from human activities. Nevertheless, de Vries et al. (2009) stressed the high uncertainties related to carbon sequestration per unit weight nitrogen addition.

Pussinen et al. (2009) saw a meaningful increase in growth rates in forests in Scandinavian countries and Finland of up to 75 % in their models compared to a scenario without climate change. In more detail, Karjalainen (1996a) described a model-estimated (gap-type forest model interfaced with a wood product model) increase in carbon sequestration in Scots pine *Pinus sylvestris*, Norway spruce *Picea abies* and silver birch *Betula pendula* stands over 150 years on medium fertile sites in northern and southern Finland (Table 16.2). It is clear that in relative terms, boreal forest stands benefit more from climate change than temperate forest stands. Similar results were found by Alam et al. (2008) who showed, using an ecosystem model, that climate change may substantially increase the growth, timber production and carbon stocks of forest stands. They found the greatest relative change in northern Finland (north of 64°N) with growth increasing by 75–78 %, timber production by 59–70 % and carbon stocks by 21–23 %. Increases in southern Finland (south of 64°N) were lower at 37–45, 23–40 and 8–10 %, respectively, although the absolute (mean) values were higher.

Table 16.2 Projected changes in net carbon sequestration (150-year average value, Mg C $ha^{-1}year^{-1}$) due to climate change over the period 1990–2140 (Karjalainen 1996a)

Species	Northern Finland		Southern Finland	
	Current climate	Future climate	Current climate	Future climate
Scots pine	1.10	1.42 (+29 %)	0.78	0.84 (+8 %)
Norway spruce	0.69	0.99 (+44 %)	0.96	0.32 (−67 %)
Silver birch and Pubescent birch	0.43	0.60 (+40 %)	0.64	0.92 (+44 %)

Nabuurs et al. (2008) and the first assessment of climate change in the Baltic Sea area (BACC Author Team 2008) reviewed the sensitivity of the main Baltic tree species to climate change. The BACC assessment concluded that coniferous species (Scots pine and Norway spruce) respond positively to rising temperature by showing higher stem growth. This is especially the case for the higher latitudes where temperature and nutrient deficiencies are limiting factors, and in general where precipitation is not limiting (Mäkipää et al. 1998; Bergh et al. 2003; Lindner et al. 2010). However, in the temperate forest zone (in the southern part of the Baltic Sea basin), a potential fertilisation effect by increased CO_2 concentrations and enhanced nitrogen availability seems to be dominated by water stress, although the effect of elevated CO_2 concentrations may counteract such potential negative effects (BACC Author Team 2008). A stronger potential fertilisation effect of increased CO_2 concentrations can be expected in northern latitudes, where increased precipitation is projected.

Birch (*Betula* spp.), the main deciduous species in the boreal zone responds to rising temperature in the same manner as spruce and pine, but is less sensitive to changes in precipitation. In contrast, oak (*Quercus* spp.) and common beech (*Fagus sylvatica*) show a generally strong response to precipitation changes in temperate zones. While the growth of beech generally responds negatively to increased temperature, oak shows a weak positive response (BACC Author Team 2008).

Karjalainen (1996b) reported similar conclusions, reporting increased forest ecosystem production under a changing climate in northern Finland and a decline in Norway spruce in southern Finland. The carbon sequestration potential of Norway spruce declined sharply in southern Finland. This corresponds to the results of the first BACC assessment (BACC Author Team 2008) and of Kellomäki and Kolström (1994) and reflects the major impact of water stress (compare Bergh et al. 2003; Lindner et al. 2010).

Similar results were reported by Köhl et al. (2010) for temperate zones in north-eastern Germany: the authors found severe challenges to tree species sensitive to water stress (i.e. in this study, Norway spruce) with model simulations over a 100-year period projecting a decline in those species in north-eastern Germany for all scenarios.

With a focus on changing patterns of precipitation and temperature, Lasch et al. (2002) reported the likely future effects of climate change on forests in the Federal State of Brandenburg (Germany)—which can be used as a proxy for the south-westernmost part of the Baltic Sea basin. The results suggested that warming would lead to a shift in the natural species composition towards more drought tolerant species, that the diversity of the forests would decline and that groundwater recharge would decrease.

16.4.2 Effects on Natural Vegetation

To investigate the possible impacts of climate change on natural vegetation, studies were undertaken using the IPCC SRES scenarios for the twenty-first century. The studies indicated that species could shift, with a tendency for species to shift northwards and eastwards in the Baltic Sea basin (Ohlemüller et al. 2006; Wolf et al. 2007). Hickler et al. (2012) used a vegetation model to project changes in Europe's natural vegetation and showed that a considerable part of the Baltic Sea basin could undergo a change in species composition, with a predominant northward shift in the hemiboreal and temperate mixed forests. Arctic and alpine regions showed particularly large changes. Similarly, Thuiller et al. (2005) reported high sensitivity for plant diversity in the Arctic and mountainous regions. Species loss was projected to be greater in the southern part of the Baltic Sea basin than the northern, but the region has high potential for species to migrate in from the south (Thuiller et al. 2005). However, a study by Chytrý et al. (2012) which separated different land-use classes showed that forests were generally least sensitive to plant species migration.

16.4.3 Effects on Carbon Storage

For managed forests, the Baltic Sea basin has been projected to see a rise in annual increment in carbon sequestration for all SRES scenarios (Nabuurs et al. 2002; Eggers et al. 2008). This would cause the relatively low net ecosystem productivity of forestry in the region (Luyssaert et al. 2010; Bellassen et al. 2011) to approach that of central and southern Europe.

In general, terrestrial carbon storage is anticipated to increase in the Baltic Sea catchment (Wolf et al. 2007; Zaehle et al. 2007; Roderfeld et al. 2008) although higher temperatures have the potential to increase respiration and thereby loss of carbon (Piao et al. 2008). The greatest uncertainties for the carbon balance concern interactions with nitrogen (Churkina et al. 2010), although reduced nitrogen deposition might be offset by more favourable climatic conditions for plant growth in the future (de Vries and Posch 2011).

Land-use change can play an important role in the terrestrial carbon cycle (Zaehle et al. 2007) and can have both positive and negative impacts on carbon storage.

16.5 Conclusion

This chapter describes the impacts of climate change on the coastal and terrestrial ecosystems of the Baltic Sea basin. The presence of fragmented and geomorphologically complex archipelago coasts is a particular feature of the central and northern part of the Baltic Sea. Archipelagos have high value for biodiversity due to the 'edge effect' and to the high number of different habitats. On the other hand, archipelago ecosystems are vulnerable to climate-mediated changes in the environment. A special characteristic of the Baltic Sea region is post-glacial land uplift. Together with climatic change, land uplift creates a complex situation where plant and animal communities must adapt to a new environment modified by both of these factors. The combined effects of climate change and land uplift on coastal ecosystems have been little studied and need particular emphasis in the future.

Warmer terrestrial ecosystems and warmer coastal sea water affect the northward migration of terrestrial and aquatic species and result in longer reproductive periods for coastal fauna and flora. The biodiversity of the Baltic Sea is particularly sensitive to changes in salinity, which can have a cascading effect on food webs and interaction between aquatic and terrestrial ecosystems. The effects of climate change on salinity and water temperature can facilitate invasion by non-indigenous aquatic bird species, such as cormorants, which can cause major changes in coastal bird communities. The climate-mediated changes can also facilitate the invasion of mammalian predators which can cause major changes in coastal and archipelago ecosystems.

The positive impacts of climate change on forest growth will continue. In relative terms, boreal forest stands benefit more from the climate change than temperate forest stands.

The Baltic Sea drainage basin is expected to undergo a change in the species composition of natural vegetation, with a predominantly northward shift of the hemiboreal and temperate mixed forests. Projected losses of species are greater in the southern part of the Baltic Sea basin than in the north.

Terrestrial carbon storage is likely to increase in the Baltic Sea catchment area. However, land-use change can play an important role in the terrestrial carbon cycle and have both positive and negative impacts on carbon storage.

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Environmental Impacts—Freshwater Biogeochemistry

17

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Abstract

Climate change effects on freshwater biogeochemistry and riverine loads of biogenic elements to the Baltic Sea are not straight forward and are difficult to distinguish from other human drivers such as atmospheric deposition, forest and wetland management, eutrophication and hydrological alterations. Eutrophication is by far the most well-known factor affecting the biogeochemistry of the receiving waters in the various sub-basins of the Baltic Sea. However, the present literature review reveals that climate change is a compounding factor for all major drivers of freshwater biogeochemistry discussed here, although evidence is still often based on short-term and/or small-scale studies.

17.1 Introduction

The aim of this chapter is to summarise current knowledge on freshwater biogeochemistry within the Baltic Sea with a focus on riverine nutrient and carbon fluxes in the Baltic Sea catchment. Wherever possible, the chapter outlines current knowledge on the effect of climate change and its interplay with the relevant human drivers of change in freshwater biogeochemistry. Quantifying the effects of individual drivers is a challenging task. The chapter covers the background sources and loads of biogenic elements as well as the additional human sources and loads, transformation of dissolved, particulate and gaseous constituents along the aquatic continuum, and possible coeffects of human and climate drivers on sources and transformation. The current and possible future export patterns of biogenic elements to the Baltic Sea are also addressed.

17.2 Overview of Freshwater Biogeochemistry and the Baltic Sea Catchment

17.2.1 The Baltic Sea Catchment

The Baltic Sea drainage basin comprises a northern boreal part that drains into the Gulf of Bothnia (Bothnian Bay and

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_17

Bothnian Sea) and a south-eastern part that drains into the southern basins of the Baltic Sea (Baltic Proper, Gulf of Finland, Gulf of Riga, Danish Sounds, Kattegat), see also Chap. 5.

The northern watersheds, draining into the Gulf of Bothnia, are generally sparsely populated (Table 17.1) and less eutrophic than the cultivated watersheds of the south-east. The dominant land cover in the north is boreal forest and wetlands. Bedrock is dominated by acid volcanic and plutonic acid rocks (mainly granites); soil types are dominated by till (Durr et al. 2005). The mean slope of the western Swedish boreal watersheds draining mainly Sweden is much steeper, and the specific run-off (i.e. the water volume per unit area and time) is roughly double that of the eastern Finnish watersheds (Table 17.1).

The watersheds of the southern and eastern parts of the catchment are dominated by agriculture (Table 17.1; Fig. 17.1). Sedimentary rocks dominate the south-eastern part of the Baltic Sea catchment, whereas non- to semi-consolidated sedimentary rocks dominate the watersheds of the Oder and Vistula and consolidated sedimentary rocks dominate the watershed of the Nemunas and Daugava (Durr et al. 2005). Most rivers are lowland rivers with a mean slope of less than 1°. River nutrient loads, especially from the Neva, Oder, Vistula, Daugava and Nemunas rivers, contribute most to riverine mass fluxes to the central and southern basins of the Baltic Sea (Stålnacke et al. 1999a). The greatest nutrient mass fluxes come from the rivers Oder and Vistula, draining Poland and with its 38 million inhabitants, about half the population of the entire Baltic Sea catchment (Hannerz and Destouni 2006). Specific discharge is much less compared to the boreal watersheds.

17.2.2 Changes Shaping the Baltic Sea

The Baltic Sea is an estuarine system with water residence times of around 30 years and is highly susceptible to changes in riverine loads of biogenic elements (carbon, C; nitrogen, N; phosphorus, P; silicon, Si) (Wulff et al. 1990; Humborg et al. 2007; Meier 2007; Conley et al. 2009; Eilola et al. 2009). In the central parts of the Baltic Sea, a mean salinity of 7 corresponds to about 80 % freshwater and 20 % marine water from the Atlantic. Two major drivers—human lifestyles and global warming—which strongly influence agricultural practices and eutrophication and hydrological patterns for example, could significantly alter the transport of biogenic elements to the Baltic Sea over the near term (Arheimer et al. 2005; Hägg et al. 2010). These changes could be potentially more significant than the variations in

Table 17.1 Major catchment characteristics of the main Baltic Sea sub-basins: population statistics (2005), land cover (2000) and hydrological properties (average for 1980–2006)

	Bothnian Bay	Bothnian Sea	Baltic Proper	Danish Straits	Gulf of Finland	Gulf of Riga	Kattegat
Area (ha)	27,029,891	23,023,056	57,336,809	2,735,671	41,897,960	13,617,922	9,008,076
Total population	1,356,575	2,639,277	54,100,727	4,528,719	10,525,186	3,794,896	3,321,259
Urban population	468,890	771,649	18,932,166	621,633	3,199,319	1,252,689	609,971
Rural population	887,685	1,867,628	35,168,561	3,907,086	7,325,865	2,542,207	2,711,288
Run-off (km^3 $year^{-1}$)	105.471	96.733	110.105	6.522	114.406	33.405	33.866
Precipitation (mm $year^{-1}$)	528.8	586.8	557.2	635.9	644.7	647.0	707.3
Run-off ratio	0.7379	0.7160	0.3446	0.3749	0.4235	0.3791	0.5315
Temperature (°C)	0.23	2.80	7.67	8.84	4.29	6.17	6.21
Deciduous forest (%)	4.24	3.34	4.87	4.81	10.07	15.59	2.67
Coniferous forest (%)	30.97	47.56	21.52	2.74	28.54	14.53	44.0
Mixed forest (%)	17.89	7.91	7.75	2.9	28.43	19.36	3.9
Shrub and herbaceous (%)	21.89	19.20	5.59	1.02	8.63	16.6	5.83
Inland wetlands (%)	11.59	6.02	0.52	0.55	2.53	1.93	2.46
Maritime wetlands (%)	0.01	0.02	0	0.21	0.01	0	0.09
Cultivated areas (%)	3.5	6.76	54.11	77.14	6.9	29.63	26.38
Bare areas (%)	3.37	0.9	0.12	0.03	0.01	0.02	0.18
Inland water (%)	5.68	6.86	2.46	1.45	13.96	1.37	11.58
Snow and ice (%)	0.15	0	0	0	0	0	0
Artificial surfaces (%)	0.6	1.04	3.01	8.76	0.78	0.94	2.88

The run-off ratio gives the share of the annual precipitation discharged by rivers to the Baltic Sea. From the Baltic Environmental Database (www.balticnest.org/bed)

Fig. 17.1 Landcover in the Baltic Sea drainage basin

riverine fluxes observed over the past 35 years (HELCOM 2004), since changes in lifestyle translate directly into anthropogenic nutrient emissions and riverine fluxes (Howarth et al. 1996; Hägg et al. 2010) and the projected changes in temperature and precipitation are expected to result in fundamental changes within the Baltic Sea catchment (Graham 1999; Graham and Bergstrom 2001; Weyhenmeyer and Karlsson 2009). To date, observations on the discharge regime of major Finnish boreal rivers reveal no changes in mean annual flow during the period 1912–2004 (Korhonen and Kuusisto 2010), but the seasonal distribution of streamflow has changed (see also Chap. 5). Winter and spring mean monthly discharge increased at most observation sites, and the spring peak has become earlier at a third of sites (Korhonen and Kuusisto 2010). However, additional drivers such as atmospheric deposition (Monteith et al. 2007; Weyhenmeyer 2008), management of forestry and wetlands as well as damming and other types of hydrological alteration (Dynesius and Nilsson 1994; Nilsson et al. 2005; Humborg et al. 2006, 2008a) are compounding factors affecting freshwater biogeochemistry, especially in the boreal watersheds.

17.2.3 Drivers of Change in Sources, Transformation and Export of Biogenic Elements to the Baltic Sea

Freshwater biogeochemistry in relatively unperturbed aquatic systems within the Baltic Sea catchment and the background load of the biogenic elements C, N, P and Si is the result of its weathering regime, which characterises total ionic strength, acidity (pH) and alkalinity as well as vegetation cover and vegetation type. Generally, weathering reactions charge rainwater with basic cations and anions including dissolved inorganic carbon (DIC), orthophosphate and silicic acid when infiltrating natural soils (Drever 1997). Nitrogen enters the system through biological N fixation, and organic C stems from recently produced biomass (mainly litter and root exudates) (Froberg et al. 2003; Karltun et al. 2005; van Hees et al. 2005; Giesler et al. 2007; Jonsson et al. 2007) and older stored soil organic C (Vonk et al. 2008; Vonk and Gustafsson 2009). However, the bedrock dominated by acid volcanic and plutonic acid rocks as well as the occurrence of coniferous forests and wetlands storing huge amounts of organic C leads to freshwaters in boreal watersheds characterised by low ionic strength, low alkalinity and high concentrations of humic and fulvic acids that form the major pool of dissolved organic carbon (DOC), nitrogen (DON) and phosphorus (DOP). Background loads and concentrations in the cultivated watersheds are difficult to estimate because humans have influenced these landscapes over many centuries. However, the occurrence of sedimentary bedrock in cultivated watersheds and the higher temperatures that increase weathering reactions lead to a higher ionic strength and higher alkalinity.

Relatively natural unperturbed conditions can be found in the well-studied River Kalixälven (Ingri et al. 1997; Humborg et al. 2004), in the Simojoki river basin (Lepisto et al. 2008) and in unmanaged headwater catchments (Mattsson et al. 2003; Finer et al. 2004; Kortelainen et al. 2006a). In most other watersheds around the Baltic Sea, freshwater biogeochemistry is affected by following human drivers: atmospheric deposition, forestry and wetland management, eutrophication, and damming and other types of hydrological alteration.

17.2.3.1 Human Drivers

Atmospheric deposition. Atmospheric deposition of acids, metals and nutrients peaked in the 1970s and 1980s with the notorious acidification effects observed in lakes and streams of Sweden and Finland (Weyhenmeyer 2008). The effects of atmospheric deposition and acidification are generally more significant in the northern boreal part of the Baltic Sea catchment than in the southern cultivated areas, because of the lower buffer capacity of the surface waters and soils. Moreover, atmospheric deposition is also more significant compared to other drivers in relatively unperturbed watersheds with low population densities. For more detailed discussion, see Chap. 15.

Forestry and wetland management. Forests and wetlands are the dominant landscape forms in the northern boreal part of the Baltic Sea catchment and cause the high concentrations of dissolved organic matter (DOM) comprising humic and fulvic acids that colour the surface waters brownish (Laudon et al. 2011). Forest and wetland management (i.e. clear-cutting, ditching, and peat mining) have influenced the hydrology and biogeochemistry of streams, lakes and rivers for centuries (Löfgren et al. 2009; Nieminen et al. 2010). For more detailed discussion, see Chaps. 21 and 25.

Eutrophication. Eutrophication is by far the most investigated and well-understood driver of change in freshwater biogeochemistry. Numerous studies illustrate the effects of agricultural practices and urban and industrial point sources on nutrient concentrations in lakes and rivers (Larsson et al. 1985; Rheinheimer 1998; Stålnacke et al. 1999b; Raike et al. 2003; Arheimer et al. 2004; Lysiak-Pastuszak et al. 2004; Ekholm et al. 2007; Humborg et al. 2007; Lindgren et al. 2007; Kronvang et al. 2009; Iital et al. 2010b) and subsequent effects on aquatic ecosystems, such as increased turbidity, anoxia and loss of biodiversity (Blenckner et al. 2006; for more detailed discussion, see Chap. 18). Long-lasting effects due to sediment nutrient release cause less efficient nutrient sequestration and retention of biogenic elements.

Hydrological alterations. Damming is more frequent in the boreal rivers owing to its higher effectiveness in terms of power generation (Humborg et al. 2000, 2008b). Major reservoirs located in their headwaters can hold between 30 and 70 % of the annual water discharge (Dynesius and Nilsson 1994; Nilsson et al. 2005). In contrast, damming is much less frequent in the lowland rivers of the south-eastern parts of the Baltic Sea catchment and it is mostly small dams and reservoirs with short water residence times that have been built there (Humborg et al. 2006). Nevertheless, there is a considerable body of literature reporting 'oligotrophication' of river systems as an effect of damming, which is the process of nutrient depletion caused by reduced contact of surface and groundwater with vegetated soils. Similar patterns have been recorded for lakes in the watershed—the higher the lake percentage of the catchment area, the lower the total organic carbon (TOC), N and DOP concentrations/fluxes (Mattsson et al. 2005; Lepisto et al. 2006).

17.2.3.2 Climate Drivers

Another driver affecting river biogeochemistry is climate, principally temperature and precipitation patterns. Although it is still difficult to quantify the extent to which human activities have affected climate in the Baltic Sea catchment, increased temperature affects hydrological properties such as evapotranspiration (positive effect) and river discharge (negative effect), so this is a vital variable for net export of dissolved and particulate matter to the Baltic Sea.

Temperature. Biogeochemical fluxes are strongly influenced by temperature. For example, stream temperature affects the uptake of nutrients and thereby their concentration in streams (Rasmussen et al. 2011). Furthermore, temperature regulates microbial activity in soils and waters, which transforms and degrades biogeochemical components. Changes in the timing of ice break-up and snow melt may alter the timing of the spring flood, which has been suggested to cause changes in DOC over time (Hongve et al. 2004; Weyhenmeyer 2008). Areas south of 61°N show high interannual variability and are especially sensitive to warming (Weyhenmeyer et al. 2011). It can be hypothesised that in such climate-sensitive areas daily freeze-thaw of surface soils activates different soil layers and could even increase soil erosion.

In the northern Baltic Sea catchment, permafrost thaw in peatlands is accelerating (Christensen et al. 2004; see also Chap. 6). Kokfelt et al. (2010) found that continued permafrost thaw and related vegetation changes towards minerotrophy (i.e. water supply through groundwater/streams/springs as opposed to precipitation) may increase C and nutrient storage in mire deposits and reduce nutrient fluxes in run-off. They concluded that rapid permafrost degradation may lead to widespread mire erosion and to relatively short periods of significantly increased nutrient loading to streams and lakes (Kokfelt et al. 2010). However, due to the relatively small areas containing permafrost, permafrost degradation is unlikely to cause much change in river discharge (see also Chap. 6).

Precipitation. Interannual variability in precipitation can have a large effect on biogeochemical fluxes. Drier-and-wetter-than-normal years have a large effect on the export of nutrients and C within Baltic Sea catchments. At the large scale, total nitrogen (TN) and total phosphorus (TP) river loads to the Baltic Sea correlate well with precipitation and discharge patterns and vary by up to 30–40 % between years (Humborg et al. 2007). Higher (wet years) and lower (dry years) export of DOC has been observed in Finland (Jager et al. 2009). Such patterns are further influenced by changes in snow pack dynamics within the catchment. Pärn and Mander (2012) concluded that the main factor driving an increase in TOC export in Estonia between 1992 and 2007 was the deepening of droughts, that is rising trends in hydrological drought days driven by climate change and magnified by man-made drainage.

17.2.3.3 Net Export of Biogenic Elements

The net export of biogenic elements as particulate, dissolved or gaseous constituents to the Baltic Sea is the result of release from natural and human sources and the transformation of matter along the aquatic continuum formed by streams, lakes and rivers (Fig. 17.2). These transformations include biological processes (formation and degradation of biomass in the widest sense) and physicochemical processes (particle formation, sedimentation, burial and gas exchange) leading to an overall retention of biogenic elements (Behrendt and Opitz 1999; Kortelainen et al. 2004, 2006b; Venohr et al. 2005), and only a proportion of the nutrients originally released from the natural and human sources will finally reach the Baltic Sea. The human drivers (Sect. 17.2.3.1) and climate drivers (Sect. 17.2.3.2) will influence the sources and transformations in different ways. Atmospheric deposition, forestry and eutrophication will clearly influence point and diffuse sources of biogenic elements, whereas these human drivers will have both positive and negative feedbacks on transformation processes and retention. Damming and other hydrological alterations including ditching as part of forestry and wetland management will directly affect water residence times in watersheds, which is a major variable determining the total period over which transformation processes can occur during passage from land to the Baltic Sea and is positively linked to the retention of biogenic elements (Valett et al. 1996; Behrendt and Opitz 1999; Søndergaard 2007). Temperature will affect weathering processes, N fixation rates, terrestrial primary production and the overall kinetics of transformation processes, while an increase in precipitation will increase weathering up to a certain threshold (Kump et al. 2000) and increase nutrient loads to the Baltic Sea, that is decrease the time available for transformation processes and retention (Hong et al. 2012).

17.3 Atmospheric Deposition and Freshwater Biogeochemistry

17.3.1 Atmospheric Deposition and Waterborne Fluxes

Atmospheric deposition (see also Chap. 15) implies the input of acids, metals, nutrients, particulates and pollutants from the atmosphere to terrestrial and aquatic ecosystems. The far-reaching impact of atmospheric deposition on freshwaters was first recognised in the 1960s. In the Nordic countries, Svante Odén was the first to warn about the consequences of atmospheric deposition, publishing an article in the largest circulation Swedish national newspaper Dagens Nyheter on 24 October 1967 entitled *The acidity of precipitation*. Odén related fish kills to sulphate deposition caused by the burning of fossil fuels. Since then, increasing numbers of freshwater quality problems in the countries

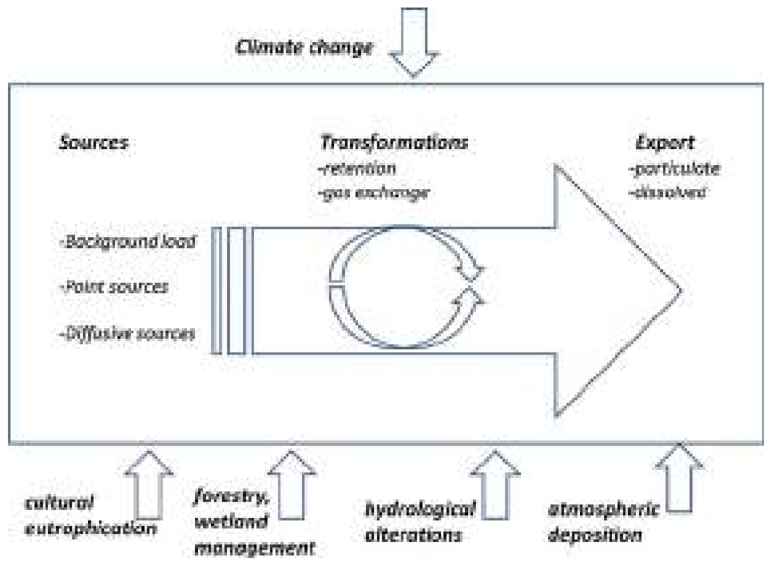

Fig. 17.2 Conceptual scheme showing how human and climate drivers influence the sources, transformations and export patterns of biogenic elements to the Baltic Sea

surrounding the Baltic Sea have been attributed to atmospheric deposition, from increased nutrient concentrations (particularly N) to increased concentrations of metals and persistent organic pollutant (Hessen et al. 1997; Agrell et al. 1999; Holt 2000; Bindler et al. 2009). Even high levels of radioactive substances accumulated in biota and sediments have their origin in atmospheric deposition. A particularly well-known example is the deposition of radioactive substances after the accident at the Chernobyl nuclear power plant on 26 April 1986, which are still being detected in lake sediments (Lusa et al. 2009). One of the most dramatic changes in freshwaters as a consequence of atmospheric deposition is probably the shift from natural N limitation to P limitation in lakes (Bergström et al. 2005; Bergström and Jansson 2006) as well as C limitation (Weyhenmeyer and Jeppesen 2010). Such shifts have caused substantial change in the biogeochemical cycles of freshwaters, and probably also the Baltic Sea.

The reasons why it took so long to identify the effects of atmospheric deposition on freshwaters are twofold. Traditionally, freshwaters (especially lakes) were studied as individual systems, with little connection to large-scale driving forces (Livingstone and Hari 2008), and the impact of atmospheric deposition on freshwaters is often masked by other drivers and therefore often overlooked.

As background levels began to need quantifying, increasing interest was shown in atmospheric deposition. Background levels for substances originating in atmospheric deposition are generally very low and still relatively uncertain, especially for those substances that are transformed as they move through the environment. Most information on background levels is available from sediment cores. Analysing sediment cores from freshwaters and taking lead (Pb) as an example indicates a significant rise in atmospheric Pb fallout from about AD 1000, followed by rapid increase during the Industrial Revolution. Concentrations peaked in the 1970s and then declined (Branvall et al. 2001). Background levels can also be quantified by methods other than sediment analyses. For example, Hägg et al. (2010) used a statistical approach to estimate a background flux of about 100 kg N km^{-2} $year^{-1}$ from boreal catchments.

Atmospheric deposition to freshwaters is best studied in remote regions. Freshwaters in these areas may be seen as mirrors of the atmosphere and so may be used as reference systems that are highly sensitive to changes in atmospheric deposition and climate. A substantial proportion of the freshwaters draining into the Baltic Sea originates in remote regions. Although atmospheric deposition shows a strong north–south gradient with the deposition of all substances increasing southwards (Weyhenmeyer 2008), the relative importance of atmospheric deposition on freshwater biogeochemical cycling increases towards the remote northern region, where nutrient and pollutant inputs from other sources are small. Atmospheric deposition generally reflects industrial development. At present, atmospheric deposition of metals and sulphate is declining not just over the Baltic Sea area but also over other regions (Ruoho-Airola and Salminen 2003; Harmens et al. 2010; Slemr et al. 2011). The decline in sulphate deposition has resulted in a rapid recovery of freshwaters from acidification (Evans et al. 2001). This recovery has reduced the need for intense liming activities with consequent impacts on the bioaccumulation of mercury (Hg) (Shastria and Diwekar 2008). Nitrogen deposition has also decreased over the northern regions, which has lowered nitrate concentrations in freshwaters (Weyhenmeyer et al. 2007; Kothawala et al. 2011). Some of the freshwaters surrounding the Baltic Sea currently experience nitrate-depleted conditions during summer resulting in periodic N limitation (Weyhenmeyer et al. 2007). If N deposition continues to decrease, a substantial proportion of freshwaters in the Baltic Sea drainage area may shift back towards N-limited systems with consequent impacts on biogeochemical cycling (Weyhenmeyer and Jeppesen 2010).

17.3.2 Transformations of Nutrients Along the Aquatic Continuum

Background levels assume that transformations in the landscape have reached steady state, that is their natural release is in balance with natural removal processes. Increasing numbers of studies, however, show that transformations can either decrease or increase in their efficiency with changing environmental conditions. For example, Weyhenmeyer and Jeppesen (2010) reported that the efficiency of nitrate removal from freshwaters varies with changes in N deposition. Since substantial fractions of substances are retained or lost in boreal catchments by a wide range of transformation processes, changes in transformation processes within the landscape will have major impacts on river export. Hägg et al. (2010) estimated that about 75 % of the anthropogenic N deposited from the atmosphere was retained within the boreal landscape before entering the sea. Retention capacity varies depending on background levels and the biogeochemical process in question. There are indications that southern parts of the boreal region surrounding the Baltic Sea are more N-saturated than northern parts (Weyhenmeyer and Jeppesen 2010), resulting in a more effective N retention capacity in northern landscapes than southern. This pattern is probably also valid for other substances and will be affected by changes in trends in atmospheric deposition.

17.3.3 Climate Impacts on Atmospheric Deposition and the Effect on Waterborne Fluxes

To distinguish the effects of changes in atmospheric deposition from the effects of climate change is a major challenge, mainly because changes in climate strongly covary with changes in atmospheric deposition. There are two reasons for this strong covariation. Over the long term, changes in climate follow changes in carbon dioxide (CO_2) emissions that themselves follow changes in the emissions of other substances (Lamarque et al. 2010). While over the short term, deposition patterns for substances from the atmosphere are determined by atmospheric circulation patterns in a similar way to meteorological variables (such as winds and precipitation); thus, short-term variations in atmospheric deposition and climate follow each other (Dayan and Lamb 2005; Weyhenmeyer 2008; Shi et al. 2011). While physical freshwater variables such as temperature, run-off and ice cover can easily be attributed to climate change, it is more difficult to assess climate change effects on chemical and biological freshwater variables (Adrian et al. 2009). Carbon is probably one of the most debated freshwater variables where agreement has not yet been reached on whether it is changes in atmospheric deposition or in climate that are mainly responsible for the recent wide-scale increases. A variety of studies attribute the increase in C concentrations and/or loads mainly to changes in climate variables such as run-off (Andersson et al. 1991; Schindler et al. 1997; Freeman et al. 2001; Worrall et al. 2004; Erlandsson et al. 2008), temperature, growing season length and run-off season duration (Weyhenmeyer and Karlsson 2009), solar radiation (Hudson et al. 2003), the timing of ice break-up and snowmelt (Hongve et al. 2004; Weyhenmeyer 2008), and soil moisture (Worrall et al. 2006). It has been proposed that concentrations and fluxes of DOC are more strongly related to climate and landscape topography than to internal properties of aquatic ecosystems (Mulholland 2003). Long-term trends in the timing of ice freeze-up (getting later) and ice break-up (getting earlier) towards shorter periods of ice cover have been reported for lakes and streams around the Northern Hemisphere (Palecki and Barry 1986; Magnuson et al. 2000; Blenckner et al. 2004, 2007; Prowse and Brown 2010) and can profoundly affect biogeochemical components. Interannual variability in freeze-up and break-up dates is also important (Weyhenmeyer et al. 2011). Other studies attribute the increase in DOC mainly to changes in atmospheric deposition chemistry (Freeman et al. 2004; Bragazza et al. 2006; Evans et al. 2006; Monteith et al. 2007). As Roulet and Moore (2006) concluded, it is probably a combination of different drivers that are responsible for increasing C levels in freshwaters (Fig. 17.3). When a variable is affected by several drivers, patterns become complex and the possibility of additive, synergistic and antagonistic effects occurs. In this context, pH is a good example. In freshwaters, pH has a strong negative relationship with sulphate deposition (Evans et al. 2001) and a strong positive relationship with temperature (Houle et al. 2010). Since temperature is currently rising and sulphate deposition currently falling, the accelerated recovery of pH in freshwaters must be attributed to both.

17.3.4 Current and Future Export to the Baltic Sea

Over previous decades and especially in the 1970s and 1980s, atmospheric deposition seems likely to have had a stronger effect on freshwater biogeochemical conditions in the Baltic Sea drainage area than climate. However, this pattern seems to weaken as atmospheric deposition declines. A shift back to climate regulation of freshwater

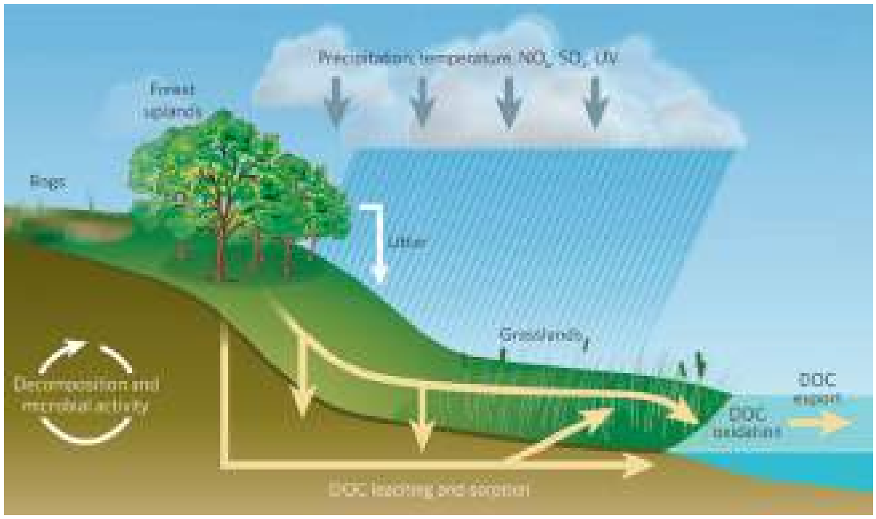

Fig. 17.3 The decomposition and subsequent leaching of organic matter in bogs, forests and wetlands are the principal sources of dissolved organic carbon (*DOC*) in the terrestrial landscape. Production is mediated by several physical and biogeochemical factors, such as the atmospheric deposition of nitrates and sulphates, moisture and temperature. The rate of export of terrestrial DOC is determined by the rate of production combined with the rate of sorption by mineral soils, and the availability of pathways for water through the environment (Roulet and Moore 2006)

biogeochemistry has already been observed in the United States (Mitchell and Likens 2011). This suggests that the biogeochemical conditions of freshwaters and the Baltic Sea could change rapidly.

17.4 Forestry, Wetland Management and Freshwater Biogeochemistry

17.4.1 Forest, Wetlands and Waterborne Fluxes

Forests and wetlands dominate the landscape in northern Fennoscandia accounting for more than 85 % of the total land area in northern Sweden, of which wetlands account for 13 % (SLU 2010; see also Chap. 21). In Finnish watersheds, the proportion of upland forests is 29–64 % (average 49 %) and the proportion of peatlands 3–60 % (average 22 %). The percentage of peatlands is highest in the band between 63° and 66°N, whereas the proportion of forests increases towards the south. Over half of the Finnish peatlands, which originally covered a third of the land area has been ditched, mostly for forestry (Aarne 1994). In Estonia, forestry is the dominant land use occurring on 21,974 km^2 (50.3 %) of the total territory. Wetlands are extensive landscapes covering 25–30 % of the country's territory, including a substantial fraction of agricultural and forest land. About 70 % of Estonian peatlands are drained or influenced by drainage to an extent that presumably no longer allows peat to accumulate (Ilomets and Kallas 1997).

As the major northern Swedish rivers pass through the landscape from their headwaters in the Scandinavian mountains, they get progressively enriched in TOC and silica (Si); the latter associated with weathering release (Humborg et al. 2004). This enrichment is directly coupled to forest and wetland cover at the landscape level (Humborg et al. 2004; Smedberg et al. 2006). The importance of boreal forests was also emphasised in a study of DOM export along a European climate gradient; the study found that the export of DOC was highest from the Finnish boreal forest sites and clearly associated with forest and wetland catchment coverage (Mattsson et al. 2009). Forests affect the biogeochemical cycles of elements in several ways; trees and in particular coniferous trees are efficient filters for dry and wet deposition increasing the airborne loads of elements to the soil (Robertson et al. 2000). Trees promote soil formation, weathering and C accumulation in soils and also transfer a large amount of current photosynthates to the soil, the latter being an important source of DOC via rhizodeposition (Giesler et al. 2007). Wetlands have a specific role in the boreal landscape since they are hotspots for waterborne organic C (Fig. 17.3). For instance, wetland coverage in the boreal landscape has been found to be positively related to organic C export in streams (Laudon et al. 2011). In Finland, high peatland proportion has been shown to increase average annual leaching of TOC both in managed (Kortelainen et al. 1997, 2006a; Kortelainen and Saukkonen 1998; Mattsson et al. 2003) and unmanaged catchments (Mattsson et al. 2003; Kortelainen et al. 2006a).

The export of DOC and a number of elements is highly seasonal and differs between forest and wetland-dominated sites (Laudon et al. 2011). Long-term monitoring of a boreal mixed forest/wetland drainage basin has shown that wetland-dominated stream catchments are characterised by low DOC concentrations during peak flow events and high DOC concentrations during baseflow conditions (Giesler et al. 2007; Laudon et al. 2011). In forested catchments, concentrations and element fluxes of DOC and elements associated with DOC increase during spring snowmelt (Dyson et al. 2011), whereas other elements are mostly diluted. Typically, elements linked to weathering such as Si, calcium (Ca), magnesium (Mg), potassium (K) and sodium (Na) are highest at baseflow conditions (Smedberg et al. 2006). However, in two forested peat-dominated (drained and undrained) catchments in eastern Finland, DOC concentrations were positively correlated with cation concentrations in both catchments indicating a common peat/groundwater flow path (Dinsmore et al. 2011). Different landscape sources also affect the 'quality' of exported DOC—during baseflow conditions, there is a greater export of older more recalcitrant DOC compared to peak flow conditions where the flux is dominated by younger DOC, that is from surface soil horizons of forest soils (Laudon et al. 2011).

In larger catchments with diverse land cover types, water mixing masks clear patterns; there is also a tendency for decreasing baseflow DOC concentrations with increasing catchment size, which is attributed to the increased influence of deep groundwater or shifts in soil texture from unsorted tills to more sorted fine materials in the lower parts of large catchments (Laudon et al. 2011). A study of water chemistry within the River Kalix catchment showed that spring flood events were dominated by TOC input from upper forest soil horizons and peatlands, while storm peak flow events were dominated by TOC flushed from peatlands. Most of the seasonality and patterns observed in headwater streams are significantly reduced with increasing distance downstream, similar to DOC export and concentrations (Wolock et al. 1997; Humborg et al. 2004; Mattsson et al. 2005; Temnerud and Bishop 2005). This may be caused by a number of factors such as mixing of different sources and degradation of DOM and/or sedimentation of elements along the flowpath. Furthermore, large catchments with longer water retention times are more buffered against variations than headwater systems. Overall, it is clear from many studies that forests and wetlands play a key role in the terrestrial export of a number of elements (Fig. 17.3).

17.4.2 Influence of Management Practices

17.4.2.1 Clear-Cutting

The effects of forestry practices on freshwater quality have been investigated in a number of studies in Sweden and Finland (Ahtiainen 1992; Ahtiainen and Huttunen 1999; Finer et al. 2004; Laudon et al. 2009; Löfgren et al. 2009; Nieminen et al. 2010). One of the most important effects of clear-cutting is the reduction in evapotranspiration, resulting in increased run-off, elevated groundwater levels and a change towards shallower flow pathways. Another effect results from the decreased competition from trees after a clear-cut, which increases nutrient availability and nitrification rates. Increased insolation to the soil caused by the absence of a forest canopy will also promote degradation of soil organic matter and mineralisation rates. The impact on the export of elements is, however, linked to the intensity, extent and duration of forest practices, but also reflects climate, topography and soil properties.

In northern Sweden, logging resulted in increased run-off and increased concentrations of Na, K, chloride, TN, TP and suspended material from the two study catchments, whereas nitrate leaching increased only from the catchment without a forest buffer (Löfgren et al. 2009). A high frequency of water sampling during a clear-cut catchment experiment in northern Sweden one year after harvesting showed increased streamwater DOC concentrations during the growing season. This study supports the hypothesis that a raised groundwater level following harvesting caused the increased DOC concentration during hydrological episodes and low-flow conditions (Laudon et al. 2009). In Finland, clear-cutting and subsequent scarification increased total and inorganic P and N, total iron (Fe) concentrations and suspended solids (Ahtiainen 1992). Nieminen (2004) showed that clear-cutting significantly increased the export of DOC and N from drained productive peatlands, while only small increases in P export were found. Buffer zones have shown to be efficient in retaining inorganic nutrients (Silvan et al. 2005; Vaananen et al. 2008; Vikman et al. 2010). Koskinen et al. (2011) showed that the calculated mean annual leaching of P, N and TOC from post-restoration treatment areas was high in comparison with average leaching from undisturbed catchments (Mattsson et al. 2003; Kortelainen et al. 2006a) and average leaching from managed forested catchments (Kortelainen et al. 1997).

17.4.2.2 Site Preparation

Site preparation may add to the effect of clear-cutting; in an experiment by Löfgren et al. (2009), site preparation caused an additional increase in DOC losses of 79 %. A Finnish study (Piirainen et al. 2007, 2009) showed similar results and suggests that site preparation after forest harvesting can increase C, N, P and cation leaching from soils more than the clear-cutting itself. Although this illustrates the effect over the short term, short-term and long-term effects of forestry can be significantly different. Clear-cutting often results in increased concentrations and leaching some years after treatment, but over the long term and after re-establishments of forests this is likely to result in a lower water table and decreasing leaching.

17.4.2.3 Ditching

Ditching has been a common forest practice especially on peat soils and has been found to result in a short-term increase in TOC concentrations (Heikurainen et al. 1978; Moore 1987). Over the long term, ditching lowers the groundwater level and can result in decreased TOC leaching. However, Sallantaus (1994) found no differences in the leaching of TOC between natural fen, natural bog, drained fen and drained bogs several years after treatment. Rantakari et al. (2010) studied long-term effects of ditching in small headwater catchments and showed decreased total inorganic carbon (TIC) and TOC concentrations, but no significant effects on lateral C export due to increased run-off patterns. Reditching, that is managing ditch networks, has been shown to result in decreasing DOC and DON concentrations, and increasing inorganic N, suspended solids and base cation concentrations, whereas no significant changes were found in TN and TP concentrations (Joensuu et al. 2001, 2002). In an experiment carried out in nine pairs of treated and control (no maintenance) catchments located in southern and central Finland, a significant increase in the export of suspended solids for the four-year study period following the ditch network maintenance and aluminium (Al) export increased for one year. The export of N, P and Fe was not significantly changed, and DOC and manganese (Mn) export decreased after the ditch maintenance operation (Nieminen et al. 2010).

17.4.2.4 Multi-Stressors

There are relatively few long-term studies on the effect of different forest practices. However, Kortelainen and Saukkonen (1998) studied average long-term leaching of C, N, P and Fe from 20 headwater catchments representing Finnish forestry land, including the most important forest practices (ditching, clear-cutting, scarification and fertilisation). Ditching was the largest-scale forestry practice in the study catchments. In many catchments, ditching was already ongoing in the early 1960s, although these catchments were not monitored until the 1970s. Considering the differences in catchment size, location, forest type and peatland type as well as different forest practices, the regional differences in the average long-term leaching of total organic nitrogen (TON) and TOC were not large. Interannual variation in TOC in single headwater streams was shown to be greater than spatial variation in average annual TOC fluxes between the catchments (Kortelainen et al. 1997). Regional variation was reduced because concentrations were generally lower in northernmost catchments, while run-off from the study catchments increased to the north. Moreover, the concentrations were higher in the catchments with a high peatland proportion, while run-off from these catchments was slightly lower compared to the catchments with a low peatland proportion. The average annual leaching of TOC, TN, TP and total Fe was greater in southern catchments than northern catchments. Furthermore, high peatland proportion increased average annual TOC, TN and total Fe export both from managed and unmanaged catchments (Kortelainen and Saukkonen 1998; Kortelainen et al. 1997, 2006a; Mattsson et al. 2003).

17.4.3 Transformations of Carbon Along the Aquatic Continuum

Receiving lakes and streams are described as 'active pipes' in the export from land to ocean (Cole et al. 2007), and their role as regulators of C cycling has been depicted in several recent studies (Cole et al. 2007; Tranvik et al. 2009; Humborg et al. 2010; Lyon et al. 2010; Einola et al. 2011). Estimates by Tranvik et al. (2009) demonstrated that the global annual emission of CO_2 from inland waters is of the same magnitude as CO_2 burial by the oceans and boreal lakes seem to be especially important in this context. This has also been emphasised in a number of studies showing that lakes are supersaturated in CO_2 (Jansson et al. 2000; Sobek et al. 2005; Kortelainen et al. 2006b). A recent budget calculates the total CO_2 efflux from all Swedish lakes and streams to be 2.58 Tg C year^{-1} (Humborg et al. 2010; Fig. 17.4). In-lake respiration, mainly derived from

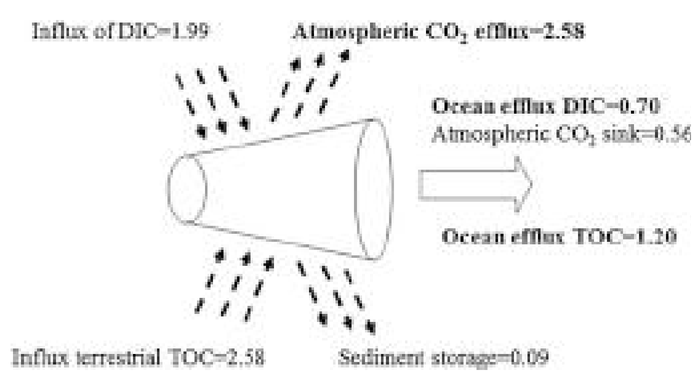

Fig. 17.4 Schematic view of the major inorganic and organic carbon pathways along the aquatic continuum. Fluxes represent Sweden as a whole and are expressed in Tg C year^{-1} (Humborg et al. 2010)

terrestrial C, has been suggested as an important contributor to the CO_2 emitted from lakes (del Giorgio et al. 1999), but inflow of DIC and CO_2-rich groundwater can also contribute to lake CO_2 (Striegl et al. 2001). The latter was stressed in a study by Humborg et al. (2010) illustrating that the partial pressure of CO_2 (pCO_2) in a large number of Swedish lakes and streams was strongly related to factors indicative of groundwater influence, that is typically weathering products such as Si, Mg and DIC. The significance of the weathering component in C sequestration has been widely overlooked and may be vital for interpreting future C budgets of major boreal and Arctic watersheds. Weathering reactions consume atmospheric CO_2 and form DIC that is locked over geological timescales in the aqueous phase and finally in ocean sediments. The weathering sink of atmospheric CO_2 constitutes a negative feedback on atmospheric CO_2 due to climate change at northern latitudes (Smedberg et al. 2006; Lyon et al. 2010). Thus, the overall feedback of boreal groundwaters, streams and lakes to atmospheric CO_2 is heavily debated. Quantitative estimates are still lacking of the relative importance of groundwater versus in-lake or stream processes contributing to CO_2 emissions. Nevertheless, it is clear that future changes in the terrestrial DOC export and changes in the influx of groundwater to the receiving lakes and streams can potentially affect CO_2 emissions in aquatic systems. Such changes could be additive in a scenario where both the groundwater influx and DOC export increase.

Dinsmore et al. (2011) found the snowpack to represent a potentially important, and often overlooked, transient C store in boreal snow-covered catchments. Meltwater from the snowpack represented an important source of streamwater CO_2 in two forested peatland (drained and undrained) catchments in eastern Finland, contributing up to 49 % of total downstream CO_2 export during the snowmelt period in April/May.

17.4.4 Climate Impacts on Waterborne Losses from Forests and Wetlands

There are a number of possible effects of changing climatic conditions that may impact waterborne losses from catchments in northern Scandinavia. The two overarching factors are likely to be the amount and seasonality of precipitation and temperature. In the northern Baltic Sea catchment, the spring flood associated with peak snowmelt mainly occurs in mid-May for the Taiga zone and in June for the Fennoscandian mountains (Ingri et al. 2005), but the timing and magnitude of this main hydrologic event may shift under a changing climate. Korhonen and Kuusisto (2010) showed that although no overall changes have been observed in mean annual stream discharge for a large number of Finnish streams, hydrological regimes during winter and spring have changed significantly (see also Chap. 5). This is mainly attributed to winters and springs becoming milder and in consequence late-winter and early-spring discharges increasing. Studies combining several hydrological model simulations (to the end of this century) in the Swedish Regional Climate Modelling Programme show that while results varied depending on the climate change scenario and model boundary conditions, some projections were consistent between runs, for example an overall increased autumn and winter run-off and increased annual run-off volume in northern Sweden (Andreasson et al. 2004). In a similar study focusing on river run-off for the entire Baltic Sea drainage basin, Graham (2004) found a general trend of reduced river flow from the south-eastern Baltic Sea catchment together with increased river flow from the north.

The magnitude and timing of the spring flood is important for the annual fluxes of dissolved and particulate matter in boreal river systems (Woo et al. 2008). Higher spring temperatures and a higher proportion of winter precipitation falling as rain could result in earlier snowmelt and thinner snowpacks. This would lead to spring floods occurring earlier in the year and having a lower magnitude (Andreasson et al. 2004; Woo et al. 2008). However, in high-latitude catchments, sustained sub-zero winter temperatures coupled with increased precipitation may lead to maintained or even increased spring floods (Dankers and Middelkoop 2008). Rantakari et al. (2010) compared TOC fluxes in headwater streams between two climatically different years and found decreased TOC export during the spring ice melt period and increasing export during the rest of the year including snow-cover and snow-free periods. Wet years have been shown to favour the export of TOC from forest-dominated areas (Kohler et al. 2009) and imply that future wetter conditions may increase the TOC export as well as many elements associated with TOC such as Al, Fe, trace elements and potentially harmful elements such as Hg. Empirical models of streamwater fluxes of DOC including both soil temperature and water fluxes have been able to predict seasonal variation in streamwater DOC concentrations reasonably accurately showing that both parameters are important (Kohler et al. 2009). Simulations of a climatic scenario with an average temperature increase of about 2.5 °C and increase in precipitation of 25 % for a boreal headwater stream suggest an increase in the annual TOC export of approximately 15 % (Kohler et al. 2009). The model simulations also indicate that the autumn months are particularly sensitive and that wetter and warmer conditions could cause a TOC increase of up to 5 mg l^{-1} (Kohler et al. 2009).

Temperature effects related to snow cover may also have profound effects on waterborne DOC export from forests (Agren et al. 2010). Higher temperature in organic soils has been shown to increase DOC export, not because of temperature control on production rates but because temperature affects C consumption and microbial activity (Moore and Dalva 2001; Pietikainen et al. 2005). Stedmon et al. (2006) found that seasonal DOM export patterns to a temperate Danish estuary reflected temperature fluctuations in more natural subdrainage areas, whereas precipitation controlled export patterns in sub-drainage areas dominated by agriculture. While increased soil temperature is likely to lead to enhanced export of DOC, it will not necessarily lead to increased fluvial export of nutrients from mire ecosystems (depending on the efficiency of internal nutrient cycling in mires). In a study of Alaskan boreal peatlands, water table depth and soil temperature were found to be significant factors influencing DOC and DON concentrations in streamwater, but it was also shown that bog peatlands retain N (D'Amore et al. 2010).

A snow manipulation experiment in northern Sweden showed that cold winters with a deeper seasonal freeze-thaw layer increased streamwater DOC concentrations during spring snowmelt (Haei et al. 2010). The experiment is interesting in that it shows the importance of winter climate conditions for streamwater DOC export. There is less information available on potential effects of climate change on the export of elements not directly associated with DOC. Concentrations of elements related to weathering normally show an inverse relationship to DOC in forested catchments (Smedberg et al. 2006). The highest concentrations are generally found during baseflow conditions with concentrations diluted during flow events. Potentially, increased weathering or changes in hydrological flow paths with a greater contribution of groundwater could lead to an increased export of elements that are dominant during baseflow conditions. Such changes have been reported from permafrost-affected areas where permafrost thaw is predicted to shift hydrological pathways from being surface water dominated to groundwater dominated (Frey and McClelland 2009). More groundwater formation indicates more weathering and increased weathering consumption of CO_2, because more CO_2 is transported to deeper groundwater flow depths due to permafrost thaw. This may increase the transport of DIC and the weathering sink of atmospheric CO_2, and thus constitute a negative feedback on atmospheric CO_2 due to climate change at northern latitudes (Smedberg et al. 2006; Lyon et al. 2009). Actually, permafrost thaw rates of 0.7–1.3 cm $year^{-1}$ have been reported for watersheds in northern Sweden (Lyon et al. 2010). For these tundra catchments, an even contribution of DOC and DIC to the net mass flux of C from the terrestrial environment to and through the surface water system has been suggested. Under future potential scenarios, there could be a corresponding increase in the flux of both DOC and DIC from this landscape due to increased advective travel times associated with deeper flow pathways (Lyon et al. 2009, 2010). Such a potential increase and this particular balance between DOC and DIC are crucial in understanding the relevant feedbacks between future climatic change effects and the hydrological and biogeochemical system. In the Baltic Sea basin, the spatial distribution of permafrost is essentially limited to high-alpine landscapes and sub-Arctic peatlands in Sweden and Finland (Christiansen et al. 2010). There is no quantitative estimate of how this projected permafrost thaw could affect fluvial transport to the northern Baltic Sea basin, but considering the limited spatial extent of permafrost the shifts in export of major ions, nutrients and organic matter are likely to be limited. Climate-related changes in the distribution of mires may also affect landscape fluxes of elements. Fennoscandian mire types can be widely divided into four types: raised bogs, aapa mires, blanket bogs and palsa mires (Pajunen 2005; Parviainen and Luoto 2007). Parviainen and Luoto (2007) investigated the climatic envelopes of these mire types and found that the distributional limits of aapa mires, palsa mires and raised bogs were primarily associated with mean annual air temperature, while blanket bogs were also largely dependent on high levels of precipitation. Only palsa mires were found to have a very narrow climate envelope indicating short-term sensitivity to climate change. Over the short term, climate change is thus unlikely to affect the spatial distribution of wetlands.

Changes in forest practices related to improved climatic growth conditions such as a longer growing season might also affect the biogeochemistry of forest stands. This could include changes in tree species, shorter times between plantation and harvest, or more intense forest management. However, all these changes will only take effect over the long term since forest growth is still relatively slow at these latitudes. The effects of climate drivers on terrestrial ecosystems were reviewed in the first assessment of climate change in the Baltic Sea basin, and it was concluded that *It is apparent that trees are growing taller and lusher compared with a few decades ago; net primary productivity has increased, the ecosystems are net carbon sinks. The likely cause is an extended growing season associated with higher average temperatures* (Smith et al. 2008). An overall increase in forest productivity may, however, have a more direct effect since it may increase litter deposition and change the flow of current photosynthates to the soil (Olsson et al. 2005). Whether this also has implications for waterborne export to aquatic ecosystems is still unclear.

17.4.5 Current and Future Export to the Baltic Sea

Most studies investigating the effects of geochemical fluxes in relation to landscape properties such as forest or wetlands or forest practices have mainly focused on smaller catchments or headwater streams. The implications at a larger scale may be less pronounced. For instance, Futter et al. (2010) estimated the contribution of short-term increases in nitrate leaching following stem-only harvesting at a larger scale and suggested that this effect accounted for about 3 % of the overall Swedish N load to the Baltic Sea, despite the fact that short-term increases in nitrate leaching can be very pronounced in headwater streams after harvesting (Rosen et al. 1996). Model simulations of N and P fluxes from Swedish forest land to the marine environment suggest that over 93 % of the leaching losses can be attributed to background loads and the small remainder to forest practices (Brandt and Rapp 2008). Similarly, forestry in Finnish river basins was estimated to contribute 9 % of the total N export to the Baltic Sea on average (Lepisto et al. 2006). Kenttämies (2006) estimated the loading of P and N from forestry to be 8 and 5 %, respectively, of the total anthropogenic load in Finland. A major reason for the relatively minor contribution of forest practice to large-scale losses is that the extent of forested land area that is annually affected by forest practices is relatively low; in Finland, about 2.5 % of the entire country (Kortelainen and Saukkonen 1998). Lepistö et al. (1995) showed that within both Swedish and Finnish headwater catchments large-scale forest management practices were needed before any clear effect on spatial variability of N leaching could be detected. Although forestry is often the largest-scale human impact in headwater streams, it is thus reasonable to assume that any effect on element fluxes to the Baltic Sea related to forest management can only relate to large-scale change in forest practices.

Climate is the overall factor that might impact waterborne fluxes of elements. Although several studies from forested headwater streams indicate that, for instance, change in winter conditions affect DOC losses, it still remains to be shown that climate change will also have large-scale impacts in large catchments flowing to the Baltic Sea. Over the coming century, climate change is unlikely to affect the spatial distribution of wetlands in the Baltic Sea basin. In general, however, increased ambient and soil temperatures (including permafrost thaw) and increased precipitation could lead to increased DOC export from wetlands, especially during baseflow conditions. This increase is likely to be more pronounced in the boreal/sub-Arctic regions of the Baltic Sea catchment. In fact, recent scenario studies indicate that DOC production from terrestrial vegetation, modelled by the LPJ-GUESS ecosystem model, could increase by 30–43 % (Omstedt et al. 2012). Rising temperatures causing an increase in net ecosystem production, increasing both the available substrate and the rate of decomposition of plant biomass derived organic matter, provided the most important explanation for the increase in DOC export from wetlands and forests. In the scenario calculations of riverine fluxes, DOC fluxes to the Baltic Sea generally increased, especially in the northern catchments, in the range of 20–50 %, with the greatest increase in the Gulf of Finland. The increasing fluxes resulted mainly from the increasing run-off, since modelled concentration changes in river water were about 10 % (Omstedt et al. 2012).

17.5 Eutrophication and Freshwater Biogeochemistry

17.5.1 Agriculture, Urban Areas and Waterborne Fluxes

Agriculture dominates the southern and eastern areas of the Baltic Sea drainage basin (Fig. 17.1; Table 17.1) and covers from 7 % of total land area in Finland and Sweden to 60 % in Denmark, Table 17.2. Agricultural production is intensive in large parts of Denmark, Germany, Poland, and southern Sweden and Finland (Table 17.2), whereas more extensive agricultural production is seen in the Baltic States. The agricultural structure differs markedly between countries: in Poland, 70 % of the area is owned by small farms of less than 10 hectares, whereas in Denmark most farms are larger than 100 hectares (FAO 2003; Benoist and Marquer 2006a, b, c, d, e, f, g; see also Chap. 21).

Urban areas and industry which are also concentrated in the southern and eastern parts of the Baltic Sea catchment (Table 17.1) are major point sources for nutrients. Although several countries have already implemented effective wastewater treatment plants (Table 17.3), poor wastewater treatment is still an issue in rural areas and effective sewage treatment would still have a significant potential in the transitional countries (Humborg et al. 2007; Iital et al. 2010b).

The most recent compilation of nutrient loads to the Baltic Sea reports loadings via rivers and coastal point sources of respectively 638,000 t TN and 28,370 t TP (HELCOM 2011, see Table 17.4). The rivers Vistula (Poland), Nemunas (Belarus, Lithuania and Russia), Oder (Poland and Germany) and Neva (Russia) all drain the southern cultivated part of the Baltic Sea catchment and account for the majority of the nutrient inputs to the Baltic Sea (HELCOM 2004). Many rivers of northern Sweden and Finland are still largely unperturbed by human activities (Humborg et al. 2003). The difference between the northern boreal and southern cultivated systems is illustrated by comparing the Kemijoki and Oder rivers. The Kemijoki drains the northern part of Finland and discharges into the Gulf of Bothnia with average concentrations of 0.4 mg TN l^{-1} and 0.02 mg TP l^{-1} for the period 1994–2008. The Oder, draining a large part of Poland, has TN and TP concentrations 10-fold higher (HELCOM 2011). Source apportionment for the riverine nutrient loading of the Baltic Sea in 2000 indicates that TN natural background losses

Table 17.2 Agricultural land subdivided by crop type, average consumption of fertiliser and manure, and livestock density in countries within the Baltic Sea drainage basin in 2005 (Benoist and Marquer 2006a, b, c, d, e, f, g) including Belarus and Russia (Sweitzer et al. 1996)

	Denmark	Sweden	Finland	Poland	Lithuania	Latvia	Russia[a]	Belarus[a]	Estonia
Agricultural area, 1000 ha	2588	3096	2262	13,132	2338	1302	4803	4081	764
Share of total land cover (%)	60	7	7	42	36	20	15	45	17
Cover of agricultural area (%)									
Arable land	69	41	60	72	45	42	31	58	72
Forage plants	17	33	28	6	26	23		21	6
Permanent grass and meadows	7	15	1	19	26	28	69[b]	21	19
Fallow	7	10	11	1	3	6		0	1
Consumption in 2005, kg N ha^{-1} *agricultural area*									
Mineral fertiliser	74.0	51.0	74.2	68.2	Nd	25.9	Nd	55[c]	26.3
Manure[d]	72.2	25.4[e]	42.9	38.3	27.2	17.9	Nd	42[c]	25.7
Livestock, head ha^{-1} *agricultural area*									
Dairy cows	0.14	0.13	0.14	0.21	0.18	0.14	Nd	0.22	0.15
Pigs	3.68	0.58	0.64	1.42	0.48	0.33	Nd	0.46	0.46

For Denmark, the area that drains into the North Sea is also included
[a] Agricultural area for that part of Belarus which drains into the Baltic Sea (Sweitzer et al. 1996)
[b] Cover of forage plants, permanent grass and meadows
[c] Consumption of fertiliser and manure is average numbers from National Statistic of Belarus (2010)
[d] NH_3 volatilisation from storage and stable excluded
[e] Input to the soil–ammonia (NH_3) volatilisation excluded

Table 17.3 Levels of sewage treatment by country in 2004 (Humborg et al. 2007)

Country	Primary (%)	Secondary (%)	Tertiary (%)
Belarus	0	50	0
Czech Republic	0	61	0
Denmark	2	5	81
Estonia	2	34	34
Finland	0	0	80
Germany	0	9	85
Lithuania	33	6	18
Latvia	2	35	33
Poland	2	23	34
Russia	0	50	0
Sweden	0	6	86

Primary treatment (lowest effect) is mainly through sedimentation, while secondary and tertiary treatments (highest effect) are mainly through biological filtration and chemical precipitation

Table 17.4 Total nutrient load by country via rivers and coastal point sources for 2006 according to data reported to HELCOM (HELCOM 2011)

Country	Total nitrogen load (t $year^{-1}$)	%	Total phosphorus load (t $year^{-1}$)	%
Germany	16,900	3	490	2
Denmark	53,000	8	1520	5
Estonia	20,400	3	790	3
Finland	79,000	12	3490	12
Lithuania	28,000	4	1240	4
Latvia	59,500	9	2800	10
Russia	107,600	17	4070	14
Poland	152,600	24	10,240	36
Sweden	121,000	19	3730	13
Total	638,000	100	28,370	100

account for 28 %, diffuse losses for 64 %, and point source discharges for 8 %. For TP, the contributions are 26, 55 and 19 %, respectively (HELCOM 2004). Agriculture accounts for the majority of the diffuse losses: 70–90 % for TN and 60–80 % for TP (HELCOM 2011).

17.5.1.1 Long-Term Trends in Land Use and Nutrient Loads

The N and P content in rivers increased steadily through the twentieth century with the highest concentrations measured during the 1980s and 1990s. Even 100 years ago, the impact from human activities on nutrient losses was substantial. Natural fertilisers were used in agriculture, and the construction of water supply and sewage systems increased the output of waste to inland and coastal waters (Savchuk et al. 2008). In addition, the widespread draining of lakes and wetlands which led to a loss of nutrient retention capacity mainly occurred at the end of the nineteenth century and in the first decades of the twentieth century (Hoffmann et al. 2000; Schernewski and Neumann 2005). Using a modelling approach, Schernewski and Neumann (2005) calculated that the overall nutrient loads to the Baltic Sea have increased by

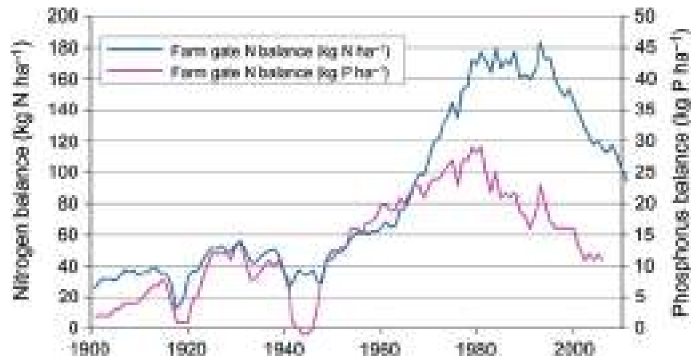

Fig. 17.5 The farm gate nitrogen (N) and phosphorus (P) balance for Denmark over the period 1900–2010 (Kyllingsbæek and Hansen 2007; Kyllingsbæk 2008, redrawn)

a factor of 2.4 (TN) and 3.1 (TP) over the past 100 years. Savchuk et al. (2008) reconstructed external nutrient inputs from various literature and data sources and estimated a similar increase over the past century: factors of 2 (TN) and 3 (TP). Gadegast et al. (2012) and Behrendt et al. (2008) using the Moneris model found TN loadings from the Oder system increased by a factor of 4.6 between 1880 and 1980.

The nutrient surplus, the balance between the application and harvest offtake of N or P on agricultural land, is an indicator for nutrient losses from agriculture. Figure 17.5 illustrates the dramatic intensification of Danish agriculture, especially following the Second World War. The graphic also shows the successful regulation of nutrient surplus in Danish agriculture over the past three decades. For a detailed assessment of long-term change in land use see Chap. 25.

In Europe, change in land use over the past few centuries has been characterised by alternating expansion and contraction of agricultural areas. Expansion was a consequence of increasing food demand caused by a growing population. By the end of the seventeenth century, owing to better agricultural methods (new rotation systems and manure use), productivity increased and so a period of contracting cultivated area began. This lasted until the mid-eighteenth century when continued population growth meant the agricultural area again expanded, continuing for the next 200 years. Since the late 1950s, the agricultural area has been relatively constant, although productivity has still increased owing to new agricultural methods and technological improvements (Rabbinge and Vanlatesteijn 1992; Rabbinge and van Diepen 2000).

Land use changes in the southern Baltic Sea catchment over the past 200 years are comparable to the developments in Europe and are strongly influenced by the industrialisation and fundamental socio-economic transformation after the Second World War and after the European socialist period. However, there are no studies on land use changes during the past 100 years for the Baltic Sea basin as a whole. For the southern, cultivated part of the basin, however, existing studies (Behrendt et al. 2008; Gadegast et al. 2012) of the Oder catchment may typify trends: land use was relatively constant before 1950. Agriculture and forests occupied about 66 and 25 % of the land area, respectively. Due to population growth, the urban areas increased slightly to 4 %. After the Second World War, a decreasing trend in agricultural land use began, and by the end of the communist era in Poland, agriculture accounted for 59 % of the land area. Areas with high yield were used more intensively, and the derelict land was converted to forest. Forested areas thus increased from 22 % in 1950 to 28 % in 1990 (Hirschfeld et al. 2009; Fig. 17.6). Following the transition from a state-

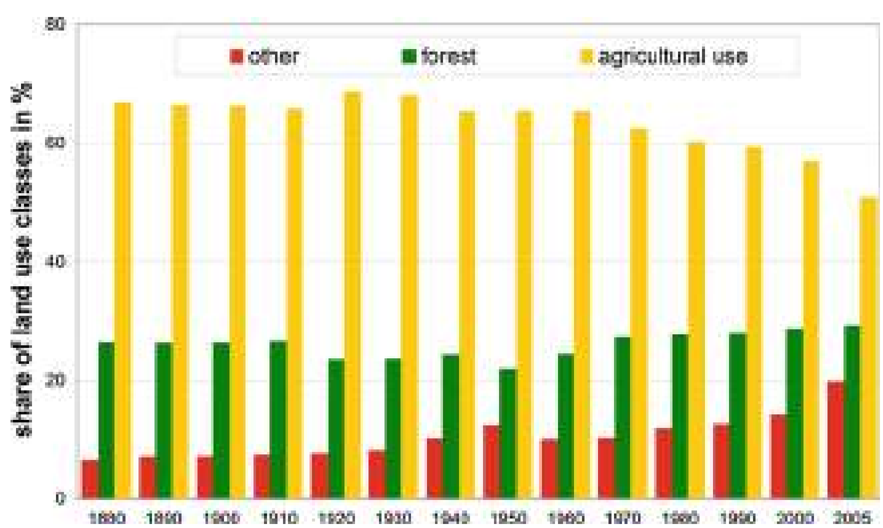

Fig. 17.6 Land use in the Oder River catchment over 1880–1940 (Gadegast et al. 2012) and for the whole of Poland over 1950–2005 (Hirschfeld et al. 2009)

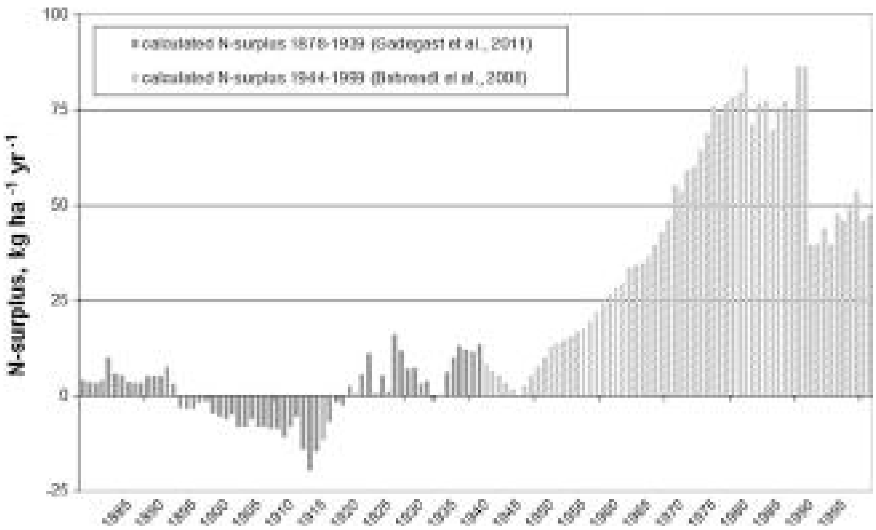

Fig. 17.7 Nitrogen (N) surplus on agricultural areas in the Oder River catchment, 1878–1939 (Gadegast et al. 2012) and 1944–1999 (Behrendt et al. 2008)

controlled to a free-market economy, agriculture experienced a recession. However, after Poland and the Czech Republic joined the EU in 2004, agricultural land use has again intensified.

The long-term N surplus in the Oder River catchment is shown in Fig. 17.7 for 1878–1939 (Gadegast et al. 2012) and 1944–1999 (Behrendt et al. 2008). Up to 1890, the mean N surplus was around 5 kg N ha^{-1} year^{-1}. Between 1890 and 1920, the N balance was negative for three reasons: higher crop yields and therefore a higher N uptake; use of human excrement as an organic fertilizer ceased; and a reduced input of mineral fertiliser. During the First World War especially, blockades prevented the import of products such as 'Chile saltpetre' and 'Guano'. Since 1920, the N surplus has increased. This is due to the invention of the Haber–Bosch process which enables the manufacture of inorganic fertiliser. The average application of mineral fertiliser on agricultural areas in the Oder River system increased from 9 kg ha^{-1} year^{-1} (1921) to 26 kg ha^{-1} year^{-1} (1938/39). After the end of the Second World War, N surpluses increased constantly due to intensification of agricultural production and increased use of mineral fertiliser until the collapse of agriculture in Poland and the Czech Republic in 1989.

Long-term change in DIN loads entering the Baltic Sea via the Oder River is shown in Fig. 17.8. DIN loads increased initially by about 43 % mainly due to growth in the urban population. Waste water treatment became necessary, and nutrient release from point sources such as sewer systems and waste water treatment plants increased. The increasing deposition of nitrogen oxides (NO_X) and NH_Y from the atmosphere due to growing industrialisation and the increasing N surplus in agriculture since 1920 influenced DIN loads to the Baltic Sea. The increase in DIN loads continued, almost tripling during 1955 to 1980 with a

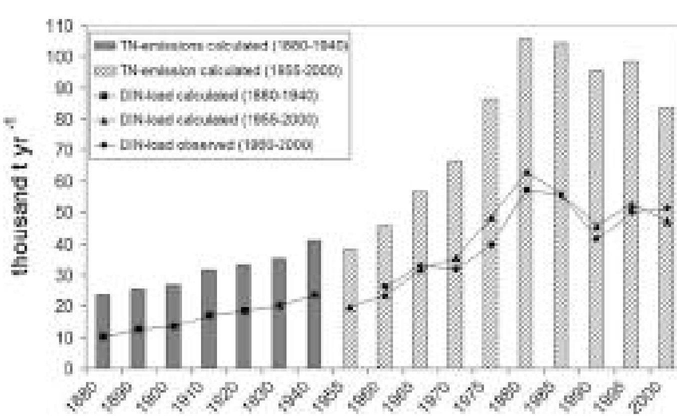

Fig. 17.8 Long-term change (1880–2000) in total nitrogen (TN) emissions to surface waters and in the calculated and measured loads of dissolved inorganic nitrogen (DIN) in the Oder River at the station Krajnik Dolny. 1880–1940 (Gadegast et al. 2012) and 1955–2000 (Behrendt et al. 2008; Venohr et al. 2010)

maximum of 63,000 t year^{-1}. DIN loads decreased around 1990 due to reduced N surpluses following the collapse of agriculture in Poland and the Czech Republic. From 1990 to 2000, DIN loads showed a slight increase as a result of a revitalisation of agriculture (Behrendt et al. 2008; Venohr et al. 2010).

17.5.1.2 Recent Trends in Nutrient Loads

Decreasing trends in riverine nutrient concentrations or loads have been reported. Raike et al. (2003) observed decreasing P concentrations in Finnish rivers and lakes formerly heavily polluted by industrial and municipal discharges. Similar observations were made in Estonia (Iital et al. 2010a, 2010b), Latvia (Stålnacke et al. 2003) and for the Nemunas river (Sileika et al. 2006) where the dissolved inorganic phosphorus (DIP) load decreased 31–86 % between 1986–1991 and 1997–2002. In Denmark, the TN load from point sources has reduced by 74 % since 1985, and the TN loads in 86 streams draining smaller agricultural catchments by 32 % (Kronvang et al. 2008). The decreases in TN concentrations in Estonian rivers relate to substantial reductions in fertiliser use, decreased agricultural land area, decreased point source load and increased self-purification capacity of soil water systems (Iital et al. 2010b). However, increasing TN concentrations are also being observed in some rivers due to higher diffuse loading, for example Finnish rivers (Raike et al. 2003) and the Nemunas river (Sileika et al. 2006). Overall, a trend analysis for the Baltic Sea on total waterborne annual loads (riverine loads plus direct coastal loads) indicates an increase in TN loads from 1994 to 2008, although this was not statistically significant, and a significant decrease (442 t year^{-1}) for TP loads (HELCOM 2011).

17.5.1.3 Agriculture and Weathering

Agricultural activities affect chemical weathering and with this probably the release of the nutrient Si from minerals in several ways. First, practices such as tilling and changes in land use alter chemical weathering fluxes positively (Paces 1983; Pierson-Wickmann 2009). An additional driver of increased chemical weathering fluxes is the increasing application of mineral fertilisers, specifically lime or carbonates (Tilman et al. 2001). Evidence for long-term (decades) trends in DIC fluxes have been identified for the Mississippi catchment, but to date have not been observed in the Baltic Sea area. Agricultural acidification, for example, associated with the application of N fertilisers may have increased fluxes of cations and decreased DIC fluxes in the past (Semhi et al. 2000; Perrin et al. 2008; Pierson-Wickmann et al. 2009). However, no studies have been identified showing this for the Baltic Sea area in detail. Mineral fertilisation increases crop production and the C-pool in plants (Ma and Takahashi 1990; Alvarez and Datnoff 2001). A hypothetical increase in terrestrial C-pools of standing stocks may contribute to an increased particulate organic carbon (POC) efflux to the Baltic Sea, specifically if precipitation patterns change towards those enhancing soil erosion. However, no data are known which justify this hypothesis. Moreover, there is no large-scale estimate on how agriculture has affected the fluxes of biogenic elements from sulphidic soils in the boreal watersheds of Sweden and Finland. On the coastal plains of Finland, approximately 3000 km^2 of acid sulphate soils have developed as a result of intensive agricultural drainage of waterlogged sulphide-bearing sediments (Åström et al. 2007).

17.5.2 Transformations of Nutrients Along the Aquatic Continuum

Retention is the permanent removal or temporary storage of nutrients and other biogenic elements within a system (von Schiller et al. 2008). Depending on the hydrological pathways along which biogenic elements are routed through the catchment, retention processes may significantly alter the concentration of these elements before they reach the marine recipient (Stålnacke et al. 2003).

In the terrestrial part of the hydrological cycle, retention processes include deposition of eroded soil and associated biogenic elements such as in buffer zones (Uusi-Kämppä 2006; Pärn et al. 2011), sequestration of C and nutrients into the organic soil pool (Lal et al. 2011), adsorption of dissolved P onto inorganic soil constituents (Litaor et al. 2003) and denitrification: a microbial dissimilation process in which dissolved nitrate is reduced to gaseous forms of N (Seitzinger 1988). Several studies have demonstrated very high N removal rates and high efficiency (up to 100 % removal) due to denitrification in groundwater-fed wetlands and wetlands subject to overland flow (Haycock and Pinay 1993; Sabater et al. 2003). N retention in groundwater is strongly dependent on hydraulic residence time. For 17 Danish catchments, Andersen et al. (2001) reported groundwater retention of 20–80 % along a gradient of increasing retention time. Minerogenic soils in general have a high affinity for P, and excess P is usually strongly sorbed in soils until a critical degree of saturation is reached (Hooda et al. 2000). Most north-western European countries, however, experience a net input of P to agricultural land (Leinweber et al. 2002), which makes the soil more vulnerable to P losses via erosion and leaching (Sharpley and Rekolainen 1997). As an example, the average P content in Danish agricultural soils increased from 3200 to 4600 kg P ha^{-1} between 1900 and 2000 due to the intensification of agriculture and increased use of fertilisers (Rubæk et al. 2005).

In the aquatic part of the hydrological cycle, that is river systems including rivers, lakes, riparian areas and floodplains, N retention processes include biotic assimilation, denitrification and sorption (Herrman et al. 2008). Recent studies on agricultural streams with high nitrate concentrations indicate that in-stream nitrate removal may not increase proportionately with nitrate availability due to nitrate saturation of the microbial community responsible for denitrification (Bernot et al. 2006; Herrman et al. 2008). The study by Herrman et al. (2008) suggested a concentration of 2 mg NO_3–N l^{-1} as a threshold, above which the streams become increasingly saturated with nitrate and export substantial N. Other research has identified hydraulic retention time (in-stream water residence time) as a key stream characteristic controlling N removal (Valett et al. 1996). In lakes, permanent N retention occurs as denitrification but also by incorporation in sedimenting organic matter that is permanently buried on the lake bottom (Søndergaard 2007). As for streams, hydraulic retention time is found to be a key factor controlling N retention in lakes (Windolf et al. 1996; Søndergaard 2007; Herrman et al. 2008). On average for 69 Danish lakes, 43 % of the N input was permanently retained (Jensen et al. 1990). Based on data reported by contracting countries, HELCOM (2004) estimated a 30 % N retention of the gross load entering river systems in the Baltic Sea drainage basin.

Retention processes for P in rivers, riparian areas and floodplains comprise sorption to suspended solids and bottom sediment, deposition of particulate matter on the river bed, biotic assimilation by algae and macrophytes, and sedimentation on inundated riparian areas and floodplains (Kronvang et al. 1999; Pärn et al. 2011). Storage of P within the river is often considered a temporary sink only, as the P build-up in biomass and sediment during summer is flushed out during high winter flows (De Witt 1999; Schulz et al. 2003). Permanent in-stream retention was explained by Svendsen et al. (1995) as sorption of DIP to Fe and Al oxides and hydroxides. In-stream net retention is probably of minor importance (Vassiljev and Stålnacke 2005), whereas P storage by sedimentation on inundated riparian areas and floodplains can be considerable; rates of up to 127 kg P ha^{-1} $year^{-1}$ are reported (review by Hoffmann et al. 2009). In lakes, P retention occurs via sedimentation of particulate-bound forms or via uptake and incorporation of dissolved P by plants and subsequent sedimentation (Søndergaard 2007). In the drainage areas of entire river systems, P retention in lakes is often considered the most important permanent sink (Svendsen et al. 1995; Vassiljev and Stålnacke 2005). Kronvang et al. (1999) found an average retention of 3 kg P ha^{-1} $year^{-1}$ for 18 shallow Danish lakes. Vassiljev and Stålnacke (2005) using a model approach estimated lake P retention to be around 30–35 % in the 44,000 km^2 Lake Peipsi catchment. However, the retention rate in some lakes is currently negative due to high internal P loading from the sediment following a reduction in the external load (Jeppesen et al. 1999). For the Baltic Sea drainage basin, HELCOM (2004) estimated that on average 31 % of the gross P load to river systems is retained.

Retention is not constant over time. Hoffmann et al. (2000) estimated that in Sweden a retention capacity of 30,000 t N has been lost since 1865 due to extensive drainage of wetlands and lakes. A similar loss of nutrient retention capacity is seen in other cultivated areas of the drainage basin (Brookes 1987; Andersen and Svendsen 1997). Over the past 30 years, recreation of the natural nutrient retention capacity of river systems by remeandering of streams allowing temporary inundation of riparian areas and floodplains, and recreation of wetlands and lakes have attracted much attention as means to reduce diffuse nutrient loading (Cooke et al. 1993; Hoffmann et al. 2011).

17.5.3 Climate Impacts on Waterborne Losses from Agriculture and Urban Areas

Rising temperatures have prolonged the growing season by 4 weeks or more across the southern part of the Baltic Sea drainage basin since 1982 (Høgda et al. 2007). There have been more droughts in large parts of western and eastern Europe (Trenberth et al. 2007); however, in Denmark, precipitation has increased by 100 mm over the past 50 years (Larsen et al. 2005). Rainfall intensity has increased (Frich et al. 2002), and this has led to severe summer flooding in Europe (Christensen and Christensen 2003). Analysis of data from 17 catchments across Europe (Bouraoui et al. 2009) revealed that climatic variables and in particular total precipitation explained most of the variance found in the nutrient load measured at the catchments outlet. DIN concentration was mainly controlled by the extent of the agricultural area, whereas P concentration was mostly controlled by precipitation intensity (calculated here as the amount of rainfall during a rainy day) and population density. Farmers are currently adapting to climate change, particularly in terms of changing the timing of cultivation and selecting other crop species and cultivars (Olesen et al. 2011). For example, the cultivation of maize, a heat-demanding, warm-season crop, is expanding northwards and increased by a factor of 9 in Denmark from 1990 to 2010 (Fig. 17.9).

The projected climate change—higher temperatures, greater precipitation in some areas, and more extreme events—will affect all hydrologic pathways for biogenic elements and thus loading to the Baltic Sea.

Agro-ecosystems are strongly affected by environmental conditions and thus by climate change. Plants respond to rising atmospheric CO_2 concentration by increasing resource use efficiencies for radiation, water and N (Olesen and Bindi 2002), which reduces the risk of N leaching. In experimental studies, Downing et al. (2000) showed a wheat grain yield

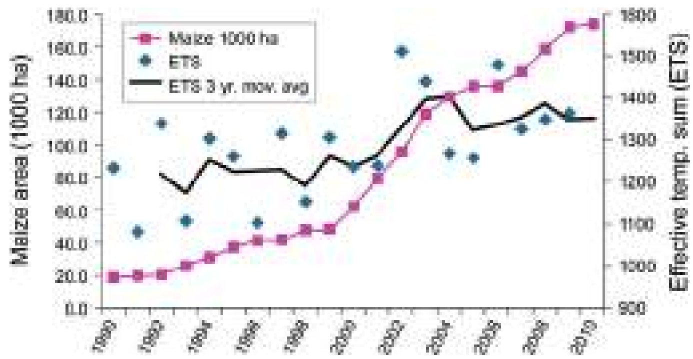

Fig. 17.9 Area with maize in Denmark and effective temperature sum (*ETS*). ETS is calculated as the sum of daily mean temperatures above 6 °C from 15 April to 30 September. Maize requires an ETS above 1200 °C (redrawn from Olesen 2008)

increase of 28 % for a doubling of current CO_2 concentration. However, increased temperature reduces crop duration for many annual crops and hence yields. This could lead to a further expansion of warm-season crops (e.g. maize and sunflower) into areas currently dominated by small-grain cereals and oilseed crops (Olesen and Bindi 2002). Earlier harvest of crops and later planting of winter crops may result in a prolonged period of bare soil in autumn (Jeppesen et al. 2011), which would increase the risk of nutrient loss by leaching and surface loss processes. In addition, soil organic matter turnover would increase under higher temperatures, which would increase the risk of leaching of nutrients and DOC (Patil et al. 2010), particularly in connection with heavy precipitation (Eckersten et al. 2001). Soil erosion and surface run-off are also expected to increase (Michael et al. 2005; Nearing et al. 2005) and thus loss of nutrients and other biogeneic elements to river systems.

The current intensive agricultural production of cereals in most Danish riparian areas is becoming increasingly difficult to sustain due to increased autumn and winter precipitation (Jeppesen et al. 2011). The problems could increase considerably under the projected climate change, and in many riparian areas, intensive agricultural production may cease in the future (Andersen et al. 2006). Abandoning cultivation and artificial drainage of riparian areas would decrease nutrient losses and aid C sequestration (Lal et al. 2011).

17.5.4 Current and Future Export to the Baltic Sea

As temperature is one of the parameters controlling denitrification, the rate of denitrification in soils, streams, wetlands and lakes will increase with increasing temperature (Veraart et al. 2011). Using a nutrient accounting approach covering all major watersheds of the Baltic Sea basin, Hong et al. (2012) found that the fraction of net anthropogenic N and P inputs exported as riverine fluxes decreased with increasing temperature (Fig. 17.10), that is the overall watershed nutrient retention which is about 70 % for all anthropogenic N and about 90 % for all anthropogenic P increased with temperature. On the other hand, Hong et al. (2012) found the fraction of net anthropogenic N and P inputs exported as riverine fluxes increased with river discharge, and that is projected to increase in future (Graham and Bergstrom 2000). Similar observations were found by Howarth et al. (2012) using a global dataset of 154 watersheds, of which 36 were in Sweden. Thus, the effect of higher nutrient retention in a warmer climate may be counteracted by a shorter hydraulic retention time following increased percolation through soils and run-off through river systems (Jeppesen et al. 2011). Increased run-off during winter and more extreme rainfall events will affect the interaction between streams and their riparian areas. Andersen et al. (2006) projected a 50 % increase in the number of days with overbank flooding by a lowland river by 2071–2100 under the IPCC A2 SRES scenario, relative to the control period (1961–1990). Because the P deposition rate increases with the magnitude of the inundation event (Kronvang et al. 2007), P retention may increase in a warmer climate. In lakes, a temperature-mediated release of Fe-bound P from the sediment has been observed (Jeppesen et al. 2009). Longer stagnation periods in lakes could increase anoxia and sediment P release. Nutrient depletion in the epilimnion in summer can shift phytoplankton blooms towards autumn when a deepening thermocline allows nutrient inputs from deeper layers (Kangro et al. 2005). As a result, the internal P loading of lakes may increase and the net P retention in lakes may decrease in a warmer climate.

To date, there are few large-scale studies evaluating possible future N loadings from the Baltic Sea catchment area as a whole. Using a multivariate statistical approach, Eriksson-Hägg et al. (2012) estimated that the net effect of changes in climate (i.e. hydrology) and in animal protein consumption may be a possible increase in TN flux ranging from 3 to 72 % (Fig. 17.11) between the major watersheds,

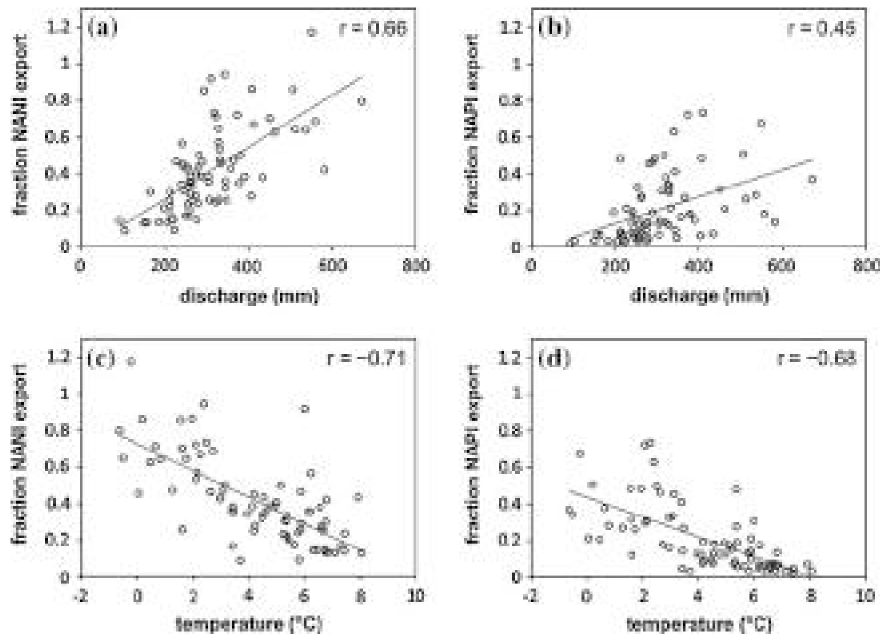

Fig. 17.10 Fraction of net anthropogenic nitrogen inputs (NANI) and net anthropogenic phosphorus inputs (NAPI) exported as riverine fluxes controlled by discharge and temperature (Hong et al. 2012)

on average 20–30 %. Overall, the authors concluded that demand for animal protein will be fundamental to controlling nutrient loading in the Baltic Sea ecosystem and may be a major block to fulfilling the environmental goals of the Baltic Sea Action Plan.

17.6 Hydrological Alterations and Freshwater Biogeochemistry

17.6.1 Hydrological Alterations and Waterborne Fluxes

Many rivers in the Baltic Sea catchments have been regulated. In general, the boreal/sub-Arctic rivers entering the Gulf of Bothnia have a higher specific run-off, especially the northern Swedish rivers which have a steeper catchment slope compared to the rivers draining the south-eastern catchments of the Baltic Sea. Damming is therefore more frequent in the boreal rivers owing to its higher effectiveness in terms of power generation; major reservoirs located in the headwaters can hold 30–70 % of their annual water discharge (Dynesius and Nilsson 1994). In contrast, damming is less frequent in the lowland rivers of the south-eastern catchment of the Baltic Sea, and minor dams and reservoirs with short water residence times were mostly built there. Studies in major global rivers have demonstrated that river regulation and damming lead to decreased nutrient transport to coastal seas, although the mechanisms vary (Humborg et al. 2000). The primary mechanism responsible for the decrease in nutrient loads was initially thought to be the trapping and subsequent sediment burial of nutrients in the form of diatoms that were autochthonously produced in the reservoirs due to decreasing water currents and improved light conditions, termed the 'artificial lake effect' (Van Bennekom and Salomons 1981). Most studies focussed on dissolved Si (DSi) fluxes, because this nutrient removal might be overcompensated by anthropogenic N and P inputs downstream of the reservoir and from other sources, and no such compensation occurs for DSi (Fig. 17.12). Thus, river regulation and damming lead mainly to lower N:Si ratios in the receiving coastal water bodies favouring non-siliceous phytoplankton blooms (Humborg et al. 1997; Dai et al. 2011).

Initial studies from the early 1990s comparing unregulated and regulated boreal rivers in the Baltic Sea catchment showed that overall the element concentrations (i.e. base cations and anions, trace metals and nutrients) are much lower in regulated rivers than unregulated rivers (Brydsten et al. 1990). From later studies in boreal and sub-Arctic Swedish watersheds, it was hypothesised that perturbed surface water–groundwater interactions as a consequence of hydrological alterations in the rivers of the northern Baltic Sea watershed lead to changes in weathering conditions. Thus, less weathering may be the major reason for the reduced DSi loads observed in these oligotrophic boreal rivers (Humborg et al. 2000, 2002, 2006). Particle trapping of biogenic Si (BSi)—mainly diatom shells—behind dams is the main reason for the reduced Si loads in the cultivated watersheds of the southern catchment area of the Baltic Sea (Humborg et al. 2006).

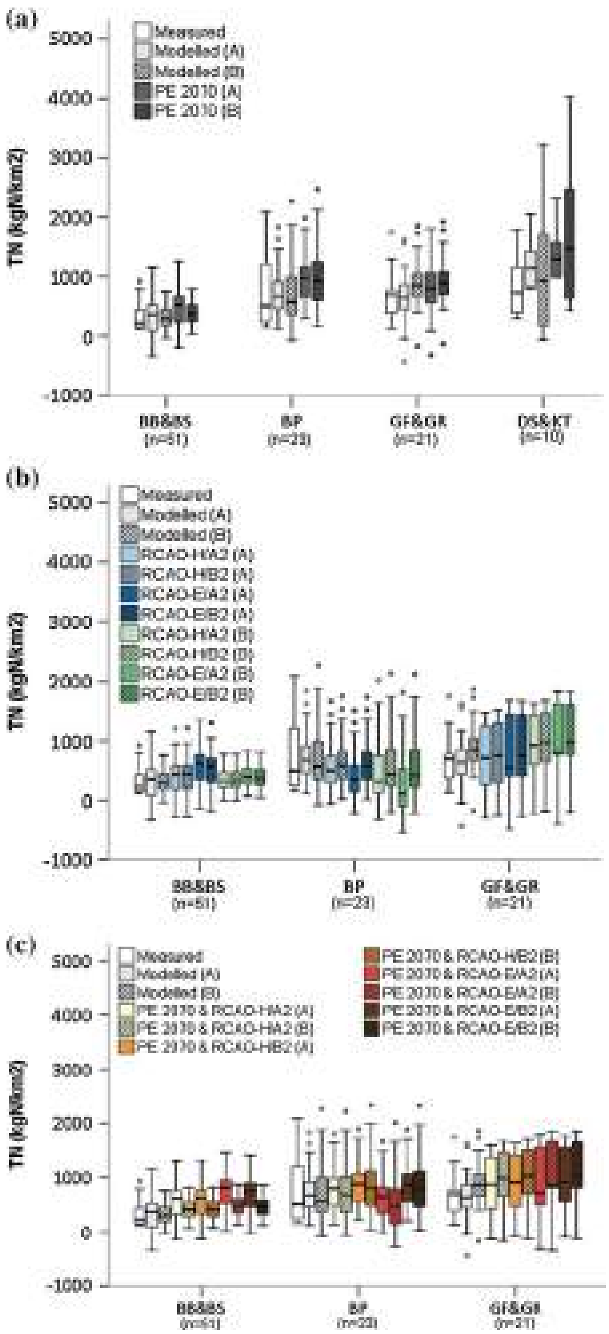

Fig. 17.11 Box and whisker plots with measured (mean 1992–1996), modelled (1992–1996) and projected future total nitrogen (TN) fluxes (kg N km^{-2} year^{-1}) for the four sub-basins: Bothnian Bay and Bothnian Sea (BB & BS), Baltic Proper (BP), Gulf of Finland and Gulf of Riga (GF & GR), and Danish straits and Kattegat (DS & KT) using the all-catchment regression (A) or the basin-specific regression (B), non-area weighted data. **a** Primary emissions scenario PE 2070. **b** Four climate scenarios: the regional climate model RCAO driven by two general circulation models (HadCM, ECHAM5) and two IPCC SRES scenarios (A2 and B2). **c** The net scenarios with both changed primary emissions and changed climate (Hägg et al. 2010)

17.6.2 Climate Impacts on Regulated Rivers and the Effect on Water-borne Fluxes

Damming of rivers has led to a moderate increase in lake area; for most Swedish watersheds, the share of the total watershed covered by man-made lakes has increased by only a few percentage (Smedberg et al. 2009). Most regulated rivers show reduced seasonality in water discharge compared to non-regulated rivers due to the controlled use of reservoir water, but total water discharge has not decreased despite the potential increase in evaporation as a result of increased lake area (Carlsson and Sanner 1994). Whether a change in climate will alter evapotranspiration patterns in regulated watersheds is not yet known, but with the projected increase in precipitation in the boreal part of the Baltic Sea catchment which is heavily dammed, overall water discharge to the Baltic Sea may also increase in the regulated rivers. Warmer summers in a future climate could cause a significant increase in evaporation and affect the seasonal discharge patterns of the regulated watersheds.

Hydropower accounts for 19, 47 and 77 % of electricity generation in Finland, Sweden and Latvia, respectively; for the other riparian countries, hydropower is less important, generating less than 5 % of electricity (Lehner et al. 2005). Model studies using different regional climate model (RCM) simulations to examine impacts on hydrology for the Luleälven watershed in Sweden revealed an overall increase

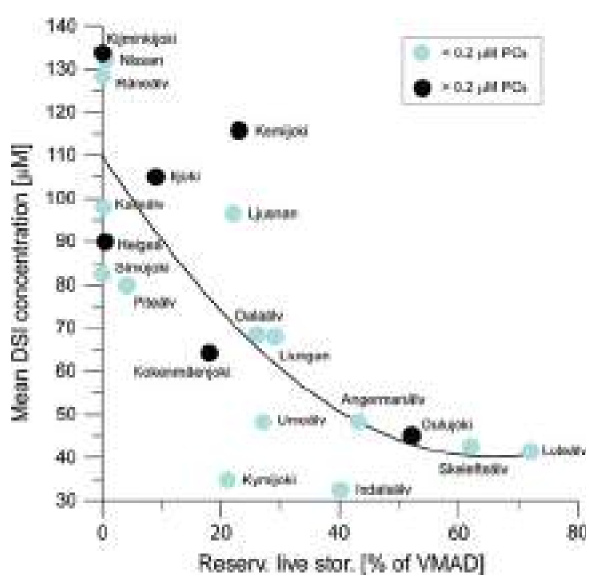

Fig. 17.12 Dissolved silicon (DSi) concentration versus reservoir life storage in regulated Swedish and Finnish rivers (redrawn from Humborg et al. 2000)

in river flow, earlier spring peak flows and an increase in hydropower potential (Graham et al. 2007). A pan-European study presented a model-based approach for analysing the possible effects of global climate change on Europe's hydropower potential at a country scale (Lehner et al. 2005). The results indicated that even following moderate climate and global change assumptions, major change in discharge regimes could be expected, leading to a potential increase in hydropower potential of 15–30 % for Finland, Sweden and Latvia. It is not yet known whether this potential could be exploited by means of new reservoirs and dams or by increasing existing dams.

17.6.3 Current and Future Export from Regulated Rivers

Annual riverine DSi loads to the Baltic Sea may have decreased by about 400,000 t year^{-1} or by a third over the past 100 years. This estimate is based on two independent approaches, one addressing the hydraulic load or water residence time as a key variable for nutrient retention in aquatic systems (Humborg et al. 2008b) and the other comparing DSi yields (t DSi km^{-2} year^{-1}) in relatively unperturbed watersheds and regulated watersheds (Conley et al. 2008). In addition to deposition of BSi behind dams, river regulation leads to less weathering as shown in a model study by Sferratore et al. (2008). Lower Si fluxes were observed in River Luleälven compared to the neighbouring River Kalixälven, although specific discharge is much higher in the former despite a similar geological setting. Since damming of the River Luleälven changed the pathways of waters, that is third- and fourth-order streams were affected by regulation, surface water–groundwater interactions in these streams were interrupted, leading to less DSi input from groundwater; diatom blooms reducing DSi concentrations behind dams could be ruled out as diatom growth is relatively low in these ultra-oligotrophic rivers. However, in the southern eutrophic watersheds, the BSi (silica bound in diatom shells) concentration can be up to 100 μM and, thus, significant for the overall riverine Si transport to the Baltic Sea (Humborg et al. 2006). For other dissolved constituents such as nutrients and DOC, no large-scale estimates of changes in river load as an effect of hydrological alterations exist.

17.7 Conclusion

Although the effect of climate change cannot yet be quantified on a Baltic Sea basin-wide scale, some key findings can be drawn:

- Long-term trends in the timing of ice freeze-up and ice break-up indicate shorter ice-cover duration in many boreal watersheds. Observations on the discharge regime of major boreal rivers reveal no change in mean annual flow but a change in the seasonal distribution of streamflow with potentially large impacts on the redistribution of organically bound C and nutrients from land to the Baltic Sea. Winter and spring mean monthly discharge increased at many observation sites, and the peak flow in spring has become earlier. Areas with a mean annual temperature around 0 °C (i.e. around 61°N) are most sensitive to further warming.
- Over past decades and especially in the 1970s and 1980s, atmospheric deposition probably had a stronger effect on freshwater biogeochemical conditions in the Baltic Sea drainage area than climate. This pattern seems to change as atmospheric deposition decreases, however. It is likely that control of freshwater biogeochemistry will shift back from dominance by atmospheric deposition to dominance by climate. If this occurs, biogeochemical conditions in freshwaters and the Baltic Sea could change rapidly.
- Numerous studies indicate that forests and wetlands play a key role in the terrestrial export of organic-bound nutrients and C. Ditching and clear-cutting often result in increased concentrations and leaching some years after the treatment has ended, but over the long term and after re-establishing forest, this could result in lower water table levels and decreased leaching. However, current forest management only affects small parts of the overall area covered by boreal forests each year and the effect on fluxes of organic-bound nutrients and C to the Baltic Sea is probably minor. Over the short term, climate change is unlikely to affect the spatial distribution of wetlands, except for palsa mires that cover too small an area in the boreal watersheds to be significant for element fluxes to the Baltic Sea. Results from small-scale field and modelling studies indicate that increased temperature and precipitation could increase DOM transport to the Baltic Sea significantly. However, the large-scale impacts on the Baltic Sea basin are still unknown. Even a northward shift in boreal forest (i.e. Norwegian spruce) with climate change may alter quite significantly the biogeochemistry of the northernmost rivers over the long term. Modelling studies of the effect of changes in vegetation cover and structure for river loads to the Baltic Sea are ongoing but have not yet been published.
- Agricultural practices and urban sources significantly increased N and P concentrations in the rivers draining the cultivated watersheds of the southern Baltic Sea catchment. Nutrient loads from these rivers to the Baltic Sea increased several fold over the past 150 years, peaking in the 1970s and 1980s. A slight decrease is seen in P loads only over recent years probably due to improved sewage treatment from urban areas, especially in Poland. Changes in climate are not uniform across the

cultivated southern catchment area; in the south-western part (i.e. Denmark and western parts of Germany), precipitation has increased since the 1980s and farmers are currently adapting to a warmer and wetter climate by selecting heat-demanding and nutrient-demanding crops like maize. Whether increased fertiliser use, which may even occur in the transitional countries like Poland and the Baltic States, as a result of the European Common Agriculture Policies or changes in lifestyle will lead to an increased nutrient flux to the Baltic Sea is still unknown. This is because water discharge especially in the south-eastern part of the catchment is projected by some modelling studies to decrease, and the retention of nutrients (denitrification of N in soils and groundwater, river bank deposition of P) may increase. Both processes would largely compensate for the projected increase in fertiliser applications that may increase nutrient leaching. However, there are still too few catchmentwide modelling studies on water discharge and observation and modelling of retention patterns are too small scale to allow overall estimates of these processes for the Baltic Sea. Initial studies (that still need further scientific elaboration) indicate that TN fluxes to the Baltic Sea may increase up to 70 % as a result of changes in water discharge and changes in lifestyle (including change in the demand for animal protein).
- Hydrological alterations have lowered the nutrient flux to the Baltic Sea, especially for Si. Overall, the dissolved Si fluxes to the Baltic Sea have decreased by a third over the past 100 years as a result of damming and eutrophication of natural lakes. The projected increase in precipitation in the boreal part of the Baltic Sea catchment could increase hydropower potential by 15–30 %. It is unlikely that the potential formation of new lakes by damming would change the amount of water entering the Baltic Sea through rivers; however, seasonal patterns in water discharge could change significantly as they become smoothed through the operation of the dams (i.e. saving water during spring freshet and using water during winter).

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18 Environmental Impacts—Marine Biogeochemistry

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Abstract

Marine biogeochemistry deals with the budgets and transformations of biogeochemically reactive elements such as carbon, nitrogen and phosphorus. Inorganic nitrogen and phosphorus compounds are the major nutrients and control organic matter (biomass) production in the surface water. Due to various anthropogenic activities, the input of these nutrients into the Baltic Sea has increased drastically during the last century and has enhanced the net organic matter production by a factor of 2–4 (eutrophication). This has led to detrimental oxygen depletion and hydrogen sulphide production in the deep basins of the Baltic Sea. Model simulations based on the Baltic Sea Action Plan (BSAP) indicate that current eutrophication and thus extension of oxygen-depleted areas cannot be reversed within the next hundred years by the proposed nutrient reduction measures. Another environmental problem is related to decreasing pH (acidification) that is caused by dissolution of the rising atmospheric CO_2. Estimates indicate a decrease in pH by about 0.15 during the last 1–2 centuries, and continuation of this trend may have serious ecological consequences. However, the concurrent increase in the alkalinity of the Baltic Sea may have significantly counteracted acidification.

18.1 Introduction

Marine biogeochemistry deals with the internal transformation of sea water constituents and their transport into and out of the sea area in question. Internal processes take place against the background of the hydrographic setting and are mainly controlled by biological production and the decomposition of organic matter. The major elements involved in these processes and addressed in this chapter are carbon (C), nitrogen (N), phosphorus (P) and oxygen (O_2). Their distributions and concentrations strongly influence the Baltic Sea ecosystem. On an annual timescale, the formation of organic matter is widely limited by the availability of N and P, which are the major nutrients. This in turn affects the O_2 conditions in deeper water layers, where organic matter is mineralised and thus consumes O_2 and may eventually lead to the formation of highly toxic hydrogen sulphide (H_2S). Hence, eutrophication is a major issue when assessing past and possible future changes. Carbon not only plays a central role as backbone for any organic material, but its inorganic form, carbon dioxide (CO_2), largely controls the marine acid/base balance, which due to the ongoing rise in atmospheric CO_2 levels is gradually shifting towards more acidic conditions.

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_18

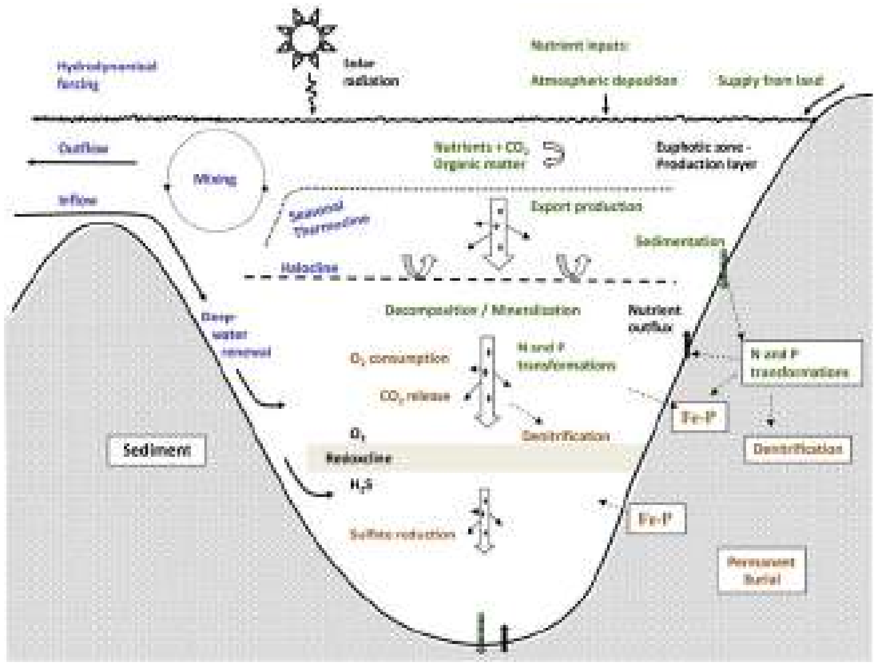

Fig. 18.1 Schematic overview of important biogeochemical and forcing processes within the Baltic Sea

18.2 Major Biogeochemical Fluxes and Transformations

This section describes the main components of the biogeochemical cycles of nutrients, carbon and oxygen in the Baltic Sea (Fig. 18.1) and includes a brief introduction to available observations and biogeochemical models. The biogeochemical cycle starts with photosynthetic primary production by phytoplankton in the sunlit surface layer (euphotic zone). Production is controlled by the availability of nutrients supplied by internal recycling and input from surrounding land areas, the atmosphere and the North Sea. The productive season and the biogeochemical cycles in the Baltic Sea are also regulated by the physical processes of transport, mixing and stratification (such as deep-water renewal and the development of a seasonal thermocline). The essential nutrients are dissolved inorganic nitrogen (DIN: ammonia NH_3 (at equilibrium with ammonium, NH_4^+), nitrite NO_2 and nitrate NO_3) and dissolved inorganic phosphorus (DIP: phosphate PO_4). However, some evidence also exists for the use of dissolved organic N and dissolved organic P as nutrients. In addition, nitrogen (N_2) fixation by cyanobacteria may add new bioavailable N to the Baltic Sea.

Sinking organic particles export nutrients below the productive surface layer. When this organic matter is decomposed, O_2 in the deeper parts of the Baltic Sea is gradually consumed and the water becomes anoxic if it is not renewed. In the absence of O_2, sulphate reduction produces H_2S that is toxic to higher life forms. The interplay between production and mineralisation of organic matter is also reflected in the cycling of CO_2 since the formation of organic matter uses up CO_2, while, conversely, mineralisation of organic matter produces CO_2 that accumulates in the deep water. A fraction of the export production reaches the sediments before being completely decomposed and mineralised, and permanent burial of organic material is an important sink especially for P. Depending on the redox conditions, P may also be bound to hydrated metal oxides (e.g. iron, Fe) in the sediments. Currents and waves may cause resuspension of sediments in shallow areas of the Baltic Sea. This process leads to a transport of lighter material and organic matter towards deeper parts of the Baltic Sea.

The major N sink in the Baltic Sea is denitrification, an anaerobic mineralisation of organic matter that transforms NO_3 to N_2. While the removal of N is enhanced at lower O_2 concentrations, the sediment P sink is weakened and may even become a source under anoxic conditions, when older mineral-bound P can be released to the overlying water. Finally, nutrients supplied from external sources that are not permanently removed or accumulated inside the Baltic Sea will be exported to the Kattegat and Skagerrak.

Box 18.1 Basis for current knowledge
For over a century, the Baltic Sea has been one of the most heavily investigated seas in the world and much of the information summarised, for example in the books by Wulff et al. (2001), Feistel et al. (2008) and Leppäranta and Myrberg (2009), is based on analyses of historical data. Information on the physical, chemical

and biological variables of the Baltic Sea is regularly gathered by national monitoring programmes in Denmark, Estonia, Finland, Germany, Latvia, Lithuania, Poland, Russia and Sweden. International cooperation between these countries and the European community is governed by the Baltic Marine Environment Protection Commission (HELCOM, www.helcom.fi) that provides data services and conducts environmental assessments based on the data collected. Other international sources of data and information concerning the Baltic Sea include the International Council for the Exploration of the Sea (ICES, www.ices.dk) and the European Monitoring and Evaluation Programme (EMEP, www.emep.int) that provides information about atmospheric emission and deposition of nutrients. In 2009, the Baltic Nest Institute, Stockholm University (www.balticnest.org), and several other institutes started a partnership on a system of distributed databases where each partner (see http://nest.su.se/bed/hydro_chem.shtml) makes the data hosted in its databases publicly available via a data portal (http://apps.nest.su.se/dataPortal/). The Baltic Nest Institute also provides access to the data portal via Data Assimilation System (DAS), http://nest.su.se/das/) and a decision support system Nest (http://apps.nest.su.se/nest/). The portal also provides access to all data collected in the Baltic Environmental Database (BED), maintained by the Baltic Nest Institute. Near-real-time observations are also available and distributed within the countries bordering the Baltic Sea by the partners participating in the Baltic Operational Oceanographic System (BOOS, www.boos.org).

The national and international funding of climate- and eutrophication-related research has been important drivers for an increased understanding of Baltic Sea biogeochemistry. For example, Wulff et al. (2001) presented results from international research projects that were funded by the Baltic Sea System Study (BASYS) and the EU Marine Science and Technology programme (EU/MAST). The combined scientific and political efforts have resulted in strategies to combat eutrophication in the Baltic Sea, for example a recent agreement to reduce nutrient loads (BSAP), HELCOM 2007). A current major international Baltic Sea research collaboration is the Joint Baltic Sea Research and Development Programme (BONUS, www.bonusportal.org), funded by the EU's Seventh Framework Programme for Research (FP7, http://ec.europa.eu/research/fp7/) and national research-funding institutions in the Baltic Sea countries. Key results of the 16 projects of the BONUS 2007 are summarised on the website www.bonusportal.org/about_bonus/bonus_and_era-net/bonus_2009-2011/bonus_projects. The overall biogeochemical cycling of nutrients and carbon in the Baltic Sea is especially addressed by the BONUS projects ECOSUPPORT and BALTIC-C.

Current understanding of how the interactions between the external forcing and the internal physical and biogeochemical processes regulate the ecological status of the Baltic Sea has been synthesised, quantified and evaluated by model experiments. Numerical models (e.g. Fennel and Neumann 2004) are also used to study the effects of climate change on biogeochemical cycles and to evaluate various management strategies (see Sect. 18.5). The development of more complex coupled physical–biogeochemical models started with horizontally integrated basin models (Stigebrandt and Wulff 1987), and coupled models dividing the Baltic Sea into a number of dynamically interconnected sub-basins are still used because of their computational efficiency (Gustafsson 2012; Gustafsson et al. 2012). With the increasing availability of computational power, three-dimensional coupled models for the Baltic Sea that produce high-resolution results on long timescales (centuries) are now state of the art at the larger institutes (e.g. Neumann et al. 2002; Eilola et al. 2009). Generally, the biogeochemical functions of the coupled models are similar in that they describe transformations of N and P including inorganic nutrients and particulate organic matter comprising phytoplankton, dead organic matter and zooplankton. Organic matter is typically produced from the inorganic nutrients by several functional groups of phytoplankton that may differ between models, but in the Baltic Sea usually include diatoms, flagellates and cyanobacteria. Sinking organic material enters the model sediment as benthic N and P. Differences between biogeochemical models may, for example, be due to parameterisations of nutrient fluxes between sediment and water and also in the rates and parameterisations of transformations of nutrients within the water column.

18.3 Changes in External Forcing

External forcing has a strong impact on biogeochemical processes and internal fluxes. This refers both to the climate-driven physical forcing and to the input of biogeochemically reactive substances via rivers, water exchange with the North Sea and the atmosphere. The physical forcing controls the water transport, stratification, ice coverage, temperature and salinity in the Baltic Sea, which in turn affect the distribution

Table 18.1 Waterborne and atmospheric input of total nitrogen and total phosphorus to the Baltic Sea for year 2000

	Waterborne input (kt $year^{-1}$)	Atmospheric input (kt $year^{-1}$)	Total input (kt $year^{-1}$)
Total nitrogen	641[a]	198[b]	839[a]
Total phosphorus	28[a]	1.5[a]	30[a]

([a] HELCOM 2011; [b] Bartnicki et al. 2011)

of nutrients and C and thus have an impact on biogeochemical processes. Changes in hydrographic conditions, such as rising surface water temperature and changes in ice cover, are discussed in Chaps. 7 and 8 and will be addressed here only if necessary for interpreting changes in biogeochemical cycles. Changes in the biogeochemical forcing mainly relate to the input of the main biogeochemical actors: nutrients, organic C and CO_2 (see also Chap. 17).

18.3.1 Nutrient Inputs

Compilations of data concerning the waterborne input (riverine input and point sources) and the atmospheric input of N and P into the entire Baltic Sea including the Kattegat are available through various HELCOM publications. The input data refer to both the inorganic and organic forms of N and P. Whereas the inorganic species are readily available for biological production, it is still an open question how much of the organic fraction is bioavailable and thus contributes to eutrophication. A summary of an input estimate (HELCOM 2005) based on 2000 data was given in the first assessment of climate change in the Baltic Sea (BACC Author Team 2008). An updated estimate based on data for 2001–2006 was recently published (HELCOM 2011). According to this later report, the annual waterborne inputs are 641,000 t N $year^{-1}$ and 28,700 t P $year^{-1}$. Adding an atmospheric deposition of 198,000 t N $year^{-1}$ (Bartnicki et al. 2011) yields a total N input to the Baltic Sea of 840,000 t N $year^{-1}$ (Table 18.1). According to HELCOM (2011), the atmospheric P deposition contributes 5 % to the total P input (1500 t P $year^{-1}$) resulting in a total input of 30,200 t P $year^{-1}$ (Table 18.1). These estimates differ only slightly from the values given in the previous HELCOM (2005) report and presented in the first assessment of climate change in the Baltic Sea (BACC Author Team 2008).

Attempts have been made to estimate nutrient inputs at the beginning of the last century when the influence of human activities was low (Larsson et al. 1985; Savchuk et al. 2008). Gustafsson et al. (2012) used data from different sources to reconstruct the N and P inputs to the various basins of the Baltic Sea including the Kattegat since 1850 (Fig. 18.2). For the period 1970–2006, annual land-borne loads for total N and total P, which include the input from rivers and point sources, are based on data compiled in the BED (Savchuk et al. 2012). Riverine inputs for the years before 1970 were reconstructed from historical data and present-day nutrient concentrations in pristine rivers in northern Scandinavia. The atmospheric deposition of N compounds for 1970–2006 was obtained both from measurements and model calculations. Prior to 1970, historical data were used to roughly estimate the atmospheric N input. Since neither regular monitoring data nor model simulations exist for the atmospheric deposition of P to the Baltic Sea, the values presented by Gustafsson et al. (2012, Fig. 18.2), which are based on deposition estimates for northern Europe (Ruoho-Airola et al. 2012 and references therein), differ from the HELCOM (2011) estimates (Table 18.1). Owing to the sparse database, the reconstruction of historical nutrient inputs is necessarily associated with a considerable uncertainty which is difficult to quantify.

Figure 18.2 indicates that N and P inputs have both increased dramatically during the last 150 years. This is due to human activities such as intensified agriculture and increasing industrialisation. The rate of increase became stronger after the Second World War and peaked around 1980 at about three (N) and seven (P) times higher than in 1850. However, inputs have decreased significantly since the 1980s due to emission reduction measures and are presently at a level comparable to those around 1960. When evaluating land-borne and

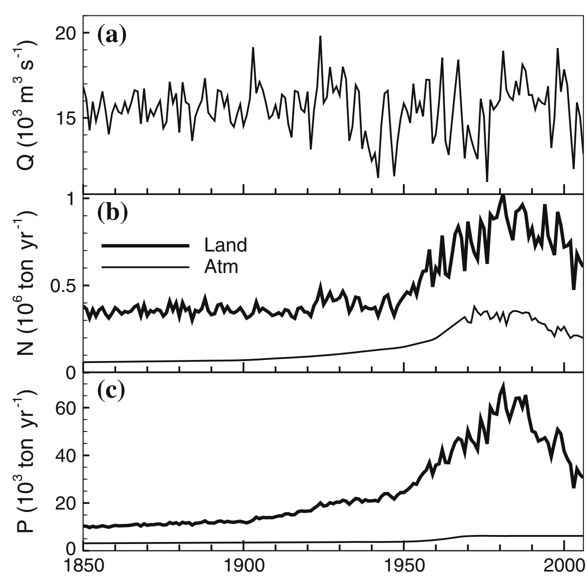

Fig. 18.2 Historical development of river water discharge (**a**), waterborne and atmospheric inputs of total nitrogen (**b**) and total phosphorus inputs (**c**) to the Baltic Sea (Gustafsson et al. 2012)

atmospheric inputs of nutrients to the Baltic Sea, the exchange with the North Sea must also be taken into account. According to budget calculations for the last 20 years, about 130 kt N year^{-1} is exported to the North Sea via the Danish Straits. This corresponds to about 15 % of the current total N inputs, while the loss of P to the North Sea amounts to 12 kt P year^{-1} corresponding to 25 % of the total input.

18.3.2 Carbon Inputs

Huge amounts of inorganic and organic carbon enter the Baltic Sea via rivers, affecting both the CO_2 system and the oxygen budget. The inputs were estimated by Kulinski and Pempkowiak (2011) using the mean concentrations of total organic carbon (particulate and dissolved) and inorganic carbon (total CO_2) in more than 60 rivers in connection with flow data. The input of total organic carbon amounted to 4090 kt C year^{-1} and consists mainly of dissolved organic carbon (DOC). More than 50 % of the annual DOC input is mineralised in the Baltic Sea (Kulinski and Pempkowiak 2011) and thus converted to CO_2. It is possible that the organic carbon input increased during the last century due to the intensification of agriculture and possibly also to rising temperatures. However, this cannot be substantiated by data.

For the input of inorganic C, Kulinski and Pempkowiak (2011) reported a value of 6810 kt C year^{-1}. Since the concentrations of total CO_2 in river water are mainly controlled by alkalinity which is primarily a consequence of limestone dissolution, changes during the last century are very likely. Hjalmarsson et al. (2008) analysed historical alkalinity data from the northern Baltic Sea which go back to 1911. They concluded that the alkalinity input into the Gulf of Finland by the River Neva increased during the last century. This was attributed to acidic precipitation which in a limestone-rich catchment such as the River Neva may intensify weathering and thus increase alkalinity. An alkalinity increase also occurred in the eastern Gotland Sea. Surface alkalinities measured during the last decade under the framework of the Swedish Monitoring Programme (SMHI) were about 100 µmol kg^{-1} higher than that reported by Buch (1945) for 1928–1938. This constitutes a significant change and is equivalent to an increase in pH of 0.02–0.03 units.

The rising levels of atmospheric CO_2 that cause enhanced CO_2 dissolution and drive the pH in sea water to lower values must also be taken into account. Concentrations increased by about 40 % from a pre-industrial level of about 280 ppm to about 390 ppm in 2008 (Fig. 18.3). The rate of CO_2 increase has grown steadily since 2000 and currently averages 2.0 ppm year^{-1}. The historic development of the atmospheric CO_2 level was reconstructed by analysing air occluded in ice cores (Fig. 18.3, Neftel et al. 1994). Direct measurement of CO_2 started in the 1950s with the well-known time series of

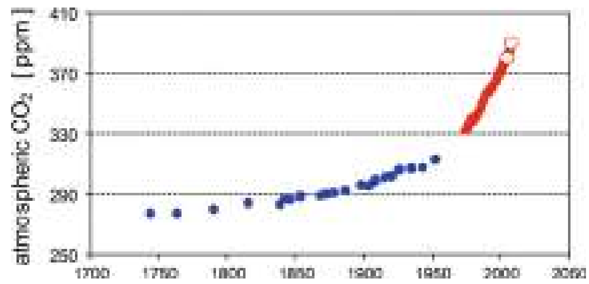

Fig. 18.3 Annual mean atmospheric carbon dioxide (CO_2) concentration obtained from analysing air in ice cores (*blue*) (Neftel et al. 1994) and direct atmospheric measurements at the Northern Hemisphere background station Barrow in Alaska (*full red circles*) (Keeling et al. 2008) and over the Baltic Sea (*open red circles*) (Schneider 2011)

measurements at Mauna Loa on Hawaii (Keeling et al. 2008), and data from a northern hemisphere background station (Barrow, Alaska) are available since 1974 at a monthly resolution (Fig. 18.3). High-resolution measurements of atmospheric CO_2 were also performed in the central Baltic Sea. A full annual cycle was obtained for 2005 and 2008 and indicated that the annual means over the Baltic Sea are almost identical to those at the Northern Hemisphere background station (Fig. 18.3, Schneider 2011).

18.4 Changes in Internal Fluxes and Transformations

18.4.1 Organic Matter Production and Nutrient Availability

18.4.1.1 Current Process Understanding

The southern and central sub-basins of the Baltic Sea are eutrophic and mesotrophic systems (Wasmund et al. 2001) that mostly process C derived from phytoplankton primary production (Elmgren 1984; Shaffer 1987). In the northern boreal regions, where phytoplankton primary production is low, organic matter input from terrestrial sources becomes increasingly important for the marine C fluxes. Food web models (Sandberg et al. 2004; Sandberg 2007) and CO_2 budgets have shown Bothnian Bay, the northernmost Baltic sub-basin, to be a net heterotrophic system (Algesten et al. 2004, 2006), whereas the Bothnian Sea oscillates between net autotrophy and heterotrophy (Algesten et al. 2004). In contrast, a CO_2 air–sea flux balance (Löffler et al. 2012) indicated that on average the Bothnian Sea acts as a sink for atmospheric CO_2 and must thus be considered as a net autotrophic system.

Baltic Sea phytoplankton communities undergo a distinct seasonal succession controlled by the temporal dynamics of stratification, light and nutrients and characterised by several blooms, that is periods of temporary mass occurrence of

Table 18.2 Winter nutrient concentrations and nutrient limitation in Baltic Sea sub-basins. Concentrations refer to mean values for 2000–2005 (Diekmann and Möllmann 2010, BED)

Basin	Winter DIN (mmol m^{-3})	Winter DIP (mmol m^{-3})	DIN:DIP (mol mol^{-1})	Spring limitation
Kattegat	6.46	0.56	11.5	N
Bornholm Basin	3.30	0.60	5.5	N
Gotland Sea	3.50	0.60	5.8	N
Bothnian Sea	3.07	0.20	15.4	N
Bothnian Bay	6.79	0.06	113	P
Gulf of Riga	11.45	0.93	12.3	P/N
Gulf of Finland	7.02	1.25	5.6	N

phytoplankton (Wasmund and Siegel 2008). Phytoplankton development starts when thermal surface layer stratification stabilises and light conditions improve, with a spring bloom dominated by diatoms and dinoflagellates. The spring bloom consumes nutrients within the euphotic zone that were accumulated during winter and ends when its limiting nutrient is exhausted. The amount of winter nutrients available to the phytoplankton spring bloom as well as the limiting nutrient differs between sub-basins. For present conditions (Table 18.2), winter nutrient concentrations are greatest in the Gulf of Riga and Gulf of Finland, which are sub-basins with high riverine nutrient loads relative to their surface area. The term nutrient limitation is often used quite loosely (Howarth 1988), but refers in aquatic ecology mostly to the nutrient that determines net primary production (Granéli et al. 1990). Assuming an N and P demand for primary production according to the Redfield ratio (N:P = 16), the winter DIN: DIP ratio is traditionally used to assess the limiting nutrient (Table 18.2). While the spring communities in the Kattegat (Granéli 1987; Granéli et al. 1990), the Baltic Proper (Granéli et al. 1990; Wasmund et al. 1998), the central Gulf of Finland and the Bothnian Sea are generally N limited (Tamminen and Andersen 2007), P limitation prevails in the Bothnian Bay (Tamminen and Andersen 2007). In the proximity of freshwater sources, limitation patterns can change spatially and interannually, depending on run-off (Pitkänen and Tamminen 1995; Seppälä et al. 1999; Tamminen and Seppälä 1999; Yurkovskis 2004).

The extent to which different nutrient sources sustain phytoplankton growth after the winter nutrients are depleted is under debate, but in the central Baltic Sea, first DIN (April) and then DIP (May to July) are depleted to the detection limit of the analytical methods (Nausch 1998; Nausch et al. 2004; Raateoja et al. 2011). Mineralisation of particulate organic P in the euphotic zone may in part cover the nutrient requirements of phytoplankton in early summer (Nausch and Nausch 2007). In addition, dissolved organic P (Nausch and Nausch 2007) and dissolved organic N (Eilola and Stigebrandt 1999; Jørgensen et al. 1999; Berg et al. 2001) deliver nutrients to pelagic communities. Preferential mineralisation of N and P compared to C might further contribute to sustain phytoplankton growth (Heiskanen et al. 1998; Osterroht and Thomas 2000; Lund-Hansen et al. 2004; Schneider et al. 2006; Riemann et al. 2008). Additional inorganic nutrients are to some extent also provided by vertical mixing (Stipa 2004; Reissmann et al. 2009). Nutrients below the thermocline can also be reached by motile phytoplankton, which migrate actively into the nutrient-rich waters below the thermocline (e.g. Olli 1999; Hajdu et al. 2007; Lips et al. 2011). Furthermore, it has been suggested that an early start to N fixation already in late spring may counteract N limitation during the decline phase of the spring bloom (Schneider et al. 2003, 2006).

During summer, when thermal stratification separates the euphotic zone entirely from the nutrient-rich waters below the thermocline, a diverse phytoplankton community is established and phytoplankton growth relies mainly on regenerated nutrients. However, in the Baltic Proper, the Gulf of Finland, and to some extent also in the Gulf of Riga and the Bothnian Sea (Wasmund et al. 2011), new production also takes place because N-fixing cyanobacteria blooms occur during summer. The magnitude of these blooms has been explained by excess inorganic dissolved phosphorus (eDIP), which due to the low N:P ratio in the winter nutrient pool still exists after the spring bloom NO_3 depletion (Janssen et al. 2004; Kahru et al. 2007; Lips and Lips 2008; Lilover and Stips 2008; Eilola et al. 2009). However, eDIP is already exhausted before the summer cyanobacteria bloom starts. The mechanism by which eDIP is transferred to summer communities and utilised by cyanobacteria is complex and potentially involves P storage in particulate (Nausch et al. 2008) and dissolved organic matter (Nausch and Nausch 2007). In addition, cyanobacteria blooms also benefit from additional P in frontal and upwelling regions (Kononen et al. 1999; Nausch et al. 2009). Atmospheric deposition is a minor source of P to cyanobacteria blooms (Rolff et al. 2008).

Blooms of N-fixing cyanobacteria provide an important input of N to the Baltic Sea ecosystem, thus increasing organic matter production and sedimentation (Vahtera et al. 2007). Nitrogen fixation also circumvents N limitation in the central Baltic Sea, and P concentrations become the major control of the annual net production. Baltic cyanobacteria blooms are formed by the heterocystous species *Nodularia* spp., *Aphanizomenon* spp. and *Anabaena* spp. (Sohm et al. 2011), which are favoured by high water temperatures (Kanoshina et al. 2003; Laamanen and Kuosa 2005; Mazur-Marzec et al. 2006; Lips and Lips 2008; Jaanus et al. 2011) and calm conditions (Kanoshina et al. 2003; Mazur-Marzec et al. 2006). Estimates of the N input by cyanobacteria to the Baltic Sea cover a wide range. Based on field observations, values between 20 and 792 kt N year^{-1} were reported (median 173 kt N year^{-1}, based on sources in Degerholm et al. 2008), whereas estimates derived from biogeochemical modelling resulted in 200–800 kt N year^{-1} (Savchuk and Wulff 2007; Neumann and Schernewski 2008). These numbers indicate that N input by N fixation is comparable to the annual riverine DIN load to the Baltic Proper (286 kt N year^{-1} in 2000, HELCOM 2004). This additional N input not only sustains the growth of cyanobacteria in summer communities, but is also transferred by exudation and organic matter mineralisation to non-N-fixing phytoplankton (Ohlendieck et al. 2007). In the central Baltic Sea, N fixation contributes a substantial part of N export from the euphotic zone (Larsson et al. 2001; Struck et al. 2004).

The amount of C converted into biomass by primary producers depends on the uptake ratio of C and nutrients during phytoplankton growth. However, under P limitation, the C:P and N:P ratios in Baltic plankton may exceed the Redfield ratio of C:N:P = 106:16:1 (e.g. Engel et al. 2002; Nausch et al. 2004; Kangro et al. 2007; Walve and Larsson 2007, 2010). However, lower C:P ratios are also observed in cases where a PO_4 excess with regard to the Redfield N:P ratio exists in the nutrient pool (Wasmund et al. 1998). Therefore, the amount of C formed per unit N or P available to producers varies, which further influences C sedimentation and O_2 consumption in sediments and bottom waters.

Measurements of phytoplankton primary production using the ^{14}C method are associated with large methodological differences in the Baltic Sea (Andreasson et al. 2009; Larsson et al. 2010). Therefore, phytoplankton primary production is poorly known on a Baltic scale. Including river plumes and coastal areas, phytoplankton net primary production in the central Baltic Sea was estimated at about 200 g C m^{-2} year^{-1} by the end of the 1990s (Wasmund et al. 2001). A substantial part of this production is mineralised already in the euphotic zone. For the spring and early summer period in the eastern Gotland Sea, Wasmund et al. (2005) estimated that mineralisation reached 52 % of phytoplankton production. Consequently, net organic C production, which is the difference between primary production and respiration within the euphotic zone, is substantially less than phytoplankton primary production. Based on oxygen budgets, Stigebrandt (1991) estimated net production in the period before 1980 to range between 41 g C m^{-2} year^{-1} in the Kattegat, 38–49 g C m^{-2} year^{-1} in the Baltic Proper, and 32 g C m^{-2} year^{-1} in the Bothnian Sea. Direct measurements of the amount of C sinking out of the euphotic zone (export production) can be obtained using sediment traps. However, interpreting the data is complicated due to resuspension effects and organic matter degradation in the traps. Heiskanen and Leppänen (1995) estimated that only 32 % of phytoplankton primary production sinks out of the euphotic zone in the northern Baltic Sea, with most of the C export occurring during the spring bloom. A C budget model for the Kattegat (Carstensen et al. 2003) indicated that 54 % of net phytoplankton production is exported from the euphotic zone. However, the relationship between phytoplankton primary production and C export from the euphotic zone depends not only on nutrient supplies and the magnitude of primary production itself, but also on the structure of the pelagic ecosystem (e.g. Boyd and Trull 2007 and references therein).

18.4.1.2 Past Changes

Since the beginning of systematic monitoring in the 1970s, surface winter DIN and—with the exception of the Bothnian Bay—DIP have increased in all sub-basins of the Baltic Sea (see Fig. 18.4 for the central Baltic Sea) and reached a peak between 1980 and 1990 (HELCOM 2009). In sub-basins with a short residence time like the Danish Straits, DIN and DIP closely follow nutrient inputs, with pronounced concentration declines since 1990 (Carstensen et al. 2006). Decreasing winter DIN concentrations after the mid- or late 1980s have also been observed in all other Baltic Sea sub-basins except the Bothnian Bay (HELCOM 2009). The decline in DIN concentrations may be due to the decreased N inputs but also may be due to higher denitrification rates caused by increased hypoxia in the Baltic Sea. Winter DIP concentrations in the Baltic Proper and the Gulf of Finland do not show a distinct trend. They are more closely related to P concentrations beneath the Baltic Proper halocline and are thus linked to inflow dynamics and P release from bottom sediments (Conley et al. 2002; HELCOM 2009, see also Sect. 18.4.2.4). Consequently, winter DIP concentrations in the central Baltic Sea increased again after the inflow in 2003 and in the Gulf of Finland after the inflow in 1996 (HELCOM 2009). In the Gulf of Riga, winter DIP concentrations remained high during the 1990s due to release from bottom sediments (Müller-Karulis and Aigars 2011). The decline in winter DIN levels, while DIP remained high in the Baltic Proper, the Gulf of Finland and the Gulf of Riga, also led to a marked decrease in DIN:DIP ratios, which favours the growth of N-fixing cyanobacteria.

Estimates of winter nutrient concentrations prior to the 1970s are available from a few modelling studies. A time slice experiment (Schernewski and Neumann 2005) as well as a transient simulation (Gustafsson et al. 2012) suggested that Baltic Proper winter nutrient concentrations in 1900

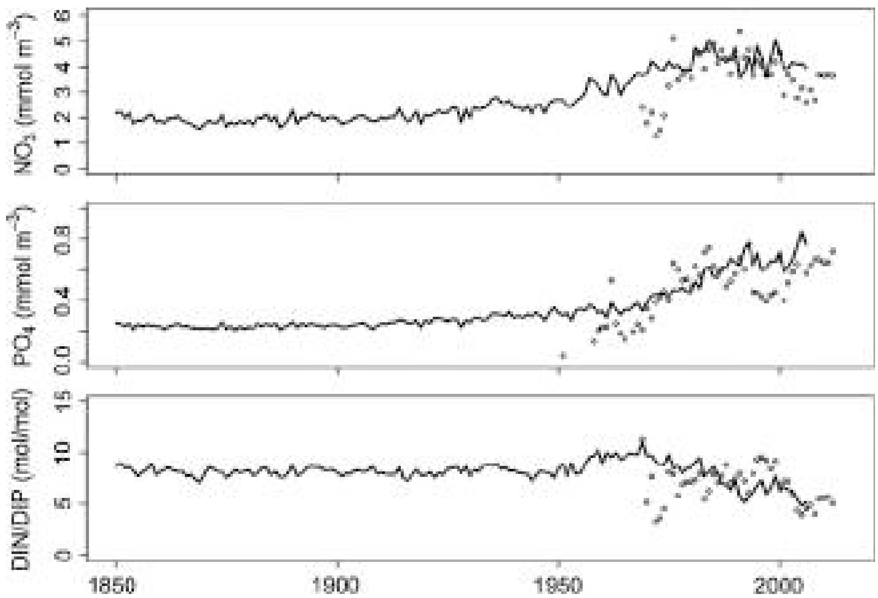

Fig. 18.4 Simulated (*line*) and observed winter nutrient concentrations (*dots*) in the surface layer of the central Baltic Sea (Gustafsson et al. 2012)

were 2.0–3.7 mmol m^{-3} for DIN and 0.23–0.35 mmol m^{-3} for DIP. Current DIN concentrations (Table 18.2) are thus within the upper range of past DIN concentrations, whereas the past DIP concentration was about a factor of two lower than present values. Nutrient concentrations remained near pristine values until the 1950s and then rose quickly (Gustafsson et al. 2012, Fig. 18.4).

The past increase in winter nutrient concentrations is also reflected in changes in primary production. In the Kattegat and Belt Sea area, where consistent time series of phytoplankton primary production measurements by the ^{14}C method are available, eutrophication led to a doubling of primary production between 1950 and 1980 (Rydberg et al. 2006, Fig. 18.5). Estimates for the increase in primary production for the Baltic Proper range from an increase of 30–70 % since the beginning of the twentieth century (Elmgren 1989) to a doubling in the 1970s and 1980s (Wasmund et al. 2001). Also, decreasing $\delta^{13}C$ values in bottom sediments indicate an increase in primary production due to eutrophication (Voss et al. 2000). Results from modelling studies differ. An increase in primary production since 1900 by a factor of three and even six were reported by Savchuk et al. (2008) and Gustafsson et al. (2012), respectively.

Net organic C production in the Gotland Basin was estimated by CO_2 measurements and yielded an increase by a factor of about 2.5 since the 1920s (Schneider and Kuss 2004), which is in close agreement with the model estimate by Gustafsson et al. (2012). Higher primary production is also reflected in enhanced deposition of organic matter at the sediment surface. According to Jonsson and Carman (1994),

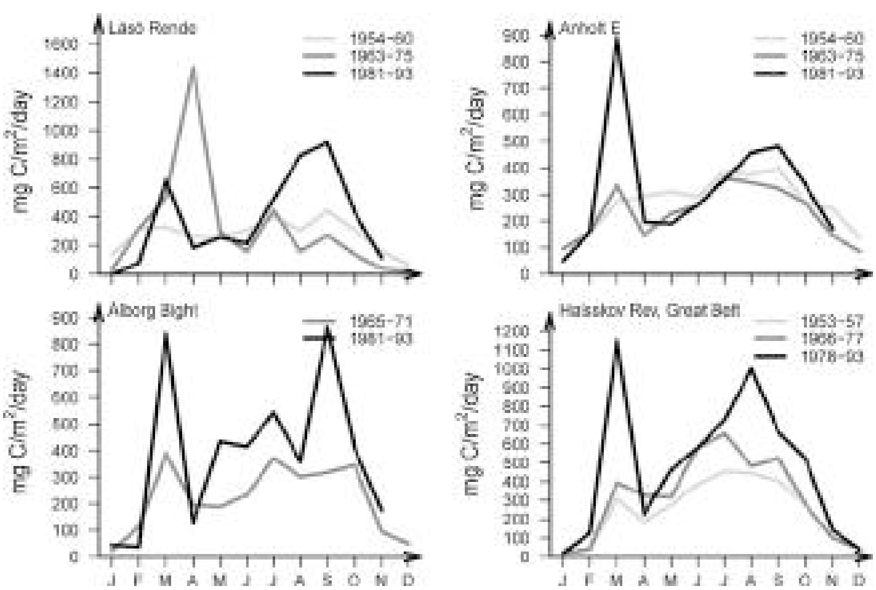

Fig. 18.5 Changes in phytoplankton primary production in the Kattegat and the Belt Sea (Rydberg et al. 2006)

Box 18.2 Main biogeochemical transformations during primary production and organic matter mineralisation (in the case of mineralisation by denitrification and sulphate reduction, the organic matter produced is given without the nitrogen and phosphorus content)

Process	Reaction
Photosynthetic primary production	$106CO_2 + 16NO_3^- + H_3PO_4 + 122H_2O + 16H^+ \rightarrow (CH_2O)_{106}\ (NH_3)_{16}\ H_3PO_4 + 138O_2$
Oxic mineralisation	$(CH_2O)_{106}\ (NH_3)_{16}\ H_3PO_4 + 106O_2 \rightarrow 106CO_2 + 16NH_3 + H_3PO_4 + 106H_2O$
Nitrification	$NH_3 + H_2O + 2O_2 \rightarrow NO_3^- + H^+ + 2H_2O$
Denitrification	$5CH_2O + 4NO_3^- + 4H^+ \rightarrow 5CO_2 + 2N_2 + 7H_2O$ (heterotrophic)
	$8NO_3^- + 5HS^- + 3H^+ \rightarrow 4N_2 + 5SO_4^{2-} + 4H_2O$ (chemolithotrophic)
Sulphate reduction	$2CH_2O + SO_4^{2-} + H^+ \rightarrow 2CO_2 + HS^- + 2H_2O$

the sedimentation rate increased by 70 % between the late 1920s and 1980s in the Baltic Proper, whereas Hansson and Gustafsson (2012) reported an increase of only 25 % after 1950. A larger increase was obtained from model simulations which yielded a rough doubling of sedimentation since the 1950s (Hansson and Gustafsson 2012) and an almost fourfold increase since 1900 (Gustafsson et al. 2012).

Cyanobacteria blooms have occurred in the Baltic Sea since the early brackish Littorina Sea stage (Chap. 2), in particular during hypoxic bottom water conditions (Bianchi et al. 2000). Observations suggest an increase in cyanobacteria after the 1960s (Finni et al. 2001). Furthermore, budgets for CO_2 and total N in the surface water of the central Baltic Sea indicated an increase in N fixation by cyanobacteria of 30 % within the past two decades (Schneider et al. 2009). Model studies suggest a fourfold (Gustafsson et al. 2012) to 10-fold (Savchuk et al. 2008) increase in N fixation since 1900.

18.4.2 Organic Matter Decomposition and Nutrient Recycling

Decomposition of organic matter refers to both the decay of large organic compounds into smaller molecules and the oxidation of organic matter. The latter process is also referred to as mineralisation and plays a key role in the cycling of C and nutrients. Mineralisation is a biologically induced process that can be expected wherever organic matter exists. This text focuses on mineralisation in water layers and at the sediment surface below the permanent halocline that may be cut off from O_2 supply for many years and thus generates an O_2-depleted or anoxic environment. Under such conditions, several important biogeochemical reactions occur (Box 18.2) that will be discussed in the following subsections.

18.4.2.1 Hydrographic Forcing of the Deep-Water Biogeochemistry

The O_2 conditions in the Baltic Sea are the result of physical transport of O_2 and consumption of O_2 by biogeochemical processes. Owing to the positive freshwater budget and the bathymetry of the Baltic Sea, a strong salinity gradient generates a permanent density stratification (pycnocline) that inhibits the vertical exchange of O_2 between the surface and deep waters. Lateral intrusions and inflows below the pycnocline, which is located at about 70 m in the central Gotland Basin, are the only effective means of O_2 transport. The inflowing water requires a density high enough to sink down to the near-bottom layers of the central Baltic Sea deep basins. Those inflows occur irregularly, sometimes at intervals of many years. Matthäus and Franck (1992) used an intensity index to classify the inflow events during the last century (Fig. 18.6). During longer periods without a sufficient supply of O_2 to the deep water, the continuous consumption of O_2 by decomposition of organic matter results in O_2 depletion and eventually in the development of anoxic conditions. Thus, the Baltic Sea is vulnerable to hypoxia and anoxia due to its hydrography (Conley et al. 2009a).

The most northern parts of the Baltic Sea, the Bothnian Sea and Bothnian Bay, are characterised by weaker density stratification, and vertical ventilation by deep convection may occur in particular years. Owing to the vertical ventilation and a lower primary production, this part of the Baltic Sea is not affected by O_2 deficiency. A detailed description of Baltic Sea hydrography is given in Chap. 7.

18.4.2.2 Oxygen Depletion and H_2S Formation

Current process understanding: The main reason for O_2 demand in sea water is the mineralisation of organic matter. This is an oxidation process performed by heterotrophic organisms that use the chemical energy stored in organic matter and thus reverse chemically the primary production

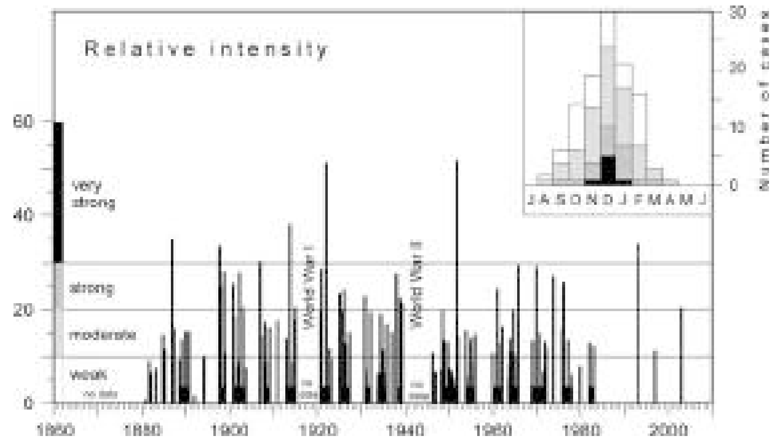

Fig. 18.6 Intensity index for major Baltic Sea inflows during the last century (Schinke and Matthäus 1998) and the seasonal frequency distribution of different inflow categories (*inset*)

process (see Box 18.2). A large fraction of the organic matter produced in the euphotic zone is sinking into deeper water layers, and by mineralisation, inorganic nutrients and CO_2 are again released. In addition to vertical transport, a considerable lateral transport of organic matter, produced in shallower areas, contributes to the C flux into the deep basins of the Baltic Sea and intensifies the mineralisation process (Emeis et al. 2000; Hille et al. 2006; Schneider et al. 2010).

Mineralisation of organic matter requires an electron acceptor (oxidant), which is provided first by O_2. Once O_2 and other oxidants such as NO_3 and manganese oxides are depleted, sulphate will be used instead. Sulphate reduction produces H_2S that at the deep-water pH mainly occurs as HS^- (see Box 18.2) and is a toxin for many forms of life. These processes may occur in the water column, but mainly take place at the immediate sediment surface (fluffy layer) and in the upper layers of the surface sediment where particulate organic matter is accumulated. The primary mineralisation products such as ammonia, PO_4, CO_2 and H_2S diffuse into the water column and may undergo further transformations depending on the redox conditions (see Sects. 18.4.2.3 and 18.4.2.4). A special mineralisation process, methanogenesis, takes place only in deeper organic-rich sediment layers where sulphate as an oxidant is entirely consumed. In this case, an internal oxidation/reduction (disproportionation) of organic matter is mediated by microorganisms and results in the formation of CO_2 and methane (CH_4). However, only a minor fraction of CH_4 reaches the surface water and can escape into the atmosphere because of oxidation processes both in the sediment and in the water column (Schmale et al. 2010).

Past changes: Eutrophication is considered the main cause of low O_2 concentrations in the deep water of the present-day Baltic Sea (Larsson et al. 1985; Conley et al. 2009b). The observed deep-water O_2 conditions during the last 100 years (Fig. 18.7) show that the occurrence of H_2S became much more frequent during recent decades. The increased availability of organic C due to eutrophication required more electron acceptors, and therefore, the depletion of O_2 and formation of H_2S were accelerated.

Strong eutrophication started in the 1960s and peaked in the 1980s (Elmgren 2001; Gustafsson et al. 2012). In combination with a long-lasting stagnation period from 1976 to 1993 (Fig. 18.6) without any significant dense inflows from the North Sea, eutrophication caused high concentrations of H_2S in the deep water of the Gotland Basin. Other locations such as the Landsort Deep, however, showed an increase in O_2 concentration. In this case, the missing inflow events weakened the stratification due to decreasing salinity and thus enhanced the vertical O_2 flux. Similarly, in the Gotland Basin, the decreasing stratification caused a deepening of the permanent halocline (Diekmann and Möllmann 2010), thus exposing larger bottom areas to ventilated surface waters, while H_2S concentrations in bottom waters below the ventilation depth increased (Fig. 18.7). The shrinking of the hypoxic area during long-lasting stagnation periods is illustrated in Fig. 18.8. It shows that the hypoxic area was much smaller at the end of the 1976–1993 stagnation period than in 2006 after some recent inflows (Conley et al. 2002, 2009a; Savchuk 2010).

Oxygen consumption by organic matter mineralisation is accompanied by the release of CO_2. Hence, the accumulation of total CO_2 in stagnant deep water can be used to estimate O_2 conditions. The oldest deep-water data on total inorganic carbon (C_T) originate from 1928–1938 (Buch 1945) and indicate a lower C_T accumulation and thus a lower O_2 deficiency than at present. This is certainly a consequence of the lower input of particulate organic matter into deeper water layers due to the lower biomass production at the beginning of the last century.

Sedimentary records suggest that periods of O_2 deficiency and anoxic conditions in the Baltic Sea also occurred in the historic past (Conley et al. 2009a; Virtasalo et al. 2011) during warmer climate conditions such as the medieval climate anomaly (MCA, ca. 900–1350, also known as

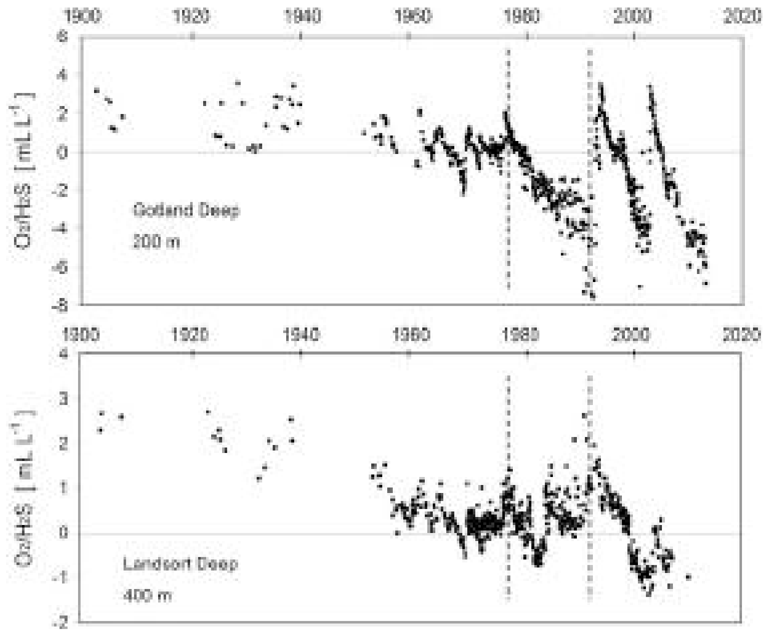

Fig. 18.7 Time series of oxygen and hydrogen sulphide (shown as negative oxygen equivalents) for the Gotland Deep and Landsort Deep. The *dashed lines* mark the long-lasting stagnation period from 1976 to 1993 (adopted from Fonselius 1969 and updated to 2013 by monitoring data from the Swedish Meteorological Hydrological Institute and the Leibniz Institute for Baltic Sea Research, Germany)

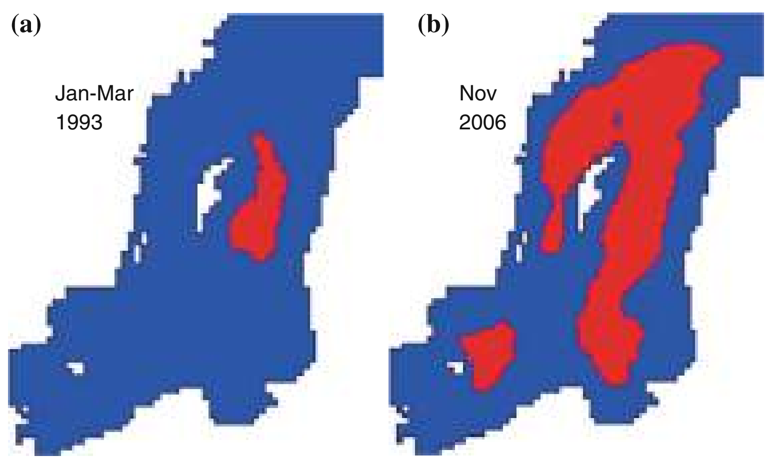

Fig. 18.8 Sediment area (*red*) covered by hypoxic waters containing less than 2 mL L^{-1} dissolved oxygen: **a** at the end of a long-lasting stagnation period in 1993 (11,050 km^2) and **b** in 2006 subsequent to some inflow events (67,700 km^2) (Conley et al. 2009a)

Medieval Warm Period, see Chap. 3). During colder periods, for example the Little Ice Age (ca. 1550–1850, see Chap. 3), the O_2 supply of the deep water improved due to more intense ventilation. The mechanism behind the correlation between large-scale meteorological conditions in the different climate periods and O_2 conditions in the Baltic Sea is not well understood and subject to ongoing research. It has also been hypothesised that human activities already had an impact on the O_2 conditions of the Baltic Sea during the MCA. At that time, a doubling of the population in the catchment area occurred that led to more intense land use and possibly increased the release of soil nutrients (Zillén et al. 2008; Zillén and Conley 2010; see also Chap. 25).

18.4.2.3 Nitrogen Transformations

Current process understanding: Transformations between N species depend on the O_2 conditions (redox potential) and are mediated by microbial activity. In sea water, the most important N compounds are ammonia, NO_2 and NO_3. This text considers those reactions that take place during the mineralisation of organic matter below the well-oxygenated surface layer. Ammonia is the primary product of the mineralisation process. In oxic conditions, nitrification occurs by which ammonia is rapidly oxidised first to NO_2 and finally to NO_3 (Box 18.2).

By vertical mixing, NO_3 may become available for new production in the surface water. An N transformation with more far-reaching consequences for the N balance is denitrification, by which NO_3 is reduced to elemental N_2 and thus removed from the DIN pool. This occurs at low O_2 concentrations (<5 µmol L^{-1}, Devol 2008) by the use of NO_3 for the oxidation of organic matter (Box 18.2). This heterotrophic denitrification is most effective in the organic-rich upper sediment layers that are in contact with an oxic water column where NO_3 is produced by nitrification. But denitrification may also take place in the water column, in particular if a pelagic redoxcline exists. In this case, denitrification occurs mainly by oxidation of H_2S through NO_3 (Box 18.2) during mixing of oxic and anoxic waters at the redoxcline (chemolithotrophic denitrification).

Most of the studies on denitrification focused on the pelagic redoxcline. Incubation experiments have been used to determine denitrification rates (Rönner and Sörensen 1985; Brettar and Rheinheimer 1991, 1992; Hannig et al. 2007), which, however, did not allow an estimate of the large-scale effect of denitrification on the Baltic Sea DIN budget. Other studies used measurements of the N_2 production in order to quantify the effect of denitrification (Rönner and Sörensen 1985; Löffler et al. 2011). Mass balances were used by Gustafsson and Stigebrandt (2007) and Schneider et al. (2010) to quantify the denitrification that takes place during the transition from oxic to anoxic conditions after renewal of the Gotland Sea deep water.

In addition to the process-orientated studies, attempts have been made to quantify the large-scale denitrification. Based on a mass balance model, Shaffer and Rönner (1984) reported a total denitrification for the Baltic Proper of 470 kt N $year^{-1}$ which corresponds to a removal of about 50 % of the total annual N input to the Baltic Sea. Furthermore, they estimated that 80–90 % of the denitrification occurred in surface sediments and that denitrification at the pelagic redoxcline is thus of minor importance. Voss et al. (2005) concluded from isotopic measurements ($\delta^{15}N$) that sediments of the coastal rims play an important role in the removal of DIN. They found that the annual denitrification in coastal areas of the Baltic Proper may range between 580 and 855 kt N. This implies that the river N load is largely removed close to the coast and has only a minor effect on eutrophication of the central basins. However, this conflicts with the strong increase in DIN in the central Baltic Sea (Fig. 18.4) that roughly coincides with the increasing N loads during 1950–1980 (Fig. 18.2). Deutsch et al. (2010) investigated denitrification in sediments with different physical and chemical properties. The results were extrapolated to the entire Baltic Sea by the use of maps describing sediment characteristics for the entire Baltic Sea. For the Baltic Proper, sediment denitrification amounted to 191 kt N $year^{-1}$. Together with denitrification at the pelagic redoxcline calculated from the rate measurements of Brettar and Rheinheimer (1991), this implies a total denitrification of 397 kt N $year^{-1}$ for the Baltic Proper. For the remaining regions where no pelagic redoxcline exists, Deutsch et al. (2010) reported a sediment denitrification of 235 kt N $year^{-1}$. Hence, the total denitrification in the Baltic Sea (823 kt N $year^{-1}$) agrees with the upper limit given by Voss et al. (2005).

Past changes: Although reliable and consistent estimates of the present-day denitrification are not yet available, the annual rates under discussion indicate the paramount importance of this process for the Baltic Sea N budget and ultimately for many related biogeochemical processes. Hence, long-term changes in denitrification may have changed the cycling of nutrients and thus had an effect on the productivity of the Baltic Sea. Direct evidence for past changes does not exist, but it is reasonable to hypothesise that denitrification has increased concurrently with increasing production because organic matter provides the substrate and generates the necessary redox gradient for denitrification. This hypothesis is supported by observations by Vahtera et al. (2007) who found a negative correlation between the Baltic Sea DIN pool and the volume of hypoxic water. Since eutrophication has increased the extent of hypoxia during the last century, denitrification may constitute a negative feedback on eutrophication by partly removing the increasing input of DIN. On the other hand, higher denitrification rates and a concurrent increased PO_4 release at hypoxic conditions (see Sect. 18.4.2.4) are decreasing the N:P ratio in the nutrient pool (Savchuk 2010). This will favour N fixation, which may then at least partly compensate DIN losses by denitrification.

18.4.2.4 Phosphorus Transformations

Current process understanding: Organic P compounds in the water column are subject to mineralisation and other biological and abiotic transformation processes during transportation and settling. Decomposition of organic P results in small organic molecules and PO_4 which is used by algae and bacteria. Although phosphate exists mainly as HPO_4^{2-} at sea water pH around 8, it is commonly depicted by 'PO_4'. PO_4 can also be adsorbed onto particulate material and settle to the sea floor (Froelich 1988). Mineralisation and transformation processes continue at the sediment–water interface and in surface sediment layers (e.g. Sundby et al. 1992).

In oxic conditions and Fe-rich systems, PO_4 is adsorbed onto amorphic oxyhydroxides of ferric iron (Fe^{3+}) (Mortimer 1941; Hingston et al. 1967). In the sediment pore water, PO_4 diffuses upwards into the oxic sediment layer and can be bound to Fe-oxyhydroxides at the surface (Mortimer 1971; Krom and Berner 1981). Fe-bound P is abundant in surface sediments overlain by oxic bottom waters in the Baltic Sea (Balzer 1986; Jensen and Thamdrup 1993; Lukkari et al. 2009; Mort et al. 2010). Furthermore, in oxic conditions, activity of benthic fauna oxygenates the sediment and enhances binding of PO_4. Benthic fauna can also mix PO_4 from pore water to the water column and accelerate the release of P from the organic phase by enhancing microbial activity (Aller 1988; Kristensen 1988).

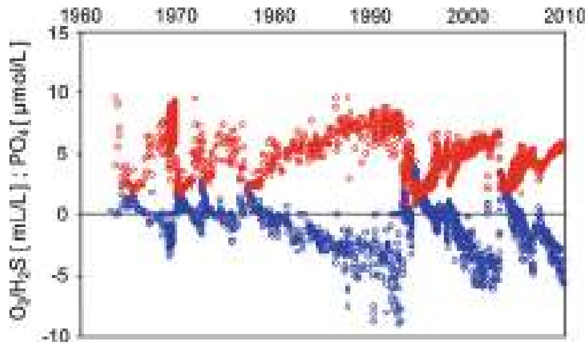

Fig. 18.9 Time series of phosphate (PO_4) and oxygen (O_2) concentrations in the deep water of the central Gotland Sea. Negative O_2 values represent hydrogen sulphide (H_2S) oxidation equivalents (Swedish National Monitoring Programme, SMHI)

In sediments overlain by anoxic water, Fe in the oxyhydroxides is reduced resulting in the release of associated PO_4 (Einsele 1936; Mortimer 1941). In organic-rich sediments common in the Baltic Sea, O_2 consumption by mineralisation of organic matter can create reduced conditions, if the O_2 is not replenished. When Fe-oxyhydroxides are not able to form, or they form only a thin layer on the sediment surface, PO_4 is not efficiently retained (Mortimer 1971). Furthermore, in brackish and marine environments, sulphide can restrict participation of Fe in PO_4 binding by precipitating ferrosulphides (Berner 1970; Caraco et al. 1989).

According to the 'vicious circle' hypothesis, eutrophication nourishes itself by increasing production and sedimentation of organic matter which enhances the formation of anoxia (e.g. Vahtera et al. 2007). When anoxia extends to previously oxic bottom areas rich in Fe-bound PO_4, it releases more PO_4 from the sediment. If vertical mixing in the water column supplies released PO_4 to the productive layer, it further enhances eutrophication, leading to yet another progression of anoxia. However, if the sea area has been anoxic for a long period, the amount of Fe-bound PO_4 sensitive to reduction-induced release may have already been diminished (Lukkari et al. 2009; Jilbert et al. 2011; Malmaeus and Karlsson 2012). Furthermore, under anoxia, regeneration of organic P may be enhanced (Ingall et al. 1993; Ingall and Jahnke 1994), for example via release of organic P compounds from Fe-oxyhydroxides (Suzumura and Kamatani 1995) and release of PO_4 from microbial sources (Gächter et al. 1988; Ingall and Jahnke 1994; Hupfer et al. 2004).

Past changes: Important long-term changes affecting P transformations and cycling within the Baltic Sea are linked to increasing P inputs during the last century. This has a direct effect on the total PO_4 inventory in the Baltic Sea and is reflected in the increase in surface water PO_4. Hence, organic matter production was enhanced and resulted in an accelerated O_2 depletion and finally the intensified formation of anoxia in the deeper below-halocline water layer since the 1970s (Conley et al. 2009a). This has restricted the binding of PO_4 to Fe-oxyhydroxides in the sediments and PO_4 recycles longer in the system maintaining higher PO_4 levels in the water (Ingall et al. 1993; Hille et al. 2005; Jilbert et al. 2011). Figure 18.9 presents observations of PO_4 concentration in the bottom water of the central Gotland Sea since about 1960. A trend cannot be detected because large changes occurred on short timescales. These can be explained by concomitant variations in O_2 concentrations and by the abundance of H_2S, which are controlled by irregular saltwater inflows from the North Sea.

18.4.3 The Marine CO_2 (Acid/Base) System

The marine CO_2 system is of biogeochemical importance as it regulates the pH of sea water and at the oceanic scale is a major control for atmospheric CO_2. This has stimulated intensive research on the marine CO_2 system during the last 20–30 years when the role of the oceans for the uptake of anthropogenic CO_2 and the consequences for marine biogeochemistry became increasingly evident. Investigations of the marine CO_2 system can also be used as a tool to study biogeochemical processes since biological production and decomposition of organic matter are intimately connected with the consumption or release of CO_2.

18.4.3.1 Current Process Understanding

The marine CO_2 system constitutes the major component of the acid/base balance in sea water. It is characterised by thermodynamic equilibria between hydrogen ions (pH) and the different CO_2 species (Box 18.2 and Schneider 2011). The latter are dissolved CO_2, carbonic acid (H_2CO_3), hydrogen carbonate (HCO_3^-) and carbonate ions (CO_3^{2-}). The sum of the different CO_2 species is called total CO_2, C_T or dissolved inorganic carbon (DIC). Another important variable of the CO_2 system is the CO_2 partial pressure ($p$$CO_2$), which is proportional to the CO_2 concentration. Finally, another variable to be considered is the alkalinity, A_T, which is defined as the excess of proton acceptors such as HCO_3^- and CO_3^{2-} ions over proton donors (H^+). The A_T controls the change in pH upon the addition of CO_2 or other acidic substances and is thus regarded as the buffer capacity of sea water (Omstedt et al. 2010).

Box 18.3 The marine CO_2 system: variables and equilibria (square brackets indicate concentrations)

Total CO_2 (dissolved inorganic carbon)	$C_T = [CO_2] + [H_2CO_3] + [HCO_3^-] + [CO_3^{2-}]$
Alkalinity (oxic waters)	$A_T = [HCO_3^-] + 2 \cdot [CO_3^{2-}] + [B(OH)_4^-] + [OH^-] - [H^+]$
CO_2 partial pressure	pCO_2
Hydrogen ions	$pH = -\log[H^+]$
CO_2 solubility constant (k_0)	$[CO_2^*] = k_0.pCO_2$
1. Dissociation constant (k_1)	$CO_2 + H_2O \leftrightarrow HCO_3^- + H^+ : \frac{[H^+] \cdot [HCO_3^-]}{[CO_2^*]} = k_1 \quad [CO_2^*] = [H_2CO_3] + [CO_2]$
2. Dissociation constant (k_2)	$HCO_3^- \leftrightarrow CO_3^{2-} + H^+ : \frac{[H^+] \cdot [CO_3^{2-}]}{[HCO_3^-]} = k_2$
Calcium carbonate solubility product, k_{sp}	Saturation : $([Ca^{2+}] \cdot [CO_3^{2-}]) = k_{sp}$

The large-scale distribution of total CO_2 in the Baltic Sea surface water is widely controlled by alkalinity, which originates mainly from the weathering of limestone in the catchment. The soils in the southern and eastern catchment are rich in limestone, and thus, their rivers carry large amounts of A_T into the Baltic Sea. In contrast, the igneous rocks prevailing in the Scandinavian catchment result in low river water A_T.

Data for total CO_2 in surface water along a transect through the entire Baltic Sea (Beldowski et al. 2010) show highest values (>2000 µmol kg^{-1}) in the Kattegat and minimum values of about 600 µmol kg^{-1} in the Bothnian Bay. However, due to uptake and release of CO_2 during biological production and mineralisation of organic matter, a distinct seasonal cycle is superimposed on the A_T-controlled background C_T and is reflected in the surface water pCO_2. The pCO_2 seasonality of the surface water in the central Baltic Sea obtained from automated measurements on a cargo ship (Schneider 2011) reveals two typical minima (Fig. 18.10). These are caused by CO_2 consumption during the two major productive periods which are the spring bloom and the N-fixing cyanobacteria bloom in midsummer.

Later in the year, the pCO_2 recovers and exceeds the atmospheric level. This is mainly due to the deepening of the mixed layer which leads to the transport of CO_2-enriched deeper water layers to the surface. The pCO_2 is directly linked to the pH which accordingly shows pronounced seasonality. High pH (of about 8.5) is observed during the spring/summer productive period, and low pH (around 7.9) can occur during the deepening of the mixed layer in autumn/winter (Wesslander et al. 2010).

The mainly biologically driven seasonal cycle in the surface water pCO_2 also controls the annual CO_2 gas exchange balance since the gas exchange is driven by the difference between the surface water pCO_2 and the relatively stable atmospheric pCO_2. It is not known whether the Baltic Sea is a net sink or a net source of atmospheric CO_2. Whereas Thomas and Schneider (1999) reported a CO_2 uptake of 0.9 mol m^{-2} $year^{-1}$ by the central Baltic Sea, Wesslander et al. (2010) concluded that the Gotland Sea and Bornholm Sea are sources of atmospheric CO_2 and release 1.6 and 2.4 mol m^{-2} $year^{-1}$ CO_2, respectively, to the atmosphere. These discrepancies are due to the different data sources, all showing insufficient temporal and spatial coverage in view of the huge daily, seasonal, interannual and regional variability in surface water pCO_2.

The deep water of the major basins is subject to continual mineralisation of organic matter that leads to the accumulation of total CO_2 during periods of stagnation. To estimate the effect of increasing C_T on pH, concurrent alkalinity changes must be taken into account (Edman and Omstedt 2013). During oxic conditions, nitrification of ammonia

Fig. 18.10 Typical seasonality in the CO_2 partial pressure (pCO_2) of surface water (*blue line*) and in the atmosphere (*green line*) in the central Baltic Sea (adopted from Schneider 2011)

takes place and results in a decrease in alkalinity. This slightly amplifies the strong decrease in pH caused by the gain in C_T. As a consequence, the pH at O_2 concentrations close to zero may be about one unit lower than the mean surface pH. When the water is turning to anoxic conditions, both denitrification and especially sulphate reduction, together with the associated release of ammonia from organic matter, strongly increase alkalinity. This damps a further decrease in pH during mineralisation at anoxic conditions and stabilises pH at about 7.0 (Ulfsbo et al. 2011) after a longer period of stagnation.

18.4.3.2 Past Changes

A large set of historical CO_2 data for the northern Baltic Sea including the Gulf of Bothnia and Gulf of Finland is available for 1927–1938 when the first pioneering studies on the marine CO_2 system were performed by Kurt Buch and co-workers at the Finnish Institute of Marine Research (FIMR) in Helsinki. They determined equilibrium constants for the CO_2 system that agreed reasonably well with those determined by more sophisticated analytical techniques in recent years. This indicates that their data can be compared against the present-day status of the marine CO_2 system. The current text restricts discussion on possible change in the CO_2 system to the surface water because interpreting the deep-water CO_2 data requires detailed information about the chronology of water mass stagnation and renewal events. This is not available for the Buch (1945) data, which give only seasonal mean values.

Surface water pCO_2 reflects the interplay between biological production, mineralisation and mixing, and has undergone large changes during the past 80 years. In Buch's (1945) data, the pCO_2 minimum after the spring bloom in the eastern Gotland Sea was about 270 µatm. Taking into account the lower atmospheric pCO_2 during the 1930s (about 320 µatm, Callendar 1940), a CO_2 partial pressure difference, ΔpCO_2, of about −50 µatm is obtained which indicates moderate undersaturation of the surface water with regard to atmospheric CO_2. Recent pCO_2 measurements show a spring bloom minimum of about 150 µatm which corresponds to a ΔpCO_2 of −250 µatm (Schneider 2011). This undersaturation is about fivefold larger than at the beginning of the last century and can be attributed to increased organic matter production as a consequence of eutrophication. Although this is consistent with other historic production estimates, it must be taken into account that Buch's (1945) spring value refers only to a single year and so its representation for the spring bloom can be questioned.

The changes in surface water ΔpCO_2 during the past 80 years are even more pronounced in midsummer. Due to the intense N_2 fixation in July, the ΔpCO_2 in the central Baltic Sea currently amounts to almost −300 µatm (Schneider et al. 2009), whereas the mean ΔpCO_2 at this time of year for 1927–1935 was less than −40 µatm (Buch 1945). This indicates low or even absent N fixation activity before the onset of the eutrophication.

Changes in the marine CO_2 system are also due to the uptake of anthropogenic CO_2. The increase in atmospheric CO_2 from 280 ppm during the pre-industrial era to almost 400 ppm in 2010 has led to a corresponding increase in the mean surface water pCO_2. According to the thermodynamics of the marine CO_2 system, a decrease in pH of about 0.15 units can be expected if no other changes in the biogeochemical cycles had occurred. This phenomenon is called 'ocean acidification' and has stimulated many research projects on the ecological consequences of the decreasing pH. Monitoring programmes have included pH measurements for several decades, and large amounts of data are currently stored at national and international data centres. But due to methodological shortcomings, many of the older data are highly uncertain and are not suitable for trend analysis. Data with a higher accuracy and a monthly resolution are available for the early 1990s onwards at the SMHI and FIMR data centres. However, the expected trend of 0.02 pH units per decade, which corresponds to the current increase in atmospheric CO_2 of 2.0 ppm $year^{-1}$, could not be definitively identified. This may be due to the large seasonal amplitude of the pH signal and/or to the considerable interannual variability (Wesslander et al. 2010). It is also possible that other changes in the biogeochemical conditions may have counteracted the pH decrease. With reference to the data of Buch (1945), alkalinity in the eastern Gotland Sea increased by about 100 µmol kg^{-1} during the last century. The effect on pH is roughly an increase of 0.03 units by which the effect of the increasing atmospheric CO_2 levels since the pre-industrial era (0.15 units) was reduced.

The increasing uptake of anthropogenic CO_2 will also decrease the concentrations of carbonate ions and thus decrease the state of the calcium carbonate saturation (Ω) which is defined by

$$\Omega = \left[CO_3^{-2}\right] \times \left[Ca^{2+}\right] / k_{sp}$$

where k_{sp} is the solubility product. This may affect the growth and possibly the survival of organisms that form shells consisting of calcite or aragonite which are the two biogenic $CaCO_3$ crystal forms. Historic calcite and aragonite saturation values can be calculated from Buch's (1945) measurements during 1928–1938. For surface water in the eastern Gotland Sea, a maximum oversaturation is obtained for calcite during midsummer (1.65), whereas a slight undersaturation (0.92) was found for aragonite. Despite an increase in atmospheric CO_2 by about 80 ppm since Buch's measurements, the surface water saturation for calcite and aragonite in midsummer did not decrease, but instead increased by a factor of 2.6 during the last 80 years according to pCO_2 measurements performed in 2005

(Schneider et al. 2009). This is mainly a consequence of the low pCO_2 that is caused by the eutrophication-related high production and that has completely overridden the effect of the rising atmospheric pCO_2. However, enhanced production implies increased mineralisation and high CO_2 concentrations in the deeper layers. Increased vertical mixing during winter therefore adds CO_2 to the elevated atmospheric CO_2 (Fig. 18.10). According to measurements in 2005 (Schneider et al. 2009), this results in a distinct calcite and aragonite undersaturation during winter of 0.56 and 0.31, respectively.

18.4.4 Burial of Carbon, Nitrogen and Phosphorus

18.4.4.1 Current Process Understanding

The biogeochemical cycling of C, N and P is linked via biological productivity (Mackenzie et al. 1993; Anderson et al. 2001), and their removal from the system through burial in the sediment is important for eutrophication and the CO_2 balance. When C, N and P in the sediment are no longer significantly interacting with the water column, they can be considered as buried. Buried C and N in the sediments of the Baltic Sea are mainly in the form of organic compounds, whereas a substantial part of P occurs as inorganic forms such as associates with Fe-oxyhydroxides and various calcium compounds (Conley et al. 1997; Carman 1998; Aigars and Carman 2001; Mort et al. 2010). Burial of P formed in the marine system removes P from the nutrient cycle, while refractory P transported from land is buried passively (Ruttenberg and Berner 1993; Delaney 1998). In the Baltic, reactive P (i.e. potentially bioavailable that can promote primary productivity, Anderson et al. 2001) is mainly buried in organic form (Mort et al. 2010).

It has been hypothesised that N and P are preferentially released during the mineralisation of organic matter and that the remainder that is deposited on the sea floor and buried is enriched in C (e.g. Froelich et al. 1982; Andersen and Jensen 1992; Ingall et al. 1993; Anderson et al. 2001). Hence, atomic ratios of organic C, N and P in sediment, compared to those in plankton (Redfield et al. 1963), have been used to draw conclusions regarding the decomposition or origin of the buried material (e.g. Froelich et al. 1982; Ruttenberg and Goñi 1997). However, atomic ratios are affected by many biogeochemical processes and properties of the sedimentation environment (e.g. Ingall and Van Cappellen 1990; Ingall and Jahnke 1994; Anderson et al. 2001; Jilbert et al. 2011). C:N:P ratios presented in the literature for the Baltic Sea sediments vary depending on sea area, bottom type and sediment depth (e.g. Carman and Cederwall 2001), the variation being clearly higher in the C:P ratio than that in the C:N ratio.

In the Baltic Sea, characteristic conditions that affect sedimentation and burial of C, N and P are, for example, high input of terrestrial material, shallow water column, intense production of organic matter and variable topography of the sea floor (Winterhalter et al. 1981; Carman and Cederwall 2001). Burial is most effective in the deep basins, which accumulate resuspended material from erosion and transportation areas. Carman and Cederwall (2001) estimated the contributions of erosion (19–33 %), transportation (38–46 %) and accumulation (28–41 %) areas to the sea floor of the Baltic Proper, the Gulf of Bothnia, the Gulf of Finland and the Gulf of Riga.

Figure 18.11 shows the spatial variation in particulate organic C content in the sediments (Leipe et al. 2011). Sedimentation rates of organic matter vary spatially, especially in the archipelagos because of the fractured topography (Winterhalter et al. 1981) and temporally because of the seasonality of plankton growth and senescence (Bianchi et al. 2002). Anoxia which is common in the Baltic Sea can favour burial of organic matter but diminish the burial of P because the formation of Fe-oxyhydroxides that bind PO_4 is inhibited (Van Cappellen and Ingall 1994; Ingall et al. 1993; Mort et al. 2010; Jilbert et al. 2011). In the Baltic Sea, Edlund and Carman (2001) reported higher organic C:P ratios for anoxic than for oxic sites and Jilbert et al. (2011) suggested increasing rates of P mineralisation in more reduced conditions. However, it is also conceivable that the shift to higher C:P ratios is caused by a slower organic C mineralisation. In oxic areas, benthic fauna affect burial of C, N and P (Aller 1988; Kristensen 1988; Andersen and Jensen 1991; Andersen and Kristensen 1992; Norkko et al. 2011).

18.4.4.2 Past Changes

It has been estimated that the sedimentation of organic matter and thus the burial of C, N and P increased in the Baltic Sea during the latter half of the twentieth century as a result of eutrophication (Jonsson and Carman 1994; Bonsdorff et al. 1997; Emeis et al. 2000; Vaalgamaa and Conley 2008). Jonsson and Carman (1994) suggested an increase in organic matter deposition by a factor larger than 1.7 between the 1920s and 1980s in the northern Baltic Proper and Emeis et al. (2000) estimated a fourfold increase in sediment accumulation in the Bornholm, Gdansk and Gotland basins since 1900. Shaffer (1987) estimated that in 1900, the burial rate of C may have been about 0.2 mol m^{-2} $year^{-1}$ in the Baltic Proper, compared to about 0.8 mol m^{-2} $year^{-1}$ in 1980. More recent literature presents variable C burial rates of up to about 5 mol m^{-2} $year^{-1}$ for the Baltic Proper. Burial rates reported for N and P in the Baltic Sea vary from 77 to 1286 mmol m^{-2} $year^{-1}$ and 2 to 51 mmol m^{-2} $year^{-1}$, respectively. According to Mort et al. (2010), burial of P has not clearly increased in the Baltic Proper, and Jilbert et al. (2011) suggested that burial of organic P has increased in hypoxic basins in the Baltic Proper because of enhanced burial of organic matter.

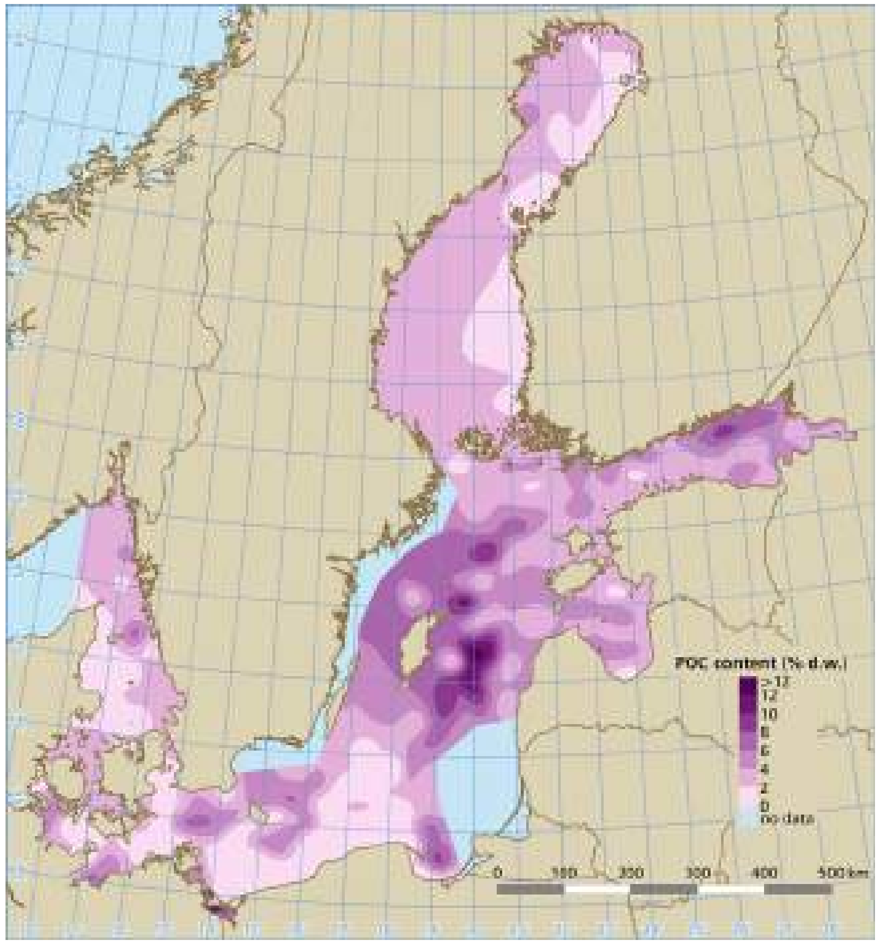

Fig. 18.11 Distribution of the percentage of particulate organic carbon (POC) in surface sediments in the Baltic Sea (Leipe et al. 2011)

Several authors have reported that increasing nutrient inputs and anoxia have altered the C:N:P ratios in the Baltic Sea sediments (e.g. Emeis et al. 2000; Hille 2006; Mort et al. 2010). For example, Emeis et al. (2000) reported a twofold increase in the C:P ratios in the Gdansk basin since 1950. Similar findings were made by Hille (2006) who analysed the elemental ratios of C, N and P in sediment cores from the Gotland Sea. The C:P ratios (Fig. 18.12a) increased by a factor of more than two during the last century. In contrast, the C:N ratios (Fig. 18.12b) remained relatively stable with values around 10 which is consistent with measurements performed at the beginning of the last century in different regions of the northern Baltic Sea (Gripenberg 1934).

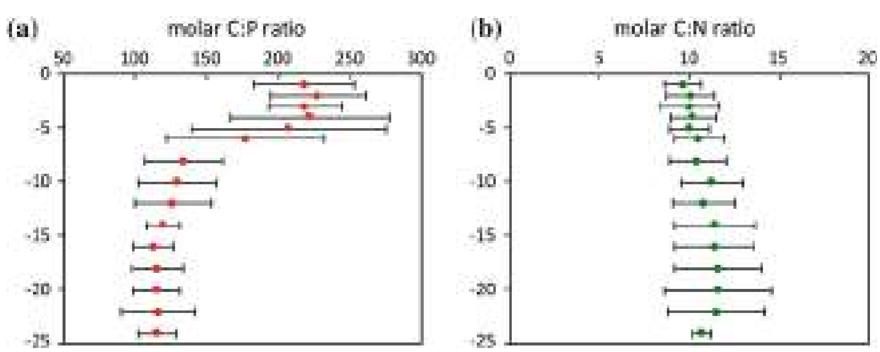

Fig. 18.12 Mean molar C:P (**a**) and C:N (**b**) ratios and standard deviation in sediments from the Gotland Basin. Sediment above 10–16 cm depth (*grey bar*) represents material accumulated during approximately the last 100 years (data from Hille 2006)

18.5 Response to Potential Future Changes

18.5.1 Nutrient Inputs

The HELCOM has adopted several recommendations since the 1970s to reduce nutrient loading of the Baltic Sea. A first attempt was to implement a 50 % reduction for N and P loads until 1995 (HELCOM 1988). This reduction target was not achieved (HELCOM 2009). With eight coastal countries being members of the European Union (EU), EU policies have become important for the Baltic Sea. Following the Water Framework Directive (WFD), nutrient reduction measures are to be installed that lead to a 'good ecological status' (European Parliament 2000). While the WFD is restricted to coastal waters, the EU Marine Strategy Framework Directive (MSFD, European Parliament 2008) refers to good environmental status in open sea areas as well as coastal waters.

Based on definitions of good environmental status, maximum allowable nutrient loads have been estimated. These loads are formulated as targets in the Baltic Sea Action Plan. The BSAP was adopted by the HELCOM member states in 2007. Compared to the period 1997–2003, a 19 % reduction for N and a 42 % reduction for P loads should be reached by 2021. However, the BSAP is currently (as of 2013) under revision and new reduction targets may be defined. While the P loads to the Baltic Sea can probably be reduced in the future, the perspective is less optimistic for the N loads. Even if the BSAP aims for a reduction in N loads, an increase in livestock may counteract this effort. With further economic development especially in some eastern European countries, animal protein consumption by humans is expected to increase. Since it is unlikely that increasing demand for animal proteins will be covered entirely by imports, a more likely scenario is an intensive development of the agricultural sector in the transitional countries. There is potential for large animal farms to be established in regions with low or inefficient current use of agricultural land. Such development could have a strong impact on the N flux from the catchment (see also Chap. 17). Eriksson Hägg et al. (2010) calculated that an increased protein consumption scenario would lead to a 16–39 % increase in N fluxes from the catchment. Taking into account climate-induced changes in river discharge, the increase in the N flux could be 3–72 % depending on the climate scenario. If N fluxes increase and P fluxes decrease, this would probably increase primary production in spring, but decrease the midsummer cyanobacteria blooms (Humborg et al. 2007).

Atmospheric deposition of N may increase in the future due to increased precipitation and increased shipping and agriculture. The recent share of atmospheric deposition in N loads is about 25 %, while the atmospheric load of P accounts for only a few percentage of the total load (HELCOM 2009).

18.5.2 Biogeochemical Cycles

Understanding of the impact of future climate change and other anthropogenic drivers on biogeochemical cycles is still limited. Several factors may influence the biogeochemical cycles in different ways. Hydrographic conditions affect the mixing depth, vertical exchange of nutrients and O_2, and inflow dynamics. Higher temperatures decrease the solubility of O_2 in sea water and accelerate many biogeochemical processes. Changing run-off and precipitation patterns influence the nutrient loads to the Baltic Sea. Taking into account that the Baltic Sea ecosystem is a complex, non-linear system, the effect of changes in external forcing is best quantified by numerical models. The BONUS (www.bonusportal.org) programme supports several projects that aim to quantify the response of the Baltic Sea ecosystem to climate change.

One focus of climate change effects on the Baltic Sea is the future development of O_2 conditions because it is a key parameter for biogeochemical cycles. Future warming will increase hypoxia through several interacting processes. Temperature controls stratification, mineralisation of organic matter and the solubility of O_2. Conley et al. (2009b) postulated that higher temperatures are likely to exacerbate hypoxia. Meier et al. (2011) showed with the aid of a model ensemble that future climate change could expand O_2-depleted regions. Reasons for increasing hypoxic and anoxic areas (Fig. 18.13) are increasing nutrient loads due to enhanced river run-off, reduced O_2 solubility and accelerated recycling of organic matter. The same model ensemble projected that hypoxic and anoxic periods in the main southern and central basins of the Baltic Sea could become more frequent and last longer (Neumann et al. 2012). This may lead to intensified P mobility and reduced denitrification efficiency. As a consequence, the pelagic DIN and DIP concentrations as well as primary production would increase (Meier et al. 2012). Such changes in biogeochemical fluxes imply that even if nutrient loads are reduced according to the BSAP, the Baltic Sea ecosystem would not recover but would stabilise close to its present state.

Furthermore, warming will preferentially favour cyanobacteria blooms. Blooms are expected to start earlier in summer, and N fixation might increase (Neumann 2010; Neumann et al. 2012; Meier et al. 2012). It was also shown that increasing nutrient loads due to climate-induced change in river run-off have a smaller impact on the modelled phytoplankton development than changes caused by socio-

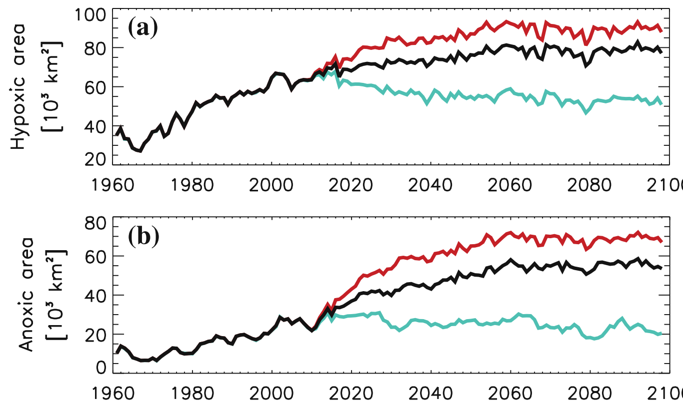

Fig. 18.13 Simulated development of (**a**) hypoxic (<2 mL O_2 L^{-1}) and (**b**) anoxic (<0 mL O_2 L^{-1}) areas within the Baltic Sea for three nutrient load scenarios: the BSAP (*blue*), reference (see Sect. 18.3.1) (*black*) and business-as-usual (*red*) (Meier et al. 2011)

economic development. The pessimistic 'business-as-usual' scenario for nutrient loads could lead to increasing adverse effects on the marine environment in a future climate (Meier et al. 2011).

Omstedt et al. (2012) used a complex model system that included biogeochemical processes in the catchment area (see Chap. 17) to simulate the future development of the hypoxic areas. For both the business-as-usual and the BSAP scenarios, they found almost the same long-term trend pattern as Meier et al. (2011) (Fig. 18.13), however, with an overall offset of about 20,000 km^2.

18.5.3 Acidification

In addition to eutrophication, ocean acidification (decreasing pH) caused by increasing atmospheric CO_2 is considered a major potential threat for the Baltic Sea. Taking into account only the effect of the CO_2 increase and ignoring other potential changes in the marine CO_2 system, the pH decrease may be calculated on the basis of the acid/base equilibria in sea water. Assuming the worst-case scenario, which is an increase from about 400 ppm CO_2 at present to 950 ppm at the end of the century (van Vuuren et al. 2011), yields a decrease in the mean pH in the Baltic Sea surface water of about 0.4 pH units. Since pH is a logarithmic quantity, this is equivalent to an increase in hydrogen ion (H^+) concentration by a factor of 2.5. This corresponds to a decrease in calcium carbonate saturation by a factor of approximately 2 and may affect the survival of organisms such as mussels that form calcite or aragonite shells (Brander and Havenhand 2011).

To estimate the possible future development of pH in the Baltic Sea until 2100, Omstedt et al. (2012) developed a biogeochemical model that included the cycling of C and thus the marine CO_2 system which mainly controls the acid/base system in sea water. The model was coupled with a catchment model in order to account for climate-induced change in the riverine inputs of total CO_2 and alkalinity and of dissolved organic C. Both these classes of C compound affect sea water pH, either directly or in case of organic C by mineralisation. Their release by weathering and soil organic matter transformation and transport in a changing climate were simulated by sub-models for the catchment area. Simulations with the coupled models were performed for a variety of different climate, nutrient load and land cover scenarios. The results were used to identify the contributions of the different forcing factors to a possible change in sea water pH.

Figure 18.14 shows the simulated pH development until 2100 for a best-case and a worst-case forcing scenario (green and red line, respectively). The best case refers to a climate change scenario based on CO_2 emission reductions resulting in atmospheric CO_2 concentrations of about 550 ppm in 2100 (ECHAM, B1). It is combined with reduced nutrient loads to the Baltic Sea according to the BSAP. For the worst case, a climate change scenario (ECHAM, A2) is used that yields final CO_2 concentration of about 850 ppm and the nutrient loads are assumed to increase according to a business-as-usual scenario. A decrease in pH of about 0.40 and 0.26 in surface waters of the central Baltic Sea is obtained for the worst-case and best-case scenarios, respectively (Fig. 18.14). Analysis of the model simulations showed that changes in surface water pH are mainly due to rising atmospheric CO_2 concentrations. Increasing nutrient loads that cause enhanced biological production only increased the seasonal pH amplitude; they had no effect on mean pH.

Temporal change in the pH of the Gotland Sea deep water is less regular because of the interplay between stagnation periods and water renewal events. Nevertheless, the model simulations show a pH decrease for both scenarios. However, in the best-case scenario, pH declines only until 2000, then shows no further change. In contrast, for the worst-case scenario, the simulations show a continuous decrease in deep-water pH (Fig. 18.14). Furthermore, model experiments indicate that pH in the deep water results not only from increasing atmospheric CO_2 but is also partly controlled by rising temperatures and changing inputs from the catchment area.

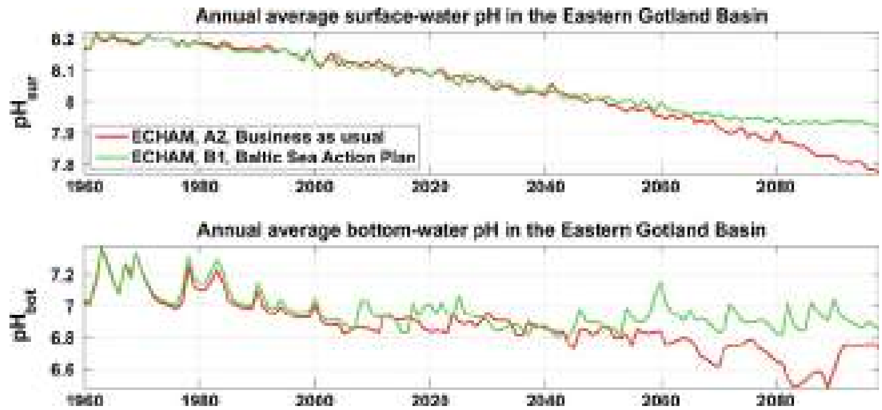

Fig. 18.14 Simulation of surface water and deep-water pH in the eastern Gotland Sea. The temporal development was calculated for (1) a worst-case scenario: atmospheric CO_2 increase to 850 ppm and nutrient inputs according to business-as-usual (*red line*) and (2) a best-case scenario: moderate increase in atmospheric CO_2 to 550 ppm and nutrient inputs according to the Baltic Sea Action Plan (*green line*) (Omstedt et al. 2012)

18.6 Conclusion

The first assessment of climate change in the Baltic Sea (BACC Author Team 2008) did not include a chapter explicitly dedicated to Baltic Sea biogeochemistry. However, some aspects such as nutrient inputs, the relationship between nutrient availability and biomass production, and the consequences of eutrophication for deep-water O_2 conditions were addressed in Chap. 5 *Climate-related Marine Ecosystem Change*. Since then, much progress has been achieved, particularly in the reconstruction of past changes and model simulations of future eutrophication scenarios. This is mainly due to BONUS-funded research and is presented here. In general, this chapter has focussed on the description of the basic processes controlling the complex relationship between nutrient transformations and organic matter production and mineralisation. This implies a consideration of both the O_2/H_2S budget and the marine CO_2 system, and so is related to both eutrophication and acidification.

Data on the concentrations of nutrients, O_2 and H_2S in the Baltic Sea are available for about the last 40 years and clearly indicate that major biogeochemical changes have occurred. Nitrate and PO_4 concentrations in the winter surface water of the Baltic Proper increased by a factor of approximately 3. This is consistent with the enhanced nutrient inputs by river water and atmospheric deposition during the last century. However, inputs have decreased since about 1980 and are currently at a level that is roughly comparable to that in 1960. This is reflected in a decrease in NO_3 concentrations in the winter surface water of the Baltic Proper since 1980. A similar trend does not exist for PO_4 concentrations. This is partly explained by the long residence time of P in the Baltic Sea and its temporary storage as PO_4–Fe-oxyhydroxide. But considerable uncertainties in the input estimates must also be taken into account. As a consequence of the increasing nutrient concentrations that mainly control the net organic matter production and the export into deeper water layers, the frequency of H_2S occurrence has increased. This may have enhanced the recycling of PO_4 and favoured blooms of N-fixing cyanobacteria. Data concerning the marine CO_2 system are available for the 1920s and 1930s. They indicate that net biomass production has increased since then by a factor of about 2.5. The data also show that alkalinity in the Baltic Proper has increased and has considerably damped the decline in pH caused by rising levels of atmospheric CO_2.

Model simulations of scenarios concerning the future biogeochemistry of the Baltic Sea indicate that climate change is likely to exacerbate eutrophication effects in the Baltic Sea. However, they also indicate that the implementation of the BSAP could help to decrease slightly the area covered with hypoxic and anoxic waters despite climate change effects. In contrast, the business-as-usual nutrient input scenario increased the hypoxic area by about 30 % and more than doubled the anoxic area. Simulations of the future development of pH in the Baltic Sea were performed with a model system that included the cycling of organic C and CO_2 in the Baltic Sea as well as the catchment area. It was shown that the rising atmospheric CO_2 mainly controls future pH changes in Baltic Sea surface water and that eutrophication and enhanced biological production are not affecting the mean pH. The worst-case CO_2 emission scenario (resulting in an atmospheric concentration of 850 ppm) projected a fall in pH in Baltic Sea surface water of about 0.40 by 2100. A more optimistic future CO_2 emission scenario (550 ppm) yielded a decrease in pH of approximately 0.26.

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19 Environmental Impacts—Marine Ecosystems

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Abstract

Increase in sea surface temperature is projected to change seasonal succession and induce dominance shifts in phytoplankton in spring and promote the growth of cyanobacteria in summer. In general, climate change is projected to worsen oxygen conditions and eutrophication in the Baltic Proper and the Gulf of Finland. In the Gulf of Bothnia, the increasing freshwater discharge may increase the amount of dissolved organic carbon (DOC) in the water and hence reduce phytoplankton productivity. In winter, reduced duration and spatial extent of sea ice will cause habitat loss for ice-dwelling organisms and probably induce changes in nutrient dynamics within and under the sea ice. The projected salinity decline will probably affect the functional diversity of the benthic communities and induce geographical shifts in the distribution limits of key species such as bladder wrack and blue mussel. In the pelagic ecosystem, the decrease in salinity together with poor oxygen conditions in the deep basins will negatively influence the main Baltic Sea piscivore, cod. This has been suggested to cause cascading effects on clupeids and zooplankton.

Keywords

Baltic Sea • Climate change • Benthic and pelagic communities • Biodiversity • Biogeography • Regime shifts • Cascading effects

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_19

19.1 Introduction

The Baltic Sea has undergone rapid post-glacial succession over the past 10,000 years (see Chap. 2). This has meant a succession of different types of communities, resulting in changes in the marine ecosystem. Superimposed on this geological-scale shift, there have also been more rapid changes in the physics, chemistry and biology that can be connected to climate (BACC Author Team 2008). Most importantly, the Baltic Sea system responds to climatological variations in the North Atlantic region, characterised by the North Atlantic Oscillation (NAO, see Chap. 4) and its local counterpart the Baltic Sea Index (BSI) (Matthäus and Schinke 1994; Hänninen et al. 2000; Lehmann et al. 2002). The Baltic Sea ecosystem is also influenced by human-induced global warming, that has been projected to affect various oceanographic parameters in the Baltic Sea in the coming decades (e.g. Meier et al. 2012 and Chap. 13).

The complexity of the interactions and feedbacks between the atmosphere, watershed and marine ecosystem make it difficult to distinguish anthropogenic effects, such as eutrophication, from those caused by changes in climate. This distinction is important to make when designing measures to improve the status of the marine environment by adjusting human actions.

Climate effects may be subdivided into direct and indirect effects. *Direct effects* influence the individual organisms, their metabolism, growth, survival and productivity, through changes in the properties of the water surrounding them (temperature, salinity, chemical composition and other properties, such as stratification and mixing). *Indirect effects* influence the structure of communities by changing species interactions, or by shaping temporal and spatial match or mismatch of populations. Such interactions affect the intensity of grazing, predation and competition, and shape the structure of communities and thus affect biodiversity, trophic relationships and ecosystem functioning (Sommer and Lengfellner 2008; Sommer and Lewandowska 2011). Both types of effects may result in changes in species' ecology and eventually evolution. The responses will together define the realised niches of species in a world affected by changing climate (Lavergne et al. 2010).

Among the climate-driven factors that directly affect individuals, temperature is fundamental. It drives biological processes and ultimately underpins life history traits, population growth and ecosystem processes. Effects range from those directly affecting metabolism of bacteria (e.g. Autio 1998) to those affecting the development of invertebrates (O'Connor et al. 2007) and fish larvae (e.g. Karås and Neuman 1981; Hakala et al. 2003). Temperature may also affect populations indirectly by influencing their food supply (Hoegh-Guldberg and Bruno 2010).

Salinity is also a fundamental factor, because most of the Baltic Sea species are of either marine or freshwater origin and many live at the edge of their salinity tolerance. Other climate-driven environmental factors that have direct effects on individuals include oxygen concentration and the acidity (pH) of the seawater. Climate also indirectly affects species and populations by affecting water stratification, mixing depth and availability of nutrients, thus influencing the primary producers as well the animals feeding on them. For a summary of effects see Table 19.1.

19.2 Community-Level Variations in the Past

19.2.1 Phytoplankton

Evidence for community-level changes in phytoplankton exists for the northern Baltic Sea. For example, a 30-year dataset from the Gulf of Finland showed that the spring bloom (measured as chlorophyll *a*) declined from 1986 to 2002, while the summer phytoplankton biomass increased from 1990 to 2002 (Raateoja et al. 2005). The decrease in the spring bloom was probably caused by decreased availability of nitrogen in spring, which resulted in less phosphorus consumed, and hence larger phosphorus reserves for the summer. This in turn benefited the nitrogen-fixing summer cyanobacteria.

Table 19.1 Summary of short-term effects of freshwater discharge and temperature on the microbial food web and phytoplankton as reported in the current literature

Food web component	Higher freshwater discharge	Higher temperature
Phytoplankton	Reduced productivity	Lower biomass and size/no consensus
Bacterioplankton	Maintained growth	Increased growth
Protozooplankton	Maintained growth	Increased growth
Sedimentation	Reduced amount and quality	Reduced amount and quality
Food web efficiency	Reduced efficiency	No consensus

Where effects of climate drivers are similar, synergistic consequences for the food web are expected

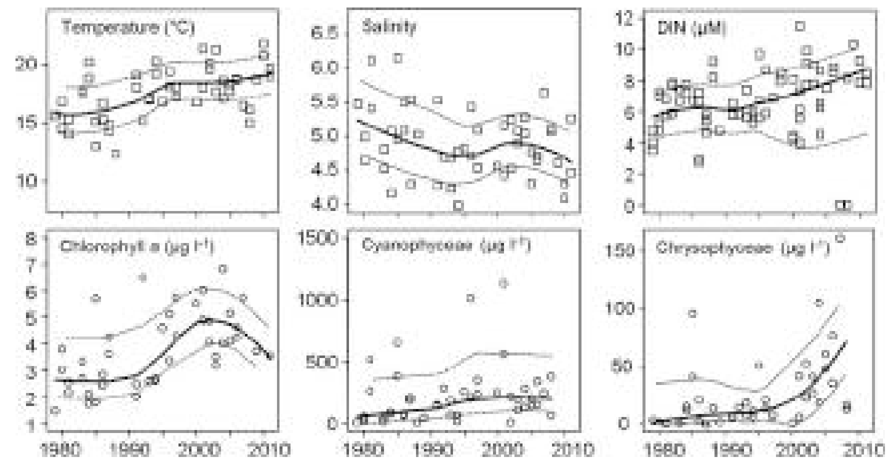

Fig. 19.1 Trends in sea surface temperature, salinity, dissolved inorganic nitrogen (DIN), chlorophyll a, Cyanophyceae and Chrysophyceae in the Gulf of Finland, 1979–2008. Redrawn from Suikkanen et al. (2013)

In a study covering 1979–2008, the summer biomass of cyanobacteria, chrysophytes and chlorophytes increased in the northern Baltic Proper and the Gulf of Finland, while the biomass of cryptophytes decreased (Suikkanen et al. 2007, 2013; Fig. 19.1). Increase in sea surface temperature in summer, decline in salinity and an increase in the winter dissolved inorganic nitrogen to phosphorus ratio (DIN:DIP) were the most important factors explaining the changes observed in the phytoplankton community. Both climate-induced and anthropogenic factors, especially nutrient loading, were concluded to have contributed to the changes.

Further evidence for climate effects on phytoplankton comes from studies of dominance changes between diatoms and dinoflagellates. In the Baltic Sea, the spring bloom is dominated by just a few species of diatoms and cold-water dinoflagellates (Kremp et al. 2008). Their relative proportions depend on temperature and ice conditions, with warm winters favouring dinoflagellates (Wasmund et al. 2011; Klais et al. 2011, 2013). Consequently, it has been suggested that warming promotes a dominance by dinoflagellates over diatoms in the Baltic Sea spring bloom (Kremp et al. 2008; Olli and Trunov 2010). However, although dinoflagellates have increased in the central and northern Baltic Proper (Wasmund and Uhlig 2003; Suikkanen et al. 2007; Klais et al. 2011), the data do not support an increase in spring bloom dinoflagellates for the Baltic Sea as a whole. The response of dinoflagellates to large-scale climate patterns seems to depend on local hydrography and community composition (Klais et al. 2013).

There are only a few studies of the actual mechanisms by which dinoflagellates are favoured in warm conditions. In the northern Baltic Sea, the dinoflagellate *Biecheleria baltica* has expanded its range in recent decades. This expansion has been linked to the exceptionally efficient benthic cyst production of this species (Olli and Trunov 2010). Given the rise in deep-water temperatures in the Baltic Proper, the benthic cyst germination of *B. baltica* may have been enhanced and contributed to the spring dominance of this species (Kremp et al. 2008).

19.2.2 Zooplankton

It is well known that oceanic zooplankton respond to variations in climate (e.g. Beaugrand and Reid 2003; Hays et al. 2005). The response has usually been explained as a bottom-up process: the atmospheric forcing influences primary productivity through hydrography and stratification of water, which changes mixing depth, light and availability of nutrients. This in turn affects grazing zooplankton, planktivorous fish and even seabirds (Aebischer et al. 1990; Frederiksen et al. 2006).

Similar responses to climate or weather parameters have also been shown in the Baltic Sea. For instance in the Gulf of Finland, there was a significant positive correlation between westerly winds and the abundance of marine copepods, whereas easterly winds promoted a high abundance of small euryhaline cladocerans (Viitasalo et al. 1995). This could have been caused by westerly winds advecting saline water—and marine species—from the northern Baltic Proper into the Gulf of Finland, and it remained unclear whether any trophic effects were involved. In contrast, rotifers and cladocerans were positively associated with high sea surface temperatures in summer (Viitasalo et al. 1995). In this case, the warm water probably increased growth rates of these surface-dwelling taxa and also increased the availability of suitable small-sized food in the strongly stratified water (Kivi et al. 1993).

In an analysis of open-sea monitoring data for the Gulf of Finland and northern Baltic Proper in 1979–2008, rotifers also increased at the expense of crustacean zooplankton, apparently due to the combined effects of changes in hydrography, eutrophication and top-down pressure (Suikkanen et al. 2013). This resulted in a shift towards a food web structure with smaller sized organisms.

Few studies have investigated the effects of climate-related parameters on shallow-water zooplankton dynamics. Scheinin and Mattila (2010) concluded that temperature and total phosphorus (an indicator of eutrophication) were the most important factors shaping littoral zooplankton communities in the northern Baltic Sea, which suggests that they could be affected by climate-induced warming and eutrophication. Kotta et al. (2009) and Põllumäe et al. (2009) suggested, however, that small-scale environmental variability probably masks the response of shallow pelagic communities to climatic variation.

In addition to trophic effects, climate-related parameters also have direct physiological effects on Baltic Sea zooplankton. For example, *Pseudocalanus acuspes*, a marine species, responded negatively to the 1980s decline in salinity (Lumberg and Ojaveer 1991; Vuorinen et al. 1998; Möllmann et al. 2000), whereas the more euryhaline copepods *Acartia* spp. and *Temora longicornis* increased, apparently due to rising spring temperature in the central Baltic Sea (Möllmann et al. 2000).

Other mechanisms by which climate change may affect zooplankton are life cycle effects. Many Baltic Sea copepod species overwinter as resting eggs in the sediments, and a connection between the timing of spring warming of the water and timing of peak population in spring has been detected (Viitasalo 1992). In many coastal areas, like the Gulf of Riga (Kotta et al. 2009), Darß-Zingst lagoon (Feike et al. 2007), Swedish east coast (Hansson et al. 2010) and Archipelago Sea (Dippner et al. 2001), mesozooplankton has been shown to respond positively to mild winters. This suggests that climate change may influence the benthic germination of copepods in the same way as shown for certain dinoflagellates (Kremp et al. 2008).

19.2.3 Open-Sea Benthic Communities

The benthic community of the open Baltic Sea is a mix of species with marine, brackish water and limnic origins. Their latitudinal distribution and species diversity are limited by the gradient of decreasing salinity towards the north (Elmgren 1989; Rumohr et al. 1996; Bonsdorff and Pearson 1999). Spatial studies of species turnover (β-diversity), in the transition zone between the North Sea, Skagerrak, Kattegat and the Belt Sea (Josefson 2009) and along the salinity gradients of the Baltic Sea (Laine et al. 1997; Bleich et al. 2011; Villnäs and Norkko 2011), all demonstrate the key role of salinity in the distribution of macrobenthic animals. The distribution of benthic species is also driven by strong vertical gradients: shallow-water soft bottom communities have higher habitat diversity and thus more species than the sub-halocline communities (e.g. Andersin et al. 1978).

Climate affects both salinity and stratification and consequently the distribution of hypoxia in the Baltic Sea, and it is therefore natural that climatic variation also affects zoobenthos. The earliest evidence for such an interaction comes from reported responses to saline water inflows that enter the Baltic Sea through the Danish Straits (see also Chap. 7). In the mid-1950s, the large saltwater inflow was accompanied by significant range expansions of many benthic marine species into the Baltic Proper (Segerstråle 1969). This process has reversed in the 1980s and 1990s when a decrease in salinity has led to a dominance shift from marine to brackish water taxa in the southern and western parts of the Baltic Sea (Villnäs and Norkko 2011). As the hypoxia is currently the most widespread on record, due both to eutrophication and to climatic conditions favouring stagnation (Conley et al. 2009, see also Chap. 18), benthic communities in the central Baltic Proper are presently in a poor state.

The saline water inflows have different effects in different basins of the Baltic Sea. For example, in the late 1980s and early 1990s, the benthic communities were in a poor state due to long-lasting stagnation. Meanwhile the Gulf of Finland had relatively good oxygen conditions and abundant macrozoobenthos. The major saline water inflow through the Danish Straits in 1993 again improved oxygen conditions in the Baltic Proper, but simultaneously pushed the stagnant oxygen-depleted water from the central Baltic Sea basin into the Gulf of Finland. This water reached the eastern Gulf of Finland by autumn 1995 to spring 1996 and caused a dramatic decrease in the macrozoobenthos (Laine et al. 2007; Savchuk 2010). The situation later reversed: in the Gulf of Finland, the benthic communities recovered, while in the stagnant Baltic Proper, anoxia gradually developed and killed the benthic communities.

Benthic conditions in the Baltic Proper and the shallower Gulf of Finland thus respond to climate-induced oceanographic variations with different time lags.

19.2.4 Shallow-Water Benthic Communities

Benthic communities on shallow-water hard and soft bottoms in the Baltic Sea differ from the deeper communities owing to light conditions enabling an abundance of algae and vascular plants. The communities are often based on a few structurally and functionally important species, such as fucoids, seagrasses, blue mussels and a few other habitat-forming species (e.g. Wallentinus 1991; Kautsky and Kautsky 2000). As in the case of deep benthos, the composition and species richness of shallow benthic communities decline from the Kattegat and southern Baltic Sea towards the north and east (Nielsen et al. 1995; Ojaveer et al. 2010).

The spatial distributions of shallow-water benthic species are determined foremost by their ability to adapt to low salinity, but also to variations in sea level, wave exposure, light and oxygen availability, grazing, predation and competition for space (Kautsky and Kautsky 2000 and references therein).

To date, very few studies have explicitly investigated the mechanisms by which climatic variations can affect shallow benthic and algal communities in the Baltic Sea. Mild winters have been shown to result in a denser growth of *Fucus* near the surface (Kiirikki and Ruuskanen 1996), at the same time promoting a higher production of associated invertebrate fauna (Wikström and Kautsky 2007). This was explained by less ice scraping in milder winters: ice scraping effectively removes key species in the uppermost part of the algal belt and thus delays development of the algae and associated fauna. Also in Haapsalu Bay, eastern Baltic Proper, changes in the macrophyte communities over the past 50 years have been affected by weather-induced changes in salinity and ice conditions (Kovtun et al. 2009).

The reproductive phase of many macroalgae is especially sensitive to seasonal changes in salinity. Laboratory experiments have shown that fertilisation of Baltic *Fucus serratus* is high at a salinity of 9, but very low at a salinity of 6, typical of the coasts in the northern Baltic Proper (Malm et al. 2001). Also *Fucus vesiculosus* needs relatively high salinity during its main reproductive periods in May–June and September–October (Serrão et al. 1999; Malm et al. 2001). However, no clear link between climate-induced changes in salinity and long-term changes in *Fucus* populations has to date been demonstrated.

19.3 System-Level Variations in the Past

19.3.1 Regime Shifts

Theory, experiments and field data all indicate that a gradual change in external drivers can result in an abrupt, nonlinear change in the ecosystem—a regime shift (e.g. Scheffer et al. 2001; Folke et al. 2004; Andersen et al. 2009). In several sea areas of the northern hemisphere, populations of predatory fish have declined, often due to overfishing, causing a large-scale reorganisation of the ecosystem in question (e.g. Myers and Worm 2003; Collie et al. 2004). Similar system-level changes have also taken place in the Baltic Sea during the past century, but these shifts have probably been caused by a combination of climatic and anthropogenic effects, including overfishing and eutrophication.

Around 1935–1955, an ‘oceanisation’ of the Baltic Sea took place: salinity increased and various pelagic and benthic taxa such as marine copepods, the jellyfish *Cyanea capillata*, the barnacle *Balanus improvisus*, and cod (*Gadus morhua*), garfish (*Belone belone*) and mackerel (*Scomber scombrus*) spread hundreds of kilometres northwards, while the ranges of species preferring low salinity retreated (Segerstråle 1969). The system was characterised by marine species and relatively good oxygen conditions.

In the late 1950s, the oceanisation ended and, as the anthropogenic nutrient load increased, primary production started to increase (Larsson et al. 1985; Elmgren 1989; Stigebrandt 1991). This resulted in increased sedimentation and deep-water oxygen consumption from the early 1960s onwards (Elmgren 1984; Conley et al. 2002). A particularly severe hypoxia of the deep waters developed in the 1980s, during a long ‘stagnation period’ with less inflow of oxygen-rich water from the North Sea (Fonselius and Valderrama 2003). Due to the hypoxia, the macrobenthic bottom fauna below the halocline was eliminated, which caused a major disruption of the benthic food web (Elmgren 1989; Norkko et al. 2010).

The hypoxia also changed the food availability for demersal fishes. While cod in the early twentieth century fed mainly on benthic organisms, once the macrobenthos disappeared the cod switched to pelagic fish, mainly sprat (*Sprattus sprattus*) and herring (*Clupea harengus membras*) (Eero et al. 2011). Cod reproduction declined (due to declining egg survival in low oxygen conditions), while cod fisheries remained intense, and cod stocks consequently collapsed. This process fundamentally changed the structure and functioning of the upper pelagic trophic levels of the Baltic Sea in the late twentieth century (see also Sects. 19.3.2 and 19.4.4).

A multivariate time-series analysis suggested that feedback loops were established in the biotic part of the ecosystem, preventing the system from switching back to its previous state (Möllmann et al. 2008, 2009; Casini et al. 2009)—a phenomenon referred to as hysteresis (Scheffer et al. 2001). In the other Baltic Sea areas, the situation may be different. Lindegren et al. (2010a) demonstrated a shift in ecosystem composition in the Sound (the transition area between the western Baltic Sea and the Kattegat). Here, the shift did not show signs of trophic cascade or hysteresis and may therefore be more easily reversible than in areas where the system has entered a new stable state.

19.3.2 Cascading Effects in the Pelagic Ecosystem

In the central and northern Baltic Sea, upper trophic levels of the pelagic ecosystem, including zooplankton, planktivores and piscivores, are all influenced by hydrography. Marine zooplankton typically declines during periods of low salinity, while certain species are favoured by warm water. Temperature also influences sprat recruitment (e.g. MacKenzie and Köster 2004; Cardinale et al. 2009) because sprat eggs survive better during mild winters (Nissling 2004). Baltic Sea cod reproduction, in turn, is dependent on the volume of sufficiently saline and oxygenated water, termed 'cod reproductive volume' (RV) (MacKenzie et al. 2000). After spawning, cod eggs sink to the depth where they are neutrally buoyant, i.e. at a salinity of ~11, and if this water is too low in oxygen, the eggs die (Wieland et al. 1994). The RV is dependent on the water balance as well as density stratification in the Baltic Sea. All these factors are influenced by climatic processes in the Baltic Sea and in the North Atlantic (e.g. Matthäus and Franck 1992).

The climatic and hydrographic control of the upper trophic levels of the pelagic ecosystem is exemplified by the 'regime shift' described in the previous section. From the early 1980s to mid-2000s, only one major inflow took place (in 1993), salinity gradually declined, the anoxic layer expanded and cod reproduction collapsed (1986–1993) in the central Baltic Sea (Köster et al. 2005). The sprat stocks

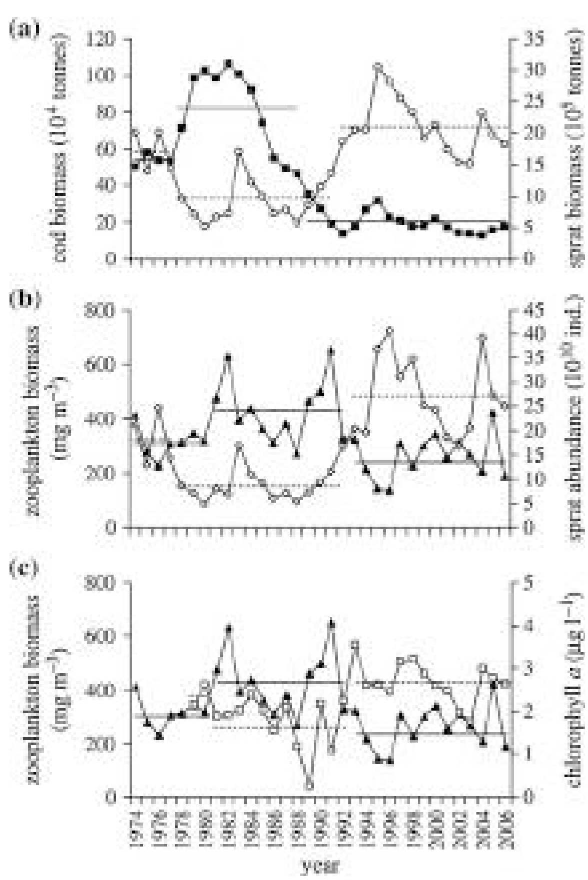

Fig. 19.2 Trends in **a** cod biomass (*squares*) and sprat biomass (*circles*); **b** zooplankton biomass (*triangles*) and sprat abundance (*diamonds*) (ind., individuals); and **c** zooplankton biomass (*triangles*) and chlorophyll *a* (*squares*). The *horizontal lines* indicate periods of different average levels in the biological time series as detected by the cumulative z-scores. Casini et al. (2008)

expanded sixfold soon after the collapse of cod stocks, during 1988–1995. A simultaneous decrease in zooplankton suggests that the effects of the cod collapse cascaded down from planktivores to zooplankton (Casini et al. 2008; Fig. 19.2). The marine zooplankton also declined because of direct physiological stress caused by declining salinity, and consequently, there was less food available for planktivores. This eventually caused starvation and low growth of clupeids, especially herring, in the late 1980s (Flinkman et al. 1998; Rönkkönen et al. 2003; Möllmann et al. 2005) (Fig. 19.3).

It is notable that the state of the cod stock in the Baltic Sea is not only driven by salinity and oxygen variation. This became clear when the eastern Baltic Sea cod stock started to recover in 2005, after more than two decades of low biomass and productivity, and despite the continuing 'cod hostile' (i.e. low oxygen) environment (Cardinale and Svedäng 2011). The recovery was mainly driven by a sudden reduction in fishing mortality and occurred in the absence of any exceptionally large year classes. This suggests that effects of fisheries may at times override environmental factors as controllers of cod stocks in the Baltic Sea. The observation launched a vivid debate on the causes and consequences of the variations in the cod stocks in the Baltic Sea (Möllmann et al. 2011).

19.3.3 Microbial Food Web

Enhanced freshwater and nutrient discharge to the central Baltic Sea is usually thought to increase primary production and phytoplankton biomass. This is probably the case in areas where phytoplankton production is limited by the availability of inorganic nutrients. In contrast, the effects of climate change on the dynamics of the microbial food web may be less straightforward.

In the Gulf of Bothnia, increased freshwater runoff may lead to enhanced microbial activity and a *decrease* in phytoplankton primary production. This is because, in addition to nutrients, river discharge in this area carries a large load of dissolved organic carbon (DOC). The DOC reduces the amount of light reaching the phytoplankton (Pettersson et al. 1997) and also serves as a substrate for bacteria which at carbon substrate sufficiency may outcompete phytoplankton for nutrients (Mindl et al. 2005).

Such an outcome was demonstrated in a field study in the Gulf of Bothnia where the effect of river discharge was investigated for 13 years (1994–2006) (Wikner and Andersson 2012). A marked increase in the ratio of bacterioplankton production to phytoplankton production was observed when river discharge was elevated. This happened despite the increased availability of inorganic nutrients. In addition to lower light availability and competition for

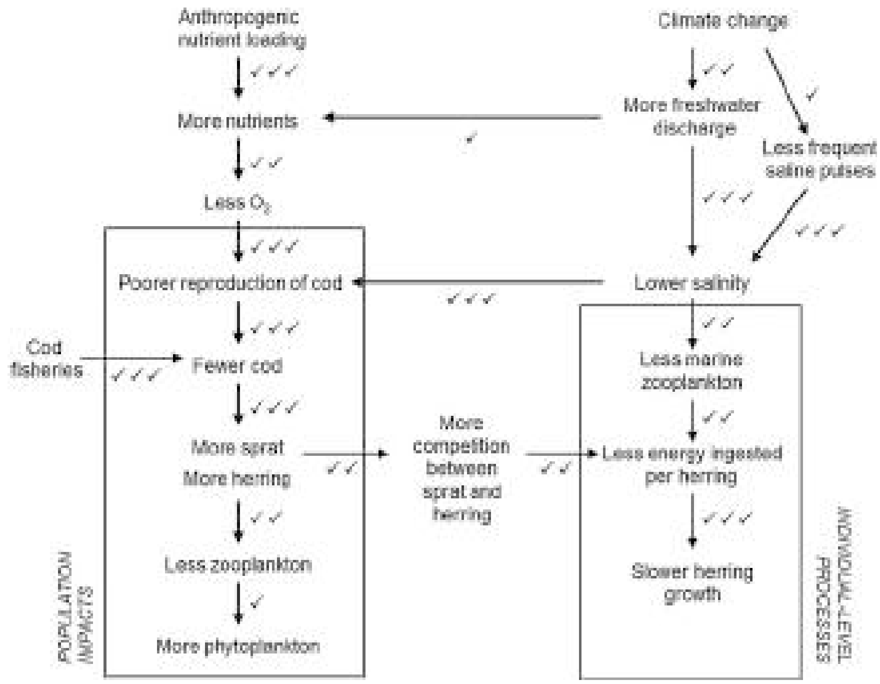

Fig. 19.3 A schematic representation of the interactions between different trophic levels and some environmental factors in the pelagic ecosystem of the central Baltic Sea. Estimated confidence levels depicted as ✓ low confidence, ✓✓ medium confidence, ✓✓✓ high confidence (based on the judgement of the chapter authors)

nutrients, this may have been caused by a shortage of the limiting nutrient phosphorus due to chemical binding to iron (Wikner and Andersson 2012). This field observation was supported by mesocosm studies with natural plankton communities, where addition of a DOC source and a corresponding reduction in light resulted in a net heterotrophic food web (Berglund et al. 2007; Dahlgren et al. 2011).

19.4 Potential Future System-Level Responses to Climate Change

19.4.1 Pelagic Dynamics

Biogeochemical models project an increase in freshwater discharge and associated nutrient loads (e.g. Meier et al. 2011). Consequently a 5 % increase in total phytoplankton biomass in the Baltic Sea has been projected by one ecosystem model (Neumann 2010). Models suggest that freshwater discharge may increase more in the forested northern catchments, where run-off concentrations of nutrients are low, and may decrease from the agriculture-dominated and nutrient-rich southern rivers (HELCOM 2007). This would suggest that climate change enhances primary production in the northern basin (Gulf of Bothnia) but may alleviate eutrophication in the more southern basins. However, the modelling results of Meier et al. (2012) suggest a larger increase in spring phytoplankton biomass (in 'business-as-usual' and 'reference' nutrient load scenarios) in the Gulf of Finland and Baltic Proper than in the Gulf of Bothnia.

As explained in the previous section, the Gulf of Bothnia is less prone to discharge driven eutrophication because of the characteristic dynamics of DOC. Future primary production does not depend only on the climate-induced increase in external nutrient loading, but also on the nutrient ratios of the discharge in each catchment, as well as on the internal biogeochemical processes and feedbacks (e.g. internal loading from anoxic sediments) in each basin.

The environmental factor that will change most due to climate change is probably sea surface temperature. Higher surface-water temperatures will favour taxa thriving in warmer stratified water, such as the cyanobacterium *Nodularia spumigena* (Kononen and Nômmann 1992), and prolonged cyanobacterial blooms have been projected (Neumann 2010). Such a shift could change the main productive period from spring to summer.

Changes in zooplankton communities are also expected. Viitasalo et al. (1995) and Suikkanen et al. (2013) suggested that an increase in water temperature will favour the smaller sized 'surface zooplankton community', especially rotifers and small cladocerans.

Climate change may also influence zooplankton by affecting the quality of their food. In a mesocosm study simulating effects of climate change, increasing temperature negatively affected food quality, since heterotrophic production increased and zooplankton fed more on items deficient in important fatty acids (Dahlgren et al. 2011).

The increase in cyanobacteria may decrease food quality for zooplankton, because cyanobacteria are low-quality food for zooplankton (Karjalainen et al. 2007 and references therein). On the other hand, mesocosm studies have shown that copepods can feed and reproduce in a decaying cyanobacteria bloom, apparently by feeding on ciliates and other organisms of the microbial loop thriving in such conditions (Engström-Öst et al. 2002; Koski et al. 2002). The net outcome for zooplankton production in the field is difficult to predict.

In mesocosm experiments simulating climate change effects on light conditions, temperature and grazing pressure on phytoplankton dynamics, a reduced phytoplankton biomass and cell size was observed, partly due to increased activity of overwintering zooplankton (Lewandowska and Sommer 2010; Sommer and Lewandowska 2011). Climate-induced changes in the zooplankton community may thus indirectly shape the future phytoplankton community composition by changing the type of phytoplankton species consumed.

Furthermore, as cyanobacteria are considered poor food for benthic invertebrates (Karlson et al. 2008), an increase in cyanobacteria and changes in the timing and species composition of the spring bloom may influence the quality of the sinking organic matter (Spilling and Lindström 2008) and thus affect the supply of good quality food for the benthic communities.

19.4.2 Benthic Dynamics

The projected decrease in salinity would have a major effect on benthic species distributions. A retreat of marine species from the north towards the south can be expected (Bleich et al. 2011), whereas non-indigenous species with tolerance to low salinity and high temperature are favoured. However, bottom-water oxygen conditions, rather than salinity, would continue to limit the distribution of benthic species. Neumann's (2010) model projection up to year 2100 suggested that the area covered by hypoxic water will decrease, while the modelling studies of Meier et al. (2011, 2012) suggested a worsening of anoxia, at least under a business-as-usual nutrient scenario. If climate change enhances eutrophication, hypoxia would increase in areas of limited water exchange, further limiting macrobenthos. In contrast, if increasing freshwater discharge leads to decreased primary production, as suggested for the Gulf of Bothnia (Wikner and Andersson 2012), food supply for benthos may decrease (Timmerman et al. 2012). Estimates of the final outcome vary, depending on the oceanographic processes involved (Viitasalo 2012).

Owing to the high natural variability of shallow-water environments, littoral organisms are well adapted to variations in temperature, light and wave exposure. Therefore, climate-induced changes in temperature and, for example, pH are suggested to have less effect in structuring littoral communities than communities inhabiting more stable environments (Thomsen and Melzner 2010).

On the other hand, if climate change imposes multiple new stresses on a species, its tolerance limits might be surpassed. For example, shoot densities of eelgrass (*Zostera marina*) decreased when experimentally subjected to simulated summer heat waves (Ehlers et al. 2008), and if salinity also decreased, this marine species would probably disappear from the margins of its distribution area.

Increased temperature may have important indirect effects in the littoral zone by enhancing the growth of micro- and macroepiphytes on *Fucus* (Wahl et al. 2010). Grazing by isopods may also intensify during warm summers, because the levels of defence chemicals in macroalgae decrease in high temperatures (Weinberger et al. 2011).

If surface-water salinity declines due to climate change, the geographical distribution limits of all salinity-dependent species will change accordingly. In the littoral environment, this may affect key marine species like fucoids and eelgrasses. The distribution limit of the eelgrass *Z. marina* approximately follows the surface 5 isohaline and will therefore probably disappear from areas with lower future salinity, such as the Gulf of Finland. Salinity decline is also expected to shift the northern distribution limits of the *Mytilus* community further south (Wikström and Kautsky 2007) and the growth rate also declines with lower salinity (Westerbom et al. 2002). Decline of the mussel beds may affect the associated flora and fauna and may also have an indirect effect on the coastal phytoplankton community, because *Mytilus* can filter a large fraction of the phytoplankton in the water over the mussel beds (Kautsky 1981; Norén et al. 1999).

Ocean acidification is projected to have severe implications for calcifying organisms such as bivalves and corals (Green et al. 2004; Orr et al. 2005). In addition to calcification, key physiological processes, such as growth, metabolism, reproduction and, hence, diversity and functioning of the benthic communities, could be affected (Widdicombe and Spicer 2008; Widdicombe et al. 2009). Reports from the Swedish Environmental Protection Agency (Naturvårdsverket 2008) and Perttilä (2012) suggest that acidification of the Baltic Sea proceeds at least at the same rate as in the ocean. Between 1993 and 2007, the Baltic Sea declined by 0.06–0.44 pH units, with the greatest changes in the Bothnian Sea and southern Baltic Proper, while globally there has been an estimated decrease of 0.1 pH units since 1750 (Naturvårdsverket 2008 and references therein).

If benthic species are adversely affected by acidification in the Baltic Sea, this would also affect the recovery potential of communities after catastrophic events. This is of concern in the Baltic Sea, where the benthic communities are frequently disturbed by eutrophication-induced hypoxia and anoxia. For further information about acidification of the Baltic Sea and potential impacts, see also Chap. 18 and Havenhand (2012).

19.4.3 Sea-Ice Dynamics

A major feature defining wintertime ecology of the Baltic Sea is the recurring formation and melt of the seasonal sea ice (Chap. 8). The sea ice contains a semi-enclosed brine channel system, comprising small pockets and elongated vertical channels, which is the primary habitat for sea-ice biota such as pennate diatoms, dinoflagellates, flagellated protists, heterotrophic flagellates and ciliates, as well as bacteria (Kaartokallio 2004; Granskog et al. 2006, 2010; Kuparinen et al. 2007).

Both the maximum extent and duration of sea ice in the Baltic Sea are projected to decrease due to climate change (Vihma and Haapala 2009; see also Chap. 8) and consequences for Baltic Sea sea-ice ecosystems are anticipated. The main changes expected are linked to direct habitat loss for ice-dwelling or ice dependent organisms; changes in seasonality (e.g. through light-field changes in spring, or increased wintertime mixing); changes in ice-modulated land-ocean interactions, such as spreading of river water plumes; and changes in nutrient deposition onto and incorporation into the growing ice sheet as well as their release into the water column upon ice melt.

Changes in food web structure and function are also possible. Under-ice phytoplankton blooms, formed by dinoflagellates or haptophytes (*Chrysochromulina birgeri*) (Larsen et al. 1995; Spilling 2007), contribute to the onset of the phytoplankton spring bloom after ice break-up (Spilling 2007). Thus, changes in sea-ice dynamics could alter the seasonal phytoplankton succession, and hence, also other ecosystem compartments. In polar environments, sea-ice dynamics influence the quality, quantity and timing of the input of organic matter to the benthos (e.g. Norkko et al. 2007; Renaud et al. 2007). These mechanisms and their potential importance to benthos appear completely unexplored in the Baltic Sea.

Reduced ice cover would alter water mixing conditions during winter. Also, because sea ice allows river water plumes to spread long distances underneath the ice, a lack of sea ice could cause the substances carried by rivers to mix into the water column closer to the river mouths. Furthermore, the seasonal sea-ice cover accumulates atmospheric deposition during winter and acts as a significant source of these substances in spring upon ice melt (Granskog and Kaartokallio 2004; Granskog et al. 2006). Sea-ice organisms can also utilise and accumulate inorganic nutrients from under-ice water and thus influence land-ocean transport pathways (Granskog et al. 2005, 2006; Kaartokallio et al. 2007). Such processes would be altered if the ice cover shortens seasonally or disappears altogether.

While there is evidence that the thickness and duration of ice cover influence phytoplankton communities (Kononen and Niemi 1984; Klais et al. 2013), there are to date very few studies attempting to model the effects of climate change on the Baltic sea-ice ecosystem. A sea-ice biogeochemical model (Tedesco et al. 2010, 2012) projected no change in the timing of the spring phytoplankton bloom with nearly ice-free conditions compared to full ice cover, but did project changes in bloom magnitude and phytoplankton community composition.

19.4.4 Regime Shifts and Cascading Effects

Several regime shifts have been identified in the central Baltic Sea during the past three decades, with the most pronounced during the late 1980s to early 1990s (Möllmann et al. 2008, 2009).

It seems clear that cod stocks are influenced by climate, that cod stocks can influence clupeids (herring and sprat) and that clupeids can influence copepod populations in the different basins of the Baltic Sea. At least in late summer and autumn, predation pressure by clupeids and mysids has been calculated to surpass the zooplankton production (Hansson et al. 1990; Rudstam et al. 1992). The effects of cod stocks diminishing may even cascade through planktivores and zooplankton to coastal fish, because in some areas, the open-sea predation by zooplanktivores may also affect the coastal zooplankton (Ljunggren et al. 2010).

Furthermore, it has been suggested that a low summer biomass of zooplankton may increase the probability of cyanobacterial blooms in the Baltic Sea (Casini et al. 2009) and that return of the cod could improve the status of the Baltic Sea through the cod–sprat–copepod–phytoplankton cascade. The key question is whether grazing by zooplankton can control the different phytoplankton and cyanobacteria groups.

In the Baltic Sea, there are several factors that decouple copepods from phytoplankton (see Kiørboe 1998). First, the pelagic ecosystem of the Baltic Sea is not based on a simple grazing chain: production is largely based on organisms of the microbial loop (Kuosa and Kivi 1989; Kivi et al. 1993; Sandberg et al. 2004) and many dominant copepods are omnivores that switch opportunistically between phytoplankton and protozoan food (Kiørboe et al. 1996). This alleviates both grazing and predation pressure by copepods and reduces the possibility of zooplankton cropping down either group. Second, several studies suggest that in spring, there is a temporal mismatch between phytoplankton and zooplankton and that this mismatch will increase due to climate change (Winder and Schindler 2004; Sommer et al. 2007; Daufresne et al. 2009). Third, in summer a large component of the primary production is produced by toxic filamentous cyanobacteria, which are not preferred food by copepods (e.g. Engström et al. 2000).

Such considerations make it unlikely that the effects of a potential increase in cod stocks would cascade down to phytoplankton or especially cyanobacteria biomass. This also makes sense because top-down forces are considered most important for the higher trophic levels (piscivores and planktivores), whereas for zooplankton and especially the primary producers, bottom-up forces (light and nutrient availability) are relatively more important (McQueen et al. 1989). Interestingly, Stige et al. (2009) showed in the south-western Barents Sea that the climatic forcing of zooplankton was stronger when the density of planktivorous fish was low. Whether this is also the case in the Baltic Sea remains to be studied.

It is notable that ecosystem regime shifts took place in both the Baltic Sea and North Sea at approximately the same time (1982–1988) (Beaugrand 2004). This suggests that regime shifts may be induced and driven by large-scale atmospheric variation. At present, there is no evidence that the regime shifts have been caused by the quasi-linear anthropogenic climate change. Rather, the regime shifts in the North Sea have been linked to natural atmospheric variability, such as the NAO (Beaugrand 2004 and references therein). Therefore, Hänninen et al. (2000) suggested that zooplankton and cod stocks, as well as growth rate of herring can be predicted from climatic variation in the North Atlantic (i.e. the NAO).

If the consequences of anthropogenic climate change alter the basic oceanography and biogeochemistry of the Baltic Sea, as suggested by the modelling results of Neumann (2010), Meier et al. (2011, 2012), system-level changes may be expected that can be associated to climate change per se.

19.4.5 Food Web Efficiency

The projected increase in nutrient discharge has been projected to enhance phytoplankton biomass in the Baltic Sea (e.g. Neumann 2010; Meier et al. 2012). In principle, this should also lead to an enhanced secondary production. The consequences of climate change on the productivity and interactions between the trophic levels are, however, difficult to project. As noted in the Gulf of Bothnia (Müren et al. 2005; Wikner and Andersson 2012) and Kiel Fjord (Hoppe et al. 2008), elevated sea surface temperature and increased freshwater and DOC discharge can make the pelagic community net heterotrophic (Fig. 19.4). Similar responses have been reported from the Hudson River estuary, USA (Howarth et al. 2000), and a subtropical estuary in Brazil (Barrera-Alba et al. 2009). Despite the higher resource (nutrient) availability, an increase in heterotrophy can be hypothesised to result in a lower fish production, because a larger proportion of the energy is consumed in the longer food chain of the microbial loop.

The results of Baltic Sea studies are ambiguous. Dahlgren et al. (2011) reported increased food web efficiency due to temperature increase in a mesocosm experiment. Also, results from mesocosms with water from the Gulf of Bothnia (Lefébure 2012) do not give a clear result. Fish production increased when a pelagic food web including three-spined stickleback (*Gasterosteus aculeatus*) was subject to realistic climate-induced rises in temperature and DOC. This was explained by higher zooplankton production in warmer

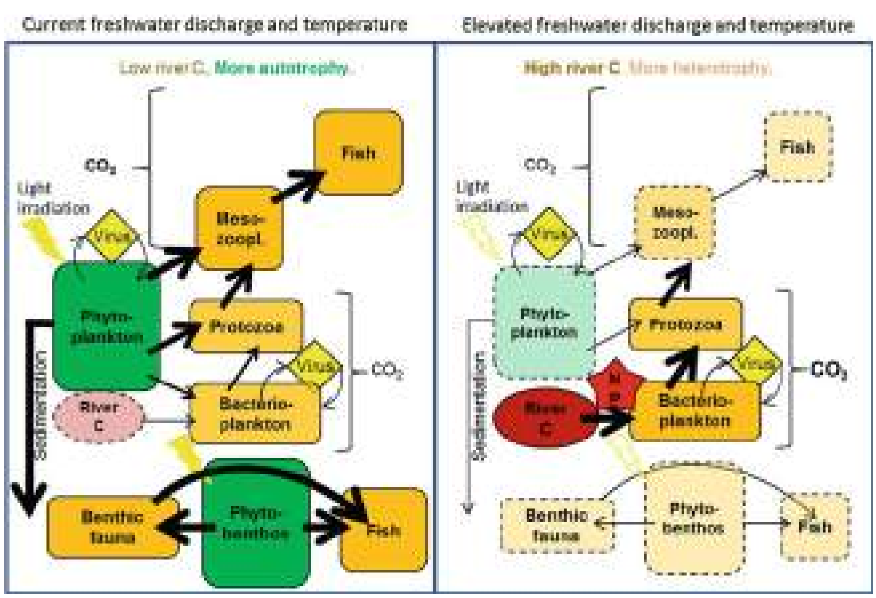

Fig. 19.4 Tentative model of food web effects resulting from increased river discharge and higher temperature. *Boxes* show biomass and *arrows* show flow of material. *Bright colours indicate* higher biomass or light irradiance, and *fat arrows* show higher flow of material. *Red star indicates* competition of inorganic nutrients between bacterioplankton and phytoplankton. Redrawn from Wikner and Andersson (2012)

water, sustained by an increased production of microzooplankton. Although the number of trophic linkages increased with increasing DOC in the water, the positive effects of increased zooplankton production overrode the negative effects of decreasing food web efficiency (Lefébure 2012).

19.4.6 Biodiversity

One of the main consequences of global warming in terrestrial and marine ecosystems is a poleward shift in the distribution limits of both southern and Arctic-boreal species. It has been suggested that more species may move in from lower latitudes than will be lost, resulting in a net increase in the species diversity of northern seas (Hiddink and Coleby 2012). However, in poorly connected areas, such as a narrow strait, colonisation by warm-water organisms may slow or be prevented (Jackson and Sax 2010). For the Baltic Sea, the geographical restriction and strong salinity gradient in the Danish Straits could slow dispersion of warm-water species into the Baltic Sea. It has also been suggested that salinity decline could decrease the diversity of the marine community component. What then is the likely net outcome of climate change on Baltic Sea biodiversity?

Hiddink and Coleby (2012) compared trends in fish species richness outside and inside the Danish Straits (i.e. in the Kattegat and southern Baltic Sea), during a warming period in 2001–2008. They expected the warming to result in a greater increase in fish diversity in the better connected Kattegat than in the more isolated southern Baltic Sea. Unexpectedly, fish species richness increased in both the Kattegat and the Baltic Proper, but the effect was probably more connected with salinity change than increasing temperature.

While Remane's (1934) *artenminimum* concept (minimum β-diversity at a salinity of 5–8) partly explains why benthic and fish diversity decreases with a decrease in salinity, the effects of salinity changes on planktonic diversity are less clear. Telesh et al. (2011a, b) challenged Remane's concept by suggesting that protistan diversity does not decline but peaks at intermediate salinities of the *horohalinicum*. This idea was in turn questioned by Ptacnik et al. (2011), who claimed that Telesh et al. (2011a) derived their diversity patterns largely from coastal bays and lagoons, which are not representative for the Baltic Sea as a whole. The effects of climate-induced salinity shifts on plankton diversity thus remain a matter of debate.

Increasing sea surface temperature and decreasing salinity would make the Baltic Sea a more suitable habitat for species originating from warmer areas with lower salinity. Thus, more non-indigenous species could be expected. The role of non-indigenous species in the Baltic Sea is, however, contentious. Although non-indigenous species are often seen as one of the major threats to marine biodiversity (Costello et al. 2010), the 'indigenous' species in the Baltic Sea are of mixed origin and have all invaded since the last glaciation. While the rate at which non-indigenous species have established has increased due to human vectors, it is partly a natural process of post-glacial succession in the Baltic Sea (Bonsdorff 2006) that may be enhanced by climate change.

In addition to species diversity, genetic diversity may also be affected by climate-induced changes in environmental parameters. At the entrance to the Baltic Sea, a steep cline in intra-population genetic diversity has been documented for several marine taxa (Johannesson and André 2006). For instance, diatoms show lower genetic diversity in the Baltic Proper than in the Skagerrak and Kattegat (Härnström et al. 2011) and the brown alga *F. vesiculosus* as well as most marine red algae mainly reproduce asexually in the northern Baltic Sea. In extreme cases, such as the eelgrass *Z. marina*, whole meadows may consist of a few clones only (Reusch et al. 2005; Ehlers et al. 2008); along the Swedish coast of the Bothnian Sea, much of the population of *F. radicans* is a single female clone (Johannesson et al. 2011).

To sum up, the main climate-related threats to Baltic Sea biodiversity include changes in water temperature, salinity and perhaps acidification. In communities with naturally high biodiversity, the large number of species with different functional roles performs a stabilising role and provides resilience against perturbations, while in species-poor communities, genetic diversity is particularly important as a source of variability in functional traits. Low genetic diversity may therefore make the Baltic Sea species vulnerable to external pressures, including climate change.

19.5 Modelling Climate Change—What Can Be Learnt from Simulating Future Ecosystems?

Potential effects of future climate change on species, food webs and ecosystems in the Baltic Sea have been investigated with a range of model types, from conceptual models (MacKenzie et al. 2007), single species population dynamic models (MacKenzie et al. 2011), multispecies models (Heikinheimo 2011), models of simple (Lindegren et al. 2009, 2010b) or complex food webs (Österblom et al. 2007), to coupled physical-biogeochemical models (Meier et al. 2011, 2012). Most studies focussed on the offshore central Baltic Sea (Österblom et al. 2007; Lindegren et al. 2009; Heikinheimo 2011; MacKenzie et al. 2007, 2011; Meier et al. 2011). One reason for this may be that the available physical-biogeochemical models do not seem to work as well in the Bothnian Sea and Bothnian Bay as in the central Baltic Sea (Eilola et al. 2011).

The diversity of ecological models raises the question as to which models are best suited for analysing ecological effects of climate change. Most earlier modelling studies investigated the direct responses of single (often exploited) marine species to climate change, or to general environmental variation (Kuikka et al. 1999; Rahikainen et al. 2003). However, species responses to climate change depend on the interplay between ecological interactions and climate variation, because climate affects species both directly and via their interactions with other species.

Which food web setting should be used for studying climate effects on a particular species? Climate effects on a given species may depend on interactions between species because of the feedbacks these create (Stenseth et al. 2002). Therefore, food web models that consider feedbacks between species, for example by incorporating both predation effects on prey and the energy gained by the predators (Heikinheimo 2011; MacKenzie et al. 2011), probably give more realistic results than models that do not include such feedbacks (Österblom et al. 2007; Lindegren et al. 2009).

Furthermore, it is not known how increasing the complexity of the food webs would affect the simulated responses to climate change. Comparative studies of ecological responses to climate change from models of different complexity, which have been subject to the same forcing scenarios (‘ensemble modelling’), are needed (see Stock et al. 2011). Studies of the responses of species or food webs in the Baltic Sea to climate change rely on models dealing with populations or even species groups (Österblom et al. 2007; Lindegren et al. 2009; Heikinheimo 2011; MacKenzie et al. 2011; Meier et al. 2011), although some ecological processes (especially in the upper trophic levels of the food web) occur between individuals. Similarly, climate forcing on the model species or food webs has often been based on observed correlations of climate and, for example, recruitment of species (Lindegren et al. 2009; Heikinheimo 2011; MacKenzie et al. 2011) rather than on actual mechanisms.

In addition to ecological–climate interactions, marine food web processes are also linked to geochemical cycles through, for example, nitrogen fixation by cyanobacteria and nutrient release during decomposition of organisms. Simulated ecological responses to climate change may again be different if these processes are accounted for, as is done in coupled physical-biogeochemical models (Meier et al. 2011, 2012).

At present, it is not possible to compare the results of biogeochemical models and food web models, because the biogeochemical models developed for the Baltic Sea do not account for the dynamics of trophic levels above phytoplankton (Meier et al. 2011) and the latter rarely extend below zooplankton (Lindegren et al. 2009). To meet this need, so-called end-to-end models have been developed for other marine systems (Fulton and Smith 2004). These are highly detailed biogeochemical ecosystem models that couple physical, biological, social, economic, and management modules. Although it is unclear how underlying climate variation propagates through the coupled modules, inclusion of such models may prove useful in comparative model studies of climate change impacts on marine food webs. The way forward includes scenario analyses with sets of ecological models of varying complexity as well as extending modelling approaches beyond the central Baltic Sea.

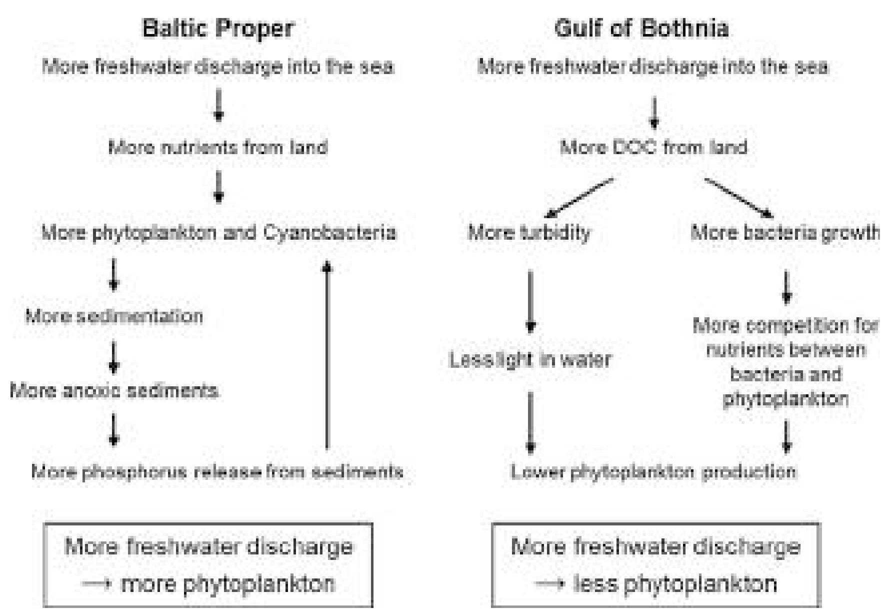

Fig. 19.5 A schematic representation of the ecosystem consequences of increasing freshwater discharge into the Baltic Proper and the Gulf of Bothnia

19.6 Conclusion

Over the past few years, significant steps have been taken in understanding the consequences of climate change on the Baltic Sea ecosystem. Advances have been made in, for example, studies on how climatic variation contributes to regime shifts and cascading trophic effects. Most of the conclusions to date are also based on observations from the open-sea system of the central Baltic Sea, and less is known about regime shifts in coastal and northern systems. Projecting the occurrence and pattern of such system-level shifts in the whole Baltic Sea thus remains a challenge.

Increase in sea surface temperature has been suggested to change seasonal succession and induce dominance shifts in primary producers in spring. Shifts in dominant species may affect the biogeochemistry and functioning of the pelagic ecosystem in the following summer (Sommer and Lengfellner 2008; Spilling and Lindström 2008). As increasing temperature and stratification also favour cyanobacteria, rotifers and small cladocerans, the plankton community is projected to shift towards smaller sized organisms.

Mesocosm studies also suggest that climate change may influence the seasonal succession of phytoplankton and zooplankton, potentially increasing the temporal mismatch between these groups in spring. Such changes may have negative consequences on zooplankton production and thus food conditions of planktivorous fish. A climate-induced decrease in salinity together with poor oxygen conditions in the deep basins would negatively influence the main Baltic Sea piscivore, cod. Several studies have confirmed that this causes cascading effects on clupeids and zooplankton. It is less clear whether the effects cascade from zooplankton to phytoplankton.

Reduced duration and spatial extent of sea ice would cause habitat loss for ice-dwelling organisms, affect the ice-modulated land–ocean interactions and probably induce changes in nutrient dynamics within and under the sea ice. There are, however, no estimates of the effects of declining sea ice on the overall productivity and pelagic-benthic coupling of the Baltic Sea ecosystem.

Modelling efforts suggest that climate change could worsen eutrophication by increasing freshwater discharge and thereby nutrient loads from land. An increase in sea surface temperature would probably also favour cyanobacteria that bind nitrogen from the atmosphere and increase the supply of nitrogen to the nitrogen-limited phytoplankton. Summer primary production and sedimentation would then increase, worsening oxygen conditions and inducing the release of phosphorus from sediments. On the other hand, increasing the supply of freshwater and associated DOC may also reduce phytoplankton productivity, at least in the Gulf of Bothnia (Fig. 19.5). Thus, it is clear that the effects of climate change on the productivity of the marine ecosystem vary from basin to basin.

Some of the most profound effects of the projected salinity decline involve losses in functional diversity that would accompany the loss of marine elements in the fauna. Also, the potential increase in primary production and sedimentation of organic matter in the northern Baltic Proper, as well the climate-driven decrease in trophic efficiency, as suggested for the Gulf of Bothnia, are potentially important factors for benthic communities.

Human-induced pressures, such as overfishing and eutrophication, may erode the resilience of the Baltic Sea ecosystem, thereby making it more vulnerable to climatic variations. The Baltic Sea communities, that are poor in both species and genetic diversity, may therefore be particularly vulnerable to external forcing factors caused by the climate change.

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20 Environmental Impacts—Coastal Erosion and Coastline Changes

Tomasz A. Łabuz

Abstract
Climate change is having an undeniable influence on coastal areas. This chapter describes the growing threat of climate change on the Baltic Sea coastline, with an emphasis on field research focused on storm surges and coastal retreat. The main climatic factors driving change in the Baltic Sea coastal zone are wind, waves, storm surges, ice jams and flooding. The cumulative effects of these drivers are also important. For example, a costly coastal protection scheme in one area may result in coastal erosion in another. Natural and man-made coastal features are experiencing unprecedented change; important natural habitats, coastal settlements and local economies are all being affected. The extent of storm surge impacts depends on the exposure of a shoreline to a surge event. The submergent and soft coastal relief of the southern Baltic Sea area is under most threat; the rate of retreat depending on the frequency and strength of the storm surges. The rate of coastal retreat has also increased in recent years due to sea-level rise and loss of beaches.

Keywords
Coast erosion • Threats for coast • Coast type response for climate change

20.1 Introduction

The coastal zone is one of the most dynamic environments in the world, because it is where different geospheres interact. Taking place over a range of timescales, these interactions cause dynamic coastal rebuilding, referred to coastal morphodynamics. The factors responsible for change in the coastal zone may be grouped into geological and geomorphological, hydrodynamic, biological, climatic and anthropogenic factors. The main geological factors—sediment type, arrangement and resistance of sediment structures, and isostasy—are the basis of the morphological processes and the development of coastal relief. Geomorphological processes are influenced by many external forces, including climatic factors such as precipitation rain and wind. They are responsible for the development of typical relief forms and sediment supply. Biological factors mainly concern the influence of plants and are responsible for the development of particular coastal types. Anthropogenic factors refer to many human activities taking place within the coastal zone: settlements, industrial development, agriculture, deforestation and coastal protection. The coastal system is a dynamic complex of sensitive factors and typically responds in a nonlinear morphological manner (Dronkers 2005). The wide range of natural and human influences on the Baltic Sea coastline makes it difficult to identify the specific effects of climate change. Exposure to wind is a key factor for coastal development. Climatic conditions greatly influence the inflow of material to the coastal zone, by determining the amount and origin of terrigenic material supplied to the coastal zone as well as the amount of biogenic material produced and supplied to the sediments.

The combination of sea-level rise, land subsidence and isostatic rebound creates both emergent and submergent coasts. Emergent coasts are identifiable by accumulative

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_20

landforms or ancient coastlines recorded in the relief or sediment structures on land. Submergent coasts occur where rising sea level and subsidence cause coastal retreat, and where water is entering and eroding land features. Crest movements show the 'sinking' of the southern Baltic Sea coast and uplift in Scandinavia. For instance, glacial isostatic adjustment (GIA) has caused the coastal zone in Estonia to emerge throughout the Holocene (last 10,000 years), with current uplift rates of 1.0–2.8 mm year^{-1} and with maximum uplift located on the north-western coast (Vallner et al. 1988; Ekman and Mäkinen 1996). Transgression is expected with the sinking of the coast (Harff and Meyer 2007) (for more information on land movements due to GIA see Chap. 9).

The main coastal features in the Baltic Sea region are sand or gravel spits with diversified dunes, cliffs cut in a variety of sediments and low-lying areas such as lagoons, wetlands and salt marshes (Fig. 20.1). Accumulative features such as sand spits or beaches are formed by currents and waves. Dunes developed on sandy spits are higher origin forms accumulated in the past or now. They are developed as aeolian forms shaped by wind. Cliffs may be cut into hard rock or into soft clays, tills and sands and are usually higher-order relief forms shaped by erosion processes such as waves or landslides.

Different types of Baltic Sea coast may be distinguished based on the strength of the factors affecting the various geological structures of the coastal zone. Most Baltic Sea coastlines and coast types were formed during the last deglaciation and the subsequent Holocene sea transgression. Unconsolidated glacial sediments were shaped into moraine hills separated by low-lying fluvioglacial valleys. High, soft moraine hills attacked by waves were shaped into relatively steep cliffs. Sea transgression flooded low-lying areas, creating shallow bays, lagoons and swampy areas. In such areas, sediment accumulation by long-shore currents often created emerging barriers. The southern Baltic Sea coast is mainly composed of moraine cliffs (made by glacial tills, clays and fluvioglacial sands) and sandy or gravel coasts developed as spits. Abundant sand sediment has led to the creation of various sandspits, with the longest occurring on the southern Baltic Sea coast, such as the Vistula and Curonian sandspits. Lagoons, shallow sea bays and coastal lakes were formed on low-lying coast, where organic accumulation and sedimentation dominated. Spits such as islands or barriers develop where sediment, moving alongshore, is deposited at the mouth of estuaries and other places with an abrupt change in the direction of the coast (Trenhaile 1997). Several sandy barrier islands occur on the German Baltic Sea coast (e.g. Hiddensee), others built by gravel or pebbles, in Estonia.

Owing to long-shore transport and the origin of land features, the Polish, Lithuanian and Latvian coastlines are effectively aligned, with sandy barriers separating valleys, deltas, lakes and lagoons from the open sea. The German and Danish soft coasts are more circuitous, due to numerous bays, peninsulas and islands.

The northern part of the Baltic Sea coast is mainly formed by isostatic uplift due to the loss of glacial cover in Scandinavia. In some places, accumulative forms occur in bays or river deltas. Finland and Sweden have emergent coasts where abundant skerries form archipelagos. Some U-shaped fjords were formed by glaciers in rocky, mainly igneous beds (Fig. 20.2b) with the presence of some sandy, accumulative barriers or beaches. Rocky limestone or chalk beds, found in Estonia and on the Baltic Sea islands, have formed cliff coasts (Fig. 20.2e, f).

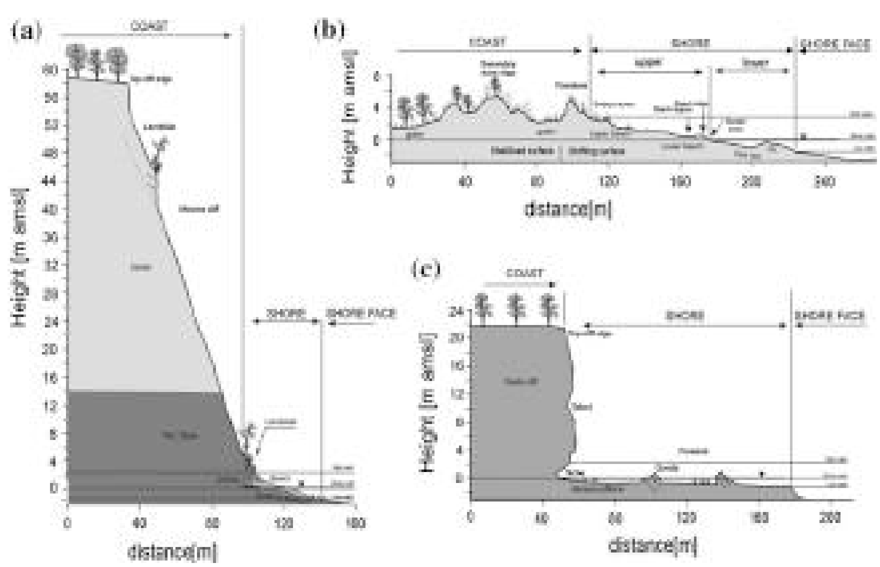

Fig. 20.1 Features and zones of the main Baltic Sea coast types. **a** Soft cliff coast, **b** Sandy coast with foredunes, **c** Rocky coast

Fig. 20.2 Coastal types of the Baltic Sea. **a** Skerries in the Gulf of Finland, Finland; **b** Igneous rock, low coast shaped by glaciers, Helsinki, Finland; **c** Low coast of glacier boulders, northern Estonia; **d** Wetlands in southern Sweden near Malmö; **e** Rocky, chalk cliff covered by moraine sediments with an abrasive platform instead of beach, Rügen Island, Germany; **f** High-eroded cliff–klint coast in Estonia; **g** High soft moraine cliff coast of tills and fluvioglacial sands, Wolin Island, Poland; **h** Typical cliff coast of tills and clays of moraine deposits, landslides on the beach caused by ongoing storm surge, Sambian Peninsula, Kaliningrad region, Russia; **i** Low eroded coast without dune ridges in Heiligendamm, Germany; **j** Low coast of end moraine deposits covered by sands, Latvia; **k** Retreating dune coast entered and covered peatbog, Kolobrzeg, middle Polish coast; **l** Sandspit dunes: shifting inland with typical accumulative foredunes, Lebsko Lake Sandbar, Poland (*Photos* T.A. Łabuz)

20.2 Characteristics of Coastal Types Around the Baltic Sea

The Scandinavian coast is mainly covered by hard rock, such as igneous, sandstone or limestone forms. It has been shaped by glaciers that have formed U-shaped fjords or flat coastal forelands. In the Bothnia Gulf, there is a skerry-type coast, which due to uplift is formed by many small islands. Skerries are commonly formed by rocky blocks that emerge from the water, separated by ice-shaped channels. The bedrock geology of the south-west archipelago consists of fractured and faulted Precambrian granites and schists of the Fennoscandian Shield that comprise the peneplain surface that slopes gently south-west (Schwartz et al. 1989). The following sections describe the coastal types by country. See Figs. 20.1 and 20.2 for the main the coastal types occurring in the Baltic Sea region.

20.2.1 Finland

Small patches of dune landscape are widespread on the Finnish coast, formed mainly from skerries (Fig. 20.2a) and post-glacial beaches on rapidly rising coastlines (Hellemaa and Doody 1991). The uplift rate is up to 9 mm $year^{-1}$, particularly in the Gulf of Bothnia. The dune landscape consists of parallel ridges separated by low, wet depressions arranged

Fig. 20.2 (continued)

parallel to the coastline. The ridges are generally no higher than 8 m, but where uplift is slower, a few reach 20 m. The south-western coast of Finland near the Åland Islands has many skerries, forming a huge archipelago. In total, there are some 73,000 islands along the coast of Finland.

20.2.2 Sweden

The Swedish coastline is over 13,500 km long. Due to continuous uplift in the northern section, only the southern part of the Skåne coast is at risk of erosion (Fig. 20.2d). The present rate of land uplift by isostatic rebound varies from zero in southern Skåne to a maximum of 0.9 mm year^{-1} along the coast of the Bothnian Bay (Hanson 2002, see also Chap. 9 and references therein). Low coasts dominate in Sweden and active cliffs cut into tills and fluvioglacial sediments are found in the south of the country. The rest of the coastline comprises low or average height cliffs made of hard rock, partly covered by glacier deposits such as till, sand or gravel. The northern part of the Bothnian coast is mainly low lying and comprises igneous rocks and gravel. In Sweden, coastal dunes mainly occur on the southern coast in Skåne and Halland (Malmö-Ystad) and in spots along the Gulf of Bothnia on low coastal areas and small barrier islands or spits among the cliffs. The dunes reach heights of 10–15 m as ridges parallel to the coastline. The counties most affected by coastal erosion are Skåne, Blekinge and Halland (European Commission 2004).

20.2.3 Denmark

The Danish coastline of the Baltic Sea varies in shape and origin. In some places, the coast is formed by cliffs (‘klints’), as in the southern part of the island of Zeeland. One of the famous Cretaceous–Tertiary klints occurs south of Copenhagen—a 14-km-long rocky cliff formation, mainly of chalk and limestone with Quaternary deposits on top. In other places, Quaternary deposits form the cliff with klint structures below ground (Prior 1977). Salt marshes, also called meadows, sink at a rate of 1–2 mm year^{-1}. Danish coastal systems are subject to erosion at rates of up to 3–5 m year^{-1} (Clemmensen et al. 2011). The islands of Gotland (Sweden) and Bornholm (Denmark) are mainly formed from rock with small areas of coastal sand dunes.

20.2.4 Germany

Two-thirds of the German Baltic Sea coastline is being eroded. The coastline is divided between the two German states of Schleswig-Holstein in the west and Mecklenburg-Vorpommern in the east. The Schleswig-Holstein coastline is 637 km long with 148-km-long soft cliffs and is generally in a retreat state due to sea-level rise (Hofstede 2008). The Mecklenburg-Vorpommern coastline is almost 1945 km long in total with an external coast of 376 km; of the latter, steep cliffs cover 128 km and flat sections 248 km. Approximately 70 % of this coast is in recession (Gurwell 2008). More than half of the coastline is 'bodden' coast—shallow bays and inlets cut-off from the sea by islands, peninsulas and spits (Sterr 2008). This coast was shaped during and after the last glaciation, with prevailing glacial and postglacial features and sediments, with low and soft cliffs. Natural coastal dunes are rare on the German Baltic Sea coast, and those that do occur are mostly found on the Darss Peninsula, Usedom Island and Rügen Island. The Darss Peninsula is a sandy spit covered by parallel ridges separated by wet, low depressions. Among these are sandy barriers that have developed as separate islands, like Hiddensee which is built from marine and aeolian deposits. These structures of marine origin have developed by long-shore transport of eroded soft moraine cliffs. Small dunes have developed on low parts of Rügen Island, mainly on the beaches, and are derived from fluvioglacial deposits and chalk rocks. The chalk cliffs are unstable (Wahle and Obst 2009) and mainly covered by typical moraine deposits (Fig. 20.2e). The south of Rügen Island is formed by low, soft cliffs made from tills and clays, covered by sands. Some cliffs on Usedom Island reach heights of 50 m. In other places, low coastal cliffs are formed from bottom moraine deposits. The remainder of the German Baltic Sea coast is formed by moraine cliffs with an average height of 5–15 m and fronted by narrow beaches. Among them are lowland areas with wetlands—salt marshes or coastal meadows. This type of coast has vegetation typical of swampy saline areas. In coastal areas with low hydrodynamic stress (common in bodden areas), peatlands may develop and reed beds may replace the beach (Lampe and Janke 2002). In places where the coast is formed of low-lying valleys, as between Kiel and Heiligendamm on the western German Baltic Sea coast, many coastal sections are protected by artificial dykes (Fig. 20.2i).

20.2.5 Poland

The Polish coast is aligned and formed of loose sediments with beaches up to 35 m wide on average. Extensive dune fields formed after the glacial recession can be found on the Polish coastal plains, where sandy dunes cover over 80 % of the 500-km-long coastline (Łabuz 2005). The dune fields are mainly situated in the direct vicinity of the coast but may reach 4 km in width and incorporate inland dunes, formed by sands transported by wind from the beach or fluvioglacial deposits. These are usually aeolian sand dunes forming transgressive ridges. The highest dune is 56 m, with the highest wandering dune 42 m; both on the Łebska Sandspit (Fig. 20.2l). The coastal dune fields usually have a cliffed seaward slope with beaches in front (Fig. 20.2k). In several places, the depressions separating individual dunes on the dune coast are over 5 m deep. In such places, the land is threatened by flooding due to heavy precipitation, ice melt or storm surges. To protect low-lying areas, coastal dunes in Poland need to be 4.6–4.8 m in height and beaches need to be 2–2.5 m in height after nourishment (Boniecka and Zawadzka 2006). One-third of the sandy coast is formed by typical parallel coastal ridges, most of them eroded. Elsewhere, accumulation due to long-shore currents transporting sediments from eroded cliffs results in the development of beaches 60 m wide or more. Accumulative tendencies are seen on the Swina Gate Sandspit, at the end of the Hel Peninsula, at the mouth of the Vistula River or on the Łebsko Lake Sandspit (Łabuz 2005). Foredunes of less than 10 m in height are still developing in these places. Soft Quaternary moraine cliffs cover 65 km of the Polish coast, among them high formations (up to 95 m) of glacial deposits (tills and clays) covered by fluvioglacial sand (Fig. 20.2g) and lower formations of till covered by limnic and organic deposits (Subotowicz 1982). The longest sections of moraine cliff occur along the western part of the coast, and locally on its eastern part. Reed habitats are found on the banks of lagoons and lakes and also along Puck Bay. The Vistula Sandspit is bordered by lowland depressions and the mouth of the largest southern Baltic river, creating a typical sandy deltaic coast in Gdańsk Bay. Low-lying areas of 3 m above mean sea level (amsl) or less are vulnerable to sea-level rise and are at risk of flooding or ground water soaking (Rotnicki and Borówka 1990). Washover fans marking storm surges develop up to 3.5 m amsl (Łabuz 2012).

20.2.6 The Kaliningrad Region (Russian Federation)

The eastern Baltic Sea coast comprises a range of soft, low moraine cliffs and low sandy coasts with sand dune fields. The only significant sand barriers are the Vistula and

Curonian sandspits, the longest on the Baltic Sea coast, which have largely eroded foredunes and partly stabilised large sandy dunes (Boldyrev et al. 2010). The Vistula Sandspit is shared by Poland and the Kaliningrad region of the Russian Federation. The Curonian Sandspit stretches from the high, unstable moraine coastal cliffs of the Sambian Peninsula (Fig. 20.2h) in the Russian Federation to the channel of the Curonian Lagoon near Klaipeda in Lithuania. This part of the Russian Baltic Sea coast is mostly being eroded, with the rate of erosion accelerating in recent years (Chubarenko et al. 2009; Boldyrev et al. 2010).

20.2.7 Lithuania

The Lithuanian coastline is 160 km long and has large shifting sand dunes stretching for about 70 km, mainly on the Curonian Sandspit, bordering low-lying land along the Curonian Lagoon. The sea front is a sandy beach up to 80 m wide. The proper coastline of 90 km in length (Povilanskas 2002b) is mainly formed of coastal dune ridges up to 10 m high. Some dune sections are being eroded by storms. Between 1990 and 2003, accumulative coasts decreased from 40 to 11 %, while erosive stretches increased to 27 % (Milerienė et al. 2008; Jarmalavičius et al. 2012). The highest annual retreat exceeds 2.2 m.

20.2.8 Latvia

The Latvian Baltic Sea coast is about 497 km long. The coast on the Baltic Proper is 183 km long and mainly exposed to the west, with the remainder in the Irbe Strait and Gulf of Riga (Eberhards 1998, 2003). The Baltic Proper coast and the western and southern coasts of the Gulf of Riga comprise soft cliff face formed by Quaternary deposits (Eberhards 2003). Hard, rocky cliffs of 4–6 m in height formed of Devonian sandstone and siltstone and covered by Quaternary deposits are found along the Vidzeme coast of the Gulf of Riga (Eberhards 2003; Koltsova and Belakova 2009). Dune coasts (Fig. 20.2j) can also be found, mainly in the Gulf of Riga. Almost 69 % of the Latvian coast is highly vulnerable to erosion (Eberhards 1998, 2003).

20.2.9 Estonia

Land uplift has created a series of different landforms along the 3800 km of Estonia's coast. Most of the coastal area of Estonia emerged from the regressive sea and had a deficiency of sediment drift, and coastal formations are therefore low (Ratas et al. 2011). Coastal dune landscapes covering about 200 km^2 are found behind sandy beaches along 340 km of the coast (Ratas et al. 2008). Dunes mainly develop where river mouths or deltas provide sand and are typically low ridges 2–8 m high and 30 m wide. Due to land uplift (2–2.5 mm year^{-1}), coastal dunes are found inland of the present coastline (Vallner et al. 1988). Most of the Estonian coast is low-lying and covered by moraine deposits with developing coastal meadows, gravel shores and a silty and low limestone coast (Fig. 20.2c). The largest Estonian islands, Hiiumaa and Saaremaa, are 1–10 m amsl with narrow gravel beaches and spits, with visible erosion and movement of low spits (Suursaar et al. 2008). The northern coast located on the Gulf of Finland comprises numerous peninsulas and bays, shaped in rocks with many small beaches of pebbles and sand. The north-eastern coast, where the coastline is mainly straight with some forelands building shore and shore face, is mostly sandstone and limestone cliffs of up to 56 m—klint (Fig. 20.2f). The Estonian coast is young and develops quickly due to uplift. Prograding beaches are stable but lacking sediment and are vulnerable to storm surges.

20.2.10 Russian Federation

The Russian coast to the south of the Gulf of Finland is mainly formed of low peninsulas built of moraine deposits with boulders in some places, and covered by unconsolidated sand. Beaches are narrow, consisting mainly of gravel and boulders. The northern coast is characterised by a very stable coastline mainly formed of bedrock formations. The southern sedimentary coast developed rapidly under wave action. Significant amounts of ice form every winter in the Gulf of Finland. Neva Bay, 21 km long, is a naturally low area of swampy meadows. Due to risk of flooding, a dam was built across Neva Bay to protect the low coast around St. Petersburg (Orviku et al. 2003). The northern coast of the gulf is higher with some sandy deposits on moraine and older bedrock.

20.3 Erosion of Specific Coastal Types in Relation to Climate Change

This section describes the erosion of different coastal types in the Baltic Sea region: forelands, beaches, coastal meadows, wetlands, sand spits, dunes, moraine soft cliffs and rocky coasts. Figure 20.3 shows the geographical distribution of the main coastal types.

20.3.1 Forelands and Beaches

Forelands (or promontories) are low-lying areas between the waterline and higher elevations further inland that are extended towards the sea. They may be rocky, till, gravel or sandy platforms that are formed by storm surges cutting into land deposits. Forelands are mainly present on rocky coast and accompanied by bays indented in the land. They can be found in the south-western and northern parts of Baltic Sea basin. The beach is an accumulative landform, formed by silt, sand or gravel deposited in the coastal zone. Beach shape and morphological parameters such as width or height provide information about coastal processes and development.

The low-lying part of the Baltic Sea coast is vulnerable to sea-level rise (see Chaps. 9 and 14). For instance, the spectacular sandy beaches of Estonia between Narva-Joesuu and Merikbila have already suffered damage from strong storms in recent decades. One likely cause is the decline in sand supply from the Narva River since it was dammed in 1956 and a hydropower plant built (Kont et al. 2008). Estonian beaches are already shrinking due to an increase in storm frequency (Orviku et al. 2003, see also Chap. 4). Spits still develop occasionally, but erosion now prevails due to major storms such as those in 2005 and 2007 (Tõnisson et al. 2008). Gravel and pebble beaches on Saarema Island were strongly affected by the January 2005 storm Gudrun and the storm resulted in beach deposits being moved inland by 15–30 m (Tõnisson et al. 2008). The shoreface was covered in boulders and accumulation was observed in shallow lagoons.

Despite isostatic uplift, the erosion of sandy beaches has prevailed throughout Estonia in recent decades: in some places the sea has even advanced inland, with beach erosion reaching 1.5 m year^{-1} (Orviku et al. 2003). With no evidence of rising sea level, the observed beach erosion must be largely due to the recent increase in storm frequency in the eastern Baltic Sea and the decline in protective sea ice. Westerly and south-westerly storm winds in autumn and winter can raise sea level up to 2.6 m above the summer level (Kont et al. 2008).

In Latvia, cuts in coastal Quaternary deposits and sandstone also mark an increase in erosion processes (Eberhards et al. 2006). Some Lithuanian beaches have recently vanished due to westerly storm surges (Povilanskas 2002a; Dailidiene et al. 2006). Near Palanga and Olando Cape, coastline sediment loss reaches maximum values of almost 180 m^3 m^{-1} (Žilinskas 2005). The height of Lithuanian beaches is 2.6–3.2 m (Jarmalavičius et al. 2012). Between 1993 and 2008, there was no visible beach recession but the coastline retreated by 10 m on average (Jarmalavičius et al. 2012). Coastal protection measures are currently used to

Fig. 20.3 Coastal types in the Baltic Sea region. *A* Soft moraine cliffs; *B* Sandy barriers and sandy dunes; *C* Rocky cliffs; *D* Skerries; *E* Low coast, meadows, organic/wetlands

prevent land erosion near Palanga and other coastal resorts (Žilinskas et al. 2006).

Beaches in the Kaliningrad region have become narrower: decreasing from 25–40 m in 1984 to 5–15 m in 2000 (Gurova 2004). Near Sokolniki, there was an accumulative coast with a 20-m-wide beach and dunes of up to 5 m high. Now there is serious erosion of the coast and a vanishing beach (Shishkina 2010). Average beach width in the Russian part of the Curonian Sandspit shrank by 10–20 m between 1984 and 2003 and height lowered by 0.2–0.5 m (Gurova 2004).

On the Polish coast, beaches lower than 3.5 m amsl are washed away by heavy storm surges each year (Łabuz 2009a, b). The storm surge of 14 January 2012 reached 4 m amsl (Łabuz 2014). Due to increasing storm frequency in recent decades (see Chap. 4), longer sections of beaches require replenishment. Coastal towns which experienced annual erosion of 0.3–0.7 m now have a nourished beach (Boniecka and Zawadzka 2006). Storms with water levels higher than 0.7 m amsl erode the entire beach as well as the top of coastal dunes or cliffs. For storms such as those in 2001, 2004 and 2006, erosion can amount to 2–5 m (Musielak et al. 2007). During a storm surge with a water level 1.7 m amsl, the beaches of the Hel Peninsula were completely flooded. Almost every year, two-thirds of the Hel Peninsula beaches need replenishing. Between 2002 and 2012, beaches on the Polish part of the Wiślana Sandspit shrank in width by 12 m in total (Łabuz 2012), and in the preceding 19-year period (1984–2001), beaches on the Russian part of the Wiślana Sandspit shrank by up to 18 m (Kobelyanskaya et al. 2009). Many sandy beaches along the Gulf of Finland have recently been severely damaged by frequent storm surges, despite extensive protective measures (Ratas et al. 2011).

At Flakket on the Danish island of Anholt in the Kattegat, the low foreland formed as a flat platform covered by gravel and sand had been enlarging up until 1934 due to uplift. Since then it has eroded at rates of up to 6.5 m year^{-1}. Beach-ridge sediments, formerly up to 500 m wide, have mostly disappeared on the north-west side of Flakket. Between 2006 and 2010, erosion rates of up to 10 m year^{-1} were recorded (Clemmensen et al. 2011). Water levels of 1.02 m amsl reach parts of the foreland once a year; while levels of 1.63 m amsl during storm surges (return period 20 years) affect the entire structure. Wave erosion has led to rapid coastal retreat and in the area immediately east of the local harbour a short stretch of shoreline is now only 25 m from the island's main road. Erosion rates elsewhere along the dune shores in the Kattegat are up to 2 m year^{-1} (Clemmensen et al. 2011).

20.3.2 Coastal Meadows and Wetlands

Coastal meadows and wetlands (marshes and mires) are mainly found in Estonia and to a lesser extent in Sweden, Latvia and Finland. An increase in sea level should theoretically cause a change in the composition of the vegetation of Baltic Sea coastal wetlands (Vestergaard 1997) and gradually displace them in an inland direction. No data are available on accretion processes in low-tidal meadows like those found along the Baltic Sea coast. Loss of meadow area will depend on the relation between the rate of sea-level rise and the rate of accretion. If accretion keeps pace with sea-level rise, then no loss will occur; if accretion is less than sea-level rise, then the loss will start from the seaward edge of the meadow. A third consequence of sea-level rise may be increased wave erosion along the seaward edge of the meadow that will tend to narrow the lower meadow (Vestergaard 1991).

Sea-level rise could strongly affect Estonia, due to its long coastline and extensive low-lying coastal areas, although the land uplift rate is higher here than in the south-western part of the Baltic Sea basin. Estonian coastal wetlands and mires are already shrinking due to climate change (Kont et al. 2007). Flat- and low-lying coastal zones that experience isostatic and tectonic uplift should be stable, despite the activation of coastal processes, presumably due to global climate change, that has been observed in Estonia since the mid-1970s (Orviku et al. 2003). A hypothetical 1 m rise in global sea level would roughly reinstate the coastline position of the 1700 s, forcing local plant communities to migrate inland (Kont et al. 2008) and would submerge many small islands. The Estonian coast shows frequent sea-level fluctuations due to changes in the precipitation/evaporation balance, river discharge and storms, but no evidence of a long-term rise in sea level (Kont et al. 2008). Westerly and south-westerly storms can cause short-term surges and extensive flooding on the Estonian coast (with maximum storm surges of 2.53 m above the Kronstadt zero), whereas the highest water-level rise in the open sea rarely exceeds 2.0 m (Kont et al. 2008, see also Chap. 9). On erodible shores, the fine-grained fractions are entirely washed out during a surge. However, the shingle ridges and boulder fields that remain are extremely resistant to further erosion. Virtsu, an important port connecting the Estonian mainland with the islands of Saaremaa and Muhu, is expected to undergo significant change. The peninsula upon which Virtsu is situated could eventually be divided into many islets. A rise in sea level of 72 cm (in the north) to 80 cm (in the south) would inundate more than 76 km^2 of Estonian territory, including all reed beds and most coastal meadows. Several plant communities and biotopes of rare species—including unique orchids—would disappear (Kont et al. 2008). In the absence of coastal protection measures, an inland migration of coastal landforms and plants would be inevitable. The areas most vulnerable to sea-level rise in the coastal wetlands near Pärnu are flooded during storms (Kont et al. 2007). The Estonian lowland may be affected by flooding up to 300 m inland. Such areas may need special protection, as sea-level rise would have socio-economic impacts, particularly on recreational areas. Water impact on the sandy coastal zone has long been observed in many sections of the eastern Gulf of Finland (Orviku et al. 2003). The high rate of coastal erosion results from the specific features of this water area, such as the morphology, shallowness and local wave field (Ryabchuk et al. 2009).

Rising sea level causes a landward shift in the meadows and a growing thickness of the organic material, whereas a fall in sea level leads to surface drying and mineralisation and a lowering of its surface; described for the German coast by Lampe and Janke (2002). Moreover, black peat layers have developed due to peat oxidation caused by a fall in either groundwater or sea level. Coastal meadows located no higher than 0.5 m amsl on the German coast seem to be in greater danger than those in Estonia.

The eastern part of the Gulf of Finland is especially prone to coastal flooding during westerly storms. The highest recorded flood in St. Petersburg occurred on 7 November 1824, when the water level reached 4.21 m amsl (Kurennoy and Ryabchuk 2011). During the storm Gudrun (January 2005, Tõnisson et al. 2006), an early forecast for the maximum surge in St. Petersburg was 3.7 m. Even floods not exceeding 1.9–2.2 m caused extensive coastal erosion and sediment resuspension in many coastal areas near St. Petersburg when accompanied by strong wind waves. Such events are particularly damaging to the coastal zone as they hinder the recovery of beaches, destroy buildings along the coast and lead to unrecoverable dune erosion, as seen in the eastern Gulf of Finland (Ryabchuk et al. 2009). A significant increase in annual and winter storm frequency in the latter half of the twentieth century has been observed in the Gulf of Finland (Orviku et al. 2003, see also Chap. 4). Low-lying coastal areas in Finland (e.g. Helsinki) are also at risk of flooding during heavy storm surges like the one in 2005.

20.3.3 Sandspits and Coastal Dunes

Sandspits are coastal deposition features commonly found from the south-western to the south-eastern parts of the Baltic Sea coastline. The predominance of westerly winds is shaping coastal dunes that develop on sandy shores from sand delivered from wide beaches. The last extreme event caused by westerly winds was in December 2013, when hurricane Xavier rebuilt southern Baltic Sea coastal beaches and dunes. These east-blowing winds are also responsible for developing storm surges causing coast erosion on the Curonian Spit (Lithuania and Russia); near Zelenogradsk the erosion rate is 0.69–0.79 m year^{-1} (Fedorova et al. 2010). Heavy erosion was recorded after hurricane Anatoly on 4–5 December 1999 on the west-facing sandy coasts in Lithuania, Latvia, Russia and Estonia.

Lithodynamical processes along the sandy Lithuanian coast have changed over the past couple of decades due to ever-stronger storm surges (Visakavičius et al. 2010). The storms of 2–6 October 2006 washed away the foredune north of Palanga Pier. In Poland, this storm caused heavy coastal erosion on 1 October, with a dune retreat of 5–8 m (Łabuz and Kowalewska-Kalkowska 2010). Estonian coastal dunes erode due to coastal retreat, as shown by several studies (Ratas et al. 2008). The storm Gudrun (9 January 2005) caused a surge of 2.75 m in Pärnu, Estonia. Erosion on Saaremaa Island, Pärnu Bay and the Gulf of Finland was significant after this event (Tõnisson et al. 2006). Near Tallinn, the beach lowered and dune retreat was up to 5 m. The erosion due to this storm was ten times greater than that during the previous 20 years. High surges on the Russian coast of the Gulf of Finland in autumn 2006 and winter 2007, with water levels 2 m amsl, totally changed the dunes in cut-off cliffs of up to 2 m high (Ryabchuk et al. 2011).

The sandy Latvian coastal zone has changed considerably in recent decades, presumably due to global warming, anthropogenic factors and increasing storm damage (Eberhards and Saltupe 1995). The sandy coastline of the Gulf of Riga was particularly strongly affected by erosion and recession during extreme storms in 1992 and 2001, with dune erosion of 20–30 m in a single storm event (Povilanskas 2002a). Detailed studies show that erosion is more frequent than accumulation along the Latvian coast, with loss rates of 3–14 m^3/m per year (Soomere et al. 2011). For the Baltic Proper, the mean rate of shore retreat is 3–8 m year^{-1}, with a maximum of 10–18 m year^{-1} or more (Eberhards 2003). Coastal dune retreat in Latvia following Hurricane Erwin in January 2005 was 1.5–7 m on average, with a maximum of 15 m north of Liepaja (Eberhards et al. 2006). The only and longest section with accumulative tendencies is on the western coast of the Kurzeme Peninsula.

The Polish coast is mainly formed of soft sandy sediments, with an average rate of retreat of 0.5–1.5 m year^{-1} (Zawadzka-Kahlau 1999). Dunes of up to 4 m in height are washed out or reduced by storm surges along the Polish coast. Behind them, storm washover fans in interdune depressions are a strong indicator of storm erosion (Łabuz 2009a). In 1995–2009, two to three storms per season completely washed away the low, new foredunes in places where the beach was less than 3 m amsl (Łabuz 2009a). In 2001–2004, storms caused greater dune retreat than the long-term annual mean for the Polish Baltic Sea coast (Musielak et al. 2004). The storm surge of 23 November 2004 caused severe erosion of up to 7 m in the western part of the Polish dune coast (Łabuz and Kowalewska-Kalkowska 2011). Dune erosion during the storm surge in October 2009 exceeded 5–9 m (Łabuz and Kowalewska-Kalkowska 2010). In recent years, foredunes have been heavily eroded along the western Polish coast (Dudzińska-Nowak and

Furmańczyk 2009; Łabuz 2009b; Furmańczyk et al. 2011). On the central and eastern coast, the maximum retreat of sand dunes was 1–2 m year^{-1} (Zawadzka-Kahlau 1999). Today erosion after each storm surge reaches 3–6 m; extreme erosion after storms in 2009 and 2012 reached 6 m year^{-1}. Low coastal dunes were completely washed away and in some places need replacing by artificial dikes or concrete walls. The situation is similar along more than two-thirds of the Hel Peninsula, where beach nourishment, groynes and other measures protect the narrow land against flooding. Coastal dunes on the Wiślana Sandspit along the Polish and Russian section were eroded in autumn and winter of 2006 and 2007 (Kobelyanskaya et al. 2009). Other studies show the Wiślana and Curonian spits to be continuously eroded, with a recent increase near the Baltyjsk Strait (Volkova et al. 2004; Chubarenko et al. 2009). The heavy storm surge of 2012 caused dune retreat of up to 5 m on the Wiślana Sandspit near the Polish–Russian border, and substantial erosion in the eastern and middle parts of the Polish coast (Łabuz 2014). The exposed western Lithuanian coast is also strongly threatened by storm surges (Dailidiene et al. 2006). Erosion has totally consumed the foredunes along the Curonian Sandspit over past couple of decades.

The German Baltic Sea sandy coast between Warnemünde and Darss Peninsula is eroding rapidly (Kortekaas et al. 2010). Near Warnemünde and on the west coast of Darss, erosion has reached 3 m since the mid-twentieth century (Kortekaas et al. 2010), probably due to heavy waves attacking the exposed west coast in winter. The November 1995 storm event caused material loss from the sandy shores of Usedom Island (Schwarzer et al. 2003). The situation was similar on the western and central Polish coast, where sea level reached 2 m amsl and coastal dunes retreated 5–10 m (Łabuz 2005). Erosion along the dune shores in the Kattegat is up to 2 m year^{-1} (Clemmensen et al. 2011).

20.3.4 Moraine Soft Cliffs

The moraine soft cliffs of the Baltic Sea region were formed during the last ice age and mainly extend from the southern (Denmark, Germany, Poland and southern Sweden) to the south-eastern coastline (Kaliningrad region and part of Latvia and Estonia coast).

The high soft cliffs of the Sambian Peninsula (northwest of Kaliningrad) erode at least 1–1.5 m year^{-1}, with a maximum of up to 10 m year^{-1}, and rates have increased over the past decade (Chubarenko et al. 2009). Along the Latvian or Lithuanian coast, the average rate of retreat may be 1–2 m year^{-1} (Povilanskas 2002a; Eberhards 2003). In Latvia, the long-term mean rate of cliff retreat was 0.5–0.6 m year^{-1} in the latter half of the twentieth century, reaching a maximum of 1–1.5 m year^{-1} along particular stretches 10–20 m high (Jūrkalne area). After 1980/1981, rates of coastal retreat became two to five times higher, reaching 1.5–4 m year^{-1} (Eberhards and Saltupe 1995). Soft cliffs in Latvia retreated 0.5–4 m on average after the storm surge of Hurricane Erwin in January 2005 (Eberhards et al. 2006). The severe storms of 1993, 1999 and 2001 each caused a coastal retreat of 3–6 m, with a maximum of up to 20–30 m at the dune coast of Cape Bernāti (Eberhards 2003).

Cliff dynamics along the Polish Baltic Sea coast indicate that the main erosion factors are rainfall and storm surges, with cliff erosion rates of up to 1.5 m year^{-1} (Subotowicz 1982; Zawadzka-Kahlau 1999; Florek et al. 2008). Cliffs may become unstable when the toe is undercut by waves (Prior 1977). Cliff retreat is also related to intense rainfall triggering landslides. Mudslide debris discharged from the cliff slope channels onto the foreshore and forms elongated rounded lobes, helping cliff retreat. Slides are usually caused by high water content in the soil after heavy rain. Landslides caused by heavy rain occurred in 2007 and 2009 on the western Polish coast. On the German Baltic Sea coast, the average rate of shoreline retreat is about 0.4 m year^{-1} (Sterr 2008). Sea-level rise is expected to increase cliff erosion on the German coast and to increase the supply of sediment to the coastal zone (Hoffmann and Lampe 2007). Today, the maximum retreat on the western part of the coast is 2.5 m year^{-1} at Heiligenhafen, near which marine accumulation creates a sandy spit (growth 2–3 m year^{-1}, Hofstede 2008). One of the highest rates of cliff abrasion in Germany is in the area of the Streckelsberg scarp, on the outer shoreline of Usedom, as shown by comparing the old Swedish map of 1695 with a map for 1986. Maximum coastal retreat here exceeds 300 m (Schumacher 2002). The mudslides of the Eocene cliff clays in Denmark are also connected to climatic events (Prior 1977). Landslides extend up to 300–400 m inland. The combination of high water levels and strong wind has often resulted in severe damage to the soft cliffs coast of the Ystad municipality. In southern Sweden, the soft moraine cliffs have retreated 1–1.5 m year^{-1} over the past 150 years (Hanson 2002). The highest Baltic moraine cliffs of Wolin Island in Poland are retreating due to storm surges by 0.3–1.5 m year^{-1} (Kostrzewski and Zwoliński 1995).

20.3.5 Rocky Coasts and Hard Cliffs

Rocky cliffs in Sweden and Finland are resistant to marine erosion and are mainly formed from hard granites. The Latvian low cliffs of hard sandstone, covered with a thin layer of Quaternary deposits, are eroded on average by over 0.5 m year^{-1} by westerly storm surges (Eberhards 2003). Danish and Estonian klints are eroded by storm surges but also by weathering of chalk and limestone. In Denmark, cliff erosion reaches 0.5 m year^{-1}. The German chalk cliffs on Rügen Island also erode, with landslides occurring after heavy rain. The maximum retreat of the Jasmund chalk cliff is over 1 m year^{-1} (Lampe 1996).

20.4 Potential Climatic Threats to Coastal Areas

Climate-related ocean–atmosphere oscillations, such as the North Atlantic Oscillation (NAO) in the northern hemisphere, can lead to coastal changes (Viles and Goudie 2003). In recent decades, an increase in North Atlantic storm frequency has been linked to a strongly positive NAO index (Chap. 4). Many coasts experience erosion and habitat loss but only a few studies clearly define the relationship between observed coastal land loss and the rate of sea-level rise (Zhang et al. 2004). Although many studies have shown that periods of increased storminess increase coastal erosion and change, it is not always possible to relate increases in coastal storm frequency to specific modes of climatic variability (Viles and Goudie 2003).

20.4.1 Sea-Level Rise

Sea-level rise within the Baltic Sea is well documented (Richter et al. 2007, Chap. 9). Sea-level rise is being observed at coastal stations along the southern Baltic coast up to the southern part of Estonia in the east and in southernmost Sweden in the west (Harff and Meyer 2007, Chap. 9). On the Polish coast, sea-level rise was 0.6–1.5 mm year^{-1} in the last century but increased to 2.2–3.9 mm year^{-1} during the last 20 years (Pruszak and Zawadzka 2008, Chap. 9 and Fig. 9.7). The southern Baltic coast is sinking at a rate of 1–2 mm year^{-1} (Siegel et al. 2004; Sterr 2008; Pruszak and Zawadzka 2008; Clemmensen et al. 2011).

Rising sea levels are expected to exacerbate coastal erosion in southern Sweden and to increase flood risk along its western and southern coasts. Rocky northern Sweden is less prone to flooding and erosion as the rise in sea level is being counteracted by land uplift. The settlements most affected by increased flood risk are along the Skåne coast and the town of Göteborg in Västra Götaland (European Commission 2004). In Poland, over half of the densely populated low-lying coast is threatened by sea-level rise (Rotnicki and Borówka 1990; Zeidler et al. 1995; Łabuz 2012).

Sand dunes may be affected by rising sea level, changes in erosion/accretion patterns and changes in groundwater level, as reviewed by Carter (1991). For dunes, the most likely response to changes in sea level is a two-way redistribution of the sand: seaward movement and deposition of eroded material below the new sea level, and landward sand drift resulting in new foredune growth behind the eroded dune as well as sand deposition on older dune land further inland. Groundwater level is a major factor in plant distribution in coastal dunes (Carter 1991) and may strongly influence dune slack vegetation. The sequence of soil moisture changes in dune slacks may, however, be modified by sand movements that change the shape of the dune system. As sea level rises, estuaries and lagoons attempt to maintain equilibrium by raising their bed elevation. This will change environmental conditions, increasing the threat to low-lying coastal areas such as marshes and wetlands. Some current areas of salt marsh and wetlands could decrease or even disappear. Saltwater intrusion and the resulting changes in coastal ecosystems are a problem in the northern Curonian Lagoon. An important area in this lagoon is the Nemunas River Delta, with its ecologically significant wetlands as well as agricultural land. Low coastal areas in Estonia are permanently under threat from rising sea level (Kont et al. 1997; Orviku et al. 2003). In Poland, low coastal areas behind dune belts on postglacial valleys or surrounding coastal lakes are threatened by future sea-level rise (Rotnicki and Borówka 1990). The land area up to 3–3.5 m amsl could be gradually flooded as sea level rises. These areas are widely distributed along the Polish and German coast and are used for agriculture and human settlements. Many lowlands on the German coast are vulnerable to sea-level rise; some densely populated (Sterr 2008). Coastal settlements are also at risk in Poland, Lithuania, Latvia and Estonia.

20.4.2 Storm Surges

The greatest storm surges on the southern Baltic Sea coast come from the north-east (Schwarzer et al. 2003; Kowalewska-Kalkowska and Kowalewski 2005; Pruszak and Zawadzka 2005). On the southern Baltic Sea coast, severe surges were recorded in January 2002 and 2012; November 1995, 2004 and 2006; and October 2009 (Schwarzer et al. 2003; Florek et al. 2008; Gurwell 2008; Łabuz 2009b, 2012). The most destructive storm surges to coastal areas are those with a water level 2 m amsl. The most catastrophic sea surges on the Polish coast overflow all relief forms lower than 3.5 m. The eastern Baltic Sea coast is mainly affected by westerly storms (Žilinskas 2005; Visakavičius et al. 2010; Kelpšaitė and Dailidienė 2011; Soomere et al. 2011). The Lithuanian and Latvian coasts are exposed to the west and vulnerable to erosion. In Estonia, more extreme sea levels and storm surges are associated with warmer winters (Suursaar et al. 2003). Storms, driven mainly by westerly winds, change the position of pebble capes and other spits. Neva Bay in the eastern Gulf of Finland is open to the west, with one of the longest fetches in the Baltic Sea—up to 400 km.

Strong westerly storms cause a predominant drift to the east and erode the coast (Ryabchuk et al. 2011). At present, the coast in these countries is threatened more by major storm surges than by sea-level rise. Such surges were reported in November 2001, September 2004 and January 2005 and 2007 (Tõnisson et al. 2006, 2008; Suursaar et al. 2003, 2008; Koltsova and Belakova 2009; Dailidienė et al. 2011).

In the Gulf of Finland, the frequency of stormy days varied greatly during the latter half of the twentieth century, from a minimum in the 1960s to a maximum in the last two decades (Orviku et al. 2003). Repeated floods were reported in 2004–2008, with strong winds of 20–23 m s^{-1} causing ice jams, water damming and high waves (Kurennoy and Ryabchuk 2011). In Denmark, most storm damage occurs when water level rises at least 2 m (Sorensen et al. 2009). The Russian coast of the Gulf of Finland was totally changed after storms on 29 November 2006 and 11 January 2007 (Ryabchuk et al. 2011). Each storm surge reached a water level of 1.5–2 m amsl. In Lithuania, changes in the wind and wave regime in the latter half of the twentieth century intensified the natural sediment transport and accelerated coastal erosion or stopped accumulation processes (Kelpšaitė and Dailidienė 2011). Consequently, Lithuania is now more at risk of flooding due to storms (Visakavičius et al. 2010).

The extent of coastline erosion and retreat depends both on the height and duration of the sea surge. Storms of similar magnitude coming immediately after other storms may have much less geomorphic effect than those which occur in isolation (as all the geomorphic work will have been done by the preceding storms). This effect was observed in Poland between November 2001 and February 2002 with several storm surges in which shoreline retreat and dune sand volume erosion was in total comparable to the first two events in November. Similar situations occurred in 2006/2007 in Poland, Germany and in 2007 in Estonia and Russia. Alternatively, long clusters of storms may have a compounding, more serious geomorphic effect than would the same number of events operating in isolation (Viles and Goudie 2003). The combination of a major rise in water level in winter months together with the shortening of the ice season may significantly increase the frequency of events that had previously been very rare (Räämet and Soomere 2011). Specifically, an increase in the frequency of three factors together—high water level, rough seas and the absence of ice in the nearshore zone—may be responsible for the intense erosion of depositional coasts in the eastern Baltic Sea.

20.4.3 Land Erosion and Habitat Loss

Owing to storm effects, almost two-thirds of Polish coastal dunes are eroded (Zawadzka-Kahlau 1999; Pruszak and Zawadzka 2005; Łabuz 2009b). The 75 % of the sandy coast with low-lying hinterland is at risk of flooding (Rotnicki and Borówka 1990; Łabuz 2012). Almost 75 % of German sandy coasts are threatened by erosion (Sterr 2008). The situation is similar in Lithuania, Latvia and Estonia (Eberhards 2003; Milerienė et al. 2008; Ratas et al. 2008). Over 65 % of the Latvian coast is eroded (Eberhards 2003). About 40 % of the Russian Neva Bay coast has recently been heavily eroded (Ryabchuk et al. 2011). With no data between storm events it is hard to estimate when coastal changes occurred (Suursaar et al. 2008).

Land erosion is highlighting the need for effective coastal protection measures. However, in some areas, so far accumulated coastline (mainly sandbars) is shrinking due to the use of coastal protection measures. This is because coastal protection measures used in one place may reduce the transfer of sand from land to longshore drift and thus its accumulation in another. This problem will increase, for example, in Poland where almost 6 % of the coast has been stabilised over the past ten years.

Erosion reduces the area available for plant and animal communities. As a result, coastal habitats are likely to be highly sensitive to sea-level rise (Vestergaard 1991, Chap. 16). The entire vegetation of cliff slopes is threatened by repeated landslides. Many such habitats are listed in Natura 2000 (www.natura.org). Dune erosion also causes loss of valuable coastal habitats (Łabuz 2012, Chap. 16). As sand dunes develop beyond the direct influence of seawater, dune vegetation is less vulnerable to sea-level rise than coastal meadow vegetation.

20.4.4 Precipitation and River Discharge

Increasing precipitation will increase river flows and thus flooding in low-lying coastal areas. Levels of coastal lakes can vary greatly in response to changes in rainfall in their catchment and to changes in evapotranspiration. After the spring melt of 2010, large amounts of water from coastal swamps broke through the narrow dune belt near Kołobrzeg on the Polish west coast. Seawater entered the brackish environment, completely destroying this fragile habitat (Łabuz 2012). Furthermore, increasing precipitation may increase river flows and encourage vegetation growth (Viles and Goudie 2003). Heavy rainfall is a destabilising factor for soft cliffs and caused landslides in Poland in 2007 and 2008 (Łabuz 2012).

On the other hand, reduced river input to the southern part of the Baltic catchment in future (see Chaps. 5 and 12) could reduce the supply of sediment for beaches or spit rebuilding (Kont et al. 2008). River mouths reinforced by breakwaters hinder long-shore sediment movement, leading to sediment build-up on one side of the river mouth and a lack of sediment on the other. In recent years, beaches in western Estonia have become narrower due to a decline in run-off from rivers (Kont et al. 2008). As the beach is lost, fixed structures or coastal land nearby are increasingly exposed to the direct impact of storm waves and may be damaged or destroyed unless new protective measures are taken to disrupt these natural coastal processes.

20.4.5 Ice-Cover Melt

Decreased ice-cover duration in the Gulf of Finland in the recent past (see Chap. 8) has reduced the protection of coastal sediments during the stormy period (Ryabchuk et al. 2011), even though ice jams pushed onto land by waves may also cause erosion of dunes and cliffs. Ice jams were observed on the southern Baltic Sea coast in the winters of 1995, 2003, 2010 and 2011. Their impact was seen in narrower and lower beaches after ice melted in late spring. On the northern Baltic Sea coast, ice sheets may have a bigger impact on the coast. In February 2008 in Neva Bay, ice strongly eroded coastal land and human structures (Ryabchuk et al. 2011). On the other hand, declining ice cover in the northern Baltic Sea would mean greater potential for coastal erosion due to wave action.

20.5 Conclusion

The Baltic Sea coasts are already experiencing the adverse impacts of climate change, mainly due to changes in sea level and storminess. Coasts are particularly vulnerable to extreme events. The response of the coastal system varies over different timescales, owing to changes in local dynamics or geo-morphological conditions as well as to the intensity of the driving forces. Identifying the contribution of climatic change to geomorphic change can be difficult, and its importance will also vary regionally (Viles and Goudie 2003). The main drivers of change in coastal geomorphology are geological structural resistance, changes in sea level, long-shore currents and storm surges. These factors are responsible for both coastal erosion and accumulation, as well as for the emergence and variability of beaches. Coastal protection measures are major factors at the local scale. At the regional and global scale, coastal erosion is induced by an increasing number of severe storm surges and sea-level rise. These factors are leading to the following:

- loss of sediment for coastal rebuilding
- loss of valuable dynamic coastlines
- loss of coastal resilience
- loss of valuable natural habitats
- loss of economic value and private property
- unpredictable land change due to extreme storm events
- increasing cost to society in terms of coastal protection measures.

The impact of climate change on coasts is exacerbated by the increasing need for coastal protection by human activities (Bardach 1989; Beukema et al. 1990; van der Meulen et al. 1991). There are two main types of management strategy. The first is to stabilise the existing coastline and thus protect coastal and neighbouring areas from future change. The second, which is more resilient and depends on human adaptation to coastal change, accepts that in some places the coast would be not protected and would be left in its natural state. The costs for adapting to climate change are generally much lower than the costs of damage incurred through a lack of adaptation. Local communities are currently better prepared for extreme events; but still very few local administrations are attempting to estimate costs associated with the risks of future climate change. Projecting future costs is also complicated because different local authorities may prefer different coastal development and management policies, for example, for protection against coastal erosion. European projects such as ASTRA (Astra-project 2007), EUROSION (European Commission 2004), SINCOS (Richter et al. 2007) or MICORE (Ferreira et al. 2009) have provided new data and insights regarding storm effects on the Baltic Sea coast. Administrative borders in coastal areas and efforts for coastal protection differ among Baltic Sea countries, even between neighbouring nations. Some management actions are complicated by the morphodynamic changes caused by artificial coastal stabilisation in some places and its natural disruption in others. The coastal zone is a linear environment that knows no national or decision-making boundaries. Decadal-scale climatic change and variability may initiate a pulse of activity that results in a complex, nonlinear landscape response.

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Part V

Socio-economic Impacts of Climate Change

Introduction

The socio-economic impacts of climate change on the scale of the Baltic Sea basin have been the subject of a limited number of international and national studies. Two aspects are discussed here: impacts on the managed rural landscape (i.e. agriculture and forestry) and impacts on urban settlements (i.e. cities and towns). The following chapters focus purely on the impacts of climate change on these human-generated systems; how these systems may affect the regional climate is addressed in Part VI.

Chapter 21 assesses how the productivity and vulnerability of forestry and agricultural systems are affected by climate change in the Baltic Sea basin and concludes that the northern and southern parts of the catchment are impacted quite differently. It is also clear that the potential human responses to climate change, such as changes in forest management schemes or crops, are themselves likely to affect socio-economic conditions in the Baltic Sea region. Chapter 22 examines observed and potential climate change impacts on urban settlements. There is a limited body of literature available on climate change impacts on urban complexes in the Baltic Sea basin, and that which does exist focuses on a few cities located around the margins of the Baltic Sea. Impacts relevant for cities and towns include rising temperatures, changing patterns of precipitation and sea-level rise, and the extent of these impacts differs depending on the location of the urban complex.

The Baltic Sea basin is home to a diverse range of human activities and the socio-economic dimension of climate change in this region deserves significant attention in the future.

21 Socio-economic Impacts—Forestry and Agriculture

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Abstract

Climate change affects the vulnerability and productivity of forestry and agricultural systems, predominantly by changes in precipitation and temperature patterns. Indirect impacts are altered risk of damage, for example, by longer periods of drought stress and other biotic and abiotic disturbances. While southern and eastern parts of the Baltic Sea basin are likely to experience a net impact of climate change that is negative for production, northern and western regions are likely to experience a general increase in production. As a result, land-use potentials will change and will foster adaptation and mitigation measures. In the northern region, forest management adaptation may lead to substantial yield increases, while in the south management, adaptation may be required to counter deteriorating conditions. Comparable conclusions can be drawn for agricultural management: if adaptation potentials are fully exploited, substantial yield increases can be expected for certain crop species. In the southern areas and for certain species, deteriorating conditions and possibly increasing climatic variability are projected. Both climate change impacts and human responses will affect socio-economic conditions in the Baltic Sea basin.

21.1 Introduction

Climate change affects the vulnerability and productivity of agricultural and forestry systems predominantly through changes in precipitation and temperature patterns and by changes in the frequency and intensity of risk factors for damage such as droughts, floods, storms and biotic disturbances such as pest infestations. In addition to changes in environmental factors, changing energy policies may influence agricultural and forestry systems through changes in demand for biomass for use as a biofuel. While southern and eastern Europe are likely face a net effect of climate change that is negative for production, northern and western regions are likely to see a general increase in production (EEA 2006). As a consequence, land-use potentials will change and will foster the need for adaptation and mitigation measures. Both climate change impacts and human responses will affect socio-economic conditions in the region.

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_21

This chapter focuses on managed forest land and agricultural land, while the effects of current and future climate change on forest growth in general are covered in Chap. 16.

21.2 Climate Change and Forest Management

21.2.1 Forest Management in the Baltic Sea Basin

The main forest types in the Baltic Sea basin are boreal coniferous forests north of 60°N and temperate deciduous forests south of 60°N (EEA 2007). Climatic conditions in boreal forests are characterised by a growing season of 3–6 months with a mean temperature of about 5 °C and a water surplus (precipitation exceeds evapotranspiration, Otto 1994). There is a clear north–south gradient in temperature and an east–west gradient in humidity. A short growing season and low nitrogen supply are the main factors limiting forest growth in the boreal forests, whereas low water availability periodically limits forest growth over large areas in the temperate southern parts of the Baltic Sea basin (BACC Author Team 2008; Gundale et al. 2011). Information on the forests of the Baltic Sea basin is provided in the report on the status of forests in Europe 2011 (Forest Europe, UNECE and FAO 2011). Figure 21.1 and Table 21.1 indicate forest areas and their share of the land area by country (Forest Europe, UNECE and FAO 2011).

Table 21.1 Basic forest data for countries of the Baltic Sea basin (excluding Russia) in 2010 (Forest Europe, UNECE and FAO 2011)

Country	Total land area (1000 ha)	Forest[a] and OWL[b] (1000 ha)	Percentage of total land area	Forest and OWL per inhabitant (ha)
Denmark	4242	635	15	0.1
Estonia	4239	2337	55	1.7
Finland	30,408	23,116	76	4.3
Germany	34,877	11,076	32	0.1
Latvia	6229	3467	56	1.5
Lithuania	6268	2249	36	0.7
Poland	30633	9316	30	0.2
Sweden	41031	30,625	75	3.3

It should be noted that Forest Europe (2012) uses specific classifications of forest land and other wooded land that result in slightly different figures to those from the FAO used in Chap. 25

[a] Forests: Land spanning more than 0.5 ha with trees higher than 5 m and a canopy cover of more than 10 %, or trees able to reach these thresholds in situ. It does not include land that is predominantly under agricultural or urban land use (FAO 2004)

[b] Other wooded land (OWL): Land not classified as forest, spanning more than 0.5 ha; with trees higher than 5 m and a canopy cover of 5–10 %, or trees able to reach these thresholds in situ; or with a combined cover of shrubs, bushes and trees above 10 %. It does not include land that is predominantly under agricultural or urban land use (FAO 2004)

Most forests in the Baltic Sea basin are managed, and forestry is mainly based on native tree species that invaded the region after the last glaciation. However, many forests in the area have been cleared for agriculture, and forests still

Fig. 21.1 Total and percentage forest area by country in 2010 (Forest Europe, UNECE and FAO 2011)

dominate the landscape only in northern Europe (e.g. Sweden, Finland and north-western Russia). In the temperate parts of the Baltic Sea basin, the current tree species composition is determined by past land use and management activities rather than by natural factors (Ellenberg 1986).

The total area of forest and other wooded land in the Baltic Sea basin (excluding Russia) is about 82 million ha. The total volume of stem wood is about 11,290 thousand m^3 and is dominated by the native tree species Scots pine (*Pinus sylvestris*) and Norway spruce (*Picea abies*). Together, these two species account for more than 70 % of the total stem volume, with smaller contributions by deciduous trees (mainly Birch, *Betula pendula*) (BACC Author Team 2008). The share of deciduous trees is greater in the temperate part of the Baltic Sea basin, which forms the transition from the temperate deciduous forest zone to the boreal coniferous forest. The role of exotic species is most important in the temperate zone, but even there their share is small. Forests in Finland and Sweden comprise about 43 % (by volume) of the total forest resources in the Baltic Sea basin (excluding Russia).

Table 21.2 provides an overview of the increment (net growth) and felling in 2010 (Forest Europe, UNECE and FAO 2011). These data indicate the productivity and utilisation rates of the forests. All countries of the Baltic Sea region manage their forests sustainably from a wood stock perspective, that is, felling does not exceed increment. However, sustainable forest management usually includes ecological and social aspects (Forest Europe 2012). In regions where biodiversity is being lost, sustainability criteria are not being met.

According to Forest Europe, UNECE and FAO (2011) the value of roundwood removals from forests in 2010 was almost EUR 10,000 million in Estonia, Finland, Germany, Lithuania, Poland and Sweden, with a corresponding employment of roughly 180,000 FTE (full-time equivalents) in the forest sector and 591,000 FTE in the manufacturing of wood and paper. Throughout the region, increment exceeds felling through increased stocking and maturing of forest resources. In the near future, forest resources are expected to increase further due to afforestation of agricultural land and the projected increase in forest growth under a warmer climate.

21.2.2 Impacts on Forest Management

Model simulations indicate that rising temperatures could improve tree growth in the northern boreal zone, while changes in precipitation are not likely to have a major effect on growth at these latitudes (Bergh et al. 2003, 2007; Ge et al. 2011b). In the southern parts of the Baltic Sea basin, tree growth is strongly water limited (Lasch et al. 2002, 2005). Here, an increase in temperature but without an increase in precipitation could further exacerbate the water deficit and thus decrease growth. In general, the temperature response optimum (the ability of tree species to manage higher temperatures) is higher and the effect of rising temperature is more positive, if precipitation also increases, whereas the main effect of higher temperatures is negative and the temperature response optimum lower if precipitation is reduced (Lindner et al. 2010). For example, growth of Norway spruce in the southern boreal zone is projected to increase up to 2050, but then decline due to more frequent dry spells during the growing season (Kellomäki et al. 2008; Ge et al. 2011b; see also Chap. 16). In the continental temperate zone, forest growth in general is limited more by water than by temperature, but the effect of increased levels

Table 21.2 Increment (net growth) and felling in forests available for wood supply, 2010 (Forest Europe, UNECE and FAO 2011, excluding Russia)

Country	Net annual increment		Felling			Value of roundwood removals
	(1000 m^3)	(m^3 ha^{-1})	(1000 m^3)	(m^3 ha^{-1})	(Percentage of annual increment)	(million EUR)
Denmark	5176	9.5	2371	4.1	40.9	–
Estonia	11,201	5.6	5714	2.8	51.0	196
Finland	91,038	4.6	59,447	3.0	65.3	1858
Germany	107,000	10.3	59,610	5.6	55.7	3003
Latvia	16,500[a]	5.5[a]	12,421	4.0	–	–
Lithuania	10,750	5.7	8600	4.6	80.0	181
Poland	67,595[b]	8.0[b]	40,693	4.8	–	1291
Sweden	96,486	4.7	80,900	3.9	83.8	2656

[a]2000
[b]2005

of atmospheric carbon dioxide (CO_2) may partly offset the potential negative effects of climate change (Freeman et al. 2005; Lindner et al. 2005).

The human response to climate-related impacts on forest management is likely to concern deteriorating condition and the possibilities for adapting forest management practices to address, for example, increased water stress or higher temperatures. This could comprise changes in stand structure (e.g. wider spacing), thinning measures, potential underplanting or selection of more suitable tree species and provenances.

Across much of the Baltic Sea basin, climate change has the potential to improve site and growing conditions, such as by removing formerly limiting conditions through rising temperature or nitrogen availability in the northern latitudes and by extending growing periods (Linderholm 2006). Climate change effects on forest growth (see Chap. 16) may lead to changes in forest yield, but may also affect management practices. The changing conditions could potentially allow an intensification of management (e.g. a shortening of rotation length and adjustment of thinning schedules), possibly even on formerly marginal sites and the introduction of new tree species. However, an intensification of forest management may have consequences affecting other goals, such as carbon sequestration and biodiversity protection. Furthermore, and in terms of practical silviculture, warming could make forest resources in wet areas inaccessible, due to shorter periods of frozen ground, thus reducing the availability of such timber. This could increase harvesting and transport costs and threaten supply for the wood industry (Lindner et al. 2010).

21.2.2.1 Adaptive Forest Management May Support Higher Yields

In general, given adequate precipitation and accessibility, changes in management have the potential for effective adaption to climate change. Regardless of the future climate scenario, it was found that shifting from current practices to thinning regimes that allowed higher stocking of trees resulted in an increase of up to 11 % in carbon uptake by the forest ecosystem. It also increased the carbon content in timber yield by up to 14 % (Garcia-Gonzalo et al. 2007b). Briceño-Elizondo et al. (2008) supported this conclusion for a revision of forest management practices not only for timber production, but also for benefits such as carbon sequestration and other amenities including biodiversity.

Kellomäki et al. (1997) simulated the impact of higher CO_2 concentrations, temperature and precipitation on Scots pine in southern Finland (61°N). They reported an increase in timber yield of up to 30 % and through that a potential shortening of rotation periods by 9 years (for a temperature rise of 0.4 °C decade^{-1} and a precipitation increase of 9 mm decade^{-1}), 17 years (for a CO_2 elevation of 33 µmol mol^{-1} decade^{-1}) and 23 years when all three factors are increased. The authors concluded that increased timber supply and profitability of forest management could be expected under a future climate. Similar results were found by Karjalainen (1996) who reported that timber production could increase substantially for 30 mixed species on medium-fertility stands in southern Finland over a 300-year period. More recently, Garcia-Gonzalo et al. (2007a) simulated different management approaches for Finland. They found the greatest increase in timber yield and percentage of saw logs to occur for a thinning regime with high stocking over 100 years. A gradual rise in temperature and precipitation and an elevation in CO_2 enhanced growth by about 24 %, resulting in a 12–13 % increase in timber yield. Bergh et al. (2007) estimated that Swedish forest growth could increase 10–50 % by 2070–2100 under the SRES A2 scenario (IPCC 2000), more in the north and less in the south and central-west. Norway spruce would be favoured in the north and Scots pine in the south, suggesting a corresponding change in preferred species for regeneration at sites suitable for both species (Swedish Forest Agency 2007).

Other studies also project increased growth and underline the need for adaptive management practices. Kärkkäinen et al. (2008) estimated the recovery of industrial wood and raw material for wood energy (biofuel) under two different cutting scenarios, contrasting 'current' and 'climate change' conditions for the next 50 years. The results indicated an average increase of about 10 % for industrial wood and 12 % for wood energy under a sustainable cutting scenario for Finland. A maximum cutting scenario would give increases of 33 and 32 %, respectively.

Pussinen et al. (2002, 2009) also found evidence that future climate change is likely to increase harvest removals and economic profitability in Finnish forestry. For forest management, this would allow shorter optimum rotations based on mean annual yields, for example for Scots pine in southern Finland. The highest mean annual carbon stock in forests over a rotation period, however, was achieved with longer rotation periods and higher nitrogen deposition (Pussinen et al. 2002; De Vries et al. 2009). In contrast, further warming may lead to reduced forest carbon stocks mainly due to increased decomposition of soil organic matter and thus lower forest soil carbon stocks. However, at the centennial perspective, the rate of delivery of bioenergy, which is largely correlated to harvesting rates, might be more important for climate change mitigation than potential changes in carbon stocks. In Sweden, the share of harvested biomass largely used directly for energy production has increased from ~40 % to near 50 % (bark, sawdust, lignin, branches and tops, wood from early thinning, etc.) over recent decades. This could also be the case in other countries due to the adoption of energy policies that restrict the use of fossil fuels and/or nuclear power.

According to Kellomäki et al. (2008), forest growth may increase by 44 % in Finland with an increase of 82 % in the potential cutting drain (maximum sustainable removals under a given management). They stressed the need to choose appropriate species and rotation periods and to consider changing forest structures and the requirement to sustain the productivity of forest land under climate change. Garcia-Gonzalo et al. (2007a) stated that both the climate change scenario and management regime influenced the profitability of timber production for a process-based ecosystem model applied to analyse the effects of climate change and management on timber yield for a forest management unit in Finland (63°N). The authors indicated that choosing the 'wrong' management regime, instead of the best one, could lead to an average economic loss of EUR 166 ha^{-1}. The highest species-specific opportunity costs (as lost potential benefit) were found for Scots pine (EUR 227 ha^{-1}) and the lowest for silver birch (*Betula pendula*) (EUR 53 ha^{-1}). They concluded by stressing the need to adapt future management to utilise the increase in growth under climate change (see also Matala et al. 2009).

Further results on climate change implications for forest management were provided by Briceño-Elizondo et al. (2006a, b), who tested the effect of eight different thinning regimes on Scots pine, Norway spruce and silver birch stands in the southern and northern boreal areas of Finland for a 100-year simulation. Results indicated that thinning regimes that increased the stocking of the tree population increased the mean carbon stock in the forest and timber yield, compared to the current thinning guidelines, regardless of tree species and climate scenario. Climate change enhanced stocks more in the north than the south. The results indicated that carbon sequestration in the ecosystem may be enhanced with no loss in timber production (Briceño-Elizondo et al. 2006a). According to Briceño-Elizondo et al. (2006b), thinning regimes that increased mean stocking over the rotation all increased total growth and timber yield, regardless of tree species and site. The authors highlighted the potential to exploit the benefits that climate change seems to provide in the form of increased growth and timber yield in the boreal conditions and suggested that current management rules be revised.

These findings support the results provided by Karjalainen (1996) on the effect of forest management on carbon sequestration in a 300-year simulation for 30 mixed species stands in southern Finland. Karjalainen explained that the total carbon balance (vegetation, litter, soil organic matter and products) was higher in unmanaged stands during the first 100 years, but not in the second or third. Under climate change conditions, results projected substantially enhanced timber production and carbon sequestration. This is supported by Garcia-Gonzalo et al. (2007a) who reported a 12–13 % increase in timber yield for an adapted thinning regime with high stocking over a 100-year rotation period under climate change impact. Matala et al. (2009) described the effect of forest management on carbon sequestration and increased production potential due to climate change over a 50-year period (2003–2053) in the growing stock of trees in Finland compared to current values (an initial amount of carbon in the growing stock of 765 million tonnes). They found an increase of ~17 % for growing stock without climate change, but an increase of about 38 % under sustainable production and assuming a gradually warming climate until 2053. Another simulated management strategy, the maximum net present value (NPV) of wood production, resulted in an increase of 18 and 34 %, respectively, compared to the initial growing stock. The results show that future development of carbon sequestration and growing stock is not only dependent on climate change scenarios but on forest management adapting to changing conditions (Matala et al. 2009). Similar conclusions were drawn by Köhl et al. (2010) for, among others, temperate regions in north-eastern Germany.

21.2.2.2 Adaptive Forest Management is Required to Counteract Negative Impacts

In the southern part of the Baltic Sea basin, reduced precipitation in combination with higher temperatures is likely to result in reduced growth and increased risk of fire and pest outbreaks (Kellomäki and Kolström 1994; Lasch et al. 2002, 2005; BACC Author Team 2008; Köhl et al. 2010). However, adaptive forest management may counteract these unfavourable conditions.

In southern Finland, reduced precipitation may lead to lower productivity of Norway spruce. Ge et al. (2011a, b) examined the potential for different thinning regimes to improve carbon uptake, stem growth and timber yield. Again, the necessity for adaptive management systems was highlighted. Similarly, based on an ecosystem model for southern Finland in 2010–2099, Alam et al. (2010) reported a stronger productivity effect on forest structure than changing climate.

Peltola et al. (2010) supported the need for adapted forest management when considering forest damage. They explained that changing forest structure, such as birch (*Betula* spp.) replacing Norway spruce in southern Finland, can reduce the risk of wind damage in winter but increase risk during periods of unfrozen soil. Increasing rotation length, for example by 20 years, can increase carbon stocks in the living biomass for Scots pine in southern Finland and north-eastern Germany, but would simultaneously lead to decreased timber harvests (Kaipainen et al. 2004), which in turn can reduce potential for bioenergy delivery.

Similar results were found for the southernmost part of the Baltic Sea Basin. Köhl et al. (2010) modelled different climate change scenarios and management types using the

German national forest inventory data and two climate change scenarios from the IPCC's Special Report on Emission Scenarios (IPCC 2000): A1B, rapid and successful economic growth; B1: high level of environmental and social consciousness combined with a globally coherent approach to a more sustainable development. Three management types were used—'maximum profit oriented', 'diameter limit cut' and 'maximum net annual forest rent'—to evaluate their effects on future productivity and species composition of German forests. The results were based on changing precipitation and temperature patterns and show clear north–south differences. Overall, Köhl et al. (2010) concluded that the effects of different climate change scenarios on the future productivity and species composition of German forests are minor compared to the effects of forest management. Garcio-Gonzalo et al. (2007b), Briceño-Elizondo et al. (2008a, b) and Alam et al. (2008) also reported increased benefits from management schemes adapted to climate change.

Adaptive management requires consideration given to changing the species composition, especially when lower precipitation is expected. Lasch et al. (2002, 2005) reported for north-eastern Germany (Federal State of Brandenburg) aims of increasing the share of deciduous and mixed forests as an adaptation to climate change. While climate change led to a reduction in groundwater recharge of about 40 %, more intensive management slightly increased groundwater recharge (Lasch et al. 2005). Simulation studies with three management scenarios indicated that the short- to mid-term effects of climatic change in terms of species composition were less severe than expected. However, comparing diversity measures indicated a decrease in species diversity in contrast to an increase in habitat diversity under climate warming (Lasch et al. 2002).

Lasch et al. (2005) concluded that the potential for adaptive management based on changes in rotation length and thinning is very limited in the Federal State of Brandenburg, which is characterised by poor sites and dry conditions. They also concluded that it is necessary to include forest transformation strategies in management impact analyses for forest planning under global climate change. In contrast, Köhl et al. (2010) demonstrated for all the north-eastern Federal States of Germany that management matters more than climate change and concluded that the negative effects of climate change can be reduced by adaptive management.

The potential northwards shift in tree species is important with respective to adaptation potential. Birch (*Betula* spp.), already the main deciduous species in the boreal zone, shows a positive response to rising temperature and low sensitivity to precipitation (Truon et al. 2007; Lindner et al. 2010). In contrast, oak (*Quercus* spp.) and beech (*Fagus sylvatica*) respond strongly to changes in precipitation in temperate zones. While temperature increase was generally negative for the growth of beech, oak showed a weak positive response (Lindner et al. 2010). With sufficient precipitation, both species seem capable of a northwards shift in abundance (Kramer et al. 2010). On this basis, the BACC Author Team (2008) recommended that consideration be given to incorporating other indigenous tree species, currently of minor importance in forestry, but with high potential for timber production or carbon sequestration under climate change. Further recommendations included an increased share of those broadleaved trees species considered to perform better under climate change, substitution of sensitive species by better adapted provenances and replacement of low-productivity tree populations (BACC Author Team 2008). The importance of choosing suitable tree provenances was emphasised by Kellomäki et al. (2008), who showed that southern provenances of Norway spruce would be less sensitive to climate change in southern Finland.

21.2.3 Concluding Comments on Management Implications

Considering the long time scales of forestry (rotation lengths of 40–140 years depending on species and region), it is clear that major climate change impacts could occur within the lifetime of existing tree stands. To a certain extent, this would limit the adaptive capacity of tree species to the variability of the existing generation (Köhl et al. 2010). However, the genetic variability of most common tree species is probably large enough to accommodate the mean changes in temperature and precipitation (Beuker et al. 1996; Persson and Beuker 1997).

For the Baltic Sea basin as a whole, there are likely to be shorter winters, longer growing seasons, changes in precipitation (Forest Europe 2012) and, potentially, changes in storm patterns (see also Chap. 11). Conditions for pest outbreaks and tree damage will change under a warmer climate, more often for the worse. In Fennoscandia, some of the economically most damaging pests could be favoured: spruce beetle (*Ips typographus*), pine weevil (*Hylobius abietis*) and root rot fungus (*Heterobasidion annosum*) (Swedish Forest Agency 2007). The likely effects of climate change on insect damage and major pest outbreaks are still largely unknown. Nevertheless, many damaging fungi and insects may expand their occurrence from Central Europe and further south to the Baltic Sea basin (Parry 2000). On the other hand, there is empirical evidence to suggest that elevated CO_2 and higher temperatures may increase the resistance of deciduous species to herbivore browsing and thus reduce the risk of forest damage (Mattson et al. 2004).

Wind felling may increase as winter soils are frozen for shorter periods, and water tables are higher. The species most sensitive to wind, Norway spruce, would be favoured if deer populations survived winters better and browsed more than today on other plant species. Unless hunting increased, this would increase the sensitivity of Fennoscandian forests to wind felling (Swedish Forest Agency 2007). There is also a risk that pathogens and insects having marginal impacts under present-day conditions could become more important and that new insect pests could move in from the south.

To maintain resilience, in terms of production, biodiversity and other forest uses, adaptive strategies should be considered at regeneration. For example, planting more tree species and favouring a higher number of species when cleaning and thinning than is usual today (Swedish Forest Agency 2007).

Problems may also be encountered due to the higher frequency and intensity of extreme events, such as droughts, storms and spring and summer freezing (CCIRG 1996; Nikulin et al. 2011), with consequent damage to forests. The Baltic Sea basin has only a few tree species of economic importance in forestry, such as Scots pine, Norway spruce, birch and oak. However, changing tree species composition may be an appropriate adaptive management strategy (Ge et al. 2011a). The following changes in tree species composition are possible adaptive management strategies:

- A shift from mono-species to mixed species stands (see Kolström et al. 2011).
- Incorporating other indigenous tree species, currently of minor importance in forestry, but with high potential for timber production or carbon sequestration under climate change.
- Increasing the share of broadleaved species assumed to perform better under climate change.
- Substituting species sensitive to drought and late spring frosts by drought-tolerant and frost-resistant tree species or provenances.
- Replacing low-productivity tree populations with high productivity ones when the current population does not make full use of the potential productivity of a site.

Changing tree species can be an appropriate adaptive management strategy for improving productivity, which also includes the adjustment of thinning (intensity, interval, pattern: from above/below). In this context, adjusting the rotation period is an effective means of managing timber production and the carbon budget of forests. Over the rotation, the timing and intensity of thinning determine the growth rate and stocking, which control the rate of carbon sequestration and the amount of carbon retained in trees and soils. In most European countries, growing stock is still increasing, because timber harvest (thinning, final felling) is less than the increment. This means that the total carbon storage in the forest is increasing. On the other hand, the age-class distribution of the forests in the Baltic Sea basin is shifting towards the older age classes, and the overall length of rotation is increasing. This means that the rate of carbon sequestration is declining, even though the carbon stores are large. Harvesting of over-mature old forests with subsequent regeneration with productive stands would make better use of the substitution potential of the forest, even though average carbon stocks are reduced. However, to meet the overall criteria for sustainable forest management (Forest Europe 2012), a sufficient area of old forest for conserving biodiversity must be retained in all countries.

21.3 Climate Change and Agricultural Ecosystems

21.3.1 Agricultural Production in Europe

Agriculture is the most important force driving land use globally. Nearly half of the total EU-27 land area is devoted to agriculture (Green et al. 2005; Stoate et al. 2009) and the productivity of European agriculture is among the highest in the world (Olesen et al. 2011). Despite a wide range of climatic conditions, soils, urbanisation, land use, infrastructure, economic and political conditions across Europe, rapid modernisation and intensification of farming systems have led to an unprecedented increase in agricultural productivity after the Second World War, particularly in western Europe (Bouma et al. 1998; Olesen et al. 2011). For example, Europe accounts for about one-fifth of global meat and cereal production, and average cereal yields in EU countries are more than 60 % above the world average (Olesen et al. 2011). Such agricultural intensification has dramatically simplified landscapes, affected carbon and nutrient cycling, facilitated species invasions, decreased native biodiversity and increased herbicide, pesticide and fertiliser use in recent decades (Matson et al. 1997; Tscharntke et al. 2005; Stoate et al. 2009; Flohre et al. 2011). These changes have had profound and far-reaching effects on ecosystem functions and services, also extending to terrestrial and aquatic ecosystems outside agro-ecosystems (Green et al. 2005; Stoate et al. 2009).

Environmental and socio-economic conditions largely determine agriculture in Europe (e.g. Olesen and Bindi 2002). Climatic and soil conditions of the great European plain extending from south-east England through France, Benelux, Germany, Poland, Hungary, Ukraine and Belarus to Russia provide the most productive conditions in Europe. Agricultural policies and socio-economic conditions have hampered production in eastern Europe, however. In northern Europe, agriculture is mainly limited by climatic and soil conditions. Consequently, less than 10 % of land is cultivated in the Nordic countries (Olesen and Bindi 2002). It is noteworthy, however, that agriculture extends exceptionally far north in the Nordic countries because the Gulf Stream comparable climatic zones are present at higher latitudes in western Europe than in North America (Saikkonen et al. 2012). Seasonal variation in day length, length of the growing season, late spring and early autumn frosts and cold winters are the main climatic constraints on agriculture in the northern Baltic Sea area.

21.3.2 Agricultural Management in the Baltic Sea Basin

Agriculture in the Baltic Sea basin is characterised by different types of land use in the countries surrounding the Baltic Sea, with Germany, Poland and Denmark having the highest proportion of utilised agricultural area (UAA) (see Table 21.3). In all countries, especially Denmark and Finland, the share of UAA that is cropping land (used mainly for the production of cereals) is higher than that of grassland. The composition of land use is crucial regarding the likely effect of climate change and the management implications (see also Chaps. 17 and 25).

21.3.3 Impacts on Agricultural Production

In general, assuming no adaptation, Alcamo et al. (2007) saw the likely effect of climate change on agriculture in northern Europe as positive, both for summer and winter crops. A positive effect of climate change is also expected for the cultivation of bioenergy crops. For livestock, the effect may be either positive or negative. The influence of climate change on crop yield depends on crop species and regional characteristics. Thus, effects must be considered separately for regions, winter and summer crops, and particular crop species. Supit et al. (2012) evaluated the effect of climate change on crop yields without regard to adaptation strategies and found that winter wheat and all other winter and autumn crops may show increased yield under higher temperatures and elevated CO_2 levels until 2050. Crops planted in summer, especially maize and other C4 plants, may also show increased yield as water efficiency is improved by higher CO_2 concentrations. In contrast, Ewert et al. (2005) indicated that in some regions, technological improvements may have greater effects on crop yield than climate change and CO_2 increase.

Table 21.3 Utilised agricultural area (UAA) in 2010 in Baltic Sea coastal states of the EU (Eurostat 2011)

Country	Total land area (1000 ha)	UAA (1000 ha and as % of total land area)	Grassland of UAA (%)	Cereal land of UAA (%)
Denmark	4242	2676 (63)	8	55
Estonia	4239	948 (22)	31	39
Finland	30,408	2291 (8)	1	29
Germany	34,877	16,704 (48)	28	30
Latvia	6229	1805 (29)	35	38
Lithuania	6268	2772 (44)	22	54
Poland	30,633	15,709 (51)	20	44
Sweden	41,031	3073 (7)	15	31

21.3.4 Management Implication for Agricultural Production in the Baltic Sea Basin

Future climate change, especially increased climate variability, poses challenges for agricultural management in the Baltic Sea basin. Adaptation strategies, such as the adoption of agro-ecological techniques, diversified production to increase crop resilience, improvements in crop water-use efficiency and promotion of drought and flood insurance, are options to consider (Trnka et al. 2011). More frequent summer droughts may necessitate additional irrigation in order to take advantage of the potential for increased crop yields created by higher temperatures and increased CO_2 concentration (Supit et al. 2012). Additional adaptation strategies include soil conservation, changes in the timing of sowing and crop rotation, cultivation of different species or different crops and changes in the use of fertilisers and pesticides. Research in plant breeding for increased heat and drought tolerance is another option (Schaller and Weigel 2007).

21.4 Conclusion

This chapter reviews the changes in growing conditions likely to result from climate change in the Baltic Sea basin and related consequences for forest management and agricultural production. Effects differ with location, with growing conditions tending to improve in the northern boreal zone, with reduced precipitation and higher temperatures tending to result in deteriorating growing conditions in the southern temperate zone. Changing growing conditions are likely to cause shifts in forest structure and diversity. The importance of adapting management practices to altered conditions is clear and may allow increased yields and economic benefits as well as climate mitigation through substitution of fossil fuel energy with bioenergy. Evidence suggests that this is particularly the case for the northern parts of the Baltic Sea basin, while in the south, the potential for improved growing conditions might be counteracted by water stress and reduced growth in sensitive species, such as Norway spruce. The need for management adaptation is especially clear in the south, in terms of change in thinning regimes, rotation periods and species selection.

Conclusions on socio-economic impacts cannot be generalised, as potential yield increases as well as loss risks from more unfavourable conditions must be considered. On the other hand, investment in better transport infrastructure in the north and the higher risk of storm damage, with market distortion, risk of species die back, and more frequent bark beetle damage necessitating costly salvage cuttings, would be a considerable burden to forest management, increasing the need for planting where natural regeneration of current species is no longer suitable. A general decrease in tree age at harvesting may also decrease risk, irrespective of whether clear-cutting or selective cutting is practised.

Overall, the results highlight the importance of adaptive forest management strategies in the Baltic Sea basin and show positive benefits for forest management and conserving biodiversity. This could be of particular importance as management practises become more intensive, increasing the need to consider other aims (such as biodiversity and carbon mitigation.).

This chapter also outlines the likely effects of climate change on agricultural production in the Baltic Sea basin. It is clear from several studies that climate change is likely to have mainly positive effects on crop yield, especially for winter crops. However, increasing climate variability will lead to a need for adaptation measures. These are manifold and will differ among region and crop species.

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Socio-economic Impacts—Urban Complexes

22

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Abstract

This chapter examines observed and potential future climate change impacts on socio-economic fields concerning urban complexes in the Baltic Sea basin. This is based on the literature review that focused mainly on English publications on climate change impacts, but included some publications in other languages on adaptation. In the Baltic Sea basin, there appears to be an imbalance between cities and towns that have been well studied with reference to climate change impacts, and cities or even regions for which there is hardly any published literature. For those publications that do exist, most concern the impact of a specific climate change effect (temperature rise, extreme events, sea-level rise) on a particular socio-economic field of an urban complex. The results of the literature review indicate that urban complexes in the Baltic Sea catchment are likely to experience climate change impacts within wide-ranging contexts: from urban services and technical infrastructure, to buildings and settlement structures and to the urban economy or population. Impacts will differ depending on the location of the urban complex: northern versus southern and coastal versus inland.

22.1 Introduction

This chapter examines observed and potential future climate change impacts on socio-economic fields concerning urban complexes. Urban complexes are human-dominated settlements with relatively higher population density than rural settlements. The term comprises cities and towns. Urban complexes are further characterised by high concentrations of buildings and built-up areas with consequent soil sealing, high concentrations of people and infrastructure as well as specific economic and cultural roles and activities. These factors render urban complexes particularly vulnerable to climate change impacts (Hunt and Watkiss 2011).

As every urban complex is characterised by a specific mix of social, ecological and economic interdependencies and its own settlement and building structure, it is difficult to generalise on scientific findings concerning urban complexes. Moreover, in the Baltic Sea basin, there seems to be an imbalance between cities and towns that have been well studied with reference to climate change impacts, and cities or even regions for which there is hardly any published literature. For those publications that do exist, most concern the impact of a specific climate change effect (temperature rise, extreme events, sea-level rise) on a particular socio-economic field of an urban complex. Systematic case studies are available but are mostly on the impacts of climate on human health (Analitis et al. 2008; Hajat and Kosatky 2009; Michelozzi et al. 2009; Rocklöv and Forsberg 2010; Baccini et al. 2011) and do not address the impact of sea-level rise on the various regions and cities of the Baltic Sea basin

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_22

(Schmidt-Thomé 2006), or are global studies that happen to include cities from the Baltic Sea region (Nicholls et al. 2007). The impacts of climate change on the different socio-economic elements of urban complexes were mainly assessed through studies using global and regional climate models driven by the Intergovernmental Panel on Climate Change (IPCC) SRES scenarios. These scenarios were mainly combined with specific impact models. Only few studies drew conclusions from qualitative statements about how the future climate could evolve.

Climate change impacts are determined not only by specific features of the urban complexes themselves, but also by the general vulnerability of urban society, its socio-economic and institutional structure as well as infrastructure and its capacity to cope with impacts (IPCC 2007). A key challenge in the urban context is to identify the main future climate risks and vulnerabilities, both physical and social (Hallegatte et al. 2011b) and to develop and implement adaptation measures. The IPCC defines adaptation as 'adjustment in natural or human systems in response to actual or expected climatic stimuli or their effects, which moderates harm or exploits beneficial opportunities' (IPCC 2007). Thus, adaptation is a response strategy to climate change that involves a reduction in vulnerabilities and avoids potential damage, but also takes advantage of opportunities that arise from climate change (Smit and Pilifosova 2001). Because this chapter focuses on climate change impacts, it mainly assesses literature concerning impacts; less focus is given to adaptation.

Climate change is not the only phenomenon affecting climate in urban areas of the Baltic Sea region; the urban areas themselves influence their climatic conditions. Average temperatures rose due to urbanisation in the twentieth century in Uppsala (Bergström and Moberg 2002), Stockholm (Moberg et al. 2002) and St. Petersburg (Jones and Lister 2002). Furthermore, city growth, together with more heavy rainfall, can increase flood risk in urban areas, as in Helsingborg (Semadeni-Davies et al. 2008a). Urban areas are usually characterised by higher temperatures than the surrounding countryside; this urban heat island effect has been identified for various cities such as Stockholm (Moberg and Bergström 1997; Bolund and Hunhammar 1999; Gustavsson et al. 2001; Moberg et al. 2002), Malmö (Bärring et al. 1985), Gothenburg (Svensson 2002) and Uppsala (Moberg and Bergström 1997). Urban cold islands, where built-up areas were colder than areas outside the city core, have also been observed, for example in Gothenburg (Svensson and Eliasson 2002). These effects depend on urban land use. In Gothenburg, there are differences of up to 6.8 °C between land-use categories (Eliasson and Svensson 2003).

Moreover, climate change is not the only driver of change in urban complexes; they are also affected by demographic change, land-use change (see Chap. 25) and political and economic change, which themselves interact with climate change impacts. This chapter reviews the literature on climate change impacts on urban complexes; non-climatic drivers of change are not considered. Section 22.2 assesses the literature addressing past, current and future climate change impacts on urban complexes. Consideration is given first to the literature concerning urban services and technical infrastructure with an emphasis on wastewater management, drinking water supply systems and transport (Sect. 22.2.1), this is followed by an assessment of the small volume of data on climate change impacts on buildings, housing and settlement structure (Sect. 22.2.2), and then climate change impacts on different sectors of the urban economy and the urban population, with a brief consideration of adaptation measures, acknowledging that this is not the focus of the chapter (Sect. 22.2.3). The final conclusions of the assessment are outlined in Sect. 22.3.

22.2 Past, Current and Future Impacts of Climate Change on Urban Complexes

22.2.1 Urban Services and Technical Infrastructure

Infrastructure is the collective term for systems designed to meet human needs or, in other words, perform a service for the urban population by for instance delivering drinking water, carrying electricity or disposing of wastewater (IPCC 2007). The vulnerability of infrastructure to climate change depends on its state of development, resilience and adaptability (IPCC 2007). The climate change impacts reported to affect technical infrastructure the most (by physical damage) are sea-level rise, extreme events such as storm surges and changing precipitation patterns, particularly flooding caused by more frequent heavy precipitation events. Less research has been undertaken on the impacts of heat and drought in cities in the Baltic Sea basin, possibly due to the lack of risk for urban services and technical infrastructure in this region of moderate climate.

Sea-level rise is expected to be greater in the southern Baltic Sea than in the northern part (see Chaps. 9 and 14), and coastal cities such as Gdansk are expected to be particularly at risk. Dikes, port facilities, industrial areas, warehouses, transportation routes, drainage water systems, sewage plants, ground water recharge areas and energy infrastructure (such as heating pipes and power plants) are vulnerable (Schmidt-Thomé et al. 2006; Staudt et al. 2006; Virkki et al. 2006; Hilpert et al. 2007). Due to isostatic uplift in northern Baltic Sea areas (Chap. 9), net sea-level rise along northern coastlines over the next few decades is expected to be less than along southern coastlines, with less risk for cities such as Stockholm (Meier and Broman 2003; Graham et al. 2006; Viehhauser et al. 2006), Helsinki (Lehtonen and Luoma 2006), Pärnu (Klein and Staudt 2006) and Loviisa (Virkki et al. 2006). The projected increase in heavy precipitation and rapid snow melt events (Chap. 11) could cause surface flooding due to undersized urban drainage and sewage systems such as in Porvoo (Virkki et al. 2006), Loviisa (Virkki et al. 2006), Helsingborg (Semadeni-Davies et al. 2008b), Kalmar (Olsson et al. 2009), Lund (Niemczynowicz 1989), Stockholm (Sverige 2007) and Uppsala (Viehhauser et al. 2006). For Lithuania, as well as snow, frozen ground was considered to affect roads, communication infrastructure and buildings (Taminskas et al. 2005). Less attention has been paid to the impacts of storms and heat on urban infrastructure and services. Much of the literature on climate change impacts on urban infrastructure in the Baltic Sea region concerns coastal cities. This is probably due to concern over the combined threat of climate impacts such as sea-level rise and storm surges.

22.2.1.1 Wastewater Management

In Scandinavia, the projected increase in maximum precipitation would cause an increase in maximum discharge from urban areas (Arnbjerg-Nielsen 2011, see also Chap. 11) and the projected increase in mean temperature would alter the hydrological cycle (Chap. 5) owing to the higher water-carrying capacity of warmer air. Consequently, urban areas, which are characterised by fewer storage elements than rural basins, would respond with further decreases in storage capacity and enhanced run-off (Niemczynowicz 1989). The impacts of changing precipitation patterns on drainage and sewage systems have mainly been assessed by studies using hydrological modelling systems with various assumptions about future precipitation amount (Niemczynowicz 1989; Semadeni-Davies et al. 2008a, b; Nie et al. 2009). The modelling studies usually use climate scenarios based on different global and regional climate model results (Semadeni-Davies et al. 2008a, b; Olsson et al. 2009). Time scales range from single rainfall events to annual precipitation amounts (Niemczynowicz 1989; Semadeni-Davies 2004; Semadeni-Davies et al. 2008a, b; Nie et al. 2009; Olsson et al. 2009). In addition to climate change, other developments such as population growth and ongoing urbanisation can influence urban drainage systems due to increased soil sealing and removal of vegetation, among others (Semadeni-Davies et al. 2008a).

Inflow volumes to drainage and wastewater systems are likely to increase under a warmer climate and to increase the occurrence of surface flooding and overflow of sewage systems and associated environmental problems if current systems remain unchanged (Niemczynowicz 1989; Semadeni-Davies et al. 2008a; Nie et al. 2009; Olsson et al. 2009; Plósz et al. 2009). For Lund, although the storm water system is designed to accommodate further development of the city, increased rainfall intensity of 20–30 % would result in significant flooding problems for the city's sewage network (Niemczynowicz 1989). Semadeni-Davies et al. (2008b) showed that for Helsingborg, the impacts of urbanisation and climate change could be more than met by implementing current developments in urban water management. For coastal cities, high rainfall in combination with sea-level rise can cause further problems, for example sea water inflow into sewer and drainage water networks and into wastewater treatment plants (Lehtonen and Luoma 2006).

22.2.1.2 Drinking Water Supply Systems

Climate change may affect drinking water supply and lead to reductions in river flow, lower groundwater tables and, in coastal areas, to saline intrusion into surface water and groundwater systems. Climate change could also affect the system itself, including damage to pipelines through erosion caused by unusually heavy rainfall (IPCC 2007). Drinking water supply systems are at risk in several urban areas of the Baltic Sea catchment. Research has largely been undertaken for coastal cities only (Klein and Staudt 2006; Lehtonen and Luoma 2006; Schmidt-Thomé et al. 2006; Staudt et al. 2006; Virkki et al. 2006). This is possibly due to the threat of saltwater intrusion (through sea-level rise) into coastal aquifers serving as drinking water reservoirs. The level of risk depends on geographical location and thus on the rate of sea-level rise and distance of the drinking water aquifers from the coast. These risks were assessed using different sea-level rise scenarios (Klein and Staudt 2006; Staudt et al. 2006; Virkki et al. 2006). Studies did not identify risk of salt water intrusion into groundwater reserves important for drinking water supply in Helsinki (Lehtonen and Luoma 2006), but did identify a risk of contamination of drinking water wells in Pärnu (Klein and Staudt 2006).

Water quality and availability can also be affected by flooding, changing precipitation patterns and higher temperatures. Increasing precipitation amounts could be expected to result in a higher rate of groundwater recharge. Yet seasonal conditions can be very different. Less rainfall combined with higher temperatures in summer could affect the quality of drinking water; another risk factor is the impact of flood waters on water quality (Meier et al. 2006;

Ekelund 2007). More frequent heavy rainfall and rapid snow melt events are considered major risk factors for water quality in many Baltic cities due to strong soil erosion and the possibility of contaminants entering drinking water reservoirs (Klein and Staudt 2006; Meier et al. 2006; Schmidt-Thomé 2006).

22.2.1.3 Transport Infrastructure and Services

Transport and its various sub-sectors—road, rail, air and sea —are affected by changes in precipitation, thunderstorms, temperature, winds, visibility and sea-level rise. Within the transport sector, the main impacts of climate change are likely to be felt through extreme events, more so than through a steady rise in temperature (Love et al. 2010). To date, research on the impacts of climate change on transportation has been overshadowed by research on the mitigation of greenhouse gas emissions from this sector. Few studies specifically address impacts of climate change on transport in the urban context within the Baltic Sea catchment. Consequently, the impacts of climate change on transport infrastructure considered here are limited to sea-level rise and flooding caused by heavy precipitation or storm surges. Possible impacts from higher temperatures, such as damage to rail and road surfaces are conceivable, but according to the IPCC (2007), among all possible impacts on transportation, the greatest in terms of cost concerns flooding.

An ongoing rise in sea level would first affect roads and railways situated near the coastline of the southern Baltic Sea where the topography is low and flat, as in Gdansk (Schmidt-Thomé et al. 2006) or Malmö (City of Malmö 2011), but also in Tallin and Pärnu (for an 0.85- to 0.95-m sea-level rise in Pärnu; Kont et al. 2008), although Klein and Staudt (2006) stated that most roads and railways were safe in the 'high case' sea-level rise scenario for 2071–2100 (+1.04 m) in Pärnu (IPCC 2007 projected lower rises, see also Chap. 14). Sea-level rise combined with higher storm surges may cause more serious damage in coastal cities than a gradual rise in sea level alone (Klein and Staudt 2006; Schmidt-Thomé et al. 2006; Staudt et al. 2006). In more northern cities such as Stockholm, where sea-level rise is not considered a major problem, floods caused by heavy rainfall or rapid snow melt can inundate parts of the traffic infrastructure, independent of the rate of sea-level rise (Graham et al. 2006; Meier et al. 2006; Viehhauser et al. 2006; Ekelund 2007). Even for Gdansk, where the rate of sea-level rise is projected to be relatively high compared to more northern coastal cities, infrastructure is more vulnerable in the case of river floods or flash floods as experienced in 2001 when the central railway station and main streets were seriously affected (Staudt et al. 2006).

There may also be benefits associated with higher temperatures. Less salting and gritting would be required, and railway points would be likely to freeze less often (IPCC 2007). This may reduce municipal costs in winter.

22.2.2 Buildings, Housing, Settlement Structure

Extreme weather events associated with climate change pose particular challenges to human settlements because assets and populations are increasingly located in coastal areas, slopes, ravines and other risk-prone regions (IPCC 2007). The current and future location as well as the extent of settlement structures and infrastructure depends on many factors such as land-use planning, policy and jurisdictional decisions, and demographic and socio-economic developments. Such factors affect the scale of climate change impacts on settlement structures. In terms of buildings, housing and settlement structure in cities of the Baltic Sea basin, sea-level rise and changing precipitation patterns are likely to be the most significant impacts of climate change. In combination with potentially more frequent and intense storm surges, several cities are at risk. Some cities are already experiencing storm and flood damage (Virkki et al. 2006); for example, in 2005, a surge during the storm 'Gudrun' flooded densely populated areas of Pärnu and Haapsalu for about 12 h (Tonisson et al. 2008). Housing facilities and residential areas are at risk in cities where buildings are located close to the shore, such as Tallin (Hilpert et al. 2007), Malmö (City of Malmö 2011), Loviisa (Virkki et al. 2006), Gdansk (Staudt et al. 2006) and Aalborg (Hansen 2010). Cities such as Helsinki (Lehtonen and Luoma 2006) and Pärnu (Klein and Staudt 2006) may be only slightly affected by sea-level rise, if at all. In Helsinki, the impact of rising sea level on new housing areas is expected to be low because future sea-level rise is already being factored into the planning processes (Lehtonen and Luoma 2006). In Pärnu, the 'Shores and Banks Protection Act' helps to lower the impact of sea-level rise on housing (Klein and Staudt 2006), because construction closer than 50 m to the shoreline is prohibited in cities (HELCOM 1996). With regard to existing structures, the example of Pärnu demonstrates that the combination of sea-level rise and storm surges will increase vulnerability. If the projected sea-level rise is combined with a projected 100-year flood, 21 % of the area for single family houses and 9 % of the apartment house area of Pärnu would be in the flood affected area (Klein and Staudt 2006). Independent of sea-level rise, heavy rains and rapid snow melt are also expected to put settlements at risk due to the lower water infiltration capacity in urban areas and associated rapid surface run-off and/or overloaded drainage systems for cities such as Odense (Zhou et al. 2012), and as is already the case in Stockholm (Viehhauser et al. 2006).

Climate change could also have some benefits for this sector. For example, lower heating costs for buildings in winter. In Finland, winter heating costs are projected to decrease by 10 % in the period 2021–2050 (Venäläinen et al. 2004) and 20–30 % by the end of the century (Kirkinen et al. 2005). For urban areas, these cost savings could become even higher due to the urban heat island effect.

22.2.3 Socio-economic Structure

22.2.3.1 Impacts of Climate Change on Different Sectors of the Urban Economy

Estimating the impact of climate change on the urban economy is complex, and scientific studies on this subject are few. Hallegatte et al. (2011b) distinguished between direct and indirect impacts; the former related to the change in mean temperature or increases in extreme weather events. Indirect effects, such as disruption of the transport system, have knock-on effects on other economic sectors, often with economic losses. Indirect impacts of climate change are difficult to assess, even when the direct impacts of climate change can be estimated with some level of confidence. Although a systemic approach to assessing the impacts of climate change in cities is advocated, very few examples of this exist (Mechler et al. 2010). See Box 22.1 for a case study on the economic costs of climate change in Copenhagen.

Box 22.1 Case study of Copenhagen

Systematic assessments of the costs of climate change impacts are rare. A study by Hallegatte and co-workers illustrated the methodology for assessing the costs associated with climate change impacts in the urban context (Hallegatte et al. 2011a). The researchers adopted a simplified catastrophe risk assessment to calculate the direct costs of storm surges in Copenhagen, coupling this with an economic model.

The analysis concluded that, at present, Copenhagen is not highly vulnerable to coastal flooding owing to its high level of defence, while in the absence of protection, the losses would increase in the future. Thus, with no protection and a rise in mean sea level of 25 cm, total losses due to a 100-year event could be close to EUR 4 billion. With 100-cm mean sea-level rise, the costs could increase to EUR 8 billion.

Primary Sector

The primary sector of the Baltic Sea area economy is relatively unimportant in the urban context, as little agriculture, mining and forestry takes places in the cities. *Fisheries* have been an important source of economic revenue in the Baltic Sea region, although the importance of this sector has decreased. There are many uncertainties concerning the impact of climate change on fisheries in the Baltic Sea that relate to the physiological, ecological and social response (Mackenzie et al. 2007). *Agriculture* is not a significant source of economic revenue in the urban context but is important in the Baltic Sea region as a whole (see Chap. 21).

Secondary Sector

The secondary sector includes those industries that create a finished or a usable product, such as production or construction.

The *Industrial sector* can be impacted by climate change. A study of regional industries and the economic effects of climate change in the coastal and estuarine zone of north-western Germany found that industries were vulnerable to ongoing climate change impacts (Wang et al. 2010). These can affect regional gross net production through primary and secondary impacts. Extreme weather events and sea-level rise can have severe effects on industrial sites located at the coastline; an example being the former uranium enrichment plant in Sillamäe in Estonia which can pose a threat to society in a storm event (Kont et al. 2003).

The *Energy sector* is expected to be impacted by climate change on both the supply side and the demand side, although it is acknowledged that the impact of climate policy is likely to be greater than changes in mean temperature, for example (Mideksa and Kallbekken 2010). Impacts on the supply side are likely to be regionally specific (Mideksa and Kallbekken 2010). The availability of renewable resources is also affected by climate change, and the trend may be positive or negative. For example, the availability of forest biomass may increase (Lundahl 1995; see also Chaps. 16 and 21). Changes in wind and solar energy are not easy to project due to uncertainties concerning variability in winds and changes in cloud cover (Lundahl 1995, Chap. 11). Vulnerability of electricity companies to climate change impacts varies by country and is determined, among other things, by differences in national regulation (Inderberg and Løchen 2012).

Tertiary Sector

The tertiary sector consists of the service part of the economy.

Tourism is an important sector impacted by climate change. Urban complexes contain either integrated or remote areas for recreational activities as an intrinsic part of their system. Within the Baltic Sea region, the number of overnight stays is clearly linked to urban complexes and regions containing larger urban complexes show a significant increase in the number of overnight stays. Climate change has direct impacts on the physical, environmental and social resources for tourism as well as on the comfort, perception and safety of tourists (Patterson et al. 2006; Moreno and Amelung 2009). This affects the economic value of tourism. Tourism is of major importance for many coastal areas especially in the southern and south-western Baltic Sea area. For Mecklenburg-Vorpommern, for instance, tourism is the most important economic sector (StatA-MV 2010). As southern and south-western Baltic Sea regions dominate this sector, climate change impacts and the tourism-relevant characteristics such as sandy beaches are best studied in these regions. The literature mentions the following impacts in relation to the tourism sector:

- Beach wrack: organic material such as kelp and sea grass left on beaches by surf, tides and wind impairs the quality of tourist beaches. Beach cleaning is common but expensive (Dugan et al. 2003; Fanini et al. 2005; Davenport and Davenport 2006). Changes in water temperature may increase the occurrence of such material on beaches (Björk et al. 2008), but this has not yet been shown for the Baltic Sea.
- Demand: some aspects of future climate change within the Baltic Sea region (Matzarakis and Amelung 2008) might increase visitor numbers to Baltic Sea resorts. Demographic change may counteract this development but may also lead to more elderly and more climate-sensitive visitors (Coombes et al. 2009).
- Erosion: relative sea-level rise together with coastal abrasion and accumulation processes could lead to changes in the Baltic Sea coastlines (BACC Author Team 2008; Harff and Meier 2011; Chap. 20). These changes would differ from region to region (influenced by vertical crustal movement, sediments, coastal typology, exposure and protection status, for example) and may decrease the attractiveness and tourism capacity of coastal areas and/or increase costs for coastal protection measures.
- Changes in the terrestrial cryosphere (see Chap. 6) in terms of extreme winter conditions (both extremely high and extremely low temperatures, strong winds, rain or severe snowfall) would affect winter recreation and winter sports and are a problem for tourism operators in the northern Baltic Sea basin, as it was shown in a Finnish study (Tervo 2008).

The relation of tourism and recreational activities to climate change shows a broad variety of direct (e.g. sunbathing) and indirect links (e.g. habitat change for ecotourism). To date, there has been little research on this and especially on threats and opportunities for tourism due to climate change at the regional level (Moreno and Amelung 2009), including for the Baltic Sea region. As a result, quantitative statements concerning the effects of climate change on further development of the tourism sector are not available. In general, it is considered likely that the impacts of climate change for the tourism industry will be largely positive in the Baltic Sea region although the costs of maintaining the attractiveness of this region as a tourist destination may increase in the southern and south-western coastal areas (e.g. costs for beach cleaning, coastal protection and beach nourishment).

Private tourism operators have short planning horizons of 1–5 years. Adaptation to long-term processes such as climate change is therefore rarely discussed. At the same time, tourism plays a major role in the spatial development of many Baltic Sea regions, which have planning horizons of 10–30 years. According to Schumacher and Stybel (2009), tourism adaptation strategies are discussed at senior policy level, but there are few examples of adaptation strategies in practice. The wide range of possible adaptation measures (Schumacher and Stybel 2009) necessitates cooperation with adjacent sectors such as urban and regional planning, coastal protection, nature protection, health, water management and forestry.

Transport and its various sub-sectors of road, rail, air and sea play an important role in all economic sectors and are all affected by climate change. The costs of climate change on the transport sector are uncertain and difficult to calculate (Koetse and Rietveld 2009). There are likely to be changes in passenger and freight transport, as well as damage to existing transport infrastructure causing congestion in urban complexes. For example, a heat wave may have costly impacts on rail systems. The costs of a heat wave in London and surrounding areas in 2003 were estimated at GB 750,000 (Love et al. 2010). Maritime transport within the Baltic Sea would also be affected. Coastal transportation infrastructure is vulnerable to sea-level rise and more frequent storm surges. Disruption in supply chains can have a devastating impact on the regional economy. For example, it is estimated that Hurricane Katrina caused damage estimated at USD 1.7 billion and affected business in over 30 US states (Becker et al. 2012). In a global review of sea ports, port authorities stated that sea-level rise is likely to have negative impacts on their port that systematic adaptation is not taking place and that more research is needed (Becker et al. 2012).

The *insurance sector* is the part of the *financial sector* that is likely to be most important in terms of climate change as it can be a means to limit damage and spread risk. The insurance sector can help society to reduce the costs of adaptation by risk sharing, but climate change can also affect the insurance sector itself, causing the sector to consider heightened risk in setting premiums and risk management (Mills 2005). Insurers must therefore set premiums in a way that is financially viable by attempting to predict the frequency and severity of insured losses based on expected, rather than historical risk (Gasper et al. 2011). Insurance practices vary by country. Within the Baltic Sea region, flood damage is privately insurable in Germany, while there is no insurance for storm surges (Botzen and van den Bergh

2008). Although private flood insurance is common, most of the flood risk is carried by the government. In Germany, the government paid out EUR 9.1 billion for flood damage in 2002, while the private sector paid out EUR 1.8 billion (Botzen and van den Bergh 2008).

Retail and commercial services are likely to be affected by climate change through effects on the efficiency of the supply chain and distribution network, as well as on the health and comfort of the workforce (Gasper et al. 2011). To date, case studies have focused on effects on the transportation network with little research on the retail or consumer sectors (Gasper et al. 2011).

22.2.3.2 Impacts of Climate Change on Urban Population

The urban population is vulnerable to the impacts of climate change around the Baltic Sea. This section reviews the current literature, with a focus on human health and the well-being of different groups of society. Vulnerability differs between these groups, based on gender, age and race (Gasper et al. 2011). For example, children and the elderly can be more vulnerable to natural hazards. Health is an important factor of social well-being, and climate change can have both immediate and lasting impacts on the urban population with the main stressors being severe weather events and extreme heat and disease; for example an increasing incidence of tickborne encephalitis in endemic regions (Lindgren 1998; Gasper et al. 2011; Jaenson et al. 2012).

Thermal Stress

Even if cold stress seems more likely for northern countries such as Sweden (Svensson et al. 2003), under a changing climate, heat stress and the demand for air conditioning in houses, health-care institutions, schools and work places may increase in summer (Svensson et al. 2003; Rocklöv et al. 2009; Baccini et al. 2011; Thorsson et al. 2011). In general, the rate of climate change projected for the coming decades under most scenarios makes future acclimatisation uncertain and society may become more vulnerable due to increasing urbanisation and an ageing population (Hajat and Kosatky 2009).

Baccini et al. (2011) analysed the impact of heat on mortality in 15 European cities including Helsinki and Stockholm. They concluded that high summer temperatures affected European population health and that this impact is likely to increase in the future due to the projected increase in ambient temperature and the frequency, intensity and duration of heat waves. Several studies in the Baltic Sea catchment examined the effects of heat on human health and well-being. These include the impacts of moderate changes in ambient temperature on human health in Copenhagen (Wichmann et al. 2011), the relationship between high temperature and hospitalisation for cardiovascular and respiratory disease in the elderly population of twelve European cities including Stockholm (Michelozzi et al. 2009), the effect of heat on mortality in Stockholm and the Stockholm area (Rocklöv et al. 2009; City of Malmö 2011) and the link between the intensification of thermal and humidity conditions with unfavourable bioclimatic effects including heat stroke on the urban population of Cracow (Piotrowicz 2009). Seasonal differences were examined for Cracow (Piotrowicz 2009) and Gothenburg (Thorsson et al. 2011). Results indicated more problems with excessive temperature in summer and different outcomes for other seasons. A future decrease in temperature stress and better outdoor thermal comfort in winter, spring and autumn was announced for Gothenburg (Thorsson et al. 2011). A study in Cracow found that more frequent mild winters could cause the human body to lose the ability to adapt in colder winters (Piotrowicz 2009).

The *costs of heat events* can be significant. For example, preventing the loss of 18,000 lives in the 2003 European heat wave could have rendered benefits of up to USD 72 billion (Halsnæs et al. 2007). The heat waves across Europe in 2003 and 2010 indicate the vulnerability of urban populations. This is despite the high levels of development in Europe (Lass et al. 2011).

Identifying *social vulnerability* to heat waves is important. Urban populations are generally more vulnerable to heat waves than rural populations. Within the urban population, particularly vulnerable groups include elderly people and those suffering from cardiovascular and respiratory diseases (Piotrowicz 2009). A study on the effects of temperature on the elderly in Sweden confirmed the impact of heat on mortality (Rocklöv and Forsberg 2010). The study also showed that the effects of high relative humidity and high temperature were greatest in the most densely populated area, Stockholm (Rocklöv and Forsberg 2010). In circumpolar regions, the populations at greatest risk from the adverse health effects of cold are children, elderly people or people suffering from cardiovascular or respiratory diseases. Factors such as ageing and urbanisation may contribute to the rise in cold-induced health problems (Mäkinen 2007). Svensson et al. (2003), referring to Gothenburg, argued that climate at high latitudes presents a number of particular bioclimatic problems, both physiological and psychological, and cold stress is perhaps more important, since people in Scandinavia seldom complain that conditions are too hot.

Changes in Air Quality

A decrease in air quality is a health-related effect of climate change in cities. Cities often have higher concentrations of air pollutants than surrounding rural areas (Eliasson and Holmer 1990; Jacob and Winner 2009). During 2005–2007 in Sczcecin, thermal stress and overheating, together with high levels of air pollutants, accounted for a significant and direct threat to human health, especially for people suffering from circulatory diseases and problems with blood pressure (Czarnecka et al. 2011). Climate change may also have positive effects. For example, Tang et al. (2011) reported a trend towards lower average nitrogen dioxide concentrations in Gothenburg and concluded that changes in weather and climate had contributed to this reduction. See Chap. 15 for information on changes in air quality.

Storms and Floods

Storms can affect the urban population through property damage, through injury and loss of life, and may leave sections of the population homeless (Gasper et al. 2011). Early studies suggested that wind damage and related economic losses in Europe over the coming decades are likely to be caused by rare events rather than interannual variability (Schwierz et al. 2010). Insurance can act as a buffer to reduce losses. Insured storm-related losses depend on the frequency, nature and dynamics of storms, the vulnerability of values at risk, the geographical distribution of these values and the conditions of the risk transfer (Schwierz et al. 2010).

Flood events also pose a threat to urban populations. A social vulnerability index of river floods in Germany demonstrated that irrespective of climate change, the vulnerable populations currently include the elderly, the financially weak and urban populations (Fekete 2009). A study of two cities in Denmark concluded that increasing the urban drainage infrastructure was financially beneficial but that larger adaptation measures were not yet financially viable (Arnbjerg-Nielsen and Fleischer 2009).

22.2.3.3 Adaptation to Climate Change in the Urban Context

Many cities in the Baltic Sea region, such as Stockholm (Deppisch and Albers 2012) and Malmö (City of Malmö 2011), have identified particular climate change impacts and are now in the process of developing adaptation strategies, as for Rostock (Deppisch and Albers 2012).

Adaptation policy affecting urban areas around the Baltic Sea has been developed at the EU, national and regional level. An EU White Paper on adaptation was published in 2008 with the aim of developing a knowledge base for adaptation (Commission of the European Communities 2009). With the exception of Lithuania, Poland and Estonia, countries around the Baltic Sea have all published a national adaptation strategy. The extent to which national strategies affect urban areas depends on the legislation and planning systems of the individual countries.

Regional or local adaptation strategies are most important for planning adaptation at the city level. The regional level is important because this is the level at which policy is developed to regulate issues related to the built environment, buildings and the maintenance of infrastructure, especially in terms of drainage and piped water and the provision of services, such as fire protection, public transport and disaster response (Gagnon-Lebrun and Agrawala 2006).

Many Baltic Sea regions have developed adaptation strategies, but the processes involved can be hindered by uncertainties in the timing of climate change impacts as well as by a lack of knowledge and expertise (Ribeiro et al. 2009; Dannevig et al. 2012; Hedensted Lund et al. 2012; Jonsson et al. 2012; Juhola et al. 2012a; Nilsson et al. 2012). Further barriers include social and cultural inertia on individual and collective action and low priority of adaptation in relation to other policy agendas (Carter 2011) such as financial, economic and social aspects, as well as climate change mitigation (Deppisch and Albers 2012). In addition, weak links across policy and administrative levels and lack of a coherent policy agenda can also slow the adaptation process, as in Finland (Juhola and Westerhoff 2011) and Sweden (Storbjörk and Hedrén 2011), see also Annex 2.

Adaptive capacity plays an important role and is a prerequisite for a region's ability to adapt to climate change (Brooks et al. 2005). Adaptive capacity—the ability or potential of a system to respond successfully to climate variability and change—includes adjustments in behaviour, resources and technologies (IPCC 2007). This varies greatly at the regional level within the Nordic countries (Juhola et al. 2012b). Knowledge facilitated by links with universities and research projects funded by the EU plays an important part in enabling adaptation at the sub-national level, for example in Germany (Frommer 2001), Finland (Westerhoff and Juhola 2010) and Sweden (Deppisch et al. 2011). Although this has enabled some regions to pursue adaptation strategies, the uncertainty of climate change impacts has slowed overall action on adaptation (Storbjörk 2007; Carter 2011). Tools such as participatory vulnerability assessments and scenario workshops have been useful in communicating adaptation challenges to local stakeholders (Baard et al. 2012; Jonsson et al. 2012).

Examples of city level strategies in the Baltic Sea area are summarised in Table 22.1. A review of adaptation strategies at the city level found that climate change is likely to exacerbate existing urban problems as well as to create new ones (Committee of the Regions 2011). The report made several recommendations: (1) urban adaptation strategies must be developed and build on existing sectoral and cross-sectoral agendas; (2) urban adaptation requires new governance structures and innovative new solutions with the help of a wide range of stakeholders; and (3) no single type of measure is able to eliminate vulnerability to climate change, but a portfolio approach is likely to be most effective (Committee of the Regions 2011). Finally, the report emphasised that that approach should be staggered and iterative to achieve progress in the short term.

There are many gaps in research concerning adaptation in urban complexes. These include how to better understand

Table 22.1 Examples of city adaptation strategies in the Baltic Sea region (City of Copenhagen 2011; Committee of the Regions 2011; Albers et al. 2013)

City	Details
Copenhagen	Main climate change impacts and consequences: sea-level rise, intense precipitation, severe drainage problems and flash flooding. The city adaptation strategy is comprehensive and cross-sectoral. Key measures include expanding the sewer grid, establishing a sustainable drainage system, reservoirs to store rain and wastewater and green roofing. Copenhagen developed its climate adaptation plan in 2011
Helsinki	Main climate change impacts and consequences: sea-level rise, intense precipitation, drainage problems and flash flooding, wind and storm damage. The adaptation strategy for the metropolitan area of Helsinki is comprehensive and cross-sectoral. The strategy was approved in 2012
Stockholm	Main climate change impacts and consequences: river floods, intense precipitation, drainage problems and flash floods, drought and potential drinking water problems, higher temperature, heat waves and wind and storm damage. The city adaptation strategy, which goes further than an existing policy paper from 2007, will be comprehensive and cross-sectoral. Key measures are not yet agreed and the strategy is still pending (as of 2013)

Table 22.2 Expected impacts on urban complexes

Climate change effects	Northern Baltic Sea region		Southern Baltic Sea region	
	Coast	Inland	Coast	Inland
Ongoing change				
Temperature rise	Drinking water quality and availability affected			
	Less salting and gritting necessary and railway points will freeze less often			
	Lower costs for heating buildings			
	Greater heat stress and demand for cooling in summer			
	Affected concentration of air pollutants; may be positive or negative for population			
Temperature rise			Positive development of tourism sector but increasing costs for coastal protection	
Precipitation patterns	Drinking water quality and availability affected			
Sea-level rise	Lower net sea-level rise with resulting minor challenges		Increased vulnerability of main infrastructure	
			Further problems in combination with high rainfalls (e.g. sea water inflow into sewer and drainage water networks)	
			Drinking water quality and availability affected by saline intrusion in freshwater reservoirs	
			Roads, railways, housing facilities, residential and industrial areas near shoreline may be affected, with economic consequences	
Extreme events				
Heavy precipitation	Surface floods due to undersized urban drainage and overflow of sewage systems			
	Drinking water quality and availability affected by flooding			
	Threats to settlements due to missing water infiltration capacity of the urban ground			
Heavy precipitation	Inundation of traffic infrastructure		Inundation of traffic infrastructure	
Heat waves	Specific vulnerable groups in urban complexes, especially in bigger and more densely populated cities			
	Effects more severe in the southern Baltic Sea region			
Storm surges	Threats to industrial sites		Serious damage of transport infrastructure, especially in combination with higher sea levels, with economic consequences	
			Threats to infrastructure, buildings and life	

the governance frameworks and stakeholder networks to implement adaptation measures as well as addressing barriers to adaptation, while keeping in mind the possibility of maladaptations (Carter 2011).

22.3 Conclusion

Urban complexes in the Baltic Sea catchment are likely to experience climate change impacts within wide-ranging contexts: to urban services and technical infrastructure, to buildings and settlement structures and to the urban economy or population. Impacts will differ depending on the location of the urban complex: northern versus southern and coastal versus inland. Table 22.2 summarises the expected impacts on urban complexes.

It is difficult to draw general conclusions from the available literature. This is for two main reasons. First, in the Baltic Sea region, there is a lack of studies on specific topics that cover a number of urban complexes. Second, because every urban complex is a unique mix of infrastructure and urban services, inhabitants, natural resources and green spaces, built structures, location and economic and societal factors, it is difficult to generalise on the potential extent of climate change impacts from single case studies. Further generalisation would need an analytical reference framework that makes it possible to categorise the urban complexes under study.

It would be very useful to develop a typology for cities in the Baltic Sea basin that identifies the main potential climate change threats for each. This would imply cross-city studies focusing on the most comparable cases or on the most different cases, using typologies of cities built in line with specific characteristics such as location, size or socio-economic character. Such cross-city studies might have the goal of identifying the specific characteristics that render them more or less vulnerable to climate change impacts in the Baltic Sea basin.

It would be good for any follow-up assessment of climate change impacts on urban complexes to include aspects of urban ecology, as this is a very specific habitat. Also, food supply for urban complexes could be integrated into the socio-economic impact studies, as urban food systems, are likely to be affected by climate change, not just in relation to primary production (Tirado et al. 2010). For example, impacts on food processing, transport and trading are largely unknown, as are issues related to food security (Miraglia et al. 2009).

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Part VI

Drivers of Regional Climate Change

Introduction

Variability and change in the climate system is partly due to random fluctuations in the climate system ('internal variability') and partly to the response of the climate system to changes in conditions external to the climate system (such as solar radiation reaching the Earth and human activities that alter atmospheric composition and land surface properties). The observed records thus always contain an imprint of the response of the climate system to changes in external forcings (the signal) superimposed with variability (the noise). Chapters 23–25 aim at identifying or detecting the signal and attributing it to specific causes.

Formal detection and attribution analyses provide the foundation for statements such as this by the Intergovernmental Panel on Climate Change in 2007: *Most of the observed increase in global average temperatures since the mid-twentieth century is very likely due to the observed increase in anthropogenic greenhouse gas concentrations*, because the observed global warming is very unlikely to be due to natural variability alone (detection) and no other known and physically plausible external forcing mechanism than human influence can explain the observed global warming (attribution). Successful attribution works by exclusion or falsification of alternative hypotheses rather than verification of the forcing mechanism under investigation. Therefore, attribution is conditional on the current state of knowledge and must be revisited as new information arises. This is because new insights may change the balance of evidence.

The following chapters discuss to what extent the building blocks for a formal attribution analysis are available to attribute recent regional climate change to human influence and other causes. Particular focus is on the external forcing mechanisms that have been identified to cause recent global warming (mainly anthropogenic greenhouse gas emissions, Chap. 23), natural and anthropogenic emissions of aerosols (Chap. 24), and changes in land use and land cover (Chap. 25).

23 Regional Evidence of Global Warming

Jonas Bhend

Abstract

This chapter assesses to what extent the factors causing global warming affect the Baltic Sea area. Summertime near-surface warming in northern Europe exceeds natural internal variability of the climate system, and the observed warming cannot be explained without human influence. Regional changes in extreme temperatures, growing-season length and timing of the onset of spring are consistent with the large-scale signal of a human influence (mainly greenhouse gases). Shifts in large-scale circulation in the Northern Hemisphere and precipitation changes in northern Europe and the Arctic have been detected to exceed natural internal variability, but the climate models used to assess these quantities seem to underestimate the observed changes. To what extent this discrepancy between simulated and observed changes also affects the attribution of regional warming to human influence is still a matter of debate. Other aspects of regional climate change including changes in storminess, snow properties, run-off and the changing physical properties of the Baltic Sea have not been formally attributed to human influence yet.

Keywords

Regional climate change • Detection and attribution

23.1 Introduction

This chapter assesses how the factors causing global warming (mainly anthropogenic greenhouse gases) affect climate in the Baltic Sea area. In contrast to the following chapters on the effect of anthropogenic aerosols (Chap. 24) and land-use and land-cover changes (Chap. 25), this chapter focuses on globally uniform or at least large-scale forcing including changes in greenhouse gases, solar irradiance and stratospheric volcanic aerosols.

To demonstrate an external influence on the observed climate change, the concept of detection and attribution is often used. Formal detection and attribution imply (i) the demonstration that the recent observed change is different from natural internal variability—the detection—and (ii) the comparison of different combinations of external forcing and assessment of their relative contribution in explaining the detected change—the attribution (see also Annex 1). Such a framework has been successfully applied at the global and continental scale to detect and attribute anthropogenic near-surface and upper-level warming, as well as large-scale changes in other climatic parameters (Hegerl et al. 2007a). At the regional scale, however, there are only very few formal detection and attribution studies available (see Stott et al. 2010 for a review of recent advances).

Climate change detection and attribution at the regional scale is complicated by various factors. First, variability increases with decreasing area of aggregation, that is the influence of small-scale phenomena does not average out. This generally leads to a decrease in the signal-to-noise ratio of externally forced changes and thus reduces the detectability of regional climate change (Stott 2003; Zwiers and Zhang 2003). Second, model biases play a more important

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_23

role at the regional scale as model performance depends on the spatial resolution of quantities under analysis (Masson and Knutti 2011). This influences our ability to detect and attribute changes at the regional scale as the simulations of recent change and variability deteriorate at smaller spatial scales. Third, forcing mechanisms that are less well understood such as changes in anthropogenic aerosols or that are thought to have a negligible effect on recent global warming such as land-cover changes may be relatively more important at the regional scale (Stott et al. 2010, see Chaps. 24 and 25 for further discussion). In addition, regional detection and attribution in northern Europe seems to be especially difficult compared with other subcontinental regions worldwide. Owing to the position of northern European landmasses at the end of the North Atlantic storm track and due to the complex land–sea distribution, interannual variability in most climatic parameters is very strong, thus masking external influences (see Fig. 9.12 in Hegerl et al. 2007a). In these cases, all that is possible is to assess whether the simulated response to external forcing is consistent with the observed change.

23.2 Causes of Change in Temperature

23.2.1 Mean Near-Surface Temperature

A cascade of evidence from global to subcontinental scales illustrates the anthropogenic influence on recent observed warming (see also Chap. 4, Sect. 4.4). The human influence on global warming (Hegerl et al. 1997), continental warming in Europe (Stott 2003; Christidis et al. 2010a), northern Europe in all seasons combined (Bhend 2010) and warming in northern Europe in summer (Jones et al. 2008) has been successfully detected. Jones et al. (2008) further concluded that anthropogenic warming has already increased the likelihood of occurrence for very warm summers in northern Europe. Using occurrence probabilities as the detection variable, Stott et al. (2011) found mixed evidence of an anthropogenic effect on the frequency of very warm seasons with a detectable anthropogenic influence in spring (MAM) and autumn (SON) and no detectable natural influence. In summer (JJA) and winter (DJF), the detectability of external influences depends on the climate model used.

Zorita et al. (2008) used an alternative approach to the detection problem by computing the likelihood of clusters of record warm years in a stationary climate. Depending on the model for long-term memory of regional temperature time series, they found very low probability of the observed cluster of record-breaking warm years in northern Europe during the past decades. Their approach, however, does not explicitly address potential causes of the warming needed to explain the recent cluster of record-breaking years.

The influence of external forcing is more difficult to detect at smaller spatial scales (Stott and Tett 1998; Zwiers and Zhang 2003). At the grid-box scale of global climate models (approx. 300 × 300 km), the observed warming up to 2002 in the Baltic Sea region is not significantly different from changes due to internal variability alone, and therefore, an anthropogenic influence is not detectable (Karoly and Wu 2005). However, the simulated warming in the Baltic Sea area is consistent with the observed warming when anthropogenic forcing (changes in greenhouse gas and sulphate concentrations) is included in the simulations with three different global climate models (the GFDL R30, HadCM2 and PCM models, Karoly and Wu 2005). The observed warming is also found to be consistent with anthropogenic signals derived from simulations with a coupled regional atmosphere–ocean model (Bhend and von Storch 2009), whereas van Oldenborgh et al. (2009) found significant differences between observed and simulated warming in spring (MAM) using global climate models. They identified the misrepresentation in circulation and snow cover changes as the main reason for the underestimation of warming in spring in global climate models. A recent study by Flanner et al. (2009) suggested that the underestimation of springtime warming at mid-latitudes is due to the lack of carbonaceous aerosols in climate model simulations. These particles darken the snow surface and thus increase the albedo, leading to a warming especially in spring when the snow surface is exposed to intense solar radiation (see Chap. 24 for discussion of the local effects of aerosols). Whereas this forcing mechanism was not included in previous climate model simulations, most models submitted to the World Climate Research Programmes's CMIP5 (Taylor et al. 2011) database include carbonaceous aerosols, which implies that the role of carbonaceous aerosols may be addressed in future attribution studies.

In addition to analyses using local information only (e.g. observed and simulated data for the Baltic Sea area), evidence for a regional anthropogenic warming is also found in formal detection and attribution studies using global constraints (Christidis et al. 2010b). Pooling climate data information across the globe helps to significantly increase the signal-to-noise ratio of externally forced changes. The additional information available when carrying out analyses simultaneously for multiple regions further helps to distinguish responses to different forcings that may be indistinguishable in individual regions. Using such an approach, Christidis et al. (2010b) detected and attributed human influence on the observed warming in northern Europe from 1950 to 1997 (see Fig. 23.1). The authors also computed the fraction of attributable risk (FAR, Stott et al. 2004) of the observed warming in northern Europe. FAR is a measure of the change in likelihood of a given warming in a world with human influence compared to the likelihood of such a

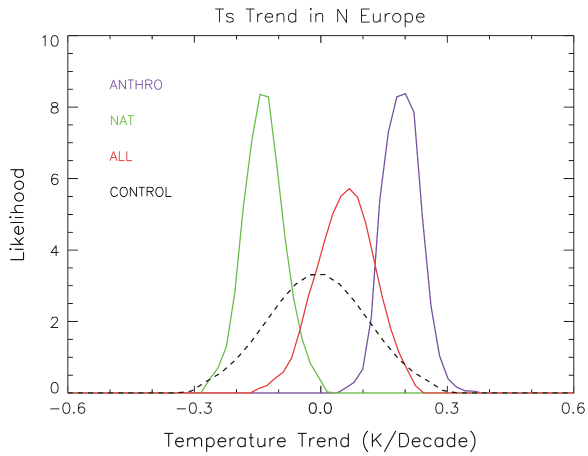

Fig. 23.1 Distribution of attributable trends in near-surface temperature in northern Europe (NEU) from 1950 to 1997 based on optimal fingerprint analysis using global constraints (reproduced from Christidis et al. 2010b). Shown are the trends due to anthropogenic forcing (ANT, *purple line*), natural forcing (NAT, *green line*) and all forcings (ALL, *red line*). The *black dashed line* denotes the distribution of trends due to internal variability as estimated from control simulations (CONTROL). The observed trend of 0.056 °C per decade (not shown) is consistent with the all-forcing signal (ALL, *red line*)

warming in a hypothetical world free of anthropogenic influence. The authors concluded that the likelihood for a warming from 1950 to 1997 in northern Europe more than doubled with a central estimate of a fivefold increase due to human influence. In a world without human influence, a cooling of −0.15 °C per decade could be expected in northern Europe mainly due to the cooling effect of volcanic eruptions during that time.

Min and Hense (2007a) investigated evidence for competing forcing hypotheses in a Bayesian framework. The authors concluded that the combination of anthropogenic and natural forcing better explains the observed change in annual and seasonal temperature in Europe from 1900 to 1999 than any of the forcings separately. They found strong evidence for a forced change over natural internal variability alone (detection) in the periods 1900–1999 and 1950–1999 but less so at the beginning of the twentieth century. Treating the detection problem in a Bayesian framework allowed the authors to also quantify the effects of different prior beliefs. These prior beliefs reflect our understanding of the problem before analysing the data. In the analysis of Min and Hense (2007a), only a strong prior belief that recent climate change is due to natural internal variability would favour this hypothesis over others. Compared to other continental-scale regions, the observational evidence of a man-made warming in Europe is less decisive. Min and Hense (2007b) extended the analysis to subcontinental regions and using different models individually as opposed to using the multi-model mean.

Christidis et al. (2007) found a detectable change in growing-season length, the onset of spring and a marginally detectable change in the end of the growing season in autumn for Europe. Moreover, Gillett et al. (2008) found a detectable anthropogenic influence on Arctic warming. Their study region encompasses the very northern part of the Baltic Sea catchment but includes land temperature measurements on all continents and islands north of 65°N.

Another source of evidence for anthropogenic warming in northern Europe stems from palaeorecords. Hegerl et al. (2011) investigated the effect of external forcing on European temperature using reconstructed temperature back to AD 1500. They detected external influences in all seasons, with the response to external forcing explaining about 30 % of the interdecadal variance.

23.2.2 Temperature Extremes

Climate change not only affects the mean climate but also affects all properties of the distribution including the frequency, intensity and spatio-temporal pattern of extreme events. Extreme events are rare, and thus, fewer data are available to make inference about extreme events and their changes. Attributing causes of change in extreme events is thus generally more difficult than for changes in mean climate. Methods involving very large ensembles of climate model simulations are being developed to quantify the human contribution to individual extreme events such as the autumn 2000 floods in the UK (Pall et al. 2011). In contrast, changes in moderately extreme events such as the coldest night or hottest day in any given year can be attributed using standard approaches.

Kiktev et al. (2003) found positive trends in warm nights and frost days for the Baltic Sea area, although significant only for parts of the area (see also Chap. 4, Sect. 4.4). They also found better agreement with simulated trends in those temperature extreme indices when atmosphere-only general circulation model (GCM) simulations include anthropogenic forcing. Morak et al. (2011) got mixed results for detectability of the observed change in warm nights in northern Europe, whereas this change was robustly detectable in other regions. These findings are corroborated by Zwiers et al. (2010) who obtained a detectable anthropogenic signal on changes in waiting times of long return-period extremes (20-year return values in the 1960s) of daily temperature in northern Europe. A combined anthropogenic and natural influence is only detectable for the warmest night per year (see Fig. 23.2). Formal attribution of the changes in temperature extremes to natural and/or anthropogenic causes is not yet fully achieved, but the consistency of observed changes with both all-forcing and anthropogenic-only forcing simulations indicates that changes in extreme events may be attributed to human influence. Zwiers et al. (2010) further estimated the attributable change in waiting times for the

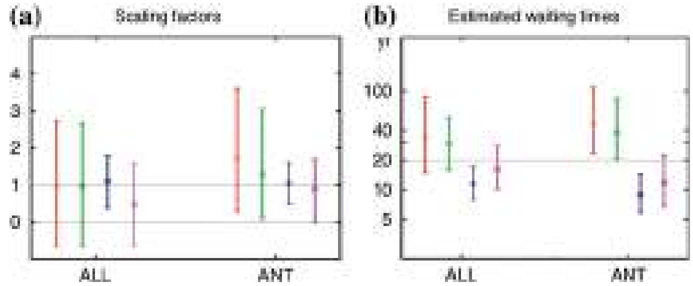

Fig. 23.2 **a** Scaling factors for annual temperature extremes and **b** estimated waiting times of 1960s 20-year return values in the 1990s. The colours refer to extreme value distributions fitted to different annual temperature extremes, namely (per year) the coldest night in *red*, the coldest day in *green*, the warmest night in *blue* and the hottest day in *purple*. The symbols indicate the central estimate, and the bars denote the 90 % confidence interval. Detection of an external influence is claimed at the 10 % level if the confidence interval around the scaling in (**a**) does not include zero (reproduced from Zwiers et al. 2010)

different temperature extremes. They found that cold events that used to occur on average every 20 years in the 1960s were roughly twice as rare in the 1990s due to human influence. Similarly, waiting times for hot extremes are approximately halved; that is, an extremely hot day with a probability of occurrence of once in 20 years in the 1960s is expected to occur on average once in 10 years in the 1990s due to human influence.

Christidis et al. (2005) found detectable changes in the warmest nights and coldest nights and days of the year globally. Using a more sophisticated measure of extremes, Christidis et al. (2011) also detected an anthropogenic influence on changes in the hottest day of the year. Regional detail of the changes in the warmest nights is given by Christidis et al. (2010b); their analysis, however, identified a dependence of the first-guess anthropogenic signal in the Baltic Sea area on the model used. This illustrates the importance of taking intermodel differences such as differences in the regional response to external forcing into account in regional detection and attribution analyses.

23.2.3 Potential Influence of Circulation Changes on the Detectability of Warming

A major caveat of attributing causes for the observed warming in northern Europe is the poor understanding of the possible influence of changes in Northern Hemisphere circulation (Gillett 2005). It is known that modes of natural variability such as the North Atlantic Oscillation (NAO) have a strong influence on temperature in the Baltic Sea region (e.g. Hurrell et al. 2003, Chap. 4, Box 4.1 and Sect. 23.3.1), but the uncertainties in model-simulated circulation and the lack of understanding in the related key processes represent a major source of uncertainty in model predictions over Europe (Woollings 2010).

Gillett et al. (2000) investigated the effect of changes in the Arctic Oscillation (AO) on detection results and found that for Northern Hemisphere temperatures, the exclusion of the AO-related temperature variability has a negligible effect on the outcome of an optimal detection analysis. These findings are corroborated by Wu and Karoly (2007), who found significant trends in the observed warming time series even after removing warming related to changes in circulation. For the Baltic Sea area, however, exclusion of warming related to circulation changes reduces the significance, thus indicating that model biases in representing circulation changes may more strongly affect regional attribution studies. In addition, the follow-up study of Wu (2010) illustrated that while the above results may hold for seasonal mean daily mean temperatures, modes of atmospheric variability have a strong influence on trends in seasonal mean daily maximum and minimum temperature in northern Europe in winter (JFM). The significant observed warming in these variables from 1951 to 2000 can be explained by changes in the leading modes of circulation variability in the Northern Hemisphere alone.

23.3 Causes of Change in Circulation and the Hydrological Cycle

23.3.1 Large-Scale Circulation

Northern European climate is strongly related to the NAO (Hurrell et al. 2003). Changes in this mode of atmospheric variability have been shown to not be well simulated by present-day climate models (Gillett 2005; Miller et al. 2006). The sign of the simulated change generally corresponds with the observed change, but the simulations seem to underestimate the magnitude of the observed change.

Gillett et al. (2005) detected an external influence on the observed global sea level pressure (SLP) changes in winter

(DJF) and Wang et al. (2009) on North Atlantic SLP changes in winter (JFM). The fingerprint used in Gillett et al. (2005) featured positive trends in the southern North Atlantic and negative trends around Iceland, thus representing an increase in the NAO index. In a more recent study, Gillett and Stott (2009) were able to attribute zonally averaged SLP changes to anthropogenic influence; however, detection and attribution using northern mid- to high-latitude data alone fails. In contrast to earlier work, the authors found that the magnitude of observed global SLP variability and change is consistent with simulations with the HadGEM1 model. Gillett and Stott (2009) analysed SLP changes from 1959 to 2009 and therefore also included the recent return of the NAO index to neutral and negative conditions (see Chap. 4, Box 4.1 and Fig. 4.1), which may reduce the discrepancy between simulations and observations. Nonetheless, there is still indication of an underestimation of the recent decrease in winter SLP in high northern latitudes. So far, it remains unclear to what extent changes in Northern Hemispheric SLP and NAO in particular are consistent with simulated anthropogenic signals.

23.3.2 Near-Surface Wind and Storminess

A variety of indices to characterise various aspects of storm climate can be found in the literature (see Chap. 4, Sect. 4.3). For the investigation of long-term changes, storminess indices based on pressure readings seem to provide the necessary homogeneous time series (see Krueger and von Storch 2011, for a discussion of related issues). In contrast, direct measurements of winds often suffer from inhomogeneity in the records related to site and instrument changes and build-up of surrounding areas (Trenberth et al. 2007; Lindenberg et al. 2012); they are thus not suited for attribution studies.

Wang et al. (2009) analysed geostrophic wind energy (derived from SLP) and ocean wave height across the Northern Hemisphere in a formal detection and attribution assessment and found a detectable external influence in winter (JFM). Their ensemble mean fingerprint includes increasing wind energy in north-western Europe. The models are able to reproduce the basic pattern, but simulated changes are smaller than the observed changes, and there is indication that models tend to underestimate internal variability compared with observations.

Using a different definition of storminess derived from pressure triangles, Matulla et al. (2008), Bärring and Fortuniak (2009) and Wang et al. (2011) confirmed the recent increase in storminess in northern Europe. Putting the recent increase in the historical context, however, Matulla et al. (2008) and Bärring and Fortuniak (2009) found little evidence of an emerging anthropogenic signal in the Baltic Sea area as similar increases have been observed in the past. The long-term series of Wang et al. (2011) showed weak trends or a slight decrease in storminess in north-western Europe from 1875 to 2005. In contrast, Donat et al. (2011) identified a long-term increase in storminess in the twentieth-century reanalysis (20CR, Compo et al. 2011), but Krueger et al. (2013) showed that long-term changes in storminess in 20CR are not consistent with observations and may be an artefact of the temporally varying number of stations assimilated in the reanalysis.

Even though the recent observed changes in storminess in the North Atlantic and Northern Hemisphere have been detected, to what extent human influence has caused these changes is still a matter of debate as similar changes have been observed in the past. Therefore, more work is needed to reconcile differences between the various approaches. Evaluating climate models against the existing long observational series may offer additional insights into causes of recent and past changes in storminess and careful validation of the models' ability to reproduce low-frequency variability may help to strengthen future work on detection and attribution of changes in storminess.

23.3.3 Hydrological Cycle

Changes in the hydrological cycle are much more complex than changes to thermal quantities, as both thermodynamics and circulation changes play an important role in shaping changes in the hydrological cycle. The characteristics of the global response of the hydrological cycle can be summarised as follows. An absolute humidity increase of 7 % per °C is expected, as a consequence of the increased water holding capacity of the atmosphere with rising temperature, and given relative humidity patterns stay approximately constant (Allen and Ingram 2002; Held and Soden 2006). The observed increase in global surface-specific humidity has been attributed to human influence (Willett et al. 2007). In the last decade, however, surface humidity has not increased further in concert with global temperature (Simmons et al. 2010), but levelled off after the 1997/98 El Niño. The authors speculated that surface humidity is instead dominated by ocean temperature.

In contrast to absolute humidity, precipitation increases by only 1–3 % per °C (Allen and Ingram 2002; Held and Soden 2006; Wentz et al. 2007). This apparent discrepancy is resolved in models by decreasing convective mass flux and a slowdown of atmospheric circulation mainly in the tropics. Further consequences include an increase in the pattern of evaporation minus precipitation; that is, wet regions get wetter, dry regions get drier (Held and Soden 2006).

Global equilibrium precipitation depends on the perturbation of the tropospheric energy balance rather than the availability of moisture (Allen and Ingram 2002). Global precipitation is less sensitive to greenhouse gas forcing than shortwave forcing such as volcanic eruptions or solar irradiance changes. Increasing atmospheric greenhouse gas concentrations reduce the ability of the troposphere to radiate away latent heat from precipitation and thus counteract the precipitation increase due to surface warming (Allen and Ingram 2002). Therefore, shortwave forcing such as changes in volcanic aerosols should be easier to detect in global precipitation than the effect of the long-term increase of atmospheric greenhouse gas concentrations. Indeed, Lambert et al. (2004, 2005) detected an influence of the combined natural and anthropogenic forcing and concluded that most of the forced signal in global mean precipitation is due to natural forcing. These findings are in line with Gillett et al. (2004) who detected the influence of volcanic eruptions on the observed global land precipitation change over the twentieth century. It is important to note that the main process leading to precipitation changes is the radiative forcing leading to a perturbation of the tropospheric energy balance and not a potential effect of volcanic ash particles on cloud properties.

Recently, the influence of long-term forcing has also been detected. Zhang et al. (2007) found a detectable anthropogenic influence on recent observed trends in annual zonal mean precipitation. Their anthropogenic fingerprint features the well-known pattern of moistening in the tropics and high latitudes, and drying in the subtropics. The climate models, however, significantly underestimate the observed changes. The authors concluded that anthropogenic forcing contributed about 50–85 % to the observed increase in precipitation in the northern mid-latitudes of 6.2 mm per decade. Noake et al. (2012) also detected external forcing in seasonal zonal mean precipitation over land in all seasons except boreal summer. Furthermore, they found that using relative anomalies reduces underestimation of observed precipitation trends. For northern Europe, Bhend (2010) detected external forcing on seasonal area-average precipitation in the Baltic Sea area. The analysis, however, also reveals that global climate models are not able to reproduce the observed variability in regional precipitation and the model-derived signals have to be inflated significantly to best fit the observations. Therefore, the confidence in this detection finding is low.

Bhend and von Storch (2008) found that the pattern of recent observed changes in winter precipitation is consistent with the anthropogenic signal—mainly greenhouse gases—derived from regional climate model simulations. The magnitude of the change, however, is much smaller in the simulations compared with the observations in line with other studies and the discussion about understanding regional circulation changes. Van Haren et al. (2012) also identified significant discrepancies between observed and simulated precipitation changes in Europe in winter (Fig. 23.3a). They further established that biases in simulated sea-surface temperature (SST) and circulation changes are the main causes of the inconsistencies between simulated and observed changes in regional precipitation in winter. In summer, the simulated precipitation changes correspond well with the observations (Fig. 23.3b).

Min et al. (2008a) found a detectable human influence on Arctic moistening. Their analysis indicates that anthropogenic forcing has led to an increase in precipitation over land north of 55°N (including the northern Baltic Sea area), whereas natural forcing has led to a decrease in precipitation from 1950 to 1999. Their analysis further suggests that the simulated changes underestimate the observed moistening considerably, and their best-guess anthropogenic signal has to be scaled up significantly to match the observed change; moreover, the simulated internal variability is not consistent with the observed residual variability, which indicates that the models underestimate both forced and internal variability in precipitation. If AO-related variability is removed from the observations, simulated changes including all forcing mechanisms and the simulated variability are consistent with the observations.

Min et al. (2009) investigated potential detectability of change in extreme precipitation in simulations. They found that detectability of changes in extreme precipitation in Europe is low for the twentieth century. Recently, Min et al.

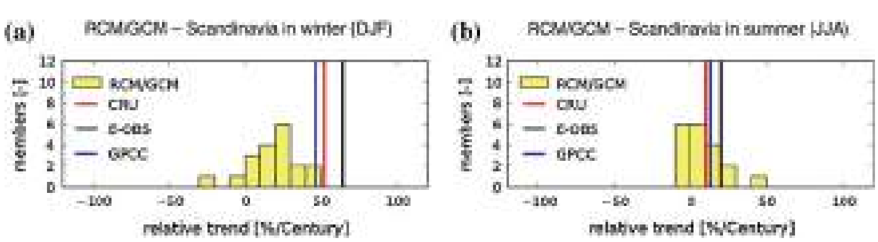

Fig. 23.3 Observed relative precipitation trends in winter (DJF, **a**) and summer (JJA, **b**) from 1961 to 2000 (*vertical lines*) along with the distribution of simulated precipitation trends in an ensemble of 19 regional climate model simulations (*yellow bars*, updated from van Haren et al. 2012). In contrast to the original manuscript, the observed trends are derived from the precipitation dataset of the Global Precipitation Climatology Centre (GPCC) and the updated CRU_TS3.10.01 and E-OBS v7.0 datasets. The ensemble of regional climate model simulations includes different regional/global model combinations (van Haren et al. 2012). Most of the models fail to reproduce the observed increase in precipitation in winter

(2011) detected an anthropogenic influence on annual maxima of daily and five-day consecutive precipitation changes from 1951 to 1999 over land in the northern mid-latitudes, but the signal is not detectable for Eurasia alone. As with mean precipitation, their results suggest that models underestimate the observed change.

In climates with intermittent snow cover, such as the Baltic Sea catchment, run-off regimes change with temperature. Earlier snowmelt due to warming leads to peak river run-off earlier in the year. Recent studies demonstrate a detectable anthropogenic influence on regional run-off for the north-western USA (Barnett et al. 2008; Hidalgo et al. 2009). However, Hansson et al. (2011) concluded that recent changes in river run-off into the Baltic Sea are not exceptional compared to changes over the past 500 years according to their reconstruction of river run-off. A formal attribution analysis for river run-off in Europe, however, is not available so far. The above studies and the often long observational records of river run-off illustrate the potential for regional attribution. On the other hand, the complex processes involved in run-off generation and human influence through land-use change in the catchment area, withdrawal and regulation are major challenges when trying to attribute run-off changes.

There is some evidence for a human influence on run-off from global and continental studies. Milly et al. (2005) illustrated that observed changes in run-off around the world are significantly correlated with simulated changes due to anthropogenic forcing—thus indicating that the global pattern of observed streamflow changes during the twentieth century is unlikely due to internal variability alone. Stahl et al. (2010) and Wilson et al. (2010) concluded that streamflow changes in near-natural catchments in northern Europe in winter and spring are congruent with expected streamflow changes due to human influence. Furthermore, Wilson et al. (2010) found that recent streamflow changes are in line with expected future changes in seasons when the temperature signal dominates (winter and spring). In contrast, the expected future increase in summer and autumn streamflow due to increasing precipitation and the spatial characteristics of the human signal in streamflow do not manifest in recent observed changes.

A vast body of literature examines the proximate causes of interannual snow cover variability. Large-scale atmospheric circulation—and the NAO in particular—have been shown to be a strong determining factor of changing snow conditions in the Baltic Sea region. The large-scale circulation affects the extent of European snow cover (Henderson and Leathers 2010), snow amounts (Kohler et al. 2006; Falarz 2007; Popova 2007; Bednorz and Wibig 2008), the occurrence of heavy snowfall (Bednorz and Wibig 2008) and snow cover duration (Klavins et al. 2009). However, formal assessments of the ultimate causes, such as increasing atmospheric greenhouse gas concentrations, for changes in snow conditions are still rare. A first formal detection and attribution assessment has been carried out for the north-western USA identifying a human influence on the decline in snowpack measured as the snow-water equivalent in spring (Pierce et al. 2008). For the Baltic Sea area, no such analysis is available so far.

The response of snow to observed recent warming and increasing precipitation in winter is non-trivial and depends on various factors such as the climatology and, therefore, varies with elevation and continentality (Räisänen 2008). Furthermore, not all snow properties will be equally sensitive to warming. Snow cover duration has been identified to be most sensitive to warming (Brown and Mote 2009), and earlier snow melt and decreasing snow cover duration have been found in the Baltic Sea area (see Chap. 6; Brown and Mote 2009; Choi et al. 2010). These changes are in line with the expected response to anthropogenic forcing. On the other hand, changes in maximum snow depth or maximum snow-water equivalent are very variable over the Baltic Sea area (see Chap. 6; Liston and Hiemstra 2011), which is due to the strong interannual variability; an anthropogenic signal may not have emerged yet (Räisänen and Eklund 2011).

Box 23.1 Event attribution: proximate versus ultimate causes

When identifying causes of individual weather- and climate-related extreme events, typically two distinct approaches are taken. Often, attempts are made to identify proximate causes such as the large-scale circulation or specific sea-surface temperature patterns favouring the occurrence and intensity of the event under consideration. For example, the NAO and the El Niño/Southern Oscillation (ENSO) have been identified as key players for the cold and snowy winter of 2009/2010 in Europe (Cohen et al. 2010; Seager et al. 2010). In contrast, there are very few studies attempting to identify the contribution of external forcing (ultimate causes) to extreme events (Stott et al. 2004; Pall et al. 2011). In the context of the cold winter of 2009/2010 in Europe, Cattiaux et al. (2010) concluded that the winter was not exceptional compared to past winters and was also considerably less cold than expected due to the record-breaking circulation indices (NAO and blocking) alone. This illustrates the combined effect of circulation anomalies (proximate causes) and background warming (here a proxy for external forcing causing global warming). In this case, it could be concluded that winter 2009/2010 was a cold winter *despite* global warming and an exceptionally cold winter *given* global warming.

Identified proximate causes of extreme events are often perceived as contradicting a potential human influence. This is generally not the case. An extreme event is by definition rare and thus will only occur under particular conditions. Therefore, a large fraction of the intensity of an extreme event may be due to internal variability or, equivalently, an extreme event may not occur due to external forcing alone. This gave rise to the widespread public belief that it is not possible to attribute a single event to climate change. External forcing, however, can have a strong influence on the frequency of occurrence of extreme events as events that are exceedingly rare (or common) in the pre-industrial climate may become more (less) frequent due to external forcing.

The subtleties and importance of clarity in relation to the question asked in event attribution are illustrated by the case of the Russian heat wave of 2010. Dole et al. (2011) identified atmospheric blocking and thus natural internal variability as the main contributor to the intensity of the heat wave. In contrast, Rahmstorf and Coumou (2011) concluded that with a probability of 80 % the Russian heat wave would not have occurred without the recent large-scale warming (most of which is attributable to human influence). The apparent contradiction is resolved by Otto et al. (2012), pointing out that whereas anthropogenic forcing only contributed relatively little to the intensity of the heat wave, it increased the estimated return time of such an event considerably (see Fig. 23.4).

In addition to the challenging distinction between the roles of proximate and ultimate causes, attribution of extreme events may be subject to selection bias. This selection bias relates to the fact that it is usually time series with recent extreme events that are studied (see for example Stott et al. 2004 and Coumou and Rahmstorf 2012). Selection biases could be avoided by using operationalised systems that routinely assess the attribution question for a set of pre-specified indices.

Attribution of extreme events (and climate change in general) to both proximate and ultimate causes is important to improve understanding of the climate system and to enhance predictability of extreme events. The communication of often seemingly contradictory findings arising from the two different approaches, however, will remain challenging.

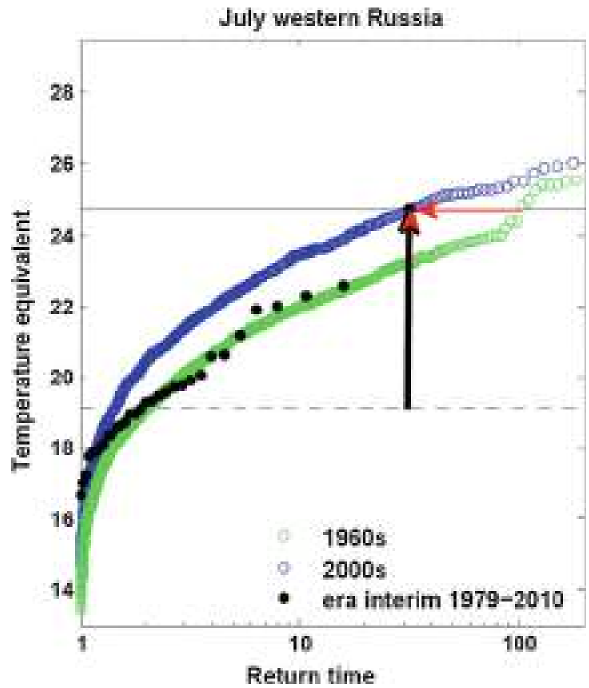

Fig. 23.4 Return periods of an index describing combined temperature and geopotential height conditions in the model for the 1960s (*green*) and the 2000s (*blue*) and in ERA-Interim for 1979–2010 (*black*). The vertical black arrow shows the anomaly of the Russian heat wave in 2010 (*black horizontal line*) compared to the July mean temperatures of the 1960s (*dashed line*). The *vertical red arrow* indicates the increase in the magnitude of the heat wave due to the shift of the distribution, whereas the *horizontal red arrow* shows the change in the return period (reproduced from Otto et al. 2012)

23.4 Causes of Change in the Baltic Sea

The human influence on ocean heat content has been detected (Barnett et al. 2001), and changes in all of the major ocean basins are found to be different from internal variability (Pierce et al. 2006; Palmer et al. 2009). Due to the complex and small-scale bathymetry of the Baltic Sea, however, the available atmosphere–ocean general circulation model (AOGCM) simulations cannot be used for detection and attribution studies in this area. Therefore, no formal detection and attribution assessment for the Baltic Sea is yet available.

Hansson and Omstedt (2008) investigated maximum ice extent and horizontally and vertically integrated temperature in the Baltic Sea using the PROBE-Baltic model. They used observations and climate proxy reconstructions as boundary conditions for their model and concluded that the recent warming and recent rate of warming does not stand out in the context of the past 500 years and thus cannot be detected. Furthermore, they do not recommend using GCM data as boundary conditions for their ocean model, thus inhibiting a formal detection and attribution approach.

Salinity in major ocean basins has been used to infer changes in freshwater run-off and precipitation (Hegerl et al. 2007b). Temporal variability of the salinity in the Baltic Sea, however, is strongly dependent on Major Baltic Inflow events of highly saline water from the North Sea (Matthäus et al. 2008). These events in turn depend on the large-scale circulation and on the salinity of the Baltic Sea (among other factors). Meier and Kauker (2003) found, based on hindcasts of Baltic Sea salinity, that salinity changes in the Baltic Sea are linked partly to changes in freshwater influx and precipitation and partly to changes in the large-scale atmospheric circulation. Due to the coupling of Baltic Sea salinity with the salinity of the North Sea and the dependence of the exchange on properties in both basins, attribution of changes in Baltic Sea salinity may only be partially achieved.

As the density of water decreases (and thus its volume increases) with increasing temperature, global sea level generally rises with increasing ocean heat content (among other factors, see Chap. 9, Sect. 9.3). Global sea level changes since 1960 agree well with simulations including anthropogenic and volcanic forcing (Domingues et al. 2008). The regional pattern of sea level rise, however, is only partly understood. Changes in sea level in the Baltic Sea have been shown to vary in concert with circulation and precipitation changes (Hünicke and Zorita 2006). The contribution of these regional effects to sea level in the Baltic Sea are of the same order of magnitude as the global sea level rise (Hünicke 2010) and thus have to be accounted for in future detection and attribution studies.

Finally, changes in sea ice are to be expected with global warming. Arctic sea ice changes have been attributed to human influences (Min et al. 2008b) and are in fact attributable since 1992. Sea ice formation in the Baltic Sea, however, is hardly comparable to Arctic sea ice, as the brackish water, the complex bathymetry and the limited extent of the Baltic Sea lead to distinct features of ice formation. The annual maximum Baltic Sea ice extent is decreasing, and from 1987 to 2009, all winters have been average or below average with regard to maximum ice extent (Vihma and Haapala 2009). The general tendency towards milder winters is masked by considerable interannual variability; the most recent winters of 2009/2010 and 2010/2011, for example, have been judged to be severe winters by the Swedish Meteorological and Hydrological Institute (SMHI). The long time series of ice extent and break-up dates available across the Baltic Sea would lend itself to attribution assessments (see Chap. 8). A formal detection and attribution assessment of Baltic Sea ice, however, is not available so far.

23.5 Causes of Climate Change Impacts

Attributing climate change impacts to causes is often complicated by the multitude of confounding factors acting on the system. Changes in marine biodiversity, for example, have been linked to climate (Hiddink and Coleby 2011), and ecosystem changes have been linked to atmospheric and direct human interference with the system (Moellmann et al. 2009). As in other attribution assessments, these linkages are based on multivariate linear regression and the results can be strongly dependent on the number of potential drivers included in the analysis. While statistical approaches provide readily available tools to explore relationships in complex systems, avoiding selection bias is crucial to avoid misattribution in systems with a multitude of potential causes.

There is a multitude of approaches for attributing causes to observed changes in climate change impacts. Hegerl et al. (2010) suggested making the distinction between four different approaches:

1. *Single-step attribution to external forcing* These methods assess the influence of external forcing onto an observed quantity with an integrated modelling system that explicitly simulates the effect of all plausible drivers on the respective variable.
2. *Multi-step attribution to external forcing* consists of several independent but linked attribution assessments. First, changes in impacts or changes in a biological system are attributed to changes in climatic conditions. In a second step, the changes in climatic conditions are attributed to changes in external forcing. The two independent assessments and their respective uncertainties are then combined to describe the resulting effect of the external forcing on the target quantity. Such an approach may be advantageous if modelling systems are incapable of faithfully reproducing the link between the target quantity and the climatic conditions.
3. *Associative pattern attribution to external forcing* is a 'meta-analysis' to characterise the sensitivity of systems to changes in external forcing based on correspondence or disagreement in the relative response across a large number of studies (in different regions and/or systems).
4. *Attribution to a change in climatic conditions* (but not explicitly to changes in external forcing) can be the last step in a multi-step attribution analysis but is more often found as a stand-alone analysis (see Box 23.1).

An example of an associative pattern attribution analysis is the study of Rosenzweig et al. (2008). Based on a compilation of significant findings from a wide range of analyses of change in physical and biological systems, the authors calculated the fraction of the findings that are consistent with the local warming. For Europe, they found that 94 % of the studies investigating change in the physical systems and 90 % of the studies on changes in the biological systems find changes that are consistent with the observed warming. While such an approach is valuable in providing an overview of findings across systems, there is the potential danger of sampling issues influencing the results.

If process models of the system are available, the uncertainties in the response to changes in the various drivers (climate and other) can be fully explored. Such an end-to-end attribution analysis, however, is so far not available for the Baltic Sea area.

23.6 Conclusion

There is a wealth of evidence documenting human influence on the observed warming globally, on all continents and in many subcontinental regions. For the last assessment report (BACC Author Team 2008), no formal assessments on the potential causes of the observed climate change in the Baltic Sea area had been available. Since then, a few such studies have been published. There is some indication of an emerging anthropogenic signal, which is detectable in thermal quantities such as seasonal temperature, but evidence for detection of changes in non-thermal quantities such as circulation and precipitation is weak.

Although human influence (mainly increasing concentrations of greenhouse gases) has been identified as a cause of the recent warming in the Baltic Sea area, there are caveats. The causes of recent circulation changes in the Euro-Atlantic sector are not yet understood, and therefore, attribution of changes especially in winter and spring are to be treated with caution. Furthermore, quantification of the contribution of individual forcings has not been accomplished. Better understanding of the regional effects of natural forcings and the effect of anthropogenic aerosols (see Chap. 24) may help to achieve quantified attribution statements. Finally, detection and attribution efforts are often subject to selection and publication biases. That is, time series with 'interesting' behaviour are preferentially studied and positive findings (of a detectable anthropogenic effect) are more likely to be published. To avoid these biases, systems to routinely issue statements on the contribution of external forcing to the observed climate across a range of pre-specified variables and regions are needed.

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Causes of Regional Change—Aerosols

24

Hans-Christen Hansson and Jonas Bhend

Abstract

This chapter starts by introducing the complex nature of atmospheric aerosols, their sources, formation and properties and describes how they interact with clouds. This is important background information for discussing how aerosols affect climate, both directly and indirectly by affecting the radiative properties of clouds. The complexity of the aerosol–cloud–climate interaction causes large uncertainty in the projections of future climate. Results from different modelling studies on the European region are presented, and these show that the large spatial and temporal variations in atmospheric aerosol concentrations and properties have large regional differences in their effect on climate. This chapter concludes with an example of a co-beneficial global air quality and climate change mitigation scenario.

24.1 Introduction

This chapter starts by introducing the complex nature of atmospheric aerosols, their sources, formation and properties and describes how they interact with clouds. This is important background information for discussing how aerosols affect climate, both directly and indirectly by affecting the radiative properties of clouds. The complexity of the aerosol–cloud–climate interaction causes large uncertainty in the projections of future climate. Results from different modelling studies on the European region are presented, and these show that the large spatial and temporal variations in atmospheric aerosol concentrations and properties have large regional differences in their effect on climate. This chapter concludes with an example of a co-beneficial global air quality and climate change mitigation scenario.

24.2 The Basics About Aerosols and Climate

24.2.1 Influence of Aerosols on Climate

Atmospheric aerosols, particles and droplets in the atmosphere influence the Earth's radiation balance directly and indirectly. Aerosols interact directly with incoming solar radiation by scattering and absorbing the sunlight. Aerosols can be observed on satellite images. Saharan dust, for example, is often easily observed in the trade wind belt when transported over the Atlantic towards South America, so are anthropogenic aerosols in the Po Valley (Italy) and in the outflow of air masses from the Chinese mainland. Changes in aerosol concentration or changes in its composition affecting its optical properties influence the radiation balance. The total direct effect of aerosols has been investigated in many studies starting with that by Charlson et al. (1991). Estimates of direct radiative forcing due to anthropogenic

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_24

aerosols range up to about −1 W m^{-2}. The Intergovernmental Panel on Climate Change (IPCC) gave −0.5 ± 0.4 W m^{-2} as a best estimate (IPCC 2007).

Clouds have a large effect on the radiation balance. Thus, changes in their albedo or lifetime can have a substantial influence on climate. There are several process chains through which aerosols are suggested to affect cloud albedo and lifetime. The first indirect aerosol effect, also called the Twomey effect, is caused by an increasing number of cloud droplets due to an increasing number of particles resulting in a higher cloud albedo. Marine stratus clouds are the most susceptible to changing particle number (Andreae and Rosenfeld 2008). This effect is well known and is observed by satellite. However, there is large uncertainty in the estimates of the global Twomey effect with a best estimate of −0.7 W m^{-2} (range −0.3 to −1.8 W m^{-2}) (IPCC 2007).

Other indirect effects mainly concern cloud lifetime often as a consequence of the change in droplet size and number described in the context of the Twomey effect. They concern the influence of aerosols on the coalescence processes initiating rain, as for example increased amount of absorbing material evaporating the droplets, effects on ice nuclei formation and other processes in ice or mixed clouds affecting the formation of precipitation. Even though large positive and negative climate forcing caused by these effects has been reported, they are not well known and there is no scientific consensus on their climate effect (Lohmann and Hoose 2009; Stevens and Feingold 2009).

The total aerosol effect, including direct and indirect effects, yields an estimate of −1.2 W m^{-2} (5–95 % range −0.6 to −2.4 W m^{-2}) since preindustrial times (IPCC 2007). This can be compared to the forcing by carbon dioxide (CO_2) of 1.66 ± 0.17 W m^{-2} and total greenhouse gas forcing of 2.9 ± 0.3 W m^{-2} (IPCC 2007). The estimated total aerosol effect is considerable; aerosols counteract almost half of the greenhouse gas-induced warming since preindustrial times. As the aerosols have a limited atmospheric lifetime, their influence is mainly limited to the source region. The regional patterns of climate change can thus be expected to vary widely and in places to be considerably greater than the global average climate change.

The magnitude of the total aerosol climate effect has been shown to be crucial in making better projections of future climate. The present total aerosol effect, estimated at about −1.2 W m^{-2}, is considerable compared to the present greenhouse gas effect, estimated at about +2.9 W m^{-2}; that is, aerosols currently mask more than a third of the greenhouse gas-induced warming. Schwartz et al. (2010) showed that uncertainty in the total anthropogenic aerosol effect dominates uncertainty in projecting future warming, giving a range of 1.5–4.5 °C for the projected temperature increase at double the preindustrial CO_2 concentration. Considering the possible regional climate effects at a global temperature increase of 4.5 °C compared to a temperature increase of 1.5 °C, the uncertainty must be seen as very large which would seriously affect the confidence in the projections as well causing a large uncertainty in climate change mitigation strategies.

Box 24.1 Sources of aerosols

Atmospheric aerosols originate from a variety of natural and anthropogenic sources. These can be classified as primary sources that emit aerosols directly and secondary sources that emit precursor gases that form particles in the atmosphere through physical and chemical reactions.

The total mass of the global natural aerosol emissions is dominated by primary sources emitting particles larger than 1 µm (such as desert dust and sea spray). Secondary sources contribute less than 10 % to the global natural aerosol mass with mainly fine particles less than 1 µm in aerodynamic diameter (Kiehl and Rodhe 1995). Although contributing much less in terms of mass, secondary particles contribute a much greater number of particles and hence dominate the aerosol effect on climate. This is because radiation and cloud formation depend more on the number of particles than on the mass.

In contrast to natural emissions, anthropogenic emissions mainly contribute to the fine particle fraction and have increased the global fine particle mass loading considerably since preindustrial times (Table 24.1). It should be noted that although anthropogenic activities are responsible for a large global emission of aerosols, their effect is mostly regional as their atmospheric lifetime is fairly short; only a few days to a week giving a transport range of typically 1000–2000 km (Tunved et al. 2003). This means that the southern Baltic Sea area aerosol is dominated by particles from anthropogenic sources, while natural aerosols dominate the northern Baltic Sea area.

Owing to the multitude of sources and processes involved, measuring natural aerosol emissions is difficult and estimates vary widely (e.g. Kiehl and Rodhe 1995; Andreae and Rosenfeld 2008). As a result, natural emission estimates used in climate models differ considerably, whereas estimates for anthropogenic sources are generally similar across models (Textor et al. 2006). The combined natural and anthropogenic aerosol burden determines the aerosol optical depth, which is measured globally and with high spatial and temporal resolution by satellites.

Table 24.1 Global emission estimates of the major aerosol sources (based on Andreae and Rosenfeld 2008)

Global aerosol emissions (Tg $year^{-1}$)			Total particles	Coarse particles	Fine particles	Natural fine particles	Anthropogenic fine particles
Carbonaceous aerosols							
	Primary organic	0–2 μm	95		95		
		Biomass burning					54
		Fossil fuel					4
		Biogenic				35	
	Black carbon	0–2 μm	10		10		
		Open burning, biofuel					6
		Fossil fuel					4.5
	Secondary organic		28		28		
		Biogenic				25	
		Anthropogenic					3.5
Sulphates			200		200		
	Biogenic sulphates					57	
	Volcanic sulphates					21	
	Anthropogenic sulphates						122
Nitrates			18		18	9	9
Industrial dust			100	70	30		30
Sea salt			10,130				
	<1 μm				180	180	
	1–16 μm			9,940			
Mineral dust			1,600				
	<1 μm				165	165	
	1–2.5 μm			496			
	2.5–10 μm			992			
Total			12,181	11,498	726	492	233

Climate models generally show fair agreement with these measurements (Kinne et al. 2006). European sources affecting the Baltic Sea area are discussed in detail in Chap. 15. The past and projected future emissions in Europe and their effect on regional climate are discussed in Sect. 24.6.

24.2.2 Atmospheric Particle Size Distribution

Atmospheric particles range in size over about five orders of magnitude. Observations of the particle number size distributions show that generally four modes can be identified (see Fig. 24.1). The fine mode particles include nuclei mode particles in the 1–10 nm range, Aitken mode particles from 20–80 nm and accumulation mode particles from 100 to 200 nm. Coarse mode particles are found in the range 1–100 μm. The nuclei mode emerges when new particles form, that is nucleation. Nuclei-mode-sized particles also emerge as primary particles from high-temperature combustion (such as engines), and 10-nm sea spray particles have even been observed. The Aitken mode consists of particles grown out of nuclei mode due to coagulation and condensation of condensable atmospheric gases originating from gaseous emissions and gas phase reactions in the atmosphere. Further growth into accumulation-mode-sized particles requires other processes such as liquid phase processes in clouds. Liquid phase chemistry occurs when the absorption of atmospheric gases into cloud droplets forms a very reactive environment, for example sulphur dioxide (SO_2) is mainly oxidised to sulphate (SO_4) in cloud droplets. Primary particles are observed in all size fractions, but in the sub-micron

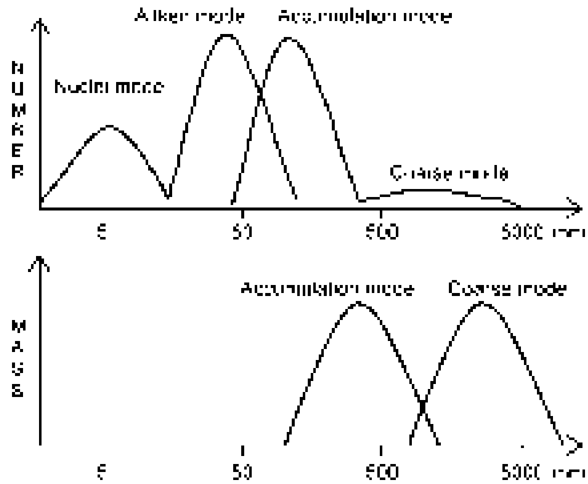

Fig. 24.1 Schematic illustration of particle number and mass distribution with particle size for atmospheric aerosol. Number is dominated by sub-micron particles in the size range 10–200 nm, observed mainly in three modes depending on source and formation process. Particle mass is mainly found in larger sub-micron and super-micron particles

size range are mainly from high-temperature processes and in the coarse mode mainly from mechanical processes.

Not only do emissions influence the number of particles in different size fractions, but sink processes are also important. Although diffusion losses are an important sink process, mainly occurring through collision with other particles or droplets, precipitation (rain or snow) is the main sink for atmospheric particles larger than about 50–70 nm. Such processes affect the fine mode particles mainly through in-cloud scavenging where particles act as cloud condensation nuclei and eventually get scavenged in a precipitating cloud. Whether a particle acts as a cloud condensation nucleus depends on its size as well as its chemical composition. Generally, particles in the upper Aitken size range and larger with a major fraction of hygroscopic compounds such as inorganic salts are good condensation nuclei in warm clouds. In ice or mixed clouds, the ice and snow formation processes are still not well known and information on which particles act as good ice nuclei is poor. Formation of ice crystals in a cloud strongly influences the onset of precipitation. For coarse particles, losses due to sedimentation start to become significant at about 5 μm.

The atmospheric particle size distribution reflects the dynamic interaction of sources, atmospheric formation processes and sinks mixing natural and anthropogenic components. Different air masses can thus have very different particle size distributions depending on the sources over which they had passed and the meteorology they had experienced. Therefore, it is not possible to identify a specific European size distribution, but rather a more or less scavenged pollution aerosol mixed with natural emissions.

24.2.3 Formation of Aerosols in the Atmosphere

Aerosols are formed in the atmosphere either through new particle formation, condensation onto already existing particles or in-cloud liquid processes. New particle formation, often called nucleation, is the process by which molecules form unstable clusters that grow over a threshold size to form a stable particle. The threshold size is about 1–2 nm. Nucleation is followed by strong condensation causing the newly formed particles to grow rapidly in diameter by several nanometres per hour (Vehkamäki and Riipinen 2012). Nucleation has been studied extensively in the boreal forest. During a 15-year period of measurements, major nucleation events occurred about 100 times a year (Kulmala et al. 2004). The boreal forest that almost completely surrounds the Baltic Sea is thus one of the most important sources of natural aerosol in the area.

A nucleation event, as seen in Fig. 24.2, is clearly visible in the particle size distribution measurements. In the morning, the aerosol is dominated by 50–70 nm particles, which slowly disappear over the course of the morning. Particles of 3 nm in diameter start to form around 10 a.m. and begin to grow. By 3–4 p.m., the particles have grown to about 30–40 nm.

Much attention has been paid to the influence of nucleation on aerosol particle size distribution and its implication for the number of cloud condensation nuclei, especially within the European framework (see the EUCAARI project; Kulmala et al. 2011). The main results of the studies were as follows:

- Nucleation is a major source of aerosol particle number concentrations, and usually, several tens of per cent of sub-micron aerosol particles have originated from condensation of atmospheric vapours—thus being of secondary origin. Roughly, half of the number of aerosol particles in the European boundary layer has originated from nucleation.
- Secondary aerosol formation is due to both natural and anthropogenic influence: while anthropogenic sulphate emissions are a major factor governing formation of new particles, natural emissions of biogenic organic vapours play an important role in defining aerosol size distributions and the climatic impact of aerosols. The results suggest that the anthropogenic contribution (both primary and secondary) dominates in most parts of Europe, the biogenic component being of less importance. However, halving SO_2 and anthropogenic primary particle emissions would only result in reductions of the order of 20 % on the total particle number concentrations—which might suggest that natural aerosol production could compensate somewhat for the reduction in anthropogenic aerosol production.

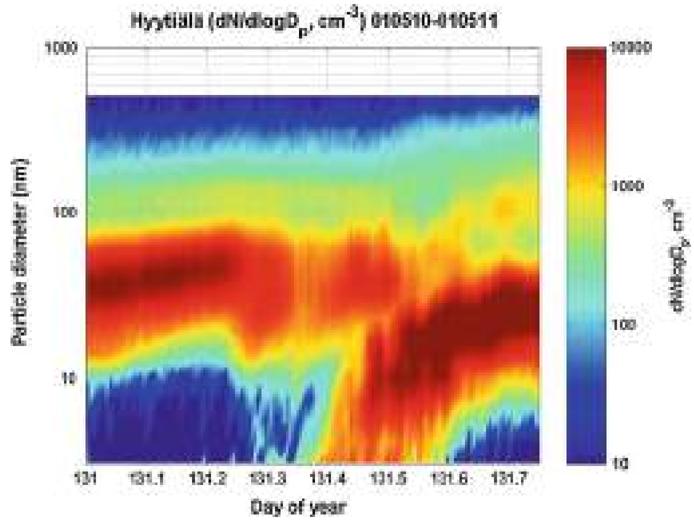

Fig. 24.2 A nucleation event occurring just before noon 10 May 2001 at Hyytiälä, Finland (Kulmala et al. 2004)

It is clear that reductions in anthropogenic primary (e.g. black carbon, also called 'soot') and secondary precursors (e.g. SO_2 and ammonia, NH_3) will not result in proportional changes in the number and size of the particles. The natural production and formation of particles might increase to reach some balance defined by atmospheric chemical and physical conditions.

24.2.4 Aerosol–Cloud Interactions

Atmospheric particles are needed as condensation sites for water vapour to form clouds. Particle size and chemistry influence the number of cloud droplets and thus the development of a cloud. Particles forming cloud droplets are strongly affected by the cloud either by being precipitated or by being transformed by fast liquid phase chemistry with absorbed gases changing the chemistry and increasing the mass of particles emerging from the dissipating cloud.

Not all particles are effective as cloud condensation nuclei, which require a certain amount of soluble hygroscopic compounds to grow into a cloud droplet (Pruppacher and Klett 1997). Such compounds are typically inorganic salts such as sulphates, nitrates and chlorides. Organic compounds are typically less hygroscopic and thus less efficient in providing good cloud condensation nuclei (Laaksonen et al. 1998; Swietlicki et al. 2008). In non-precipitating clouds, particles form droplets, pass through the cloud and exit the cloud, mainly at the top of the cloud, where the water evaporates.

Clouds even at low latitudes often contain ice crystals. The mix of droplets and ice crystals in clouds complicates the physical and chemical processes considerably and affects the formation of precipitation. Research on mixed phase or pure ice clouds is quite complicated, and understanding is not well developed. At present, dust seems to be the most important ice nucleus, while there is disagreement concerning black carbon. Biological particles such as bacteria are good ice nuclei but are often present in low numbers and so are minor ice nuclei contributors (Stratmann et al. 2010).

In conclusion, clouds affect the aerosol both in terms of formation, adding more mass to cloud condensation nuclei through liquid phase chemistry, and as the major sink of sub-micron particles. Particles affect cloud albedo and possibly cloud lifetime. However, information on the specific processes and their general effect on lifetime in, for example, mixed clouds is still limited. A better understanding of these processes and their incorporation in climate models is crucial for advancing climate projection (and hindcasting) capability.

24.3 Aerosol Processes Affecting Climate

24.3.1 Direct Effects

The direct particle effect on climate forcing is caused by scattering and absorption of sunlight. Both scattering and absorption are strongly dependent on particle size but also on the chemical composition of the particles. The total forcing effect is estimated at -0.5 ± 0.4 W m^{-2} (IPCC 2007); scattering accounting for -0.7 W m^{-2} and absorption for $+0.25$ W m^{-2}. However, there has been strong debate on the absorption estimate. Ramanathan and Carmichael (2008) argued that absorption is underestimated and a more accurate global estimate could be as large as $+0.9$ W m^{-2} due to black carbon aerosol alone. While Quaas et al. (2009) using several global models estimated the direct effect at -0.4 ± 0.2 W m^{-2}, the EUCAARI project claimed to have narrowed the range further to -0.2 ± 0.1 W m^{-2} (Kulmala et al. 2011). In an AEROCOM exercise involving nine global climate models, Schulz et al. (2006) found a total direct forcing in the range -0.2 ± 0.2 W m^{-2} including a warming by black carbon in the range $+0.2 \pm 0.15$ W m^{-2}. These findings indicate that the direct effect is relatively small, but Quaas et al. (2008) derived an estimate of the direct effect of -0.9 ± 0.4 W m^{-2} from satellite measurements. However, they recognised the discrepancy and suggested that it may be partly that satellite retrievals of aerosols are not available over bright surfaces such as deserts, snow- or ice-covered surfaces and low-level clouds where the direct forcing may even be positive. They suggested applying a reduction of 30–60 % to the satellite-based estimate, which brings the observations and model results quite close.

An important source of uncertainty in the total direct aerosol effect is the uncertainty in relative humidity, spatially and temporally, as atmospheric particles are generally hygroscopic. In measurements of how the atmospheric aerosol increases in size with increasing humidity, it is usually found that a dominating number of the particles is growing by about 30–50 % in diameter; that is, a factor of 2–3 in volume at 90 % relative humidity while scattering increases by a factor of about 3 compared to dry particles (Swietlicki et al. 2008). Growth is increasingly sensitive to relative humidity, for example, increasing humidity from 90 to 95 % results in an additional increase in scattering of roughly 30 % (Zieger et al. 2010).

When air pollution spreads to the top of the boundary layer, relative humidity usually increases and particles grow and scatter more light back to space. This strongly affects the aerosol optical depth, which is measured by sun photometers from satellites and on the ground generating an extensive database of observations across the globe that is used to control climate models. Errors in estimates of relative humidity thus strongly affect estimates of the direct effect of particles on climate. To ensure good model performance, comparisons with observation of relative humidity and aerosol optical parameters are necessary.

24.3.2 Indirect Effects

At least six aerosol–cloud interactions have been identified that indirectly affect climate. Table 24.2 summarises processes discussed in the literature (Lohmann and Feichter 2005), with newer estimates from Lohmann et al. (2010).

The first indirect aerosol effect affects cloud albedo through the increase in cloud condensation nuclei due to anthropogenic emissions. The most obvious evidence of the Twomey effect (see Table 24.2) is ship tracks easily observable from space. The ship tracks are white narrow cloud streaks resulting from ship emissions.

According to the IPCC (2007), the best estimate of a global climate effect due to the first indirect effect is about -0.7 W m^{-2} with uncertainty giving a range of -0.3 to -1.8 W m^{-2}. Lohmann et al. (2010) suggested -0.9 ± 0.4 W m^{-2} which is close to the -0.7 ± 0.5 W m^{-2} proposed by Kulmala et al. (2011). Quaas et al. (2008) found that satellite measurement showed considerably less effect, giving an estimate of -0.2 ± 0.1 W m^{-2} for the first indirect effect. They recognised this to be considerably lower than most models but consistent with estimates from models constrained by satellite observations of clouds in terms of cloud droplet number, cloud liquid water path and cloud top temperature (Lohmann and Lesins 2002; Quaas et al. 2006). However, Penner et al. (2011) argued that satellite measurements underestimate the first indirect effect by a factor of 3–6 because they typically use the present-day relationship between observed cloud drop number concentrations and

Table 24.2 Summary of different indirect aerosol effects on climate (based on Lohmann and Feichter 2005; Lohmann et al. 2010)

Effects	Cloud type	Description	Forcing, W m^{-2}
First indirect aerosol effect ('Twomey effect')	All clouds	The more numerous smaller cloud particles reflect more solar radiation	-0.9 ± 0.4
Second indirect aerosol effect ('Albrecht affect')	All clouds	Smaller cloud particles decrease precipitation efficiency, prolonging cloud lifetime	Uncertain
Semi-direct effect	All clouds	Absorption of solar radiation by black carbon may cause evaporation of cloud particles	Uncertain
Glaciation indirect effect	Mixed ice and liquid clouds	More ice nuclei increase precipitation efficiency	Uncertain
Thermodynamic effect	Mixed ice and liquid clouds	Smaller cloud droplets delay the onset of freezing	Uncertain
Riming indirect effect	Mixed ice and liquid clouds	Smaller cloud droplets decrease riming efficiency	Uncertain
Total anthropogenic aerosol effect	All cloud types	Includes the above-mentioned indirect effects plus the direct aerosol effect	0 to -1.8

aerosol optical depths, and these are not valid for the pre-industrial values of droplet numbers.

Other effects mentioned in Table 24.2 are suggested to affect cloud lifetime. As clouds generally cool the climate by having a higher albedo than the Earth's surface, shorter cloud lifetime will warm the climate, while longer cloud lifetime will cool the climate. The second indirect aerosol effect (also referred to as the Albrecht effect) is, like the Twomey effect, based on the premise that the number of cloud droplets will increase with an increasing number of available cloud condensation nuclei. In contrast to the first indirect effect, the second indirect effect concerns the processes initiating precipitation. The onset of precipitation is sensitive to the formation of a few big droplets, also referred to as precipitation embryos (Albrecht 1989). Ice nuclei are similar precipitation embryos considered to be crucial for the onset of precipitation.

Black carbon is a strong light absorber and when enclosed in cloud droplets might cause evaporation and thus dissipate the cloud prematurely. This process is often referred to as the semi-direct effect. However, studies show that partial evaporation causes multiple effects as higher albedo due to smaller droplets and fewer giant droplets suppress precipitation both with a negative forcing on climate. A review by Koch and Del Genio (2010) indicated that the semi-direct effect is most likely to produce/account for a slight negative forcing that could be large enough to eliminate the direct warming of black carbon.

The glaciation effect refers to the formation of ice nuclei in cold clouds, that is clouds at least partially containing ice crystals or frozen droplets. In cold clouds, formation of ice crystals is important for the formation of precipitation. Anthropogenic emissions enhancing the number of good ice nuclei might then increase the precipitation probability and thus decrease cloud lifetime resulting in a positive radiative forcing. Black carbon particles have been suggested to be good ice nuclei, but other reports have concluded differently (Stratmann et al. 2010). Dust particles, on the other hand, are found to be important ice nuclei, while the fraction of anthropogenic dust is very difficult to estimate. Hoose et al. (2008) investigated the effect of assuming black carbon to have favourable ice nuclei properties and found that the positive increase in ice nuclei from black carbon particles was counteracted by dust particles losing their ice nuclei capability due to a coating of anthropogenic inorganic salts. These findings highlight the complexity of the processes and the need for a better understanding of ice nuclei properties and the key processes controlling cloud lifetime.

Other processes taking place within mixed ice and liquid water clouds are the thermodynamic effect and the riming effect, both connected to the competition between ice crystals and water droplets for condensing water in the cloud. The World Meteorological Organization International Aerosol Precipitation Science Assessment Group (Levin and Cotton 2007) concluded that observations and understanding of how precipitation is affected by pollution are still lacking. Rosenfeld et al. (2008) stated that large concentrations of man-made aerosols have been shown to affect precipitation and suggested that through different radiative and cloud-mediated processes, aerosols affect the thermodynamics driving the formation of clouds and precipitation. Stevens and Feingold (2009), however, recognised the inability to detect the specific effects of aerosols on precipitation and proposed that the difficulty in disentangling relationships among the aerosol, clouds and precipitation reflects the inadequacy of existing tools and methodologies as well as a failure to account for processes that buffer cloud and precipitation responses to aerosol perturbations.

In conclusion, it is clear that the indirect climate effects of anthropogenic atmospheric aerosols are not well known. Uncertainty in describing the indirect climate effect therefore dominates the uncertainty in total aerosol forcing.

24.4 Aerosol Influence on Regional Climate

Aerosols, including black carbon particles, and ozone are air pollutants that influence climate forcing and possibly also the hydrological cycle. Particles have a short lifetime in the atmosphere, typically two to four days, giving them a transport range of 1000–2000 km (Tunved et al. 2003). This means that the southern Baltic Sea area is dominated by anthropogenic emissions from the central European continent, while the northern Baltic Sea area is dominated by natural aerosol sources (such as the boreal forests). Particle concentration decreases with distance from the source, initially due to dilution and later due to sinks such as precipitation. However, secondary processes such as condensation of organic components and cloud chemistry add mass to the aerosols, increasing their influence on climate relevant processes and so extending the range of climate influence. Tropospheric ozone has a longer atmospheric lifetime of about a month thus giving it time to spread through the whole hemisphere.

The radiation effects of air pollution were measured directly in the Asian Brown Cloud Study and showed strong heating effects due to light-absorbing black carbon (Ramanathan et al. 2005). The simulated annual mean surface heat budget from 10 S to 30 N and 60 E to 100 E, that is over southern India and the Indian Ocean, shows progressive dimming since the 1930s. The dimming is about -15 W m^{-2}, meaning that 15 W m^{-2} less heat is reaching the ground and is instead trapped within the atmosphere by heat-absorbing aerosols. Besides redistributing heat within the atmosphere (affecting the surface and atmospheric temperature), evaporation of water and thus atmospheric water vapour concentration are also affected.

The heating and cooling effects of aerosols imply a complicated mix of interacting atmospheric processes that affect radiation and heat transfer which in turn affect regional temperature and the hydrological cycle.

European anthropogenic emissions of air pollutants have varied strongly over time, changing with population growth, major changes in land use and industrialisation. Emissions grew rapidly through the mid-twentieth century (Fig. 24.3) and in London were responsible for the Great Smog of December 1952 that resulted in many thousands of premature deaths. Banning the use of coal for residential heating and introducing high stacks for major industrial sources, both elements of the UK 1956 Clean Air Act helped to reduce local air pollution but in doing so caused problems elsewhere, particularly 'acid rain'. Acid rain—a potent mix of acidifying gases and particles—travelled long distances and when it eventually came down damaged surface water, groundwater and forests soils.

The adoption of the Convention on Long-range Transboundary Air Pollution (CLRTAP) in 1979 has resulted in a strong reduction in sulphur emissions such that present emissions are now back to pre-war levels (Fig. 24.3). Present emissions are now about 25 % of those in 1990. European emissions of black carbon decreased by about 40 % between 1960 and 2000 (Fig. 24.4).

Clearly, there have been very large variations in European emissions over the past 50 years, especially for sulphur, and this should be reflected in the regional climate if anthropogenic aerosols are indeed strong climate forcers, as has been suggested by the IPCC (2007) for example.

Using a coupled atmosphere-ocean climate model (GISS-ER) to investigate the response to forcing imposed over different latitude bands, Shindell and Faluvegi (2009) showed that the mid-latitudes and polar regions are sensitive to the source location. Although mid-latitude emissions mainly affect this region, they also affect the tropics. Furthermore, the authors also concluded that decreasing sulphur emissions and increasing black carbon have substantially contributed to the rapid warming observed in the Arctic.

Booth et al. (2012) using a state-of-the-art Earth system climate model (HadGEM2-ES) showed that aerosol emissions and periods of volcanic activity explain 76 % of the simulated multi-decadal variance in detrended 1860–2005 North Atlantic sea surface temperatures. The variation in sea surface temperature mainly depends on the indirect aerosol–cloud climate effects. The authors concluded with a strong

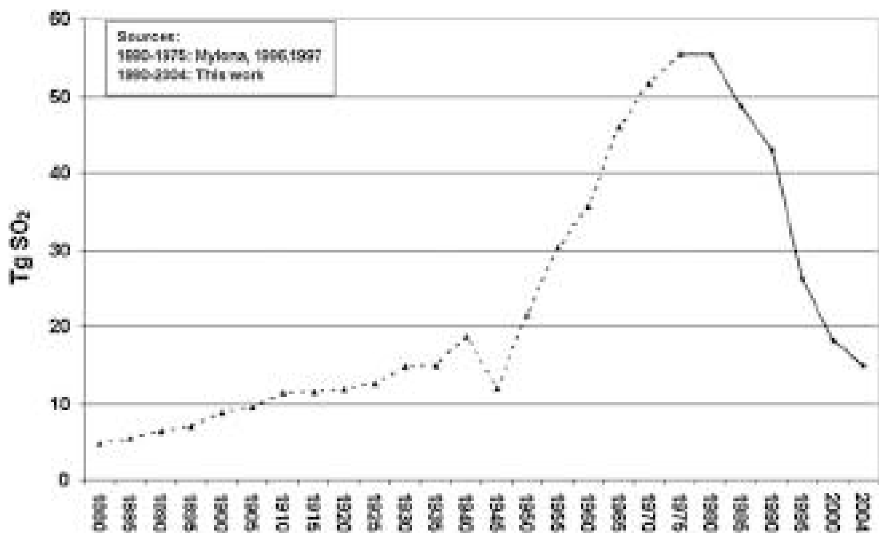

Fig. 24.3 Historical development of sulphur dioxide (SO_2) emissions in Europe (Vestreng et al. 2007)

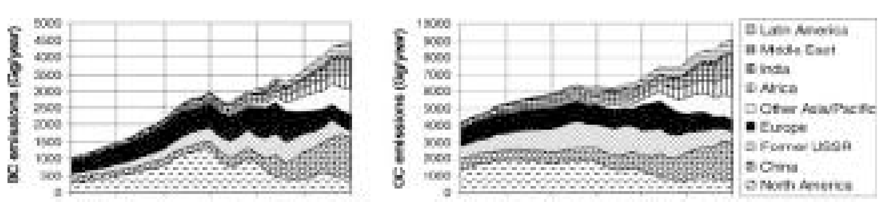

Fig. 24.4 Emissions of (*left*) black carbon and (*right*) organic carbon. Emissions are segregated by region (Bond et al. 2007)

statement based on their findings: '*Our findings suggest that anthropogenic aerosol emissions influenced a range of societally important historical climate events such as peaks in hurricane activity and Sahel drought. Decadal-scale model predictions of regional Atlantic climate will probably be improved by incorporating aerosol–cloud microphysical interactions and estimates of future concentrations of aerosols, emissions of which are directly addressable by policy actions*'.

Regional models allow higher spatial resolution than global models and a more detailed description of processes but need set boundary conditions specifying the influence of climate, meteorology and air components outside the model domain. Zubler et al. (2011) used a regional climate model that included interactive aerosol and cloud physics to investigate surface shortwave radiation in Europe over the past 50 years. This period is characterised by 'global dimming' until the mid-1980s and then 'brightening' and was fairly well described by the model in terms of the variation in aerosol emissions. The main cause of the brightening/dimming appeared to be the varying cloud fraction, that is the aerosol effect on cloud lifetime. Huszar et al. (2012) used an interactive coupling of a regional model (RegCM3) and a chemistry transport model (CAMx) to study the interaction between regional climate and air pollutants and found a perturbation of temperature in the range −1.5 to +1.5 °C due to aerosols and ozone. Interestingly, they did not find any correlation between forcing and induced temperature changes, which indicates the complexity of the climate system. However, they did identify methodological difficulties with regional models in applying realistic boundary conditions.

It seems possible that climate forcing by air pollutants strongly affects regional climate and thus global climate not only increasing global temperature but also causing changes in large-scale circulation. Even though the total anthropogenic forcing implies a warmer future climate in Europe, changing meteorology will also cause spatial variability. The main uncertainty in global climate models seems due to poor understanding of aerosol–cloud interactions. This uncertainty strongly affects the ability to make climate projections. Furthermore, there are still difficulties in achieving the necessary spatial resolution. Although the resolution of current regional climate models should be good enough, according to Huszar et al. (2012) and Zubler et al. (2011), there are methodological problems in adjusting the boundary conditions for changes in regional forcing on the general meteorology that in turn affect the boundary conditions.

24.4.1 Baltic Sea Area

The annual mean temperature in Sweden increased by about 0.9 °C between 1990 and 2005 (see climate records at www.smhi.se), which coincided with a major decrease in European sulphur emissions (see Fig. 24.3). This increase in temperature is about double that observed in the northern hemisphere over this period. The modelling studies discussed in Sect. 24.4 show that it is still difficult to determine to what extent the observed warming in Scandinavia is caused by global climate change and to what extent it is caused by the strong regional decline in aerosol concentrations. However, the Baltic Sea area is on one of the major outflow paths of air pollutants from the European landmass, and so it is feasible that a substantial fraction of the observed warming is due to the strong decline in sulphur emissions over the last 20 years. Black carbon emissions have also decreased over this period although substantially less than for sulphur. Some of the warming might have been offset by the decreasing black carbon.

Precipitation in Sweden seems to have increased between 1990 and 2005, but it is not statistically significant compared to 1960–1990 (see climate records at www.smhi.se). Current climate models have considerable uncertainties in describing precipitation and its dependence on aerosols, which complicates discussions on the influence of air quality on precipitation. A better understanding of aerosol–cloud interactions is needed for climate models to better simulate observations, not only for precipitation but also for aerosols, gases, clouds and other parameters affecting precipitation (see Chap. 10).

24.5 Climate Change Mitigation by Air Quality Mitigation

Air pollution contains components that both heat the climate (black carbon and ozone) and cool the climate (particulates). Emission abatement measures to mitigate air pollution, thus risk counteracting climate change mitigation. The UNEP Integrated Assessment of Black Carbon and Tropospheric Ozone (UNEP 2011) was performed on a mixture of available abatement measures (see Table 24.3) considered to be commonly available and possible to implement. The measures focus on black carbon and methane emissions, but co-emitted compounds such as organic carbon and ozone precursors are also considered in the analysis.

The impact of air pollutants on global climate and air quality was investigated using two different and well-established global climate models—ECHAM and GISS (Shindell et al. 2012). Besides the measures mentioned in Table 24.3, CO_2 abatement was assumed such that a maximum atmospheric concentration of 450 ppm of CO_2 would be reached, which assumes CO_2 emissions to stabilise by 2020 and then decrease to zero by 2080. The investigation was performed to evaluate the impact of black carbon and methane abatement measures in the context of a very

Table 24.3 Measures that improve climate change mitigation and air quality and that have a large emission reduction potential (Shindell et al. 2012; UNEP 2011)

Measure	Sector
Methane measures	
Extended premine degasification and recovery and oxidation of methane from ventilation air from coal mines	Extraction and transport of fossil fuels
Extended recovery and utilisation, rather than venting, of associated gas and improved control of unintended fugitive emissions from the production of oil and natural gas	
Reduced gas leakage from long-distance transmission pipelines	
Separation and treatment of biodegradable municipal waste through recycling, composting and anaerobic digestion as well as landfill gas collection with combustion utilisation	Waste management
Upgrading primary wastewater treatment to secondary/tertiary treatment with gas recovery and overflow control	
Control of methane emissions from livestock, mainly through farm-scale anaerobic digestion of manure from cattle and pigs	Agriculture
Intermittent aeration of continuously flooded rice paddies	
Black carbon technological measures (affecting BC and other co-emitted compounds)	
Diesel particle filters for road and off-road vehicles as part of a move to worldwide adoption of Euro 6/VI standards	Transport
Introduction of clean-burning biomass stoves for cooking and heating in developing countries[a,b]	Residential
Replacing traditional brick kilns with vertical shaft and Hoffman kilns	Industry
Replacing traditional coke ovens with modern recovery ovens, including the improvement of end-of-pipe abatement measures in developing countries	
Black carbon regulatory measures (affecting BC and other co-emitted compounds)	
Elimination of high-emitting vehicles in road and off-road transport	Transport
Ban on open burning of agricultural waste[a]	Agriculture
Substitution of clean-burning cook stoves using modern fuels (LPG or biogas) for traditional biomass cook stoves in developing countries[a, b]	Residential

[a] Partly motivated by its effect on health and regional climate including areas of ice and snow
[b] For cooking stoves, given their importance for black carbon emissions, two alternative measures are included

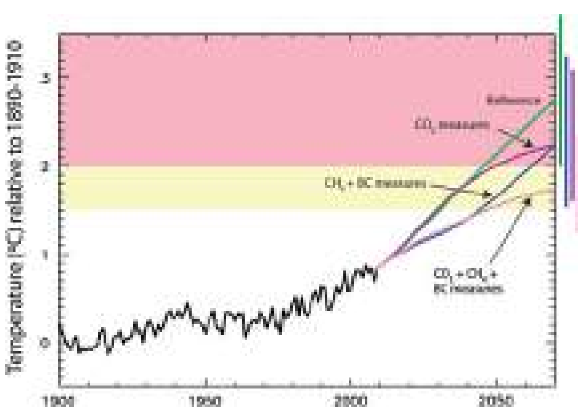

Fig. 24.5 Observed change in mean global temperature to 2009 and projected change in mean global temperature under various future scenarios, all relative to the 1890–1910 mean. *Shaded areas* show zones beyond 1.5 °C (*yellow*) and 2 °C (*pink*) (UNEP 2011)

ambitious CO_2 abatement programme. Figure 24.5 shows that abatement measures for black carbon and methane could reduce the projected global warming by 2070 by 0.5 °C. However, abatement of these compounds, often referred to as short-lived climate forcers, will only delay the rise in global warming if CO_2 emissions continue unchecked.

Abatement of CO_2 emissions is critical for limiting the rise in global temperature, but the implementation of abatement measures for black carbon and methane could also make a significant contribution. As the climate response to changes in the emissions of these short-lived climate forcers is relatively fast, reductions in the emission of black carbon and methane can be used to slow the immediate rise in global temperature and limit the level at which it would stabilise due to CO_2 emissions alone.

24.6 Conclusion

The recently revised Gothenburg Protocol to the Convention on Long-range Transboundary Air Pollution (CLRTAP) should ensure a decline in emissions of sulphur and nitrogen compounds over the next few years (see www.unece.org/env/lrtap). By 2020, emissions from the EU are required to have decreased by 59 % for SO_2, 42 % for nitrogen oxides (NO_X), 6 % for NH_3 and 28 % for volatile organic compounds (VOC) relative to 2005. Simultaneously, particle emissions ($PM_{2.5}$) must decrease by 22 %, with a voluntary emphasis on black carbon. Black carbon emissions will decrease considerably with the introduction of the new EU emission standards for vehicle and truck diesel engines (EURO 5 and EURO 6) that will take effect during the next 10–15 years as the transport fleet is gradually renewed. However, the major black carbon emission is from furnaces and stoves, for which abatement plans have not been introduced.

The major changes in emissions of sulphur and nitrogen compounds are thus already complete, while black carbon will probably decrease far more, as will organic compounds emitted from the same combustion sources. As a result, the total climate effect from the expected changes in aerosol emissions will probably be minor (Kulmala et al. 2011), while emission reductions strongly affecting ozone and methane concentrations may reduce climate warming significantly. However, ozone and methane concentrations over Europe and the Baltic Sea area are strongly dependent on emissions over the northern hemisphere and the world as a whole, respectively.

In conclusion, analyses on regional aerosol effects in northern Europe are rare and the commonly used regional climate models are mostly unable to simulate aerosol–climate interactions. However, recent modelling efforts investigating the influence of European aerosol emissions indicate an effect on large-scale circulation over Europe that is very likely to have affected the climate in the Baltic Sea region. To what extent is still not known. Development of the modelling capability and targeted analyses is urgently needed to reduce uncertainties related to the effect of changes in aerosol concentration on regional climate.

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Causes of Regional Change—Land Cover

25

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Abstract

Anthropogenic land-cover change (ALCC) is one of the few climate forcings for which the net direction of the climate response over the last two centuries is still not known. The uncertainty is due to the often counteracting temperature responses to the many biogeophysical effects and to the biogeochemical versus biogeophysical effects. Palaeoecological studies show that the major transformation of the landscape by anthropogenic activities in the southern zone of the Baltic Sea basin occurred between 6000 and 3000/2500 cal year BP. The only modelling study of the biogeophysical effects of past ALCCs on regional climate in north-western Europe suggests that deforestation between 6000 and 200 cal year BP may have caused significant change in winter and summer temperature. There is no indication that deforestation in the Baltic Sea area since AD 1850 would have been a major cause of the recent climate warming in the region through a positive biogeochemical feedback. Several model studies suggest that boreal reforestation might not be an effective climate warming mitigation tool as it might lead to increased warming through biogeophysical processes.

Keywords

land use • land cover • Holocene • land cover-climate interactions • climate forcing • Baltic Sea catchment area • Europe • northern hemisphere

25.1 Introduction

This chapter addresses several major questions. Did anthropogenic land-cover change (ALCC) occur in the Baltic Sea catchment during the last two centuries, and if so, did this play a role in the recent climate warming observed in the region? If not, did it have any other effect on climate? If recent ALCC occurred, is it unique in magnitude compared to ALCC before AD 1850 and back to the Neolithic time (6000 cal year BP—calendar years before present)? Did past ALCC have an effect on past climate?

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1_25

The chapter discusses the effects of ALCC on past climate (on timescales of decades, centuries and millennia) as well as future climate. It also reviews studies on natural climate-induced (potential) land-cover change (CLCC) and its feedbacks on climate, as it may help understanding of important processes involved in land cover–climate interactions. Here, land cover relates to vegetation cover, in particular tree cover versus low herb and low shrub vegetation, as well as snow cover. ALCC may be an external climate forcing, while CLCC is part of the climate system and may cause feedbacks on climate, whatever forcing is the cause of the initial climate change (Fig. 25.1). A feedback can be either positive (enhances the climate change responsible for the vegetation/land-cover change) or negative (mitigates the climate change). Similarly, a single forcing can enhance or mitigate the effect of other forcings; that is, the effect of ALCC may mitigate the effect of greenhouse gas emissions. The effects of ALCC or CLCC on climate (as forcing and feedbacks, respectively) are due to biogeophysical and biogeochemical processes at the boundary between vegetation and the atmosphere (Findell et al. 2007). When attributing causes of regional climate change at the scale of the Baltic Sea area, the biogeophysical effects are of particular interest since they exert a direct, measurable effect on regional climate. Biogeochemical effects are more relevant in the context of global climate change since the timescale of carbon dioxide (CO_2) mixing in the atmosphere is very short. Consequently, regional changes in the carbon balance affect regional climate only indirectly by affecting the global CO_2 concentration. Therefore, this chapter focuses primarily on biogeophysical mechanisms, although biogeochemical processes are also described.

Sensitivity studies with global Earth System Models have increased our understanding of interactions between land cover and climate over the past decade (IPCC 2007). However, the mechanisms involved in biogeophysical feedbacks are mainly regional to local in scale; therefore, use of regional climate models and vegetation models should potentially provide better insights on those feedbacks. Nonetheless, few published studies have used regional climate models, and none were specifically designed to evaluate the effects of ALCC on past, present and future climate change at the scale of the Baltic Sea basin. There are also very few attribution studies using global climate models and focusing on ALCC as a possible forcing at both global and regional scales. To date (2013), there is a single study on the effect of past ALCC on climate change at 6000 and 200 cal year BP in north-western Europe using a regional climate model (Gaillard et al. 2010; Strandberg et al. 2013). Therefore, current understanding of the role of ALCC in regional climate change at the scale of the Baltic Sea basin, and in particular as a possible forcing of the warming of the last two centuries, must rely primarily on studies at the European or northern hemisphere scale using global climate models. The largest model study to date focusing on the impacts of ALCC on the climate of the northern hemisphere is that within the Land-Use and Climate, Identification of

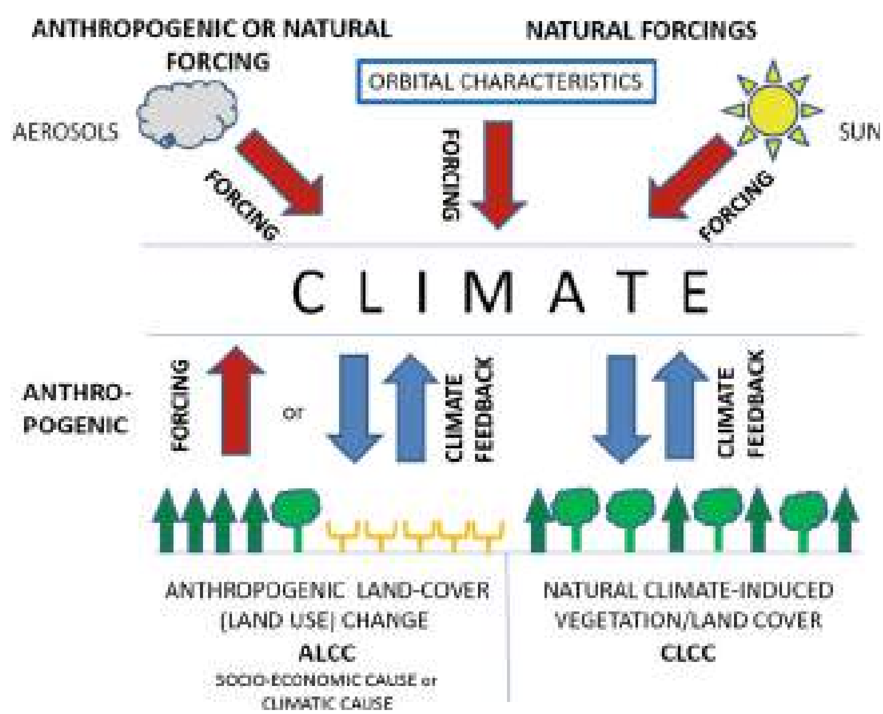

Fig. 25.1 Schematic illustration of interactions between land cover and climate: biogeochemical and biogeophysical effects of climate-induced and human-induced changes in land cover, feedbacks and forcings. CLCC (climate-induced land-cover change, i.e. change in natural (potential) vegetation); ALCC (anthropogenic land-cover change, i.e. change in human-induced vegetation due to agricultural activities)

Robust Impacts (LUCID) project (Pitman et al. 2009; de Noblet-Ducoudré et al. 2012). This was set up to study the robustness of modelled biogeophysical impacts of historical ALCC on climate (roughly AD 1850 to modern time).

This chapter first explains the processes involved in land cover–climate interactions and summarises the results from modelling studies investigating these processes (Sects. 25.2.1 and 25.2.2). It then reviews current understanding of how past and recent land-cover change, both ALCC and CLCC, might have influenced regional climate in the Baltic Sea region (Sects. 25.2.3 and 25.4). As none of the chapters in the present assessment of climate change in the Baltic Sea basin deals with Holocene ALCC, this chapter also reviews studies on ALCC since Neolithic time (about the last 6000 years) (Sect. 25.3). Finally, the chapter discusses the possible effects of future resource management on land cover and, as a result, on biogeochemical and biogeophysical processes and future climate (Sect. 25.5). Holocene CLCC is presented in Chap. 2.

'Climate change' refers to systematic changes in response to external forcing, in accordance with Chaps. 23 and 24. In order to avoid any confusion of concepts, 'Land cover–climate interaction' and 'biogeophysical (or biogeochemical) effect' refer to forcing mechanisms, while the term 'feedback' is used only for the effects of CLCC on climate (i.e. as part of the climate system). See Chap. 2 for a description of the relationships between climate change and natural vegetation during the Holocene, and Chaps. 16 and 21 for a complete account of the influence of recent climate change on vegetation.

25.2 Land Cover–Climate Interactions: What Are They and What Are the Mechanisms Involved?

25.2.1 Biogeophysical Effects/Feedbacks

Biogeophysical effects (Fig. 25.2) influence physical exchange fluxes and the energy balance between the atmosphere and land surface. The major biogeophysical feedbacks are due to (i) land-surface characteristics such as albedo (referred as the *albedo effect*; albedo is the surface reflectivity with respect to short-wave radiation) and roughness (e.g. smooth snow or rough forest), and (ii) evapotranspiration (the sum of transpiration from plant stomata and evaporation from other water sources at or below ground) (Levis 2010).

Albedo is the proportion of incoming solar radiation reflected by a surface. It strongly influences the energy available for absorption by the land surface. The greatest contrast in albedo occurs between open and forested land, especially in the presence of snow. While snow is completely exposed on open land, it is partly covered in a forested area. This is often referred to as the *snow-masking effect*. Snow masking can cause a positive feedback on climate. For example, a high-latitude northwards expansion of trees and shrubs (low albedo) due to warming will hide the snow (high albedo) on the ground and thus increase the absorption of solar radiation, which will in turn enhance the warming and lead to a further northwards expansion in tree cover, leading to further warming (Fig. 25.2). This type of positive feedback is especially strong when the forest comprises evergreen conifers that retain their needles during winter. Modelling experiments have shown that the albedo effect can be significant (e.g. Bala et al. 2007; Liang et al. 2010).

Vegetation also influences the hydrological cycle. Structural changes in vegetation, such as changes in leaf area index (LAI), roughness length, and rooting depth modify the evapotranspiration of water from the land surface. The LAI represents the amount of leaf material present in an ecosystem and is geometrically defined as the total one-sided area of photosynthetic tissue (in m^2) per unit ground surface area (in m^2). Surface roughness (often just referred to as 'roughness') is a measure of the texture of a surface. The roughness length (expressed in m) depends on the frontal area of the average element (e.g. trees in a forest) facing the wind divided by the ground width it occupies. For instance, the roughness of featureless terrain is 0.005 m (smooth), flat terrain with grass or very low vegetation 0.03 m (open), and mature forest 1.0 m (closed) (Davenport et al. 2000). While the LAI influences the amount of intercepted water and the partitioning of energy fluxes into sensible and latent heat, the roughness length affects the turbulent mixing of heat into the atmosphere. The rooting depth determines the amount of water extracted from the soils by the vegetation; a deeper and/or more extensive root system will enhance the ability of the vegetation to extract soil water. In environments where neither temperature nor water limits vegetation growth, the vegetation tends to flourish, which increases both LAI and roughness. Since vegetation transpires water through leaf stomata, an increase in LAI will be associated with increasing evapotranspiration and, as a result, an increase in latent heat at the expense of sensible heat. Sensible heat warms the atmosphere close to the vegetation surface, whereas latent heat is stored in the released water vapour and warms the atmosphere only when condensation occurs, some distance away from the vegetation and higher up. Therefore, the *hydrological cycle effect* has a dampening effect on local to regional temperature change since stronger evapotranspiration implies that more energy is required to vaporise water (Fig. 25.2). An increase in woodland cover due to climate warming or anthropogenic activities (forest planting) will also increase the land-surface roughness and, in turn, enhance the moisture convergence and lead to increased precipitation. Higher water availability triggers a

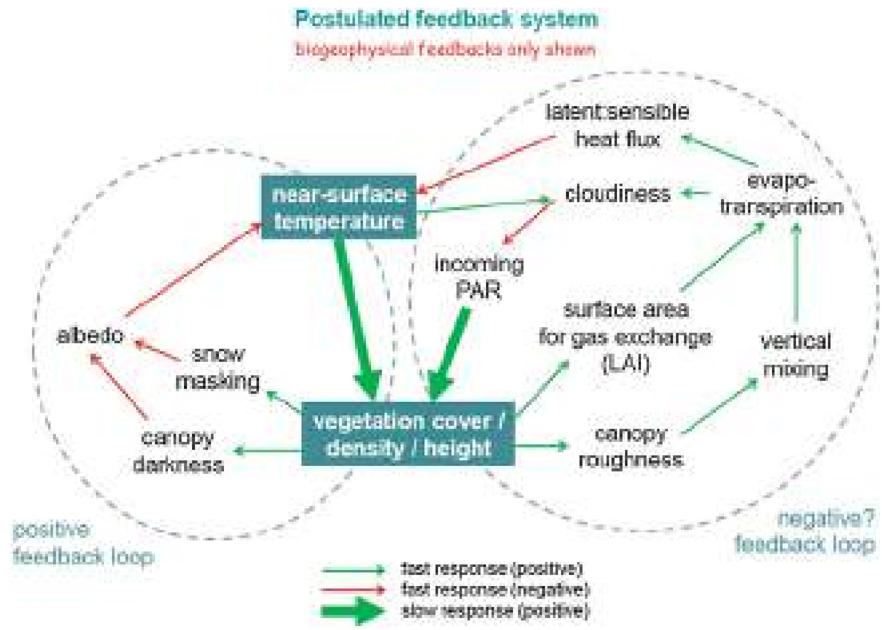

Fig. 25.2 A simplified scheme of the biogeophysical feedback system (after Ben Smith, unpublished). *LAI* leaf area index; *PAR* photosynthetically active radiation

positive *precipitation effect/feedback* by producing higher vegetation density and a further increase in land-surface roughness and precipitation, etc. Higher atmospheric CO_2 concentrations may also cause an increase in vegetation density (the *fertilisation effect*, see Sect. 25.2.2) and produce similar secondary biogeophysical effects such as the *precipitation effect* through an increase in roughness. As pointed out by Levis (2010), these feedbacks (Fig. 25.2) can all be modified or eliminated by ALCC. Moreover, it is rare that a single feedback dominates or is the only one active. Several feedbacks often occur together, which increases the difficulty of interpreting the results.

The study by Zhang (2011) is one of the few investigations using observations that focus on the Baltic Sea region. The study was based on precipitation and run-off data for 1961–2003 in southern and central Sweden and showed a trend towards increased evapotranspiration. As the major cause of the increased evapotranspiration was increased winter evaporation, the author proposed this may be related to a known land-use change in the study area, namely the replacement of deciduous trees (lose their leaves in winter) by planted coniferous forest (with evapotranspiration from needles in winter, also from intercepted water, i.e. rainfall collected on the needles).

25.2.2 Biogeochemical Effects/Feedbacks

The land surface plays a major role within the global carbon cycle: vegetation takes up atmospheric CO_2 through photosynthesis and uses the carbon to build biomass, while the oxygen is released to the atmosphere; some time later, the vegetation dies and dead biomass builds up soils; the soil organic matter is then decomposed by micro-organisms and the resulting CO_2 released to the atmosphere, thus closing the cycle. However, disturbance processes such as forest and grassland fires, a climate-induced decrease in woodland and anthropogenic deforestation will also release carbon to the atmosphere. The land surface contains significant amounts of carbon in vegetation (350–550 PgC, Prentice et al. 2001) and in soils (1500–2400 PgC, Batjes 1996). Additional carbon is stored in wetlands (200–450 PgC) and in the loess soils of permafrost areas (200–400 PgC, McGuire et al. 2009).

Owing to the general character of the Baltic Sea region with its extensive forests and substantial wetland areas, the carbon storage in vegetation and soils in the region is undoubtedly significant, although no specific regional estimates appear to be available. Both humans and climate may have a significant impact on this carbon storage. The carbon balance of the land surface depends primarily on the atmospheric CO_2 concentration and temperature. In carbon cycle models, such as those used in the Coupled Climate Carbon Cycle Model Intercomparison Project (C4MIP, Friedlingstein et al. 2006), the sensitivity of the carbon cycle to climate change can be expressed by two parameters, β and γ. β describes the sensitivity to changes in atmospheric CO_2 concentration, while γ describes the sensitivity to changes in climate, especially temperature. Vegetation experiments with elevated CO_2 concentrations provide observational evidence of enhanced net primary productivity (NPP) under increased atmospheric CO_2 (Norby et al. 2005), implying a positive β. While the experiments give a direct indication of feedback between CO_2 concentration and CO_2 uptake, i.e. the *fertilisation effect*, there is still much uncertainty about the universality of the results, especially since interactions with nutrient and water availability are likely but remain difficult to quantify (Gedalof and Berg 2010). Nevertheless, elevated CO_2 concentrations probably do enhance productivity, as long as other conditions for additional growth are met. Outputs from land-surface models (LSMs) show an increase in carbon storage under increased atmospheric CO_2 of 0.85–2.4 PgC ppm v^{-1} in early studies (Cramer et al. 2001), while later studies that consider the limitation of carbon uptake by nitrogen availability show a considerably decreased enhancement (Sokolov et al. 2008; Thornton et al. 2009; Zaehle et al. 2010). In the LSM CLM4, for example, the estimated increase in NPP when considering nitrogen availability is only 30 % of the increase without considering nitrogen dynamics (Bonan and Levis 2010).

Through the temperature sensitivity of both photosynthesis and respiration, the terrestrial carbon balance is also strongly influenced by changing temperature, although the precise response is not well known. Under water-limited conditions, an increase in temperature would lead to stronger water stress due to enhanced evapotranspiration. In contrast, an increase in temperature in cold regions would lead to a longer growing season, thereby enhancing vegetation growth. With respect to soil organic matter, an increase in temperature will lead to increased decomposition, that is enhanced carbon losses to the atmosphere (Davidson et al. 2006). Modelling studies suggest that warming will accelerate carbon losses from soils, implying a positive feedback between warming and the carbon cycle. Friedlingstein et al. (2006) found a range of −20 to −177 PgC per °C for the γ factor and Sitch et al. (2008) of −60 to −198 PgC per °C. However, the models used did not consider nutrient limitation and may have overestimated γ, since warming may increase nitrogen mineralisation and availability in soils, enhancing vegetation growth. Current climate–carbon cycle models including a nitrogen cycle show this effect (Sokolov et al. 2008; Thornton et al. 2009; Zaehle et al. 2010), but the uncertainties in γ remain very high.

25.2.3 Impact of Hypothetical Land-Cover Change on Climate: Climate Model Simulations

Ban-Weiss et al. (2011) studied climate forcing and response to idealised changes in surface latent and sensible heat. They found that globally adding a uniform 1 W m^{-2} source of latent heat flux along with a uniform 1 W m^{-2} sink of sensible heat leads to a decrease in global mean surface air temperature of 0.54 ± 0.04 °C, explained mainly by an increase in planetary albedo associated with an increase in low-elevation cloudiness caused by increased evaporation. The model results indicate that, on average, when latent heating replaces sensible heating, global and local surface temperatures decrease. Kvalevåg et al. (2010) used GCM (general circulation model) simulations to compare impacts on climate due to vegetation and albedo changes together or to albedo changes only; they concluded that effects due to changes in albedo dominate in temperate regions. The authors also claimed that divergent conclusions between similar studies are probably due to differences in specifications of albedo. Sensitivity to albedo was also shown in a model experiment where a hypothetical boreal forest expansion, decreasing the surface albedo, led to an enhancement of the summertime Arctic frontal zone and a strengthening of the jet (Liess et al. 2011). Boreal forests are characterised by lower albedo and a higher Bowen ratio (the ratio of sensible to latent heat fluxes, i.e. heat loss or gain) for similar levels of soil water availability than temperate forests (Bonan 2008). Thus, replacing boreal forests by grassland results in a cooling effect due to a decrease in both the Bowen ratio (as long as soil water is available) and net radiation (increase in albedo); the cooling effect may become higher than the warming effect of increased carbon emissions.

Eliseev (2011) showed that, at the global scale, changes in surface albedo due to the replacement of natural vegetation by agricultural land would have a greater influence on the available energy at the surface by absorbed short-wave radiation than the influence of the albedo effect. This is explained by relatively low insolation during winter at the latitudes characterised by the snow-masking effect of forest vegetation. Moreover, Cook et al. (2008) showed that feedback mechanisms including interactive vegetation and snow may be very sensitive to the parameterisation of the snow fraction. For instance, a fast-growing snow fraction produced a large-scale southward retreat of boreal vegetation and a widespread cooling. Bathiany et al. (2010) found that afforestation of all currently treeless areas north of 45°N would lead to a global mean warming of 0.26 °C due to biogeophysical effects, while the reduction in atmospheric CO_2 would only be 6.5 ppm, leading to a net warming. The albedo effect would be most significant in winter and spring when forests mask snow, causing an additional regional temperature rise. Earlier, similar studies of idealised, large-scale deforestation also found that an albedo cooling would dominate over CO_2 warming in boreal regions, indicating that boreal reforestation would probably not be an effective mitigation tool in such areas (Betts 2000; Claussen et al. 2001; Sitch et al. 2005; Bala et al. 2007).

25.3 Reconstructing Past Land-Cover Change

This section presents the various methods available to reconstruct past land-cover and their changes through time and space and describes the major ALCCs in the Baltic Sea region over about the last 6000 years, from the Neolithic (start of agriculture) until modern time. All ages are given in calibrated ^{14}C years (or calendar years) BC (Before Christ)/AD (Anno Domini = after Christ) or BP (Before Present; present = AD1950).

25.3.1 Methodology

Attempts to reconstruct past changes in land cover have been based on two major approaches: (i) interpretation of palaeoecological data, fossil pollen in particular, and (ii) use of land use and population historical records as well as archaeological records of past settlements.

Estimates of human-induced changes in land cover based on historical records, remotely sensed images, land census and modelling (Ramankutty and Foley 1999; Olofsson and Hickler 2008; Klein Goldewijk et al. 2011) were used to provide first insights into the effects of ALCC on past climate (e.g. Brovkin et al. 2006; Olofsson and Hickler 2008). The most frequently used database in climate modelling to date is the History Database of the Global Environment (HYDE) database (Klein Goldewijk et al. 2011). However, its estimates of anthropogenic land cover during key periods of the past show large discrepancies with more recently developed scenarios of ALCC by Pongratz et al. (2008), Lemmen (2009), and Kaplan et al. (2009) (see review in Gaillard et al. 2010; Fig. 25.3).

Pongratz et al. (2008) estimated the extent of cropland and pasture since AD 800 based on published maps of agricultural areas for the past three centuries and, for earlier times, a country-based method using population data as a proxy for agricultural activity. The resulting map of agricultural land was then combined with a map of potential vegetation. One of the strengths of the study is that the uncertainties associated with the approach were quantified, in particular those relating to the estimates of technological progress in agriculture and size of human populations. These ALCC scenarios were produced at a very high time resolution and used in modelling studies (see Sect. 25.4).

Lemmen (2009) developed an independent estimate of human population density, technological change and agricultural activity during the period 9500–2000 BC based on dynamical hindcasts of socio-economic development (GLUES, Global Land Use and technological Evolution Simulator; Wirtz and Lemmen 2003). The population density estimate was combined with per capita crop intensity from HYDE (version 3.1) to infer areal demand for cropping at an annual resolution in 685 world regions. Comparison of the simulated crop fraction estimate with the HYDE estimate showed large discrepancies attributed to missing local historical data in HYDE (Lemmen 2009; Gaillard et al. 2010).

Kaplan et al. (2009) created a high-resolution, annually resolved time series of anthropogenic deforestation in Europe over the past 6000 years (referred as KK10 scenarios) by using (i) a model of the forest cover–human population relationship based on estimates of human population for the period 1000 BC to AD 1850, and (ii) a model of land suitability to cultivation and pasture, assuming that high-quality agricultural land was cleared first and marginal land next. Alternative scenarios of deforestation were also produced by taking into account technological developments, which led to major differences in south-western, south-eastern and eastern Europe (Fig. 25.3). The Kaplan et al. (2009) KK10 scenarios are also different from the HYDE database (Fig. 25.3) and provide estimates of deforestation in Europe around AD 1800 that compare better to historical accounts than the HYDE scenarios (Gaillard et al. 2009; Krzywinski and O'Connell 2009). They are also closer to pollen-inferred land-cover change over the past 6000 years (see Sect. 25.3.3 and Figs. 25.4, 25.5, 25.6 and 25.7; Gaillard et al. 2010; Trondman et al. 2011, 2012). This implies that previous attempts to quantify anthropogenic perturbation of the Holocene carbon cycle based on the HYDE and Olofsson and Hickler's databases may have underestimated early human impact.

The second approach for reconstructing past land-use changes relies on quantifying and synthesising records of land-cover change based on palaeoecological proxy data. Such proxy-based reconstructions complement model-based scenarios of ALCC and are essential to evaluate and improve their reliability.

Objective long-term records of the inferred past changes in vegetation cover are limited. Palaeoecological data, particularly fossil pollen, have been used to approximate past vegetation changes at sub-continental to global scales (e.g. Prentice and Jolly 2000; Tarasov et al. 2007; Williams et al. 2008). However, these studies have focused on tree vegetation and are of little use for a quantitative assessment of human impacts on land cover (Anderson et al. 2006; Gaillard et al. 2010). They did not resolve problems related to (i) the non-linearity of pollen–vegetation relationships in percentages, (ii) the definition of the spatial scale of vegetation represented by pollen and (iii) the differences in pollen productivity between plant taxa (e.g. Sugita et al. 1999; Gaillard et al. 2008; Gaillard 2013). A new framework of vegetation reconstruction was recently developed that resolves these problems: the Landscape Reconstruction Algorithm (LRA) (Sugita 2007a, b). This consists of two separate models, Regional Estimates of VEgetation

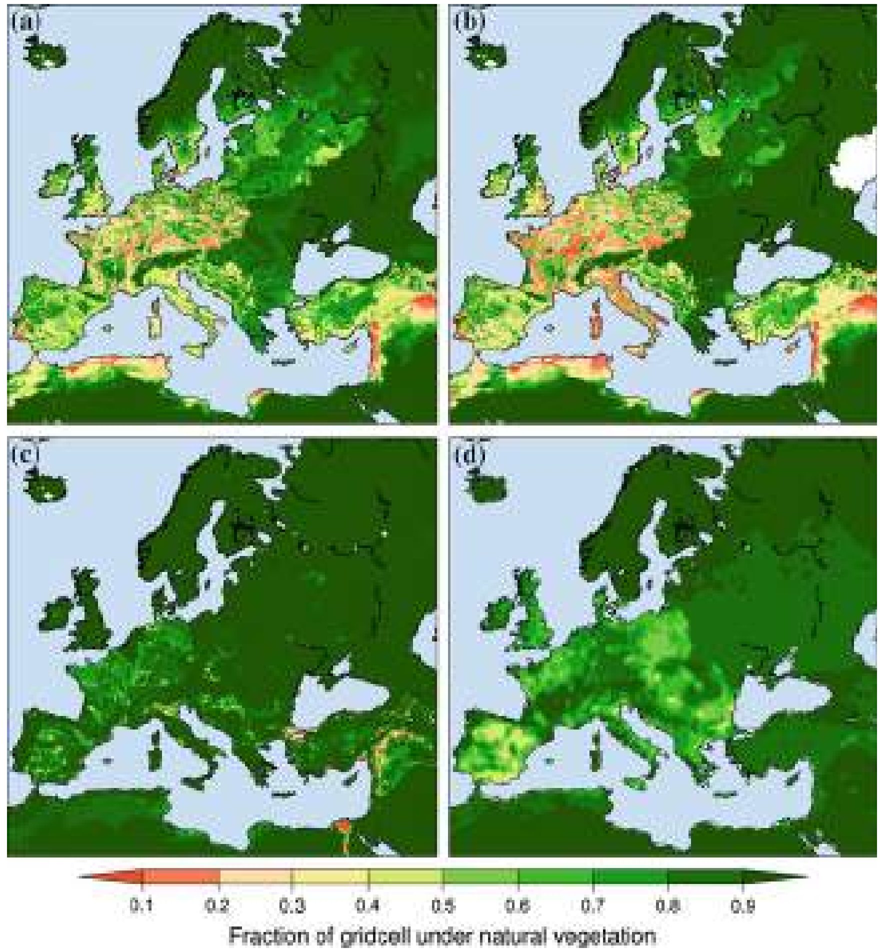

Fig. 25.3 Anthropogenic land use in Europe and surrounding areas at AD 800 simulated by four modelling approaches: **a** Kaplan et al. (2009) standard scenario; **b** Kaplan et al. (2009) technology scenario; **c** the HYDE 3.1 database (Klein Goldewijk et al. 2011); **d** Pongratz et al. (2008) maximum scenario. From Gaillard et al. (2010)

Abundance from Large Sites (REVEALS) and LOcal Vegetation Estimates (LOVE), allowing vegetation abundance to be inferred from pollen percentages at the regional (about 100 × 100 km) (REVEALS) and local spatial scales (LRA: REVEALS + LOVE), respectively. The minimum size of the area for which LRA reconstructions of local vegetation are valid is calculated by the LOVE model and varies between sites; it usually has a radius of about 0.5–3 km in southern Scandinavia (Sugita et al. 1999; Hellman et al. 2009; Fredh 2012). The LOVE model requires estimates of regional vegetation obtained using the REVEALS model. Extensive simulations support the theoretical premise of the LRA (Sugita 1994, 2007a, b). In Europe, REVEALS was empirically tested in southern Sweden (Hellman et al. 2008) and Central Europe (Soepboer et al. 2010), and the LRA (REVEALS + LOVE) in Denmark (Nielsen and Odgaard 2010; Overballe-Petersen et al. 2012) and in southern Sweden (Cui et al. 2012; Fredh 2012). The LRA approach is, to date, the best method for inferring anthropogenic land cover from pollen data (Gaillard 2013). *Human-impact pollen indicators* such as cereals, other cultivated plants, weeds and other plants favoured by human activities and cattle grazing are used widely to interpret pollen records in terms of human-induced vegetation types (such as cultivated land, fresh/dry meadows and pastureland, ruderal land) applying the *indicator species approach*. However, these interpretations can only be qualitative. Pollen-inferred reconstruction of past human impact on vegetation is often complemented by information from plant macroremains (seeds, fruits, leaves etc.), insect remains and archaeological/historical data (e.g. Greisman and Gaillard 2009; Olsson and Lemdahl 2009, 2010).

To avoid confusion below, reconstructions of the regional vegetation are referred to as REVEALS reconstructions and reconstructions of the local vegetation as LRA reconstructions.

25.3.2 Major Past Land-Use Types in the Baltic Sea Region

The major cultural landscape/land-use types in the Baltic Sea region in the past were wood-pasture, coppices and pollards (trees cut in various ways to obtain wood for daily needs and fodder for cattle, respectively), slash-and-burn cultivation,

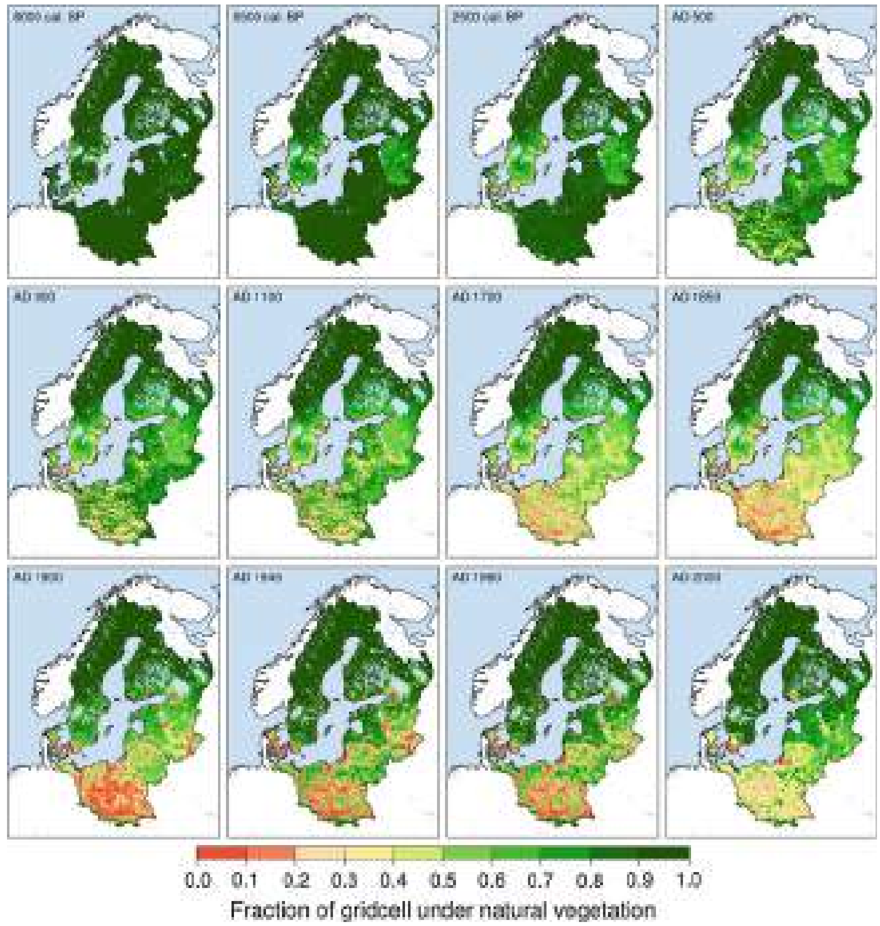

Fig. 25.4 Anthropogenic land-cover change in the Baltic Sea basin over the Holocene. The time trend of increasing population and agricultural land use can be seen in this series of maps covering the past 8000 years. KK10 scenarios extracted for the Baltic Sea area from Kaplan et al. (2009)

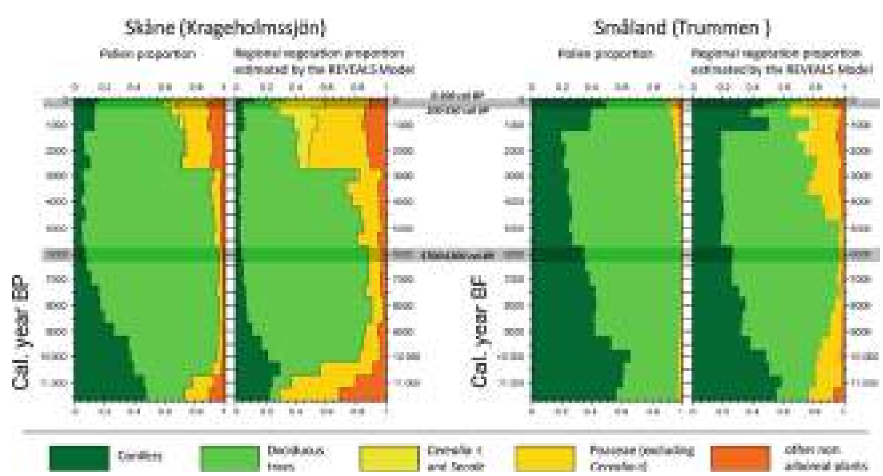

Fig. 25.5 REVEALS estimates of conifers, deciduous trees, cereals, grasses and herbs in southern Sweden, provinces of Skåne and Småland (Gaillard et al. 2010)

cultivated fields, grasslands and meadows (hay making and grazing), heathlands and summer farms (transhumance) (e.g. Gaillard et al. 2009).

Although human impact on the environment of the Baltic Sea area (and Europe in general) began in the Mesolithic (pre-Neolithic, before 6000 BP in NW Europe), it is generally accepted that farming cultures were responsible for the first major impact on European natural environments. The pre-Neolithic lowland European landscapes are generally assumed to have been densely forested, but open land undoubtedly existed in areas where soil conditions did not allow the development of dense forests (Svenning 2002) and might also have occurred through grazing by large herbivores (e.g. Vera 2000), fire (e.g. Olsson et al. 2010; Svenning

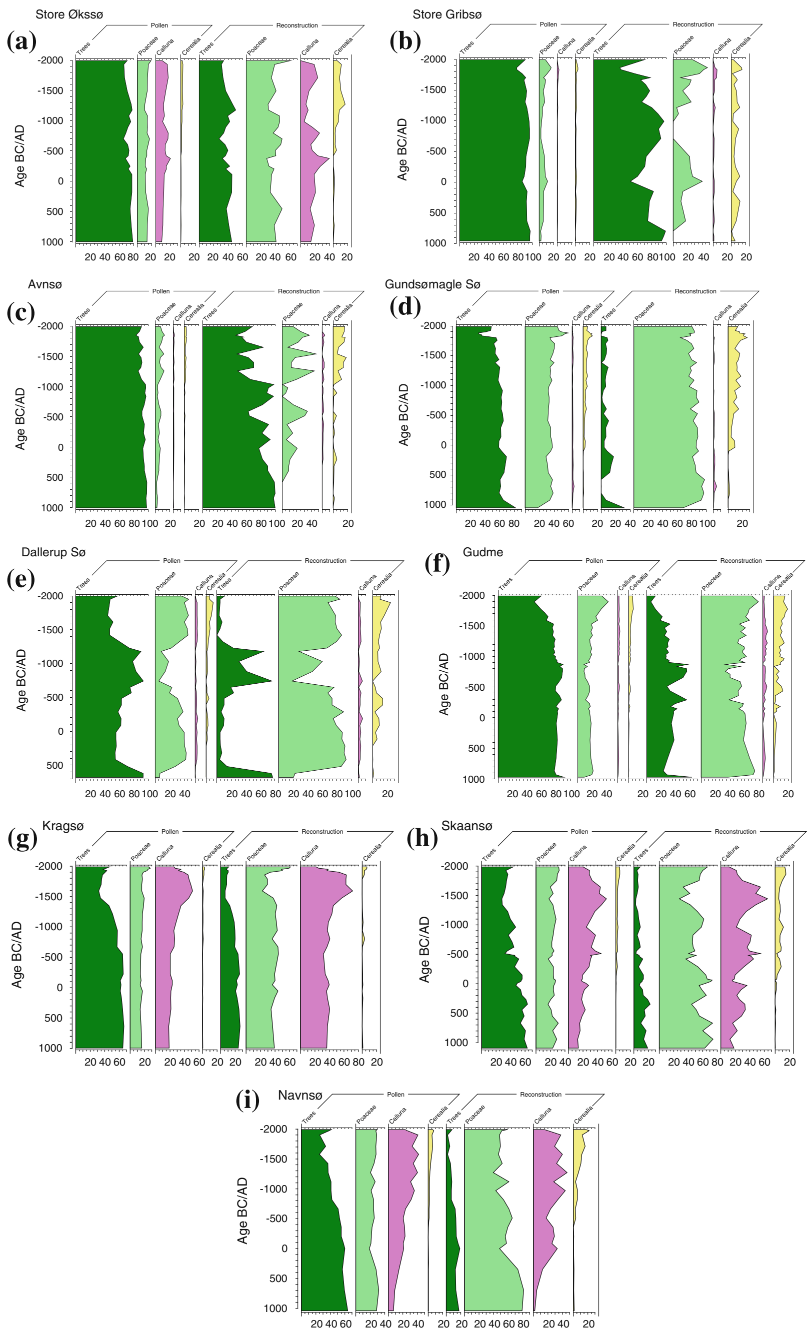

Fig. 25.6 Landscape openness in Denmark during the past 3000 years (2700 years at Dallerups Sø as modelled by the Landscape Reconstruction Algorithm (LRA)) (Sugita 2007a, b) using nine pollen records from small lakes distributed in the three major contrasting regions of the country. Pollen percentages on the *left* of each plot, LRA estimates of plant cover on the *right*. The ages are given in calibrated years BC/AD. Landscape openness is expressed by the LRA estimated cover (in percentages) of herbs (here only Gramineae (grasses, light green) and Cerealia (cereals, yellow) are shown) and low shrubs (here only Calluna (heather, violet) is shown). The LRA estimates of tree cover (*dark green*) and all herb and low shrubs (not shown here) sum up to a total of 100 %. The trees are overrepresented in the pollen percentages, while Gramineae and Cerealia (and herbs in general) are under-represented in the pollen percentages; that is, landscape openness is under-represented in the pollen percentages. The latter is true for all LRA reconstructions performed in Europe so far. In this case, heather is slightly under-represented in the pollen percentages, which is not always the case and depends on the overall vegetation composition. From Nielsen and Odgaard (2010)

2002), and the activities of Mesolithic people. However, since the beginning of the Neolithic, deforestation was a prerequisite to sow crops. Domesticated animals including cattle, sheep and pigs were introduced and also contributed to opening up the landscape. A summary of knowledge concerning the major land uses in the Baltic Sea catchment area

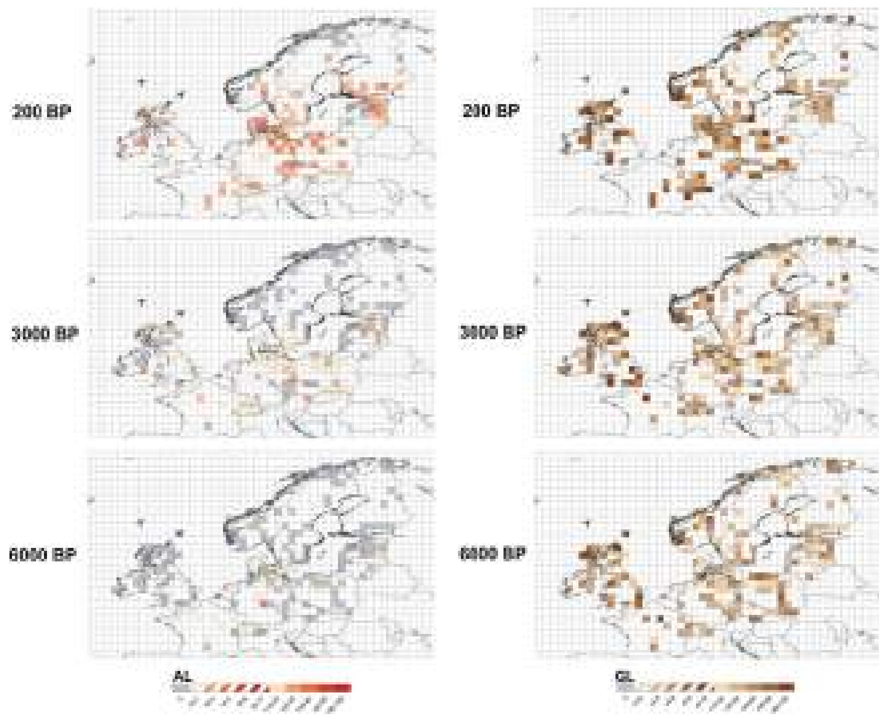

Fig. 25.7 REVEALS estimates of grassland and arable land in north-western Europe at 6000 BP, 3000 BP, and 200 BP after Trondman et al. (2011, 2012). First-generation LANDCLIM maps (produced by Anna-Kari Trondman, Florence Mazier, Anne Birgitte Nielsen, Ralph Fyfe and LANDCLIM members for the purpose of this chapter; see Gaillard et al. (2010) for a description of the LANDCLIM project). *Left* agricultural land {AL = total Cereals [Cerealia undiff., *Triticum* (wheat) type, *Avena* (oats) type, *Hordeum* (barley) type, Secale cereale (rye)]}. *Right* Grassland [GL = Cyperaceae, Filipendula, *Plantago lanceolata*, *Plantago montana*, *Plantago media*, Poaceae, *Rumex* p.p. (mainly *R. acetosa* and *R. acetosella*, i.e. *Rumex acetosa* pollen-morphological type)]. This first generation of LANDCLIM REVEALS estimates is based on all available Holocene pollen records from small and large sites (lakes and bogs) including all or part of the time windows 6000, 3000 and 200 BP (1950), and with ≥3 dates for the chronological control. These pollen records were collected from the European (EPD), Alpine (ALPADABA) and Czech (PALYCZ) pollen databases (van der Knaap et al. 2000; Fyfe et al. 2007; Kuněs et al. 2009, respectively). The pollen productivity estimates (PPEs) used are, for each taxon, the mean of all PPEs obtained within the project area (north-western Europe and western Europe north of the Alps, Broström et al. 2008). From Gaillard (2013)

over the last 6000 years follows. For more complete reviews and bibliography on the subject, see in particular Behre (1988), Berglund (1991) and Gaillard et al. (2009).

Wood-pasture was of major importance until late mediaeval times. Coppices and pollards can be traced back through written sources to the sixteenth century in north-western Europe. There is evidence from archaeological contexts for coppicing and pollarding as far back as the Mesolithic and Neolithic, respectively. Coppices and pollards were progressively abandoned during the twentieth century and replaced by open pastures and cultivated fields or developed into secondary, broadleaved forests. Slash-and-burn is also a form of woodland use. Cereals (often rye) were sown in clearings created by felling, drying and burning the woody vegetation, which enriched poor soils with ash. In Finland, slash-and-burn started with arable farming in the Neolithic and lasted until the early twentieth century. In Sweden, slash-and-burn was mainly associated with spruce forests and was common after the expansion of spruce from the north (3000–1000 BP). It was practiced so intensively in the eighteenth and nineteenth centuries that woodlands failed to fully regenerate. For that reason and the associated fire hazard, slash-and-burn was prohibited in many parts of Scandinavia in the early twentieth century.

Pastoral activity has had a fundamental influence on the landscape and vegetation of the Baltic Sea region since the Neolithic. The major factor involved in the formation and maintenance of *pastures* and *meadows* was the need for fodder. Denser settlements, rising populations and increased demand for food resulted in an increase in livestock, which in turn demanded more grazing land and meadows. The practice of hay making brought about the development of the *infield/outland* system that optimised the available land resources thanks to an efficient regime for nutrient recycling. The infield included cultivated fields and hay meadows, while the outland was grazing land (grassland, heaths and/or forest). Livestock was also necessary to fertilise soils for cultivation, and the hay meadows had to be large enough to sustain the amount of fodder necessary for the livestock required to fertilise the area of crop fields that would cover the food demand of the human population. In other words, the hay meadows were essential to the crop fields and their size had to be three to 20 times larger than the cultivated area depending on the soil conditions (Fogelfors 1997). Hay making is often associated with the introduction of the scythe in the Late Iron Age (ca. AD 1000). However, species-rich hay meadows may have already existed in southern Sweden at the end of the Late Bronze Age/beginning of the Early Iron Age (from ca. 2600 years BP; Gaillard et al. 1994). Mowing (hay making) probably started with the practice of stalling that may have first occurred in connection with a climate cooling in north-western Europe dated to ca. 3000 BP (e.g. Berglund 2000).

The history of heathland can also be traced back to the Neolithic. For example, in Denmark and southern Sweden, pollen analysis of soils beneath Neolithic mounds has shown that heathland arose from woodland clearance on poor sandy soils. Heathland was widespread on relatively poor soils (often in areas characterised by granitic bedrock or areas outside the maximal ice extent of the Weichselian) in large parts of southern Scandinavia and neighbouring countries around the Baltic Sea (e.g. Greisman 2009; Olsson and Lemdahl 2009). The development of wooded or treeless pastures and hay meadows in upland and northern regions of the Baltic Sea region is closely linked to upland summer grazing and collection of fodder. There is still very little known about the history of summer grazing in the region, except in the province of Värmland in northern Sweden (e.g. Regnéll and Olsson 1998) where remains of summer farms were dated to mediaeval time. During the twentieth century, traditionally managed hay meadows, pastures, heathland and upland summer farming decreased dramatically with the introduction of chemical fertilisers and feed concentrates, reclamation and afforestation.

25.3.3 Land-Use and Anthropogenic Land-Cover Change Since the Neolithic (6000 BP)

The account presented here is based on ALCC model scenarios (Figs. 25.3 and 25.4), recent REVEALS and LRA reconstructions (Sugita et al. 2008; Gaillard et al. 2010; Nielsen and Odgaard 2010; Nielsen et al. 2012; Trondman et al. 2012) (Figs. 25.5, 25.6 and 25.7), earlier syntheses of palaeoecological proxy records, of which the most important are those by Berglund (1991), Berglund et al. (1996, 2002) and Ralska-Jasiewiczowa et al. (2004), and a large number of palaeoecological studies of which only a very small fraction is cited below. Examples are provided from the major environmental zones (according to Metzger et al. 2005) of the region, that is (i) the Nemoral, Atlantic North and Continental zones in the south, and (ii) the Boreal and Alpine North zones in the north, each representing about 50 % of the total land cover of the Baltic Sea basin.

25.3.3.1 Neolithic to Iron Age (ca. 6000–1000 BP)

According to archaeological and palaeoecological data, arable farming was introduced in the loess areas of central Germany with the Linear Pottery culture around 7700–7500 BP (Kalis et al. 2003), which is reflected in pollen diagrams from the area (e.g. Beug 1992; Voigt 2006). However, along the coasts of the Baltic Sea in northern Germany, Denmark, southern Sweden, and northern Poland (north of the Elbe river), the Mesolithic Ertebølle culture persisted for a long time, possibly because of the good fishing and hunting possibilities (e.g. Regnéll et al. 1995; Kalis et al. 2003; Richards et al. 2003). Larger scale crop cultivation and animal husbandry occurred first with the Neolithic Funnel Beaker culture from around 6100 BP in north-eastern Germany and northern Poland, 5900 BP in Denmark (Richards et al. 2003) and 5900 BP in southern Sweden (e.g. Berglund 1991). The earliest cultural impact on the landscape consisted mainly of a change in forest composition towards more early-successional species (birch, hazel), but from Late Neolithic, (ca. 4300–3800 BP in southern Scandinavia) anthropogenic grassland increased in size, while the areas with arable fields were still relatively small (e.g. Berglund et al. 2002; Odgaard and Nielsen 2009). In most of the southern environmental zones, a gradual differentiation of the landscape into three more or less distinct types occurred from the Late Neolithic onwards: (i) flat areas on clay-rich moraine soils developed the most intensive agricultural impact; (ii) hilly areas were more marginal in terms of agricultural activities and, therefore, remained rich in forest; and (iii) poor sandy soils gradually became dominated by heathland (e.g. in Denmark and southern Sweden, Berglund 1991; Odgaard and Rasmussen 2000; Berglund et al. 2002; Lagerås 2007; Greisman 2009; Nielsen and Odgaard 2010). By the Mid-/Late Bronze Age (ca. 3000–2500 BP), this division was well established and the overall pattern remained in place until around AD 1800, although the composition and distribution of the landscape types varied in time and space, and the total landscape openness increased through time. In some marginal areas of Denmark and southern Sweden, and along the Baltic Sea coast, in particular in the north-eastern part of Germany, northern Poland and the Baltic States, regional forest regeneration occurred during the Migration period (ca. 1600–1450 BP or AD 400–550; Andersen and Berglund 1994).

According to the ALCC scenarios of Kaplan et al. (2009), the earliest significant deforestation in the Baltic Sea basin occurred in the earliest Neolithic period on fertile soils, and by the Viking Age (ca. 1200–1000 BP), large areas of present-day Denmark, southern Sweden and Poland (i.e. the southern environmental zones) were ≥50 % under human use for crop and pasture land (Figs. 25.3 and 25.4). On the other hand, the scenarios by Pongratz et al. (2008) indicate that by 1200 BP, only about 3 % of the area potentially covered by vegetation on the globe was transformed to agricultural land, almost as much for cropland as for pastureland, none of the Baltic Sea basin (and almost none of Europe) was deforested by more than 50 %, and most of the region was deforested by 20 % or less except for Denmark, northern Germany, southernmost Sweden (Skåne Province) and Poland (Fig. 25.3). Thus, although both models identify the same regions as the most deforested, discrepancies are large between the estimates of the deforested land fraction for crop cultivation and pastures. Pollen-based REVEALS and LRA estimates of regional and local openness at 1000–1200 BP (≥50 % cover) agree with the Kaplan et al. (2009) scenarios for the regions of Skåne and Småland (southern Sweden) (Sugita et al. 2008; Gaillard et al. 2010; Fig. 25.5), most of Denmark (Nielsen and Odgaard 2010; Nielsen et al. 2012; Fig. 25.6), and north-western Europe in general (Trondman et al. 2011, 2012; Fig. 25.7), although between-site differences may be large at the local spatial scale (e.g. Denmark). The REVEALS-based reconstructions suggest that changes in human impact on vegetation/land cover over the past 6000 years were much more profound than suggested by earlier interpretations of pollen percentages; that is, the share of non-forested land through the Holocene is strongly underestimated by percentages of non-arboreal pollen (NAP, i.e. pollen from herbaceous plants) (Figs. 25.5 and 25.7). For instance, the REVEALS estimates of regional openness in the southernmost province of Sweden, Skåne, are 15–30 % for 6000–2750 BP and about 60 % for 2750–1000 BP, compared to 5–10 % and 30 % herb pollen, respectively, while in the province of Småland, north of Skåne, they are <10 % for 6000–4500 BP, 10–25 % for 4500–2000 BP and 25–30 % for 2000–1000 BP, compared to <2 %, <5 % and about 5 % herb pollen, respectively. These REVEALS reconstructions suggest that by the Late Bronze Age/beginning of the Iron Age, large areas of southern Sweden were under human use for crop cultivation and pastures. In Denmark, LRA estimates of local vegetation show that, on the richest soils, the openness often reached values over 80 % from ca. 3000 BP, while hilly areas were characterised by less openness (seldom over 40 %). On poor sandy soils, open heathland was dominant (70 % to over 80 %) (Nielsen and Odgaard 2010; Fig. 25.6). The maps of Fig. 25.7 clearly show the strong increase in size of the land surfaces covered by cultivated land (here exclusively cultivated with cereals) and grassland (here mainly grasses) between 6000 BP (Neolithic) and 3000 BP (Late Bronze Age), and between 3000 and 200 BP (AD 1750–1800). The pollen-based REVEALS and LRA reconstructions indicate that the ALCC scenarios by Kaplan et al. (2009) are reasonable, except perhaps the degree of deforestation in the Neolithic time (5500 BP), which in some areas are too high compared with REVEALS reconstructions from southern Sweden (Gaillard et al. 2010; Trondman et al. 2011, 2012; Figs. 25.5 and 25.7) and from Denmark and northern Germany (Nielsen et al. 2012).

25.3.3.2 Middle Ages (AD 1050–1500)

The ALCC scenarios of Kaplan et al. (2009) showed that deforestation intensified in Poland and the Baltic countries from the mediaeval period onwards. However, the major difference is seen between AD 700 and 900 (ca. 1300–1100 BP), in particular in the southern part of the Baltic Sea basin (Denmark, northern Germany, Poland), which increases up to 10–20 % deforestation (Fig. 25.4). This is in good agreement with the REVEALS and LRA-based reconstructions of regional and local vegetation that suggest increases of deforestation in early mediaeval time up to 10–15 % in southernmost Sweden, and up to about 20–40 % on rich soils and marginal areas of Denmark.

According to pollen and other palaeoecological studies, the land under agriculture expanded in area in the Middle Ages, resulting in a significant increase in landscape openness in the southern environmental zones of the Baltic Sea catchment. This was also a period with technological advances in agriculture (Porsmose 1999) and changes in crop composition (e.g. Behre 1992; Robinson et al. 2009). In Denmark for instance, open-land areas increased especially in the period AD 1200–1400, earliest in the core agricultural areas, and about 100 years later in the more forested areas (LOVE estimates, Fig. 25.6; Odgaard and Nielsen 2009). In the heathland regions in the west, the last forests disappeared (Odgaard and Nielsen 2009). The landscape also became more open in north-eastern Germany (Nielsen et al. 2012), and in southern Sweden, regional openness reached 80 % and 35 % in Skåne and Småland, respectively (REVEALS estimates, Fig. 25.5). Nevertheless, large parts of Småland were characterised by much larger openness in areas where grazed heathland expanded (Greisman and Gaillard 2009; Marlon et al. 2010; Cui et al. 2012).

25.3.3.3 Modern Time (AD 1500–2000)

The ALCC scenarios of Kaplan et al. (2009) indicate a progressive increase in deforestation of the region, in particular its southern part, reaching a peak around AD 1900. The twentieth century in the Baltic Sea basin is characterised by a period of land abandonment that is especially marked during the period 1980–2000, however, mainly confined to Denmark, northern Germany, the Baltic States and Poland (Fig. 25.4). At a global scale, the ALCC scenarios of Pongratz et al. (2008) indicate that, around AD 1700, the agricultural area had increased to about 9 % of the area potentially covered by vegetation on the globe (PGV), of which 3.5 % was cleared forest (85 % for cropland, 15 % for pasture) and 5.5 % was grassland and shrubland under human use (30 % for the cultivation of crops). Between AD 800 and 1700, the ALCC scenarios show that natural vegetation under agricultural use had increased by about 5 million km^2 (i.e. about 6 % of PGV). Within the next 300 years, the total agricultural area increased to about 50 % of PGV (mainly pastureland), that is roughly a 5.5 times larger area than at AD 1700. This reconstruction suggests that global ALCC was small between AD 800 and 1700 compared to the industrial time, but relatively large compared to previous millennia. During the preindustrial period of the twentieth century, the reconstruction shows clear between-region differences in histories of agriculture (Pongratz et al. 2008). However, regional reconstructions for the Baltic Sea region based on pollen records, other palaeoecological proxies, and archaeological/historical data differ significantly from the global picture proposed by Pongratz et al. (2008). According to the REVEALS and LRA model-based reconstructions, the deforested area did increase between AD 800 and 1700, however, by not more than about 50 % of the earlier deforestation. The increase in deforestation between about AD 1700 and 1850/1900 does not represent more than 50 % of the landscape openness at AD 1700, in many areas much less (10–20 %).

The pollen-based reconstructions indicate that the percentage cover of cereals was very high in the eastern parts of northern Germany and northern Poland from AD 1500 onwards, and lower in north-western Germany, Denmark and southern Sweden, where grazed grassland and heathland were the dominant human-induced vegetation types (Berglund et al. 2002; Berglund 2006; Nielsen et al. 2012). This agrees with the archaeological findings and historical sources indicating that cereals were imported to Denmark and Sweden from areas south of the Baltic Sea region (e.g. Robinson et al. 2009), while cattle were exported in large numbers from Denmark and Schleswig-Holstein to other parts of northern Germany and to the Netherlands from the fourteenth to mid-eighteenth century (Gijsbers and Koolmees 2001; Bruun and Fritzbøger 2002). Grazed heathlands had their maximum extent in the entire Baltic Sea region around AD 1500–1800. Thereafter, many heathland areas—as well as permanent grasslands and meadows—were converted into arable land or planted forests, especially with conifers (e.g. Eliasson 2002; Dahlström 2008; Frederiksen et al. 2009; Gaillard et al. 2009). Urban areas also expanded, especially after AD 1900 (e.g. Frederiksen et al. 2009; Münier 2009). In southern Sweden, southern Norway and the Baltic states, the landscape openness was at its maximum around AD 1850. Since then, urbanisation, abandonment of agrarian landscapes, land-use change and modern forestry have led to reforestation of large areas formerly used for agriculture (see also Chap. 21). This trend is not unique to the Baltic Sea region, but is also characteristic of many other regions of Europe for which the nineteenth century was the time of most intensive land use with maximum landscape openness, while the twentieth century was characterised by reforestation after abandonment and/or through plantation, as for example in southern Norway, northern Italy, central France, the Pyrenees, central Spain and Portugal (Gaillard et al. 2009; Krzywinski and O'Connell 2009).

25.4 Effects of Land-Cover Change on Past Climate: Model Experiments

This section reviews available studies on the effect of long-term CLCC and ALCC on past climate in the northern hemisphere and Europe.

25.4.1 AD 1850 to Modern Time

The LUCID project compares responses to historical ALCCs in various climate models in a series of studies (Pitman et al. 2009; de Noblet-Ducoudré et al. 2012). Pitman et al. (2009) concluded that there was general agreement on the significant effect of vegetation patterns and land-cover change on regional climate, while their role on global climate was still under debate. In particular, the effect of teleconnections related to land-cover change was considered questionable; that is, whether a change in the climate in a given region could be related to land-cover change in other regions. Some climate modelling studies suggest that such teleconnections exist (Henderson-Sellers et al. 1993; Zhang et al. 1996; Gedney and Valdes 2000; Werth and Avissar 2002, 2005), while others indicate they do not (Findell et al. 2007, 2009; Pitman et al. 2009). The following text reviews some of the earlier studies addressing these questions.

At the global scale, Sheng et al. (2010) identified hot spots of climate-induced change in surface energy fluxes during the period 1948–2000, although these were not related to land-cover change but rather to variability in atmospheric–surface interactions. The hot spots were primarily found in northern high-latitude areas. Based on observations, Teuling et al. (2010) investigated how differences in water and energy exchange due to the differences in land cover in the temperate forest zone affected the European heatwave in August 2003. They concluded that grassland and forest areas react differently to changes in soil water availability. As long as water availability is high, woodland exerts a warming effect compared to grassland due to higher Bowen ratios over woodland; that is, more of the available net radiation energy at the surface is used for vertical heat transfer than for evapotranspiration (see also Bonan 2008). However, woodland can also sustain its evapotranspiration rate when water availability is low, which leads to lower Bowen ratios than in grassland in dry conditions; that is, grassland becomes the source of excess heating instead of woodland.

There are still important problems in relation to how ALCCs are explored in numerical experiments using climate models. Pielke et al. (2011) concluded that most studies were based on only one or two models, which did not reflect the uncertainty between models in their responses to increased CO_2 levels or in the strength of their land–atmosphere interactions, an uncertainty that is evident in the LUCID multi-model study of Pitman et al. (2009). That study mainly focused on the Northern Hemisphere summer season, and the key result was a statistically significant impact of ALCC on the simulated latent heat flux and air temperature over the regions where anthropogenic land cover changed, but the direction of the change in summer temperature was inconsistent among the models. In terms of rainfall, four of the coupled atmosphere–land models used showed a significant impact on rainfall over regions with ALCC, while three models did not show impacts greater than the expected random variability of model outputs (Seneviratne et al. 2010). In their review, Pitman et al. (2009) did not find statistically significant impacts of ALCC on latent heat flux, temperature or rainfall remote from the actual ALCC, that is no teleconnections. The authors also suggested that robust conclusions on the effects of ALCCs on climate can only be drawn from multi-model experiments. Studies based on a single model only provide indications of possible feedback mechanisms and their implications.

The goal of the most recently published part of the LUCID project (de Noblet-Ducoudré et al. 2012) was to provide a detailed examination of why the LSMs diverge in their response to ALCC. For this purpose, the authors used seven atmosphere–land models with a common experimental design. For the vegetation distribution, each model used as a starting point the same distribution of crop and pasture, at a resolution of 0.5° × 0.5°, as extracted from Ramankutty and Foley (1999), combined with the pasture areas from Klein Goldewijk et al. (2011) (Fig. 25.8). However, as there are between-model differences in (i) the way land information was represented, (ii) sources of information to describe present-day and potential vegetation, and (iii) strategies to implement ALCC in the model, the resulting land-cover distribution (including natural vegetation) used in each model differed (de Noblet-Ducoudré et al. 2012; Fig. 25.8). Although the areas covered by crops increased from AD 1870 to AD 1992 in all land-cover data sets used, the increase varied; all LSMs describe temperate deforestation, but at varying degrees. Within the Eurasian region studied in LUCID, western Europe is characterised by reforestation (i.e. land abandonment and forest planting) rather than deforestation, and the Baltic Sea catchment area by mixed deforestation and reforestation, while the entire Eurasian region exhibits overall deforestation (Figs. 25.8 and 25.9). These differences in ALCC implementations between the LUCID model runs had an influence on how ALCC affected the near-surface climate in the models' results; that is, there is no consistency in how ALCC influenced the partitioning of available energy between latent and sensible heat fluxes at a specific time (Boisier et al. 2012; de Noblet-Ducoudré et al. 2012).

These results highlight the urgent need to evaluate LSMs more thoroughly. However, there are some robust common features shared by all models: the amount of available energy used for turbulent fluxes and the almost linear relationship between the climate response to ALCC and the amount of trees removed. All models simulated a systematic increase in surface albedo in all seasons. For most models, this increase (7 % for a full transition from forest to crop/grassland) was proportional to the amount of deforestation imposed on the individual models. Moreover, the larger surface albedo was shown to cause a decrease in QA (computed as the sum of absorbed solar energy and incident atmospheric infrared radiation); QA decreased in all seasons everywhere in the temperate regions and was also proportional to the amount of deforestation imposed on a given model. In most cases, crops and grasslands were less efficient than trees in transferring energy to the atmosphere in the form of turbulent fluxes due to a lower aerodynamic roughness length. All models that underwent a change in their forest fraction that was greater than 15 % simulated cooler ambient air temperature in all seasons. These common features and their dependence on the ALCC descriptions prescribed in each model suggest that, for a specified amount of deforestation occurring over specific periods, the dispersion among the models would be significantly smaller if the ALCC descriptions had been exactly the same in all models. LUCID also compared the biogeophysical impacts of ALCC with the impact of elevated greenhouse gas concentrations on sea surface temperatures and sea-ice extent. The results show that ALCC had an impact of similar magnitude—but of opposite sign—to increased greenhouse gases and warmer oceans. However, it should be stressed that although this result is valid for the entire Eurasian region, this is not necessarily the case for individual parts of that large region, such as the Baltic Sea catchment area. Moreover, in view of the dominant reforestation of the catchment's western part and deforestation of its eastern part, it is not possible to predict the net effect of ALCC at the scale of the entire catchment area without modelling the effects of deforestation and reforestation at the regional scale using regional climate models.

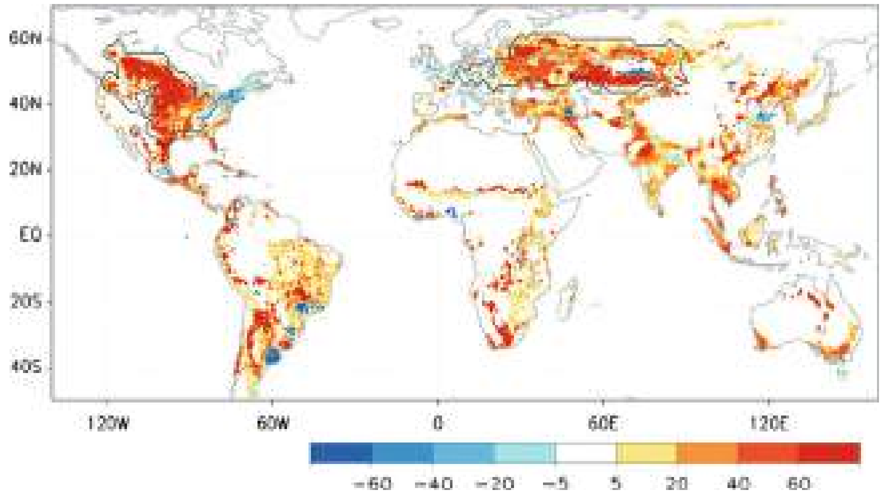

Fig. 25.8 Changes in the extent of agricultural land (crops and pastureland) between pre-industrial time (AD 1870) and present day (AD 1992). *Yellow* and *red* indicate an increase, and *blue* a decrease, in the extent of agricultural land since AD 1870. The two contours on the map indicate the regions used for specific analysis (North America and Eurasia) (de Noblet-Ducoudré et al. 2012)

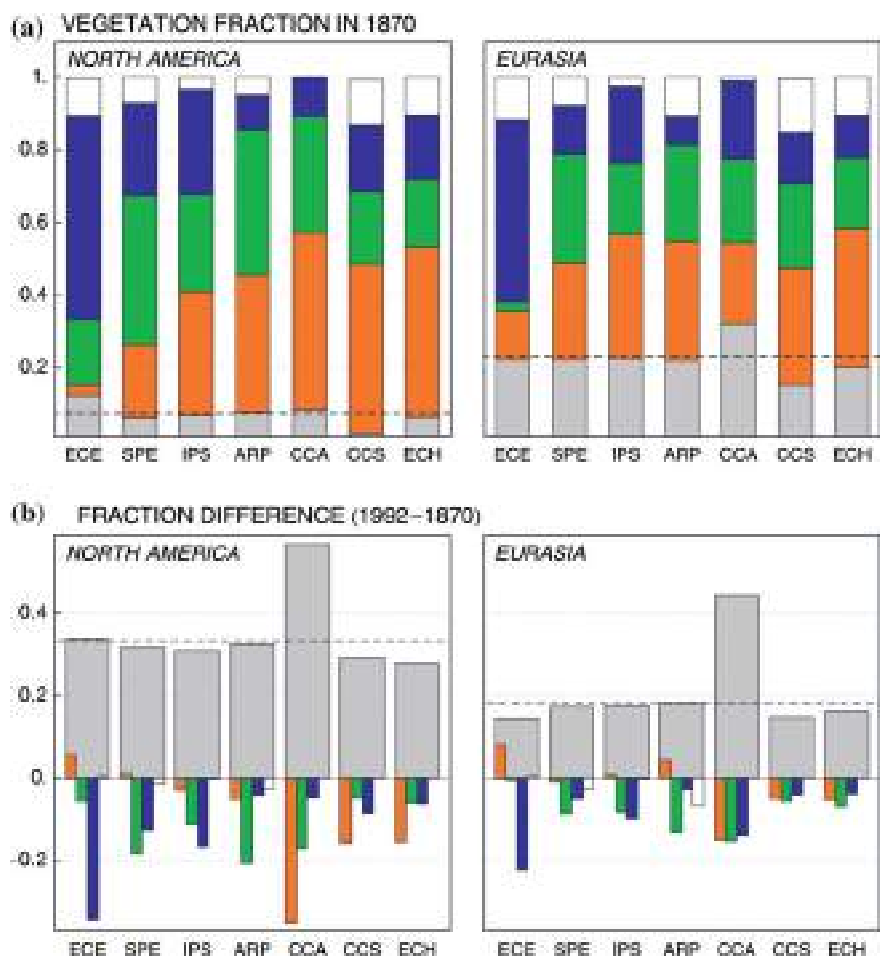

Fig. 25.9 Vegetation descriptions used in the LUCID project (de Noblet-Ducoudré et al. 2012). **a** Vegetation fraction in AD 1870 in the studied Eurasian region (shown in Fig. 25.8). Surface (in fraction of total area) covered by crops (*grey*), grassland types (*orange*), evergreen trees (*green*), deciduous trees (*blue*) and desert (*white*) for all seven models used in the LUCID project (for details on the seven models, see de Noblet-Ducoudré et al. 2012); **b** Fraction difference (AD 1992–AD 1870) (in fraction of total area) for each of the vegetation types shown in (**a**). The *dashed black line* in both graphs (**a**) and (**b**) shows the crop fraction that was finally implemented in all seven models (de Noblet-Ducoudré et al. 2012)

25.4.2 Before AD 1850

There are few studies of land cover–climate interactions before AD 1850, although the number has increased rapidly since 2009. To date, there are no published modelling studies on the possible feedbacks of CLCC on past natural climate warming such as the well-known Early Holocene increase in mean annual temperatures and on the effects of Late Holocene CLCC and/or ALCC on later climate changes such as the Medieval Climate Anomaly (warming) and the Little Ice Age (cooling, see also Chap. 3), except that of Pongratz et al. 2009b. However, simulations of the northwards expansion of trees due to a warmer climate showed that such climate-induced vegetation changes produced a positive feedback on climate (e.g. Cheddadi et al. 1997).

There is only one study of the effect of past ALCC on regional climate in Europe using a regional climate model, the LANDCLIM project (Gaillard et al. 2010; Strandberg et al. 2013). The aim of the study was to evaluate the direct effects of anthropogenic deforestation on simulated climate at two contrasting times of the Holocene ~6000 BP and ~200 BP in Europe applying RCA3, a regional climate model with 50 km spatial resolution (Samuelsson et al. 2011). Three alternative descriptions of the past vegetation were used: (i) potential natural vegetation (V) simulated by the dynamic vegetation model LPJ-GUESS (Smith et al. 2001), (ii) potential vegetation with anthropogenic land cover (deforestation) as simulated by the HYDE model (V + H), and (iii) potential vegetation with anthropogenic land cover as simulated by the KK model (V + K). The climate model results show that the simulated effects of deforestation depend on both local/regional climate and vegetation characteristics (Strandberg et al. 2013). At ~6000 BP, the extent of simulated deforestation in Europe is generally small, but there are areas where deforestation as simulated by Kaplan et al. (2009) (V + K) is large enough to produce significant differences in summer temperature of 0.5–1 °C, which is the case in southern Sweden and eastern Poland. However, the KK model overestimates deforestation in these areas compared to the pollen-based REVEALS reconstructions. At ~200 BP, simulated deforestation is much more extensive than previously assumed, in particular according to the pollen-based REVEALS reconstructions (see Sect. 25.3.3.3) and the KK model. This leads to significant temperature differences in large parts of Europe. In winter, deforestation leads to lower temperatures because of the differences in albedo between forested and unforested areas, particularly in the snow-covered regions. In summer, deforestation leads to higher temperatures in Central and eastern Europe since evapotranspiration from unforested areas is lower than from forests (*hydrological cycle effect*). Summer evaporation is already limited in the southernmost parts of Europe under potential vegetation conditions and, therefore, cannot become much lower. Accordingly, the *albedo effect* dominates also in summer, which implies that deforestation causes a decrease in temperature. Differences in summer temperature due to deforestation range from −1 °C in south-western Europe (cooling) to +1 °C in eastern Europe (warming). In the Baltic Sea area, the effects of deforestation at 200 BP are much weaker than in south-western and eastern Europe. The effect is strongest in southern Sweden where deforestation leads to lower winter temperatures by only 0.2–0.4 °C, but there is no effect in summer. The choice of anthropogenic land-cover estimate was shown to have a significant influence on the simulated climate. But the climate proxy data available for the two time windows are not precise enough to evaluate the results of the climate model runs in quantitative terms effectively.

Earlier studies also show that the *albedo effect* of historical deforestation was probably dominant among the effects of deforestation in the northern atmosphere. However, most studies suffer from the drawback that the grid resolution is rather coarse and far coarser than the scale necessary to capture local to regional processes (Hibbard et al. 2007). Brovkin et al. (2006) used the scenarios of past deforestation produced by Ramankutty and Foley (1999) for the period AD 1800 to present-day and the HYDE database to reconstruct the effects of ALCC on climate over the past 1000 years. The outputs from six different climate models showed a cooling of 0.1–0.4 °C over the Northern Hemisphere due to the biogeophysical effects (mainly increased albedo) of the estimated decrease in forest cover between AD 1000 and 2000. They also found a warming of similar magnitude due to the biogeochemical effects of ALCC, therefore a net effect close to zero. Pongratz et al. (2009b) investigated the influence of historical land-use changes on radiative forcing (RF). For all of Europe, except Scandinavia, a decrease of 0.3 W m^{-2} was found between AD 800 and 1700. At the global scale, the RF was small throughout the pre-industrial period (negative with a magnitude less than 0.05 W m^{-2}) and not strong enough to explain the cooling reconstructed from climate proxies between AD 1000 and 1900 (Little Ice Age).

To date, there are few estimates of CO_2 emissions due to historical ALCC at the sub-continental scale and none using regional climate models. In the context of this review, the most interesting study so far is that of Pongratz et al. (2010) that separated the relative strength of biogeochemical versus biogeophysical effects from ALCC during the past millennium using a coupled atmosphere–ocean general circulation model (AOGCM) and applying the reconstruction of historical ALCC of Pongratz et al. (2008). They found that biogeophysical effects had a slight cooling influence on global mean temperature (−0.03 °C in the twentieth century), while biogeochemical effects led to a strong warming (0.16–0.18 °C). During the industrial era, both effects caused significant changes in certain regions, but only a few regions experienced a biogeophysical cooling strong enough to dominate the overall temperature response. The authors concluded that the climate response to historical ALCC, both globally and in most regions, was dominated by the rise in CO_2 caused by ALCC emissions. However, the biogeophysical temperature response at the regional scale was greater than suggested by its global mean. For example, in Europe, the annual mean temperature decreased by 0.3–0.5 °C, and the cooling in northern high and mid-latitudes was found to be largely albedo-driven, leading to a winter cooling of up to 0.9 °C in north-eastern Europe, in general accordance with previous studies (e.g. Betts 2001). However, the albedo dominance over hydrological aspects in the Pongratz et al. (2010) study is only significant for the annual mean temperature, whereas transpiration effects are in some cases seasonally offsetting. The authors also concluded that strong local biogeophysical effects could substantially influence the spatial pattern of the net temperature response, as in eastern Europe for example. The global versus local effectiveness of biogeochemical versus biogeophysical effects was also demonstrated by the fact that, at the global scale, the entire land area was more strongly influenced by biogeochemical warming than the ocean, while biogeophysical cooling is particularly pronounced over agricultural areas. Pongratz et al. (2011) also quantified the contribution of local ALCC to historical global warming and showed the importance of past land-use decisions in influencing the mitigation potential of reforestation on these lands. In these simulations, they found that CO_2 warming dominated over albedo cooling at the global scale because past land-use decisions resulted in the use of the most productive land with larger carbon stocks and less snow than on average. Therefore, land-use decisions led to CO_2 warming in most agriculturally important regions of the world. This suggests that, in most places, reversion of past land-cover change may often be the most feasible step of implementing ALCC as a mitigation tool. However, because the amount of CO_2 emissions and the change in biophysical properties vary across regions and types of land-cover change, detailed analysis—that is, simulation of the regional climate response to local occurrence of ALCC—is needed for specific reforestation projects. The climate effect of past ALCC is likely to be a good indicator of the mitigation potential of reversing the area to its natural state.

The rest of this section summarises other studies of global-scale carbon emissions. Pongratz et al. (2009a) performed transient simulations over the entire last millennium with a GCM that couples the atmosphere, ocean and land surface with a closed carbon cycle. By applying the ALCC of Pongratz et al. (2008) as the only forcing to the climate system, they showed that the terrestrial biosphere experienced a net loss of 96 Gt C over the last millennium, leading to an increase in atmospheric CO_2 by 20 ppm. The biosphere–atmosphere coupling led therefore to a restoration of 37 and 48 % of the primary emissions over the industrial period (AD 1850–2000) and pre-industrial period (AD 800–1850), respectively. Atmospheric CO_2 rose above natural variability by late mediaeval times, but global mean temperatures were not significantly altered until strong population growth in the industrial period. Pongratz et al. (2009a) also found that only long-lasting epidemics or wars led to carbon sequestration because emissions from past ALCC compensate carbon uptake in 'regrowing' vegetation for several decades. Reick et al. (2010) derived the CO_2 emissions associated with ALCCs since AD 800 as reconstructed by Pongratz et al. (2008) and compared them with the pre-industrial development of atmospheric CO_2 known from Antarctic ice cores. They concluded that pre-industrial traces of CO_2 emissions from ALCC before AD 1750 was obscured by other processes of similar magnitude, while the steep increase in atmospheric CO_2 after AD 1750 and until AD 1850 (i.e. before the rise of fossil fuel emissions to significant values) was largely explained by rising emissions from ALCC. These results partly contrast with those of Kaplan et al. (2010) who found that by AD 1850, at the global scale, the cumulative CO_2 emissions due to deforestation since 6000 BC were 137–189 Pg C (using the HYDE scenarios of ALCC) and 325–357 Pg C (using the KK ALCC scenarios of Kaplan et al. (2009)). Kaplan et al. (2010) concluded that their results support the hypothesis that anthropogenic activities led to the stabilisation of atmospheric CO_2 concentrations at a level that made the world substantially warmer than it would otherwise have been. Similarly, Boyle et al. (2011) showed by using new model assumptions that the quantity of terrestrial carbon release due to early farming, even using the most conservative assumptions, greatly exceeds the net terrestrial carbon release estimated by inverse modelling of ice core data by Elsig et al. (2009). However, the conclusions of both Kaplan et al. (2010) and Boyle et al. (2011) remain an open question as the emission estimates are not compatible with current understanding of the global carbon cycle and records from ice cores.

25.5 Potential Future Trends in Land Cover and Associated Effects on Future Climate

25.5.1 Future Land-Cover Change Due to Anthropogenic Climate Change and Possible Feedbacks

Many of the existing scenarios of future trends in land-cover change are based on observations of vegetation change due to the recent climate warming, but also on socio-economic assumptions. Studies on future climate-induced (potential natural) vegetation change and related biogeophysical feedbacks to regional climate provide some indication of what to expect in the future, since the underlying mechanisms are likely to be similar for human-induced vegetation change. Such studies in Europe indicate a boreal treeline advance into the tundra regions of the northern latitudes of both the Barents Sea region (Göttel et al. 2008) and northernmost Europe (Wramneby et al. 2010; Smith et al. 2011). The most significant feedback associated with forest expansion at these latitudes is expected to be the albedo feedback (warming) that is likely to be strong enough to offset the climate gains from the increased carbon sequestration in these forests. Using the coupled regional climate–vegetation model RCA-GUESS (Smith et al. 2011) at the European scale, Wramneby et al. (2010) also showed that a future rise in the altitudinal limit of deciduous trees in the Scandinavian mountains due to increased temperature would lead to enhanced warming through the positive snow–vegetation–albedo feedback.

While the albedo feedback and its amplifying effect on climate warming is expected to be the most important biogeophysical feedback in boreal regions such as northern Europe (Strengers et al. 2010), an increase in forest cover also implies a contrasting biogeophysical feedback mechanism due to enhanced evapotranspiration. This feedback may, however, be of minor importance in boreal forests dominated by evergreen trees, since these forests have a comparatively low evapotranspiration rate (Bonan 2008). For the part of the Baltic Sea region characterised by a more temperate climate, the role of evapotranspiration might, however, be of greater importance due to the dominance of more strongly transpiring broadleaved deciduous forests, although authors disagree on the role of temperate forests in climate change (South et al. 2011). Significant feedbacks from such changes in the hydrological cycle were identified by, for example, Wramneby et al. (2010), but primarily in Central Europe. Meanwhile, there was no evidence that variations in cloudiness and precipitation over Europe could be attributed to vegetation dynamics. The lack of an established relationship between increased/reduced evapotranspiration, precipitation and cloud formation over Europe could be because these are strongly determined by the advection of moisture from the Atlantic. This is likely to overwhelm any feedback signal from vegetation-mediated changes in evapotranspiration. In addition, the ratio between sensible and latent heat exerts a strong local control on surface temperature, but effects on cloud formation and precipitation will take place at the site of condensation, further away and higher up in the atmosphere, diffusing the signal (Wramneby et al. 2010).

Climate–vegetation feedbacks not only influence the mean climate but can also affect climate variability. Seneviratne et al. (2006) performed a suite of climate model sensitivity simulations with and without soil moisture responses to infer the role of the land surface; a substantial fraction of the future temperature variability in Europe was attributed to land-surface processes mediated by soil–moisture feedbacks. In some respect, climate variability provides a better understanding of climate change, since its concrete consequences might be extreme climate events such as floods and droughts. For the European domain, and the Baltic Sea countries, such events already have severe consequences (Della-Marta et al. 2007).

25.5.2 Resource Management and Future Land-Cover Change Scenarios: Possible Effect on Future Climate

Today, the total land area of the Baltic Sea coastal countries (Russia excluded) is roughly 160 million ha, of which 79 million ha is forest and 46 million ha is agricultural land (FAO 2009a). Sweden and Finland are mainly covered by forest and constitute approximately 63 % of the total forest area in the Baltic Sea coastal countries, while Germany and Poland are dominated by agricultural land and represent about 71 % of the total agricultural land area (FAO 2009a, see also Chap. 21). In the past 20 years, forest areas have increased in western and eastern Europe (FAO 2009b). Production of food is extremely valuable for the European Union as well as forest raw material for pulp, paper and construction material. One of the most immediate challenges facing the forest and agricultural sector in EU countries is to meet the anticipated rise in demand for raw materials resulting from the promotion of renewable energy sources (e.g. EC 2009).

The forces driving future land use in Europe include agricultural policy and international markets. Other issues include global food security/scarcity; the possible development of new, alternative agricultural products; the preservation of agricultural and forest land; and urbanisation. Moreover, future climate change may also influence land use. Globally, a number of future land-use change scenarios have been explored, and over recent decades, regional scenarios have emerged for different parts of the world (Alcamo et al. 2008). Regional studies pinpointing future changes in the Baltic Sea region are very limited, but over the European domain, a growing number of future land-use scenarios are becoming available. The difficulty in moving focus from global to regional land-use scenarios lies in the variety of possible outcomes, since more details and locally specific questions need to be considered (Carter et al. 2007; Alcamo et al. 2008; Metzger et al. 2010). Scenarios on the future development of European land use generally rest on two assumptions (Alcamo et al. 2008): an increase in agricultural productivity and a decrease in European population. According to the United Nations projections, a population decline by 8 % is expected by 2030 (UN 2005), with a further decline likely for later years. At the same time, agricultural productivity is expected to increase by between 25 and 163 % depending on the technological developments assumed (Ewert et al. 2005). The net result of these trends is a decrease in the agricultural area required for food production. For Europe as a whole, the scenarios show a decline in cropland of 28–47 % by AD 2080 and a decline in grassland of 6–58 % (Rounsevell et al. 2006), the abandoned areas being reclaimed by either urban development or forestry, although some areas may be used to cultivate bioenergy crops. Bergh et al. (2010) argued that the expected climate conditions during the twenty-first century (according to IPCC SRES; Nakićenović and Swart 2000) are likely to imply improved growing conditions in boreal and cold temperate climates, but a decrease in production both in agriculture and forestry in Central and southern Europe (Lindner et al. 2010; Masters et al. 2010), which would result in increased pressure on both agriculture and forestry in northern Europe especially in the latter half of the century. The demand for crops and the importance of food security might imply that forests would need to be replaced by agricultural land. A few other studies have also indicated a sustained or even expanded agricultural fraction for some Baltic Sea countries (e.g. Denmark and Finland; by Audsley et al. 2006) during the twenty-first century.

The effect on climate of possible future regional land-use changes is to a large extent unexplored. Biogeochemical feedbacks from regional land-use changes have been discussed in the context of global climate change in some studies (Carter et al. 2007; Rounsevell and Reay 2009), but the direct biogeophysical feedbacks from expected land-use changes are yet to be addressed. Given that the majority of available future land-use scenarios at the European scale assume increasing fractions of forested areas in parallel with a reduction in agricultural land, it would imply a positive (warmer climate) albedo-mediated effect in winter when previously snow-covered agricultural land is replaced by snow-masking forested areas and, at least potentially, a negative (colder climate) effect from an enhanced hydrological cycle in summer due to higher LAI (Wramneby et al. 2010). It could also imply an increased CO_2 sink. Whether the net effect would be a warming or a cooling is not possible to assess without extensive modelling experiments. On the other hand, if future land-use change in the Baltic Sea region happens to be an increase in area of agricultural land as suggested by some studies, the different effects mentioned above would be of opposite sign, and so the net effect probably of different magnitude.

The International Union of Forest Research Organizations (IUFRO) has highlighted the immense potential for the forest sector to mitigate climate change at low cost, while agricultural production would have much smaller potential. Forestation/reforestation has been suggested as a tool to mitigate global warming because a growing forest takes up and stores carbon from the atmosphere (UNFCCC 2005). However, the scientific challenge is to understand how different land-use strategies can contribute to mitigation benefits (e.g. Canadell and Raupach 2008). Pongratz et al. (2011) have shown that reversion of past land-cover change may often be the most feasible step for implementing ALCC as mitigation tool, but that careful analysis of the possible effects of ALCC at the regional scale should be performed for each forestation/deforestation project (see Sect. 25.4.2). It is also debated whether carbon sequestration in forests is the most effective way of mitigating climate change and, therefore, whether forest management should be optimised to increase the carbon stock. Various forms of carbon-related land-use strategies, as well as carbon-accounting mechanisms, increasingly enter agricultural and forest politics and policy. Bioenergy, carbon sinks and raw wood products as substitutes for fossil-based materials (Sathre and O'Conner 2010) add to the traditional list of products and services that forests may provide. Ranges of energy-associated industry, in particular the bioenergy industry, together with other climate- and energy-related organisations, are now entering the forest sector with hopes of realising their strategies. However, it is important to note that many of the socio-economic factors and assumptions controlling future land-use policies take indirect biogeochemical processes into consideration while neglecting direct biogeophysical processes (Jackson et al. 2008), presumably because the number of studies on CO_2 is substantially larger. In other words, current understanding of land-cover changes and their biogeophysical feedbacks in regional climate change is limited in comparison with the large-scale carbon cycle feedbacks.

25.6 Conclusion

1. ALCC is one of the few climate forcings for which the net direction of the climate response (warming or cooling) over the last two centuries is still not known with

certainty. The major uncertainty is due to the often counteracting temperature responses to biogeochemical versus biogeophysical effects, but also to the difficulty of quantifying the counteracting effects of changes in albedo and hydrological cycle (both biogeophysical effects), as well as obtaining precise land-cover data for the past. However, it is recognised that it is important to attempt to quantify the contribution of local/regional ALCC to past regional and global climate change. This information is necessary (i) to identify ALCC as a forcing of past climate change (over centuries and millennia), (ii) assuming that agricultural expansion continues in many regions of the world, to understand whether it will lead to similar climatic consequences as in the past, and (iii) to understand how land-use strategies may mitigate future climate warming.

2. Deforestation or reforestation by humans may have a series of contrasting effects on regional climate, of which the net result in terms of temperature and precipitation is still not possible to establish from existing modelling studies. Although the biogeochemical effect of historical ALCC on climate is theoretically well understood, the magnitude of the increase/decrease in the forcing is not accurately quantified to date. In contrast, the influence of the biogeophysical effects of ALCC is still not fully understood because it is far more complex. While the land-surface modelling community generally agrees that ALCC may affect climate through physical effects—in particular by affecting the albedo and therefore the surface–energy balance—there is no consensus on what this proposition implies. Even though the estimated global-scale impact of historical ALCC on RF through land-surface albedo changes is small relative to the CO_2-related RF (Forster et al. 2007), it does not imply that ALCC has no significant impact on regional climate (e.g. Pielke et al. 2002; Davin et al. 2007). There is common agreement that, provided ALCC is spatially coherent at a sufficiently large scale, this would affect the regional-scale climate significantly. However, to date, three factors could not be quantified yet: (i) the scale of ALCC required to produce a significant effect on climate on global and regional spatial scales (when compared to other forcings of climate change), (ii) how large the resulting change in the regional climate can be expected to be, and (iii) how much the nature of the existing climate over a region might suppress or amplify the initial impacts of ALCC. Increased concentration of greenhouse gases in the atmosphere and the subsequent changes in sea-surface temperatures and sea-ice extent are considered today to be the main drivers of climate change also over land. However, such an assumption leads to erroneous conclusions regarding the land-surface impacts on climate change in regions where ALCCs have been significant. ALCCs have the potential to mask a regional warming signal, with the resulting risk that detection and attribution studies may miss a clear greenhouse signal or misattribute a greenhouse signal if the ALCCs are poorly accounted for in the model.
3. For most of Europe, and the southern zone of the Baltic Sea basin in particular, palaeoecological studies clearly show that the major transformation of the landscape by anthropogenic activities occurred between 6000 BP and 3000/2500 BP with major deforestation during the Late Bronze Age or Early Iron Age, depending on the area. The deforested area did increase between AD 800 and 1700, however, by not more than about 50 % of the earlier deforestation. Similarly, the increase in deforestation between about AD 1700 to AD 1850/1900 did not represent more than 50 % of the landscape openness at AD 1700, in many areas much less. The effect of this long-term deforestation on past climate, at global and regional scales, is not fully understood and still much debated.
4. There do not appear to be any model studies looking specifically at both the effects of biogeophysical and biogeochemical processes related to historical (ca. AD 1850—modern) ALCC on climate change at the scale of the Baltic Sea region. The only modelling study of the biogeophysical effects of past ALCC on regional climate in north-western Europe suggests that deforestation between 6000 and 200 BP may have produced significant changes in winter and summer surface temperatures (of the order of ±0.5–1 °C) through biogeophysical processes that vary in size and direction (decrease or increase) depending on the geographical area and season. In the Baltic Sea area, the major effect is seen as a slight cooling of 0.2–04 °C during winter in southern Sweden primarily due to the *albedo effect*. The net effect of anthropogenic deforestation and reforestation through carbon sources versus sinks (warming vs. cooling) and biogeophysical cooling versus warming is not yet quantified. Reforestation has been suggested as a tool to mitigate global warming. However, several model studies of large-scale reforestation/deforestation indicate that in boreal regions, the magnitude of the positive albedo forcing following reforestation could be larger than the magnitude of the negative forcing from CO_2 uptake, therefore leading to warming. The latter would imply that boreal reforestation might not be an effective climate warming mitigation tool as it could lead to increased warming through biogeophysical processes.
5. A conclusion from available modelling studies reviewed here is that there is no indication to date that deforestation in the Baltic Sea area since AD 1850 would have been a major cause of the recent climate warming in the region through a positive biogeochemical feedback (release of CO_2). Moreover, the southern part of the

Baltic Sea area was reforested after AD 1900, which should have resulted in a different net effect on regional climate than deforestation. The ALCC scenarios do suggest a significant increase in land use at this time, but this does not agree with palaeoecological reconstructions indicating that openness was at its maximum in many parts of Europe—and in the southern part of the Baltic Sea region—around AD 1850–1900. The question is whether the change in land cover over the transition to the industrial period was large enough to make a significant contribution to the climate warming. Theoretically, the tendency towards reforestation from the end of the nineteenth century and through the twentieth century could have implied a cooling if the biogeochemical effect was the largest of all effects or a warming if the albedo effect in winter was dominant over other biogeophysical effects and the biogeochemical cooling effect. However, the respective magnitude of each potential effect and their net result are still unknown for the Baltic Sea region. Therefore, there is still an urgent need to better understand the biogeophysical effects of reforestation in this region because of the idea still prevailing that planting trees will mitigate climate warming. Hibbard et al. (2010) recognised the importance of an accurate representation of land-use and land-cover change to understand and quantify the interactions and feedbacks between climate and socio-economic systems, respectively, and highlighted recent and innovative methods that integrate observations and modelling analyses of regional to global aspects of biogeophysical and biogeochemical interactions of land-cover change with the climate system. To conclude, a quote from de Noblet-Ducoudré et al. (2012): 'the appropriate question is not whether ALCC has a globally averaged significant impact, but is rather whether ALCC has an impact on regions that have undergone intensive ALCC (such as North America, Europe, India, China, Russia, Japan, and Indonesia) that is worth accounting for when exploring the impact of other human forcings on regional climate'.

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Annex 1: The Concept of Detection and Attribution

Armineh Barkhordarian

The main goal of 'detection and attribution' analysis in relation to climate change is to determine whether climate has changed as a direct result of human activity. The general philosophy of detection and attribution is first to assume that the variability in a dataset can be linearly decomposed into two components: variability that cannot be attributed to any known causes and variability that can in principle be attributed to externally forced climate change signals. When all physically plausible forcing mechanisms are considered to have been accounted for then the former will indicate the internal variability. This can be estimated either from control runs with global-scale climate models or from the observed record.

Detection and attribution is therefore a two-step process in which the first step—detection—establishes whether the observed change is significantly different from changes due to internal variability alone and the second step—attribution—identifies the mix of external forcing mechanisms that best explain the change detected. Detection can be phrased as a statistical test with the null hypothesis being that the observed change is due to internal variability alone. Attribution, in contrast, cannot be phrased as a statistical test but rather as a plausibility assessment.

Different statistical methods have been used in the detection and attribution context, such as correlation-based methods (Santer et al. 1993), Bayesian methods (Hasselmann 1998) and the optimal detection method (also known as 'optimal fingerprinting'), which can be considered as a generalised multivariate linear regression (Hasselmann 1997; Hegerl et al. 1997). A short description of the optimal detection method (optimal fingerprinting) is now given.

A1.1 Standard Approach

The standard detection model (e.g. Hasselmann 1997; Hegerl et al. 1997), assuming a noise-free model-simulated response pattern is a linear regression problem as follows:

$$y = \sum_{i=1}^{m} x_i a_i + u_0 = Xa + u_0 \qquad (1)$$

where $\boldsymbol{y}$ is the observed record, which is expressed as a linear sum of $\boldsymbol{m}$ model-simulated response patterns, $\boldsymbol{x_i}$, with scaling factor $\boldsymbol{a_i}$ plus internally generated variability $\boldsymbol{u_0}$. Matrix $\mathbf{X}$ contains the estimated response to the external forcings (signals) under investigation. Vector $\boldsymbol{a}$ is a vector of scaling factor that adjusts the amplitude of signal patterns by accounting for uncertainty in the amplitude of the external forcing and for the possibility that the amplitude of the climate model response to the forcing may not be correct.

Fitting the regression model (Eq. 1) requires an estimate of the climate's natural (internal) variability $\boldsymbol{u_0}$. The observed record is not long enough to give a reliable estimate of internal variability. Another difficulty is that the observed data may also contain the effect of external forcings. Thus, long control integrations of global-scale climate models, which are pre-industrial control experiments with all forcings held constant, are typically used for this purpose. Therefore, the distribution of the scaling factor $\boldsymbol{a_i}$ is assessed from fits of the regression model to non-overlapping control run segments derived from long control simulations. In this approach an ordinary least squares (OLS) method is used to estimate the scaling factors $\boldsymbol{a_i}$.

The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1

In the standard approach, detection of a climate change signal occurs if any of the scaling factors $\boldsymbol{a_i}$ are shown to be significantly different from zero. This is handled by testing the null hypothesis $\mathbf{H_{DE}}$: $\mathbf{a = 0}$ (where $\mathbf{0}$ is a vector of zeros). If the null hypothesis $\mathbf{H_{DE}}$ is rejected, it indicates that the observed representation of climate change deviates significantly from internal variability, that is, the observed change $\boldsymbol{y}$ cannot be explained by internal variability $\boldsymbol{u_0}$ alone.

Once detection has been established, attribution (consistency of observed changes with a combination of external forcing) is assessed by testing the null hypothesis $\mathbf{H_{AT}}$: $\mathbf{a = 1}$ (where 1 denotes a vector of unit). When there is insufficient evidence to reject $\mathbf{H_{AT}}$, the attribution of changes to the respective forcing is claimed.

A1.2 Accounting for Noise in Model-Simulated Response Patterns

The standard approach to optimal fingerprinting assumes that the model-simulated response pattern is known exactly, that is, it is not subject to sampling uncertainty (repeating the experiment would yield an identical pattern). Thus, the response patterns are specified by averaging the response to a particular forcing from an ensemble of runs of a climate model (Hegerl et al. 1997). The problem is that even with a few member ensembles, sampling uncertainty is still far from negligible, particularly in weak signal-to-noise situations such as the analysis of the response to solar forcing (Allen and Stott 2002). Therefore, the regression model (Eq. 1) is extended by an additional error term $\boldsymbol{u_i}$ that accounts for internal variability in the response patterns. Thus, the simple statistical model (Eq. 1) is replaced by:

$$y = \sum_{i=1}^{m} (x_i - u_i)a_i + u_0 \qquad (2)$$

where $\boldsymbol{x_i}$ is the ith model-simulated response pattern estimated from a finite ensemble and therefore contaminated with sampling noise $\boldsymbol{u_i}$ and $\boldsymbol{u_0}$ is the noise in the observation.

As described by Allen and Stott (2002), the presence of this internal variability (noise) in the signal patterns may bias OLS estimates of scaling factor ($\mathbf{a}$) towards zero slope, particularly if only a small ensemble is available to estimate signals. Using a total least squares (TLS) algorithm solves this problem. TLS minimises the perpendicular distance between the scatter points and the best-fit line, not the vertical distance minimised by OLS. Therefore, the bias of best estimate towards zero in OLS is removed by using TLS in optimal fingerprinting.

A.3 Signal-to-Noise Optimisation

As described by Hasselmann (1993), an optimal fingerprint method further requires a signal-to-noise optimisation. The scaling factors $\boldsymbol{a_i}$ are optimal estimators only if the internal variability is independent and identically distributed (iid). Therefore, the original observations and model data cannot be used in the regression model (Eq. 2) and a transformed version resulting in iid residuals is needed. A transformation is given by using a truncated set of the empirical orthogonal functions (EOFs) of the control simulations (weighted by the square root of their eigenvalues). Thus, the control simulations are used to present the original data in a dimension-reduced EOF space. A residual test proposed by Allen and Tett (1999) is usually performed to assess whether the control simulations adequately represent the variability of the data in the truncated space.

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Annex 2: Climate Science and Communication for the Baltic Sea Region

Dennis Bray and Grit Martinez

Self-criticism and reflection on the 'Assessment of Climate Change for the Baltic Sea Basin 2008' and the communication of the science to regional political decision makers

A2.1 Introduction

The first assessment of climate change in the Baltic Sea region was published in January 2008 (BACC Author Team 2008). Two years after its publication, a survey was undertaken in the form of a questionnaire (opened 5 February, closed 31 March, 2010) to assess the authors' and the BALTEX scientific membership's opinions concerning the report (Bray 2010). While the "Assessment of Climate Change for the Baltic Sea Basin" was intended for a scientific audience (it does not contain '*a summary for policy makers*'), a 2011 survey of regional decision makers on the German Baltic Sea (Bray and Martinez 2011) extended the inquiry to assess not only the perspective of the decision makers but also the awareness of the BACC report among the non-scientific decision makers. This annex presents findings from both surveys, thereby including both the production and consumption of scientific knowledge. In doing so it not only gives scope to the self-criticism of science (something beginning to be demanded by climate scientists themselves) but also provides a glimpse of the greater process of communicating science to the public (also something gaining significant attention).

A2.2 Baltic Sea Climate Scientists and the BACC Report

A2.2.1 Survey Sample

Contributors to the BACC report were all members of the BALTEX community, an open interdisciplinary research network around the Baltic Sea (The Baltic Sea Experiment, www.baltex-research.eu). The survey sample for the assessment of the BACC report consisted of the entire BALTEX mailing list, consisting of 700 scientists (of which 84 comprised the 'author team' of the BACC report and represented 13 Baltic Sea countries). From the 700 invitations, 134 scientists responded. This response rate, approximately 19 %, is similar to results of other online surveys. The respondents' participation in the BACC report is shown in Table A2.1. For the methodological background to the survey, see Bray (2010) and Bray and Martinez (2011).

Figure A2.1 depicts the national distributions of the respondents to the survey and the members of BALTEX who contributed to the BACC report. As this figure indicates, while Germany had the greatest proportion of

Table A2.1 Participation and awareness of the BACC report

Are you aware of the BACC report?		
	Number	Percentage
Yes	78	74.3
No	27	25.7
Total	105	100
Missing 32		
Following responses refer only to 78 who answered yes above		
Did you contribute as an author?		
	Number	Percentage
Yes	26	33.8
No	51	66.2
Total	77	100
Missing 60		
Were you consulted for input other than as an author?		
	Number	Percentage
Yes	21	26.9
No	57	73.1
Total	78	100
Missing 59		
Did you contribute as to what should and should not be included?		
	Number	Percentage
Yes	13	16.7
No	65	83.3
Total	78	100
Missing 59		

The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1

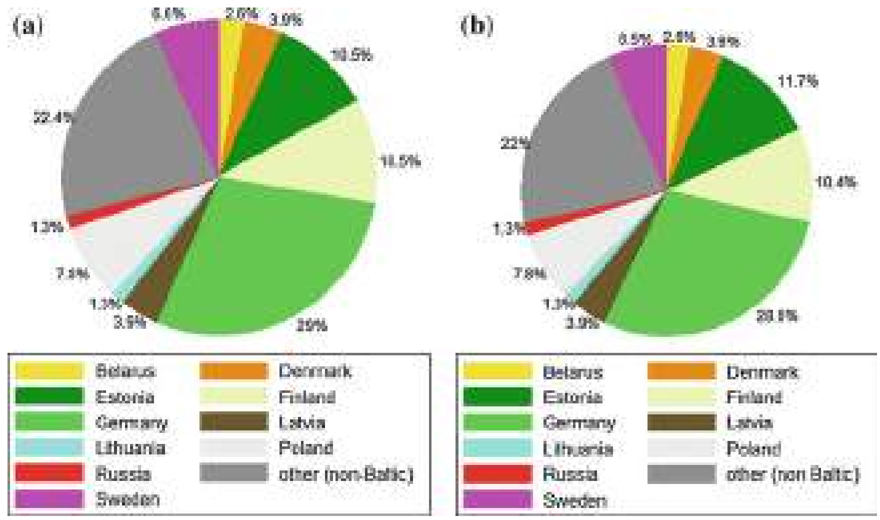


Fig. A2.1 National representation in the BACC survey. **a** Respondents aware of BACC Report by country; **b** respondents participating in production of BACC Report by country

scientists aware of the BACC report, the country with the most authors participating in the survey was Sweden. As 77 of the 105 scientists claiming to be aware of the BACC report were also authors, the subsequent comments on the BACC report will include all 105 scientists. The results are presented as box plots. Box plots illustrate the median, spread and data values, providing a visual assessment of the degree of consensus. Lowest and highest values are indicated by 'whiskers' extending from the boxes. The boxes contain the 50 % of total values falling between the 25th and 75th percentile, meaning that 50 % of the cases have values within the box, 25 % have values larger than the upper boundary and 25 % have values less than the lower boundary. The length of the box indicates how much spread there is in the data values within the middle 50th percentile. If, for example, one box is much longer than another then the data values in the longer box have more variability. The length of the box is considered to suggest consensus and the location of the box to represent assessment. The median is in the middle of the box only if the distribution is symmetric. If the median line is closer to the left of the box than to the right of the box the data are skewed in that direction, meaning that there are more cases towards that end of the distribution. If the median is closer to the right of the box then the tail of the distribution is towards those values. By focusing on the middle 50th percentile, extreme perceptions are separated from the more conservative perception represented in the shaded box.

A2.2.2 BALTEX Scientists' Assessment of the BACC Report

Scientists participating in the survey were asked to evaluate the BACC report in terms of making fair assessments of the presentation of future climatic projections for the Baltic Sea region. The results are presented in Fig. A2.2. All categories (precipitation, clouds, etc.) were taken directly from the table of contents in the BACC report so as to maintain consistency between the evaluation and the report itself. As Fig. A2.2 indicates, there is very little discrepancy among those aware of the BACC report as to the question of whether the report tended to underestimate or overestimate the possible magnitude of future climatic change. If anything, the scientists tended towards the claim that a number of climate change phenomena are slightly underestimated in the report. Only for surface air temperature and coastal erosion were there claims of slight overestimation.

Overall, it could be concluded that there is considerable agreement among the survey participants that the BACC report reflects the claims of the broader scientific community.

A2.2.3 BALTEX Scientists' Assessment of the State of the Science

For the remainder of this annex, the results refer to the entire sample of respondents, that is, all scientists who participated in the survey; both those that were aware of and those that were unaware of the BACC report. Figure A2.3 indicates the scientists' assessment of the abilities of the science to make projections of future climatic conditions.

As is evident in Fig. A2.3, BALTEX scientists recognise the limitations of global climate and ocean models, and reflect this uncertainty in their assessment of the limitations of regional models.

While there are no doubt numerous, and perhaps insurmountable, reasons for the shortcomings of both global and regional modelling, BALTEX scientists were asked their opinion concerning a few factors. Their opinions are presented in Fig. A2.4. According to the results presented, there

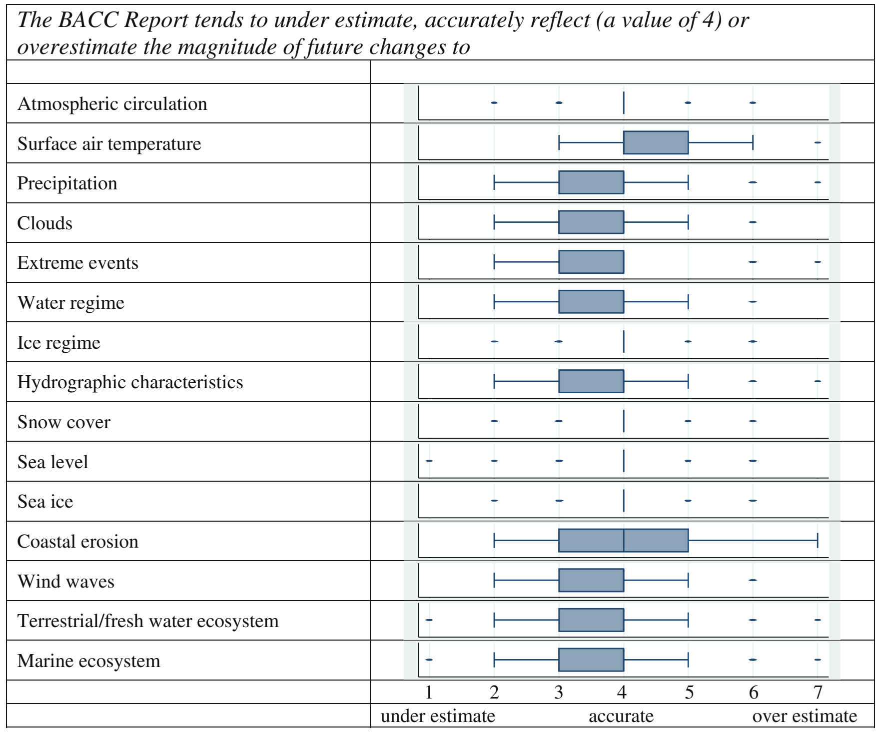


Fig. A2.2 Assessment of the scientific content of the BACC report

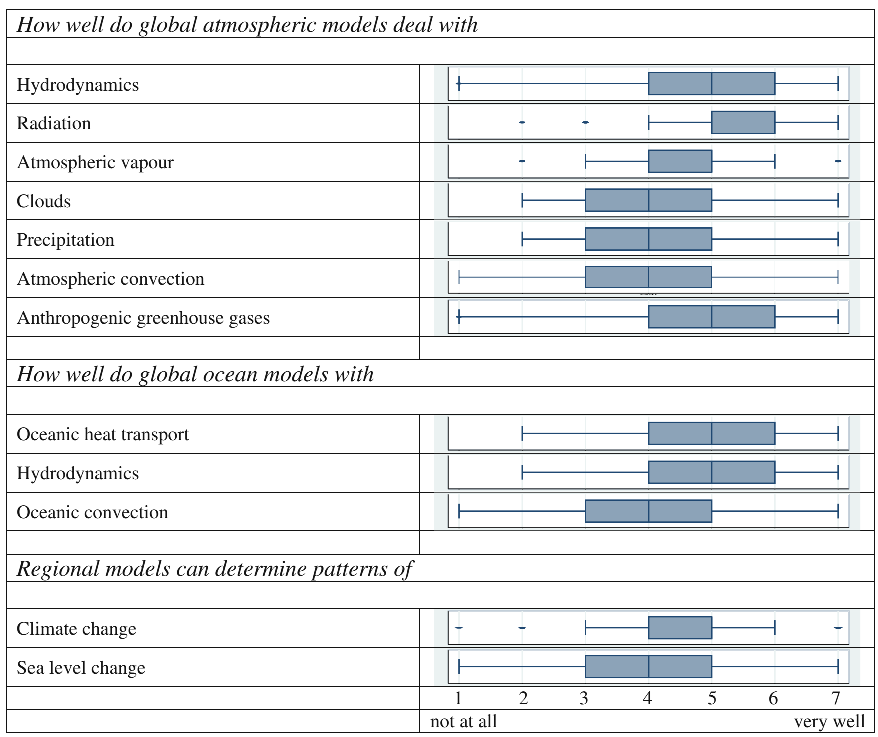


Fig. A2.3 BALTEX scientists' assessment of the ability of models (complete sample of scientists)

is room for considerably more research and the status of the science is perceived to be somewhat immature. However, the nature of the issue and the attention paid to climate change by the media means that the public is well aware of, although not necessarily well informed about, climate change.

A2.3 Message to the Public

Acknowledging a level of uncertainty, climate scientists were asked if they thought climate change and/or sea-level rise posed a threat to the Baltic Sea region. The results of the

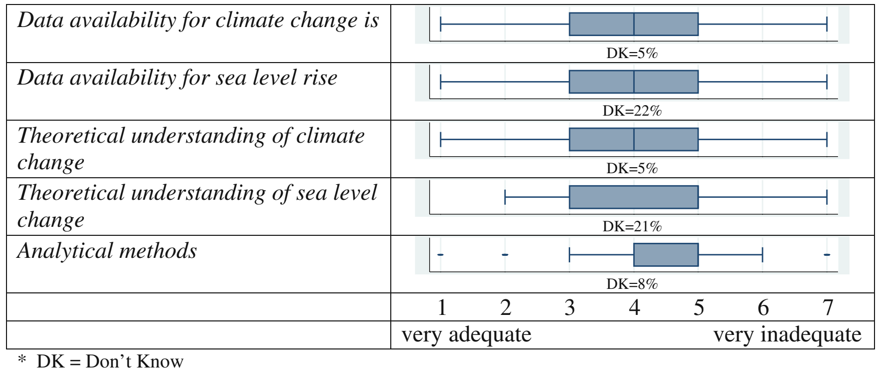


Fig. A2.4 Assessment of data, understanding and methodology for assessing climate change

survey are shown in Fig. A2.5. This figure also includes the perceptions of regional political decision makers for the German Baltic Sea region for the purposes of comparison.

Scientists expressed a wide range of perceptions concerning the level of threat from climate change and sea-level rise and significantly less than a majority is convinced that climate change or sea-level rise poses a very serious threat to the Baltic Sea region in which they live. This is even more pronounced when a time frame is introduced. Most scientists express limited concern about the next 10 years but are more concerned about the next 50 years, especially in relation to sea-level rise. When asked how the public should assess the issues of climate change and sea-level rise, scientists thought they should be concerned but not 'very worried'. As to the regional political decision makers, they thought that climate change and sea-level rise were both concerns over which the public should be worried, but again, not *very* worried. However, in terms of implementing adaptations strategies, this group expressed a sense of urgency that exceeded their declared level of worry, and far exceeded the conclusions of the scientists. This might be explained by the sources of information that regional administrations use in their decision making. Figure A2.6 shows the importance of various information sources.

After assessing the sources of the information, it is necessary to see what impact this information has on the decision makers' perceptions of future changes to the regional climate. Decision makers are most often concerned with impacts. While scientists express a considerable degree of uncertainty about the ability of regional climate models to project future conditions (see Fig. A2.3), decision makers show less uncertainty about the future climate (as shown in Fig. A2.7).

Figure A2.6 represents the lay interpretation of the scientific projections. Under conditions of climate change, warmer summer temperatures and stronger winds are perceived to be the strongest contenders for change. These are followed by warmer winter temperatures, more summer and winter rain, more snow, more storm floods, increased flooding due to precipitation and sea-level rise. It is interesting to note is that no decision makers responded 'not at all', and that a somewhat confused collective perception of the future is evident; namely that some decision makers expect summer temperatures to be warmer and some that summer temperatures will be cooler, some suggesting more rain, some suggesting less rain.

However, lay perspectives are not limited to non-scientists, as scientists themselves are often asked to comment on issues beyond their areas of expertise. While conceding that the ability to explicitly state climate change impacts on socio-economic systems is limited, climate scientists from the Baltic Sea region maintain that they should play a prominent role in the resolution of climate issues while only marginally labelling climate change as a *scientific* issue (Fig. A2.8).

There is a tendency for scientists to suggest that climate change (and sea-level rise) remains a scientific issue, suggesting the need for further investigation before placing any onus on politicians. There is also a slight tendency for scientists to favour enforced regulation and adaptation measures (as opposed to mitigation) as a means to meet the challenges of climate change and sea-level rise and to feel that that decisions regarding adaptation measures should be influenced principally by scientific expertise.

A2.4 Conclusion

Decision makers claim the need to take immediate action on meeting the challenges of climate change and sea-level rise, and claim they think the impacts of climate change and sea-level rise are something people should be worried about. Scientists, on the other hand, while agreeing that climate change and sea-level rise are potential threats, do not claim any impending catastrophe, even within the next 50 years. The direct voice of scientists saying there is no need for panic is going unheeded. In the truth-to-power model of science-policy interaction, scientists ideally produce impartial knowledge and this is given as truth to those in positions of political power, often by-passing the input of civil society (cf. Price 1965). The analysis reported here

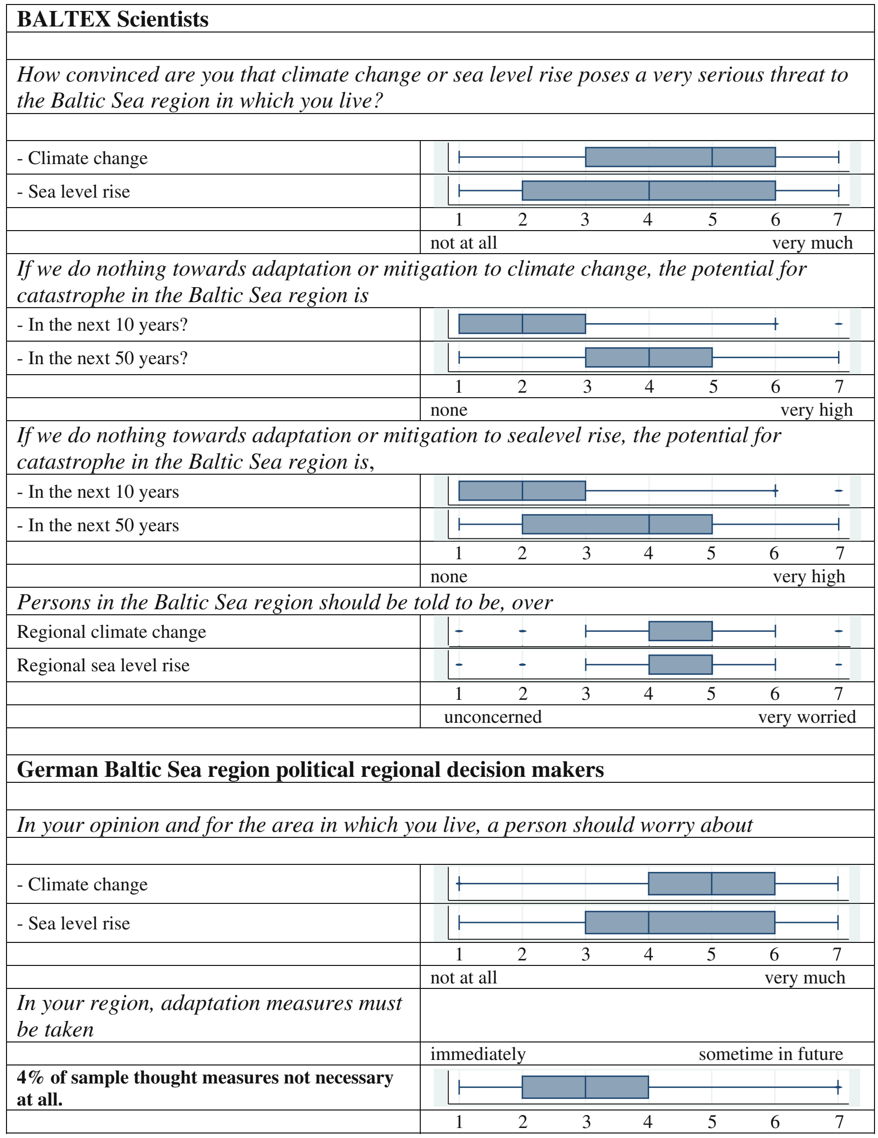


Fig. A2.5 Perceived impacts of climate change and/or sea-level rise in the Baltic Sea region

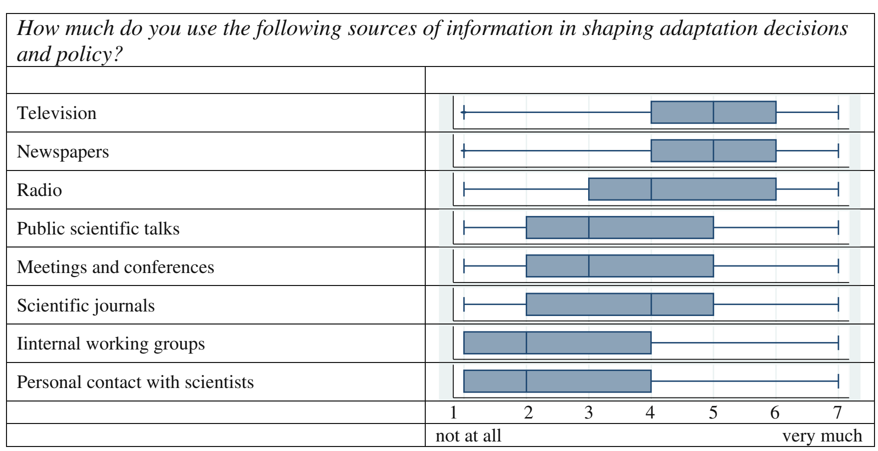


Fig. A2.6 Sources of information informing regional political decision makers for the German Baltic Sea region

Fig. A2.7 Decision makers' perceptions of future climate

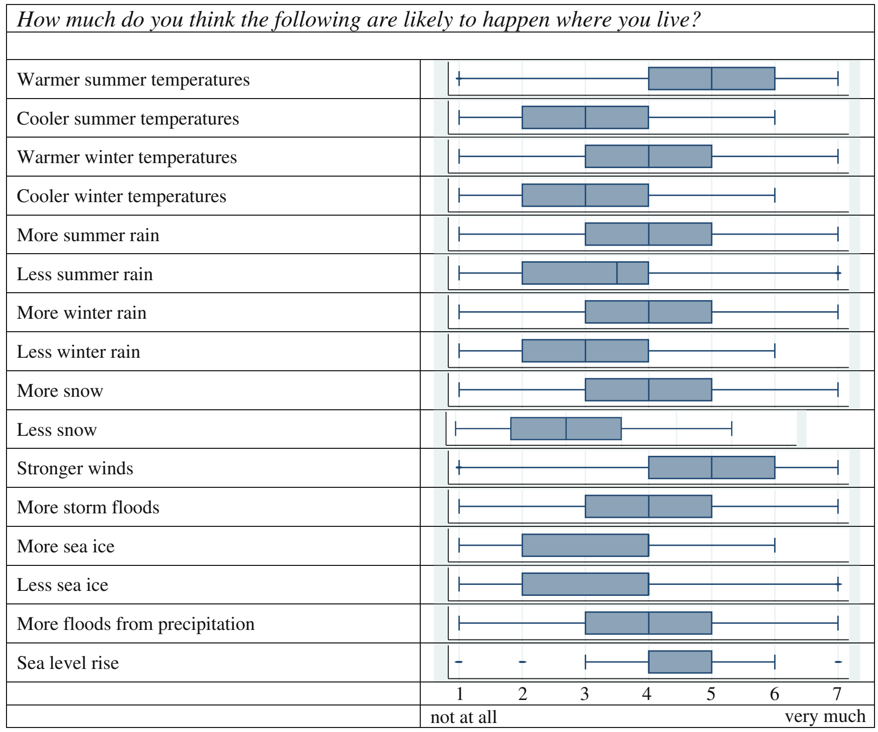
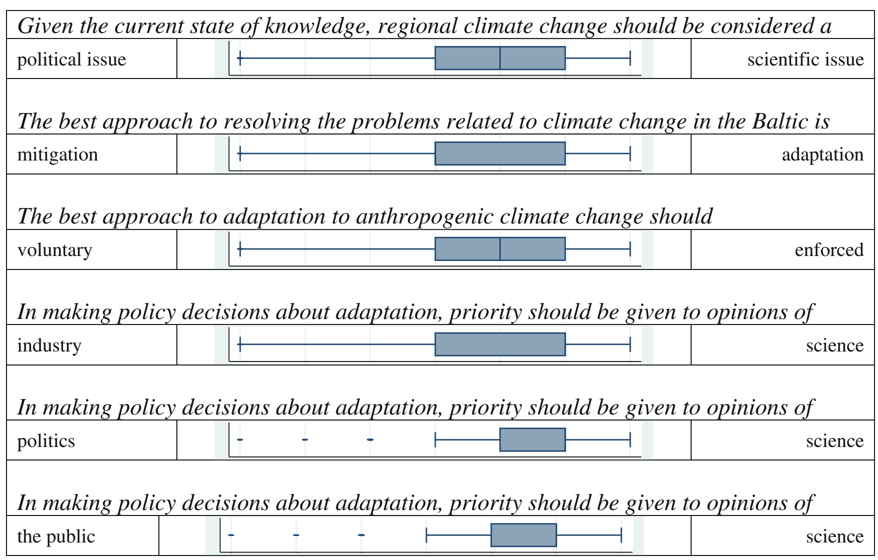


Fig. A2.8 Scientists and adaptation: normative perceptions



appears to indicate a somewhat modified version of this model, whereby truth, offered by science, states that the future will be different, but the difference as perceived by power is determined by sources other than science. In the present study, it appears to be the media and urban myths (street knowledge) that are shaping the understanding of risk and danger in the political imagination. The only conclusion that can be drawn from this is that communication between science and politics is in need of considerable improvement.

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Glossary

Acidification Decrease of pH in surface waters (soils, rivers, lakes, seas), caused by the uptake of carbon dioxide (CO_2) from the atmosphere

Active layer Top layer of soil that thaws during the summer and freezes again during the autumn

Added value Any kind of skill score measuring the relative improvement of a new forecast scheme relative to the existing one. In climate modelling it often refers to the relative improvement obtained from a downscaling procedure

Advection Transport mechanism of a substance or conserved property by a fluid due to the fluid's bulk motion

Aerosol Suspension of airborne solid or liquid particles, with a typical size between a few nanometres and 10 μm that reside in the atmosphere for at least several hours. Aerosols may be of natural or anthropogenic origin. Aerosols may influence climate directly through scattering and absorbing radiation and indirectly by acting as cloud condensation nuclei or ice nuclei, modifying the optical properties and lifetime of clouds

Albedo The fraction of solar radiation reflected by a surface or object, often expressed as a percentage. Snow-covered surfaces have a high albedo, the albedo of soils ranges from high to low and vegetation-covered surfaces and oceans have a low albedo. The Earth's planetary albedo varies mainly through varying cloudiness, snow, ice, leaf area and cover changes

Alkalinity Quantitative capacity of sea water to neutraliSe an acid

Anaerobic Oxygen-free metabolism

Anoxia Oxygen depletion, no free oxygen available for metabolic processes

Anthropogenic Resulting from human activity

Anticyclone Large-scale circulation of winds around a central region of high atmospheric pressure, clockwise in the northern hemisphere, anti-clockwise in the southern hemisphere

Arctic Oscillation (AO) Atmospheric sea-level pressure see-saw with higher surface pressure at mid-latitudes and lower pressure in polar regions, and vice versa

Autochthonous Species or property found at the place of its origin

Autotrophy Capability of organisms to produce biomass from inorganic substances using energy from sunlight or chemical reactions

Baroclinic A fluid or flow in which density depends on pressure and temperature. Isobaric surfaces cross isothermal surfaces

Barotropic A fluid or flow in which density depends only on pressure. Isobaric surfaces are also isothermal surfaces

Benthic Organisms or processes prevailing at the sea floor

Bias Difference between the simulated (modeled) value and the observed value

Bioaccumulation Accumulation of a substance in organisms though the surrounding medium or by food uptake

Black carbon Fine carbon particles (≤ 2.5 μm in aerodynamic diameter), formed through the incomplete combustion of fossil fuels, biofuel, and biomass, and emitted in both anthropogenic and naturally occurring soot

Boreal In the subarctic climate zone

Calibrated ^{14}C years (or calendar years) before present (BP) Dates derived from the radiocarbon (^{14}C) method and converted to calendar dates by comparing raw radiocarbon dates with dates derived from independent dating methods (such as tree growth-rings or sediment layers)

Catchment Drainage basin, watershed

Chemolithotrophic Deriving metabolic energy in microorganisms though inorganic chemical processes

The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1

Chironomids Taxonomic group of flies

Chlorophyll-a The molecule primarily responsible for the conversion of physical (photons) to chemical energy in photosynthetic organisms

Chlorophytes Taxonomic group of phytoplankton (green algae)

Chrysophytes Taxonomic group of phytoplankton (golden-brown algae)

Cladocerans Taxonomic group of small aquatic crustaceans (water flies), important in the Baltic Sea

Clastic Rocks or sediments composed of fragments or detrital material

Clupeids Taxonomic group of fish (e.g. herrings, sprat, sardines)

Cod reproductive volume Volume of sufficiently saline and oxygenated water so that buoyant cod eggs and larvae can survive

Coleopteran Taxonomic group of insects (beetles)

Convection Vertical movement of air or water masses driven by an unstable vertical density gradient (heavier masses over lighter masses)

Conveyor belt Characterisation of the global thermohaline circulation in the oceans

Copepods Taxonomic group of small crustaceans ubiquitous in all aquatic environments

Coriolis force Deflection of moving objects when viewed in a rotating reference frame, e.g. on the Earth´s surface

Cryosphere Portions of the Earth´s surface where water is in solid form (ice, snow cover, frozen ground)

Cyanobacteria Taxonomic group of bacteria important in the Baltic Sea and many other marine and freshwater environments(blue-green algae); many species are capable of fixing molecular nitrogen (N-autotrophs), thereby affecting the nitrogen cycle

Cyclogenesis Development or strengthening of cyclonic circulation in the atmosphere

Cyclolysis Termination or weakening of cyclonic circulation in the atmosphere

Cyclone Atmospheric low-pressure system

Dalton minimum Period of low solar activity from about 1790 to 1830

Data homogenisation Process of making data intercomparable (e.g. which were taken with different instruments)

Demersal Water column zone just above the sea bed

Dendroclimatology Methodology to determine past climates primarily from annual tree rings

Denitrification Biochemical respiratory process in microorganisms (bacteria) to convert nitrate to nitrite and ultimately molecular nitrogen at low oxygen concentrations

Diapycnal Perpendicular to a surface of constant density

Diatoms Taxonomic group of phytoplankton featuring a silica cell wall they are ubiquitous in all aquatic environments

Dicothermal Layer of cold and fresh water sandwiched by two layers of warmer wate

Dinoflagellates Taxonomic group of phytoplankton and protozoa, ubiquitous in all aquatic environments

Discharge Volume rate of water flow, usually given as cubic metres per second ($m^3\ s^{-1}$)

Dissimilation Metabolic pathways in organisms breaking down molecules into smaller units to release energy

Downscaling Methodology to increase the resolution of coarse global climate models to account for regional features such as topography and clouds

Drainage basin Watershed area

Dynamical downscaling Downscaling method where output from the GCM is used to drive a regional, numerical model in higher spatial resolution

e-folding time Time interval in which an exponentially growing quantity decreases by a factor of e (Eulerian number, $\sim$2.71828)

Eccentricity Amount by which the orbit around a body deviates from a perfect circle (e.g. the Earth's orbit of around the sun)

Eddies Parcels of swirling fluids (water or air)

Eigenvector In algebra, a vector whose direction remains unaltered by the action of a linear operator

Ekman drift or -transport The 90° net transport of the surface water layer (the layer affected by wind) by wind forcing relative to the wind direction

El Niño Warm phase of the El Niño Southern Oscillation (ENSO), associated with a band of warm ocean water that develops in the central and east-central equatorial Pacific

El Niño Southern Oscillation (ENSO) The cycle of warm and cold water temperatures of the tropical central and eastern Pacific Ocean

Ensemble Group of model runs with slightly different starting and lateral boundary conditions

Epilimnion Upper-most layer in a thermally stratified lake

Equilibrium Condition of a system in which all competing influences are balanced

Estuary Partly enclosed coastal body of brackish water with one or more rivers or streams flowing into it, and with a free connection to the open sea transition zone between freshwater (river) and maritime environments

Euryhaline Property of organisms which are able to adapt to a wide range of salinities

Eustatic sea level Global sea level defined by the volume of water in the world oceans or its basins

Eutrophication Natural or artificial addition of nutrients to bodies of water and the effects of the added nutrients

Evaporation Transfer of a liquid substance from its surface to the gaseous phase

Evapotranspiration Sum of evaporation and plant transpiration from the Earth's land and ocean surface to the atmosphere

Export production Fraction of primary production which leaves the ocean's surface layer (euphotic zone) by sedimentation

Exudation Release of cell constituents to the surrounding medium

Fast ice Stationary sea ice attached to the coastline or the sea floor

Flagellate Group of aquatic one-celled organisms able to move by action of a whip-like extrusion (flagellum)

Fluvioglacial Property created by glacial meltwater

Foehn Dry and warm down-slope wind that occurs in the lee (downwind side) of a mountain range

Fucoids Belonging to the taxonomic group of brown algae (*Fucus*)

Geodesy Scientific discipline that deals with the measurement and representation of the Earth

Geopotential Potential energy of an air mass

Geostrophic Condition for a theoretical wind or water current resulting from an exact balance between the Coriolis effect and the pressure gradient force (geostrophic balance)

Groyne Artificial hydraulic coastal structure to interrupt coastal currents and limit sediment movement (coastal protection)

Gyre Large rotating ocean current

Halocline Water layer where the salinity gradient is largest

Halophilous Referring to an aquatic organism which requires high salinity environments

Halosteric Sea water salinity contribution of sea level

Hemiboreal Climatic zone between the temperate and the boreal zone

Heterocysts Specialised cell type in nitrogen-autotroph cyanobacteria in which the nitrogen-fixing biogeochemical reactions take place

Heterogeneity Property of something distinctly non-uniform in composition or character

Heterotrophy Property of organism that cannot fix carbon (photo- or chemosynthesis) and needs organic carbon for growth

Hindcast modelling Test for a computer model by simulating past properties and comparing with the measured properties

Holocene Contemporary geological epoch, starting at about 12,000 cal yr BP and extending to the present

Homogeneity Property of something uniform in composition or character

Horohalinicum Salinity zone which corresponds to 5–8 and divides freshwater and marine faunas and flora (=critical salinity)

Hydrostatic flow A flow where the vertical acceleration is negligible or zero

Hygroscopicity Property of a substance to attract and hold water molecules from the surrounding environment

Hypoxia Low oxygen conditions in sea water (generally less than 2 ml^{-1})

Hypsographic Related to the distribution of elevations on the Earth's surface and the sea floor

Hysteresis Dependence of the output of a system not only on its current input, but also on its history of past inputs

Insolation Total amount of solar radiation energy received on a given surface area during a given time ($J\ m^{-2}$)

Interstadial Subdivision within a glacial stage marking a temporary retreat of the ice

Inundation Flooding

Irradiance Power of electromagnetic radiation per unit area (radiative flux) incident on a surface (W m^{-2})

Isopycnal Surface of constant potential density of water

Isostatic land uplift Post-glacial uplift rise of land masses that were depressed by the weight of ice sheets during the last glacial period

Isotope Variant of a chemical element; isotopes differ by the number of neutrons in the nucleus, the number of protons is not variable

Isotopic measurements Isotopes of specific elements (e.g. carbon nitrogen or oxygen) have specific properties (different atomic weight, different biogeochemical reactivity) which can be used to analyse specific environmental processes

Kelvin wave Wave in the ocean or atmosphere that balances the Earth´s Coriolis force against a topographic boundary such as a coastline or by the change in sign of the Coriolis force, like along the Equator

La Niña Cold phase of the El Niño Southern Oscillation (ENSO), associated with a band of cold ocean water that develops in the central and east-central equatorial Pacific

Lagrangian surface transport In oceanography, describing the paths that water parcels follow over time

Langmuir circulation In oceanography, a series of shallow, slow, counter-rotating vortices at the ocean´s surface, aligned with the wind

Late Maunder Minimum Minimum of observed sunspots in the period 1675-1715

Lateglacial Period in the beginning of the modern warm period when the northern hemisphere warmed substantially, causing a process of accelerated deglaciation following the Last Glacial Maximum (ca. 25,000 - 13,000 years ago)

Latent heat Amount of energy absorbed or released by a substance during a first order phase change that occurs without changing its temperature (e.g. by freezing, melting, evaporation, condensation)

Laurentide Ice Sheet Ice sheet covering most of North America several times times during the quaternary glacial periods (ca. 95,000-20,000 years ago)

Lithodynamics Sediment relocations in coastal regions (morphodynamics)

Littoral Part of a sea, lake or river that is close to the shore

Lycopod Taxonomic group of simple plants related to ferns (club moss)

Macroalgae Describing a group of macroscopic, multicellular, marine algae that live in coastal areas attached to the ground (sea weed)

Macrobenthic Describing organisms that live at the sea bottom (benthos) and are visible to the naked eye

Macrofossil Preserved remnants of extinct organisms visible to the naked eye

Macrozoobenthos Animals living on the sea bottom which are visible to the naked eye

Meridional Describing directions along the Earth's surface; meridional means 'along the meridian', i.e. the North-South direction

Mesolithic Archaeological time frame between the Palaeolithic and the Neolithic, in northern Europe dated to about between ca. 12,000 and 5000 4000 years ago

Mesoscale Intermediate scale in the study weather systems smaller than large (synoptic) scale systems but larger than micro- and storm-scale systems. Horizontal dimensions generally range from around 5 km to several hundred kilometers

Mesotrophic Intermediate level of trophic status, between oligotrophic (low in nutrients) and eutrophic (high in nutrients)

Micro- and macroepiphytes Microscopic or larger organisms that grow non-parasitically on a plant

Microzooplankton Zooplankton smaller than 200 µm in size

Mineralisation Process by which an organism produces inorganic substances

Minerogenic Formed by minerals

Momentum flux The vertical flux of horizontal momentum, equal to the force per unit area, or stress

Monitoring Regular systematic observation or survey of an environmental entity or process

Morphodynamics Sediment relocations in coastal regions

Mysids Taxonomic group of small aquatic crustaceans

Neolithic Period in the development of human technology, beginning about 10,200 BC and ending between 4500 and 2000 BC ('New Stone Age')

New production Aquatic primary production which is not fueled by nutrients regenerated within the system

Nitrification Biochemical oxidation of ammonia or ammonium to nitrite followed by the oxidation of the nitrite to nitrate. Nitrification is an important step in the nitrogen cycle

Nucleation First step in the formation of either a new thermodynamic phase or a new structure via self-assembly or self-organisation

Obliquity Tilt of the Earth; the angle between the Earth's rotational axis and its orbital axis, or the angle between its equatorial plane and orbital plane

Oligotrophic Low in nutrients

Omnivore Organism that can derive its energy and nutrients from a variety of food sources, e.g. plants and animals ('all-eater')

Orography Study of the topographic relief of mountains, and can more broadly include hills, and any part of a region′s elevated terrain

Ostracod Taxonomic group of small aquatic crustacean

Oxidant Oxidizing agent (oxidant, oxidizer or oxidiser), element or compound in an oxidation-reduction (redox) reaction that accepts an electron from another species

Palaeobotany Branch of palaeontology dealing with the recovery and identification of plant remains from geological contexts

Palaeoclimatology Study of climatic changes on the scale of the entire history of the Earth

Palaeoecology Study of fossils and subfossils to reconstruct the ecosystems of the past

Palaeoflora Plants of the geologic past

Pelagic In lakes and the oceans, the offshore deep water realm from the surface to the bottom

Pennate Subgroup of diatoms, elongated, non-centric in shape

Periglacial Geomorphic processes that result from seasonal thawing of snow in areas of permafrost, the runoff from which refreezes in ice wedges and other structures

Perihelion Point in the orbit of the Earth or another sun-orbiting body where it is nearest to the sun

Permafrost Permanently frozen soil (at least for two years)

pH Measure of the acidity or basicity of an aqueous solution

Photosynthates Products of photosynthesis (organic molecules)

Photosynthesis Process of collecting and converting photon energy to chemical energy in the chlorophyll-a molecule in plant cells to drive the production of organic matter from CO_2 and inorganic molecules, producing O_2

Phytoplankton One-celled autotrophic microscopic organisms of the plankton (i.e. free floating)

Piscivore Characterisation of fish-eating organisms

Planktivore Characterisation of plankton-eating organisms

Pleistocene Geological period which lasted from about 2588,000 to 11,700 years ago, spanning the world's recent period of repeated glaciations

Polar amplification Observation that any change in the net radiation balance tends to produce a larger change in temperature near the poles than the planetary average

Pollen Microscopic grains of seed plants, which produce the male gametes usually protected by a resistant hard coat which makes them usable in palaeoecological and biostratification studies

Primary production Process of producing organic biomass by CO_2 uptake in plants, driven by photosynthesis

Process-based model Theoretical concepts and computational methods that represent and simulate the behaviour of real-world systems derived from a set of functional components and their interactions with each other and the system environment, through physical and mechanistic processes occurring over time

Projection Potential future evolution of a quantity or set of quantities, often computed with the aid of a model. Unlike predictions, projections are conditional on assumptions concerning, for example, future socio- economic and technological developments that may or may not be realised

Protozoa One-celled heterotrophic (non-autotrophic) organisms

Proxy Record that is interpreted, using physical and biophysical principles, to represent some combination of climate-related variations back in time. Examples include pollen analysis, tree rings, data derived from marine sediments and ice cores can be calibrated to provide quantitative climate information

Pycnocline Water layer where the density gradient is largest

Radiative forcing Change in the net, downward minus upward, radiative flux (expressed in W m^{-2}) at the top of the atmosphere due to a change in an external driver such as the concentration of carbon dioxide or the output of the sun

Reanalysis Estimates of past atmospheric or oceanographic properties by processing observed meteorological or oceanographic data using fixed state-of-the-art weather forecasting or ocean circulation models with data assimilation techniques. Global reanalyses suffer from changing coverage and biases in the observing systems

Recruitment The number of new juvenile fish reaching a size/age where they represent a viable target for fishery for a given species

Redfield ratio Atomic ratio of carbon, nitrogen and phosphorus found in phytoplankton and throughout the deep oceans. This empirically developed stoichiometric ratio is found to be C:N:P = 106:16:1. It is of great importance for biogeochemical and ecological questions

Redox (Bio-) Chemical reaction by which one partner acts as electron donor (oxidation) and another as electron acceptor (reduction)

Redoxcline Water layer where the highest activity of biogeochemical redox reactions is observed

Remote sensing Environmental surveying by satellite sensors from the Earth's orbit

Resilience Capacity of a system to respond to a perturbation or disturbance by resisting damage and recovering quickly

Resolution In climate models, the physical distance (metres or degrees) between each point on the grid used to compute the equations. Temporal resolution refers to the time step or time elapsed between each computation of the equations

Return value The highest (or alternatively lowest) value of a given variable, on average occurring once in a given period of time (e.g. in 10 years return period)

Rhizodeposition Release of plant material or exudates by the root to the root-surrounding soil (rhizosphere) acts as a food source for a ecological community of microorganisms

Rossby waves Subset of inertial waves in the atmosphere and oceans that largely owe their properties to Earth´s rotation

Rotifers Taxonomic group of microscopic zooplankton (wheel animals)

Runoff Part of precipitation that does not evaporate and is not transpired but flows through the ground or over the ground surface and returns to bodies of water. Describes general, long-term and/or regional processes and is typically given as litres per second per square kilometre ($l\ s^{-1}\ km^{-2}$), allowing comparisons between rivers of different sizes, or millimetres per year ($mm\ yr^{-1}$), allowing comparisons with precipitation and evaporation

Satellite altimetry Sea surface height measured by satellite sensors

Scenario A plausible description of how the future may develop based on a coherent and internally consistent set of assumptions about key driving forces and relationships. Scenarios are neither predictions nor forecasts, but are useful to provide a view of the implications of developments and actions

Seiche A standing wave in an enclosed or partially enclosed body of water, e.g. the Baltic Sea

Semi-empirical model Model in which calculations are based on a combination of observed associations between variables and theoretical considerations relating variables through fundamental principles (e.g. conservation of energy). For example, in sea level studies, semi-empirical models refer specifically to transfer functions formulated to project future global mean sea level change, or contributions to it, from future global mean surface temperature change or radiative forcing

Sensible heat Turbulent or conductive flux of heat from the Earth´s surface to the atmosphere that is not associated with phase changes of water

Sequestration Carbon storage in terrestrial or marine reservoirs

Silviculture Cultivation, development and care of forests

Skilful scale Spatial scale resolution at which the climate models are able to capture climate features (approx. 8 grid point distances or about 1000-2500 km)

Skill Statistical evaluation of the accuracy of forecasts or the effectiveness of detection techniques

Solar constant A measure of flux density, the amount of solar electromagnetic radiation (the solar irradiance) per unit area that would be incident on a plane perpendicular to the rays, at a distance of one astronomical unit (AU) from the sun. It is measured by satellite to be roughly 1.361 $kW\ m^{-2}$ at solar minimum and approximately 0.1 % greater at solar maximum

Spectral nudging Technique for driving regional climate models (RCMs) on selected spatial scales corresponding to those produced by the driving general circulation model (GCM). It prevents large and unrealistic departures between the GCM driving fields and the RCM fields at the GCM spatial scales

Speleothem Secondary mineral deposit formed in a cave

Stationarity Stochastic process whose joint probability distribution does not change when shifted in time

Statistical downscaling Downscaling method where a statistical relationship is established from observations between large scale variables (e.g. atmospheric surface pressure) and a local variable (e.g. wind speed at a particular site)

Steric effect Change in sea level caused by changes in water density due to temperature or salinity changes

Stochastic Non-deterministic (i.e. 'random') condition so that the subsequent state of the system is determined probabilistically

Stokes drift The transfer of a specific fluid parcel as it travels with the fluid flow

Stomata In plant leaves, pores that control the leaf´s gas exchange. Important factor in evapotranspiration

Stratification Division of air or water into different layers

Stratigraphy Description of stratification

Streamflow Flow of water in streams, rivers and other channels

Surge Temporary increase, at a particular locality, in the height of the sea or streams due to extreme meteorological conditions (strong winds, heavy precipitation)

Synoptic scale In meteorology, the horizontal length scale of the order of 1000 kilometres or more

Teleconnection Climatic anomalies related to each other at large distances (typically thousands of kilometers)

Thermocline Water layer where the vertical temperature gradient is largest

Thermohaline circulation Large-scale circulation in the ocean that transforms low-density upper ocean waters to higher-density intermediate and deep waters and returns those waters back to the upper ocean. The circulation is asymmetric, with conversion to dense waters in restricted regions at high latitudes and the return to the surface involving slow upwelling and diffusive processes over much larger geographic regions. It is largely driven by high densities at or near the surface, caused by cold temperatures and/or high salinities, but also by wind and tides

Thermophilous Condition related to warmth-loving species

Thermosteric Sea water temperature contribution of sea level change

Tide gauge Device for measuring the change in sea level relative to a reference point

Transgression Retreat of a coastline due to tectonic sinking of the crust or eustatic sea-level rise

Trend Change, generally monotonic in time, in the value of a variable

Trophic level Referring to the position an organism occupies in a food chain

Tundra Environment or region where tree growth is hindered by low temperatures and short growing seasons

Undersampling Sampling at too low a frequency

Upwelling Oceanographic phenomenon that involves wind-driven motion of dense, cooler, and usually nutrient-rich water towards the ocean surface, replacing the warmer, usually nutrient-depleted surface water

Validation Part of the testing of the usability of a climate model. Refers to determining whether the model accurately represents reality. To do this, model results have to be compared with observations obtained in the same conditions

Varve Annual layer of sediment or sedimentary rock

Weathering Dissolution of silicate and carbonate rocks by action of weather. May involve physical processes (mechanical weathering) or chemical activity (chemical weathering)

Younger Dryas Period between 12,850 and 11,650 thousand years ago during the deglaciation, characterised by a temporary return to colder conditions in many locations, especially around the North Atlantic

Zoobenthos Animals living on or in the sea floor

Zooplankton Small aquatic animals usually drifting with the currents (restricted mobility)

^{14}C Unstable carbon isotope with an atomic weight of approximately 14 and a half-life of about 5700 years. It is often used for dating purposes going back some 40,000 years

δ13C Isotope measurement of the ratio between the stable carbon isotopes ^{12}C and ^{13}C in a sample. The ratio can give clues regarding past CO_2 concentrations in the atmosphere

δ15N Isotope measurement of the ratio between the stable nitrogen isotopes ^{14}N and ^{15}N in a sample. The ratio can give clues regarding the nitrogen cycle

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The BACC II Author Team, *Second Assessment of Climate Change for the Baltic Sea Basin*, Regional Climate Studies,
DOI 10.1007/978-3-319-16006-1