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Geophysics of sea ice in the Baltic Sea: A review

Timo Vihma [a,*], Jari Haapala [b,1]

[a] Finnish Meteorological Institute, POB 503, 00101 Helsinki, Finland
[b] Finnish Institute of Marine Research, POB 2, 00561 Helsinki, Finland

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ABSTRACT

With improved observation methods, increased winter navigation, and increased awareness of the climate and environmental changes, research on the Baltic Sea ice conditions has become increasingly active. Sea ice has been recognized as a sensitive indicator for changes in climate. Although the interannual variability in the ice conditions is large, a change towards milder ice winters has been detected from the time series of the maximum annual extent of sea ice and the length of the ice season. On the basis of the ice extent, the shift towards a warmer climate took place in the latter half of the 19th century. On the other hand, data on the ice thickness, which are mostly limited to the land-fast ice zone, basically do not show clear trends during the 20th century, except that during the last 20 years the thickness of land-fast ice has decreased. Due to difficulties in measuring the pack-ice thickness, the total mass of sea ice in the Baltic Sea is, however, still poorly known. The ice extent and length of the ice season depend on the indices of the Arctic Oscillation and North Atlantic Oscillation. Sea ice dynamics, thermodynamics, structure, and properties strongly interact with each other, as well as with the atmosphere and the sea. The surface conditions over the ice-covered Baltic Sea show high spatial variability, which cannot be described by two surface types (such as ice and open water) only. The variability is strongly reflected to the radiative and turbulent surface fluxes. The Baltic Sea has served as a testbed for several developments in the theory of sea ice dynamics. Experiences with advanced models have increased our understanding on sea ice dynamics, which depends on the ice thickness distribution, and in turn redistributes the ice thickness. During the latest decade, advance has been made in studies on sea ice structure, surface albedo, penetration of solar radiation, sub-surface melting, and formation of superimposed ice and snow ice. A high vertical resolution has been found as a prerequisite to successfully model thermodynamic processes during the spring melt period. A few observations have demonstrated how the river discharge and ice melt affect the stratification of the oceanic boundary layer below the ice and the oceanic heat flux to the ice bottom. In general, process studies on ice–ocean interaction have been rare. In the future, increasingly multidisciplinary studies are needed with close links between sea ice physics, geochemistry and biology.

1. Introduction

The primary reasons to study sea ice geophysics in the Baltic Sea have been related to two aspects: winter navigation and climate. The extent and thickness of the ice cover and the duration of the ice season are important for navigation and travel on the ice, and already for centuries there has been an interest to monitor these variables (e.g., Sass, 1866). Today the winter navigation in the Baltic Sea is very active. For example, in Finland more than 80% of the international trade is transported via the sea, and ships need ice-breaker assistance for 3–6 months each winter, although in the mildest winters this need is restricted to the Gulf of Finland, Gulf of Bothnia, and the Gulf of Riga. The winter navigation is strongly increasing: in the Gulf of Finland the traffic is estimated to twofold by 2010–2015 with more need for operational services. These services are based on monitoring of the actual ice conditions and forecasting of the conditions in the time scale of a few days. The latter is based on models, in which at least the ice dynamics and often also the thermodynamics are taken into account. Simultaneously with the increasing navigation, the interest in the climate and environmental changes is strongly increasing, and sea-ice research is becoming more multidisciplinary.

Through various mechanisms sea ice is an important factor in the climate system of the Baltic Sea region. First, sea ice has a high albedo. Sea ice is often covered by snow, and the albedo of a dry, new snow cover can be up to 0.9, while the albedo of melting bare ice is only of the order of 0.4 (or less, if the ice has become very

* Corresponding author. Tel.: +358 9 1929 4173; fax: +358 1929 4103.
E-mail addresses: timo.vihma@fmi.fi (T. Vihma), jari.haapala@fmi.fi (J. Haapala).
[1] Present address: Finnish Meteorological Institute, POB 503, 00101 Helsinki, Finland.

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thin). Even the latter is much higher than the albedo of the open sea, which is typically below 0.1. Changes on the snow/ice surface are accordingly associated with a strong feedback mechanism: a small reduction in the albedo can cause a large increase in the net solar radiation. Second, sea ice and its snow cover act as good insulators between the ocean and the atmosphere reducing, or even preventing, the air–sea exchange of heat, water vapour, CO_2, and other gases. In winter, the air–water temperature difference through the ice and snow can be up to 30 K. Sea-ice cover is, however, seldom uniform, but broken by cracks, leads, and polynyas, which act as pathways for heat and moisture from the ocean to the atmosphere. In winter conditions, the sum of the turbulent fluxes of sensible and latent heat over the areas of open water can reach several hundreds of Watts per square metre. This has a large effect on the thermodynamics of the atmosphere, sea ice, and the ocean. Third, the ice cover acts as a mechanical barrier between the atmosphere and ocean. Below land-fast ice this prevents the air–sea momentum flux, and below drift ice the momentum flux may be either increased or decreased, depending on the ice motion and the roughness of the ice surface and bottom. Fourth, the ice pack stores and advects fresh water, heat, atmospheric settling, and sediments, and may release them far away from their original source. The extent and thickness of sea ice are thus highly sensitive indicators of climate variability and change.

In recent decades, we have witnessed several research activities addressing the sea ice geophysics in the Baltic Sea. This review was initiated by the Baltic Sea Experiment (BALTEX), a continental scale experiment under the Global Energy and Water Cycle Experiment of the World Climate Research Programme, where sea ice has been one of the central research topics (Omstedt et al., 2004a; Vihma and Haapala, 2005). In the European project 'Baltic Air–Sea–Ice Study' (BALTEX–BASIS, Launiainen and Vihma, 2001) from 1997 to 2000, the main emphasis was on the interaction between the atmosphere, sea ice, and the ocean, as well as in sea ice thermodynamics. Sea ice was also addressed in the European projects 'Baltic Sea System Study' (BASYS; Krauss, 2000) and 'Ice State' (Riska and Tuhkuri, 1999). The focus of the sea-ice research in BASYS was to model the seasonal and long-term evolution of the Baltic Sea ice pack, while the objective of the 'Ice State' was a formulation of interconnection between the local and geophysical scales describing ice cover deformation. Sea-ice modelling issues have also been parts of the German and Swedish national research projects DEKLIM and SWECLIM (Rummukainen et al., 2004). Important field observations have been carried out in the Gulf of Finland as a part of a Finnish-Japanese co-operative program (Kawamura et al., 2001; Granskog et al., 2004). In the field of Baltic Sea ice climatology, a regular series of workshops has been arranged every 3 years since 1993 (Leppäranta and Haapala, 1993; Järvet, 1999; Sztobryn, 1999; Omstedt and Axell, 2002). A European project 'Arctic Ice Cover Simulation Experiment' (AICSEX) paid attention also to the relation between the Arctic and Baltic Sea ice conditions. From the point of view of winter navigation, Baltic Sea ice was also studied in the European project 'Ice Ridging Information System for Decision Making in Shipping Operations' (IRIS). For multidisciplinary collaboration in sea-ice research in the Baltic Sea and the Arctic, the Nordic Network for Sea-Ice Research (NetICE) was founded in 2007.

Here, we review the geophysical sea-ice research in the Baltic Sea. Sea ice is also studied in the fields of geochemistry, biology, ship engineering, and remote sensing, but we will not review these fields. A review on sea ice in the Baltic Sea, written from the point of view of geochemistry and biology, was presented in Granskog et al. (2006a), while the remote sensing issues are reviewed in Barale and Gade (2008). Our review on the Baltic Sea ice climate was included in Heino et al. (2008), and it is improved and extended here in Section 2. In Section 3, we address the observations, process studies, and modelling of sea ice structure, properties, and thermodynamics, while sea ice dynamics is addressed in Section 4. The structure, properties, thermodynamics, and dynamics of sea ice are all interrelated. Many basic studies on ice dynamics have, however, not addressed the three other topics. We therefore organize this review with separate Sections 3 and 4. Many observational studies on thermodynamics have also included process-oriented modelling or development of model parameterizations, and therefore we will include such model-related studies in Section 3. Operational and climatological modelling is addressed in Sections 5 and 6, respectively. A summary is given and perspectives for the future are discussed in Section 7.

2. Baltic Sea ice climate

The extent and thickness of the ice cover as well as the duration of the ice season are the most commonly used variables to describe the ice climate. Observed climatological trends in these variables

Table 1
Trends in the annual maximum ice extent (MIB), annual maximum ice thickness (MIT), and duration of the ice season (DIS).

	Region	Period	Trend	Significance	Reference
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	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 +	See Fig. 3	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), No (other)	Haapala and Leppäranta (1997), Seinä (1993), and Launiainen et al. (2002)
MIT	Gulf of Bothnia	1980–2000	− other	NR but evident	Seinä (1993) and Launiainen et al. (2002)
MIT	Northern Gulf of Finland	Early 1900–1990s		NR	Alenius et al. (2003)

− Denotes decreasing and + increasing trend. NR indicates that the statistical significance of the trend has not been reported.

are summarized in Table 1 and discussed below. Unfortunately, in many papers, the statistical significance of the trends observed has not been reported in detail.

2.1. Ice extent and thickness

Monitoring of the ice extent has been based on different methods. Regular ice observations in the coastal regions of the Baltic Sea started in late 1800s. The Finnish operational ice service was established in 1915, but better estimates of the ice extent became possible only with reconnaissance flights, which in Finland started in 1934. The latest milestone in the accuracy of data on sea ice extent was the start of satellite observations in 1967.

The oldest information available on sea ice extent in the Baltic Sea originated from various sources including observations at lighthouses, records on travel on the ice, old newspapers, and scientific articles (Speerschneider, 1915, 1927). Jurva (1952) collected such information from winters 1720–1940, also utilizing the correlation between the ice extent and air temperature in Stockholm and Helsinki. Due to uncertainties in the data from the 1700s and early 1800s, Jurva himself never published the whole time series. In his last paper, Jurva (1952) showed the estimated ice extent only from 1830 onwards. The whole time series from 1720 onwards was published as figures in Palosuo (1953), from where the ice extent has been later digitized by various authors (Lamb, 1977; Alenius and Makkonen, 1981; Leppäranta and Seinä, 1985). In the original figures, the ice extent is illustrated with bar diagrams, and an estimate of the uncertainty of the ice extent is denoted by dashed lines, which we reproduce in Fig. 1. The absolute uncertainty is largest for severe ice winters, such as 1739/1740, when the estimates range from 350,000 to 4,201,000 km^2, but a relative uncertainty of 15–25% is also present in mild ice winters in 1760s. Only the maximum estimates are given in many time series published. Seinä (1994) and Seinä and Palosuo (1996) have summarized the maximum annual ice extent (MIB) in the Baltic Sea utilizing the material of the Finnish operational ice service from winters 1941–1995 and the information collected by Jurva (1952) from winters 1720–1940.

The inter-annual variability in sea ice extent is large (Fig. 1). In the widely used classification of ice winters by Seinä and Palosuo (1996), mild, average, and severe winters contain the same percentage (∼33%) of the winters in the period of 1720–1995. Mild and severe winters are further classified in extremely mild, mild, severe, and extremely severe ones. The extreme categories both contain ∼10% of the winters. The sea areas covered by ice in extremely mild, mild, average, severe, and extremely severe winters are shown in Fig. 2. According to this classification, during the last 22 years all ice winters have been average, mild, or extremely mild, which may give reason to develop a new classification based on recent ice climate. The latest extremely severe ice winter occurred in 1986–1987, and the latest winters with the Baltic Sea totally frozen have been in 1941–1942 (certainly) and 1946–1947 (most probably; Simojoki, 1952). According to Haapala and Leppäranta (1997), the maximum annual ice extent in the Baltic Sea did not show clear trends during 1900s (the decreasing trend found had a statistical significance less than 90%). In the time scale from 1720 to 2008 we see, however, a decreasing trend (Fig. 1 and Table 1), and it is further discussed in Section 4.

Accurate data on the ice thickness is almost entirely restricted to the zone of land-fast ice. In the Bothnian Bay, the annual maximum level ice thickness is typically 0.65–0.80 m (Seinä and Peltola, 1991), and it reaches 0.3–0.5 m even in mild winters. In the Skagerrak and the coastal areas of Germany and Poland the annual maximum ice thickness seldom exceeds 0.5 m (BSH, 1994).

In their analysis of 37 time series from coastal stations around the Baltic Sea, Jevrejeva et al. (2004) did not find any consistent change in the annual maximum ice thickness. According to Haapala and Leppäranta (1997), the level-ice thickness in the Baltic Sea

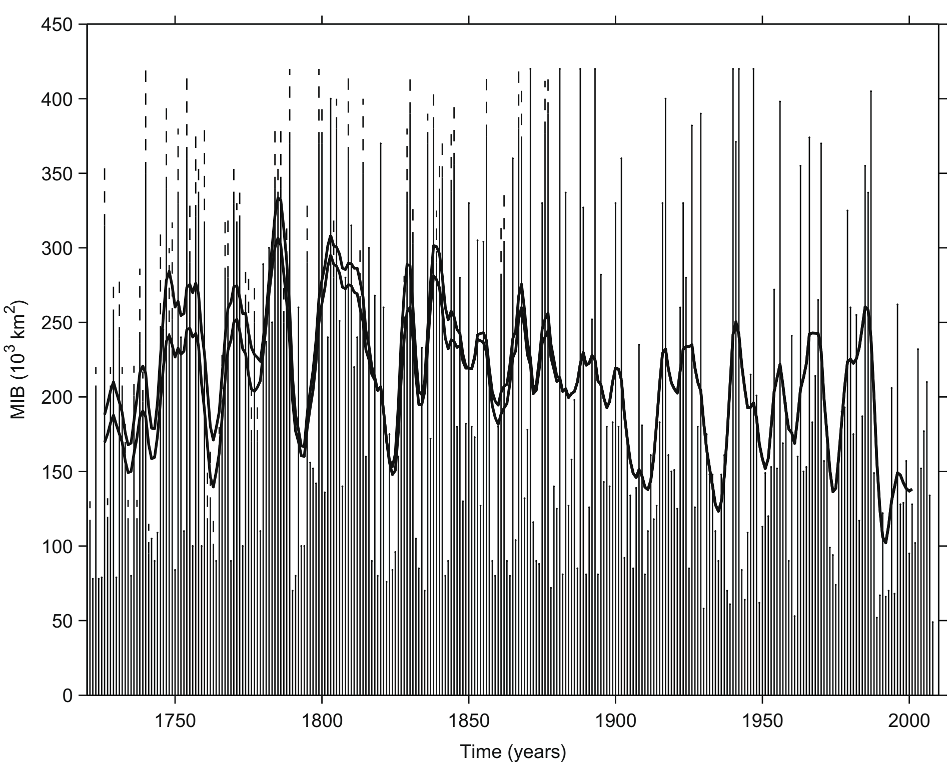

Fig. 1. Annual maximum ice extent in the Baltic Sea 1720–2008. The uncertainty of observations during the early part of the time series is indicated as dashed bars (Palosuo, 1953). The curves show the 15-year Gaussian-filtered time series for the high and low estimates of the maximum ice extent.

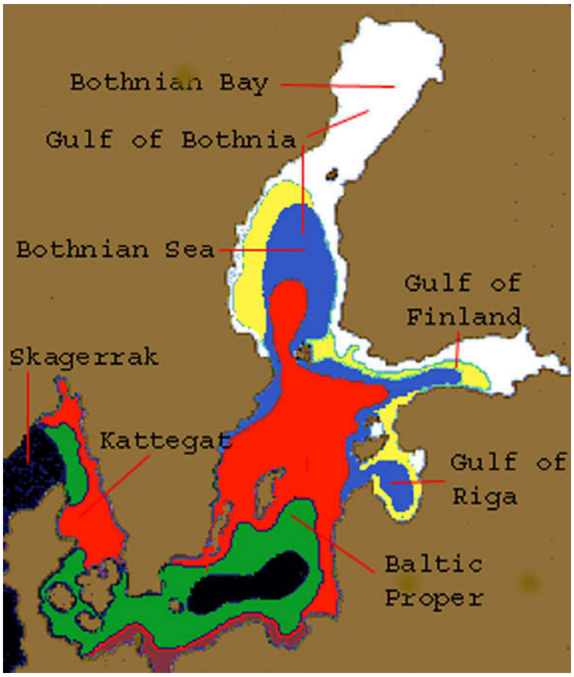

Fig. 2. Annual maximum ice extent in the Baltic Sea in winters classified as extremely mild (white and at least part of the yellow area), mild (also at least part of the blue area), average (also at least part of the red area), severe (also at least part of the green area), and extremely severe (also at least part of the black area). Redrawn from Seinä and Palosuo (1996).

did not show clear trends during the 20th century. Seinä (1993) and Launiainen et al. (2002) reported an increasing trend in the maximum annual ice thickness off Kemi (northernmost Gulf of Bothnia) during the 20th century until 1980s; in more southerly locations in the Gulf of Bothnia no clear trends were observed for the same period (Table 1). At all stations in the Gulf of Bothnia, decreasing trends have been observed since 1980s. In the Gulf of Finland, the maximum annual ice thickness has had a decreasing trend off Helsinki and Loviisa (Alenius et al., 2003). With focus on winter navigation, Launiainen et al. (2002) calculated the annual maximum distance from the harbour of Hamina (eastern Gulf of Finland) to a zone of sea ice less than 0.10 m thick. The results for the period from 1951 to 2000 strongly depended on the air temperature with short distances in 1990s.

In the drift ice regions, where the most of the sea-ice mass locates, we do not have accurate data on the ice thickness. During the last 10 years, many field studies have concentrated on the mapping of the ice thickness, but no systematic long-term measurements of the drift ice thickness have been carried out. Sea-ice thickness distribution can be observed with a fixed upward-looking sonar (which would allow long-term monitoring), from submarines, and by the airborne electromagnetic method (EM). Multala et al. (1996) showed that a fixed-wing aircraft EM-method is applicable to the Baltic Sea, but these measurements have not been continued. Haas (2004a,b) used helicopter-borne EM-instrument in the Baltic Sea during the winters of 2003 and 2004, and measured ice thickness characteristics along the Finnish coast. The results showed that, even in rather large regions, the mean deformed ice mass is often much larger than the undeformed ice mass. The observed mean sea-ice thickness averaged over an approximately 100-km-long flight track varied from 0.3 to 1.8 m, with several sections where the mean ice thickness was more than 1.5 m. The ice thickness information presented in Finnish and Swedish operational ice charts refers to level-ice thickness, and the results of Haas (2004a,b) clearly demonstrated that it should not be used as an estimate for the deformed ice thickness.

As far as we are aware, climate-scale changes in the occurrence of various ice types have not been studied, although daily ice observations were classified according to the ice type already more than 60 years ago: nine different ice types were identified in the instructions for German ice observers (Deutschen Seewarte, 1945).

2.2. Length of the ice season

The first sea ice in the Baltic Sea typically forms in November (at earliest in the beginning of October) in the shallow coastal areas in the northernmost Bothnian Bay. The maximum ice coverage is usually reached in February or March, but sometimes already in January, and sea ice remains in the Bothnian Bay typically until mid-May. Remnants of individual ice ridges have been observed as late as early July (Palmen, 1928). In the Skagerrak and the coastal areas of Germany and Poland, the probability of the sea ice occurrence is 25–75% (BSH, 1994). There the ice season is accordingly very variable from year to year: in German coastal waters, in some winters the ice cover forms as late as early March, while in some winters the last ice has disappeared already by the end of December (Schmeltzer, 1999).

Local coastal observations usually form the basis for analyses on the length of the ice season. Decreasing trends in the duration of the ice season have been reported in at least eight papers (Table 1): depending on the period and region studied, some of the trends have been statistically significant while others have remained insignificant. Although most studies have indicated decreasing trends in the duration of the ice season (Table 1), increasing trends have also been reported in several locations (Fig. 3; Jevrejeva et al., 2004).

Along the Polish coast, the length of the ice season has decreased by 1–3 days per decade in the period 1896–1993 (Sztobryn, 1994; see also Girjatowicz and Kozuchowski, 1995). Sztobryn and Stanislawczyk (2002) found, however, large spatial differences in the sea ice climate over small regions in the south-eastern Baltic Sea. Girjatowicz and Kozuchowski (1999) analyzed the ice conditions in the region of the Szczecin Lagoon in the period from 1888 to 1995, and found a statistically significant decreasing trend in the duration of the ice season. Haapala and Lepparanta (1997) concluded that at the Finnish coast the length of the ice season shows a decreasing trend (Table 1), as also does the probability of annual ice occurrence in Utö (Northern Baltic Proper). Tarand (1993) and Tarand and Nordli (2001) discovered that over the last 500 years the sea ice break-up dates in the port of Tallinn have become earlier since about the mid-19th century, and that the changes have been particularly large during the latest decades. This was confirmed by Jaagus (2006), who analyzed data from nine Estonian stations in the period 1949/50–2003/04. A decrease by more than a month in the duration of the ice season was observed in the West Estonian Archipelago, while at the southern coast of the Gulf of Finland the decrease has been insignificant. Only three stations at the west coast of Estonia showed a statistically significant change towards later dates of freezing (Jaagus, 2006).

The sea ice and air temperature time series along the Estonian coast in the period of 1900–1990 were analyzed by Jevrejeva (2000). At the end of the study period, the date of a stabilized transition of the air temperature to sub-zero values was some 8–14 days later than in early 1900s, while the onset date of melting air temperatures has become 10–15 days earlier. The number of days with sea ice has decreased by 5–7 days in a century in the Gulf of Finland, and by 5–10 days in the Gulf of Riga. These changes have been associated with a climatic warming of 0.5–1.0 °C in Estonia in November–April in the period of 1900–1990; the warming has, however, been statistically significant at 99.9% confidence level only at one of the eight stations analyzed by Jevrejeva (2000). Analyzing the historical record of ice break-up at the port of Riga in

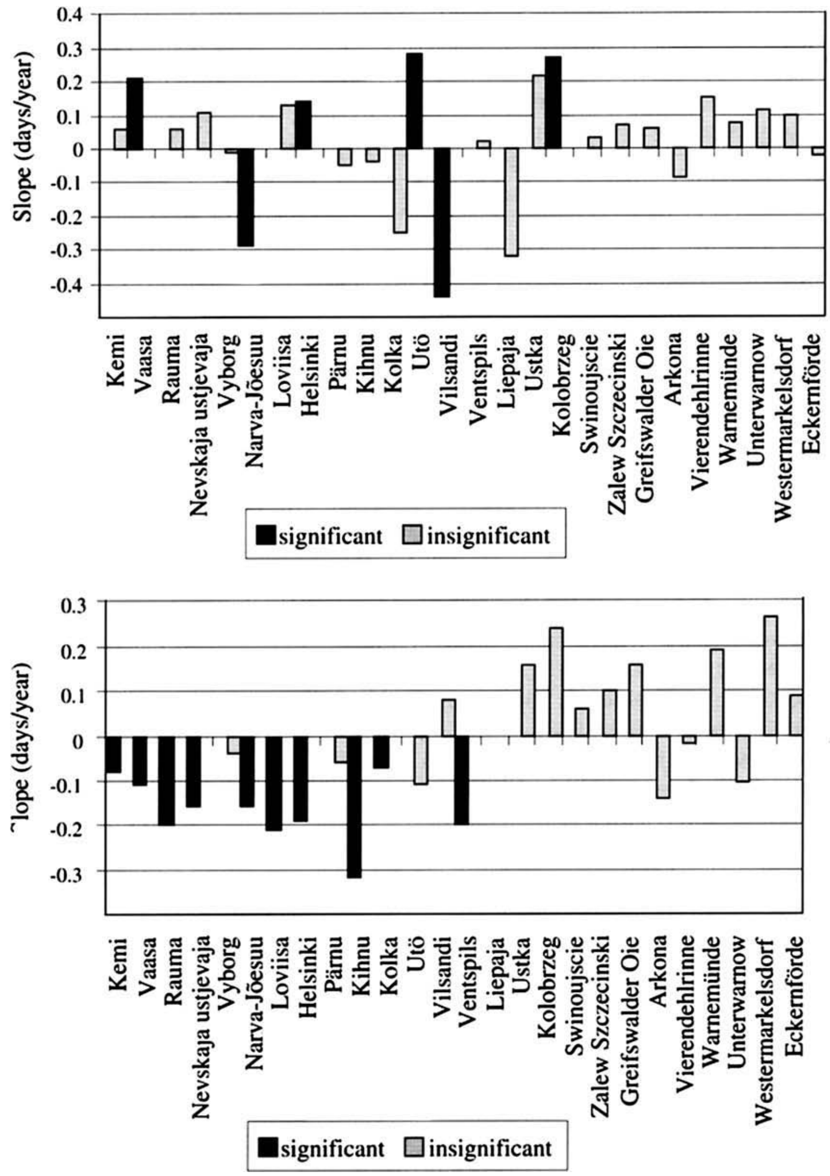

Fig. 3. Slopes of the trends in time series of date of freezing (above) and break-up (below) in the Baltic Sea (1900–2000); black and grey columns represent statistically significant ($p < 0.05$) and insignificant ($p > 0.05$) slopes, respectively. The coastal stations Kemi, Vaasa, and Rauma locate in the Gulf of Bothnia; Nevskaja ustjevaja, Vyborg, Narva-Joesuu, Loviisa, and Helsinki in the Gulf of Finland; Pärnu, Kihnu, and Kolka in the Gulf of Riga; Utö in the Archipelago Sea; Vilsandi, Ventspils, and Liepaja in the eastern coast of the Baltic Proper, while the rest of the stations locate at the southern and south-western coasts of the Baltic Sea. Redrawn from Jevrejeva et al. (2004).

1529–1990, Jevrejeva (2001) detected a decreasing trend of about 2.0 days per century for the break-up dates for severe winters (statistically significant at the 99.9% level). For mild and average winters, no statistically significant trends were detected.

Jevrejeva et al. (2004) presented a thorough analysis of twentieth-century time series at 37 coastal stations around the Baltic Sea. In general, the observations showed a tendency towards milder ice conditions. Among variables studied, the largest change has occurred in the length of the ice season, which has decreased by 14–44 days in a century, and it is mostly due to the earlier ice break-up. The trends in the time series showed, however, large spatial variations in the Baltic Sea. The trends in the dates of freezing and break-up are shown in Fig. 3; we see that statistically significant trends with an opposite sign can be found even across the Gulf of Finland (date of freezing at Helsinki and Narva-Joesuu). Break-up in the north is characterised by a statistically significant decreasing trend showing earlier (8–20 days) break-up, while in the south, trends are insignificant but have a tendency for a later date of break-up.

2.3. Large-scale atmospheric forcing on the ice climate

During the last decade, increased attention has been paid on the relation between the inter-annual variations in the Baltic Sea ice conditions and the indices of the North Atlantic Oscillation (NAO) and Arctic Oscillation (AO). We show the MIB and the fields of 2-m air temperature and 10-m wind vector over the Baltic Sea region as averaged for winters (January–March) with the NAO index $\geqslant 0.5$ (Fig. 4, upper panel) and the NAO index $\leqslant -0.5$ (Fig. 4, lower panel). The NAO indices as well as the 2-m air temperatures and 10-m winds were based on the reanalysis of the US National Center for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR). High positive NAO indices reflect strong westerlies over the North Atlantic, which bring mild and moist maritime air to the European continent. Negative NAO

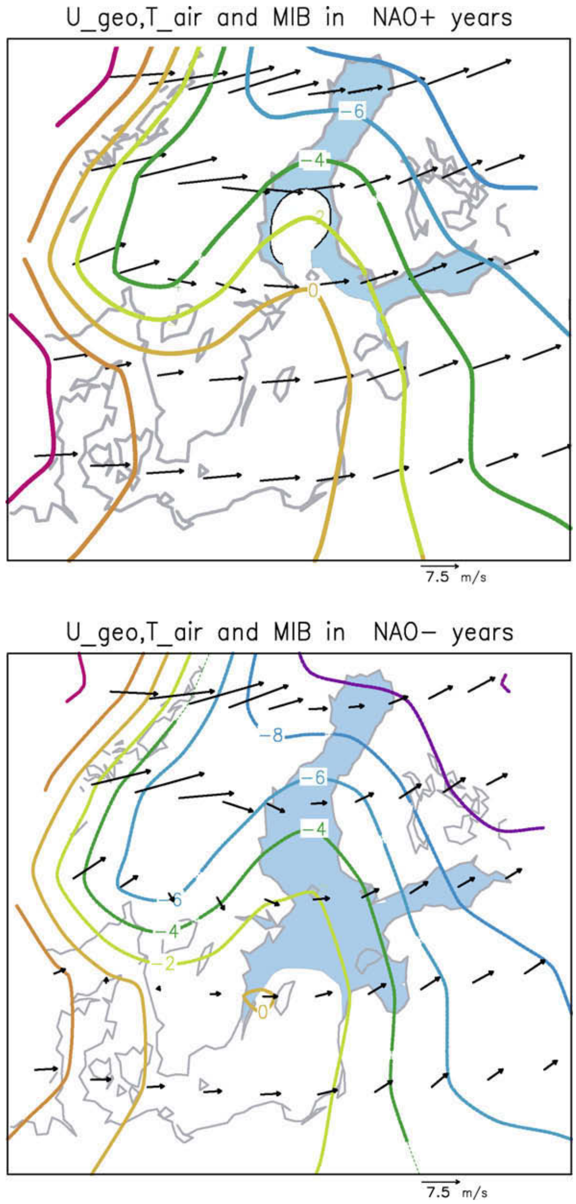

Fig. 4. Mean MIB, 2-m air temperature and 10-m wind vector over the Baltic Sea region in winters (January–March) with (above) NAO index ⩾ 0.5 and (below) NAO index ⩽ −0.5. The NAO indices, air temperatures, and winds are based on the NCEP/NCAR reanalysis.

indices express a weakening or even blocking of the westerly airflow over the Atlantic Ocean. The main difference in the 10-m wind fields over the Baltic Sea is that in years with the NAO index ⩾ 0.5 the magnitude of the vector average is approximately twice as large as in years with the NAO index ⩽ −0.5. In the latter years, there is no prevailing westerly flow over the Skagerrak and Kattegat, seen as very small vector averages, but over the Baltic Proper a westerly flow prevails even with the NAO index ⩽ −0.5. On the other hand, the effect of NAO on the 2-m air temperature is very uniform over the study region: the winters with the NAO index ⩾ 0.5 are approximately 2 K warmer than the winters with the NAO index ⩽ −0.5.

During the winters with the NAO index ⩾ 0.5, the average MIB is 121,000 km^2, with a range from 45,000 to 337,000 km^2, while during winters with the NAO index ⩽ −0.5, the average MIB is 259,000 km^2, with a range from 150,000 to 405,000 km^2. Koslowski and Loewe (1994) have pointed out that, depending on the exact location and size of the high- and low-pressure areas, in rare cases a locally severe ice winter in the south-western Baltic Sea can develop even in spite of a high NAO winter index. This is the case also for the MIB, which in some cases is very large even in winters with a positive NAO index, such as 1985/1986 (Fig. 5). In this winter, the monthly NAO indices from December to March were 0.22, 1.11, −1.00, and 1.71, respectively, i.e., the large-scale atmospheric circulation favoured extensive ice formation only in February. Also the results of Omstedt and Chen (2001) and Jaagus (2006) suggest that February is the key month, when the large-scale circulation plays a main role in determining the MIB.

In the Baltic Sea, NAO seems to affect mostly the late-winter temperature (January–March) with a significant impact also on the mid-spring (April–May) period, when the air temperature is strongly correlated to the ice break-up dates (Yoo and D'Odorico, 2002). Moving correlation analyses have demonstrated that the relationship between the NAO index and MIB is not constant in time (Omstedt and Chen, 2001; Janssen, 2002; Schrum and Janssen, 2002; Chen and Li, 2004). Meier and Kauker (2002) have interestingly pointed out that during two periods, around 1926 and 1966, the correlation has increased simultaneously with improvements in the observation methods for MIB. Changes in the NAO–MIB relationship can, however, also be solely due to changes in the location of the atmospheric pressure patterns (Koslowski and Loewe, 1994; Chen and Li, 2004).

Considering periodicity in the indices, Jevrejeva and Moore (2001) found out that the time series of ice break-up date reflect variations in the winter AO index in the 13.9-year period, but not in the NAO 7.8-year period. These analyses were extended by Jevrejeva et al. (2003), who calculated cross-wavelet power for the time series, and found out that the times of largest variance in the Baltic Sea ice conditions were in excellent agreement with significant power in the AO at 2.2–3.5, 5.7–7.8, and 12–20 year periods (previously Alenius and Makkonen (1981) had detected the most distinct cycles in the MIB at the periods of 3.5, 5.2, 8, and 13 years). It is noteworthy that Jevrejeva et al. (2003) found similar patterns also with the Southern Oscillation Index and El Nino sea surface temperature time series. Also according to Omstedt et al. (2004b), 90% of the variance of the time series is for the time scales shorter than 15 years. A concern of these studies is, however, that they have assumed the long-term time series of the MIB as homogeneous in accuracy (compare to Section 2.1).

Jaagus (2006) analyzed the freezing and break-up dates near the Estonian coast in relation to large-scale atmospheric circulation. Although a significant shift towards a later date in the first appearance of sea ice was found on the western Estonian coast, this trend could not be explained by the trends in the circulation parameters. On the other hand, the date of ice break-up and, hence, the length of the ice season were strongly related to the NAO and AO indices and the frequency of the zonal westerly circulation type. Using the conditional Mann–Kendall test, Jaagus (2006) demonstrated that the significant decreasing trends in the duration of the ice season near the Estonian coast during 1949/50–2003/04 are caused by the increasing intensity of westerlies in winter, especially in February, and by a corresponding decrease in the frequency of meridional circulation types.

So far most studies on the large-scale atmospheric forcing on sea ice have addressed the ice extent and duration of the ice season. Also the total mass of ice is a climatologically very relevant variable, but we lack good data on it over large regions. Over a small region, such as the coastal areas of Schleswig-Holstein, the ice thickness and concentration can be observed sufficiently accurately. Koslowski and Loewe (1994) calculated the areal ice volume, which can be visualized as an equivalent ice thickness in a partly ice-covered sea. Analogously to degree days (accumulated degrees of frost), they further calculated the accumulated areal

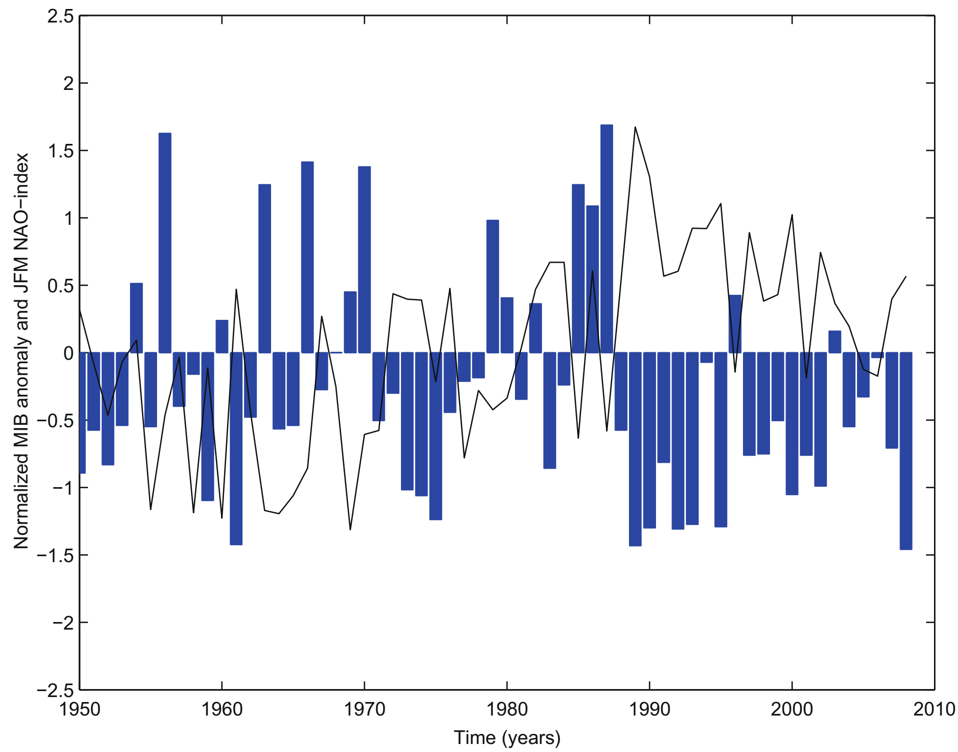

Fig. 5. Time series of the mean January–March North Atlantic Oscillation Index (line) and the annual maximum ice extent of the Baltic Sea, MIB (Seinä and Palosuo, 1996) (bars). The MIB is presented as an anomaly of the normalized time series, and the NAO indices are obtained from the NOAA Climate Prediction Center, calculated according to Barnston and Livezey (1987).

ice volume, and showed that in the period from 1879 to 1992 it was negatively correlated with the NAO winter index (compare to Fig. 5).

On the basis of the data on the accumulated areal ice volume from the 1878 to 1993 period, Koslowski and Glaser (1995) reconstructed the ice winter severity since 1701 for the south-western Baltic Sea, and Koslowski and Glaser (1999) extended the calculations for the period from 1501 to 1995. Around 1800 the ice production in the south-western Baltic Sea was three times larger than today, while the present-day ice winter regime has lasted since about 1860. This conclusion based on data from the south-western Baltic Sea is approximately in agreement with observations on the ice extent in the whole Baltic Sea: according to Omstedt and Chen (2001), the shift towards a warmer climate took place in 1877. This was associated with a period of an increased low-pressure activity (Omstedt et al., 2004b) related to the end of the Little Ice Age. Omstedt and Chen (2001) also point out that a colder climate is associated with higher variability in the ice extent (compare to Fig. 1) and with a higher sensitivity of the ice extent to changes in winter air temperature. They further developed a statistical model that links the ice extent and atmospheric circulation. The model was based on monthly indices of the westerly and southerly winds as well as the total vorticity. Despite of its simple form the statistical model achieved a skill only slightly lower than that of a numerical model in predicting the inter-annual variability. Omstedt and Chen (2001) argue that such a statistical model could be a useful tool in predicting the mean ice extent on the basis of monthly atmospheric pressure fields, which are predicted applying a general circulation model.

The MIB is also statistically related to the duration of the snow cover (Jaagus, 1999). This relationship is assumed to partially originate from the effect of the large-scale atmospheric circulation on both snow and ice cover, and it remains unknown how much the continental snow cover itself can affect the Baltic Sea ice conditions through its effects on the regional air temperature.

The presence of sea ice in the Baltic Sea affects the intensity of mesoscale cyclones as well as marine and coastal weather conditions (Gustafsson et al., 1998; Niros et al., 2002), but the Baltic Sea is not large enough to affect the large-scale atmospheric circulation. The ice cover in the Baltic Sea also reduces wind-induced water level variations (Lisitzin, 1957).

3. Thermodynamics, structure, and properties of sea ice

The structure, properties, and thermodynamics of sea ice are closely interrelated. Sea ice and snow thermodynamics are controlled by the exchange of heat at the ice–ocean and air–snow interfaces (or, in the case of bare ice, at the air–ice interface), penetration of solar radiation below the snow/ice surface, conduction of heat inside the snow and ice, melting and freezing, as well as vertical transport of liquid water and moist air in the snow pack. The heat budget equation inside the ice is:

$$\frac{\partial}{\partial t}(\rho_i c_i T) = \frac{\partial}{\partial z}\left(k_i \frac{\partial T}{\partial z}\right) - \frac{\partial F_{sw}}{\partial z} + q \tag{1}$$

where ρ_i is the ice density, c_i is the specific heat of ice, T is the temperature, k_i is the heat conductivity of ice, F_{sw} is the net solar radiation flux, and q is an internal heat source (e.g., release of latent heat of freezing). The boundary conditions are $k_i\ \partial T/\partial z = F_{net}$ at the surface and $T = T_f$ at the bottom, where F_{net} is the net surface heat flux and T_f is the freezing temperature of seawater. An analogous equation can be written for the snow cover. The net heat flux at the surface is:

$$F_{net} = (1-\alpha)(1-\beta)F_{sw} + F_{lw} + F_{lw} + F_H + F_E + F_Q \tag{2}$$

where $F_{sw\downarrow}$ is the downward shortwave radiation, α is the surface albedo, and β is the fraction of the absorbed shortwave radiation that penetrates through the surface. $F_{lw\downarrow}$ and $F_{lw\uparrow}$ are the downward and upward longwave radiation, respectively. F_H and F_E are the turbulent sensible and latent heat fluxes, respectively, and F_Q is the conductive heat flux at the surface. Ice growth and melt at the bottom are determined by the difference between the conductive heat flux and the oceanic heat flux at the ice bottom:

$$\rho_i L \frac{\partial h}{\partial t} = k_i \frac{\partial T}{\partial z} - F_{wi} \tag{3}$$

where h is the ice thickness, L is the latent heat of melting, and F_{wi} is the oceanic heat flux. For idealized conditions, analytical models for sea ice growth can be derived (Leppäranta, 1993).

3.1. Atmosphere–ice interaction

The thermal interaction between the Baltic Sea and the atmosphere is most vigorous in winter with partial or no ice cover. Large heat and moisture fluxes from the open water modify the overlying air mass, and sometimes generate convective snowbands with hazardous effects (Andersson and Gustafsson, 1994; Gustafsson et al., 1998). When a compact sea-ice cover is present, the turbulent surface fluxes are much smaller and the direct effects on the atmosphere are therefore weaker, but the interaction is essential for the ice thermodynamics.

A lot of attention has been paid on the fluxes at the air–snow interface (Eq. (2)), in particular in the project BALTEX–BASIS. These studies can be divided in three categories addressing (a) local-scale processes over land-fast ice, (b) spatial variations in different flow conditions, and (c) spatial averaging in the grid-scale of atmospheric and sea-ice models. The sensible heat flux F_H has been addressed in several recent studies. It is usually parameterized by the so-called bulk method:

$$F_H = \rho c_p C_H (\theta_S - \theta_Z) V_Z \tag{4}$$

where θ is the potential temperature, V is wind speed, ρ is the air density, c_p is the specific heat, and C_H is a heat transfer coefficient. The subscripts S and Z refer to the snow surface and height z in the air. C_H depends on the roughness lengths for momentum (z_0) and heat (z_T) and the atmospheric stratification. The stratification effects are expressed by stability functions Ψ_M and Ψ_H, which depend on the Obukhov length L_O

$$C_H = \frac{k^2}{(\log(z/z_0) - \Psi_M(z/L_O))(\log(z/z_T) - \Psi_H(z/L_O))} \tag{5}$$

where k is the von Karman constant (≈ 0.40). In the local scale over land-fast ice, Launiainen et al. (2001) analyzed the relation of the surface fluxes and the wind and temperature profiles, and obtained a new formula for the ratio of the roughness lengths: $z_0/z_T = 0.035\, Re^{0.98}$. The roughness Reynolds number is $Re = (z_0 \cdot V)v^{-1}$, where v is the kinematic viscosity of air. The result was found for smooth to moderately rough snow-covered sea ice. Based on the same data, Cheng et al. (2001) demonstrated the importance of accurate air–sea fluxes for modelling of sea ice thermodynamics.

Vihma and Brümmer (2002) studied the spatial variations in the atmosphere–surface exchange of heat and moisture (F_H and F_E) over the ice-edge zone. They showed that during a cold-air outbreak in the Gulf of Bothnia the atmospheric boundary layer (ABL) was strongly affected by the heat fluxes from leads. In such off-ice flows, the upwind snow/ice surface is typically close to a thermal equilibrium with the ABL, and in the downwind region, due to the large heat capacity of the sea, the surface temperature of the open sea is not much affected by the overflowing cold air. On the contrary, during advection of warm marine air mass over sea ice, the snow/ice surface temperature is strongly affected (Cheng and Vihma, 2002), while an ice surface of a limited fetch does not always have any large thermodynamic effect on the ABL (Vihma and Brümmer, 2002). Cheng and Vihma (2002) applied a two-dimensional, coupled, mesoscale atmosphere ice model to study the warm-air advection over sea ice. The model was run into a steady state under various idealized flow conditions with respect to season, cloud cover and wind speed. If the turbulent heat flux from air to snow was large enough to compensate the radiative cooling of the surface, a downward conductive heat flux was generated in the upper ice and snow layers. The stronger the surface heating, the larger was the region (downwind of the ice edge) where this downward flux occurred. The development of the stably stratified ABL downwind of the ice edge depended above all on the wind speed and cloud cover.

Except of the land-fast ice in coastal regions, the ice cover in the Baltic Sea is usually fractured. This reduces the representativeness of local point measurements, and aircraft observations are essential to measure spatially averaged fluxes in the ABL. The results of Brümmer et al. (2002a) revealed significant spatial and temporal variations in the surface fluxes; the fluxes depended above all on the large-scale weather conditions and on the state of the surface, i.e., either open water, compact land-fast sea ice, or broken sea ice (Fig. 6). The spatial variability of the heat fluxes in the ice-edge zone was small during warm-air advection and large during cold-air advection. Even in the latter conditions, however, the spatial variability of the surface was often exceeded by the spatial variability of the net radiation flux caused by an inhomogeneous cloud cover. The lead effect is seen also in the diurnal cycle of air temperature: Niros et al. (2002) observed a much smaller mean diurnal cycle in winter over the Bothnian Bay than over the nearby land areas.

Considering the parameterization of turbulent surface fluxes over a heterogeneous surface, the so-called mosaic method has become increasingly common, and has also been applied for partly ice-covered grid boxes in operational weather prediction models (e.g., HIRLAM). In the method, each model grid cell is divided into patches of surface types (e.g., ice and open water), and the surface temperature of each patch is calculated. The grid-averaged surface fluxes result from area-averaging of the fluxes over the individual surface patches. For the sensible heat flux:

$$\langle F_H \rangle = \rho\ c_p \sum_{i=1}^{N} f^i C_H^i \left(\theta_S^i - \langle \theta_Z \rangle \right) \langle V \rangle \tag{6}$$

where the angle brackets denote a grid average, N is the number of different surface types in the grid box, and f is the area fraction of the surface type. The fluxes of latent heat and momentum can be parameterized analogously. Sometimes the division between surface types is, however, not so clear: in addition to thick ice and open water, several thin and intermediate ice types may exist. In principle this could be taken into account in (6), but in practice we seldom have detailed information on the coverage of different ice types. A simpler approach is to apply a parameter-aggregation method, which is based on large-scale transfer coefficients, often called as effective transfer coefficients (C_H^{eff} for heat), for heterogeneous surfaces. For the sensible heat flux:

$$\langle F_H \rangle = \rho\ c_p C_H^{eff} (\langle \theta_S \rangle - \langle \theta_Z \rangle) \langle V \rangle \tag{7}$$

Schröder et al. (2003) applied the BALTEX–BASIS aircraft data in parameterizing the turbulent surface fluxes using this approach. The data suggested a value of $(0.9 \pm 0.3) \times 10^{-3}$ (mean ± standard deviation) for the 10-m neutral heat transfer coefficient (which equals C_H of Eq. (5) except that the stratification effects are not included). The aircraft measurements were made approximately at the height of the lowest atmospheric grid level of regional-scale models. The results should therefore be applicable for such models, in which the lowest grid level is located above the constant-flux

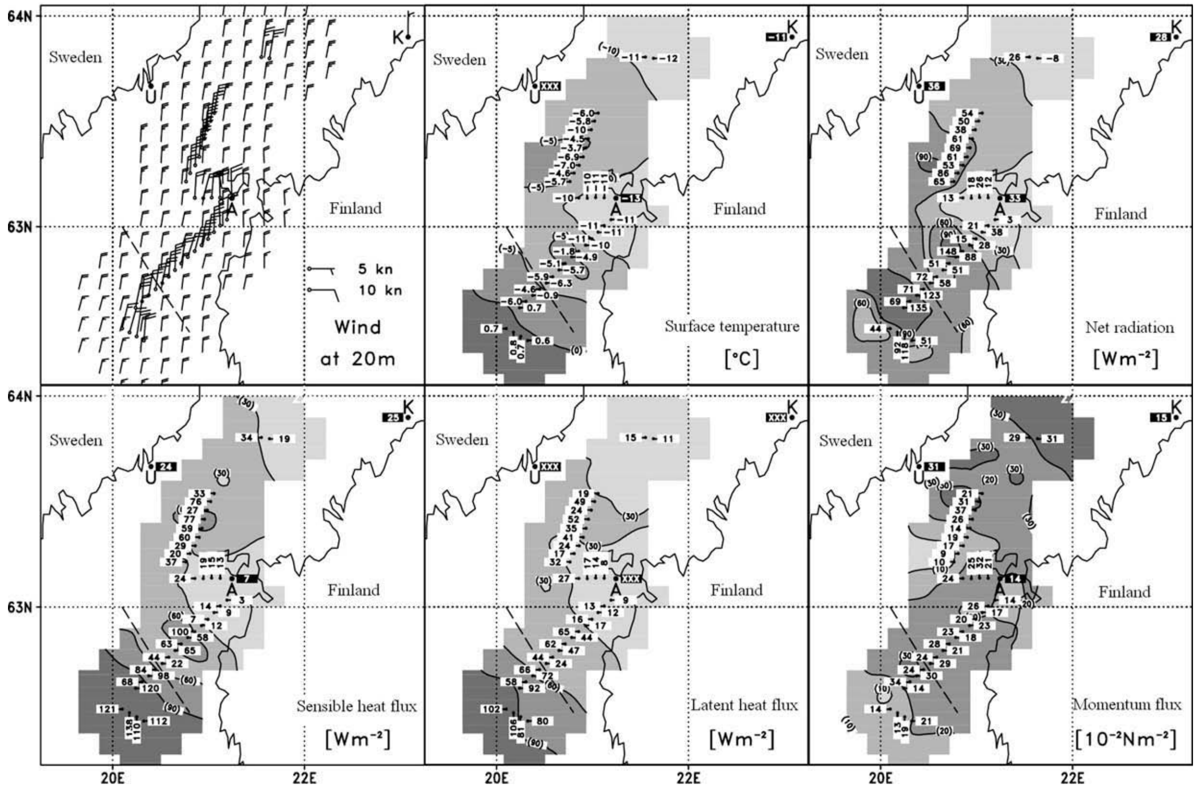

Fig. 6. Horizontal distribution of wind, net radiation, surface temperature, sensible heat flux, latent heat flux, and momentum flux over the Gulf of Bothnia during a cold-air outbreak on 5 March 1998. The results are based on aircraft observations at flight altitudes of 20–30 m, except the numbers marked on black rectangles, which represent observations at three ice stations. The dashed line marks the outermost ice edge. Redrawn from Brümmer et al. (2002a).

layer. The parameter-aggregation method may, however, sometimes yield a wrong sign for F_H (Vihma, 1995).

If non-coupled to atmospheric models, many sea-ice models still use simple parameterizations for the incoming radiative fluxes ($F_{sw\downarrow}$ and $F_{lw\downarrow}$). Considering longwave radiation, these parameterizations are typically based on easily observable quantities, such as the 2-m air temperature and humidity and the cloud fraction. In cold temperatures, however, various formulae presented in the literature differ a lot from each other (Launiainen and Cheng, 1998; Pirazzini et al., 2001). Many formulae underestimate $F_{lw\downarrow}$ under clear-skies, particularly in cold conditions, and simple parameterizations perform less well than multi-layer radiative transfer schemes (Niemelä et al., 2001a). Also for the cloud effect on $F_{sw\downarrow}$, simple schemes (usually dependent on the total cloud cover) perform poorly (Niemelä et al., 2001b). Hence, we strongly recommend that atmospheric model results, based on the multi-layer radiative transfer schemes, should be applied in sea-ice models instead of the simple parameterizations. It should be stressed that the modelled sea ice extent may strongly depend on the details of the atmospheric radiation scheme (Döscher et al., 2002).

3.2. Processes in ice and snow

During the last 10 years, significant advance has been made in studies on the snow/ice surface albedo, penetration of solar radiation in snow and ice, sub-surface melting, and formation of snow ice (snow transformed to ice due to flooding of seawater) and superimposed ice (refrozen snow melt or rain).

Saloranta (1998, 2000) was the first to model the snow-ice formation demonstrating its importance for sea-ice mass balance in the Baltic Sea. In BALTEX–BASIS, Lundin (2001) studied the snow influence on land-fast ice thickness, and found out that an increased mean snow thickness over a wide area generates flooding, which increases the ice thickness via snow-ice formation. Snow thickness variations on smaller scales do not generate flooding but have an opposite effect on the ice thickness: due to an enhanced insulation effect, a locally thicker snow cover results in locally thinner ice. Granskog et al. (2003) studied the land-fast sea ice on 15 sites along the Finnish coast during winter/spring 2000, and observed that, on average, 18–21% of the total sea-ice mass was composed of snow or precipitation. Subsequent observations have, however, indicated large inter-annual variations in the amount of snow-ice and superimposed ice in the Baltic Sea. Granskog et al. (2004) observed that the contribution of snow-ice and superimposed ice to the total land-fast ice thickness in the Santala Bay, Gulf of Finland, varied from 0% to 35% between different winters. They concluded that the contribution strongly depends on the amount and timing of snow accumulation and timing of snowmelt-freeze processes, which all exhibit large year-to-year variation. Cheng et al. (2003) applied a one-dimensional model and BALTEX–BASIS data from the winters of 1998 and 1999: the results indicated that the refreezing of the surface melt water was the primary source of superimposed ice formation. Granskog et al. (2006b) made detailed observations on the superimposed ice formation throughout the snow melt period in the Gulf of Bothnia in spring 2004. Both refreezing of melt water and freezing of wet precipitation contributed to the superimposed ice formation. Diurnal and synoptic-scale temperature variations were essential to generate melt and refreezing. In the Baltic Sea with brackish water, the ice bottom temperatures are only slightly below 0 °C, and the refreezing is therefore strongly controlled by the conductive heat flux towards the periodically colder snow surface.

Due to the brackish water, sea ice in the Baltic Sea has a low salinity and brine volume. The structure of sea ice is in any case basically similar to that in Polar Oceans with preferred horizontal *c*-axis, jagged grain boundaries, and a substructure within the grains associated with brine layers (Granskog et al., 2006a). Characteristic differences do, however, exist (Kawamura et al., 2001), and in proximity of river estuaries the Baltic Sea ice is very similar to freshwater ice (Palosuo, 1961). The highest salinities are usually observed at the uppermost parts of the ice cover (Palosuo, 1963; Granskog et al., 2004, 2006a). Ice salinity and brine volume are important for the relationships between the physical, chemical, structural, and optical properties in the ice cover. Still 10 years ago, there were, however, very few studies on these relationships. The observations in the Santala Bay and other locations on the land-fast ice have partly filled this gap. Kawamura et al. (2001) analyzed the ice, snow and water samples collected at Santala Bay once a week from January to April 1999, and found that the ice was composed of a granular upper layer, occupying approximately one-third of the entire ice thickness, and underlying columnar ice towards the bottom (Fig. 7; photography of thin vertical slides of ice have become a popular method to analyse the ice structure). The granular ice consisted of two layers with different origins, i.e., snow ice and superimposed ice. A transition layer may also exist between the granular and columnar ice layers (Granskog et al., 2003). Granskog et al. (2004) analyzed the seasonal development of the structure, salinity, and stable oxygen isotopic composition ($\delta^{18}O$) of the land-fast sea ice. The evolution of the ice salinity profile in winter 2000 is shown in Fig. 8; large and rapid variations are present especially in the uppermost parts of the ice cover. These variations were concurrent with snow-ice formation and flooding events, and accordingly strongly connected to atmospheric conditions (Granskog et al., 2004). The formation of snow ice and superimposed ice also affected the profile of $\delta^{18}O$, which showed lower values close to the surface.

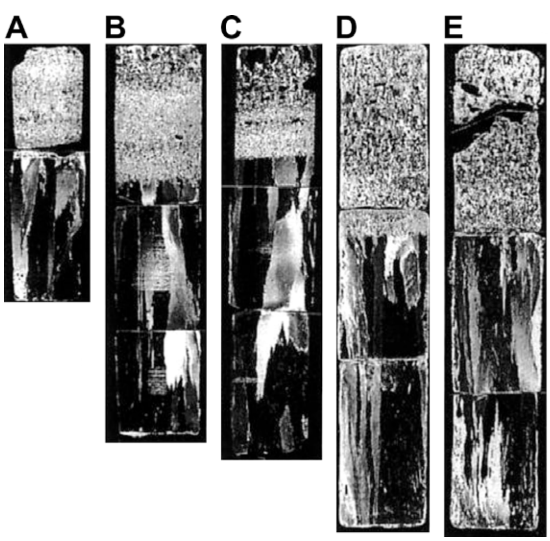

Fig. 7. Photographs of vertical thin sections of ice samples at Santala Bay, Gulf of Finland on (A) 28 January, (B) 17 February, (C) 3 March, (D) 17 March, and (E) 31 March, 1999. The largest sample (E) is 0.75 m high. Redrawn from Kawamura et al. (2001).

The shortwave radiation measurements of Rasmus et al. (2002) in spring 1998 demonstrated that the snow/ice surface albedo mostly depends on the condition of the surface, especially on the presence of liquid water or snow, and on cloud cover. Also, at this coastal measurement site, forest shading had an effect. Pirazzini et al. (2006) studied the evolution of snow and ice albedo throughout the snow melt period in the Gulf of Bothnia. The daily mean albedo ranged from 0.79 over a new snow cover to 0.30 over a bare, melting ice. The snow thickness was the most important factor affecting the albedo in the seasonal scale, but the albedo also had a large diurnal cycle (amplitude 0.07), which was generated by the snow/ice metamorphism due to melting during daylight and refreezing during night. Pirazzini et al. (2006) presented a new parameterization for albedo, dependent on snow and ice thickness, to be applied over the Baltic Sea in spring, when periods of melting and refreezing alternate, but when the ice is still relatively thick (about half a metre). Cheng et al. (2006) used the same dataset to validate a high-resolution thermodynamic model. The results indicated that with the albedo parameterized as a function of the surface temperature, which is a common practice, the modeled snow thickness is liable to became too sensitive to the atmospheric forcing. Ehn et al. (2004) made spectral irradiance measurements in the Santala Bay in March and early April 2000. The sea ice and seawater contained high amounts of dissolved and particulate matter, which effectively absorb radiation at wavelengths below 700 nm, and can potentially increase the ice melt rate. During the measurement period, the sea ice was snow-free, but highly scattering surface layers were formed in conditions of melting and refreezing. In the case of very thin (0.10 m) melting ice, the albedo was ~0.17 (Ehn et al., 2004).

The importance of vertical resolution and time step in numerical modelling of sea ice thermodynamics has been studied by Cheng (2002). In idealized cases he also compared the model results with analytical solutions. He found out that during the freezing season the influence of the resolution on model results is not significant, except for short-term predictions. During late winter and spring, when the solar radiation increases, the vertical resolution becomes much more important. In a coarse resolution model, the penetration of solar radiation into snow and ice is not described accurately; the absorbed solar radiation mostly contributes to the surface heat balance, the diurnal cycle of the surface temperature therefore becomes too large, and the sub-surface melting cannot be modelled. Cheng (2002) suggests that for process studies

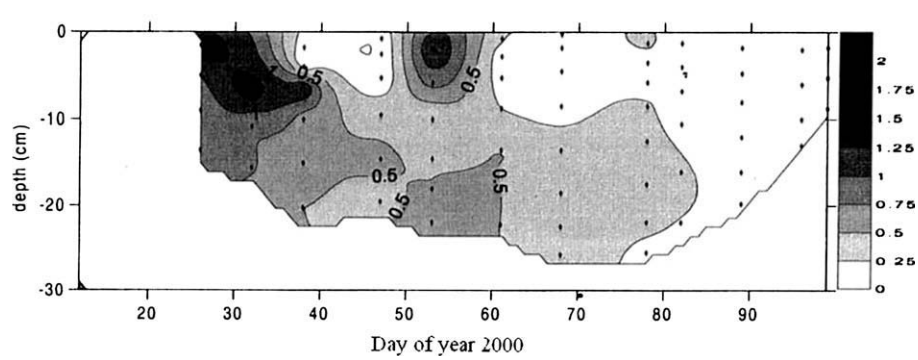

Fig. 8. Evolution of the vertical profile of ice salinity in winter 2000 in Santala Bay, Gulf of Finland. Redrawn from Granskog et al. (2004).

an ice model should apply a time step of about 10 min and a vertical resolution of 0.02–0.05 m.

In addition to model results, also observations from March 1999 in the Gulf of Bothnia have indicated a sub-surface temperature maximum at the melting point due to solar radiation penetrating below the snow surface (Cheng et al., 2003). The results of Cheng et al. (2003) suggested that sub-surface melting has an important contribution (~20%) to the total melting during early spring. In numerical modelling, the total melting is sensitive to the thermal properties of snow, while sub-surface melting is sensitive to the extinction coefficient. Launiainen and Cheng (1998) demonstrated that during the melting period in spring, a layer of new snow can enhance the melting, although it initially has a large surface albedo. The melting starts because of the high volumetric extinction of solar radiation in the new snow.

3.3. Ice–ocean interaction

Only few direct observations have been made on the ice–ocean exchange processes in the Baltic Sea. In BALTEX–BASIS, Shirasawa et al. (2001) measured turbulence below the land-fast ice in Vaasa archipelago in the Gulf of Bothnia. The results revealed some interesting features of the oceanic boundary layer (OBL): the momentum flux at the depth of 5 m from the ice bottom was ten times larger than at the depth of 0.5 m, and the heat flux from the water to the ice was very small, on average less than 1 W m^{-2}. Both these findings indicated an existence of a shallow very stable OBL below the ice. A low-salinity layer was formed below the sea ice also in the Santala Bay (Ehn et al., 2004). There it was due to discharged meltwater, which stayed below the ice until the ice ablated in April. The water-ice heat fluxes were, however, much larger at the Santala Bay: mean oceanic heat fluxes were of 38–47 W m^{-2} for the ice growing period and 54–62 W m^{-2} for the ice melting period in winters 1999–2001 (Shirasawa et al., 2002). Also basin-scale observations (Omstedt, 2001) indicated oceanic heat fluxes larger than those observed in Vaasa archipelago. The very small values there were probably due to inflow of river water. Formation of distinct stably stratified under-ice plumes of river water has been observed in the vicinity of river mouths in the Bothnian Bay (Granskog et al., 2005).

Various modelling studies have addressed the thermodynamic coupling at the ice–ocean interface. Thermodynamic sea-ice models typically prescribe the ice–ocean heat flux as a lower boundary condition, while coupled ice–ocean models predict it. An example of the latter is the one-dimensional model of Omstedt and Nyberg (1995), in which the ice–ocean heat flux is calculated on the basis of the temperature profile in the oceanic boundary layer, the laminar and turbulent Prandtl numbers, and the laminar and turbulent viscosity of sea water. Using a further developed version of the model, PROBE-Baltic, Omstedt and Rutgersson (2000) made a 15-year simulation, and obtained a result of 3 W m^{-2} for the mean water–ice heat flux. In cold winters, such as 1984/1985, 1985/1986, and 1986/1987, this value was almost 8 W m^{-2}.

The BALTEX–BASIS data yielded a mean bulk heat transfer coefficient of 3.9×10^{-4} (Shirasawa et al., 2002). Due to the large spatial variability observed in the OBL, the representativeness of this result remains so far unclear.

4. Sea Ice dynamics

4.1. Sea-ice thickness distribution

In a geophysical scale, pack ice is understood to be a continuum of fractures and mixture of several different ice types and open water. A typical ice floe is composed of undeformed and deformed ice types, and has a large variability in thickness in 10–100 m scale. The sea-ice thickness variability in a certain region is a result of the ice–ocean and ice–atmosphere heat fluxes, advection of non-uniform scalar fields, and differential ice motion. A life cycle of the pack ice begins from the cooling of the ocean, formation of the frazil ice, and initial freezing of the ocean surface due to heat loss to the atmosphere (Fig. 9). In that stage, sea ice is expanding both to the vertical and horizontal space, i.e., sea-ice thickness and area increase. In the second phase, thermodynamic growth is oriented to the vertical dimension, sea ice is thickening but the sea-ice compactness remains the same. Third stage in the life cycle is the opening, compression and deformation of the pack ice due to the differential sea-ice motion caused by winds and ocean currents. In compression, the total mass of sea ice remains unchanged, but regionally sea ice is thickening when thin undeformed sea ice is forming various thick deformed ice types (rafted, rubble, and ridged ice). Divergent ice motion generates leads into the ice pack and temporarily decreases sea-ice concentration, until new ice is formed in the leads. During winter, all these processes (thermal growth as well as opening and compression of the ice pack) vary in a synoptic scale and result in a highly variable sea-ice landscape, which is described by the ice thickness distribution function $g(h)$ (Thorndike et al, 1975). Both deformed and undeformed ice can be divided into several thickness classes and, further, each class has its inherent thickness distribution. The ice classes also differ from each other in their mechanical and thermodynamical properties. The thickness distribution is defined as:

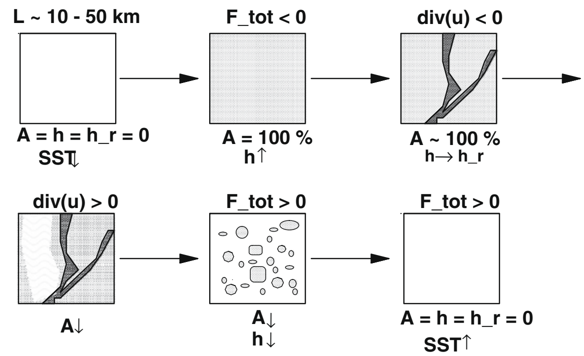

Fig. 9. The life cycle of the pack ice. The evolution is characterized by the cooling of the ocean surface, freezing of the seawater, compression and opening of the ice pack, melting of ice, and warming of the surface layer. F_{tot} denotes the total heat flux at the ocean–ice–atmosphere interface and div(u) is the divergence of the ice pack. The SST, A, h, and h_r denote the sea surface temperature, ice compactness, level-ice thickness, and ridged ice thickness; the arrows indicate whether the variable is increasing or decreasing.

$$\int_{h2}^{h1} g(h)dh = \frac{1}{R} f(h_1, h_2) \tag{8}$$

where R is the area under consideration and $f(h_1,h_2)$ is a sub-area covered by ice with thickness from h_1 to h_2. The evolution equation for $g(h)$ is,

$$\frac{\partial g(h)}{\partial t} - \vec{u} \cdot \nabla g(h) = \Theta + \Psi \tag{9}$$

where $\vec{u}$ is the horizontal ice velocity vector, Θ is the thermodynamic growth rate, and Ψ represents the redistribution of ice thickness due to deformation. The $g(h)$ can be calculated from several observational datasets (Wadhams, 1998) but only a few numerical models resolve it. This is mainly due to difficulties in determining the redistribution function Ψ. In principle, Ψ depends on $g(h)$ and the strain rate invariants.

In the classical Hibler (1979) model, $g(h)$ is approximated with two ice thickness categories: $h = h(h_0, H)$, where h_0 is the thin ice, interpreted as open water, and H is the thick ice (h > defined minimum ice thickness). Ridging of ice is taken into account since ice thickness can freely increase during the convergent ice motion, although ice concentration is constrained to be 1.0 at maximum. Most of the Baltic Sea ice models apply this approach.

To solve $g(h)$ numerically, several ice categories are needed (Hibler, 1980; Flato and Hibler, 1995). An alternative approach is to solve the ice concentration and mass for each ice category or ice type in a Lagrangian ice thickness space (Bitz et al, 2001):

$$\frac{DA_i}{Dt} = \Theta_{A_i} + \Psi_i \tag{10}$$

$$\frac{D\tilde{h}_i}{Dt} = \Theta_{h_i} + \Omega_i \tag{11}$$

where A_i is the concentration (i.e., areal fraction) of the ice category i, Θ_{A_i} denotes thermodynamical changes, and Ψ_i is the change of ice concentration due to deformation (i.e., redistribution). $\tilde{h}_i$ is the mean thickness of ice per unit area, while Θ_{h_i} and Ω_i are its changes due to thermodynamics and redistribution. Floe thickness h is obtained diagnostically since $\tilde{h}_i = hA$. Multi-category sea-ice models apply redistribution functions to describe an average evolution of the pack ice deformation processes. Several deformation processes, such as compacting, rafting, and ridging, are possible during a single time step. This mimics the real behavior of the pack ice in a continuum scale.

The first sea-ice models, in which the redistribution of ice was explicitly taken into account, were developed for operational purposes. Leppäranta (1981) made a distinction between undeformed and deformed ice. The prognostic variables of the model were the level-ice thickness, ridge density, ridge sail height, and total ice concentration. With minor modifications this scheme was used in several Baltic Sea ice models (Omstedt et al., 1994; Zhang and Leppäranta, 1995; Haapala and Leppäranta, 1996; Schrum, 1997). Shortcomings in the Leppäranta (1981) ice redistribution scheme are that the model does not include separate equations for the level ice and deformed ice concentrations, and it assumes that ridging occurs only when ice concentration reaches unity during convergence. Haapala (2000) presented a simplified ice thickness redistribution model where the pack ice was composed of open water, two different types of undeformed ice, as well as rafted ice, rubble ice, and ridged ice. The main advantage of the model is that it separates the ice types generated thermally and mechanically. The model results were compared to the operational ice charts and SSM/I remote sensing data, and the model was found to produce a realistic seasonal evolution of the pack ice. Both sub-basin and inter-basin ice characteristics were reproduced by the model. It was shown that the deformed ice production is a stepwise process related to storm activity. Most of the deformation was produced in the coastal zone, which also is an important region for thermodynamically produced ice because of the ice growth in leads. A shortcoming in the Haapala (2000) model is that it uses the Hibler (1979) parameterization for the ice strength (Eq. (12)) instead of Rothrock (1975) parameterization. The recent development (Haapala et al., 2005) overcomes this problem.

4.2. Momentum balance

A comprehensive description of sea-ice motion is given in Leppäranta (2005), which is partly based on research addressing the Baltic Sea. In the field of sea ice momentum balance, the main scientific advance during the last 10 years has been the studies by Leppäranta et al. (1998) and Zhang (2000). To better understand them, we look at the ice momentum equation in a horizontal plane: the motion of sea ice is driven by the wind stress, bottom stress due to the ocean current, and the sea surface tilt, and the motion is also affected by the internal stress of the ice pack and the Coriolis force:

$$m\left(\frac{D\vec{u}}{Dt} + f\hat{k} \times \vec{u}\right) = A(\vec{\tau}^a + \vec{\tau}^w) + mg\nabla H + \nabla \cdot \sigma \tag{12}$$

where m is the total ice and snow mass, f is the coriolis parameter, $\hat{k}$ is the upward unit vector, $\vec{\tau}^a$ is the wind stress vector, $\vec{\tau}^w$ is the water stress vector, g is the gravitational acceleration, ∇H is the sea surface tilt, and σ is the internal stress tensor. According to the scaling arguments (Leppäranta, 1998), the nonlinear advection terms and the sea surface tilt can be neglected in the calculation of the momentum balance.

The determination of the internal stress of the ice pack is the major problem in (12). The simplest assumption is the free drift law, i.e., there is no internal stress. It may hold locally but, if used in a numerical model, it leads to a large overestimation of the ice velocity and dynamic growth (Leppäranta et al., 1998). The viscous-plastic model (Hibler, 1979) is the most widely used scheme. The constitutive law is:

$$\sigma = 2\eta\dot{\varepsilon} + (\xi - \eta)tr\dot{\varepsilon}I + \frac{1}{2}PI \tag{13}$$

where η is the shear viscosity, ξ is the bulk viscosity, $\dot{\varepsilon}$ is the strain rate tensor, I is the unit tensor, and P is the ice strength. The viscous-plastic ice rheology implies that under very low strain rates the bulk and shear viscosities are constant and the model produces linear viscous behavior; otherwise the viscosities are calculated according to the plastic flow rule (Hibler, 1979). The ice strength parameter links the ice dynamics to ice thickness and compactness. For a two-level model with $h = h(h_0, H)$ it is

$$P = P^*\tilde{h}e^{-C(1-A)} \tag{14}$$

where P^* is the ice strength constant, $\tilde{h}$ is the mean ice thickness over the grid cell, and C is the compaction constant. A major difference between the two-level and multi-category ice models is that the ice strength parameter P is in multi-category models directly related to the energy consumed during deformation (Rothrock, 1975).

The ice strength constant P^* and the aspect ratio of the yield curve are important model parameters but their values can only be determined experimentally. Zhang and Leppäranta (1995) and Leppäranta et al. (1998) showed that P^* can vary from 1.0×10^4 to 5.0×10^4 N m^{-2} depending on the ice conditions in the Baltic Sea. Leppäranta et al. (1998) used radar satellite (ERS-1 SAR) data on ice motion for the verification of the modelled ice velocity fields, and noticed considerable stiffening of the ice pack as the minimum ice thickness increased from 0.10 to 0.30 m. The results supported the assumption of a plastic rheology for thick (more than 0.30 m) and compact ice, and Leppäranta et al. (1998) recommended a value of 2.5×10^4 N m^{-2} for a Baltic Sea ice model with a 10-km spatial resolution. Zhang (2000) applied a viscous-plastic sea-ice model for the Bothnian Bay and validated the results against remote sensing data and the drift of five GPS-tracked buoys in March 1997 (Fig. 10). He concluded that 3.0×10^4 N m^{-2} is the most representative value for P^*. Leppäranta et al. (2001) analyzed the same dataset, and ended up with a value of 4×10^4 N m^{-2} for P^*. Leppäranta and Wang (2002) present additional aspects on high-resolution sea-ice modelling.

The parameterization of the water and wind stress vectors forms another problem in (12). The water stress vector can be presented as (Leppäranta, 1981):

$$\vec{\tau}^w = \rho_w C_{wi}|\vec{u}_{wg} - \vec{u}|\left[\cos\theta_w(\vec{u}_{wg} - \vec{u}) + \sin\theta_w\ \hat{k} \times (\vec{u}_{wg} - \vec{u})\right] \tag{15}$$

where ρ_w is the water density and C_{wi} is the water–ice drag coefficient. $\vec{u}_{wg}$ is the geostrophic current velocity, and θ_w is the deviation

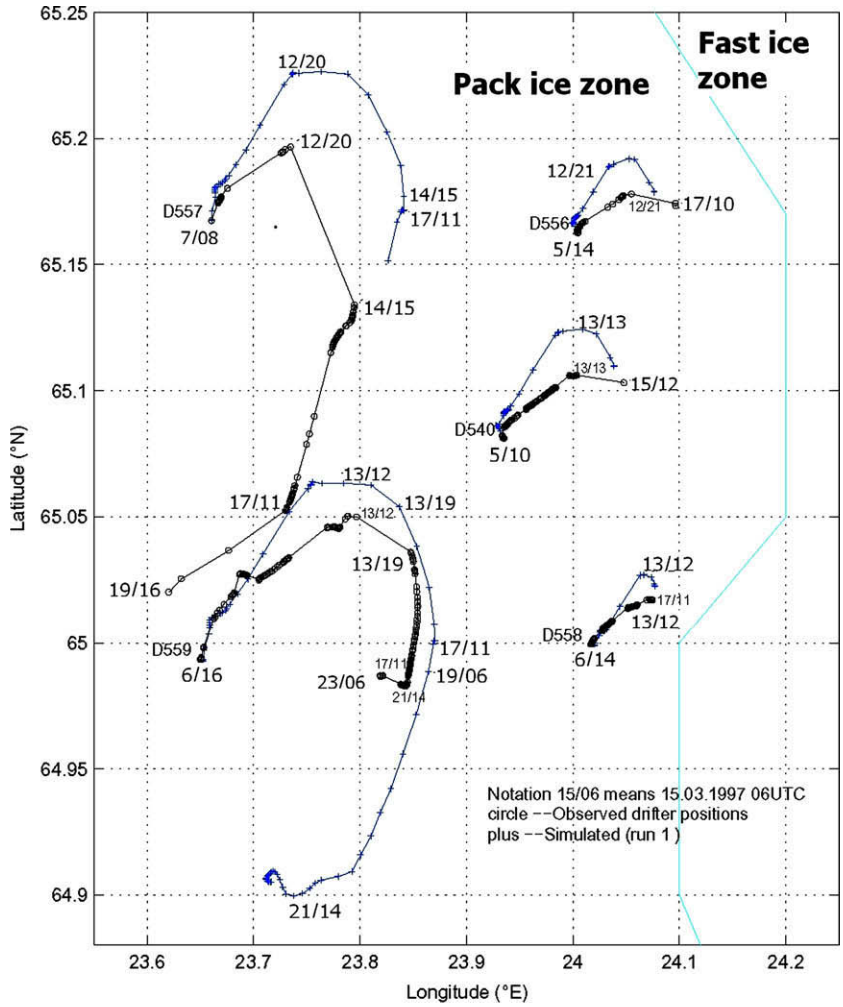

Fig. 10. Observed (circle marks) and simulated (plus marks) GPS drifter trajectories in the Bothnian Bay in March 1997. Redrawn from Zhang (2000).

angle between the current and the ice drift. Leppäranta and Omstedt (1990) found 3.5×10^{-3} as a representative value for C_{wi} in the Baltic Sea, and this value has been subsequently applied in modelling studies (Uotila, 2001; Zhang, 2000). No newer attempts have been made to estimate the value of C_{wi}.

Despite of the recent advance, the modelling of sea ice dynamics in the Baltic Sea is still challenging; this is partly due to the small size of the basins, and consequently a strong influence of boundaries (coastline and the zone of land-fast ice) restricting the ice drift. Even with a state-of-the art model, differences between the observed and modeled drift trajectories can be significant (Fig. 10). In ocean conditions with a divergent mean ice velocity field, as in the Antarctic, even very simple parameterizations can result to a closer agreement of observed and calculated trajectories (Vihma and Launiainen, 1993).

4.3. Atmospheric forcing

In a local scale, the wind stress $\vec{\tau}^a$, i.e., the air–ice momentum flux, depends on the wind speed, the surface roughness length z_0, and the thermal stratification in the atmospheric surface layer:

$$\vec{\tau}^a = \rho C_{DZ} V_Z^2; \quad C_{DZ} = \left(\frac{k}{\ln(z/z_0) - \psi_M(z/L_O)}\right)^2 \tag{16}$$

where C_{DZ} is the air–ice drag coefficient, and z indicates the reference height for C_{DZ} and V_{DZ}. The results of Launiainen et al. (2001) from the land-fast ice in the Gulf of Bothnia indicated that the local z_0 did not depend on the wind speed. In a model parameterization of the momentum flux, the ice conditions have to be taken into account in the grid-scale, and C_{DZ} does not depend solely on the local z_0 over ice. Modelling ice drift in the Gulf of Bothnia, Uotila (2001) calculated C_{DZ} on the basis of the ice concentration, surface roughness of ice floes, form drag due to floe edges (the freeboard depends on the ice and snow thickness), and the thermal stratification in the atmospheric boundary layer (ABL). The stratification effect was essential: the improvement in the model results by including this effect was comparable to the improvement achieved by increasing the grid resolution from 18 to 5 km. Uotila (2001) also showed that in the centre of the Gulf of Bothnia the ice drift was highly wind-dependent, and a linear relationship between the wind and drift velocities explained 80% of the drift's variance. Omstedt et al. (1996) studied the ice–ocean response to wind forcing using both an analytical and a numerical model. The numerical predictions agreed well with observations, but during conditions of variable winds the analytical model did not capture the wind-dependency properly. This was due to an application of linear stress laws.

Over sea ice, the stratification in the ABL is typically stable, which reduces $\vec{\tau}^a$. Localized convection may, however, occur over leads and polynyas (Vihma and Brümmer, 2002), and this enhances the turbulent mixing and $\vec{\tau}^a$. In the case of inaccurate model results for near-surface winds (as often in conditions of a stable background stratification with localized convection), Vihma

(1995) proposed to parameterize $\vec{\tau}^a$ on the basis of the atmospheric pressure field and a geostrophic drag coefficient. Mesoscale circulations (Magnusson, 2001) may also contribute to the subgrid-scale air–ice momentum transfer. Vihma (1999) summarized various aspects of mesoscale variations in the wind stress over sea ice.

Recent observations of ice velocity in the Bothnian Bay provide insight into the ice kinematics in highly compact ice conditions. Uotila (2001) showed that internal ice stresses were important close to the coast, and the modelling of the coastal ice motion was only successful by using a high-resolution (5 km) model with a realistic ice rheology. In the Baltic Sea, the main factors in the ice momentum balance are the wind stress and the internal stress of the ice pack (Leppäranta et al., 2001). When the compactness of the ice pack is high, the internal stress plays a major role, while it is negligible when the compactness is less than 0.8; then the ice velocity is close to the free drift velocity (Leppäranta et al., 2001). According to the plastic law, the internal stress of ice is linearly proportional to the ice strength (which is related to the ice thickness), and the wind stress is proportional to square of the wind velocity. This implies that, in addition to low compactness situations, there is a possibility for the ice velocity to reach the free drift velocity, if the wind stress is considerably larger than the internal stress. Leppäranta et al., 2001 found out that for wind speeds exceeding 15 m/s the observed ice velocities were closer to the free drift speed, and the deviation angle between the ice and wind directions reached a relatively constant value in agreement with the free drift law.

5. Operational modelling

In the Baltic Sea, the time scale of operational modelling of sea ice conditions is up to 5 days, with most focus in the scale of 1–2 days, because ships usually need not to navigate in ice-covered regions for more than 2 days.

Operational numerical modelling of the Baltic Sea ice conditions began in the early 1970s. Leppäranta and Zhang (1992) implemented the viscous-plastic model of Hibler (1979) to the Baltic Sea. Zhang and Leppäranta (1995) coupled the ice model to the storm surge model, and clearly demonstrated how variations in water elevations are reduced due to the internal friction of the ice pack (reported already by Lisitzin (1957)). The model of Zhang and Leppäranta has been used for operational purposes in Finland during the latest years (Bai et al., 1995; Cheng et al., 1999). The same ice model was used in Sweden (Omstedt et al., 1994; Omstedt and Nyberg, 1995), but it was coupled to the ice–ocean box model of Omstedt (1990).

Another forecasting model, HIROMB (Kleine and Skylar, 1995; Wilhelmsson, 2001), is presently used in several institutes around the Baltic Sea. Operational sea-ice models were further developed in the IRIS project, where the main objective was to provide enhanced ice information for ship route selection, with a particular focus on ice ridges. In the project, the approach of Lensu (2003) to calculate ridge statistics has been implemented to the HIROMB and the Zhang (2000) model. In addition, a multi-category model of Haapala et al. (2005) has been applied in high-resolution sea-ice predictions.

Accurate meteorological forecasts form a prerequisite for a successful modelling of ice conditions. On the other hand, accurate numerical weather predictions are only possible if the surface boundary conditions, in particular the ice concentration, are described accurately enough (Gustafsson et al., 1998). Drusch (2006) showed that in the numerical forecasts of the European Centre for Medium-Range Weather Forecasts (ECMWF), the turbulent surface fluxes, ABL height, cloud cover, and locally also the ABL temperatures and humidities were strongly affected when the sea-ice concentration was based on the high-resolution (1 km) analyses of the SMHI instead of the operationally used global National Center for Environmental Prediction dataset with a 0.5° resolution. The cloud cover changed by up to 40% locally. The ice concentration is important particularly during cold-air outbreaks. A few recent studies focused on the validation of operational meteorological models over the Baltic Sea ice cover. Ganske et al. (2001) compared the BALTEX–BASIS rawinsonde and aircraft data with the analyses and 24-h forecasts of the regional model HIRLAM. The differences were largest during passages of cold and warm fronts. The model errors were largest near the surface, and the vertical gradients of air temperature and wind speed were too small in HIRLAM. Pirazzini et al. (2002) applied a dataset from the Gulf of Bothnia in March 1999 for HIRLAM validation. The comparison indicated that the main discrepancies were related to the snow surface and 2-m temperatures: in cold nights the temperature inversions were too weak and delayed in HIRLAM. Model validation made during the IRIS project suggests that HIRLAM underestimates near-surface wind speeds over the frozen Bothnian Bay. Brümmer et al. (2009) applied the BALTIMOS data from winter 2001 (Brümmer et al., 2002b) for validation of the atmosphere–ice interaction in the regional model REMO. The main results were that the vertical temperature stratification in the lowest 200 m over sea ice is too stable and the horizontal inhomogeneity of sea ice is insufficiently represented in REMO.

6. Climatological modelling

The first climatological sea-ice models concentrated on the thermodynamic growth of ice. A pioneering work for climatological modelling in the Baltic Sea was done by Omstedt (1990), who developed a box model for the Baltic Sea dividing the sea into various sub-basins. The vertical structure of the temperature and salinity was calculated in detail, and the horizontal advection of heat and salt was solved diagnostically. The ocean model was coupled to a one-dimensional ice model. The model was further developed by Omstedt and Nyberg (1996). Due to its simplicity, the model allowed decadal-scale simulations of ice conditions. Omstedt and Nyberg (1996) showed that ice conditions are largely controlled by the atmospheric forcing; even minor changes in the air temperature can lead to large changes in the ice extent. The model was also applied by Omstedt et al. (1997) to estimate the climatology of evaporation. They concluded that sea ice reduces evaporation from the Baltic Sea (by 8% for the period 1981–1994) and therefore increases the net precipitation. We agree on this direct effect of sea ice but point out that the presence of sea ice also reduces convection and convective precipitation in winter, which reduces the increase in the net precipitation. Hence, the comprehensive effects of sea ice on net precipitation could only be studied with a coupled atmosphere–ice–ocean model. Simulating the net precipitation accurately enough is, however, a challenge; Rutgersson et al. (2002) showed that two present day regional climate simulations for the Baltic Sea resulted in too high precipitation, too low evaporation, and thus excessive net precipitation.

Omstedt and Rutgersson (2000) simulated a 15-year period (1980–1995) with and without considering sea ice in the calculations, and found that the net effect of sea ice is to reduce the total heat loss from the Baltic Sea by 0–15 W m^{-2}; the effect was largest in coldest winters. With the sea-ice cover, there was a small (1 W m^{-2}) mean net heat input from the atmosphere to the Baltic Sea.

Haapala and Leppäranta (1996) presented the first seasonal simulations of the Baltic Sea ice conditions with the evolution of sea ice calculated in two dimensions. Their model was based on the Hibler (1979) viscous-plastic rheology, the Semtner (1976) thermodynamic model, and the Leppäranta (1981) ice thickness redistribution schema. The ice model was coupled to a simple ocean model. In the most advanced Baltic Sea models, a three-dimensional primitive-equation ocean model is coupled to a two-dimensional dynamic–thermodynamic ice model. The first such modelling results were presented by Lehmann and Krauss (1995), who coupled a high-resolution free-surface ocean model to a viscous-plastic ice model. The model (BSIOM) applications have not particularly focused on sea ice, but have demonstrated that the inclusion of a realistic description of the sea-ice cover in the Baltic Sea is important in simulating the transports of heat, salt and water, the wind-driven and thermohaline circulation (Lehmann and Hinrichsen, 2000), and the effects of atmospheric forcing on upwelling (Lehmann et al., 2002). Other advanced models were presented by Schrum (1997) and Meier (1999). Schrum (1997) coupled a three-dimensional shelf ocean model to an extended version of Leppäranta and Zhang's (1992) sea-ice model. The model has been used in analyzing the influence of the NAO on the circulation of the North Sea and the Baltic Sea: a high NAO index is related to an intensification of the cyclonic circulation in both seas (Schrum, 2001). The model was later coupled to an atmospheric model (Schrum et al., 2003), which generated a clear improvement compared to an uncoupled atmospheric model. The sea surface temperature connected to sea ice was among the variables most sensitive to the coupling.

The approach of Meier (1999) was to develop a numerical model suitable for parallel computing. They coupled a highly optimized primitive-equation ocean model to an elastic-viscous-plastic ice model. Meier (1999) presented results of a 13-year hindcast simulation, and showed that the ocean temperature and salinity fields and ice conditions were generally well reproduced by this Rossby Center ocean model (RCO). Ice dynamics was found to redistribute the mean ice thickness and concentration from south-western to north-eastern parts of the Gulf of Bothnia. A need to include more ice classes in the model was identified, instead of only having level ice and open water.

The RCO has been used for estimation of the future hydrographic and ice conditions in the Baltic Sea (Meier, 2002a,b). In these predictions, the RCO was forced by the Rossby Center regional atmospheric climate model (Rummukainen et al., 1999). Comparable simulations were made with the Haapala (2000) model by Tuomenvirta et al. (2000, 2001). A comparison of these two model predictions was presented in Haapala et al. (2001). Present-day climatological ice conditions and inter-annual variability were realistically reproduced by the models, except that the production of deformed ice was underestimated due to the underestimation of surface winds in the forcing data. The simulated MIB ranged from 180×10^3 to 420×10^3 km^2 in the control simulation of both models, and from 45×10^3 to 270×10^3 km^2 in the scenario simulation with a 150% increase in the atmospheric CO_2 concentration. The range of the maximum annual ice thickness was from 0.32 to 0.96 m and from 0.11 to 0.60 m in the control and scenario simulations, respectively. In contrast to earlier climate change estimates (Tinz 1996; Omstedt et al., 2000), sea ice was still formed every winter in the Northern Bothnian Bay and in the easternmost parts of the Gulf of Finland (Fig. 11). Overall, the simulated changes in quantities such as the ice extent and thickness, as well as their inter-annual variations, were relatively similar in both models. This is remarkable, because the two coupled ice–ocean model systems were independently developed and different in many aspects. This increases the reliability of future projections of ice conditions in the Baltic Sea. The most recent predictions with the Rossby Center coupled atmosphere-ocean model do not change these conclusions (Meier et al., 2004).

RCO has also been used for a long-term hindcast simulation of the Baltic Sea. Meier and Kauker (2002) simulated the period of 1902–1998 with a reconstructed atmospheric forcing. Their results are outstanding in respect of the reproduction of the vertical structure of the hydrography, salt water inflows, and the inter-annual variability of the ice conditions in the Baltic Sea.

In several regional modelling studies the main focus has been in aspects other than the ice conditions, although the sea-ice cover has been taken into account in the coupled models (Schrum and Backhaus, 1999; Rummukainen et al., 2001; Schrum et al., 2003; Döscher et al., 2002).

7. Summary and perspectives

In recent decades, with improved observation methods, increased winter navigation, and increased awareness of the climate and environmental changes, research on the Baltic Sea ice conditions has become increasingly active. Considering the ice climate, the main findings can be summarized as follows.

(1) A change towards milder ice winters has been detected from the time series of the maximum annual extent of sea ice and the length of the ice season. On the basis of the ice extent, the shift towards a warmer climate took place in the latter half of the 19th century. During the last 21 years, all ice winters in the Baltic Sea have been average, mild, or extremely mild. Since the beginning of available data from 1720, this is the longest period without severe ice winters. The record of the length of the ice season shows a decreasing trend by 14–44 days in the latest century, the exact number depending on the location.
(2) The MIB generally decreases with increasing indices of AO and NAO, but the MIB can be very large even in winters with a positive seasonal NAO index, such as 1985/1986. At least in the northern Baltic Sea, NAO affects the length of the ice season via its influence on the break-up date, but it does not correlate with the freezing date.
(3) Data on the ice thickness mostly originates from the landfast ice zone, and basically do not shows clear trends during the 20th century, except that during the last 20 years the ice thicknesses have decreased. In the northernmost Bothnian Bay, the ice thickness showed an increasing trend until 1980s.

Considering physical processes related to sea ice, we have obtained an improved understanding on various complex interactions:

(1) The atmosphere, sea ice, and the sea are closely coupled via thermodynamic and dynamic processes. BALTEX field experiments and modelling studies have demonstrated, among others, the high variability of thermodynamic surface conditions, which cannot be described by two surface types (such as ice and open water) only. From the point of view of the surface fluxes, however, the temporal and spatial variations in the cloud cover are often even more important than the spatial variations in the surface conditions.
(2) Sea ice thermodynamics and dynamics are closely interrelated. Sea ice dynamics results in opening and closing of leads, while thermodynamics results in ice formation, growth, and melt. Ice dynamics depends on the ice thickness distribution, and in turn redistributes the ice thickness via rafting and ridging. First models for this have been developed.
(3) The structure, physical properties, and thermodynamics of sea ice are closely interrelated. The penetration of solar radiation into snow and ice has been addressed by observations

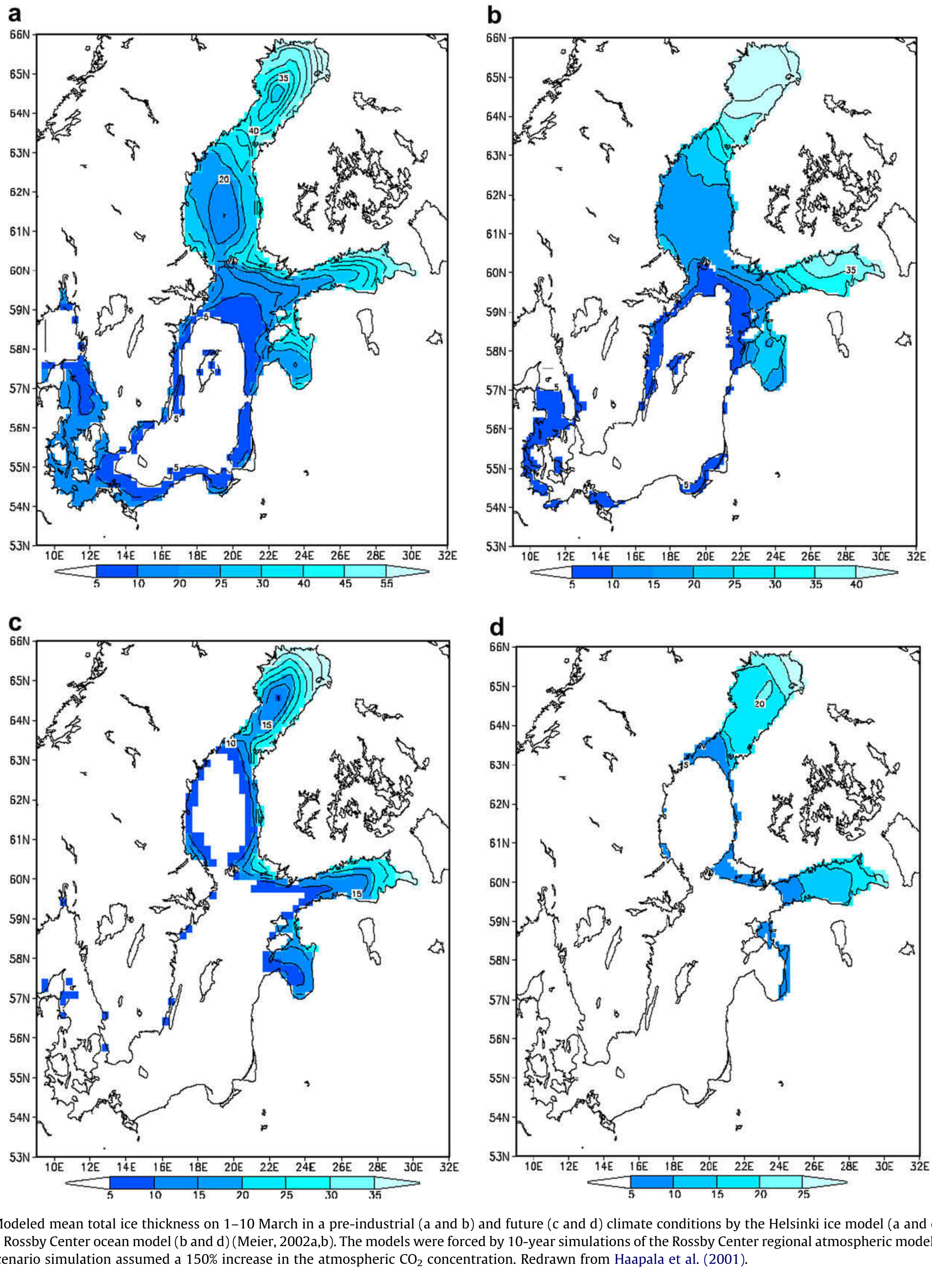

Fig. 11. Modeled mean total ice thickness on 1–10 March in a pre-industrial (a and b) and future (c and d) climate conditions by the Helsinki ice model (a and c) (Haapala, 2000) and Rossby Center ocean model (b and d) (Meier, 2002a,b). The models were forced by 10-year simulations of the Rossby Center regional atmospheric model. The future climate scenario simulation assumed a 150% increase in the atmospheric CO_2 concentration. Redrawn from Haapala et al. (2001).

and modelling, and its importance for sub-surface melting has been demonstrated. The formation of granular layers of superimposed ice and snow ice has been better quantified, both ice types being common in the Baltic Sea. Observations have indicated the importance of snow and ice thickness as well as the diurnal cycle of snow/ice metamorphism on the surface albedo, and model parameterization schemes have been improved.

(4) A few observations have demonstrated the stabilizing effect of river discharge and ice melt on the oceanic boundary layer below the ice. This strongly reduces the oceanic heat flux to the ice bottom. In general, process studies on ice–ocean interaction have been rare.

A major challenge for the future is to deepen and further integrate the knowledge on these processes, and to well organize further research efforts. Our perspectives on the future research can be summarized as follows.

(a) Increasing amounts of remote sensing data on sea ice will be available, and the most effective utilization of these data in climate-related studies is a challenge. Above all, we need to study all possibilities of applying remote sensing data to estimate the sea-ice thickness distribution. Remote sensing data can also be utilized to detect individual ice ridges or clusters of them, which is important for navigation. In addition, high-resolution data on sea-ice concentration are essential for successful modelling of the ocean and the atmosphere.
(b) The snow cover on sea ice deserves more attention, in particular during the spring melt season, which is often interrupted by periods of refreezing. Snow-ice and superimposed ice are essential for the total ice thickness in the Baltic Sea, and we need more observations and modelling efforts focusing on them. Also the permeability of sea ice deserves more attention; it is important for the infiltration of seawater through the ice, and therefore affects snow-ice formation.
(c) The snow/ice surface albedo is a critical parameter for climate modelling. In addition to its dependence on the state of the snow and ice cover, the snow/ice albedo interacts with the cloud radiative forcing, the partitioning between direct and diffuse radiation, and the multiple reflections between the snow/ice surface and the cloud base. More field experiments that address all the factors affecting snow and ice albedo are needed.
(d) The main bottleneck of the development of the sea-ice models is the lack of proper validation data. Operational ice charts contain very good information on the ice extent and the thickness of coastal land-fast ice. The latter can be used for validation of thermodynamic models. Two important ice variables are, however, missing from the monitoring activities, namely the ice velocity and the thickness of drifting ice. In principle both variables could be observed with a reasonable accuracy. Ice velocity can be measured with drifters or derived from consecutive satellite images. Deformed ice thickness is more difficult to map, but an upward-looking sonar in a fixed position would provide valuable statistics, and airborne electromagnetic mapping surveys can provide synoptic data. Utilizing such data, climate-scale studies should have more focus on deformed ice. In addition, more climate-scale studies could be made on the occurrence of various ice types.
(e) An obvious shortcoming in the present-day Baltic Sea ice and ice–ocean models is their horizontal resolution, which does not enable to resolve ocean eddies and fine structure of the pack ice. The Finnish operational sea-ice model presently uses a resolution of 1 nm ($\sim$2 km). In this scale the model produces much stronger gradients in the ice thickness field than the same model used with a resolution typical for climate simulations. Compared with satellite data and ice charts, the model with a 1-nm resolution is, however, still too coarse. To capture coastal leads and narrow deformation zones, the ice model should have a resolution of the order of 200 m. Within the next years, nested models or models with curvi-linear co-ordinates can be developed with a certain region modelled with a very high resolution. We may foresee that within a few years operational mesoscale atmospheric models will reach a horizontal resolution of the order of 1 km, which will provide more accurate forcing fields for sea-ice models. Improved vertical resolution in the atmosphere, snow, sea ice, and ocean will provide better possibilities to model stably stratified ABL and OBL, as well as the details of snow and ice thermodynamics.
(f) All sea-ice models presently used in the Baltic Sea are finite-difference models. Sea-ice models based on an alternative approach are, however, already under development for the Arctic Ocean. For example, Lindsay and Stern (2003) have developed a sea-ice model based on Lagrangian approach, and Hopkins et al. (2004) have presented a finite-element sea-ice model, which explicitly calculates the evolution of single ice floes, interaction between the ice floes, and ridge build-up. Applying these kinds of models in the Baltic Sea would most probably yield significant advance for both research and operational services.
(g) Surprisingly, despite of the computational power already available, the operational sea ice and atmospheric models applied in the Baltic Sea have so far remained uncoupled, although coupled models have been developed for climate applications. Coupling of operational models is assumed to yield better forecasts, at least in cases when large sea areas are rapidly opened or closed due to ice formation, advection, or melt, but so far there is not much basis to quantify the probable benefits.

Even if in the coming decades the ice cover in the Baltic Sea will be strongly reduced, the experiences obtained in the Baltic Sea will be very useful for research and operational activities in the Arctic and Antarctic.

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