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Changes in Climate Extremes and their Impacts on the Natural Physical Environment

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Executive Summary

This chapter addresses changes in weather and climate events relevant to extreme impacts and disasters. An extreme (weather or climate) event is generally defined as the occurrence of a value of a weather or climate variable above (or below) a threshold value near the upper (or lower) ends ('tails') of the range of observed values of the variable. Some climate extremes (e.g., droughts, floods) may be the result of an accumulation of weather or climate events that are, individually, not extreme themselves (though their accumulation is extreme). As well, weather or climate events, even if not extreme in a statistical sense, can still lead to extreme conditions or impacts, either by crossing a critical threshold in a social, ecological, or physical system, or by occurring simultaneously with other events. A weather system such as a tropical cyclone can have an extreme impact, depending on where and when it approaches landfall, even if the specific cyclone is not extreme relative to other tropical cyclones. Conversely, not all extremes necessarily lead to serious impacts. [3.1]

Many weather and climate extremes are the result of natural climate variability (including phenomena such as El Niño), and natural decadal or multi-decadal variations in the climate provide the backdrop for anthropogenic climate changes. Even if there were no anthropogenic changes in climate, a wide variety of natural weather and climate extremes would still occur. [3.1]

A changing climate leads to changes in the frequency, intensity, spatial extent, duration, and timing of weather and climate extremes, and can result in unprecedented extremes. Changes in extremes can also be directly related to changes in mean climate, because mean future conditions in some variables are projected to lie within the tails of present-day conditions. Nevertheless, changes in extremes of a climate or weather variable are not always related in a simple way to changes in the mean of the same variable, and in some cases can be of opposite sign to a change in the mean of the variable. Changes in phenomena such as the El Niño-Southern Oscillation or monsoons could affect the frequency and intensity of extremes in several regions simultaneously. [3.1]

Many factors affect confidence in observed and projected changes in extremes. Our confidence in observed changes in extremes depends on the quality and quantity of available data and the availability of studies analyzing these data. It consequently varies between regions and for different extremes. Similarly, our confidence in projecting changes (including the direction and magnitude of changes in extremes) varies with the type of extreme, as well as the considered region and season, depending on the amount and quality of relevant observational data and model projections, the level of understanding of the underlying processes, and the reliability of their simulation in models (assessed from expert judgment, model validation, and model agreement). Global-scale trends in a specific extreme may be either more reliable (e.g., for temperature extremes) or less reliable (e.g., for droughts) than some regional-scale trends, depending on the geographical uniformity of the trends in the specific extreme. '*Low confidence*' in observed or projected changes in a specific extreme neither implies nor excludes the possibility of changes in this extreme. [3.1.5, 3.1.6, 3.2.3; Box 3-2; Figures 3-3, 3-4, 3-5, 3-6, 3-7, 3-8, 3-10]

There is evidence from observations gathered since 1950 of change in some extremes. It is *very likely* that there has been an overall decrease in the number of cold days and nights, and an overall increase in the number of warm days and nights, at the global scale, that is, for most land areas with sufficient data. It is *likely* that these changes have also occurred at the continental scale in North America, Europe, and Australia. There is *medium confidence* of a warming trend in daily temperature extremes in much of Asia. *Confidence* in observed trends in daily temperature extremes in Africa and South America generally varies from *low* to *medium* depending on the region. Globally, in many (but not all) regions with sufficient data there is *medium confidence* that the length or number of warm spells or heat waves has increased since the middle of the 20th century. It is *likely* that there have been statistically significant increases in the number of heavy precipitation events (e.g., 95th percentile) in more regions than there have been statistically significant decreases, but there are strong regional and subregional variations in the trends. There is *low confidence* that any observed long-term (i.e., 40 years or more) increases in tropical cyclone activity are robust, after accounting for past changes in observing capabilities. It is *likely* that there has been a poleward shift in the main Northern and Southern Hemisphere extratropical storm tracks. There is *low confidence* in observed trends in

small-scale phenomena such as tornadoes and hail because of data inhomogeneities and inadequacies in monitoring systems. There is *medium confidence* that since the 1950s some regions of the world have experienced a trend to more intense and longer droughts, in particular in southern Europe and West Africa, but in some regions droughts have become less frequent, less intense, or shorter, for example, in central North America and northwestern Australia. There is limited to medium evidence available to assess climate-driven observed changes in the magnitude and frequency of floods at regional scales because the available instrumental records of floods at gauge stations are limited in space and time, and because of confounding effects of changes in land use and engineering. Furthermore, there is low agreement in this evidence, and thus overall *low confidence* at the global scale regarding even the sign of these changes. It is *likely* that there has been an increase in extreme coastal high water related to increases in mean sea level in the late 20th century. [3.2.1, 3.3.1, 3.3.2, 3.3.3, 3.4.4, 3.4.5, 3.5.1, 3.5.2, 3.5.3; Tables 3-1, 3-2]

There is evidence that some extremes have changed as a result of anthropogenic influences, including increases in atmospheric concentrations of greenhouse gases. It is *likely* that anthropogenic influences have led to warming of extreme daily minimum and maximum temperatures at the global scale. There is *medium confidence* that anthropogenic influences have contributed to intensification of extreme precipitation at the global scale. It is *likely* that there has been an anthropogenic influence on increasing extreme coastal high water due to an increase in mean sea level. The uncertainties in the historical tropical cyclone records, the incomplete understanding of the physical mechanisms linking tropical cyclone metrics to climate change, and the degree of tropical cyclone variability provide only *low confidence* for the attribution of any detectable changes in tropical cyclone activity to anthropogenic influences. Attribution of single extreme events to anthropogenic climate change is challenging. [3.2.2, 3.3.1, 3.3.2, 3.4.4, 3.5.3; Table 3-1]

The following assessments of the likelihood of and/or confidence in projections are generally for the end of the 21st century and relative to the climate at the end of the 20th century. There are three main sources of uncertainty in the projections: the natural variability of climate; uncertainties in climate model parameters and structure; and projections of future emissions. Projections for differing emissions scenarios generally do not strongly diverge in the coming two to three decades, but uncertainty in the sign of change is relatively large over this time frame because climate change signals are expected to be relatively small compared to natural climate variability. For certain extremes (e.g., precipitation-related extremes), the uncertainty in projected changes by the end of the 21st century is more the result of uncertainties in climate models rather than uncertainties in future emissions. For other extremes (in particular temperature extremes at the global scale and in most regions), the emissions uncertainties are the main source of uncertainty in projections for the end of the 21st century. In the assessments provided in this chapter, uncertainties in projections from the direct evaluation of multi-model ensemble projections are modified by taking into account the past performance of models in simulating extremes (for instance, simulations of late 20th-century changes in extreme temperatures appear to overestimate the observed warming of warm extremes and underestimate the warming of cold extremes), the possibility that some important processes relevant to extremes may be missing or be poorly represented in models, and the limited number of model projections and corresponding analyses currently available of extremes. For these reasons the assessed uncertainty is generally greater than would be assessed from the model projections alone. Low-probability, high-impact changes associated with the crossing of poorly understood climate thresholds cannot be excluded, given the transient and complex nature of the climate system. Feedbacks play an important role in either damping or enhancing extremes in several climate variables. [3.1.4, 3.1.7, 3.2.3, 3.3.1, 3.3.2; Box 3-2]

Models project substantial warming in temperature extremes by the end of the 21st century. It is *virtually certain* that increases in the frequency and magnitude of warm daily temperature extremes and decreases in cold extremes will occur through the 21st century at the global scale. It is *very likely* that the length, frequency, and/or intensity of warm spells or heat waves will increase over most land areas. For the Special Report on Emissions Scenarios (SRES) A2 and A1B emission scenarios, a 1-in-20 year annual hottest day is *likely* to become a 1-in-2 year annual extreme by the end of the 21st century in most regions, except in the high latitudes of the Northern Hemisphere where it is *likely* to become a 1-in-5 year annual extreme. In terms of absolute values, 20-year extreme annual daily maximum temperature (i.e., return value) will *likely* increase by about 1 to 3°C by mid-21st century and by about 2 to 5°C by the late 21st century, depending on the region and emissions scenario (considering the B1, A1B,

and A2 scenarios). Regional changes in temperature extremes will often differ from the mean global temperature change. [3.3.1; Table 3-3; Figure 3-5]

It is *likely* that the frequency of heavy precipitation or the proportion of total rainfall from heavy rainfalls will increase in the 21st century over many areas of the globe. This is particularly the case in the high latitudes and tropical regions, and in winter in the northern mid-latitudes. Heavy rainfalls associated with tropical cyclones are *likely* to increase with continued warming induced by enhanced greenhouse gas concentrations. There is *medium confidence* that, in some regions, increases in heavy precipitation will occur despite projected decreases in total precipitation. For a range of emission scenarios (SRES A2, A1B, and B1), a 1-in-20 year annual maximum 24-hour precipitation rate is *likely* to become a 1-in-5 to 1-in-15 year event by the end of the 21st century in many regions, and in most regions the higher emissions scenarios (A1B and A2) lead to a greater projected decrease in return period. Nevertheless, increases or statistically non-significant changes in return periods are projected in some regions. [3.3.2; Table 3-3; Figure 3-7]

There is generally *low confidence* in projections of changes in extreme winds because of the relatively few studies of projected extreme winds, and shortcomings in the simulation of these events. An exception is mean tropical cyclone maximum wind speed, which is *likely* to increase, although increases may not occur in all ocean basins. It is *likely* that the global frequency of tropical cyclones will either decrease or remain essentially unchanged. There is *low confidence* in projections of small-scale phenomena such as tornadoes because competing physical processes may affect future trends and because climate models do not simulate such phenomena. There is *medium confidence* that there will be a reduction in the number of mid-latitude cyclones averaged over each hemisphere due to future anthropogenic climate change. There is *low confidence* in the detailed geographical projections of mid-latitude cyclone activity. There is *medium confidence* in a projected poleward shift of mid-latitude storm tracks due to future anthropogenic forcings. [3.3.3, 3.4.4, 3.4.5]

Uncertainty in projections of changes in large-scale patterns of natural climate variability remains large. There is *low confidence* in projections of changes in monsoons (rainfall, circulation), because there is little consensus in climate models regarding the sign of future change in the monsoons. Model projections of changes in El Niño-Southern Oscillation variability and the frequency of El Niño episodes as a consequence of increased greenhouse gas concentrations are not consistent, and so there is *low confidence* in projections of changes in the phenomenon. However, most models project an increase in the relative frequency of central equatorial Pacific events (which typically exhibit different patterns of climate variations than do the classical East Pacific events). There is *low confidence* in the ability to project changes in other natural climate modes including the North Atlantic Oscillation, the Southern Annular Mode, and the Indian Ocean Dipole. [3.4.1, 3.4.2, 3.4.3]

It is *very likely* that mean sea level rise will contribute to upward trends in extreme coastal high water levels in the future. There is *high confidence* that locations currently experiencing adverse impacts such as coastal erosion and inundation will continue to do so in the future due to increasing sea levels, all other contributing factors being equal. There is *low confidence* in wave height projections because of the small number of studies, the lack of consistency of the wind projections between models, and limitations in the models' ability to simulate extreme winds. Future negative or positive changes in significant wave height are *likely* to reflect future changes in storminess and associated patterns of wind change. [3.5.3, 3.5.4, 3.5.5]

Projected precipitation and temperature changes imply possible changes in floods, although overall there is *low confidence* in projections of changes in fluvial floods. *Confidence* is *low* due to limited evidence and because the causes of regional changes are complex, although there are exceptions to this statement. There is *medium confidence* (based on physical reasoning) that projected increases in heavy rainfall would contribute to increases in local flooding, in some catchments or regions. Earlier spring peak flows in snowmelt and glacier-fed rivers are *very likely*. [3.5.2]

There is *medium confidence* that droughts will intensify in the 21st century in some seasons and areas, due to reduced precipitation and/or increased evapotranspiration. This applies to regions including southern

Europe and the Mediterranean region, central Europe, central North America, Central America and Mexico, northeast Brazil, and southern Africa. Definitional issues, lack of observational data, and the inability of models to include all the factors that influence droughts preclude stronger *confidence* than *medium* in the projections. Elsewhere there is overall *low confidence* because of inconsistent projections of drought changes (dependent both on model and dryness index). There is *low confidence* in projected future changes in dust storms although an increase could be expected where aridity increases. [3.5.1, 3.5.8; Box 3-3; Table 3-3; Figure 3-10]

There is *high confidence* that changes in heat waves, glacial retreat, and/or permafrost degradation will affect high-mountain phenomena such as slope instabilities, mass movements, and glacial lake outburst floods. There is also *high confidence* that changes in heavy precipitation will affect landslides in some regions. There is *low confidence* regarding future locations and timing of large rock avalanches, as these depend on local geological conditions and other non-climatic factors. There is *low confidence* in projections of an anthropogenic effect on phenomena such as shallow landslides in temperate and tropical regions, because these are strongly influenced by human activities such as land use practices, deforestation, and overgrazing. [3.5.6, 3.5.7]

The small land area and often low elevation of small island states make them particularly vulnerable to rising sea levels and impacts such as inundation, shoreline change, and saltwater intrusion into underground aquifers. Short record lengths and the inadequate resolution of current climate models to represent small island states limit the assessment of changes in extremes. There is insufficient evidence to assess observed trends and future projections in rainfall across the small island regions considered here. There is *medium confidence* in projected temperature increases across the Caribbean. The *very likely* contribution of mean sea level rise to increased extreme coastal high water levels, coupled with the *likely* increase in tropical cyclone maximum wind speed, is a specific issue for tropical small island states. [3.4.4, 3.5.3; Box 3-4]

This chapter does not provide assessments of projected changes in extremes at spatial scales smaller than for large regions. These large-region projections provide a wider context for national or local projections, where these exist, and where they do not exist, a first indication of expected changes, their associated uncertainties, and the evidence available. [3.2.3.1]

3.1.   Weather and Climate Events Related to Disasters

A changing climate leads to changes in the frequency, intensity, spatial extent, duration, and timing of weather and climate extremes, and can result in unprecedented extremes (Sections 3.1.7, 3.3, 3.4, and 3.5). As well, weather or climate events, even if not extreme in a statistical sense, can still lead to extreme conditions or impacts, either by crossing a critical threshold in a social, ecological, or physical system, or by occurring simultaneously with other events (Sections 3.1.2, 3.1.3, 3.1.4, 3.3, 3.4, and 3.5). Some climate extremes (e.g., droughts, floods) may be the result of an accumulation of weather or climate events that are, individually, not extreme themselves (though their accumulation is extreme, e.g., Section 3.1.2). A weather system such as a tropical cyclone can have an extreme impact, depending on where and when it approaches landfall, even if the specific cyclone is not extreme relative to other tropical cyclones. Conversely, not all extremes necessarily lead to serious impacts. Changes in extremes can also be directly related to changes in mean climate, because mean future conditions in some variables are projected to lie within the tails of present-day conditions (Section 3.1.6). Hence, the definition of extreme weather and climate events is complex (Section 3.1.2 and Box 3-1) and the assessment of changes in climate that are relevant to extreme impacts and disasters needs to consider several aspects. Those related to vulnerability and exposure are addressed in Chapters 2 and 4 of this report, while we focus here on the physical dimension of these events.

Many weather and climate extremes are the result of natural climate variability (including phenomena such as El Niño), and natural decadal or multi-decadal variations in the climate provide the backdrop for anthropogenic climate changes. Even if there were no anthropogenic changes in climate, a wide variety of natural weather and climate extremes would still occur.

3.1.1.   Categories of Weather and Climate Events Discussed in this Chapter

This chapter addresses changes in weather and climate events relevant to extreme impacts and disasters grouped into the following categories:
1) Extremes of atmospheric weather and climate variables (temperature, precipitation, wind)
2) Weather and climate phenomena that influence the occurrence of extremes in weather or climate variables or are extremes themselves (monsoons, El Niño and other modes of variability, tropical and extratropical cyclones)
3) Impacts on the natural physical environment (droughts, floods, extreme sea level, waves, and coastal impacts, as well as other physical impacts, including cryosphere-related impacts, landslides, and sand and dust storms).

The distinction between these three categories is somewhat arbitrary, and the categories are also related. In the case of the third category, 'impacts on the natural physical environment,' a specific distinction between these events and those considered under 'extremes of atmospheric weather and climate variables' is that they are not caused by variations in a single atmospheric weather and climate variable, but are generally the result of specific conditions in several variables, as well as of some surface properties or states. For instance, both floods and droughts are related to precipitation extremes, but are also impacted by other atmospheric and surface conditions (and are thus often better viewed as compound events, see Section 3.1.3). Most of the impacts on the natural physical environment discussed in the third category are extremes themselves, as well as often being caused or affected by atmospheric weather or climate extremes. Another arbitrary choice made here is the separate category for phenomena (or climate or weather systems) that are related to weather and climate extremes, such as monsoons, El Niño, and other modes of variability. These phenomena affect the large-scale environment that, in turn, influences extremes. For instance, El Niño episodes typically lead to droughts in some regions with, simultaneously, heavy rains and floods occurring elsewhere. This means that all occurrences of El Niño are relevant to extremes and not only extreme El Niño episodes. A change in the frequency or nature of El Niño episodes (or in their relationships with climate in specific regions) would affect extremes in many locations simultaneously. Similarly, changes in monsoon patterns could affect several countries simultaneously. This is especially important from an international disaster perspective because coping with disasters in several regions simultaneously may be challenging (see also Section 3.1.3 and Chapters 7 and 8).

This section provides background material on the characterization and definition of extreme events, the definition and analysis of compound events, the relevance of feedbacks for extremes, the approach used for the assignment of confidence and likelihood assessments in this chapter, and the possibility of 'surprises' regarding future changes in extremes. Requirements and methods for analyzing changes in climate extremes are addressed in Section 3.2. Assessments regarding changes in the climate variables, phenomena, and impacts considered in this chapter are provided in Sections 3.3 to 3.5. Table 3-1 provides summaries of these assessments for changes at the global scale. Tables 3-2 and 3-3 (found on pages 191-202) provide more regional detail on observed and projected changes in temperature extremes, heavy precipitation, and dryness (with regions as defined in Figure 3-1). Note that impacts on ecosystems (e.g., bushfires) and human systems (e.g., urban flooding) are addressed in Chapter 4.

3.1.2.   Characteristics of Weather and Climate Events Relevant to Disasters

The identification and definition of weather and climate events that are relevant from a risk management perspective are complex and depend on the stakeholders involved (Chapters 1 and 2). In this chapter, we focus on the assessment of changes in 'extreme climate or weather events' (also referred to herein as 'climate extremes' see below and Glossary), which generally correspond to the 'hazards' discussed in Chapter 1.

Box 3-1 | Definition and Analysis of Climate Extremes in the Scientific Literature

This box provides some details on the definition of climate extremes in the scientific literature and on common approaches employed for their investigation.

A large amount of the available scientific literature on climate extremes is based on the use of so-called 'extreme indices,' which can either be based on the probability of occurrence of given quantities or on threshold exceedances (Section 3.1.2). Typical indices that are seen in the scientific literature include the number, percentage, or fraction of days with maximum temperature (Tmax) or minimum temperature (Tmin), below the 1st, 5th, or 10th percentile, or above the 90th, 95th, or 99th percentile, generally defined for given time frames (days, month, season, annual) with respect to the 1961-1990 reference time period. Commonly, indices for 10th and 90th percentiles of Tmax/Tmin computed on daily time frames are referred to as 'cold/warm days/nights' (e.g., Figures 3-3 and 3-4; Tables 3-1 to 3-3, and Section 3.3.1; see also Glossary). Other definitions relate to, for example, the number of days above specific absolute temperature or precipitation thresholds, or more complex definitions related to the length or persistence of climate extremes. Some advantages of using predefined extreme indices are that they allow some comparability across modelling and observational studies and across regions (although with limitations noted below). Moreover, in the case of observations, derived indices may be easier to obtain than is the case with daily temperature and precipitation data, which are not always distributed by meteorological services. Peterson and Manton (2008) discuss collaborative international efforts to monitor extremes by employing extreme indices. Typically, although not exclusively, extreme indices used in the scientific literature reflect 'moderate extremes,' for example, events occurring as often as 5 or 10% of the time. More extreme 'extremes' are often investigated using Extreme Value Theory (EVT) due to sampling issues (see below). Extreme indices are often defined for daily temperature and precipitation characteristics, and are also sometimes applied to seasonal characteristics of these variables, to other weather and climate variables, such as wind speed, humidity, or to physical impacts and phenomena. Beside analyses for temperature and precipitation indices (see Sections 3.3.1 and 3.3.2; Tables 3-2 and 3-3), other studies are, for instance, available in the literature for wind-based (Della-Marta et al., 2009) and pressure-based (Beniston, 2009a) indices, for health-relevant indices (e.g., 'heat index') combining temperature and relative humidity characteristics (e.g., Diffenbaugh et al., 2007; Fischer and Schär, 2010; Sherwood and Huber, 2010), and for a range of dryness indices (see Box 3-3).

Extreme Value Theory is an approach used for the estimation of extreme values (e.g., Coles, 2001), which aims at deriving a probability distribution of events from the tail of a probability distribution, that is, at the far end of the upper or lower ranges of the probability distributions (typically occurring less frequently than once per year or per period of interest, i.e., generally less than 1 to 5% of the considered overall sample). EVT is used to derive a complete probability distribution for such low-probability events, which can also help analyzing the probability of occurrence of events that are outside of the observed data range (with limitations). Two different approaches can be used to estimate the parameters for such probability distributions. In the *block maximum approach*, the probability distribution parameters are estimated for maximum values of consecutive blocks of a time series (e.g., years). In the second approach, instead of the block maxima the estimation is based on events that exceed a high threshold (*peaks over threshold* approach). Both approaches are used in climate research.

Hence, the present chapter does not directly consider the dimensions of vulnerability or exposure, which are critical in determining the human and ecosystem impacts of climate extremes (Chapters 1, 2, and 4).

This report defines an 'extreme climate or weather event' or 'climate extreme' as "the occurrence of a value of a weather or climate variable above (or below) a threshold value near the upper (or lower) ends of the range of observed values of the variable" (see Glossary). Several aspects of this definition can be clarified thus:

- Definitions of thresholds vary, but values with less than 10, 5, 1%, or even lower chance of occurrence for a given time of the year (day, month, season, whole year) during a specified *reference* period (generally 1961-1990) are often used. In some circumstances, information from sources other than observations, such as model projections, can be used as a reference.

- Absolute thresholds (rather than these relative thresholds based on the range of observed values of a variable) can also be used to identify extreme events (e.g., specific critical temperatures for health impacts).

- What is called an extreme weather or climate event will vary from place to place in an absolute sense (e.g., a hot day in the tropics will be a different temperature than a hot day in the mid-latitudes), and possibly in time given some adaptation from society (see Box 3-1).

- Some climate extremes (e.g., droughts, floods) may be the result of an accumulation of moderate weather or climate events (this accumulation being itself extreme). Compound events (see Section 3.1.3), that is, two or more events occurring simultaneously, can lead to high impacts, even if the two single events are not extreme per se (only their combination).

Recent publications have used other approaches for evaluating characteristics of extremes or changes in extremes, for instance, analyzing trends in record events or investigating whether records in observed time series are being set more or less frequently than would be expected in an unperturbed climate (Benestad, 2003, 2006; Zorita et al., 2008; Meehl et al., 2009c; Trewin and Vermont, 2010). Furthermore, besides the actual magnitude of extremes (quantified in terms of probability/return frequency or absolute threshold), other relevant aspects for the definition of climate extremes from an impact perspective include the event's duration, the spatial area affected, timing, frequency, onset date, continuity (i.e., whether there are 'breaks' within a spell), and preconditioning (e.g., rapid transition from a slowly developing meteorological drought into an agricultural drought, see Box 3-3). These aspects, together with seasonal variations in climate extremes, are not as frequently examined in climate models or observational analyses, and thus can only be partly assessed within this chapter.

As noted in the discussion of 'extreme weather or climate events' in Section 3.1.2, thresholds, percentiles, or return values used for the definition of climate extremes are generally defined with respect to a given reference period (generally historical, i.e., 1961-1990, but possibly also based on climate model data). In some cases, a transient baseline can also be considered (i.e., the baseline uses data from the period under examination and changes as the period being considered changes, rather than using a standard period such as 1961-1990). The choice of the reference period may be relevant for the magnitude of the assessed changes as highlighted, for example, in Lorenz et al. (2010). The choice of the reference period (static or transient) could also affect the assessment of the respective role of changes in mean versus changes in variability for changes in extremes discussed in Section 3.1.6. If extremes are based on the probability distribution from which they are drawn, then a simple change in the mean (and keeping the same distribution) would, strictly speaking, produce no relative change in extremes at all. The question of the choice of an appropriate reference period is tied to the notion of adaptation. Events that are considered extreme nowadays in some regions could possibly be adapted to if the vulnerability and exposure to these extremes is reduced (Chapters 1, 2, and 4 through 7). However, there are also some limits to adaptation as highlighted in Chapter 8. These considerations are difficult to include in the statistical analyses of climate scenarios because of the number of (mostly non-physical) aspects that would need to be taken into account.

To conclude, there is no precise definition of an extreme (e.g., D.B. Stephenson et al., 2008). In particular, we note limitations in the definition of both probability-based or threshold-based climate extremes and their relations to impacts, which apply independently of the chosen method of analysis:
- An event from the extreme tails of probability distributions is not necessarily extreme in terms of impact.
- Impact-related thresholds can vary in space and time, that is, single absolute thresholds (e.g., a daily rainfall exceeding 25 mm or the number of frost days) will not reflect extremes in all locations and time periods (e.g., season, decade).

As an illustration, projected patterns (in the magnitude but not the sign) of changes in annual heat wave length were shown to be highly dependent on the choice of index used for the assessment of heat wave or warm spell duration (using the mean and maximum Heat Wave Duration Indices, HWDImean and HWDImax, and the Warm Spell Duration Index, WSDI; see Orlowsky and Seneviratne, 2011), because of large geographical variations in the variability of daily temperature (Alexander et al., 2006). Similar definition issues apply to other types of extremes, especially those characterizing dryness (see Section 3.5.1 and Box 3-3).

- Not all extreme weather and climate events necessarily have extreme impacts.
- The distinction between extreme weather events and extreme climate events is not precise, but is related to their specific time scales:
  – An extreme weather event is typically associated with changing weather patterns, that is, within time frames of less than a day to a few weeks.
  – An extreme climate event happens on longer time scales. It can be the accumulation of several (extreme or non-extreme) weather events (e.g., the accumulation of moderately below-average rainy days over a season leading to substantially below-average cumulated rainfall and drought conditions).

For simplicity, we collectively refer to both extreme weather events and extreme climate events with the term 'climate extremes' in this chapter.

From this definition, it can be seen that climate extremes can be defined quantitatively in two ways:
1) Related to their probability of occurrence
2) Related to a specific (possibly impact-related) threshold.

The first type of definition can either be expressed with respect to given percentiles of the distribution functions of the variables, or with respect to specific return frequencies (e.g., '100-year event'). Compound events can be viewed as a special category of climate extremes, which result from the combination of two or more events, and which are again 'extreme' either from a statistical perspective (tails of distribution functions of climate variables) or associated with a specific threshold (Section 3.1.3.). These two definitions of climate extremes, probability-based or threshold-based, are not necessarily antithetic. Indeed, hazards for society and ecosystems are often extreme both from a probability and

threshold perspective (e.g., a 40°C threshold for midday temperature in the mid-latitudes).

In the scientific literature, several aspects are considered in the definition and analysis of climate extremes (Box 3-1).

3.1.3. Compound (Multiple) Events

In climate science, compound events can be (1) two or more extreme events occurring simultaneously or successively, (2) combinations of extreme events with underlying conditions that amplify the impact of the events, or (3) combinations of events that are not themselves extremes but lead to an extreme event or impact when combined. The contributing events can be of similar (clustered multiple events) or different type(s). There are several varieties of clustered multiple events, such as tropical cyclones generated a few days apart with the same path and/or intensities, which may occur if there is a tendency for persistence in atmospheric circulation and genesis conditions. Examples of compound events resulting from events of different types are varied – for instance, high sea level coinciding with tropical cyclone landfall (Section 3.4.4), or cold and dry conditions (e.g., the Mongolian Dzud, see Case Study 9.2.4), or the impact of hot events and droughts on wildfire (Case Study 9.2.2), or a combined risk of flooding from sea level surges and precipitation-induced high river discharge (Svensson and Jones, 2002; Van den Brink et al., 2005). Compound events can even result from 'contrasting extremes', for example, the projected occurrence of both droughts and heavy precipitation events in future climate in some regions (Table 3-3).

Impacts on the physical environment (Section 3.5) are often the result of compound events. For instance, floods will more likely occur over saturated soils (Section 3.5.2), which means that both soil moisture status and precipitation intensity play a role. The wet soil may itself be the result of a number of above-average but not necessarily extreme precipitation events, or of enhanced snow melt associated with temperature anomalies in a given season. Similarly, droughts are the result of pre-existing soil moisture deficits and of the accumulation of precipitation deficits and/or evapotranspiration excesses (Box 3-3), not all (or none) of which are necessarily extreme for a particular drought event when considered in isolation. Also, impacts on human systems or ecosystems (Chapter 4) can be the results of compound events, for example, in the case of health-related impacts associated with combined temperature and humidity conditions (Box 3-1).

Although compound events can involve causally unrelated events, the following causes may lead to a correlation between the occurrence of extremes (or their impacts):
1) A common external forcing factor for changing the probability of the two events (e.g., regional warming, change in frequency or intensity of El Niño events)
2) Mutual reinforcement of one event by the other and vice versa due to system feedbacks (Section 3.1.4)

3) Conditional dependence of the occurrence or impact of one event on the occurrence of another event (e.g., extreme soil moisture levels and precipitation conditions for floods, droughts, see above).

Changes in one or more of these factors would be required for a changing climate to induce changes in the occurrence of compound events. Unfortunately, investigation of possible changes in these factors has received little attention. Also, much of the analysis of changes of extremes has, up to now, focused on individual extremes of a single variable. However, recent literature in climate research is starting to consider compound events and explore appropriate methods for their analysis (e.g., Coles, 2001; Beirlant et al., 2004; Benestad and Haugen, 2007; Renard and Lang, 2007; Schölzel and Friederichs, 2008; Beniston, 2009b; Tebaldi and Sanso, 2009; Durante and Salvadori, 2010).

3.1.4. Feedbacks

A special case of compound events is related to the presence of feedbacks within the climate system, that is, mutual interaction between several climate processes, which can either lead to a damping (negative feedback) or enhancement (positive feedback) of the initial response to a given forcing (see also 'climate feedback' in the Glossary). Feedbacks can play an important role in the development of extreme events, and in some cases two (or more) climate extremes can mutually strengthen one another. One example of positive feedback between two extremes is the possible mutual enhancement of droughts and heat waves in transitional regions between dry and wet climates. This feedback has been identified as having an influence on projected changes in temperature variability and heat wave occurrence in Central and Eastern Europe and the Mediterranean (Seneviratne et al., 2006a; Diffenbaugh et al., 2007), and possibly also in Britain, Eastern North America, the Amazon, and East Asia (Brabson et al., 2005; Clark et al., 2006). Further results also suggest that it is a relevant factor for past heat waves and temperature extremes in Europe and the United States (Durre et al., 2000; Fischer et al., 2007a,b; Hirschi et al., 2011). Two main mechanisms that have been suggested to underlie this feedback are: (1) enhanced soil drying during heat waves due to increased evapotranspiration (as a consequence of higher vapor pressure deficit and higher incoming radiation); and (2) higher relative heating of the air from sensible heat flux when soil moisture deficit starts limiting evapotranspiration/latent heat flux (e.g., Seneviratne et al., 2010). Additionally, there may also be indirect and/or non-local effects of dryness on heat waves through, for example, changes in circulation patterns or dry air advection (e.g., Fischer et al., 2007a; Vautard et al., 2007; Haarsma et al., 2009). However, the strength of these feedbacks is still uncertain in current climate models (e.g., Clark et al., 2010), in particular if additional feedbacks with precipitation (e.g., Koster et al., 2004b; Seneviratne et al., 2010) and with land use and land cover state and changes (e.g., Lobell et al., 2008; Pitman et al., 2009; Teuling et al., 2010) are considered. Also, feedbacks between trends in snow cover and changes in temperature extremes have been highlighted as being relevant for projections (e.g., Kharin et al., 2007; Orlowsky and Seneviratne, 2011). Feedbacks with soil moisture

Table 3-1 | Overview of considered extremes and summary of observed and projected changes at a global scale. Regional details on observed and projected changes in temperature and precipitation extremes are provided in Tables 3-2 and 3-3. Extremes (e.g., cold/warm days/nights, heat waves, heavy precipitation events) are defined with respect to late 20th-century climate (see also Box 3-1 for discussion of reference period).

		Observed Changes (since 1950)	Attribution of Observed Changes	Projected Changes (up to 2100) with Respect to Late 20th Century
Weather and Climate Variables	**Temperature** (Section 3.3.1)	*Very likely* decrease in number of unusually cold days and nights at the global scale. *Very likely* increase in number of unusually warm days and nights at the global scale. *Medium confidence* in increase in length or number of warm spells or heat waves in many (but not all) regions. *Low or medium confidence* in trends in temperature extremes in some subregions due either to lack of observations or varying signal within subregions. [Regional details in Table 3-2]	*Likely* anthropogenic influence on trends in warm/cold days/nights at the global scale. No attribution of trends at a regional scale with a few exceptions.	*Virtually certain* decrease in frequency and magnitude of unusually cold days and nights at the global scale. *Virtually certain* increase in frequency and magnitude of unusually warm days and nights at the global scale. *Very likely* increase in length, frequency, and/or intensity of warm spells or heat waves over most land areas. [Regional details in Table 3-3]
	Precipitation (Section 3.3.2)	*Likely* statistically significant increases in the number of heavy precipitation events (e.g., 95th percentile) in more regions than those with statistically significant decreases, but strong regional and subregional variations in the trends. [Regional details in Table 3-2]	*Medium confidence* that anthropogenic influences have contributed to intensification of extreme precipitation at the global scale.	*Likely* increase in frequency of heavy precipitation events or increase in proportion of total rainfall from heavy falls over many areas of the globe, in particular in the high latitudes and tropical regions, and in winter in the northern mid-latitudes. [Regional details in Table 3-3]
	Winds (Section 3.3.3)	*Low confidence* in trends due to insufficient evidence.	*Low confidence* in the causes of trends due to insufficient evidence.	*Low confidence* in projections of extreme winds (with the exception of wind extremes associated with tropical cyclones).
Phenomena Related to Weather and Climate Extremes	**Monsoons** (Section 3.4.1)	*Low confidence* in trends because of insufficient evidence.	*Low confidence* due to insufficient evidence.	*Low confidence* in projected changes in monsoons, because of insufficient agreement between climate models.
	El Niño and other Modes of Variability (Sections 3.4.2 and 3.4.3)	*Medium confidence* in past trends toward more frequent central equatorial Pacific El Niño-Southern Oscillation (ENSO) events. Insufficient evidence for more specific statements on ENSO trends. *Likely* trends in Southern Annular Mode (SAM).	*Likely* anthropogenic influence on identified trends in SAM.[1] Anthropogenic influence on trends in North Atlantic Oscillation (NAO) are *about as likely as not.* No attribution of changes in ENSO.	*Low confidence* in projections of changes in behavior of ENSO and other modes of variability because of insufficient agreement of model projections.
	Tropical Cyclones (Section 3.4.4)	*Low confidence* that any observed long-term (i.e., 40 years or more) increases in tropical cyclone activity are robust, after accounting for past changes in observing capabilities.	*Low confidence* in attribution of any detectable changes in tropical cyclone activity to anthropogenic influences (due to uncertainties in historical tropical cyclones record, incomplete understanding of physical mechanisms, and degree of tropical cyclone variability).	*Likely* decrease or no change in frequency of tropical cyclones. *Likely* increase in mean maximum wind speed, but possibly not in all basins. *Likely* increase in heavy rainfall associated with tropical cyclones.
	Extratropical Cyclones (Section 3.4.5)	*Likely* poleward shift in extratropical cyclones. *Low confidence* in regional changes in intensity.	*Medium confidence* in an anthropogenic influence on poleward shift.	*Likely* impacts on regional cyclone activity but *low confidence* in detailed regional projections due to only partial representation of relevant processes in current models. *Medium confidence* in a reduction in the numbers of mid-latitude storms. *Medium confidence* in projected poleward shift of mid-latitude storm tracks.
Impacts on Physical Environment	**Droughts** (Section 3.5.1)	*Medium confidence* that some regions of the world have experienced more intense and longer droughts, in particular in southern Europe and West Africa, but opposite trends also exist. [Regional details in Table 3-2]	*Medium confidence* that anthropogenic influence has contributed to some observed changes in drought patterns. *Low confidence* in attribution of changes in drought at the level of single regions due to inconsistent or insufficient evidence.	*Medium confidence* in projected increase in duration and intensity of droughts in some regions of the world, including southern Europe and the Mediterranean region, central Europe, central North America, Central America and Mexico, northeast Brazil, and southern Africa. Overall *low confidence* elsewhere because of insufficient agreement of projections. [Regional details in Table 3-3]
	Floods (Section 3.5.2)	Limited to medium evidence available to assess climate-driven observed changes in the magnitude and frequency of floods at regional scale. Furthermore, there is low agreement in this evidence, and thus overall *low confidence* at the global scale regarding even the sign of these changes. *High confidence* in trend toward earlier occurrence of spring peak river flows in snowmelt- and glacier-fed rivers.	*Low confidence* that anthropogenic warming has affected the magnitude or frequency of floods at a global scale. *Medium confidence* to *high confidence* in anthropogenic influence on changes in some components of the water cycle (precipitation, snowmelt) affecting floods.	*Low confidence* in global projections of changes in flood magnitude and frequency because of insufficient evidence. *Medium confidence* (based on physical reasoning) that projected increases in heavy precipitation would contribute to rain-generated local flooding in some catchments or regions. *Very likely* earlier spring peak flows in snowmelt- and glacier-fed rivers.

Table 3-1 (continued)

		Observed Changes (since 1950)	Attribution of Observed Changes	Projected Changes (up to 2100) with Respect to Late 20th Century
Impacts on Physical Environment (Continued)	Extreme Sea Level and Coastal Impacts (Sections 3.5.3, 3.5.4, and 3.5.5)	*Likely* increase in extreme coastal high water worldwide related to increases in mean sea level in the late 20th century.	*Likely* anthropogenic influence via mean sea level contributions.	*Very likely* that mean sea level rise will contribute to upward trends in extreme coastal high water levels. *High confidence* that locations currently experiencing coastal erosion and inundation will continue to do so due to increasing sea level, in the absence of changes in other contributing factors.
	Other Physical Impacts (Sections 3.5.6, 3.5.7, and 3.5.8)	*Low confidence* in global trends in large landslides in some regions. *Likely* increased thawing of permafrost with *likely* resultant physical impacts.	*Likely* anthropogenic influence on thawing of permafrost. *Low confidence* of other anthropogenic influences because of insufficient evidence for trends in other physical impacts in cold regions.	*High confidence* that changes in heat waves, glacial retreat, and/or permafrost degradation will affect high mountain phenomena such as slope instabilities, mass movements, and glacial lake outburst floods. *High confidence* that changes in heavy precipitation will affect landslides in some regions. *Low confidence* in projected future changes in dust activity.

Notes: 1. Due to trends in stratospheric ozone concentrations.

and snow affect extremes in specific regions (hot extremes in transitional climate regions, and cold extremes in snow-covered regions), where they may induce significant deviations in changes in extremes versus changes in the average climate, as also discussed in Section 3.1.6. Other relevant feedbacks involving extreme events are those that can lead to impacts on the global climate, such as modification of land carbon uptake due to enhanced drought occurrence (e.g., Ciais et al., 2005; Friedlingstein et al., 2006; Reichstein et al., 2007) or carbon release due to permafrost degradation (see Section 3.5.7). These aspects are not, however, specifically considered in this chapter (but see Section 3.1.7, on projections of possible increased Amazon drought and forest dieback in this region). Chapter 4 also addresses feedback loops between droughts, fire, and climate change (Section 4.2.2.1).

3.1.5. Confidence and Likelihood of Assessed Changes in Extremes

In this chapter, all assessments regarding past or projected changes in extremes are expressed following the new IPCC Fifth Assessment Report uncertainty guidance (Mastrandrea et al., 2010). The new uncertainty guidance makes a clearer distinction between confidence and likelihood (see Box SPM.2). Its use complicates comparisons between assessments in this chapter and those in the IPCC Fourth Assessment Report (AR4), as they are not directly equivalent in terms of nomenclature. The following procedure was adopted in this chapter (see in particular the Executive Summary and Tables 3-1, 3-2, and 3-3.):

- For each assessment, the *confidence* level for the given assessment is first assessed (*low*, *medium*, or *high*), as discussed in the next paragraph.
- For assessments with *high confidence*, likelihood assessments of a direction of change are also provided (*virtually certain* for 99-100%, *very likely* for 90-100%, *likely* for 66-100%, *more likely than not* for 50-100%, *about as likely as not* for 33-66%, *unlikely* for 0-33%, *very unlikely* for 0-10%, and *exceptionally unlikely* for 0-1%). In a few cases for which there is *high confidence* (e.g., based on physical understanding) but for which there are not sufficient

model projections to provide a more detailed likelihood assessment (such as '*likely*'), only the confidence assessment is provided.
- For assessments with *medium confidence*, a direction of change is provided, but without an assessment of likelihood.
- For assessments with *low confidence*, no direction of change is generally provided.

The confidence assessments are expert-based evaluations that consider the confidence in the tools and data basis (models, data, proxies) used to assess or project changes in a specific element, and the associated level of understanding. Examples of cases of *low confidence* for model projections are if models display poor performance in simulating the specific extreme in the present climate (see also Box 3-2), or if insufficient literature on model performance is available for the specific extreme, for example, due to lack of observations. Similarly for observed changes, the assessment may be of *low confidence* if the available evidence is based only on scattered data (or publications) that are insufficient to provide a robust assessment for a large region, or the observations may be of poor quality, not homogeneous, or only of an indirect nature (proxies). In cases with *low confidence* regarding past or projected changes in some extremes, we indicate whether the *low confidence* is due to lack of literature, lack of evidence (data, observations), or lack of understanding. It should be noted that there are some overlaps between these categories, as for instance a lack of evidence can be at the root of a lack of literature and understanding. Cases of changes in extremes for which confidence in the models and data is rated as '*medium*' are those where we have some confidence in the tools and evidence available to us, but there remain substantial doubts about some aspects of the quality of these tools. It should be noted that an assessment of *low* confidence in observed or projected changes or trends in a specific extreme neither implies nor excludes the possibility of changes in this extreme. Rather the assessment indicates *low confidence* in the ability to detect or project any such changes.

Changes (observed or projected) in some extremes are easier to assess than in others either due to the complexity of the underlying processes or to the amount of evidence available for their understanding. This

results in differing levels of uncertainty in climate simulations and projections for different extremes (Box 3-2). Because of these issues, projections in some extremes are difficult or even impossible to provide, although projections in some other extremes have a high level of confidence. In addition, uncertainty in projections also varies over different time frames for individual extremes, because of varying contributions over time of internal climate variability, model uncertainty, and emission scenario uncertainty to the overall uncertainty (Box 3-2 and Section 3.2). Overall, we can infer that our confidence in past and future changes in extremes varies with the type of extreme, the data available, and the region, season, and time frame being considered, linked with the level of understanding and reliability of simulation of the underlying physical processes. These various aspects are addressed in more detail in Box 3-2, Section 3.2, and the subsections on specific extremes in Sections 3.3-3.5.

3.1.6.   Changes in Extremes and Their Relationship to Changes in Regional and Global Mean Climate

Changes in extremes can be linked to changes in the mean, variance, or shape of probability distributions, or all of these (see, e.g., Figure 1-2). Thus a change in the frequency of occurrence of hot days (i.e., days above a certain threshold) can arise from a change in the mean daily maximum temperature, and/or from a change in the variance and/or shape of the frequency distribution of daily maximum temperatures. If changes in the frequency of occurrence of hot days were mainly linked to changes in the mean daily maximum temperature, and changes in the shape and variability of the distribution of daily maximum temperatures were of secondary importance, then it might be reasonable to use projected changes in mean temperature to estimate how changes in extreme temperatures might change in the future. If, however, changes in the shape and variability of the frequency distribution of daily maximum temperature were important, such naive extrapolation would be less appropriate or possibly even misleading (e.g., Ballester et al., 2010). The results of both empirical and model studies indicate that although in several situations changes in extremes do scale closely with changes in the mean (e.g., Griffiths et al., 2005), there are sufficient exceptions from this that changes in the variability and shape of probability distributions of weather and climate variables need to be considered as well as changes in means, if we are to project future changes in extremes (e.g., Hegerl et al., 2004; Schär et al., 2004; Caesar et al., 2006; Clark et al., 2006; Della-Marta et al., 2007a; Kharin et al., 2007; Brown et al., 2008; Ballester et al., 2010; Orlowsky and Seneviratne, 2011). This appears to be especially the case for short-duration precipitation, and for temperatures in mid- and high latitudes (but not all locations in these regions). In mid- and high latitudes stronger increases (or decreases) in some extremes are generally associated with feedbacks with soil moisture or snow cover (Section 3.1.4). Note that the respective importance of changes in mean versus changes in variability also depends on the choice of the reference period used to define the extremes (Box 3-1).

An additional relevant question is the extent to which regional changes in extremes scale with changes in global mean climate. Indeed, recent publications and the public debate have focused, for example, on global mean temperature targets (e.g., Allen et al., 2009; Meinshausen et al., 2009), however, the exact implications of these mean global changes (e.g., '2°C target') for regional extremes have not been widely assessed (e.g., Clark et al., 2010). Orlowsky and Seneviratne (2011) investigated the scaling between projected changes in the 10th and 90th percentile of Tmax on annual and seasonal (June-July-August: JJA, December-January-February: DJF) time scales with globally averaged annual mean changes in Tmax based on the whole CMIP3 ensemble (see Section 3.2.3 for discussion of the CMIP3 ensemble). The results highlight particularly large projected changes in the 10th percentile Tmax in the northern high-latitude regions in winter and the 90th percentile Tmax in Southern Europe in summer with scaling factors of about 2 in both cases (i.e., increases of about 4°C for a mean global increase of 2°C). However, in some regions and seasons, the scaling can also be below 1 (e.g., changes in 10th percentile in JJA in the high latitudes). This is also illustrated in Figure 3-5a, which compares analyses of changes in return values of annual extremes of maximum daily temperatures for the overall land and specific regions, and shows high region-to-region variability in these changes. The changes in return values at the global scale ('Globe (Land only)') for their part are almost identical to the changes in global mean daily maximum temperature, suggesting that the scaling issues are related to regional effects rather than overall differences in the changes in the tails versus the means of the distributions of daily maximum temperature. The situation is very different for precipitation (Figure 3-7a), with clearly distinct behavior between changes in mean and extreme precipitation at the global scale, highlighting the dependency of any scaling on the variable being considered. The lack of consistent scaling between regional and seasonal changes in extremes and changes in means has also been highlighted in empirical studies (e.g., Caesar et al., 2006). It should further be noted that not only do regional extremes not necessarily scale with global mean changes, but also mean global warming does not exclude the possibility of cooling in some regions and seasons, both in the recent past and in the coming decades: it has for instance been recently suggested that the decrease in sea ice caused by the mean warming could induce, although not systematically, more frequent cold winter extremes over northern continents (Petoukhov and Semenov, 2010). Also parts of central North America and the eastern United States present cooling trends in mean temperature and some temperature extremes in the spring to summer season in recent decades (Section 3.3.1). It should be noted that, independently of scaling issues for the means and extremes of the same variable, some extremes can be related to mean climate changes in other variables, such as links between mean global changes in relative humidity and some regional changes in heavy precipitation events (Sections 3.2.2.1 and 3.3.2).

Global-scale trends in a specific extreme may be either more reliable or less reliable than some regional-scale trends, depending on the geographical uniformity of the trends in the specific extreme. In particular, climate projections for some variables are not consistent, even in the sign of the projected change, everywhere across the globe (e.g., Christensen et al., 2007; Meehl et al., 2007b). For instance, projections typically include some regions with a tendency toward wetter conditions and others with

a tendency toward drier conditions, with some regions displaying a shift in climate regimes (e.g., from humid to transitional or transitional to dry). Some of these regional changes will depend on how forcing changes may alter the regional atmospheric circulation, especially in coastal regions and regions with substantial orography. Hence for certain extremes such as floods and droughts, regional projections might indicate larger changes than is the case for projections of global averages (which would average the regional signals exhibiting changes of opposite signs). This also means that signals at the regional scale may be more reliable (and meaningful) in some cases than assessments at the global scale. On the other hand, temperature extremes projections, which are consistent across most regions, are thus more reliable at the global scale ('*virtually certain*') than at the regional scale (at most '*very likely*').

3.1.7.   Surprises / Abrupt Climate Change

This report focuses on the most probable changes in extremes based on current knowledge. However, the possible future occurrence of low-probability, high-impact scenarios associated with the crossing of poorly understood climate thresholds cannot be excluded, given the transient and complex nature of the climate system. Such scenarios have important implications for society as highlighted in Section 8.5.1. So, an assessment that we have *low confidence* in projections of a specific extreme, or even lack of consideration of given climate changes under the categories covered in this chapter (e.g., shutdown of the meridional overturning circulation), should not be interpreted as meaning that no change is expected in this extreme or climate element (see also Section 3.1.5). Feedbacks play an important role in either damping or enhancing extremes in several climate variables (Section 3.1.4), and this can also lead to 'surprises,' that is, changes in extremes greater (or less) than might be expected with a gradual warming of the climate system. Similarly, as discussed in 3.1.3, contrasting or multiple extremes can occur but our understanding of these is insufficient to provide credible comprehensive projections of risks associated with such combinations.

One aspect that we do not address in this chapter is the existence of possible tipping points in the climate system (e.g., Meehl et al., 2007b; Lenton et al., 2008; Scheffer et al., 2009), that is, the risks of abrupt, possibly irreversible changes in the climate system. Abrupt climate change is defined as follows in the Glossary: "The nonlinearity of the climate system may lead to abrupt climate change, sometimes called rapid climate change, abrupt events, or even surprises. The term abrupt often refers to time scales faster than the typical time scale of the responsible forcing. However, not all abrupt climate changes need be externally forced. Some changes may be truly unexpected, resulting from a strong, rapidly changing forcing of a nonlinear system." Thresholds associated with tipping points may be termed 'critical thresholds,' or, in the case of the climate system, 'climate thresholds'. Scheffer et al. (2009) illustrate the possible equilibrium responses of a system to forcing. In the case of a linear response, only a large forcing can lead to a major state change in the system. However, in the presence of a critical threshold even a small change in forcing can lead to a similar

major change in the system. For systems with critical bifurcations in the equilibrium state function two alternative stable conditions may exist, whereby an induced change may be irreversible. Such critical transitions within the climate system represent typical low-probability, high-impact scenarios, which were also noted in the AR4 (Meehl et al., 2007b). Lenton et al. (2008) provided a recent review on potential tipping elements within the climate system, that is, subsystems of the Earth system that are at least subcontinental in scale and which may entail a tipping point. Some of these would be especially relevant to certain extremes [e.g., El Niño-Southern Oscillation (ENSO), the Indian summer monsoon, and the Sahara/Sahel and West African monsoon for drought and heavy precipitation, and the Greenland and West Antarctic ice sheets for sea level extremes], or are induced by changes in extremes (e.g., Amazon rainforest die-back induced by drought). For some of the identified tipping elements, the existence of bistability has been suggested by paleoclimate records, but is still debated in some cases (e.g., Brovkin et al., 2009). There is often a lack of agreement between models regarding these low-probability, high-impact scenarios, for instance, regarding a possible increased drought and consequent die-back of the Amazon rainforest (e.g., Friedlingstein et al., 2006; Poulter et al., 2010; see Table 3-3 for dryness projections in this region), the risk of an actual shutdown of the Atlantic thermohaline circulation (e.g., Rahmstorf et al., 2005; Lenton et al., 2008), or the potential irreversibility of the decrease in Arctic sea ice (Tietsche et al., 2011). For this reason, *confidence* in these scenarios is assessed as *low*.

3.2.   Requirements and Methods for Analyzing Changes in Extremes

3.2.1.   Observed Changes

Sections 3.3 to 3.5 of this chapter provide assessments of the literature regarding changes in extremes in the observed record published mainly since the AR4 and building on the AR4 assessment. Summaries of these assessments are provided in Table 3-1. Overviews of observed regional changes in temperature and precipitation extremes are provided in Table 3-2. In this section issues are discussed related to the data and observations used to examine observed changes in extremes.

Issues with data availability are especially critical when examining changes in extremes of given climate variables (Nicholls, 1995). Indeed, the more rare the event, the more difficult it is to identify long-term changes, simply because there are fewer cases to evaluate (Frei and Schär, 2001; Klein Tank and Können, 2003). Identification of changes in extremes is also dependent on the analysis technique employed (X. Zhang et al., 2004; Trömel and Schönwiese, 2005). Another important criterion constraining data availability for the analysis of extremes is the respective time scale on which they occur (Section 3.1.2), since this determines the required temporal resolution for their assessment (e.g., heavy hourly or daily precipitation versus multi-year drought). Longer time resolution data (e.g., monthly, seasonal, and annual values) for temperature and precipitation are available for most parts of the world

starting late in the 19th to early 20th century, and allow analysis of meteorological drought (see Box 3-3) and unusually wet periods of the order of a month or longer. To examine changes in extremes occurring on short time scales, particularly of climate variables such as temperature and precipitation (or wind), normally requires the use of high-temporal resolution data, such as daily or sub-daily observations, which are generally either not available, or available only since the middle of the 20th century and in many regions only from as recently as 1970. Even where sufficient data are available, several problems can still limit their analysis. First, although the situation is changing (especially for the situation with respect to 'extreme indices,' Box 3-1), many countries still do not freely distribute their higher temporal resolution data. Second, there can be issues with the quality of measurements. A third important issue is climate data homogeneity (see below). These and other issues are discussed in detail in the AR4 (Trenberth et al., 2007). For instance, the temperature and precipitation stations considered in the daily data set used in Alexander et al. (2006) are not globally uniform. Although observations for most parts of the globe are available, measurements are lacking in Northern South America, Africa, and part of Australia. The other data set commonly used for extremes analyses is from Caesar et al. (2006; used, e.g., in Brown et al., 2008), which also has data gaps in most of South America, Africa, Eastern Europe, Mexico, the Middle East, India, and Southeast Asia. Also the study by Vose et al. (2005) has data gaps in South America, Africa, and India. It should be further noted that the regions with data coverage do not all have the same density of stations (Alexander et al., 2006; Caesar et al., 2006). While some studies are available on a country or regional basis for areas not covered in global studies (see, e.g., Tables 3-2 and 3-3), lack of data in many parts of the globe leads to limitations in our ability to assess observed changes in climate extremes for many regions.

Whether or not climate data are homogeneous is of clear relevance for an analysis of extremes, especially at smaller spatial scales. Data are defined as homogeneous when the variations and trends in a climate time series are due solely to variability and changes in the climate system. Some meteorological elements are especially vulnerable to uncertainties caused by even small changes in the exposure of the measuring equipment. For instance, erection of a small building or changes in vegetative cover near
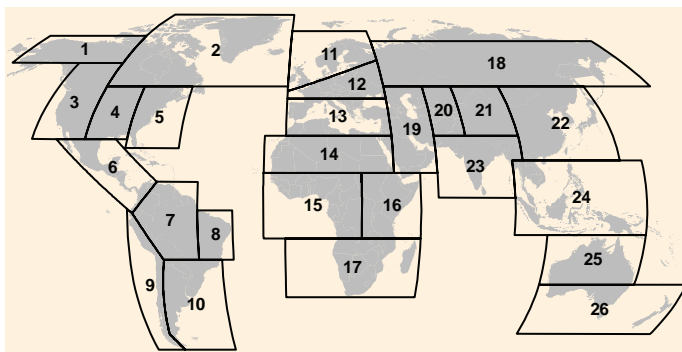
Figure 3-1 | Definitions of regions used in Tables 3-2 and 3-3, and Figures 3-5 and 3-7. Exact coordinates of the regions are provided in the on-line supplement, Appendix 3.A. Assessments and analyses are provided for land areas only.

the measuring equipment can produce a bias in wind measurements (Wan et al., 2010). When a change occurs it can result in either a discontinuity in the time series (step change) or a more gradual change that can manifest itself as a false trend (Menne and Williams Jr., 2009), both of which can impact on whether a particular observation exceeds a threshold. Homogeneity detection and data adjustments have been implemented for longer averaging periods (e.g., monthly, seasonal, annual); however, techniques applicable to shorter observing periods (e.g., daily) data have only recently been developed (e.g., Vincent et al., 2002; Della-Marta and Wanner, 2006), and have not been widely implemented. Homogeneity issues also affect the monitoring of other meteorological and climate variables, for which further and more severe limitations also can exist. This is in particular the case regarding measurements of wind and relative humidity, and data required for the analysis of weather and climate phenomena (tornadoes, extratropical and tropical cyclones; Sections 3.3.3, 3.4.4, and 3.4.5), as well as impacts on the physical environment (e.g., droughts, floods, cryosphere impacts; Section 3.5).

Thunderstorms, tornadoes, and related phenomena are not well observed in many parts of the world. Tornado occurrence since 1950 in the United States, for instance, displays an increasing trend that mainly reflects increased population density and increased numbers of people in remote areas (Trenberth et al., 2007; Kunkel et al., 2008). Such trends increase the likelihood that a tornado would be observed. A similar problem occurs with thunderstorms. Changes in reporting practices, increased population density, and even changes in the ambient noise level at an observing station all have led to inconsistencies in the observed record of thunderstorms.

Studies examining changes in extratropical cyclones, which focus on changes in storm track location, intensities, and frequency, are limited in time due to a lack of suitable data prior to about 1950. Most of these studies have relied on model-based reanalyses that also incorporate observations into a hybrid model-observational data set. However, reanalyses can have homogeneity problems due to changes in the amount and type of data being assimilated, such as the introduction of satellite data in the late 1970s and other observing system changes (Trenberth et al., 2001; Bengtsson et al., 2004). Recent reanalysis efforts have attempted to produce more homogeneous reanalyses that show promise for examining changes in extratropical cyclones and other climate features (Compo et al., 2006). Results, however, are strongly dependent on the reanalysis and cyclone tracking techniques used (Ulbrich et al., 2009).

The robustness of analyses of observed changes in tropical cyclones has been hampered by a number of issues with the historical record. One of the major issues is the heterogeneity introduced by changing technology and reporting protocols within the responsible agencies (e.g., Landsea et al., 2004). Further heterogeneity is introduced when records from multiple ocean basins are combined to explore global trends, because data quality and reporting protocols vary substantially between agencies (Knapp and Kruk, 2010). Much like other weather and climate observations,

FAQ 3.1 | Is the Climate Becoming More Extreme?

While there is evidence that increases in greenhouse gases have likely caused changes in some types of extremes, there is no simple answer to the question of whether the climate, in general, has become more or less extreme. Both the terms 'more extreme' and 'less extreme' can be defined in different ways, resulting in different characterizations of observed changes in extremes. Additionally, from a physical climate science perspective it is difficult to devise a comprehensive metric that encompasses all aspects of extreme behavior in the climate.

One approach for evaluating whether the climate is becoming more extreme would be to determine whether there have been changes in the typical range of variation of specific climate variables. For example, if there was evidence that temperature variations in a given region had become significantly larger than in the past, then it would be reasonable to conclude that temperatures in that region had become more extreme. More simply, temperature variations might be considered to be becoming more extreme if the difference between the highest and the lowest temperature observed in a year is increasing. According to this approach, daily temperature over the globe may have become less extreme because there have generally been greater increases in mean daily minimum temperatures globally than in mean daily maximum temperatures, over the second half of the 20th century. On the other hand, one might conclude that daily precipitation has become more extreme because observations suggest that the magnitude of the heaviest precipitation events has increased in many parts of the world. Another approach would be to ask whether there have been significant changes in the frequency with which climate variables cross fixed thresholds that have been associated with human or other impacts. For example, an increase in the mean temperature usually results in an increase in hot extremes and a decrease in cold extremes. Such a shift in the temperature distribution would not increase the 'extremeness' of day-to-day variations in temperature, but would be perceived as resulting in a more extreme warm temperature climate, and a less extreme cold temperature climate. So the answer to the question posed here would depend on the variable of interest, and on which specific measure of the extremeness of that variable is examined. As well, to provide a complete answer to the above question, one would also have to collate not just trends in single variables, but also indicators of change in complex extreme events resulting from a sequence of individual events, or the simultaneous occurrence of different types of extremes. So it would be difficult to comprehensively describe the full suite of phenomena of concern, or to find a way to synthesize all such indicators into a single extremeness metric that could be used to comprehensively assess whether the climate as a whole has become more extreme from a physical perspective. And to make such a metric useful to more than a specific location, one would have to combine the results at many locations, each with a different perspective on what is 'extreme.'

tropical cyclone observations are taken to support short-term forecasting needs. Improvements in observing techniques are often implemented without any overlap or calibration against existing methods to document the impact of the changes on the climate record. Additionally, advances in technology have enabled better and more complete observations. For example, the introduction of aircraft reconnaissance in some basins in the 1940s and satellite data in the 1960s had a profound effect on our ability to accurately identify and measure tropical cyclones, particularly those that never encountered land or a ship. While aircraft reconnaissance programs have continued in the North Atlantic, they were terminated in the Western Pacific in 1987. The introduction of geostationary satellite imagery in the 1970s, and the introduction (and subsequent improvement) of new tropical cyclone analysis methods (such as the Dvorak technique for estimating storm intensity), further compromises the homogeneity of historical records of tropical cyclone activity.

Regarding impacts to the physical environment, soil moisture is a key variable for which data sets are extremely scarce (e.g., Robock et al., 2000; Seneviratne et al., 2010). This represents a critical issue for the validation and correct representation of soil moisture (agricultural) as well as hydrological drought (Box 3-3) in climate, land surface, and

hydrological models, and the monitoring of ongoing changes in regional terrestrial water storage. As a consequence, these need to be inferred from simple climate indices or model-based approaches (Box 3-3). Such estimates rely in large part on precipitation observations, which have, however, inadequate spatial coverage for these applications in many regions of the world (e.g., Oki et al., 1999; Fekete et al., 2004; Koster et al., 2004a). Similarly, runoff observations are not globally available, which results in significant uncertainties in the closing of the global and some regional water budgets (Legates et al., 2005; Peel and McMahon, 2006; Dai et al., 2009; Teuling et al., 2009), as well as for the global analysis of changes in the occurrence of floods (Section 3.5.2). Additionally, ground observations of snow, which are lacking in several regions, are important for the investigation of physical impacts, particularly those related to the cryosphere and runoff generation (e.g., Essery et al., 2009; Rott et al., 2010).

All of the above-mentioned issues lead to uncertainties in observed trends in extremes. In many instances, great care has been taken to develop procedures to reduce the confounding influences of these issues on the data, which in turn helps to reduce uncertainty, and progress has been made in the last 15 years (e.g., Caesar et al., 2006;

Three types of metrics have been considered to avoid these problems, and thereby allow an answer to this question. One approach is to count the number of record-breaking events in a variable and to examine such a count for any trend. However, one would still face the problem of what to do if, for instance, hot extremes are setting new records, while cold extremes are not occurring as frequently as in the past. In such a case, counting the number of records might not indicate whether the climate was becoming more or less extreme, rather just whether there was a shift in the mean climate. Also, the question of how to combine the numbers of record-breaking events in various extremes (e.g., daily precipitation and hot temperatures) would need to be considered. Another approach is to combine indicators of a selection of important extremes into a single index, such as the Climate Extremes Index (CEI), which measures the fraction of the area of a region or country experiencing extremes in monthly mean surface temperature, daily precipitation, and drought. The CEI, however, omits many important extremes such as tropical cyclones and tornadoes, and could, therefore, not be considered a complete index of 'extremeness.' Nor does it take into account complex or multiple extremes, nor the varying thresholds that relate extremes to impacts in various sectors.

A third approach to solving this dilemma arises from the fact that extremes often have deleterious economic consequences. It may therefore be possible to measure the integrated economic effects of the occurrence of different types of extremes into a common instrument such as insurance payout to determine if there has been an increase or decrease in that instrument. This approach would have the value that it clearly takes into account those extremes with economic consequences. But trends in such an instrument will be dominated by changes in vulnerability and exposure and it will be difficult, if not impossible, to disentangle changes in the instrument caused by non-climatic changes in vulnerability or exposure in order to leave a residual that reflects only changes in climate extremes. For example, coastal development can increase the exposure of populations to hurricanes; therefore, an increase in damage in coastal regions caused by hurricane landfalls will largely reflect changes in exposure and may not be indicative of increased hurricane activity. Moreover, it may not always be possible to associate impacts such as the loss of human life or damage to an ecosystem due to climate extremes to a measurable instrument.

None of the above instruments has yet been developed sufficiently as to allow us to confidently answer the question posed here. Thus we are restricted to questions about whether specific extremes are becoming more or less common, and our confidence in the answers to such questions, including the direction and magnitude of changes in specific extremes, depends on the type of extreme, as well as on the region and season, linked with the level of understanding of the underlying processes and the reliability of their simulation in models.

Brown et al., 2008). As a consequence, more complete and homogenous information about changes is now available for at least some variables and regions (Nicholls and Alexander, 2007; Peterson and Manton, 2008). For instance, the development of global databases of daily temperature and precipitation covering up to 70% of the global land area has allowed robust analyses of extremes (see Alexander et al., 2006). In addition, analyses of temperature and precipitation extremes using higher temporal resolution data, such as that available in the Global Historical Climatology Network – Daily data set (Durre et al., 2008) have also proven robust at both a global (Alexander et al., 2006) and regional scale (Sections 3.3.1 and 3.3.2). Nonetheless, as highlighted above, for many extremes, data remain sparse and problematic, resulting in lower ability to establish changes, particularly on a global basis and for specific regions.

3.2.2.   The Causes behind the Changes

This section discusses the main requirements, approaches, and considerations for the attribution of causes for observed changes in extremes. In Sections 3.3 to 3.5, the causes of observed changes in

specific extremes are assessed. A global summary of these assessments is provided in Table 3-1. Climate variations and change are induced by variability internal to the climate system, and changes in external forcings, which include natural external forcings such as changes in solar irradiance and volcanism, and anthropogenic forcings such as aerosol and greenhouse gas emissions principally due to the burning of fossil fuels, and land use and land cover changes. The mean state, extremes, and variability are all related aspects of the climate, so external forcings that affect the mean climate would in general result in changes in extremes. For this reason, we provide in Section 3.2.2.1 a brief overview of human-induced changes in the mean climate to aid the understanding of changes in extremes as the literature directly addressing the causes of changes in extremes is quite limited.

3.2.2.1.   Human-Induced Changes in the Mean Climate that Affect Extremes

The occurrence of extremes is usually the result of multiple factors, which can act either on the large scale or on the regional (and local) scale (see also Section 3.1.6). Some relevant large-scale impacts of

external forcings affecting extremes include net increases in temperature induced by changes in radiation, enhanced moisture content of the atmosphere, and increased land-sea contrast in temperatures, which can, for example, affect circulation patterns and to some extent monsoons. At regional and local scales, additional processes can modulate the overall changes in extremes, including regional feedbacks, in particular linked to land-atmosphere interactions with, for example, soil moisture or snow (e.g., Section 3.1.4). This section briefly reviews the current understanding of the causes (i.e., in the sense of attribution to either external forcing or internal climate variability) of large-scale (and some regional) changes in the mean climate that are of relevance to extreme events, to the extent that they have been considered in detection and attribution studies.

Regarding observed increases in global average annual mean surface temperatures in the second half of the 20th century, we base our analysis on the following AR4 assessment (Hegerl et al., 2007): Most of the observed increase in global average temperatures is *very likely* due to the observed increase in anthropogenic greenhouse gas concentrations. Greenhouse gas forcing alone would *likely* have resulted in a greater warming than observed if there had not been an offsetting cooling effect from aerosol and other forcings. It is *extremely unlikely* (<5%) that the global pattern of warming can be explained without external forcing, and *very unlikely* that it is due to known natural external causes alone. Anthropogenically forced warming over the second half of the 20th century has also been detected in ocean heat content and air temperatures in all continents (Hegerl et al., 2007; Gillett et al., 2008b).

Hegerl et al. (2007) assessed literature that considered detection in temperature trends at scales as small as approximately 500 km. Recent work has provided more evidence of detection of an anthropogenic influence at increasingly smaller spatial scales and for seasonal averages (Stott et al., 2010). For instance, Min and Hense (2007) found that estimates of response to anthropogenic forcing from the multi-model Coupled Model Intercomparison Project 3 (CMIP3) ensemble (see Section 3.2.3.3) provided a better explanation for observed continental-scale seasonal temperature changes than alternative explanations such as natural external forcing or internal variability. In another study, an anthropogenic signal was detected in 20th-century summer temperatures in Northern Hemisphere subcontinental regions except central North America, although the results were more uncertain when anthropogenic and natural signals were considered together (Jones et al., 2008). An anthropogenic signal has also been detected in multi-decadal trends in a US climate extreme index (Burkholder and Karoly, 2007), in the hydrological cycle of the western United States (Barnett et al., 2008), in New Zealand temperatures (Dean and Stott, 2009), and in European temperatures (Christidis et al., 2011a).

Attribution has more stringent demands than those for the detection of an external influence in observations. Overall, attribution at scales smaller than continental has still not yet been established primarily due to the low signal-to-noise ratio and the difficulties of separately attributing effects of the wider range of possible driving processes

(either attributable to external forcing or internal climate variability) at these scales (Hegerl et al., 2007). One reason is that averaging over smaller regions reduces the internal variability less than does averaging over large regions. In addition, the small-scale details of external forcing, and the responses simulated by models, are less credible than large-scale features. For instance, temperature changes are poorly simulated by models in some regions and seasons (Dean and Stott, 2009; van Oldenborgh et al., 2009). Also the inclusion of additional forcing factors, such as land use change and aerosols that can be more important at regional scales, remains a challenge (Lohmann and Feichter, 2007; Pitman et al., 2009; Rotstayn et al., 2009).

One of the significant advances since AR4 is emerging evidence of human influence on global atmospheric moisture content and precipitation. According to the Clausius-Clapeyron relationship, the saturation vapor pressure increases approximately exponentially with temperature. It is physically plausible that relative humidity would remain roughly constant under climate change (e.g., Hegerl et al., 2007). This means that specific humidity increases about 7% for a one degree increase in temperature in the current climate. Indeed, observations indicate significant increases between 1973 and 2003 in global surface specific humidity but not in relative humidity (Willett et al., 2008), and at the largest spatial-temporal scales moistening is close to the Clausius-Clapeyron scaling of the saturated specific humidity ($\sim$7% K^{-1}; Willett et al., 2010), though relative humidity over low- and mid-latitude land areas decreased over a 10-year period prior to 2008 possibly due to a slower temperature increase in the oceans than over the land (Simmons et al., 2010). By comparing observations with model simulations, changes in the global surface specific humidity for 1973-2003 (Willett et al., 2007), and in lower tropospheric moisture content over the 1988-2006 period (Santer et al., 2007) can be attributed to anthropogenic influence.

The increase in the atmospheric moisture content would be expected to lead to an increase in extreme precipitation when other factors do not change. Min et al. (2011) detected an anthropogenic influence in annual maxima of daily precipitation over Northern Hemisphere land areas. The influence of anthropogenic forcing has been detected in the latitudinal pattern of land precipitation trends though the model-simulated magnitude of changes is smaller than that observed (X. Zhang et al., 2007). The smaller changes in model simulations may be due in part to averaging precipitation trends from different model simulations, as spatial patterns of trends simulated by different models are not exactly the same. The influence of anthropogenic greenhouse gases and aerosols on changes in precipitation over high-latitude land areas north of 55°N has also been detected (Min et al., 2008). Detection is possible there, despite limited data coverage, in part because the response to forcing is relatively strong, and because internal variability in precipitation is low in this region.

3.2.2.2.   How to Attribute a Change in Extremes to Causes

The good practice guidance paper on detection and attribution (Hegerl et al., 2010) reconciles terminologies of detection and attribution used

FAQ 3.2 | Has Climate Change Affected Individual Extreme Events?

A changing climate can be expected to lead to changes in climate and weather extremes. But it is challenging to associate a single extreme event with a specific cause such as increasing greenhouse gases because a wide range of extreme events could occur even in an unchanging climate, and because extreme events are usually caused by a combination of factors. Despite this, it may be possible to make an attribution statement about a specific weather event by attributing the changed probability of its occurrence to a particular cause. For example, it has been estimated that human influences have more than doubled the probability of a very hot European summer like that of 2003.

Recent years have seen many extreme events including the extremely hot summer in parts of Europe in 2003 and 2010, and the intense North Atlantic hurricane seasons of 2004 and 2005. Can the increased atmospheric concentrations of greenhouse gases be considered the 'cause' of such extreme events? That is, could we say these events would *not* have occurred if CO_2 had remained at pre-industrial concentrations? For instance, the monthly mean November temperature averaged across the state of New South Wales in Australia for November 2009 is about 3.5 standard deviations warmer than the 1950-2008 mean, suggesting that the chance of such a temperature occurring in the 1950-2008 climate (assuming a stationary climate) is quite low. Is this event, therefore, an indication of a changing climate? In the CRUTEM3V global land surface temperature data set, about one in every 900 monthly mean temperatures observed between 1900 and 1949 lies more than 3.5 standard deviations above the corresponding monthly mean temperature for 1950-2008.[1] Since global temperature was lower in the first half of the 20th century, this clearly indicates that an extreme warm event as rare as the November 2009 temperature in any specific location could have occurred in the past, even if its occurrence in recent times is more probable.

A second complicating issue is that extreme events usually result from a combination of factors, and this will make it difficult to attribute an extreme to a single causal factor. The hot 2003 European summer was associated with a persistent high-pressure system (which led to clear skies and thus more solar energy received at the surface) and too-dry soil (which meant that less solar energy was used for evaporation, leaving more energy to heat the soil). Another example is that hurricane genesis requires weak vertical wind shear, as well as very warm sea surface temperatures. Since some factors, but not others, may be affected by a specific cause such as increasing greenhouse gas concentrations, it is difficult to separate the human influence on a single, specific extreme event from other factors influencing the extreme.

Nevertheless, climate models can sometimes be used to identify if specific factors are changing the likelihood of the occurrence of extreme events. In the case of the 2003 European heat wave, a model experiment indicated that human influences more than doubled the likelihood of having a summer in Europe as hot as that of 2003, as discussed in the AR4. The value of such a probability-based approach – "Does human influence change the likelihood of an event?" – is that it can be used to estimate the influence of external factors, such as increases in greenhouse gases, on the frequency of specific types of events, such as heat waves or cold extremes. The same likelihood-based approach has been used to examine anthropogenic greenhouse gas contribution to flood probability.

The discussion above relates to an individual, specific occurrence of an extreme event (e.g., a single heat wave). For the reasons outlined above it remains very difficult to attribute any individual event to greenhouse gas-induced warming (even if physical reasoning or model experiments suggest such an extreme may be more likely in a changed climate). On the other hand, a long-term trend in an extreme (e.g., heat wave occurrences) is a different matter. It is certainly feasible to test whether such a trend is likely to have resulted from anthropogenic influences on the climate, just as a global warming trend can be assessed to determine its likely cause.

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[1] We used the CRUTEM3V land surface temperature data. We limit our calculation to grid points with long-term observations, requiring at least 50 non-missing values during 1950-2008 for a calendar month and a grid point to be included. A standard deviation is computed for the period 1950-2008. We then count the number of occurrences when the temperature anomaly during 1900-1949 relative to 1950-2008 mean is greater than 3.5 standard deviations, and compare it with the total number of observations for the grid and month in that period. The ratio of these two numbers is 0.00107.

by Working Groups I and II in the AR4. It provides detailed guidance on the procedures that include two main approaches to attribute a change in climate to causes. One is single-step attribution, which involves assessments that attribute an observed change within a system to an external forcing based on explicitly modelling the response of the variable to the external forcings. The alternate procedure is multi-step attribution, which combines an assessment that attributes an observed change in a variable of interest to a change in climate, with a separate assessment that attributes the change to external forcings. Attribution of changes in climate extremes has some unique issues. Observed data

are limited in both quantity and quality (Section 3.2.1), resulting in uncertainty in the estimation of past changes; the signal-to-noise ratio may be low for many variables and insufficient data may be available to detect such weak signals. In addition, global climate models (GCMs) have several issues in simulating extremes and downscaling techniques can only partly circumvent these issues (Section 3.2.3).

Single-step attribution based on optimal detection and attribution (e.g., Hegerl et al., 2007) can in principle be applied to climate extremes. However, the difference in statistical properties between mean values and extremes needs to be carefully addressed (e.g., Zwiers et al., 2011; see also Section 3.1.6). Post-processing of climate model simulations to derive a quantity of interest that is not explicitly simulated by the models, by applying empirical methods or physically based models to the outputs from the climate models, may make it possible to directly compare observed extremes with climate model results. For example, sea level pressure simulated by multiple GCMs has been used to derive geostrophic wind to represent atmospheric storminess and to derive significant wave height on the oceans for the detection of external influence on trends in atmospheric storminess and northern oceans wave heights (X.L. Wang et al., 2009a). GCM-simulated precipitation and temperature have also been downscaled as input to hydrological and snowpack models to infer past and future changes in temperature, timing of the peak flow, and snow water equivalent for the western United States, and this enabled a detection and attribution analysis of human-induced changes in these variables (Barnett et al., 2008).

If a single-step attribution of causes to effects on extremes or physical impacts of extremes is not feasible, it might be feasible to conduct a multiple-step attribution. The assessment would then need to be based on evidence not directly derived from model simulations, that is, physical understanding and expert judgment, or their combination. For instance, in the northern high-latitude regions, spring temperature has increased, and the timing of spring peak flows in snowmelt-fed rivers has shifted toward earlier dates (Regonda et al., 2005; Knowles et al., 2006). A change in streamflow may be attributable to external influence if streamflow regime change can be attributed to a spring temperature increase and if the spring temperature increase can be attributed to external forcings (though these changes may not necessarily be linked to changes in floods; Section 3.5.2). If the chain of processes is established (e.g., in this case additionally supported by the physical understanding that snow melts earlier as spring temperature increases), the confidence in the overall assessment would be similar to, or weaker than, the lower confidence in the two steps in the assessment. In cases where the underlying physical mechanisms are less certain, such as those linking tropical cyclones and sea surface temperature (see Section 3.4.4), the confidence in multi-step attribution can be severely undermined. A necessary condition for multi-step attribution is to establish the chain of mechanisms responsible for the specific extremes being considered. Physically based process studies and sensitivity experiments that help the physical understanding (e.g., Findell and Delworth, 2005; Seneviratne et al., 2006a; Haarsma et al., 2009) can possibly play a role in developing such multi-step attributions.

Extreme events are rare, which means that there are also few data available to make assessments regarding changes in their frequency or intensity (Section 3.2.1). When a rare and high-impact meteorological extreme event occurs, a question that is often posed is whether such an event is due to anthropogenic influence. Because it is very difficult to rule out the occurrence of low-probability events in an unchanged climate and because the occurrence of such events usually involves multiple factors, it is very difficult to attribute an individual event to external forcing (Allen, 2003; Hegerl et al., 2007; Dole et al., 2011; see also FAQ 3.2). However, in this case, it may be possible to estimate the influence of external forcing on the likelihood of such an event occurring (e.g., Stott et al., 2004; Pall et al., 2011; Zwiers et al., 2011).

3.2.3.   Projected Long-Term Changes and Uncertainties

In this section we discuss the requirements and methods used for preparing climate change projections, with a focus on projections of extremes and the associated uncertainties. The discussion draws on the AR4 (Christensen et al., 2007; Meehl et al., 2007b; Randall et al., 2007) with consideration of some additional issues relevant to projections of extremes in the context of risk and disaster management. More detailed assessments of projections for specific extremes are provided in Sections 3.3 to 3.5. Summaries of these assessments are provided in Table 3-1. Overviews of projected regional changes in temperature extremes, heavy precipitation, and dryness are provided in Table 3-3 (see pages 196-202).

3.2.3.1.   Information Sources for Climate Change Projections

Work on the construction, assessment, and communication of climate change projections, including regional projections and of extremes, draws on information from four sources: (1) GCMs; (2) downscaling of GCM simulations; (3) physical understanding of the processes governing regional responses; and (4) recent historical climate change (Christensen et al., 2007; Knutti et al., 2010b). At the time of the AR4, GCMs were the main source of globally available regional information on the range of possible future climates including extremes (Christensen et al., 2007). This is still the case for many regions, as can be seen in Table 3-3.

The AR4 concluded that statistics of extreme events for present-day climate, especially temperature, are generally well simulated by current GCMs at the global scale (Randall et al., 2007). Precipitation extremes are, however, less well simulated (Randall et al., 2007; Box 3-2). As they continue to develop, and their spatial resolution as well as their complexity continues to improve, GCMs could become increasingly useful for investigating smaller-scale features, including changes in extreme weather events. However, when we wish to project climate and weather extremes, not all atmospheric phenomena potentially of relevance can be realistically or explicitly simulated. GCMs include a number of approximations, known as parameterizations, of processes (e.g., relating to clouds) that cannot be fully resolved in climate models. Furthermore,

the assessment of climate model performance with respect to extremes (summarized in Sections 3.3 to 3.5 for specific extremes), particularly at the regional scale, is still limited by the rarity of extreme events that makes evaluation of model performance less robust than is the case for average climate. Evaluation is further hampered by incomplete data on the historical frequency and severity of extremes, particularly for variables other than temperature and precipitation, and for specific regions (Section 3.2.1; Table 3-2).

The requirement for projections of extreme events has provided one of the motivations for the development of regionalization or downscaling techniques (Carter et al., 2007). These have been specifically developed for the study of regional- and local-scale climate change, to simulate weather and climate at finer spatial resolutions than is possible with GCMs – a step that is particularly relevant for many extremes given their spatial scale. These techniques are, nonetheless, constrained by the reliability of large-scale information coming from GCMs. Recent advances in downscaling for extremes are discussed below.

As indicated in the Glossary, downscaling "is a method that derives local-to regional-scale (up to 100 km) information from larger-scale models or data analyses." Two main methods are distinguished: dynamical downscaling and empirical/statistical downscaling (Christensen et al., 2007). The dynamical method uses the output of regional climate models (RCMs), global models with variable spatial resolution, or high-resolution global models. The empirical/statistical methods develop statistical relationships that link the large-scale atmospheric variables with local/regional climate variables. In all cases, the quality of the downscaled product depends on the quality of the driving model. Dynamical and statistical downscaling techniques are briefly introduced hereafter. Specific limitations that need to be considered in the evaluation of projections are also discussed in Section 3.2.3.2.

The most common approach to dynamical downscaling uses high-resolution RCMs, currently at scales of 20 to 50 km, but in some cases down to 10 to 15 km (e.g., Dankers et al., 2007), to represent regional sub-domains, using either observed (reanalysis) or lower-resolution GCM data to provide their boundary conditions. Using non-hydrostatic mesoscale models, applications at 1- to 5-km resolution are also possible for shorter periods (typically a few months, a few full years at most) – a scale at which clouds and convection can be explicitly resolved and the diurnal cycle tends to be better resolved (e.g., Grell et al., 2000; Hay et al., 2006; Hohenegger et al., 2008; Kanada et al., 2010b). Less commonly used approaches to dynamical downscaling involve the use of stretched-grid (variable resolution) models and high-resolution 'time-slice' models (e.g., Cubasch et al., 1995; Gibelin and Deque, 2003; Coppola and Giorgi, 2005) with the latter including some simulations at 20 km globally (Kamiguchi et al., 2006; Kitoh et al., 2009; Kim et al., 2010). The main advantage of dynamical downscaling is its potential for capturing mesoscale nonlinear effects and providing information for many climate variables at a relatively high spatial resolution, although still not as high as some require. Dynamical downscaling cannot provide information at the point (i.e., weather station) scale (a scale at which

the RCM and GCM parameterizations would not work). Like GCMs, RCMs provide precipitation averaged over a grid cell, which means a tendency to more days of light precipitation (Frei et al., 2003; Barring et al., 2006) and reduced magnitude of extremes (Chen and Knutson, 2008; Haylock et al., 2008) compared with point values. These scaling issues need to be considered when evaluating the ability of RCMs and GCMs to simulate precipitation and other extremes.

Statistical downscaling methods use relationships between large-scale fields (predictors) and local-scale surface variables (predictands) that have been derived from observed data, and apply these to equivalent large-scale fields simulated by climate models (Christensen et al., 2007). They may also include weather generators that provide the basis for a number of recently developed user tools that can be used to assess changes in extreme events (Kilsby et al., 2007; Burton et al., 2008; Qian et al., 2008; Semenov, 2008). Statistical downscaling has been demonstrated to have potential in a number of different regions including Europe (e.g., Schmidli et al., 2007), Africa (e.g., Hewitson and Crane, 2006), Australia (e.g., Timbal et al., 2008, 2009), South America (e.g., D'Onofrio et al., 2010) and North America (e.g., Vrac et al., 2007; Dibike et al., 2008). Statistical downscaling methods are able to access finer spatial scales than dynamical methods and can be applied to parameters that cannot be directly obtained from RCMs. Seasonal indices of extremes can, for example, be simulated directly without having to first produce daily time series (Haylock et al., 2006a), or distribution functions of extremes can be simulated (Benestad, 2007). However, statistical downscaling methods require observational data at the desired scale (e.g., the point or station scale) for a long enough period to allow the model to be well trained and validated, and in some methods can lack coherency among multiple climate variables and/or multiple sites. One specific disadvantage of some, but not all, methods based on the analog approach is that they cannot produce extreme events greater in magnitude than have been observed before (Timbal et al., 2009). Moreover, statistical downscaling does not allow for the possibility of future process-based changes in relationships between predictors and predictands (see Section 3.2.3.2). There have been few systematic intercomparisons of dynamical and statistical downscaling approaches focusing on extremes (Fowler et al., 2007b). Two examples focus on extreme precipitation for the United Kingdom (Haylock et al., 2006a) and the Alps (Schmidli et al., 2007), respectively. A few hybrid statistico-dynamical downscaling methods also exist, including a two-step approach used to downscale heavy precipitation events in southern France (Beaulant et al., 2011). A conceptually similar cascading technique has also been used to downscale tropical cyclones (Bender et al., 2010; see Section 3.4.4).

In terms of temporal resolution, while GCMs and RCMs operate at sub-daily time steps, model output at six-hourly or shorter temporal resolutions, which is desirable for some applications such as urban drainage, is less widely available than daily output. Where limited studies have been undertaken, there is evidence that at the typical spatial resolutions used (i.e., non-cloud/convection-resolving scales), RCMs do not adequately represent sub-daily precipitation and the

diurnal cycle of convection (Gutowski et al., 2003; Brockhaus et al., 2008; Lenderink and Van Meijgaard, 2008). Development of sub-daily statistical downscaling methods is constrained by the availability of long observed time series for calibration and validation and this approach is not currently widely used for climate change applications, although some weather generators, for example, do provide hourly information (Maraun et al., 2010).

It is not possible in this chapter to provide assessments of projected changes in extremes at spatial scales smaller than for large regions (Table 3-3). These large-region projections provide a wider context for national or more local projections, where they exist, and, where they do not, a first indication of expected changes, their associated uncertainties, and the evidence available. Several countries, for example in Europe, North America, Australia, and some other regions, have developed national or sub-national projections (generally based on dynamical and/or statistical downscaling), including information about extremes, and a range of other high-resolution information and tools are available from national weather and hydrological services and academic institutions to assist users and decisionmakers.

3.2.3.2. Uncertainty Sources in Climate Change Projections

Uncertainty in climate change projections arises at each of the steps involved in their preparation: determination of greenhouse gas and aerosol precursor emissions (driven by socioeconomic development and represented through the use of multiple emissions scenarios), concentrations of radiatively active species, radiative forcing, and climate response including downscaling. Also, uncertainty in the estimation of the true 'signal' of climate change is introduced by both errors in the model representation of Earth system processes and by internal climate variability.

As was noted in Section 3.2.3.1, most shortcomings in GCMs and RCMs result from the fact that many important small-scale processes (e.g., representations of clouds, convection, land surface processes) are not represented explicitly (Randall et al., 2007). Some processes – particularly those involving feedbacks (Section 3.1.4), and this is especially the case for climate extremes and associated impacts – are still poorly represented and/or understood (e.g., land-atmosphere interactions, ocean-atmosphere interactions, stratospheric processes, blocking dynamics) despite some improvements in the simulations of others (see Box 3-2 and below). Therefore, limitations in computing power and in the scientific understanding of some physical processes currently restrict further global and regional climate model improvements. In addition, uncertainty due to structural or parameter errors in GCMs propagates directly from global model simulations as input to RCMs and thus to downscaled information.

These problems limit quantitative assessments of the magnitude and timing, as well as regional details, of some aspects of projected climate change. For instance, even atmospheric models with approximately 20-km

horizontal resolution still do not resolve the atmospheric processes sufficiently finely to simulate the high wind speeds and low pressure centers of the most intense hurricanes (Knutson et al., 2010). Realistically capturing details of such intense hurricanes, such as the inner eyewall structure, would require models with 1-km horizontal resolution, far beyond the capabilities of current GCMs and of most current RCMs (and even global numerical weather prediction models). Extremes may also be impacted by mesoscale circulations that GCMs and even current RCMs cannot resolve, such as low-level jets and their coupling with intense precipitation (Anderson et al., 2003; Menendez et al., 2010). Another issue with small-scale processes is the lack of relevant observations, such as is the case with soil moisture and vegetation processes (Section 3.2.1) and relevant parameters (e.g., maps of soil types and associated properties, see for instance Seneviratne et al., 2006b; Anders and Rockel, 2009).

Since many extreme events, such as those associated with precipitation, occur at rather small temporal and spatial scales, where climate simulation skill is currently limited and local conditions are highly variable, projections of future changes cannot always be made with a high level of confidence (Easterling et al., 2008). The credibility of projections of changes in extremes varies with extreme type, season, and geographical region (Box 3-2). Confidence and credibility in projected changes in extremes increase when the physical mechanisms producing extremes in models are considered reliable, such as increases in specific humidity in the case of the projected increase in the proportion of summer precipitation falling as intense events in central Europe (Kendon et al., 2010). The ability of a model to capture the full distribution of variables – not just the mean – together with long-term trends in extremes, implies that some of the processes relevant to a future warming world may be captured (Alexander and Arblaster, 2009; van Oldenborgh et al., 2009). It should nonetheless be stressed that physical consistency of simulations with observed behavior provides necessary but not sufficient evidence for credible projections (Gutowski et al., 2008a).

While downscaling provides more spatial detail (Section 3.2.3.1), the added value of this step and the reliability of projections always needs to be assessed (Benestad et al., 2007; Laprise et al., 2008). A potential limitation and source of uncertainty in downscaling methods is that the calibration of statistical models and the parameterization schemes used in dynamical models are necessarily based on present (and past) climate (as well as an understanding of physical processes). Thus they may not be able to capture changes in extremes that are induced by future mechanistic changes in regional (or global) climate, that is, if used outside the range for which they were designed (Christensen et al., 2007). Spatial inhomogeneity of both land use/land cover and aerosol forcing adds to regional uncertainty. This means that the factors inducing uncertainty in the projections of extremes in different regions may differ considerably. Some specific issues inducing uncertainties in RCM projections are the interactions with the driving GCM, especially in terms of biases and climate change signal (e.g., de Elía et al., 2008; Laprise et al., 2008; Kjellström and Lind, 2009; Déqué et al., 2011) and the choice of regional domain (Wang et al., 2004; Laprise et al., 2008).

In the case of statistical downscaling, uncertainties are induced by, *inter alia*, the definition and choice of predictors (Benestad, 2001; Hewitson and Crane, 2006; Timbal et al., 2008) and the underlying assumption of stationarity (Raje and Mujumdar, 2010). In general, both approaches to downscaling are maturing and being more widely applied but are still restricted in terms of geographical coverage (Maraun et al., 2010). For many regions of the world, no downscaled information exists at all and regional projections rely only on information from GCMs (see Table 3-3).

For many user-driven applications, impact models need to be included as an additional step for projections (e.g., hydrological or ecosystem models). Because of the previously mentioned issues of scale discrepancies and overall biases, it is necessary to bias-correct RCM data before input to some impacts models (i.e., to bring the statistical properties of present-day simulations in line with observations and to use this information to correct projections). A number of bias correction methods, including quantile mapping and gamma transform, have recently been developed and exhibit promising skill for extremes of daily precipitation (Piani et al., 2010; Themeßl et al., 2011).

3.2.3.3.   Ways of Exploring and Quantifying Uncertainties

Uncertainties can be explored, and quantified to some extent, through the combined use of observations and reanalyses, process understanding, a hierarchy of climate models, and ensemble simulations. Ensembles of model simulations represent a fundamental resource for studying the range of plausible climate responses to a given forcing (Meehl et al., 2007b; Randall et al., 2007). Such ensembles can be generated either by (i) collecting results from a range of models from different modelling centers (multi-model ensembles), to include the impact of structural model differences; (ii) by generating simulations with different initial conditions (intra-model ensembles) to characterize the uncertainties due to internal climate variability; or (iii) varying multiple internal model parameters within plausible ranges (perturbed and stochastic physics ensembles), with both (ii) and (iii) aiming to produce a more systematic estimate of single model uncertainty (Knutti et al., 2010b).

Many of the global models utilized for the AR4 were integrated as ensembles, permitting more robust statistical analysis than is possible if a model is only integrated to produce a single projection. Thus the available CMIP3 Multi-Model Ensemble (MME) GCM simulations reflect both inter- and intra-model variability. In advance of AR4, coordinated climate change experiments were undertaken which provided information from 23 models from around the world (Meehl et al., 2007a). The CMIP3 simulations were made available at the Program for Climate Model Diagnosis and Intercomparison (www-pcmdi.llnl.gov/ipcc/about_ipcc.php). However, the higher temporal resolution (i.e., daily) data necessary to analyze most extreme events were quite incomplete in the archive, with only four models providing daily averaged output with ensemble sizes greater than three realizations and many models not included at all. GCMs are expensive to run, thus a compromise is needed between the

number of models, number of simulations, and the complexity of the models (Knutti, 2010).

Besides the uncertainty due to randomness itself, which is the canonical statistical definition, it is important to distinguish between the uncertainty due to *insufficient agreement* in the model projections, the uncertainty due to *insufficient evidence* (insufficient observational data to constrain the model projections or insufficient number of simulations from different models or insufficient understanding of the physical processes), and the uncertainty induced by *insufficient literature*, which refers to the lack of published analyses of projections. For instance, models may agree on a projected change, but if this change is controlled by processes that are not well understood and validated in the present climate, then there is an inherent uncertainty in the projections, no matter how good the model agreement may be. Similarly, available model projections may agree in a given change, but the number of available simulations may restrain the reliability of the inferred agreement (e.g., because the analyses need to be based on daily data that may not be available from all modelling groups). All these issues have been taken into account in assessing the confidence and likelihood of projected changes in extremes for this report (see Section 3.1.5).

Uncertainty analysis of the CMIP3 MME in AR4 focused essentially on the seasonal mean and inter-model standard deviation values (Christensen et al., 2007; Meehl et al., 2007b; Randall et al., 2007). In addition, confidence was assessed in the AR4 through simple quantification of the number of models that show agreement in the sign of a specific climate change (e.g., sign of the change in frequency of extremes) – assuming that the greater the number of models in agreement, the greater the robustness. However, the shortcoming of this definition of model agreement is that it does not take account of possible common biases among models. Indeed, the ensemble was strictly an 'ensemble of opportunity,' without sampling protocol, and the possible dependence of different models on one another (e.g., due to shared parameterizations) was not assessed (Knutti et al., 2010a). Furthermore, this particular metric, which assesses sign agreement only, can provide misleading conclusions in cases, for example, where the projected changes are near zero. For this reason, in our assessments of projected changes in extreme indices we consider the model agreement as a necessary but not a sufficient condition for likelihood statements [e.g., agreement of 66% of the models, as indicated with shading in several of the figures (Figures 3-3, 3-4, 3-6, 3-8, and 3-10), is a minimum but not a sufficient condition for a change being considered '*likely*'].

Post-AR4 studies have concentrated more on the use of the MME in order to better characterize uncertainty in climate change projections, including those of extremes (Kharin et al., 2007; Gutowski et al., 2008a; Perkins et al., 2009). New techniques have been developed for exploiting the full ensemble information, in some cases using observational constraints to construct probability distributions (Tebaldi and Knutti, 2007; Tebaldi and Sanso, 2009), although issues such as determining appropriate metrics for weighting models are challenging (Knutti et al., 2010a). Perturbed-physics ensembles have also become available (e.g.,

Box 3-2 | Variations in Confidence in Projections of Climate Change: Mean versus Extremes, Variables, Scale

Comparisons of observed and simulated climate demonstrate good agreement for some climate variables such as mean temperature, especially at large horizontal scales (e.g., Räisänen, 2007). For instance, Figure 9.12 of the AR4 (Hegerl et al., 2007) compares the ability of 14 climate models to simulate the temporal variations of mean temperature through the 20th century. When the models included both natural and anthropogenic forcings, they consistently reproduced the decadal variations in global mean temperature. Without the anthropogenic influences the models consistently failed to reproduce the multi-decadal temperature variations. However, when the same models' abilities to simulate the temperature variations for smaller domains were assessed, although the mean temperature produced by the ensemble generally tracked the observed temperature changes, the consistency among the models was poorer than was the case for the global mean (Figure 9.12; Hegerl et al., 2007), partly because averaging over global scales smoothes internal variability or 'noise' more than averaging over smaller domains (see also Section 3.2.2.1). We can conclude that the smaller the spatial domain for which simulations or projections are being prepared, the less confidence we should have in these projections (although in some limited cases regional-scale projections can have higher reliability than larger-scale projections; see Section 3.1.6).

This increased uncertainty at smaller scales results from larger internal variability at smaller scales or 'noise' (i.e., natural variability unrelated to external forcings) and increased model uncertainty, both of which lead to lower model consistency at these scales (Hawkins and Sutton, 2009). The latter factor is largely due to the role of unresolved processes (representations of clouds, convection, land surface processes; see also Section 3.2.3). Hawkins and Sutton (2009) also point out regional variations in these aspects: in the tropics the temperature signal expected from anthropogenic factors is large relative to the model uncertainty and the natural variability, compared with higher latitudes. Figure 9.12 from AR4 (Hegerl et al., 2007) also shows that the models are more consistent in reproducing decadal temperature variations in the tropics than at higher latitudes, even though the magnitudes of the temperature trends are larger at higher latitudes.

Uncertainty in projections also depends on the variables, phenomena, or impacts considered (Sections 3.3. to 3.5.). There is more model uncertainty for variables other than temperature, for instance precipitation (Räisänen, 2007; Hawkins and Sutton, 2011; see also Section 3.2.3). And the situation is more difficult again for extremes. For instance, climate models simulate observed changes in extreme temperatures relatively well, but the frequency, distribution, and intensity of heavy precipitation is more poorly simulated (Randall et al., 2007) as are observed changes in heavy precipitation (e.g., Alexander and Arblaster, 2009). Also, projections of changes in temperature extremes tend to be more consistent across climate models (in terms of sign) than for (wet and dry) precipitation extremes (Tebaldi et al., 2006; Orlowsky and Seneviratne, 2011; see also Figures 3-3 through 3-7 and 3-10) and significant inconsistencies are also found for projections of agricultural (soil moisture) droughts (Wang, 2005; see also Box 3-3; Figure 3-10). For some other extremes, such as tropical cyclones, differences in the regional-scale climate change projections between models can lead to marked differences in projected tropical cyclone activity associated with anthropogenic climate change (Knutson et al., 2010), and thus decrease confidence in projections of changes in that extreme.

The relative importance of various causes of uncertainties in projections is somewhat different for earlier compared with later future periods. For some variables (mean temperature, temperature extremes), the choice of emission scenario becomes more critical than model uncertainty for the second part of the 21st century (Tebaldi et al., 2006; Hawkins and Sutton, 2009, 2011) though this does not apply for mean precipitation and some precipitation-related extremes (Tebaldi et al., 2006; Hawkins and Sutton, 2009, 2011), and has in particular not been evaluated in detail for a wide range of extremes. Users need to be aware of such issues in deciding the range of uncertainties that is appropriate to consider for their particular risk or impacts assessment

In summary, confidence in climate change projections depends on the (temporal and spatial) scale and variable being considered and whether one considers extremes or mean quantities. Confidence is highest for temperature, especially at the global scale, and decreases when other variables are considered, and when we focus on smaller spatial domains (Tables 3-1 and 3-3). Confidence in projections for extremes is generally weaker than for projections of long-term averages.

Collins et al., 2006; Murphy et al., 2007) and used to examine projected changes in extremes and their uncertainties (Barnett et al., 2006; Clark et al., 2006, 2010; Burke and Brown, 2008). Advances have also been made in developing probabilistic information at regional scales from the GCM simulations, but there has been rather less development extending this to probabilistic downscaled regional information and to

extremes (Fowler et al., 2007a; Fowler and Ekstrom, 2009). Perhaps the most comprehensive approach to date for quantifying the influence of the cascade of uncertainties in regional projections is that used to develop the recent United Kingdom Climate Projections (UKCP09; Murphy et al., 2009). A complex Bayesian framework is used to combine a perturbed physics ensemble exploring uncertainties in atmosphere and ocean processes, and the carbon and sulfur cycles, with structural uncertainty (represented by 12 CMIP3 models) and an 11-member RCM perturbed physics ensemble. The published projections provide probability distributions of changes in various parameters including the wettest and hottest days of each season for 25-km grid squares across the United Kingdom. These probabilities are conditional on the emissions scenario (low, medium, high) and are described as representing the "relative degree to which each climate outcome is supported by the evidence currently available, taking into account our understanding of climate science and observations, and using expert judgment" (Murphy et al., 2009).

Both statistical and dynamical downscaling methods are affected by the uncertainties that affect the global models, and a further level of uncertainty associated with the downscaling step also needs to be taken into consideration (see also Sections 3.2.3.1 and 3.2.3.2). The increasing availability of coordinated RCM simulations for different regions permits more systematic exploration of dynamical downscaling uncertainty. Such simulations are available for Europe (e.g., Christensen and Christensen, 2007; van der Linden and Mitchell, 2009) and a few other regions such as North America (Mearns et al., 2009) and West Africa (van der Linden and Mitchell, 2009; Hourdin et al., 2010). RCM intercomparisons have also been undertaken for a number of regions including Asia (Fu et al., 2005), South America (Menendez et al., 2010) and the Arctic (Inoue et al., 2006). A new series of coordinated simulations covering the globe is planned (Giorgi et al., 2009). Increasingly, RCM output from coordinated simulations is made available at the daily time scale, facilitating the analysis of some extreme events. Nevertheless, it is important to point out that ensemble runs with RCMs currently involve a limited number of driving GCMs, and hence only subsample uncertainty space. Ensuring adequate sampling of RCM simulations (both in terms of the number of considered RCMs and number of considered driving GCMs) may be more important for extremes than for changes in mean values (Frei et al., 2006; Fowler et al., 2007a). Internal variability, for example, has been shown to make a significant contribution to the spectrum of variability on at least multi-annual time scales and potentially up to multi-decadal time scales (Kendon et al., 2008; Hawkins and Sutton, 2009, 2011; Box 3-2).

3.3. Observed and Projected Changes in Weather and Climate Extremes

3.3.1. Temperature

Temperature is associated with several types of extremes, for example, heat waves and cold spells, and related impacts, for example, on human health, the physical environment, ecosystems, and energy consumption

(e.g., Chapter 4, Sections 3.5.6 and 3.5.7; see also Case Studies 9.2.1 and 9.2.10). Temperature extremes often occur on weather time scales that require daily or higher time scale resolution data to accurately assess possible changes (Section 3.2.1). It is important to distinguish between daily mean, maximum (i.e., daytime), and minimum (nighttime) temperature, as well as between cold and warm extremes, due to their differing impacts. Spell lengths (e.g., duration of heat waves) are relevant for a number of impacts. Note that we do not consider here changes in diurnal temperature range or frost days, which are not typical 'climate extremes'. There is an extensive body of literature regarding the mechanisms of changes in temperature extremes (e.g., Christensen et al., 2007; Meehl et al., 2007b; Trenberth et al., 2007). Heat waves are generally caused by quasi-stationary anticyclonic circulation anomalies or atmospheric blocking (Xoplaki et al., 2003; Meehl and Tebaldi, 2004; Cassou et al., 2005; Della-Marta et al., 2007b), and/or land-atmosphere feedbacks (in transitional climate regions), whereby the latter can act as an amplifying mechanism through reduction in evaporative cooling (Section 3.1.4), but also induce enhanced persistence due to soil moisture memory (Lorenz et al., 2010). Also snow feedbacks (Section 3.1.4), and possibly changes in aerosols (Portmann et al., 2009), are relevant for temperature extremes. Trends in temperature extremes (either observed or projected) can sometimes be different for the most extreme temperatures (e.g., annual maximum/minimum daily maximum/minimum temperature) than for less extreme events [e.g., cold/warm days/nights; see, for instance, Brown et al. (2008) versus Alexander et al. (2006)]. One reason for this is that 'moderate extremes' such as warm/cold days/nights are generally computed for each day with respect to the long-term statistics for that day, thus, for example, an increase in warm days for annual analyses does not necessarily imply warming for the very warmest days of the year.

Observed Changes

Regional historical or paleoclimatic temperature reconstructions may help place the recent instrumentally observed temperature extremes in the context of a much longer period, but literature on this topic is very sparse and most regional reconstructions are for Europe. For example Dobrovolny et al. (2010) reconstructed monthly and seasonal temperature over central Europe back to 1500 using a variety of temperature proxy records. They concluded that the summer 2003 heat wave and the July 2006 heat wave exceeded the +2 standard deviation (associated with the reconstruction method) of previous monthly temperature extremes since 1500. Barriopedro et al. (2011) showed that the anomalously warm summers of 2003 in western and central Europe and 2010 in eastern Europe and Russia both broke the 500-year long seasonal temperature record over 50% of Europe. The coldest periods within the last five centuries occurred in the winter and spring of 1690. Another 500-year temperature reconstruction was recently completed for the Mediterranean basin by means of documentary data and instrumental observations (Camuffo et al., 2010). It suggests strong natural variability in the basin, possibly exceeding the recent warming, although discontinuities in the records limit the interpretation of this finding.

The AR4 (Trenberth et al., 2007, based on Alexander et al., 2006) reported a statistically significant increase in the numbers of warm nights and a statistically significant reduction in the numbers of cold nights for 70 to 75% of the land regions with data (for the spatial coverage of the underlying data set and the definition of warm/cold days and nights, see Section 3.2.1 and Box 3-1, respectively). Changes in the numbers of warm days and cold days also showed warming, but less marked than for nights, with about 40 to 50% of the area with data showing statistically significant changes consistent with warming (Alexander et al., 2006). Less than 1% of the area with data showed statistically significant trends in cold/warm days and nights that were consistent with cooling (Alexander et al., 2006). Trenberth et al. (2007) also reported, based on Vose et al. (2005), that from 1950 to 2004, the annual trends in minimum and maximum land-surface air temperature averaged over regions with data were 0.20°C per decade and 0.14°C per decade, respectively, and that for 1979 to 2004, the corresponding linear trends for the land areas with data were 0.29°C per decade for both maximum and minimum temperature. Based on this evidence, the IPCC AR4 (SPM; IPCC, 2007b) assessed that it was *very likely* that there had been trends toward warmer and more frequent warm days and warm nights, and warmer and less frequent cold days and cold nights in most land areas.

Regions that were found to depart from this overall behavior toward more warm days and nights and fewer cold days and nights in Alexander et al. (2006) were mostly central North America, the eastern United States, southern Greenland (increase in cold days and decreases in warm days), and the southern half of South America (decrease in warm days; no data available for the northern half of the continent). In central North America and the eastern United States this partial tendency for a negative trend in extremes is also consistent with a reported mean negative trend in temperatures, mostly in the spring to summer season (also termed 'warming hole', e.g., Pan et al., 2004; Portmann et al., 2009). Several explanations have been suggested for this behavior, which seems partly associated with a change in the hydrological cycle, possibly linked to soil moisture and/or aerosol feedbacks (Pan et al., 2004; Portmann et al., 2009).

More recent analyses available since the AR4 include a global study (for annual extremes) by Brown et al. (2008) based on the data set from Caesar et al. (2006), and regional studies for North America (Peterson et al., 2008a; Meehl et al., 2009c), Central-Western Europe (since 1880; Della-Marta et al., 2007a), central and eastern Europe (Bartholy and Pongracz, 2007; Kürbis et al., 2009), the eastern Mediterranean region including Turkey (Kuglitsch et al., 2010), western Central Africa, Guinea Conakry and Zimbabwe (Aguilar et al., 2009), the Tibetan Plateau (You et al., 2008) and China (You et al., 2011), Uruguay (Rusticucci and Renom, 2008), and Australia (Alexander and Arblaster, 2009). Further references can also be found in Table 3-2. Overall, these studies are consistent with the assessment of an increase in warm days and nights and a reduction in cold days and nights on the global basis, although they do not necessarily consider trends in all four variables, and a few single studies report trends that are not statistically significant or even trends opposite to the global tendencies in some extremes, subregions,

seasons, or decades. For instance, Rusticucci and Renom (2008) found in Uruguay a reduction of cold nights, a positive but a statistically insignificant trend in warm nights, statistically insignificant decreases in cold days at most investigated stations, and inconsistent trends in warm days. Together with the previous results from Alexander et al. (2006) for southern South America (see above) and further regional studies (Table 3-2), this suggests a less consistent warming tendency in South America compared to other continents. Another notable feature is that studies for central and southeastern Europe display a marked change point in trends in temperature extremes at the end of the 1970s/beginning of 1980s (Table 3-2), which for some extremes can lead to very small and/or statistically not significant overall trends since the 1960s (e.g., Bartholy and Pongracz, 2007).

There are fewer studies available investigating changes in characteristics of cold spells and warm spells, or cold waves and heat waves, compared with studies of the intensity or frequency of warm and cold days or nights. Alexander et al. (2006) provided an analysis of trends in warm spells [based on the Warm Spell Duration Index (WSDI); see Table 3-2 and Box 3-1] mostly in the mid- and high-latitudes of the Northern Hemisphere. The analysis displays a tendency toward a higher length or number of warm spells (increase in number of days belonging to warm spells) in much of the region, with the exception of the southeastern United States and eastern Canada. Regional studies on trends in warm spells or heat waves are also listed in Table 3-2. Kunkel et al. (2008) found that the United States has experienced a general decline in cold waves over the 20th century, with a spike of more cold waves in the 1980s. Further, they report a strong increase in heat waves since 1960, although the heat waves of the 1930s associated with extreme drought conditions still dominate the 1895-2005 time series. Kuglitsch et al. (2009) reported an increase in heat wave intensity, number, and length in summer over the 1960-2006 time period in the eastern Mediterranean region. Ding et al. (2010) reported increasing numbers of heat waves over most of China for the 1961-2007 period. The record-breaking heat wave over western and central Europe in the summer of 2003 is an example of an exceptional recent extreme (Beniston, 2004; Schär and Jendritzky, 2004). That summer (June to August) was the hottest since comparable instrumental records began around 1780 and perhaps the hottest since at least 1500 (Luterbacher et al., 2004). Other examples of recent extreme heat waves include the 2006 heat wave in Europe (Rebetez et al., 2008), the 2007 heat wave in southeastern Europe (Founda and Giannakopoulos, 2009), the 2009 heat wave in southeastern Australia (National Climate Centre, 2009), and the 2010 heat wave in Russia (Barriopedro et al., 2011). Both the 2003 European heat wave (Andersen et al., 2005; Ciais et al., 2005) and the 2009 southeastern Australian heat wave were also associated with drought conditions, which can strongly enhance temperature extremes during heat waves in some regions (see also Section 3.1.4).

Some recent analyses have led to revisions of previously reported trends. For instance, Della-Marta et al. (2007a) found that mean summer maximum temperature change over Europe was +1.6 ± 0.4°C during 1880 to 2005, a somewhat greater increase than reported in earlier
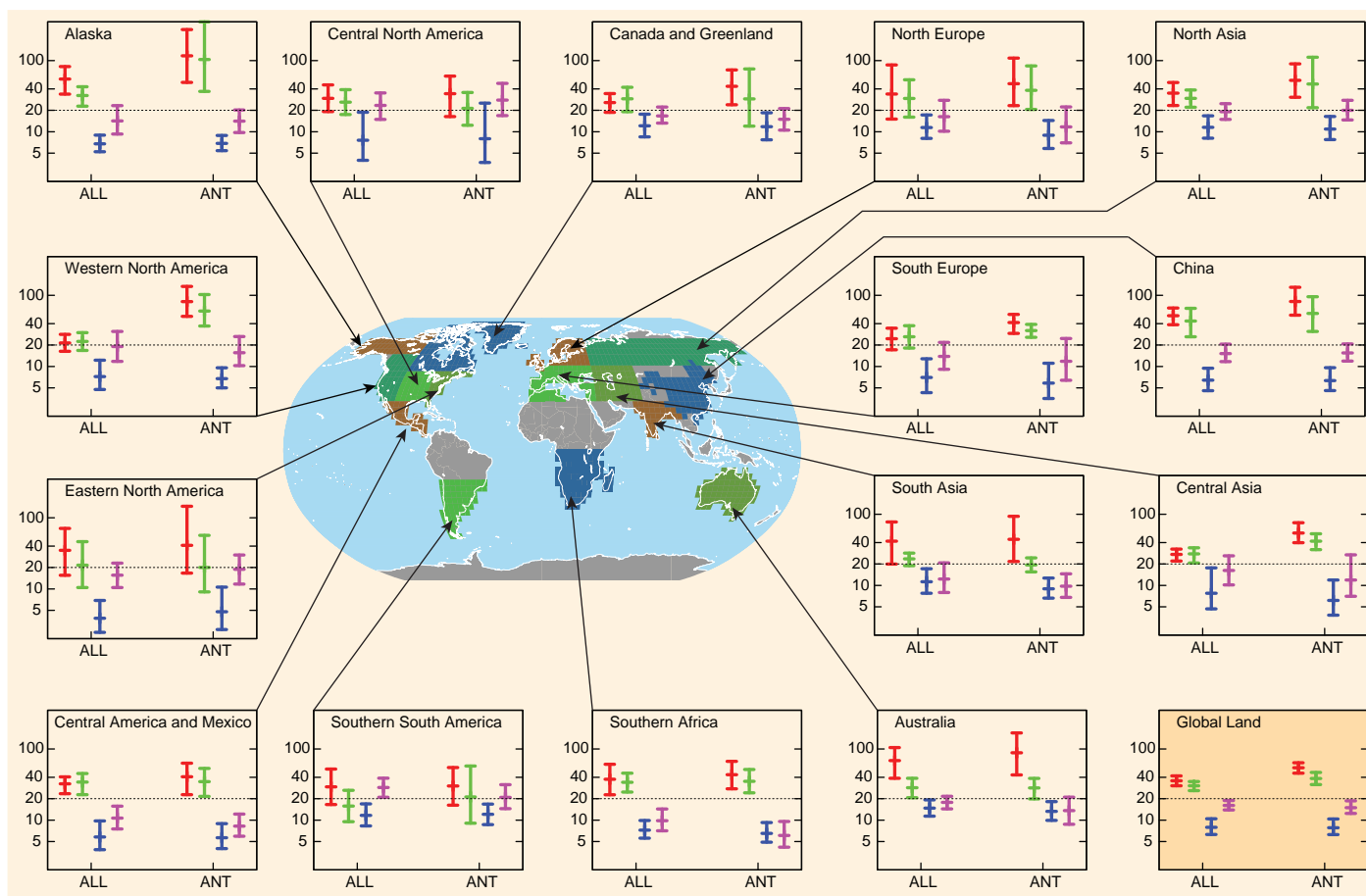
Figure 3-2 | Estimated return periods (years) and their 5 and 95% uncertainty limits for 1960s 20-year return values of annual extreme daily temperatures in the 1990s climate (see text for more details). ANT refers to model simulated responses with only anthropogenic forcing and ALL is both natural and anthropogenic forcing. Error bars are for annual minimum daily minimum temperature (red: TNn), annual minimum daily maximum temperature (green: TXn), annual maximum daily minimum temperature (blue: TNx), and annual maximum daily maximum temperature (pink: TXx), respectively. Grey areas have insufficient data. Source: Zwiers et al., (2011).

studies. Kuglitsch et al. (2009, 2010) homogenized and analyzed over 250 daily maximum and minimum temperature series in the Mediterranean region since 1960, and found that after homogenization the positive trends in the frequency of hot days and heat waves in the Eastern Mediterranean region were higher than reported in earlier studies. This was due to the correction of many warm-biased temperature data in the region during the 1960s and 1970s.

In summary, regional and global analyses of temperature extremes on land generally show recent changes consistent with a warming climate at the global scale, in agreement with the previous assessment in AR4. Only a few regions show changes in temperature extremes consistent with cooling, most notably for some extremes in central North America, the eastern United States, and also parts of South America. Based on the available evidence we conclude that it is *very likely* that there has been an overall decrease in the number of cold days and nights and *very likely* that there has been an overall increase in the number of warm days and nights in most regions, that is, for land areas with data (corresponding to about 70 to 80% of all land areas; see Table 3-2). It is *likely* that this statement applies at the continental scale in North America, Europe, and Australia (Table 3-2). However, some subregions on these continents have had warming trends in temperature extremes that were small or not

statistically significant (e.g., southeastern Europe), and a few subregions have had cooling trends in some temperature extremes (e.g., central North America and eastern United States). Asia also shows trends consistent with warming in most of the continent, but which are assessed here to be of *medium confidence* because of lack of literature for several regions apart from the global study from Alexander et al. (2006). Most of Africa is insufficiently well sampled to allow an overall likelihood statement to be made at the continental scale, although most of the regions on this continent for which data are available have exhibited warming in temperature extremes (Table 3-2). In South America, both lack of data and some inconsistencies in the reported trends imply *low confidence* in the overall trends at the continental scale (Table 3-2). In many (but not all) regions with sufficient data there is *medium confidence* that the number of warm spells or heat waves has increased since the middle of the 20th century (Table 3-2).

Causes of Observed Changes

The AR4 (Hegerl et al., 2007) concluded that surface temperature extremes have *likely* been affected by anthropogenic forcing. This assessment was based on multiple lines of evidence of temperature

extremes at the global scale including the reported increase in the number of warm extremes and decrease in the number of cold extremes at that scale (Alexander et al., 2006). Hegerl et al. (2007) also state that anthropogenic forcing may have substantially increased the risk of extreme temperatures (Christidis et al., 2005) and of the 2003 European heat wave (Stott et al., 2004).

Recent studies on attribution of changes in temperature extremes have tended to reaffirm the conclusions reached in the AR4. Alexander and Arblaster (2009) found that trends in warm nights over Australia could only be reproduced by a coupled model that included anthropogenic forcings. As part of the recent report of the US Climate Change Science Program (CCSP, 2008), Gutowski et al. (2008a) concluded that most of the observed changes in temperature extremes for the second half of the 20th century over the United States can be attributed to human activity. They compared observed changes in the number of frost days, the length of growing season, the number of warm nights, and the heat wave intensity with those simulated in a nine-member multi-model ensemble simulation. The decrease in frost days, an increase in growing season length, and an increase in heat wave intensity all show similar changes over the United States in 20th-century experiments that combine anthropogenic and natural forcings, though the relative contributions of each are unclear.

Results from two global coupled climate models with separate anthropogenic and natural forcing runs indicate that the observed changes are simulated with anthropogenic forcings, but not with natural forcings (even though there are some differences in the details of the forcings). Zwiers et al. (2011) compared observed annual temperature extremes including annual maximum daily maximum and minimum temperatures, and annual minimum daily maximum and minimum temperatures with those simulated responses to anthropogenic forcing or anthropogenic and natural external forcings combined by multiple GCMs. They fitted probability distributions (Box 3-1) to the observed extreme temperatures with a time-evolving pattern of location parameters as obtained from the model simulations, and found that both anthropogenic influence and the combined influence of anthropogenic and natural forcing can be detected in all four extreme temperature variables at the global scale over the land, and also over many large land areas. Globally, return periods for events that were expected to recur once every 20 years in the 1960s are now estimated to exceed 30 years for extreme annual minimum daily maximum temperature and 35 years for extreme annual minimum daily minimum temperature, although these estimates are subject to considerable uncertainty. Further, return periods were found to have decreased to less than 10 or 15 years for annual maximum daily minimum and daily maximum temperatures respectively (Figure 3-2).

However, the available detection and attribution studies for extreme maximum and minimum temperatures (Christidis et al., 2011b; Zwiers et al., 2011) suggest that the models overestimate changes in the maximum temperatures and underestimate changes in the minimum temperatures during the late 20th century.

Projected Changes and Uncertainties

Regarding projections of extreme temperatures, the AR4 (Meehl et al., 2007b) noted that cold episodes were projected to decrease significantly in a future warmer climate and considered it *very likely* that heat waves would be more intense, more frequent, and last longer in a future warmer climate. Post-AR4 studies of temperature extremes have utilized larger model ensembles (Kharin et al., 2007; Sterl et al., 2008; Orlowsky and Seneviratne, 2011) and generally confirm the conclusions of the AR4, while also providing more specific assessments both in terms of the range of considered extremes and the level of regional detail (see also Table 3-3).

There are few global analyses of multi-model projections of temperature extremes available in the literature. The study by Tebaldi et al. (2006), which provided the basis for extreme projections given in the AR4 (Figures 10.18 and 10.19 in Meehl et al., 2007b), provided global analyses of projected changes (A1B scenario) in several extremes indices based on nine GCMs (note that not all modelling groups that saved daily data also calculated the indices). For temperature extremes, analyses were provided for heat wave lengths (using only one index, see discussion in Box 3-1) and warm nights. Stippling was used where five out of nine models displayed statistically significant changes of the same sign. Orlowsky and Seneviratne (2011) recently updated the analysis from Tebaldi et al. (2006) for the full ensemble of GCMs that contributed A2 scenarios to the CMIP3, using a larger number of extreme indices [including several additional analyses of daily extremes (see Figures 3-3 and 3-4), and three heat wave indices instead of one; see also discussion of heat wave indices in Box 3-1], using other thresholds for display and stippling of the figures (no results displayed if less than 66% of the models agree on the sign of change; stippling used only for 90% model agreement), and providing seasonal analyses. This analysis confirms that strong agreement (in terms of sign of change) exists between the various GCM projections for temperature-related extremes, with projected increases in warm day occurrences (Figure 3-3) and heat wave length, and decreases in cold extremes (Figure 3-4). Temperature extremes on land are projected to warm faster than global annual mean temperature in many regions and seasons, implying large changes in extremes in some places, even for a global warming of 2 or 3°C (with scaling factors for the SRES A2 scenario ranging between 0.5 and 2 for moderate seasonal extremes; Orlowsky and Seneviratne, 2011). Based on the analyses of Tebaldi et al. (2006) and Orlowsky and Seneviratne (2011), as well as physical considerations, we assess that increases in the number of warm days and nights and decreases in the number of cold days and nights (defined with respect to present regional climate, i.e., the 1961-1990 reference period, see Box 3-1) are *virtually certain* at the global scale. Further, given the assessed changes in hot and cold days and nights and available analyses of projected changes in heat wave length in the two studies, we assess that it is *very likely* that the length, frequency, and/or intensity of heat waves will increase over most land areas.

Another global study of changes in extremes based on the CMIP3 ensemble is provided in Kharin et al. (2007), which focuses on changes

in annual extremes (20-year extreme values) based on 12 GCMs for temperature extremes and 14 GCMs for precipitation extremes employing the SRES A2, A1B, and B1 emissions scenarios. This analysis projects increases in the temperature of the 1-in-20 year annual extreme hottest day of about 2 to 6°C (depending on region and scenario; Figure 3-5 adapted from Kharin et al., 2007) and strong reductions in the return periods of this extreme event by the end of the 21st century. However, as noted above, the limited number of relevant detection and attribution studies suggests that models may overestimate some changes in temperature extremes, and our assessments take this into account by reducing the level of certainty in the assessments from what would be derived by uncritical acceptance of the projections in Figure 3-5. The assessments are also weakened to reflect the possibility that some important processes relevant to extremes may be missing or be poorly represented in models, as well as the fact that the model projections considered in this study did not correspond to the full CMIP3 ensemble. Hence, we assess that in terms of absolute values, the 20-year extreme annual daily maximum temperature (i.e., return value) will *likely* increase by about 2 to 5°C by the late 21st century, and by about 1 to 3°C by mid-21st century, depending on the region and emissions scenario (considering the B1, A1B, and A2 scenarios; Figure 3-5a). Furthermore,

we assess that globally under the A2 and A1B scenarios a 1-in-20 year annual extreme hot day is *likely* to become a 1-in-2 year annual extreme by the end of the 21st century in most regions, except in the high latitudes of the Northern Hemisphere where it is *likely* to become a 1-in-5 year annual extreme (Figure 3-5b, based on material from Kharin et al., 2007). Further, we assess that under the more moderate B1 scenario a current 1-in-20 year extreme would *likely* become a 1-in-5 year event (and a 1-in-10 year event in Northern Hemisphere high latitudes).

Next, regional assessments of projected changes in temperature extremes are provided. More details are found in Table 3-3. For North America, the CCSP reached the following conclusions (using IPCC AR4 likelihood terminology) regarding projected changes in temperature extremes by the end of the 21st century (Gutowski et al., 2008a):

1) Abnormally hot days and warm nights and heat waves are *very likely* to become more frequent.
2) Cold days and cold nights are *very likely* to become much less frequent.
3) For a mid-range scenario (A1B) of future greenhouse gas emissions, a day so hot that it is currently experienced only once every 20 years would occur every 3 years by the middle of the century over
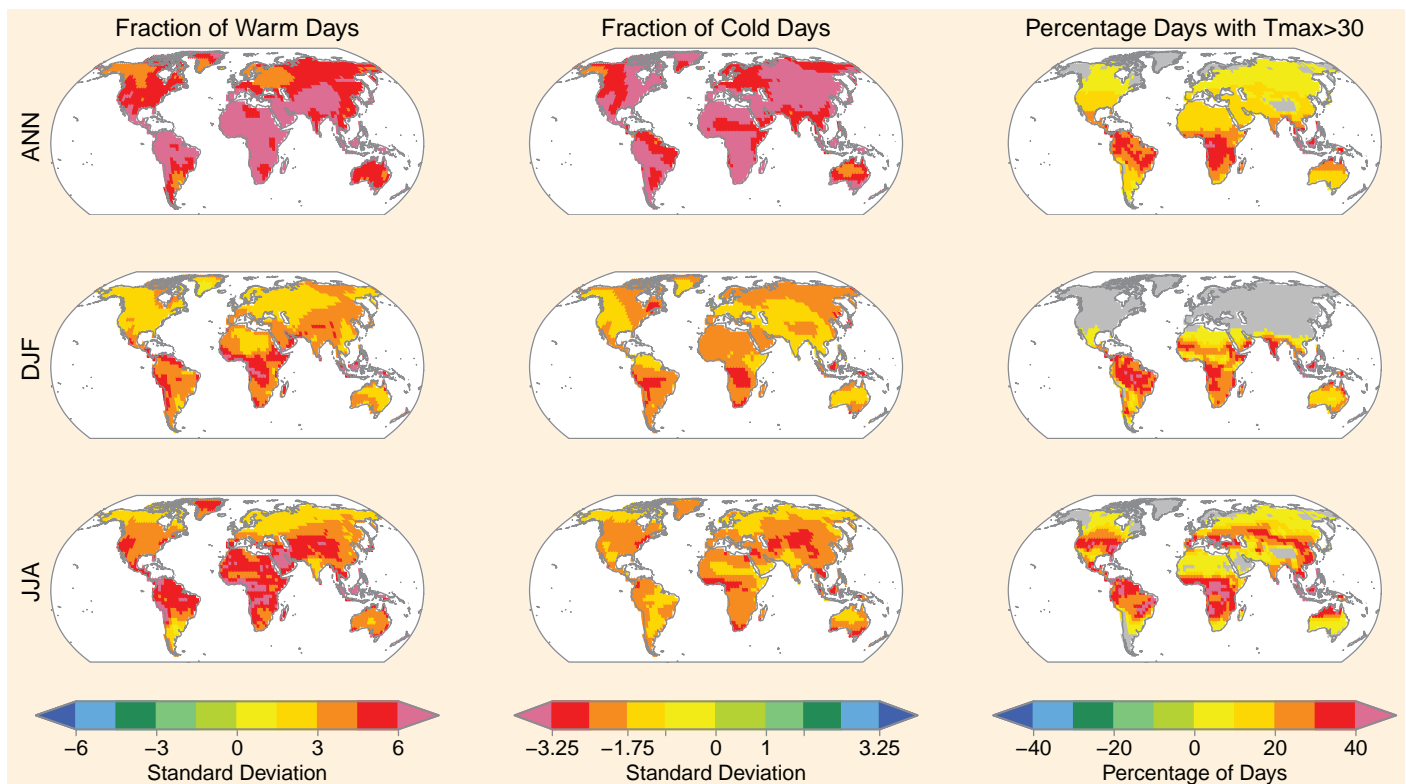
Figure 3-3 | Projected annual and seasonal changes in three indices for daily Tmax for 2081-2100 with respect to 1980-1999, based on 14 GCMs contributing to the CMIP3. Left column: fraction of warm days (days in which Tmax exceeds the 90th percentile of that day of the year, calculated from the 1961-1990 reference period); middle column: fraction of cold days (days in which Tmax is lower than the 10th percentile of that day of the year, calculated from the 1961-1990 reference period); right column: percentage of days with Tmax >30°C. The changes are computed for the annual time scale (top row) and two seasons (December-January-February, DJF, middle row, and June-July-August, JJA, bottom row) as the fractions/percentages in the 2081-2100 period (based on simulations for emission scenario SRES A2) minus the fractions/percentages of the 1980-1999 period (from corresponding simulations for the 20th century). Warm day and cold day changes are expressed in units of standard deviations, derived from detrended per year annual or seasonal estimates, respectively, from the three 20-year periods 1980-1999, 2046-2065, and 2081-2100 pooled together. Tmax >30°C changes are given directly as differences in percentage points. Color shading is only applied for areas where at least 66% (i.e., 10 out of 14) of the GCMs agree on the sign of the change; stippling is applied for regions where at least 90% (i.e.,13 out of 14) of the GCMs agree on the sign of the change. Adapted from Orlowsky and Seneviratne (2011); updating Tebaldi et al. (2006) for additional number of indices and CMIP3 models, and including seasonal time frames. For more details, see Appendix 3.A.
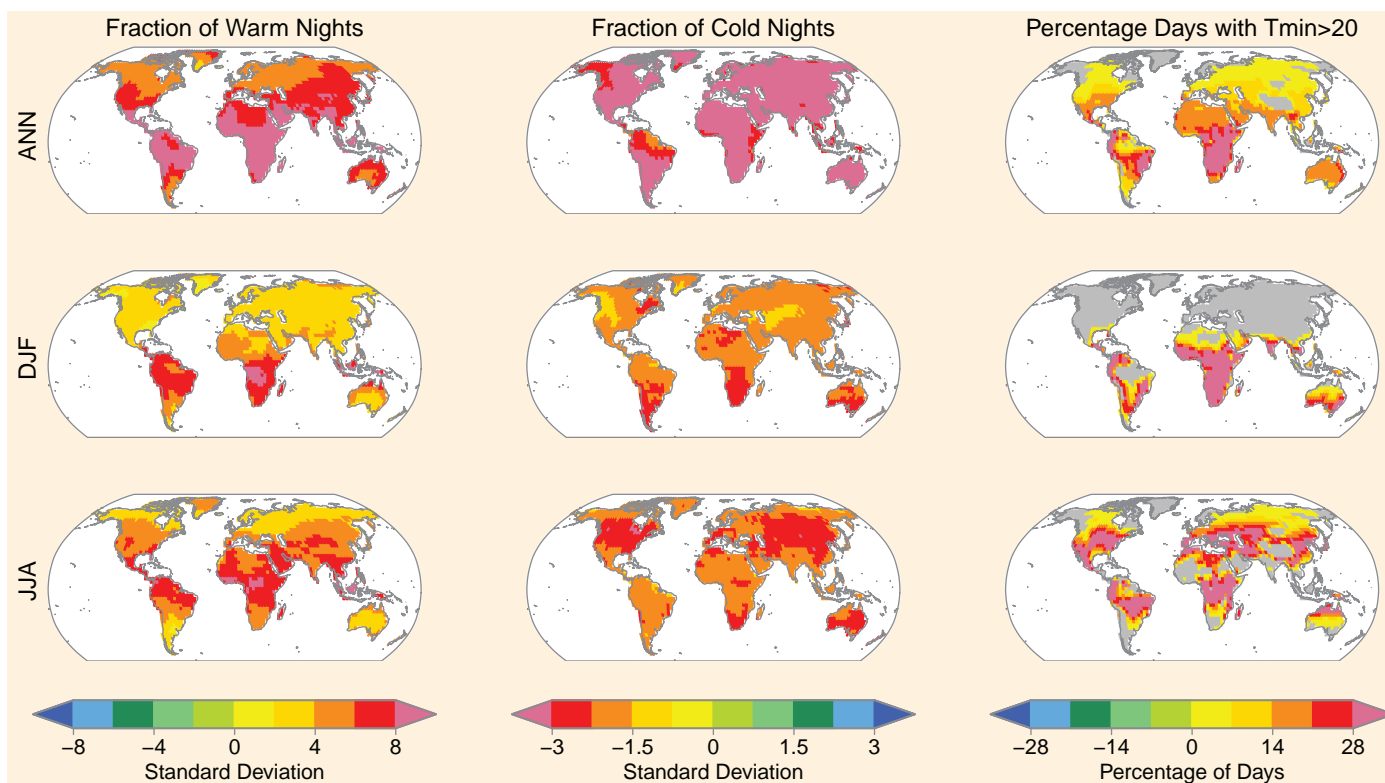
Figure 3-4 | Projected annual and seasonal changes in three indices for daily Tmin for 2081-2100 with respect to 1980-1999, based on 14 GCMs contributing to the CMIP3. Left column: fraction of warm nights (days at which Tmin exceeds the 90th percentile of that day of the year, calculated from the 1961-1990 reference period); middle column: fraction of cold nights (days at which Tmin is lower than the 10th percentile of that day of the year, calculated from the 1961-1990 reference period); right column: percentage of days with Tmin >20°C. The changes are computed for the annual time scale (top row) and two seasons (December-January-February, DJF, middle row, and June-July-August, JJA, bottom row) as the fractions/percentages in the 2081-2100 period (based on simulations under emission scenario SRES A2) minus the fractions/percentages of the 1980-1999 period (from corresponding simulations for the 20th century). Warm night and cold night changes are expressed in units of standard deviations, derived from detrended per year annual or seasonal estimates, respectively, from the three 20-year periods 1980-1999, 2046-2065, and 2081-2100 pooled together. Tmin >20°C changes are given directly as differences of percentage points. Color shading is only applied for areas where at least 66% (i.e., 10 out of 14) of the GCMs agree in the sign of the change; stippling is applied for regions where at least 90% (i.e.,13 out of 14) of the GCMs agree in the sign of the change. Adapted from Orlowsky and Seneviratne (2011); updating Tebaldi et al. (2006) for additional number of indices and CMIP3 models, and including seasonal time frames. For more details, see Appendix 3.A.

much of the continental United States and every 5 years over most of Canada; by the end of the century, it would occur every other year or more.

Meehl et al. (2009c) examined changes in record daily high and low temperatures in the United States and show that even with projected strong warming resulting in many more record highs than lows, the occasional record low is still set. For Australia, the CMIP3 ensemble projected increases in warm nights (15-40% by the end of the 21st century) and heat wave duration, together with a decrease in the number of frost days (Alexander and Arblaster, 2009). Inland regions show greater warming compared with coastal zones (Suppiah et al., 2007; Alexander and Arblaster, 2009) and large increases in the number of days above 35 or 40°C are indicated (Suppiah et al., 2007). For the entire South American region, a study with a single RCM projected more frequent warm nights and fewer cold nights (Marengo et al., 2009a). Several studies of regional and global model projections of changes in extremes are available for the European continent (see also Table 3-3). Analyses of both global and regional model outputs show major increases in warm temperature extremes across the Mediterranean region including events such as hot days (Tmax >30°C) and tropical nights (Tmin>20°C) (Giannakopoulos et al., 2009; Tolika et al., 2009).

Comparison of RCM projections using the A1B forcing scenario, with data for 2007 (the hottest summer in Greece in the instrumental record with a record daily Tmax observed value of 44.8°C) indicates that the distribution for 2007 is closer to the distribution for 2071-2100 than for the 2021-2050 period, thus 2007 might be considered a 'normal' summer of the future (Founda and Giannakopoulos, 2009; Tolika et al., 2009). Beniston et al. (2007) concluded from an analysis of RCM output that regions such as France and Hungary may experience as many days per year above 30°C as currently experienced in Spain and Sicily. In this RCM ensemble, France was the area with the largest projected warming in the uppermost percentiles of daily summer temperatures although the mean warming was greatest in the Mediterranean region (Fischer and Schär, 2009). New results from an RCM ensemble project increases in the amplitude, frequency, and duration of health-impacting heat waves, especially in southern Europe (Fischer and Schär, 2010). Overall these regional assessments are consistent with the global assessments provided above. It should be noted, however, that the assessed uncertainty is larger at the regional level than at the continental or global level (see Box 3-2). Global-scale trends in a specific extreme may be either more reliable or less reliable than regional-scale trends, depending on the geographical uniformity of the trends in the specific extreme (Section 3.1.6).
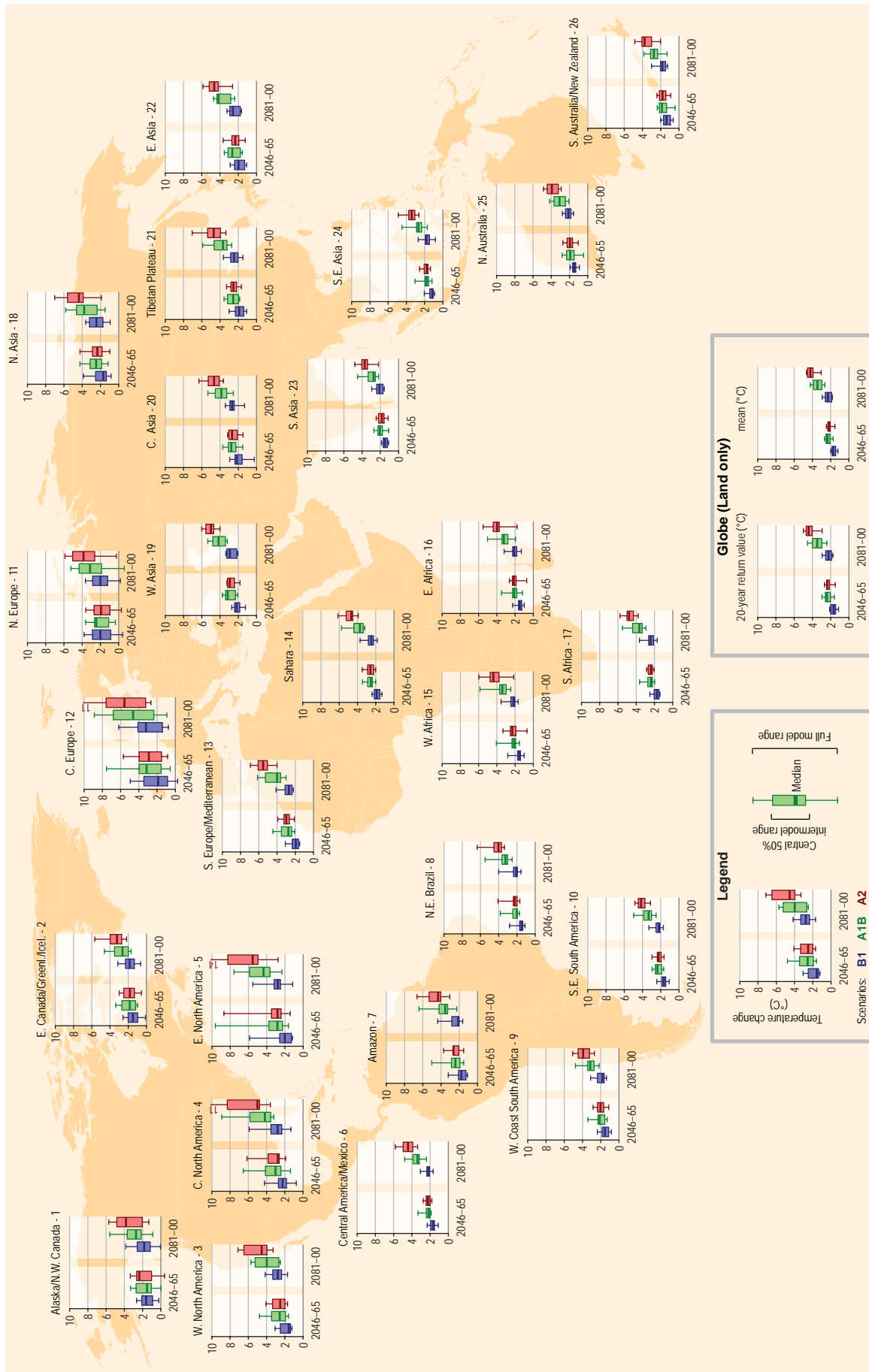
Figure 3-5a | Projected changes (in °C) in 20-year return values of the annual maximum of the daily maximum temperature. The bar plots (see legend for more information) show results for regionally averaged projections for two time horizons, 2046 to 2065 and 2081 to 2100, as compared to the late 20th century (1981-2000), and for three different SRES emission scenarios (B1, A1B, A2). Results are based on 12 GCMs contributing to the CMIP3. See Figure 3-1 for defined extent of regions. Values are computed for land points only. The 'Globe' analysis (inset box) displays the change in 20-year return values of the annual maximum of the daily maximum temperature computed using all land grid points (left), and the change in annual mean daily maximum temperature computed using all land grid points (right). Adapted from the analysis of Kharin et al. (2007). For more details, see Appendix 3.A.
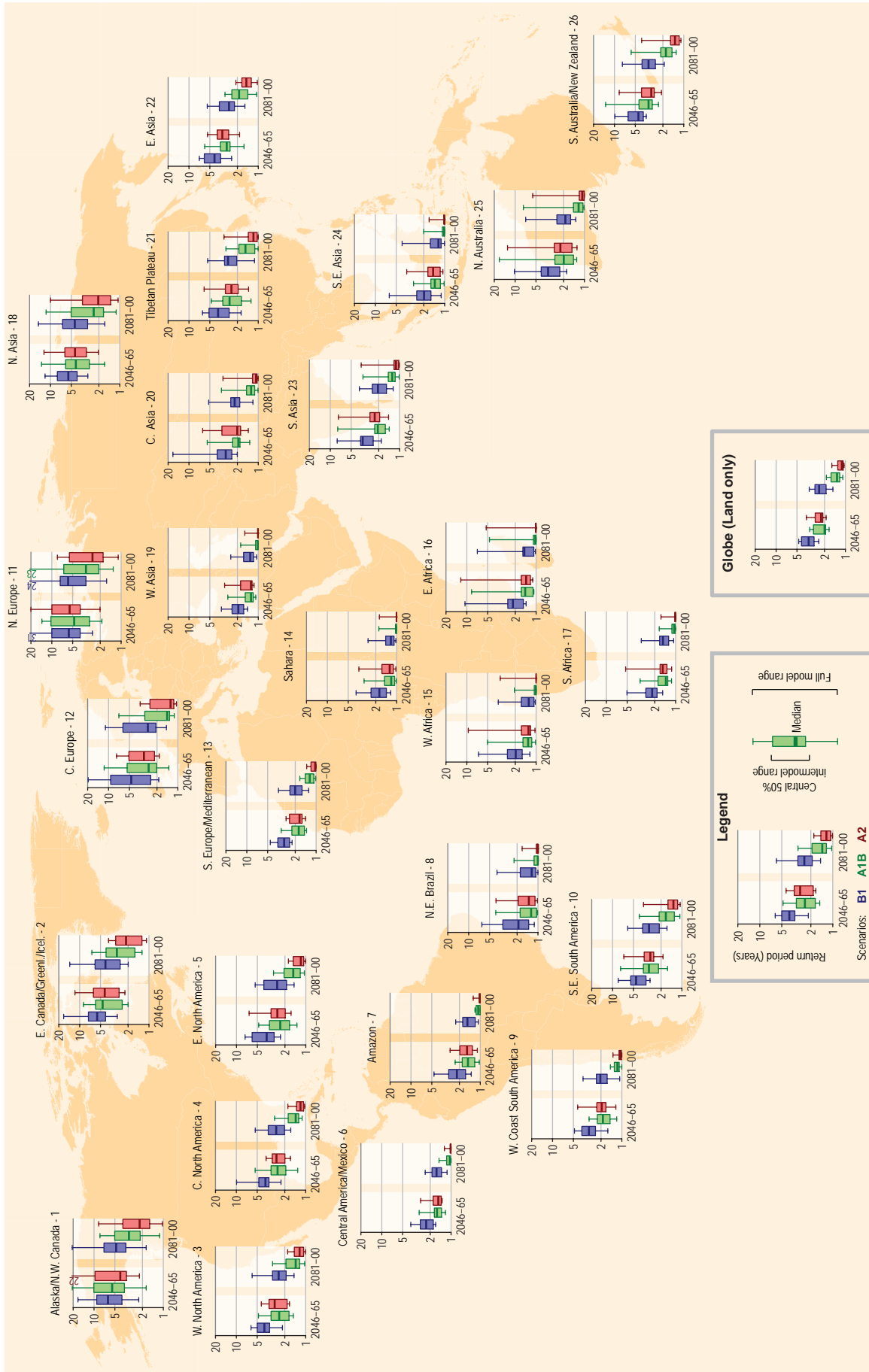
Figure 3-5b | Projected return period (in years) of late 20th-century 20-year return values of the annual maximum of the daily maximum temperature. The bar plots (see legend for more information) show results for regionally averaged projections for two time horizons, 2046 to 2065 and 2081 to 2100, as compared to the late 20th century (1981-2000), and for three different SRES emission scenarios (B1, A1B, A2). Results are based on 12 GCMs contributing to the CMIP3. See Figure 3-1 for defined extent of regions. The 'Globe' analysis (inset box) displays the projected return period (in years) of late 20th-century 20-year return values of the annual maximum of the daily maximum temperature computed using all land grid points. Adapted from the analysis of Kharin et al. (2007). For more details, see Appendix 3.A.

Temperature extremes were the type of extremes projected to change with most confidence in the AR4 (IPCC, 2007a). This is confirmed regarding the sign of change with more recent analyses (Figures 3-3 and 3-4), although there is a large spread with respect to the magnitude of changes both due to emission scenario and climate model uncertainty (Figures 3-5a,b). If changes in temperature extremes scale with changes in mean temperature (i.e., simple shifts of the probability distribution), we infer that it is *virtually certain* that hot extremes will increase and cold extremes will decrease over the 21st century with respect to the 1960-1990 climate. Changes in the tails of the temperature distributions may not scale with changes in the mean in some regions (Section 3.1.6), though in most such reported cases hot extremes tend to increase and cold extremes decrease more than mean temperature, and thus the above statement for extremes (*virtually certain* increase in hot extremes and decrease in cold extremes) still applies. Central and eastern Europe is a region where the evidence suggests that projected changes in temperature extremes result from both changes in the mean as well as from changes in the shape of the probability distributions (Schär et al., 2004). The main mechanism for the widening of the distribution is linked to the drying of the soil in this region (Sections 3.1.4 and 3.1.6). Furthermore, remote surface heating may induce circulation changes that modify the temperature distribution (Haarsma et al., 2009). Other local, mesoscale, and regional feedback mechanisms, in particular with land surface conditions (beside soil moisture, also with vegetation and snow; Section 3.1.4) and aerosol concentrations (Ruckstuhl and Norris, 2009) may enhance the uncertainties in temperature projections. Some of these processes occur at a small scale unresolved by the models (Section 3.2.3). In addition, lack of observational data (e.g., for soil moisture and snow cover; see Section 3.2.1) reduces the possibilities to evaluate climate models (e.g., Roesch, 2006; Boe and Terray, 2008; Hall et al., 2008; Brown and Mote, 2009). Because of these various processes and associated uncertainties, mean global warming does not necessarily imply warming in all regions and seasons (see also Section 3.1.6). Regarding mesoscale processes, lack of information also affects confidence in projections. One example is changes in heat waves in the Mediterranean region that are suggested to have the largest impact in coastal areas, due to the role of enhanced relative humidity in health impacts (Diffenbaugh et al., 2007; Fischer and Schär, 2010). But it is not clear how this pattern may or may not be moderated by sea breezes (Diffenbaugh et al., 2007).

In summary, since 1950 it is *very likely* that there has been an overall decrease in the number of cold days and nights and an overall increase in the number of warm days and nights at the global scale, that is, for land areas with sufficient data. It is *likely* that such changes have also occurred at the continental scale in North America, Europe, and Australia. There is *medium confidence* in a warming trend in daily temperature extremes in much of Asia. *Confidence* in historical trends in daily temperature extremes in Africa and South America generally varies from *low* to *medium* depending on the region. Globally, in many (but not all) regions with sufficient data there is *medium confidence* that the length or number of warm spells or heat waves has increased since the

middle of the 20th century. It is *likely* that anthropogenic influences have led to warming of extreme daily minimum and maximum temperatures at the global scale. Models project substantial warming in temperature extremes by the end of the 21st century. It is *virtually certain* that increases in the frequency and magnitude of warm days and nights and decreases in the cold days and nights will occur through the 21st century at the global scale. This is mostly linked with mean changes in temperatures, although changes in temperature variability can play an important role in some regions. It is *very likely* that the length, frequency, and/or intensity of warm spells or heat waves (defined with respect to present regional climate) will increase over most land areas. For the SRES A2 and A1B emission scenarios a 1-in-20 year annual hottest day is *likely* to become a 1-in-2 year annual extreme by the end of the 21st century in most regions, except in the high latitudes of the Northern Hemisphere where it is *likely* to become a 1-in-5 year annual extreme. In terms of absolute values, 20-year extreme annual daily maximum temperature (i.e., return value) will *likely* increase by about 1 to 3°C by mid-21st century and by about 2 to 5°C by the late 21st century, depending on the region and emissions scenario (Figure 3-5). Moderate temperature extremes on land are projected to warm faster than global annual mean temperature in many regions and seasons. Projected changes at subcontinental scales are less certain than is the case for the global scale. Regional changes in temperature extremes will differ from the mean global temperature change. Mean global warming does not necessarily imply warming in all regions and seasons.

3.3.2.   Precipitation

This section addresses changes in daily extreme or heavy precipitation events. Reductions in mean (or total) precipitation that can lead to drought (i.e., associated with lack of precipitation) are considered in Section 3.5.1. Because climates are so diverse across different parts of the world, it is difficult to provide a single definition of extreme or heavy precipitation. In general, two different approaches have been used: (1) relative thresholds such as percentiles (typically the 95th percentile) and return values; and (2) absolute thresholds [e.g., 50.8 mm (2 inches) day^{-1} of rain in the United States, and 100 mm day^{-1} of rain in China]. For more details on the respective drawbacks and advantages of these two approaches, see Section 3.1 and Box 3-1. Note that we do not distinguish between rain and snowfall (both considered as contributors to overall extreme precipitation events) as they are not treated separately in the literature, but do distinguish changes in hail from other precipitation types. Increases in public awareness and changes in reporting practices have led to inconsistencies in the record of severe thunderstorms and hail that make it difficult to detect trends in the intensity or frequency of these events (Kunkel et al., 2008). Furthermore, weather events such as hail are not well captured by current monitoring systems and, in some parts of the world, the monitoring network is very sparse (Section 3.2.1), resulting in considerable uncertainty in the estimates of extreme

precipitation. There are also known biases in precipitation measurements, mostly leading to rain undercatch. Little evidence of paleoclimatic and historical changes in heavy precipitation is available to place recent variations into context.

Observed Changes

The AR4 (Trenberth et al., 2007) concluded that it was *likely* that there had been increases in the number of heavy precipitation events (e.g., 95th percentile) over the second half of the 20th century within many land regions, even in those where there had been a reduction in total precipitation amount, consistent with a warming climate and observed significant increasing amounts of water vapor in the atmosphere. Increases had also been reported for rarer precipitation events (1-in-50 year return period), but only a few regions had sufficient data to assess such trends reliably. However, the AR4 (Trenberth et al., 2007) also stated that "Many analyses indicate that the evolution of rainfall statistics through the second half of the 20th century is dominated by variations on the interannual to inter-decadal time scale and that trend estimates are spatially incoherent (Manton et al., 2001; Peterson et al., 2002; Griffiths et al., 2003; Herath and Ratnayake, 2004)". Overall, as highlighted in Alexander et al. (2006), the observed changes in precipitation extremes were found at the time to be much less spatially coherent and statistically significant compared to observed changes in temperature extremes: although statistically significant trends toward stronger precipitation extremes were generally found for a larger fraction of the land area than trends toward weaker precipitation extremes, statistically significant changes in precipitation indices for the overall land areas with data were only found for the Simple Daily Intensity index, and not for other considered indices such as Heavy Rainfall Days (Alexander et al., 2006).

Recent studies have updated the assessment of the AR4, with more regional results now available (Table 3-2). Overall, this additional evidence confirms that more locations and studies show an increase than a decrease in extreme precipitation, but that there are also wide regional and seasonal variations, and trends in many regions are not statistically significant (Table 3-2).

Recent studies on past and current changes in precipitation extremes in North America, some of which are included in the recent assessment of the CCSP report (Kunkel et al., 2008), have reported an increasing trend over the last half century. Based on station data from Canada, the United States, and Mexico, Peterson et al. (2008a) reported that heavy precipitation has been increasing over 1950-2004, as well as the average amount of precipitation falling on days with precipitation. For the contiguous United States, DeGaetano (2009) showed a 20% reduction in the return period for extreme precipitation of different return levels over 1950-2007; Gleason et al. (2008) reported an increasing trend in the area experiencing a much above-normal proportion of heavy daily precipitation from 1950 to 2006; and Pryor et al. (2009) provided evidence of increases in the intensity of events above the 95th percentile during the 20th century, with a larger magnitude of the increase at the end of the

century. The largest trends toward increased annual total precipitation, number of rainy days, and intense precipitation (e.g., fraction derived from events in excess of the 90th percentile value) were focused on the Great Plains/northwestern Midwest (Pryor et al., 2009). In the core of the North American monsoon region in northwest Mexico, statistically significant positive trends were found in daily precipitation intensity and seasonal contribution of daily precipitation greater than its 95th percentile in the mountain sites for the period 1961-1998. However, no statistically significant changes were found in coastal stations (Cavazos et al., 2008). Overall, the evidence indicates a *likely* increase in observed heavy precipitation in many regions in North America, despite statistically non-significant trends and some decreases in some subregions (Table 3-2). This general increase in heavy precipitation accompanies a general increase in total precipitation in most areas of the country.

There is *low* to *medium confidence* in trends for Central and South America, where spatially varying trends in extreme rainfall events have been observed (Table 3-2). Positive trends in many areas but negative trends in some regions are evident for Central America and northern South America (Dufek and Ambrizzi, 2008; Marengo et al., 2009b; Re and Ricardo Barros, 2009; Sugahara et al., 2009). For the western coast of South America, a decrease in extreme rainfall in many areas and an increase in a few areas are observed (Haylock et al., 2006b).

There is *medium confidence* in trends in heavy precipitation in Europe, due to partly inconsistent signals across studies and regions, especially in summer (Table 3-2). Winter extreme precipitation has increased in part of the continent, in particular in central-western Europe and European Russia (Zolina et al., 2009), but the trend in summer precipitation has been weak or not spatially coherent (Moberg et al., 2006; Bartholy and Pongracz, 2007; Maraun et al., 2008; Pavan et al., 2008; Zolina et al., 2008; Costa and Soares, 2009; Kyselý, 2009; Durão et al., 2010; Rodda et al., 2010). Increasing trends in 90th, 95th, and 98th percentiles of daily winter precipitation over 1901-2000 were found (Moberg et al., 2006), which has been confirmed by more detailed country-based studies for the United Kingdom (Maraun et al., 2008), Germany (Zolina et al., 2008), and central and eastern Europe (Bartholy and Pongracz, 2007; Kyselý, 2009), while decreasing trends have been found in some regions such as northern Italy (Pavan et al., 2008), Poland (Lupikasza, 2010), and some Mediterranean coastal sites (Toreti et al., 2010). Uncertainties are overall larger in southern Europe and the Mediterranean region, where there is *low confidence* in the trends (Table 3-2). A recent study (Zolina et al., 2010) has indicated that there has been an increase of about 15 to 20% in the persistence of wet spells over most of Europe over the last 60 years, which was not associated with an increase of the total number of wet days.

There is *low* to *medium confidence* in trends in heavy precipitation in Asia, both at the continental and regional scale for most regions (Table 3-2; see also Alexander et al., 2006). A weak increase in the frequency of extreme precipitation events is observed in northern Mongolia (Nandintsetseg et al., 2007). No systematic spatially coherent trends in the frequency and duration of extreme precipitation events have been

found in Eastern and Southeast Asia (Choi et al., 2009), central and south Asia (Klein Tank et al., 2006), and Western Asia (X. Zhang et al., 2005; Rahimzadeh et al., 2009). However, statistically significant positive and negative trends were observed at subregional scales within these regions. Heavy precipitation increased in Japan during 1901-2004 (Fujibe et al., 2006), and in India (Rajeevan et al., 2008; Krishnamurthy et al., 2009) especially during the monsoon seasons (Sen Roy, 2009; Pattanaik and Rajeevan, 2010). Both statistically significant increases and decreases in extreme precipitation have been found in China over the period 1951-2000 (Zhai et al., 2005) and 1978-2002 (Yao et al., 2008). In Peninsular Malaysia during 1971-2005 the intensity of extreme precipitation increased and the frequency decreased, while the trend in the proportion of extreme rainfall over total precipitation was not statistically significant (Zin et al., 2009). Heavy precipitation increased over the southern and northern Tibetan Plateau but decreased in the central Tibetan Plateau during 1961-2005 (You et al., 2008).

In southern Australia, there has been a *likely* decrease in heavy precipitation in many areas, especially where mean precipitation has decreased (Table 3-2). There were statistically significant increases in the proportion of annual/seasonal rainfall stemming from heavy rain days from 1911-2008 and 1957-2008 in northwest Australia (Gallant and Karoly, 2010). Extreme summer rainfall over the northwest of the Swan-Avon River basin in western Australia increased over 1950-2003 while extreme winter rainfall over the southwest of the basin decreased (Aryal et al., 2009). In New Zealand, the trends are positive in the western North and South Islands and negative in the east of the country (Mullan et al., 2008).

There is *low* to *medium confidence* in regional trends in heavy precipitation in Africa due to partial lack of literature and data, and due to lack of consistency in reported patterns in some regions (Table 3-2). The AR4 (Trenberth et al., 2007) reported an increase in heavy precipitation over southern Africa, but this appears to depend on the region and precipitation index examined (Kruger, 2006; New et al., 2006; Seleshi and Camberlin, 2006; Aguilar et al., 2009). Central Africa exhibited a decrease in heavy precipitation over the last half century (Aguilar et al., 2009); however, data coverage for large parts of the region was poor. Precipitation from heavy events has decreased in western central Africa, but with low spatial coherence (Aguilar et al., 2009). Rainfall intensity averaged over southern and west Africa has increased (New et al., 2006). There is a lack of literature on changes in heavy precipitation in East Africa (Table 3-2). Camberlin et al. (2009) analyzed changes in components of rainy seasons' variability over the time period 1958-1987 in this region, but did not specifically address trends in heavy precipitation. There were decreasing trends in heavy precipitation over parts of Ethiopia during the period 1965-2002 (Seleshi and Camberlin, 2006).

Changes in hail occurrence are generally difficult to quantify because hail occurrence is not well captured by monitoring systems and because of historical data inhomogeneities. Sometimes, changes in environmental conditions conducive to hail occurrence are used to infer changes in hail

occurrence. However, the atmospheric conditions are typically estimated from reanalyses or from radiosonde data and the estimates are associated with high uncertainty. As a result, assessment of changes in hail frequency is difficult. For severe thunderstorms in the region east of the Rocky Mountains in the United States, Brooks and Dotzek (2008) found strong variability but no clear trend in the past 50 years. Cao (2008) identified a robust upward trend in hail frequency over Ontario, Canada. Kunz et al. (2009) found that both hail damage days and convective instability increased during 1974-2003 in a state in southwest Germany. Xie et al. (2008) identified no trend in the mean annual hail days in China from 1960 to the early 1980s but a statistically significant decreasing trend afterwards.

Causes of Observed Changes

The observed changes in heavy precipitation appear to be consistent with the expected response to anthropogenic forcing (increase due to enhanced moisture content in the atmosphere; see, e.g., Section 3.2.2.1) but a direct cause-and-effect relationship between changes in external forcing and extreme precipitation had not been established at the time of the AR4. As a result, the AR4 only concluded that it was *more likely than not* that anthropogenic influence had contributed to a global trend towards increases in the frequency of heavy precipitation events over the second half of the 20th century (Hegerl et al., 2007).

New research since the AR4 provides more evidence of anthropogenic influence on various aspects of the global hydrological cycle (Stott et al., 2010; see also Section 3.2.2), which is directly relevant to extreme precipitation changes. In particular, an anthropogenic influence on atmospheric moisture content is detectable (Santer et al., 2007; Willett et al., 2007; see also Section 3.2.2). Wang and Zhang (2008) show that winter season maximum daily precipitation in North America appears to be statistically significantly influenced by atmospheric moisture content, with an increase in moisture corresponding to an increase in maximum daily precipitation. This behavior has also been seen in model projections of extreme winter precipitation under global warming (Gutowski et al., 2008b). Climate model projections suggest that the thermodynamic constraint based on the Clausius-Clapeyron relation is a good predictor for extreme precipitation changes in a warmer world in regions where the nature of the ambient flows change little (Pall et al., 2007). This indicates that the observed increase in extreme precipitation in many regions is consistent with the expected extreme precipitation response to anthropogenic influences. However, the thermodynamic constraint may not be a good predictor in regions with circulation changes, such as mid- to higher latitudes (Meehl et al., 2005) and the tropics (Emori and Brown, 2005), and in arid regions. Additionally, changes in precipitation extremes with temperature also depend on changes in the moist-adiabatic temperature lapse rate, in the upward velocity, and in the temperature when precipitation extremes occur (O'Gorman and Schneider, 2009a,b; Sugiyama et al., 2010). This may explain why there have not been increases in precipitation extremes everywhere, although a low signal-to-noise ratio may also play a role. However, even in
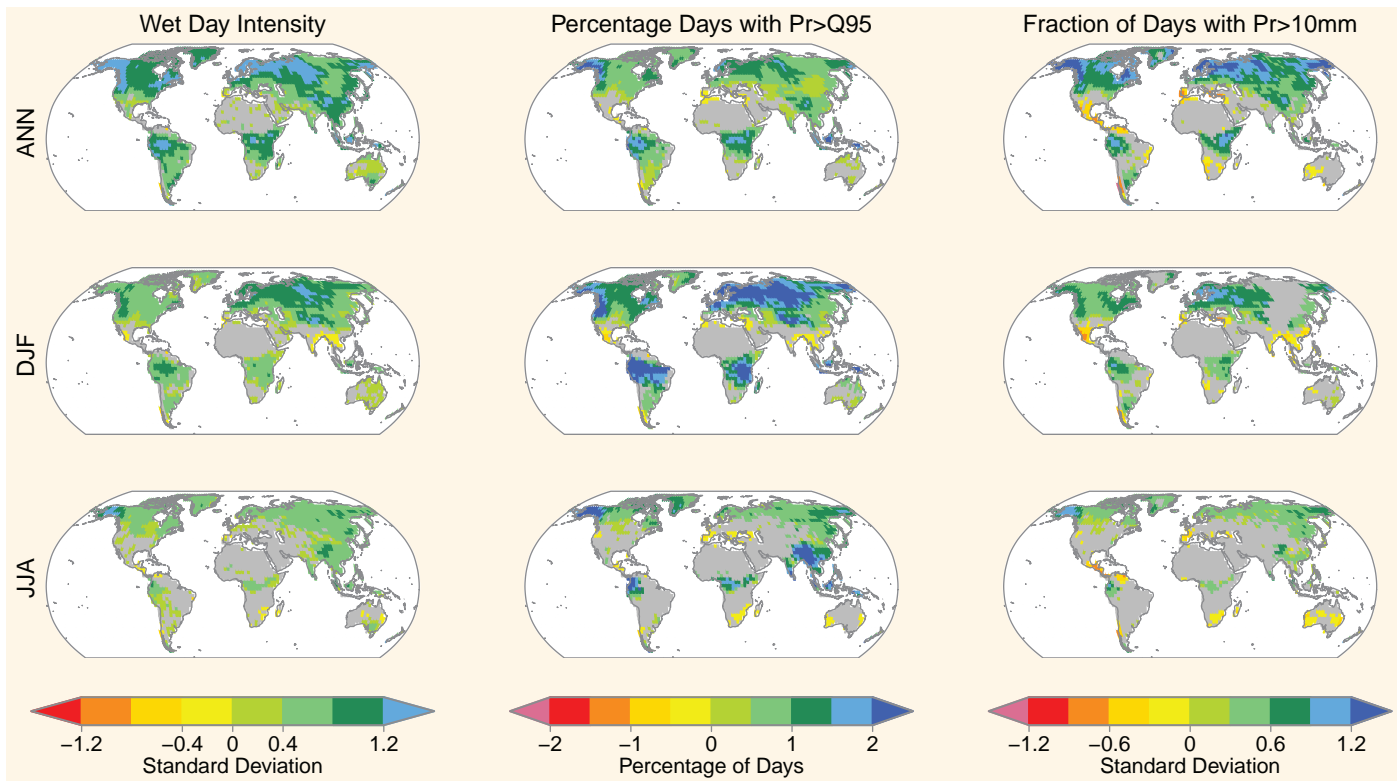
Figure 3-6 | Projected annual and seasonal changes in three indices for daily precipitation (Pr) for 2081-2100 with respect to 1980-1999, based on 17 GCMs contributing to the CMIP3. Left column: wet-day intensity; middle column: percentage of days with precipitation above the 95% quantile of daily wet day precipitation for that day of the year, calculated from the 1961-1990 reference period; right column: fraction of days with precipitation higher than 10 mm. The changes are computed for the annual time scale (top row) and two seasons (DJF, middle row, and JJA, bottom row) as the fractions/percentages in the 2081-2100 period (based on simulations under emission scenario SRES A2) minus the fractions/percentages of the 1980-1999 period (from corresponding simulations for the 20th century). Changes in wet-day intensity and in the fraction of days with Pr >10 mm are expressed in units of standard deviations, derived from detrended per year annual or seasonal estimates, respectively, from the three 20-year periods 1980-1999, 2046-2065, and 2081-2100 pooled together. Changes in percentages of days with precipitation above the 95% quantile are given directly as differences in percentage points. Color shading is only applied for areas where at least 66% (i.e., 12 out of 17) of the GCMs agree on the sign of the change; stippling is applied for regions where at least 90% (i.e., 16 out of 17) of the GCMs agree on the sign of the change. Adapted from Orlowsky and Seneviratne (2011); updating Tebaldi et al. (2006) for additional number of indices and CMIP3 models, and including seasonal time frames. For more details, see Appendix 3.A.

regions where the Clausius-Clapeyron constraint is not closely followed, it still appears to be a better predictor for future changes in extreme precipitation than the change in mean precipitation in climate model projections (Pall et al., 2007). An observational study seems also to support this thermodynamic theory. Analysis of daily precipitation from the Special Sensor Microwave Imager over the tropical oceans shows a direct link between rainfall extremes and temperature: heavy rainfall events increase during warm periods (El Niño) and decrease during cold periods (Allan and Soden, 2008). However, the observed amplification of rainfall extremes is larger than that predicted by climate models (Allan and Soden, 2008), due possibly to widely varying changes in upward velocities associated with precipitation extremes (O'Gorman and Schneider, 2008). Evidence from measurements in the Netherlands suggests that hourly precipitation extremes may in some cases increase 14% per degree of warming, which is twice as fast as what would be expected from the Clausius-Clapeyron relationship alone (Lenderink and Van Meijgaard, 2008), though this is still under debate (Haerter and Berg, 2009; Lenderink and van Meijgaard, 2009). A comparison between observed and multi-model simulated extreme precipitation using an optimal detection method suggests that the human-induced increase in greenhouse gases has contributed to the observed intensification of

heavy precipitation events over large Northern Hemisphere land areas during the latter half of the 20th century (Min et al., 2011). Pall et al. (2011) linked human influence on global warming patterns with an increased risk of England and Wales flooding in autumn (September-November) 2000 that is associated with a displacement in the North Atlantic jet stream. The present assessment based on evidence from new studies and those used in the AR4 is that there is *medium confidence* that anthropogenic influence has contributed to changes in extreme precipitation at the global scale. However, this conclusion may be dependent on the season and spatial scale. For example, there is now about a 50% chance that an anthropogenic influence can be detected in UK extreme precipitation in winter, but the likelihood of the detection in other seasons is very small (Fowler and Wilby, 2010).

Projected Changes and Uncertainties

Regarding projected changes in extreme precipitation, the AR4 concluded that it was *very likely* that heavy precipitation events, that is, the frequency of heavy precipitation or proportion of total precipitation from heavy precipitation, would increase over most areas of the globe
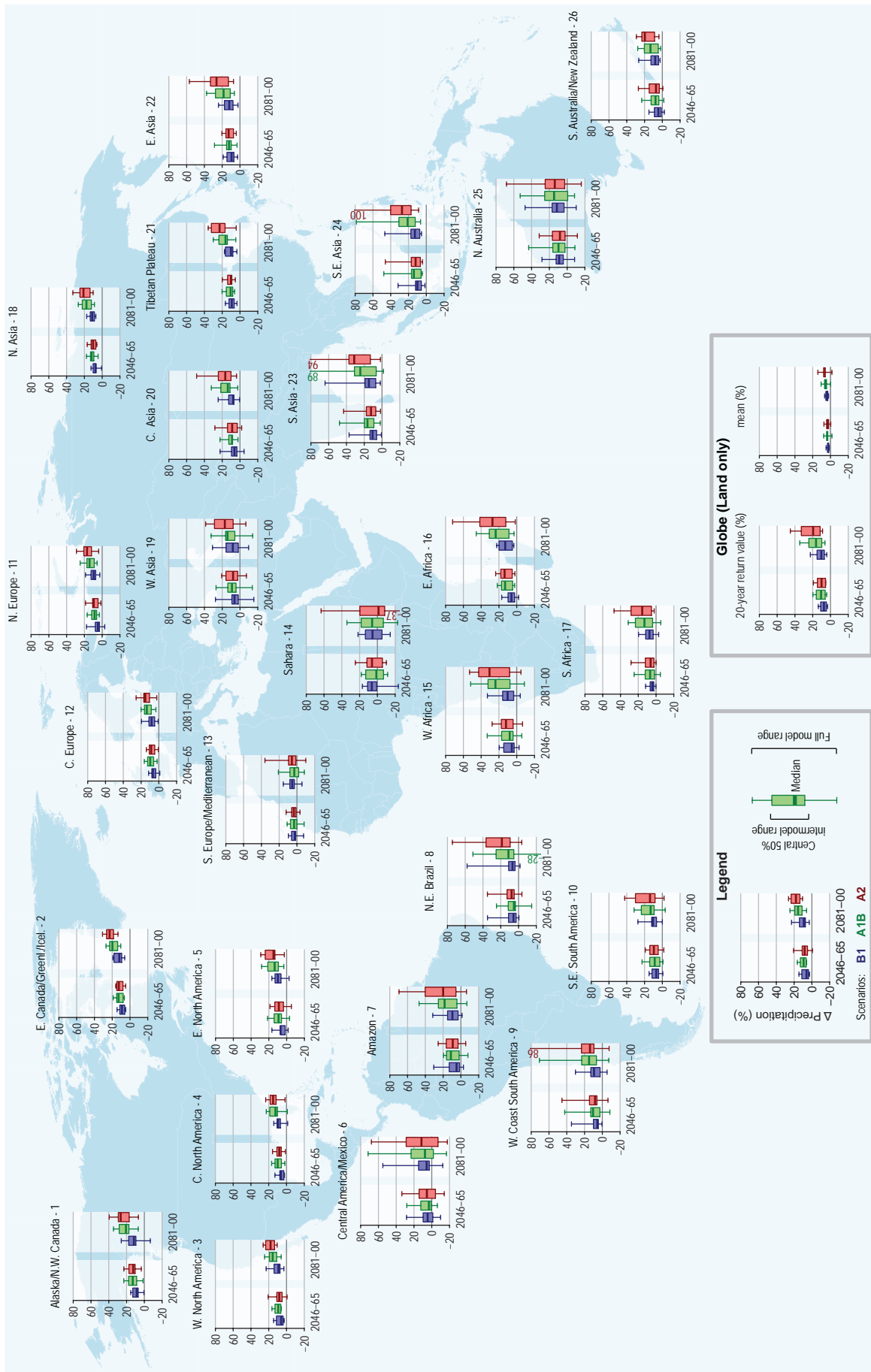
Figure 3-7a | Projected changes (%) in 20-year return values of annual maximum 24-hour precipitation rates. The bar plots (see legend for more information) show results for regionally averaged projections for two time horizons, 2046 to 2065 and 2081 to 2100, as compared to the late 20th century (1981-2000), and for three different SRES emission scenarios (B1, A1B, A2). Results are based on 14 GCMs contributing to the CMIP3. See Figure 3-1 for defined extent of regions. Values are computed for land points only. The 'Globe' analysis (inset box) displays the change in 20-year return values of the annual maximum 24-hour precipitation rates computed using all land grid points (left), and the change in annual mean 24-hour precipitation rates computed using all land grid points (right). Adapted from the analysis of Kharin et al. (2007). For more details, see Appendix 3.A.
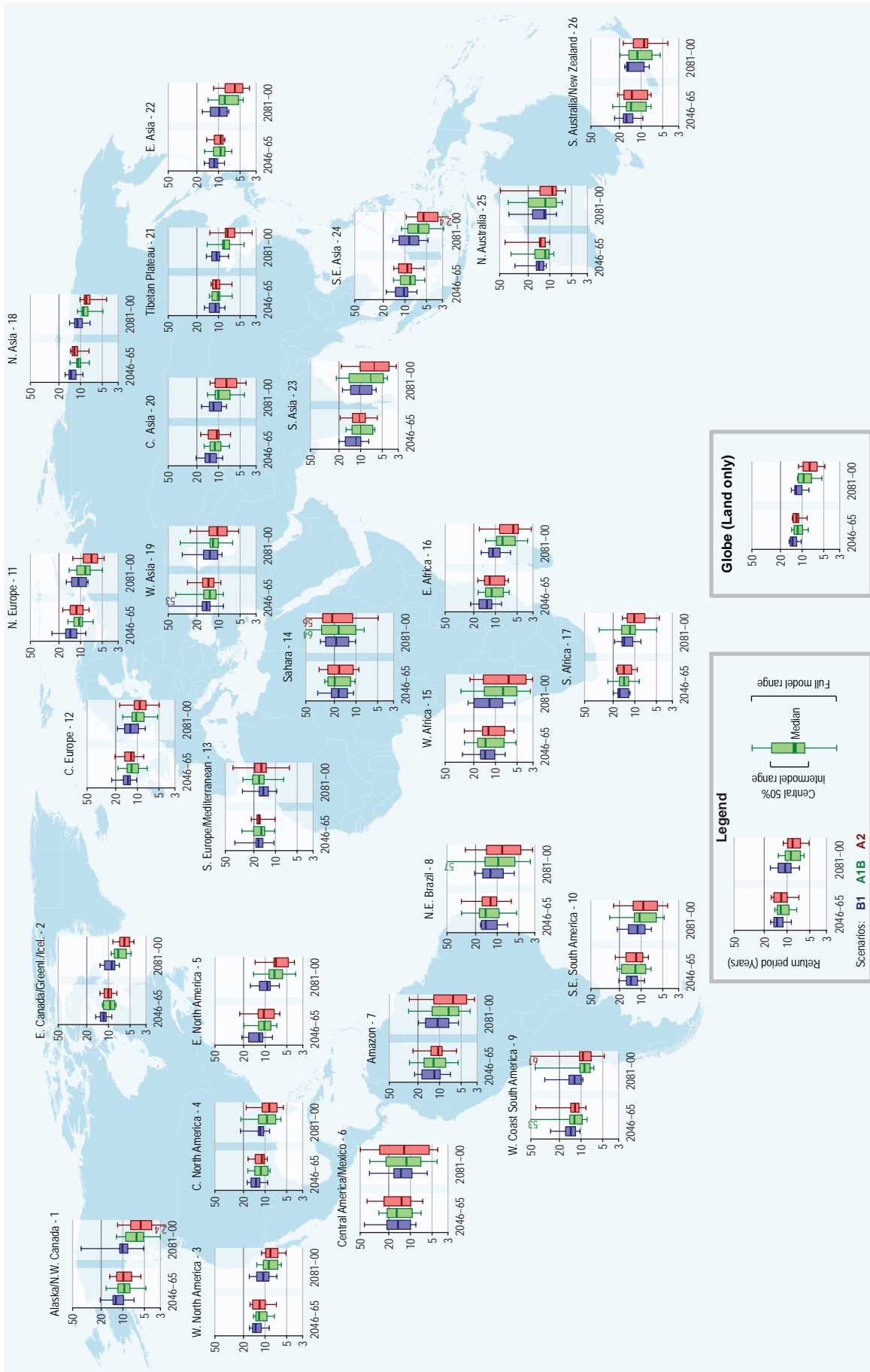
Figure 3-7b | Projected return period (in years) of late 20th century 20-year return values of annual maximum 24-hour precipitation rates. The bar plots (see legend for more information) show results for regionally averaged projections for two time horizons, 2046 to 2065 and 2081 to 2100, as compared to the late 20th century (1981–2000), and for three different SRES emission scenarios (B1, A1B, A2). Results are based on 14 GCMs contributing to the CMIP3. See Figure 3-1 for defined extent of regions. The 'Globe' analysis (inset box) displays the projected return period (in years) of late 20th-century 20-year return values of annual maximum 24-hour precipitation rates computed using all land grid points. Adapted from the analysis of Kharin et al. (2007). For more details, see Appendix 3.A.

in the 21st century (IPCC, 2007a). The tendency for an increase in heavy daily precipitation events was found in many regions, including some regions in which the total precipitation was projected to decrease.

Post-AR4 analyses of climate model simulations partly confirm this assessment but also highlight fairly large uncertainties and model biases in projections of changes in heavy precipitation in some regions (Section 3.2.3 and Table 3-3). On the other hand, more GCM and RCM ensembles have now been analyzed for some regions (Table 3-3; see also, e.g., Kharin et al., 2007; Kim et al., 2010). At the time of the AR4, Tebaldi et al. (2006) was the main global study available on projected changes in precipitation extremes (e.g., Figure 10.18 of Meehl et al., 2007b). Orlowsky and Seneviratne (2011) extended this analysis to a larger number of GCMs from the CMIP3 ensemble and for seasonal in addition to annual time frames (see also Section 3.3.1). Figure 3-6 provides corresponding analyses of projected annual and seasonal changes of the wet-day intensity, the fraction of days with precipitation above the 95% quantile of daily wet-day precipitation, and the fraction of days with precipitation above 10 mm day^{-1}. It should be noted that the 10 mm day^{-1} threshold cannot be considered extreme in several regions, but highlights differences in projections for absolute and relative thresholds (see also discussion in Box 3-1 and beginning of this section). Figure 3-6 indicates that regions with model agreement (at least 66%) with respect to changes in heavy precipitation are mostly found in the high latitudes and in the tropics, and in some mid-latitude regions of the Northern Hemisphere in the boreal winter. Regions with at least 90% model agreement are even more limited and confined to the high latitudes. Overall, model agreement in projected changes is found to be stronger in boreal winter (DJF) than summer (JJA) for most regions. Kharin et al. (2007) analyzed changes in annual maxima of 24-hour precipitation in the outputs of 14 CMIP3 models. Figure 3-7a displays the projected percentage change in the annual maximum of the 24-hour precipitation rate from the late 20th-century 20-year return values, while Figure 3-7b displays the corresponding projected return periods for late 20th-century 20-year return values of the annual maximum 24-hour precipitation rates in the mid-21st century (left) and in late 21st century (right) under three different emission scenarios (SRES B1, A1B, and A2). Between the late 20th and the late 21st century, the projected responses of extreme precipitation to future emissions show increased precipitation rates in most regions, and decreases in return periods in most regions in the high latitudes and the tropics and in some regions in the mid-latitudes consistent with projected changes in several indices related to heavy precipitation (see Figure 3-6 and Tebaldi et al., 2006), although there are increases in return periods or only small changes projected in several regions. Except for these regions, the return period for an event of annual maximum 24-hour precipitation with a 20-year return period in the late 20th century is projected to be about 5 to 15 years by the end of the 21st century. The greatest projected reductions in return period are in high latitudes and some tropical regions. The stronger CO_2 emissions scenarios (A1B and A2) lead to greater projected decreases in return period. In some regions with projected decreases in total precipitation (Christensen et al., 2007) such as southern Africa, west Asia, and the west coast of South America, heavy precipitation is

nevertheless projected to increase (Figure 3-7, Table 3-3). In some other areas with projected decreases in total precipitation (e.g., Central America and northern South America), however, heavy precipitation is projected to decrease or not change. It should be noted that Figure 3-7 addresses very extreme heavy precipitation events (those expected to occur about once in 20 years) whereas Figure 3-6 addresses less extreme, but still heavy, precipitation events. Projections of changes for these differently defined extreme events may differ.

Future precipitation projected by the CMIP3 models has also been analyzed in a number of studies for various regions using different combinations of the models (see next paragraphs and Table 3-3). In general these studies confirm the findings of global-scale studies by Tebaldi et al. (2006) and Kharin et al. (2007).

By analyzing simulations with a single GCM, Khon et al. (2007) reported a projected general increase in extreme precipitation for the different regions in northern Eurasia especially for winter. Su et al. (2009) found that for the Yangtze River Basin region in 2001-2050, the 50-year heavy precipitation events become more frequent, with return periods falling to below 25 years (relative to 1951-2000 behavior). For the Indian region, the Hadley Centre coupled model HadCM3 projects increases in the magnitude of the heaviest rainfall with a doubling of atmospheric CO_2 concentration (Turner and Slingo, 2009). Simulations by 12 GCMs projected an increase in heavy precipitation intensity and mean precipitation rates in east Africa, more severe precipitation deficits in the southwest of southern Africa, and enhanced precipitation further north in Zambia, Malawi, and northern Mozambique (Shongwe et al., 2009, 2011). Rocha et al. (2008) evaluated differences in the precipitation regime over southeastern Africa simulated by two GCMs under present (1961-1990) and future (2071-2100) conditions as a result of anthropogenic greenhouse gas forcing. They found that the intensity of all episode categories of precipitation events is projected to increase practically over the whole region, whereas the number of episodes is projected to decrease in most of the region and for most episode categories. Extreme precipitation is projected to increase over Australia in 2080-2099 relative to 1980-1999 in an analysis of the CMIP3 ensemble, although there are inconsistencies between projections from different models (Alexander and Arblaster, 2009).

High spatial resolution is important for studies of extreme precipitation because the physical processes responsible for extreme precipitation require high spatial resolution to resolve them (e.g., Kim et al., 2010). Post-AR4 studies have employed three approaches to obtain high spatial resolution to project precipitation extremes: high-resolution GCMs, dynamical downscaling using RCMs, and statistical downscaling (see also Section 3.2.3.1). Based on the Meteorological Research Institute and Japan Meteorological Agency 20-km horizontal grid GCM, heavy precipitation was projected to increase substantially in south Asia, the Amazon, and west Africa, with increased dry spell persistence projected in South Africa, southern Australia, and the Amazon at the end of the 21st century (Kamiguchi et al., 2006). In the Asian monsoon region, heavy precipitation was projected to increase, notably in Bangladesh

and in the Yangtze River basin due to the intensified convergence of water vapor flux in summer. Using statistical downscaling, Wang and Zhang (2008) investigated possible changes in North American extreme precipitation probability during winter from 1949-1999 to 2050-2099. Downscaled results suggested a strong increase in extreme precipitation over the south and central United States but decreases over the Canadian prairies. Projected European precipitation extremes in high-resolution studies tend to increase in northern Europe (Frei et al., 2006; Beniston et al., 2007; Schmidli et al., 2007), especially during winter (Haugen and Iversen, 2008; May, 2008), as also highlighted in Table 3-3. Fowler and Ekström (2009) project increases in both short-duration (1-day) and longer-duration (10-day) precipitation extremes across the United Kingdom during winter, spring, and autumn. In summer, model projections for the United Kingdom span the zero change line, although there is *low confidence* due to poor model performance in this season. Using daily statistics from various models, Boberg et al. (2009a,b) projected a clear increase in the contribution to total precipitation from more intense events together with a decrease in the number of days with light precipitation. This pattern of change was found to be robust for all European subregions. In double-nested model simulations with a horizontal grid spacing of 10 km, Tomassini and Jacob (2009) projected positive trends in extreme quantiles of heavy precipitation over Germany, although they are relatively small except for the high-CO_2 A2 emission scenario. For the Upper Mississippi River Basin region during October through March, the intensity of extreme precipitation is projected to increase (Gutowski et al., 2008b). Simulations with a single RCM project an increase in the intensity of extreme precipitation events over most of southeastern South America and western Amazonia in 2071-2100, whereas in northeast Brazil and eastern Amazonia smaller or no changes are projected (Marengo et al., 2009a). Outputs from another RCM indicate an increase in the magnitude of future extreme rainfall events in the Westernport region of Australia, consistent with results based on the CMIP3 ensemble (Alexander and Arblaster, 2009), and the size of this increase is greater in 2070 than in 2030 (Abbs and Rafter, 2008). When both future land use changes and increasing greenhouse gas concentrations are considered in the simulations, tropical and northern Africa are projected to experience less extreme rainfall events by 2025 during most seasons except for autumn (Paeth and Thamm, 2007). Simulations with high-resolution RCMs projected that the frequency of extreme precipitation increases in the warm climate for June through to September in Japan (Nakamura et al., 2008; Wakazuki et al., 2008; Kitoh et al., 2009). An increase in 90th-percentile values of daily precipitation on the Pacific side of the Japanese islands during July in the future climate was projected with a 5-km mesh cloud-system-resolving non-hydrostatic RCM (Kanada et al., 2010b).

Post-AR4 studies indicate that the projection of precipitation extremes is associated with large uncertainties, contributed by the uncertainties related to GCMs, RCMs, and statistical downscaling methods, and by natural variability of the climate. Kyselý and Beranova (2009) examined scenarios of change in extreme precipitation events in 24 future climate runs of 10 RCMs driven by two GCMs, focusing on a specific area of central Europe with complex orography. They demonstrated that the

inter- and intra-model variability and related uncertainties in the pattern and magnitude of the change are large, although they also show that the projected trends tend to agree with those recently observed in the area, which may strengthen their credibility. May (2008) reported an unrealistically large projected precipitation change over the Baltic Sea in summer in an RCM, apparently related to an unrealistic projection of Baltic Sea warming in the driving GCM. Frei et al. (2006) found large model differences in summer when RCM formulation contributes significantly to scenario uncertainty. In exploring the ability of two statistical downscaling models to reproduce the direction of the projected changes in indices of precipitation extremes, Hundecha and Bardossy (2008) concluded that the statistical downscaling models seem to be more reliable during seasons when local climate is determined by large-scale circulation than by local convective processes. Themeßl et al. (2011) merged linear and nonlinear empirical-statistical downscaling techniques with bias correction methods, and demonstrated their ability to drastically reduce RCM error characteristics. The extent to which the natural variability of the climate affects our ability to project the anthropogenically forced component of changes in daily precipitation extremes was investigated by Kendon et al. (2008). They show that annual to multidecadal natural variability across Europe may contribute to substantial uncertainty. Also, Kiktev et al. (2009) performed an objective comparison of climatologies and historical trends of temperature and precipitation extremes using observations and 20th-century climate simulations. They did not detect significant similarity between simulated and actual patterns of the indices of precipitation extremes in most cases. Moreover, Allan and Soden (2008) used satellite observations and model simulations to examine the response of tropical precipitation events to naturally driven changes in surface temperature and atmospheric moisture content. The observed amplification of rainfall extremes was larger than that predicted by models. The underestimation of rainfall extremes by the models may be related to the coarse spatial resolution used in the model simulations – the magnitude of changes in precipitation extremes depends on spatial resolution (Kitoh et al., 2009) – suggesting that projections of future changes in rainfall extremes in response to anthropogenic global warming may be underestimated.

Confidence is still *low* for hail projections particularly due to a lack of hail-specific modelling studies, and a lack of agreement among the few available studies. There is little information in the AR4 regarding projected changes in hail events, and there has been little new literature since the AR4. Leslie et al. (2008) used coupled climate model simulations under the SRES A1B scenario to estimate future changes in hailstorms in the Sydney Basin, Australia. Their future climate simulations show an increase in the frequency and intensity of hailstorms out to 2050, and they suggest that the increase will emerge from the natural background variability within just a few decades. This result offers a different conclusion from the modelling study of Niall and Walsh (2005), which simulated Convective Available Potential Energy (CAPE) for southeastern Australia in an environment containing double the pre-industrial concentrations of equivalent CO_2. They found a statistically significant projected decrease in CAPE values and concluded that "it is possible that there will be a decrease in the frequency of hail in southeastern

Australia if current rates of CO_2 emission are sustained," assuming the strong relationship between hail incidence and the CAPE for 1980-2001 remains unchanged under enhanced greenhouse conditions.

In summary, it is *likely* that there have been statistically significant increases in the number of heavy precipitation events (e.g., 95th percentile) in more regions than there have been statistically significant decreases, but there are strong regional and subregional variations in the trends (i.e., both between and within regions considered in this report; Figure 3-1 and Tables 3-2 and 3-3). In particular, many regions present statistically non-significant or negative trends, and, where seasonal changes have been assessed, there are also variations between seasons (e.g., more consistent trends in winter than in summer in Europe). The overall most consistent trends toward heavier precipitation events are found in North America (*likely* increase over the continent). There is *low confidence* in observed trends in phenomena such as hail because of historical data inhomogeneities and inadequacies in monitoring systems. Based on evidence from new studies and those used in the AR4, there is *medium confidence* that anthropogenic influence has contributed to intensification of extreme precipitation at the global scale. There is almost no literature on the attribution of changes in hail extremes, thus no assessment can be provided for these at this point in time. Projected changes from both global and regional studies indicate that it is *likely* that the frequency of heavy precipitation or proportion of total rainfall from heavy falls will increase in the 21st century over many areas on the globe, especially in the high latitudes and tropical regions, and northern mid-latitudes in winter. Heavy precipitation is projected to increase in some (but not all) regions with projected decreases of total precipitation (*medium confidence*). For a range of emission scenarios (A2, A1B, and B1), projections indicate that it is *likely* that a 1-in-20 year annual maximum 24-hour precipitation rate will become a 1-in-5 to -15 year event by the end of 21st century in many regions. Nevertheless, increases or statistically non-significant changes in return periods are projected in some regions.

3.3.3.   Wind

Extreme wind speeds pose a threat to human safety, maritime and aviation activities, and the integrity of infrastructure. As well as extreme wind speeds, other attributes of wind can cause extreme impacts. Trends in average wind speed can influence potential evaporation and in turn water availability and droughts (e.g., McVicar et al., 2008; see also Section 3.5.1 and Box 3-3). Sustained mid-latitude winds can elevate coastal sea levels (e.g., McInnes et al., 2009b), while longer-term changes in prevailing wind direction can cause changes in wave climate and coastline stability (Pirazzoli and Tomasin, 2003; see also Sections 3.5.4 and 3.5.5). Aeolian processes exert significant influence on the formation and evolution of arid and semi-arid environments, being strongly linked to soil and vegetation change (Okin et al., 2006). A rapid shift in wind direction may reposition the leading edge of a forest fire

(see Section 4.2.2.2; Mills, 2005) while the fire itself may generate a local circulation response such as tornado genesis (e.g., Cunningham and Reeder, 2009). Unlike other weather and climate elements such as temperature and rainfall, extreme winds are often considered in the context of the extreme phenomena with which they are associated such as tropical and extratropical cyclones (see also Sections 3.4.4 and 3.4.5), thunderstorm downbursts, and tornadoes. Although wind is often not used to define the extreme event itself (Peterson et al., 2008b), wind speed thresholds may be used to characterize the severity of the phenomenon (e.g., the Saffir-Simpson scale for tropical cyclones). Changes in wind extremes may arise from changes in the intensity or location of their associated phenomena (e.g., a change in local convective activity) or from other changes in the climate system such as the movement of large-scale circulation patterns. Wind extremes may be defined by a range of quantities such as high percentiles, maxima over a particular time scale (e.g., daily to yearly), or storm-related highest values. Wind gusts, which are a measure of the highest winds in a short time interval (typically 3 seconds), may be evaluated in models using gust parameterizations that are applied to the maximum daily near-surface wind speed (e.g., Rockel and Woth, 2007).

Over paleoclimatic time scales, proxy data have been used to infer circulation changes across the globe from the mid-Holocene (~6000 years ago) to the beginning of the industrial revolution (Wanner et al., 2008). Over this period, there is evidence for changes in circulation patterns across the globe. The Inter-Tropical Convergence Zone (ITCZ) moved southward, leading to weaker monsoons across Asia (Haug et al., 2001). The Walker circulation strengthened and Southern Ocean westerlies moved northward and strengthened, affecting southern Australia, New Zealand, and southern South America (Shulmeister et al., 2006; Wanner et al., 2008), and an increase in ENSO variability and frequency occurred (Rein et al., 2005; Wanner et al., 2008). There is also weaker evidence for a change toward a lower Northern Atlantic Oscillation (NAO), implying weaker westerly winds over the north Atlantic (Wanner et al., 2008). While the changes in the Northern Hemisphere were attributed to changes in orbital forcing, those in the Southern Hemisphere were more complex, possibly reflecting the additional role on circulation of heat transport in the ocean. Solar variability and volcanic eruptions may also have contributed to decadal to multi-centennial fluctuations over this time period (Wanner et al., 2008).

The AR4 did not specifically address changes in extreme wind although it did report on wind changes in the context of other phenomena such as tropical and extratropical cyclones and oceanic waves and concluded that mid-latitude westerlies had increased in strength in both hemispheres (Trenberth et al., 2007). Direct investigation of changes in wind climatology has been hampered by the sparseness of long-term, high-quality wind measurements from terrestrial anemometers arising from the influence of changes in instrumentation, station location, and surrounding land use (e.g., Cherry, 1988; Pryor et al., 2007; Jakob, 2010; see also Section 3.2.1). Nevertheless, a number of recent studies report trends in mean and extreme wind speeds in different parts of the world based on wind observations and reanalyses.

Over North America, a declining trend in 50th and 90th percentile wind speeds has been reported for much of the United States over 1973 to 2005 (Pryor et al., 2007) and in 10-m hourly wind data over 1953-2006 over western and most of southern Canada (Wan et al., 2010). An increasing trend has been reported in average winds over Alaska over 1955-2001 by Lynch et al. (2004) and over the central Canadian Arctic in all seasons and in the Maritimes in spring and autumn by Wan et al. (2010) as well as in annual maximum winds in a regional reanalysis over the southern Maritimes from 1979-2003 (Hundecha et al., 2008). Over China, negative trends have been reported in 10-m monthly mean and 95th percentile winds over 1969-2005 (Guo et al., 2011), in daily maximum wind speeds over 1956-2004 by Jiang et al. (2010a), and in 2-m average winds over the Tibetan plateau from 1966-2003 (Y. Zhang et al., 2007), confirming earlier declining trends in mean and strong 10-m winds reported by Xu et al. (2006). Over Europe, Smits et al. (2005) found declining trends in extreme winds (those occurring on average 10 and 2 times per year) in 10-m anemometer data over 1962-2002. Pirazolli and Tomasin (2003) reported a generally declining trend in both annual mean and annual maximum winds from 1951 to the mid-1970s and an increasing trend since then, from observations in the central Mediterranean region. Similar to the mostly declining trends found in Northern Hemisphere studies of surface wind observations, Vautard et al. (2010) also found mostly declining trends in surface wind observations across the continental northern mid-latitudes and a stronger decline in extreme winds compared to mean winds in surface wind measurements. In the Southern Hemisphere, McVicar et al. (2008) reported declines in 2-m mean wind speed over 88% of Australia (significant over 57% of the country) over 1975-2006 and positive trends over about 12% of the mainland interior and southern and eastern coastal regions including Tasmania. In Antarctica, increasing trends in mean wind speeds have been reported over the second half of the 20th century (Turner et al., 2005). With the exception of the robust declines in wind reported over China, studies in most areas are too few in number to draw robust conclusions on wind speed change and even fewer studies have addressed extreme wind change. Some studies report opposite trends between anemometer winds and reanalysis data sets in some areas (Smits et al., 2005; McVicar et al., 2008; Vautard et al., 2010); however, comparisons of surface anemometer data at 10 m or lower with reanalysis-derived 10-m data that do not resolve complex surface features is problematic.

Trends in extreme winds have also been inferred from trends in particular phenomena. With regards to tropical cyclones (Section 3.4.4.), no statistically significant trends have been detected in the overall global annual number although a trend has been reported in the intensity of the strongest storms since 1980 [but there is *low confidence* that any observed long-term (i.e., 40 years or more) increases in tropical cyclone activity are robust, after accounting for past changes in observing capabilities; see Section 3.4.4]. In the mid-latitudes, studies have used proxies for wind such as pressure tendencies or geostrophic winds calculated from triangles of pressure (geo-winds) over Europe (e.g., Barring and von Storch, 2004; Matulla et al., 2008; Allan et al., 2009; Barring and Fortuniak, 2009; X.L. Wang et al., 2009b) and Australia (e.g.,

Alexander and Power, 2009; Alexander et al., 2011). For Europe, these studies suggest that storm activity was higher around 1900 and in the 1990s and lower in the 1960s and 1970s, although X.L. Wang et al. (2009b) note that seasonal trends behave differently than annual trends. In general, long-term trends differ between the different available studies as well as studies that focus on the period for which reanalysis data exist (e.g., Raible, 2007; Leckebusch et al., 2008; Della-Marta et al., 2009; Nissen et al., 2010), and strong inter-decadal variability is also often reported (e.g., Allan et al., 2009; X.L. Wang et al., 2009b; Nissen et al., 2010). Over southeast Australia, a decline in storm activity since around 1885 has been reported (Alexander and Power, 2009; Alexander et al., 2011). See Section 3.4.5 for more discussion of extratropical cyclones. Regarding other phenomena associated with extreme winds, such as thunderstorms, tornadoes, and mesoscale convective complexes, studies are too few in number to assess the effect of their changes on extreme winds. As well, historical data inhomogeneities mean that there is *low confidence* in any observed trends in these small-scale phenomena.

The AR4 reported for the mid-latitudes that trends in the Northern and Southern Annular Modes, which correspond to sea level pressure reductions over the poles, are *likely* related in part to human activity, and this in turn has affected storm tracks and wind patterns in both hemispheres (Hegerl et al., 2007). The relationship between mean and severe winds and natural modes of variability has been investigated in several post-AR4 studies. On the Canadian west coast, Abeysirigunawardena et al. (2009) found that higher extreme winds tend to occur during the negative (i.e., cold) ENSO phase. The generally increasing trend in mean wind speeds over recent decades in Antarctica is consistent with the change in the nature of the Southern Annular Mode toward its high index state (Turner et al., 2005). Donat et al. (2010b) concluded that 80% of storm days in central Europe are connected with westerly flows that occur primarily during the positive phase of the NAO. Declining trends in wind over China have mainly been linked to circulation changes due to a weaker land-sea thermal contrast (Xu et al., 2006; Jiang et al., 2010a; Guo et al., 2011). Vautard et al. (2010) attribute the slowdown in mid to high percentiles of surface winds over most of the continental northern mid-latitudes to changes in atmospheric circulation (10-50%) and an increase in surface roughness due to biomass increases (25-60%), which are supported by RCM simulations. X.L. Wang et al. (2009a) formally detected a link between external forcing and positive trends in the high northern latitudes and negative trends in the northern mid-latitudes using a proxy for wind (geostrophic wind energy) in the boreal winter. Trends in mean and annual maximum winds in the central Mediterranean region were found to be positively correlated with temperature but not with the NAO index (Pirazzoli and Tomasin, 2003). Nissen et al. (2010) used cyclone tracking to identify associated strong winds in reanalysis data from 1957 to 2002 and found a positive trend in the central Mediterranean region and southern Europe and a negative trend over the western Mediterranean region.

Projections of wind speed changes and particularly wind extremes were not specifically addressed in the AR4 although references to wind speed were made in relation to other variables and phenomena such as
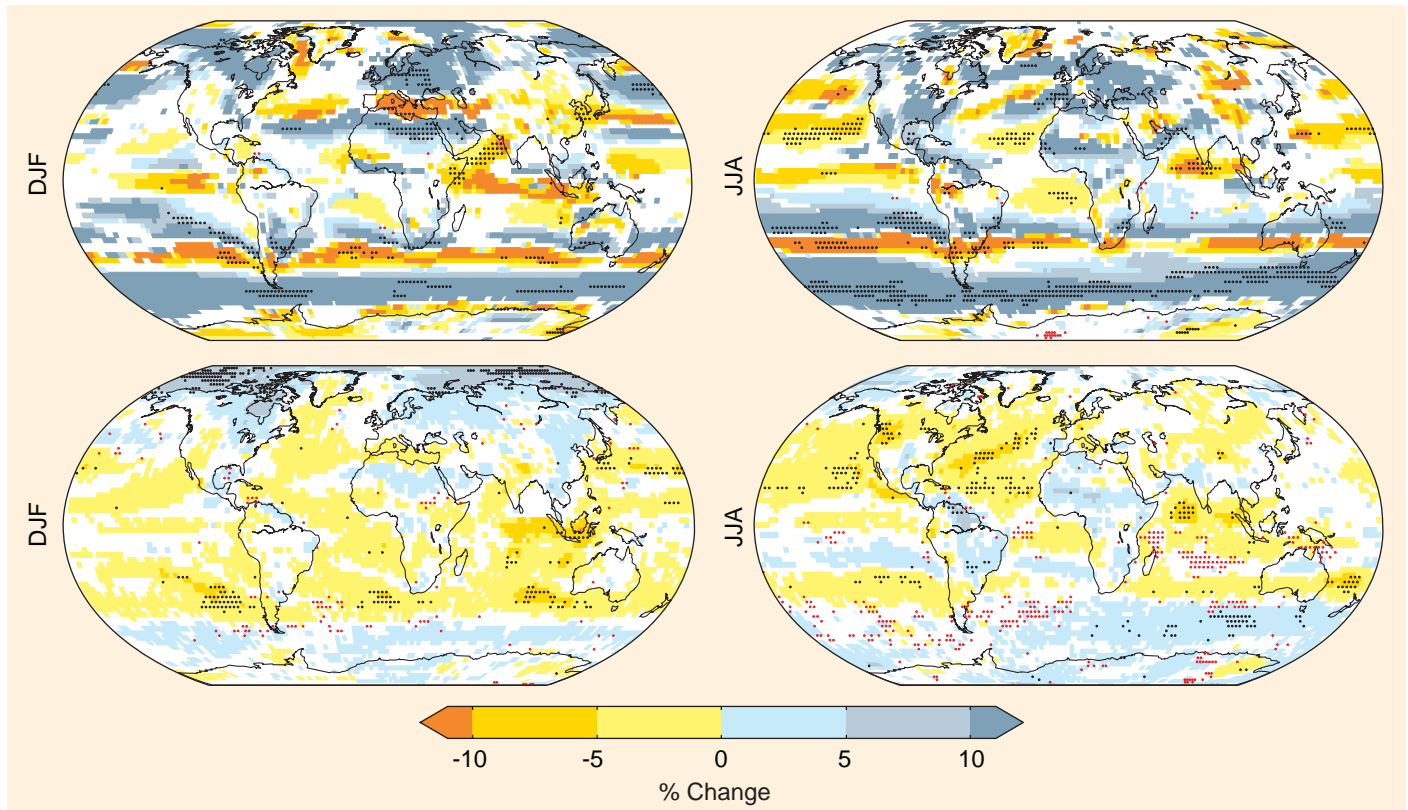
Figure 3-8 | Averaged changes from a 19-member ensemble of CMIP3 GCMs in the mean of the daily averaged 10-m wind speeds (top) and 99th percentile of the daily averaged 10-m wind speeds (bottom) for the period 2081-2100 relative to 1981-2000 (% change) for December to February (left) and June to August (right) plotted only where more than 66% of the models agree on the sign of the change. Black stippling indicates areas where more than 90% of the models agree on the sign of the change. Red stippling indicates areas where more than 66% of models agree on a small change between ±2%. Adapted from McInnes et al. (2011); for more details see Appendix 3.A.

mid-latitude storm tracks, tropical cyclones, and ocean waves (Christensen et al., 2007; Meehl et al., 2007b). Meehl et al. (2007b) projected a *likely* increase in tropical cyclone extreme winds in the future and provided more evidence for a projected poleward shift of the storm tracks and associated changes in wind patterns. Since the AR4, new studies have focused on future changes in winds. Gastineau and Soden (2009) reported a decrease in 99th-percentile winds at 850 hPa in the tropics and an increase in the extratropics in a 17-member multi-model ensemble over 2081-2100 relative to 1981-2000. McInnes et al. (2011) presented spatial maps of multi-model agreement in mean and 99th-percentile 10-m wind change between 1981-2000 and 2081-2100 in a 19-member ensemble (see Figure 3-8). These show an increase in mean winds over Europe, parts of Central and North America, the tropical South Pacific, and the Southern Ocean. Mean wind speed declines occur along the equator reflecting a slowdown in the Walker circulation (Collins et al., 2010) (and in the vicinity of the subtropical ridge in both hemispheres which, together with the strengthening of winds further poleward, reflect the contraction toward the poles of the mid-latitude storm tracks; see Section 3.4.5). Seasonal differences are also apparent with more extensive mean wind increases in the Arctic and parts of the northern Pacific in DJF and decreases over most of the northern Pacific in JJA. The 99th-percentile wind changes show declines over most ocean areas except the northern Pacific and Arctic and Southern Ocean south of 40°S in DJF, the south Pacific between about 10 and 25°S in JJA, and the Southern Ocean south of 50°S in JJA. Increases in 99th-percentile

winds occur over the Arctic and large parts of the continental area in the Northern Hemisphere in DJF and in Africa, northern Australia, and Central and South America in JJA. Despite the projections displayed in Figure 3-8, the relatively few studies of projected extreme winds, combined with shortcomings in the simulation of extreme winds and the different models, regions, and methods used to develop projections of this quantity, means that we have *low confidence* in projections of changes in strong winds.

Regional increases in winter wind storm risk over Europe due to changes in storm tracks are also supported by a number of regional studies (e.g., Pinto et al., 2007b; Debernard and Roed, 2008; Leckebusch et al., 2008; Sterl et al., 2009; Donat et al., 2010a,b, 2011). However, GCMs at their current resolution are unable to resolve small-scale phenomena such as tropical cyclones, tornadoes, and mesoscale convective complexes that are associated with particularly severe winds, although as noted by McInnes et al. (2011) these winds would typically be more extreme than 99th percentile. There is evidence to suggest an increase in extreme winds from tropical cyclones in the future (see Section 3.4.4). An increase in atmospheric greenhouse gas concentrations may cause some of the atmospheric conditions conducive to tornadoes such as atmospheric instability to increase due to increasing temperature and humidity, while others such as vertical shear to decrease due to a reduced pole-to-equator temperature gradient (Diffenbaugh et al., 2008), but the literature on these phenomena is extremely limited at

this time. There is thus *low confidence* in projections of changes in such small-scale systems because of limited studies, inability of climate models to resolve these phenomena, and possible competing factors affecting future changes. Confidence in the extreme wind changes is therefore lower in the regions most influenced by these phenomena irrespective of whether there is high agreement between GCMs on the sign of the wind speed change.

In addition to studies using GCMs there have also been several recent studies employing RCMs. Those focusing on Europe (e.g., Beniston et al., 2007; Rockel and Woth, 2007; Haugen and Iversen, 2008; Rauthe et al., 2010) also provide a general picture of an increasing trend in extreme winds over northern Europe despite a range of different downscaling models used, the different GCMs in which the downscaling is undertaken, and different metrics used to quantify extreme winds. Small-scale polar lows that typically form north of 60°N have been found to decline in frequency in RCM simulations downscaled from a GCM under different emission scenarios and this is related to greater stability over the region due to mid-troposphere temperatures warming faster than sea surface temperatures over the region (Zahn and von Storch, 2010). In other parts of the world there have been very few studies. Over China, Jiang et al. (2010b) projected decreases in annual and winter mean wind speed based on two RCMs that downscale two different GCMs. Over North America, statistical downscaling of winds from four GCMs over five airports in the northwest United States indicated declines in summer wind speeds and less certain changes in winter (Sailor et al., 2008).

A number of recent studies have addressed observed changes in wind speed across different parts of the globe, but due to the various shortcomings associated with anemometer data and the inconsistency in anemometer and reanalysis trends in some regions, we have *low confidence* in wind trends and their causes at this stage. We also have *low confidence* in how the observed trends in mean wind speed relate to trends in extreme winds. The few studies of projected extreme winds, combined with shortcomings in the simulation of extreme winds and the different models, regions, and methods used to develop projections of this quantity, mean that we have *low confidence* in projections of changes in extreme winds (with the exception of changes associated with tropical cyclones; Section 3.4.4). There is *low confidence* in projections of small-scale phenomena such as tornadoes because competing physical processes may affect future trends and because climate models do not simulate such phenomena.

3.4. Observed and Projected Changes in Phenomena Related to Weather and Climate Extremes

3.4.1. Monsoons

Changes in monsoon-related extreme precipitation and winds due to climate change are not well understood. Generally, precipitation is the most important variable, but it is also a variable associated with larger uncertainties in climate simulations and projections (Wang et al., 2005; Kang and Shukla, 2006). Changes in monsoons should be better depicted by large-scale dynamics, circulation, or moisture convergence more broadly than via precipitation only. However, few studies have focused on observed changes in the large-scale and regional monsoon circulations. Hence, in this section, we focus mostly on monsoon-induced changes in total and seasonal rainfall, with most discussions of intense rainfall covered in Section 3.3.2.

Modeling experiments to assess paleo-monsoons suggest that in the past, during the Holocene due to orbital forcing on a millennial time scale, there was a progressive southward shift of the Northern Hemisphere summer position of the ITCZ around 8,000 years ago. This was accompanied by a pronounced weakening of the monsoon rainfall systems in Africa and Asia and increasing dryness on both continents, while in South America the monsoon was weaker and drier than in the present, as suggested both by models and paleoclimatic indicators (Wanner et al., 2008).

The delineation of the global monsoon has been mostly performed using rainfall data or outgoing longwave radiation (OLR) fields (Kim et al., 2008). Lau and Wu (2007) identified two opposite time evolutions in the occurrence of rainfall events in the tropics: a negative trend in moderate rain events and a positive trend in heavy and light rain events. Positive trends in intense rain were located in deep convective cores of the ITCZ, South Pacific Convergence Zone, Indian Ocean, and monsoon regions.

In the Indo-Pacific region, covering the southeast Asian and north Australian monsoon, Caesar et al. (2011) found low spatial coherence in trends in precipitation extremes across the region between 1971 and 2003. In the few cases where statistically significant trends in precipitation extremes were identified, there was generally a trend towards wetter conditions, in common with the global results of Alexander et al. (2006). Liu et al. (2011) reported a decline in recorded precipitation events in China over 1960-2000, which was mainly accounted for by a decrease in light precipitation events, with intensities of 0.1-0.3 mm day^{-1}. Some of the extreme precipitation appeared to be positively correlated with a La Niña-like sea surface temperature (SST) pattern, but without suggesting the presence of a trend. With regard to wind changes, Guo et al. (2011) analyzed near-surface wind speed change in China and its monsoon regions from 1969 to 2005 and showed a statistically significant weakening in annual and seasonal mean wind.

For the Indian monsoon, Rajeevan et al. (2008) showed that extreme rain events have an increasing trend between 1901 and 2005, but the trend is much stronger after 1950. Sen Roy (2009) investigated changes in extreme hourly rainfall in India, and found widespread increases in heavy precipitation events across India, mostly in the high-elevation regions of the northwestern Himalaya as well as along the foothills of the Himalaya extending south into the Indo-Ganges basin, and particularly during the summer monsoon season during 1980-2002.

In the African monsoon region, Fontaine et al. (2011) investigated recent observed trends using high-resolution gridded precipitation (period 1979-2002), OLR, and reanalyses. Their results revealed a rainfall increase in North Africa since the mid-1990s. Over the longer term, however, Zhou et al. (2008a,b) and Wang and Ding (2006) reported an overall decreasing long-term trend in global land monsoon rainfall during the last 54 years, which was mainly caused by decreasing rainfall in the North African and South Asian monsoons.

For the American monsoon regions, Cavazos et al. (2008) reported increases in the intensity of precipitation in the mountain sites of the northwestern Mexico section of the North American monsoon over the 1961-1998 period, apparently related to an increased contribution from heavy precipitation derived from tropical cyclones. Arriaga-Ramírez and Cavazos (2010) found that total and extreme rainfall in the monsoon region of western Mexico and the US southwest presented a statistically significant increase during 1961-1998, mainly in winter. Groisman and Knight (2008) found that consecutive dry days (see Box 3-3 for definition) have significantly increased in the US southwest. On the other hand, increases in heavy precipitation during 1960-2000 in the South American monsoon have been documented by Marengo et al. (2009a,b) and Rusticucci et al. (2010). Studies using circulation fields such as 850 hPa winds or moisture flux have been performed for the South American monsoon system for assessments of the onset and end of the monsoon, and indicate that the onset exhibits a marked interannual variability linked to variations in SST anomalies in the eastern Pacific and tropical Atlantic (Gan et al., 2006; da Silva and de Carvalho, 2007; Raia and Cavalcanti, 2008; Nieto-Ferreira and Rickenbach, 2011).

Attributing the causes of changes in monsoons is difficult in part because there are substantial inter-model differences in representing Asian monsoon processes (Christensen et al., 2007). Most models simulate the general migration of seasonal tropical rain, although the observed maximum rainfall during the monsoon season along the west coast of India, the North Bay of Bengal, and adjoining northeast India is poorly simulated by many models due to limited resolution. Bollasina and Nigam (2009) show the presence of large systematic biases in coupled simulations of boreal summer precipitation, evaporation, and SST in the Indian Ocean. Many of the biases are pervasive, being common to most simulations.

The observed negative trend in global land monsoon rainfall is better reproduced by atmospheric models forced by observed historical SST than by coupled models without explicit forcing by observed ocean temperatures (Kim et al., 2008). This trend in the east Asian monsoon is strongly linked to the warming trend over the central eastern Pacific and the western tropical Indian Ocean (Zhou et al., 2008b). For the west African monsoon, Joly and Voldoire (2010) explore the role of Gulf of Guinea SSTs in its interannual variability. In most of the studied CMIP3 simulations, the interannual variability of SST is very weak in the Gulf of Guinea, especially along the Guinean Coast. As a consequence, the influence on the monsoon rainfall over the African continent is poorly reproduced. It is suggested that this may be due to the counteracting

effects of the Pacific and Atlantic basins over the last decades. The decreasing long-term trend in north African summer monsoon rainfall may be due to the atmosphere response to observed SST variations (Hoerling et al., 2006; Zhou et al., 2008b; Scaife et al., 2009). A similar trend in global monsoon precipitation in land regions is reproduced in CMIP3 models' 20th-century simulations when they include anthropogenic forcing, and for some simulations natural forcing (including volcanic forcing) as well, though the trend is much weaker in general, with the exception of one model (HadCM3) capable of producing a trend of similar magnitude (Li et al., 2008). The decrease in east Asian monsoon rainfall also seems to be related to tropical SST changes (Li et al., 2008), and the less spatially coherent positive trends in precipitation extremes in the southeast Asian and north Australian monsoons appear to be positively correlated with a La Niña-like SST pattern (Caesar et al., 2011).

A variety of factors, natural and anthropogenic, have been suggested as possible causes of variations in monsoons. Changes in regional monsoons are strongly influenced by the changes in the states of dominant patterns of climate variability such as ENSO, the Pacific Decadal Oscillation (PDO), the Northern Annular Mode (NAM), the Atlantic Multi-decadal Oscillation (AMO), and the Southern Annular Mode (SAM) (see also Sections 3.4.2 and 3.4.3). Additionally, model-based evidence has suggested that land surface processes and land use changes could in some instances significantly impact regional monsoons. Tropical land cover change in Africa and southeast Asia appears to have weaker local climatic impacts than in Amazonia (Voldoire and Royer, 2004; Mabuchi et al., 2005a,b). Grimm et al. (2007) and Collini et al. (2008) explored possible feedbacks between soil moisture and precipitation during the early stages of the monsoon in South America, when the surface is not sufficiently wet, and soil moisture anomalies may thus also modulate the development of precipitation. However, the influence of historical land use on the monsoon is difficult to quantify, due both to the poor documentation of land use and difficulties in simulating the monsoon at fine scales. The impact of aerosols (black carbon and sulfate) on changes in rainfall variability and amounts in monsoon regions has been discussed by Meehl et al. (2008), Lau et al. (2006), and Silva Dias et al. (2002). These studies suggest that there are still large uncertainties and a strong model dependency in the representation of the relevant land surface processes and the role of aerosol direct forcing, and resulting interactions (e.g., in the case of land use forcing; Pitman et al., 2009).

Regarding projections of change in the monsoons, the AR4 (Christensen et al., 2007) concluded: "There is a tendency for monsoonal circulations to result in increased precipitation due to enhanced moisture convergence, despite a tendency towards weakening of the monsoonal flows themselves. However, many aspects of tropical climatic responses remain uncertain." Held and Soden (2006) demonstrate that an increase in the hydrological cycle is accompanied by a global weakening of the large-scale circulation. As global warming is projected to lead to faster warming over land than over the oceans (e.g., Meehl et al., 2007b; Sutton et al., 2007), the continental-scale land-sea thermal contrast, a major factor affecting monsoon circulations, will become stronger in summer. Based on this projection, a simple scenario is that the summer

monsoon will be stronger and the winter monsoon will be weaker in the future than now. However, model results derived from the analyses of 15 CMIP3 global models are not as straightforward as implied by this simple consideration (Tanaka et al., 2005), as they show a weakening of these tropical circulations by the late 21st century compared to the late 20th century. In turn, such changes in circulation may lead to changes in precipitation associated with monsoons. For instance, the monsoonal precipitation in Mexico and Central America is projected to decrease in association with increasing precipitation over the eastern equatorial Pacific through changes in the Walker circulation and local Hadley circulation (e.g., Lu et al., 2007). Furthermore, observations and models suggest that changes in monsoons are related at least in part to changes in observed SSTs, as noted above.

At regional scales, there is little consensus in GCM projections regarding the sign of future change in monsoon characteristics, such as circulation and rainfall. For instance, while some models project an intense drying of the Sahel under a global warming scenario, others project an intensification of the rains, and some project more frequent extreme events (Cook and Vizy, 2006). Increases in precipitation are projected in the Asian monsoon (along with an increase in interannual season-averaged precipitation variability), and in the southern part of the west African monsoon, but with some decreases in the Sahel in northern summer. In the Australian monsoon in southern summer, an analysis by Moise and Colman (2009) from the entire ensemble mean of CMIP3 simulations suggested no changes in Australian tropical rainfall during the summer and only slightly enhanced interannual variability.

A study of 19 CMIP3 global models reported a projected increase in mean south Asian summer monsoon precipitation of 8% and a possible extension of the monsoon period (Kripalani et al., 2007). A study (Ashfaq et al., 2009) from the downscaling of the National Center for Atmospheric Research (NCAR) CCSM3 global model using the RegCM3 regional model suggests a weakening of the large-scale monsoon flow and suppression of the dominant intra-seasonal oscillatory modes with overall weakening of the south Asian summer monsoon by the end of the 21st century, resulting in a decrease in summer precipitation in southeast Asia.

Kitoh and Uchiyama (2006) used 15 models under the A1B scenario to analyze the changes in intensity and duration of precipitation in the Baiu-Changma-Meiyu rain band at the end of the 21st century. They found a delay in early summer rain withdrawal over the region extending from the Taiwan province of China, and across the Ryukyu Islands to the south of Japan, contrasted with an earlier withdrawal over the Yangtze Basin. They attributed this feature to El Niño-like mean state changes over the monsoon trough and subtropical anticyclone over the western Pacific region (Meehl et al., 2007b). A southwestward extension of the subtropical anticyclone over the northwestern Pacific Ocean associated with El Niño-like mean state changes and a dry air intrusion in the mid-troposphere from the Asian continent to the northwest of Japan provides favorable conditions for intense precipitation in the Baiu season in Japan (Kanada et al., 2010a). Kitoh et al. (2009) projected changes in

precipitation characteristics during the east Asian summer rainy season, using a 5-km mesh cloud-resolving model embedded in a 20-km mesh global atmospheric model with CMIP3 mean SST changes. The frequency of heavy precipitation was projected to increase at the end of the 21st century for hourly as well as daily precipitation. Further, extreme hourly precipitation was projected to increase even in the near future (2030s) when the temperature increase is still modest, even though uncertainties in the projection (and even the simulation) of hourly rainfall are still high.

Climate change scenarios for the 21st century show a weakening of the North American monsoon through a weakening and poleward expansion of the Hadley cell (Lu et al., 2007). The expansion of the Hadley cell is caused by an increase in the subtropical static stability, which pushes poleward the baroclinic instability zone and hence the outer boundary of the Hadley cell. Simple physical arguments (Held and Soden, 2006) predict a slowdown of the tropical overturning circulation under global warming. A few studies (e.g., Marengo et al., 2009a) have projected over the period 1960-2100 a weak tendency for an increase in dry spells. The projections show an increase in the frequency of rainfall extremes in southeastern South America by the end of the 21st century, possibly due to an intensification of the moisture transport from Amazonia by a more frequent/intense low-level jet east of the Andes in the A2 emissions scenario (Marengo et al., 2009a; Soares and Marengo, 2009).

There are many deficiencies in model representation of the monsoons and the processes affecting them, and this reduces confidence in their ability to project future changes. Some of the uncertainty in global and regional climate change projections in the monsoon regions results from the limits in the model representation of resolved processes (e.g., moisture advection), the parameterizations of sub-grid-scale processes (e.g., clouds, precipitation), and model simulations of feedback mechanisms at the global and regional scale (e.g., changes in land use/cover; see also Section 3.1.4). Kharin et al. (2007) made an intercomparison of precipitation extremes in the tropical region in all AR4 models with observed extremes expressed as 20-year return values. They found very large disagreement in the tropics suggesting that some physical processes associated with extreme precipitation are not well represented by the models due to model resolution and physics. Shukla (2007) noted that current climate models cannot even adequately predict the mean intensity and the seasonal variations of the Asian summer monsoon. This reduces confidence in the projected changes in extreme precipitation over the monsoon regions. Many of the important climatic effects of the Madden-Julian Oscillation (MJO, a natural mode of the climate system operating on time scales of about a month), including its impacts on rainfall variability in the monsoons, are still poorly simulated by contemporary climate models (Christensen et al., 2007).

Current GCMs still have difficulties and display a wide range of skill in simulating the subseasonal variability associated with the Asian summer monsoon (Lin et al., 2008b). Most GCMs simulate westward propagation of the coupled equatorial easterly waves, but relatively poor eastward propagation of the MJO and overly weak variances for both the easterly waves and the MJO. Most GCMs are able to reproduce the basic

characteristics of the precipitation seasonal cycle associated with the South American Monsoon System (SAMS), but there are large discrepancies in the South Atlantic Convergence Zone represented by the models in both intensity and location, and in its seasonal evolution (Vera et al., 2006). In addition, models exhibit large discrepancies in the direction of the changes associated with the summer (SAMS) precipitation, which makes the projections for that region highly uncertain. Lin et al. (2008a) show that the coupled GCMs have significant problems and display a wide range of skill in simulating the North American monsoon and associated intra-seasonal variability.

Most of the models reproduce the monsoon rain belt, extending from southeast to northwest, and its gradual northward shift in early summer, but overestimate the precipitation over the core monsoon region throughout the seasonal cycle and fail to reproduce the monsoon retreat in the fall. The AR4 assessed that models fail in representing the main features of the west African monsoon although most of them do have a monsoonal climate albeit with some distortion (Christensen et al., 2007). Other major sources of uncertainty in projections of monsoon changes are the responses and feedbacks of the climate system to emissions as represented in climate models. These uncertainties are particularly related to the representation of the conversion of emissions into concentrations of radiatively active species (i.e., via atmospheric chemistry and carbon cycle models) and especially those derived from aerosol products of biomass burning, which can affect the onset of the rainy season (Silva Dias et al., 2002). The subsequent response of the physical climate system complicates the nature of future projections of monsoon precipitation. Moreover, the long-term variations of model skill in simulating monsoons and their variations represent an additional source of uncertainty for the monsoon regions, and indicate that the regional reliability of long climate model runs may depend on the time slice for which the output of the model is analyzed.

The AR4 (Hegerl et al., 2007) concluded that the current understanding of climate change in the monsoon regions remains one of considerable uncertainty with respect to circulation and precipitation. With a few exceptions in some monsoon regions, this has not changed. These conclusions have been based on very few studies, there are many issues with model representation of monsoons and the underlying processes, and there is little consensus in climate models, so there is *low confidence* in projections of changes in monsoons, even in the sign of the change. However, one common pattern is a *likely* increase in extreme precipitation in monsoon regions (see Section 3.3.2), though not necessarily induced by changes in monsoon characteristics, and not necessarily occurring in all monsoon regions.

3.4.2.   El Niño-Southern Oscillation

The El Niño-Southern Oscillation (ENSO) is a natural fluctuation of the global climate system caused by equatorial ocean-atmosphere interaction in the tropical Pacific Ocean (Philander, 1990). The term 'Southern

Oscillation' refers to a tendency for above-average surface atmospheric pressures in the Indian Ocean to be associated with below-average pressures in the Pacific, and vice versa. This oscillation is associated with variations in SSTs in the east equatorial Pacific. The oceanic and atmospheric variations are collectively referred to as ENSO. An El Niño episode is one phase of the ENSO phenomenon and is associated with abnormally warm central and east equatorial Pacific Ocean surface temperatures, while the opposite phase, a La Niña episode, is associated with abnormally cool ocean temperatures in this region. Both phases are associated with a characteristic spatial pattern of droughts and floods. An El Niño episode is usually accompanied by drought in southeastern Asia, India, Australia, southeastern Africa, Amazonia, and northeast Brazil, with fewer than normal tropical cyclones around Australia and in the North Atlantic. Wetter than normal conditions during El Niño episodes are observed along the west coast of tropical South America, subtropical latitudes of western North America, and southeastern America. In a La Niña episode the climate anomalies are usually the opposite of those in an El Niño. Pacific islands are strongly affected by ENSO variations. Recent research (e.g., Kenyon and Hegerl, 2008; Ropelewski and Bell, 2008; Schubert et al., 2008a; Alexander et al., 2009; Grimm and Tedeschi, 2009; Zhang et al., 2010) has demonstrated that different phases of ENSO (El Niño or La Niña episodes) also are associated with different frequencies of occurrence of short-term weather extremes such as heavy rainfall events and extreme temperatures. The relationship between ENSO and interannual variations in tropical cyclone activity is well known (e.g., Kuleshov et al., 2008). The simultaneous occurrence of a variety of climate extremes in an El Niño episode (or a La Niña episode) may provide special challenges for organizations coping with disasters induced by ENSO (see also Section 3.1.1). Monitoring and predicting ENSO can lead to disaster risk reduction through early warning (see Case Study 9.2.11).

The AR4 noted that orbital variations could affect the ENSO behavior (Jansen et al., 2007). Cane (2005) found that a relatively simple coupled model suggested that systematic changes in El Niño could be stimulated by seasonal changes in solar insolation. However, a more comprehensive model simulation (Wittenberg, 2009) has suggested that long-term changes in the behavior of the phenomenon might occur even without forcing from radiative changes. Vecchi and Wittenberg (2010) concluded that the "tropical Pacific could generate variations in ENSO frequency and intensity on its own (via chaotic behavior), respond to external radiative forcings (e.g., changes in greenhouse gases, volcanic eruptions, atmospheric aerosols, etc), or both." Meehl et al. (2009a) demonstrate that solar insolation variations related to the 11-year sunspot cycle can affect ocean temperatures associated with ENSO.

ENSO has varied in strength over the last millennium with stronger activity in the 17th century and late 14th century, and weaker activity during the 12th and 15th centuries (Cobb et al., 2003; Conroy et al., 2009). On longer time scales, there is evidence that ENSO may have changed in response to changes in the orbit of the Earth (Vecchi and Wittenberg, 2010), with the phenomenon apparently being weaker around 6,000 years ago (according to proxy measurements from corals

and climate model simulations; Rein et al., 2005; Brown et al., 2006; Otto-Bliesner et al., 2009), and model simulations suggest that it was stronger at the last glacial maximum (An et al., 2004). Fossil coral evidence indicates that the phenomenon continued to operate during the last glacial interval (Tudhope et al., 2001). Thus the paleoclimatic evidence indicates that ENSO can continue to operate, although altered perhaps in intensity, in very different background climate states.

The AR4 noted that the nature of ENSO has varied substantially over the period of instrumental data, with strong events from the late 19th century through the first quarter of the 20th century and again after 1950. An apparent climate shift around 1976-1977 was associated with a shift to generally above-normal SSTs in the central and eastern Pacific and a tendency toward more prolonged and stronger El Niño episodes (Trenberth et al., 2007). Ocean temperatures in the central equatorial Pacific (the so-called NINO3 index) suggest a trend toward more frequent or stronger El Niño episodes over the past 50 to 100 years (Vecchi and Wittenberg, 2010). Vecchi et al. (2006) reported a weakening of the equatorial Pacific pressure gradient since the 1960s, with a sharp drop in the 1970s. Power and Smith (2007) proposed that the apparent dominance of El Niño during the last few decades was due in part to a change in the background state of the Southern Oscillation Index (SOI, the standardized difference in surface atmospheric pressure between Tahiti and Darwin), rather than a change in variability or a shift to more frequent El Niño events alone. Nicholls (2008) examined the behavior of the SOI and another index, the NINO3.4 index of central equatorial Pacific SSTs, but found no evidence of trends in the variability or the persistence of the indices [although Yu and Kao (2007) reported decadal variations in the persistence barrier, the tendency for weaker persistence across the Northern Hemisphere spring], nor in their seasonal patterns. There was a trend toward what might be considered more 'El Niño-like' behavior in the SOI (and more weakly in NINO3.4), but only through the period March to September and not in November to February, the season when El Niño and La Niña events typically peak. The trend in the SOI reflected only a trend in Darwin pressures, with no trend in Tahiti pressures. Apart from this trend, the temporal/seasonal nature of ENSO has been remarkably consistent through a period of strong global warming. There is evidence, however, of a tendency for recent El Niño episodes to be centered more in the central equatorial Pacific than in the east Pacific (Yeh et al., 2009), and for these central Pacific episodes to be increasing in intensity (Lee and McPhaden, 2010). In turn, these changes may explain changes that have been noted in the remote influences of the phenomenon on the climate over Australia and in the mid-latitudes (Wang and Hendon, 2007; Weng et al., 2009). For instance, Taschetto et al. (2009) demonstrated that episodes with the warming centered in the central Pacific exhibit different patterns of Australian rainfall variations relative to the east Pacific-centered El Niño events.

The possible role of increased greenhouse gases in affecting the behavior of ENSO over the past 50 to 100 years is uncertain. Yeh et al. (2009) suggested that changes in the background temperature associated with increases in greenhouse gases should affect the behavior of El Niño, such as the location of the strongest SST anomalies, because El Niño

behavior is strongly related to the average ocean temperature gradients in the equatorial Pacific. Some studies (e.g., Q. Zhang et al., 2008) have suggested that increased activity might be due to increased CO_2; however, no formal attribution study has yet been completed and some other studies (e.g., Power and Smith, 2007) suggest that changes in the phenomenon are within the range of natural variability (i.e., that no change has yet been detected, let alone attributed to a specific cause).

Global warming is projected to lead to a mean reduction in the zonal mean wind across the equatorial Pacific (Vecchi and Soden, 2007b). However, this change should not be described as an 'El Niño-like' average change even though during an El Niño episode these winds also weaken, because there is only limited correspondence between these changes in the mean state of the equatorial Pacific and an El Niño episode. The AR4 determined that all models exhibited continued ENSO interannual variability in projections through the 21st century, but the projected behavior of the phenomenon differed between models, and it was concluded that "there is no consistent indication at this time of discernible changes in projected ENSO amplitude or frequency in the 21st century" (Meehl et al., 2007b). Models project a wide variety of changes in ENSO variability and the frequency of El Niño episodes as a consequence of increased greenhouse gas concentrations, with a range between a 30% reduction to a 30% increase in variability (van Oldenborgh et al., 2005). One model study even found that although ENSO activity increased when atmospheric CO_2 concentrations were doubled or quadrupled, a considerable decrease in activity occurred when CO_2 was increased by a factor of 16 times, much greater than is possible through the 21st century (Cherchi et al., 2008), suggesting a wide variety of possible ENSO changes as a result of CO_2 changes. The remote impacts, on rainfall for instance, of ENSO may change as CO_2 increases, even if the equatorial Pacific aspect of ENSO does not change substantially. For instance, regions in which rainfall increases in the future tend to show increases in interannual rainfall variability (Boer, 2009), without any strong change in the interannual variability of tropical SSTs. Also, since some long-term projected changes in response to increased greenhouse gases may resemble the climate response to an El Niño event, this may enhance or mask the response to El Niño events in the future (Lau et al., 2008b; Müller and Roeckner, 2008).

One change that models tend to project is an increasing tendency for El Niño episodes to be centered in the central equatorial Pacific, rather than the traditional location in the eastern equatorial Pacific. Yeh et al. (2009) examined the relative frequency of El Niño episodes simulated in coupled climate models with projected increases in greenhouse gas concentrations. A majority of models, especially those best able to simulate the current ratio of central Pacific locations to east Pacific locations of El Niño events, projected a further increase in the relative frequency of these central Pacific events. Such a change would also have implications for the remote influence of the phenomenon on climate away from the equatorial Pacific (e.g., Australia and India). However, even the projection that the 21st century may see an increased frequency of central Pacific El Niño episodes, relative to the frequency of events located further east (Yeh et al., 2009), is subject to considerable uncertainty. Of

the 11 coupled climate model simulations examined by Yeh et al. (2009), three projected a relative decrease in the frequency of these central Pacific episodes, and only four of the models produced a statistically significant change to more frequent central Pacific events.

A caveat regarding all projections of future behavior of ENSO arises from systematic biases in the depiction of ENSO behavior through the 20th century by models (Randall et al., 2007; Guilyardi et al., 2009). Leloup et al. (2008) for instance, demonstrate that coupled climate models show wide differences in the ability to reproduce the spatial characteristics of SST variations associated with ENSO during the 20th century, and all models have failings. They concluded that it is difficult to even classify models by the quality of their reproductions of the behavior of ENSO, because models scored unevenly in their reproduction of the different phases of the phenomenon. This makes it difficult to determine which models to use to project future changes in ENSO. Moreover, most of the models are not able to reproduce the typical circulation anomalies associated with ENSO in the Southern Hemisphere (Vera and Silvestri, 2009) and the Northern Hemisphere (Joseph and Nigam, 2006).

There was no consistency in projections of changes in ENSO variability or frequency at the time of the AR4 (Meehl et al., 2007b) and this situation has not changed as a result of post-AR4 studies. The evidence is that the nature of ENSO has varied in the past apparently sometimes in response to changes in radiative forcing but also possibly due to internal climatic variability. Since radiative forcing will continue to change in the future, we can confidently expect changes in ENSO and its impacts as well, although both El Niño and La Niña episodes will continue to occur (e.g., Vecchi and Wittenberg, 2010). Our current limited understanding, however, means that it is not possible at this time to confidently predict whether ENSO activity will be enhanced or damped due to anthropogenic climate change, or even if the frequency of El Niño or La Niña episodes will change (Collins et al., 2010).

In summary, there is *medium confidence* **in a recent trend toward more frequent central equatorial Pacific El Niño episodes, but insufficient evidence for more specific statements about observed trends in ENSO. Model projections of changes in ENSO variability and the frequency of El Niño episodes as a consequence of increased greenhouse gas concentrations are not consistent, and so there is** *low confidence* **in projections of changes in the phenomenon. However, there is** *medium confidence* **regarding a projected increase (projected by most GCMs) in the relative frequency of central equatorial Pacific events, which typically exhibit different patterns of climate variations than do the classical East Pacific events.**

3.4.3.   Other Modes of Variability

Other natural modes of variability beside ENSO (Section 3.4.2) that are relevant to extremes and disasters include the North Atlantic Oscillation

(NAO), the Southern Annular Mode (SAM), and the Indian Ocean Dipole (IOD) (Trenberth et al., 2007). The NAO is a large-scale seesaw in atmospheric pressure between the subtropical high and the polar low in the North Atlantic region. The positive NAO phase has a strong subtropical high-pressure center and a deeper than normal Icelandic low. This results in a shift of winter storms crossing the Atlantic Ocean to a more northerly track, and is associated with warm and wet winters in northwestern Europe and cold and dry winters in northern Canada and Greenland. Scaife et al. (2008) discuss the relationship between the NAO and European extremes. Paleoclimatic data indicate that the NAO was persistently in its positive phase during medieval times and persistently in its negative phase during the cooler Little Ice Age (Trouet et al., 2009). The NAO is closely related to the Northern Annular Mode (NAM); for brevity we focus here on the NAO but much of what is said about the NAO also applies to the NAM. The SAM is the largest mode of Southern Hemisphere extratropical variability and refers to north-south shifts in atmospheric mass between the middle and high latitudes. It plays an important role in climate variability in these latitudes. The SAM positive phase is linked to negative sea level pressure anomalies over the polar regions and intensified westerlies. It has been associated with cooler than normal temperatures over most of Antarctica and Australia, with warm anomalies over the Antarctic Peninsula, southern South America, and southern New Zealand, and with anomalously dry conditions over southern South America, New Zealand, and Tasmania and wet anomalies over much of Australia and South Africa (e.g., Hendon et al., 2007). The IOD is a coupled ocean-atmosphere phenomenon in the Indian Ocean. A positive IOD event is associated with anomalous cooling in the southeastern equatorial Indian Ocean and anomalous warming in the western equatorial Indian Ocean. Recent work (Ummenhofer et al., 2008, 2009a,b) has implicated the IOD as a cause of droughts in Australia, and heavy rainfall in east Africa (Ummenhofer et al., 2009c). There is also evidence of modes of variability operating on multi-decadal time scales, notably the Pacific Decadal Oscillation (PDO) and the Atlantic Multi-decadal Oscillation (AMO). Variations in the PDO have been related to precipitation extremes over North America (Zhang et al., 2010).

Both the NAO and the SAM exhibited trends toward their positive phase (strengthened mid-latitude westerlies) over the last three to four decades, although the NAO has been in its negative phase in the last few years. Goodkin et al. (2008) concluded that the variability in the NAO is linked with changes in the mean temperature of the Northern Hemisphere. Dong et al. (2011) demonstrated that some of the observed late 20th-century decadal-scale changes in NAO behavior could be reproduced by increasing the CO_2 concentrations in a coupled model, and concluded that greenhouse gas concentrations may have played a role in forcing these changes. The largest observed trends in the SAM occur in December to February, and model simulations indicate that these are due mainly to stratospheric ozone changes. However it has been argued that anthropogenic circulation changes are poorly characterized by trends in the annular modes (Woollings et al., 2008). Further complicating these trends, Silvestri and Vera (2009) reported changes in the typical hemispheric circulation pattern related to the SAM and its associated impact on both temperature and precipitation anomalies, particularly

over South America and Australia, between the 1960s-1970s and 1980s-1990s. The time scales of variability in modes such as the AMO and PDO are so long that it is difficult to diagnose any change in their behavior in modern data, although some evidence suggests that the PDO may be affected by anthropogenic forcing (Meehl et al., 2009b). The AR4 (Hegerl et al., 2007) concluded that trends over recent decades in the NAO and SAM are *likely* related in part to human activity. The negative NAO phase of the last few years, however, with the lack of formal attribution studies, means that attribution of changes in the NAO to human activity in recent decades now can only be considered *about as likely as not* (expert opinion). Attribution of the SAM trend to human activity is still assessed to be *likely* (expert opinion) although mainly attributable to trends in stratospheric ozone concentration (Hegerl et al., 2007).

The AR4 noted that there was considerable spread among the model projections of the NAO, leading to low confidence in NAO projected changes, but the magnitude of the increase for the SAM is generally more consistent across models (Meehl et al., 2007b). However, the ability of coupled models to simulate the observed SAM impact on climate variability in the Southern Hemisphere is limited (e.g., Miller et al., 2006; Vera and Silvestri, 2009). Variations in the longer time-scale modes of variability (AMO, PDO) might affect projections of changes in extremes associated with the various natural modes of variability and global temperatures (Keenlyside et al., 2008).

Sea level pressure is projected to increase over the subtropics and mid-latitudes, and decrease over high latitudes (Meehl et al., 2007b). This would equate to trends in the NAO and SAM, with a poleward shift of the storm tracks of several degrees latitude and a consequent increase in cyclonic circulation patterns over the Arctic and Antarctica. In the Southern Hemisphere, two opposing effects, stratospheric ozone recovery and increasing greenhouse gases, can be expected to affect the modes such as the SAM (Arblaster et al., 2011). During the 21st century, although stratospheric ozone concentrations are expected to recover, tending to lead to a weakening of the SAM, models consistently project polar vortex intensification to continue due to the increases in greenhouse gases, except in summer where the competing effects of stratospheric ozone recovery complicate this picture (Arblaster et al., 2011).

A recent study (Woollings et al., 2010) found a tendency toward a more positive NAO under anthropogenic forcing through the 21st century with one model, although they concluded that confidence in the model projections was low because of deficiencies in its simulation of current-day NAO regimes. Goodkin et al. (2008) predict continuing high variability, on multi-decadal scales, in the NAO with continued global warming. Keenlyside et al. (2008) proposed that variations associated with the multi-decadal modes of variability may offset warming due to increased greenhouse gas concentrations over the next decade or so. Conway et al. (2007) reported that model projections of future IOD behavior showed no consistency. Kay and Washington (2008) reported that under some emissions scenarios, changes in a dipole mode in the Indian Ocean could change rainfall extremes in southern Africa.

In summary, it is *likely* that there has been an anthropogenic influence on recent trends in the SAM (linked with trends in stratospheric ozone rather than changes in greenhouse gases), but it is only *about as likely as not* that there have been anthropogenic influences on observed trends in the NAO. Issues with the ability of models to simulate current behavior of these natural modes, the influence of competing factors (e.g., stratospheric ozone, greenhouse gases) on current and future mode behavior, and inconsistency between the model projections (and the seasonal dependence of these projections), means that there is *low confidence* in the ability to project changes in the modes including the NAO, SAM, and IOD. Models do, however, consistently project a strengthening of the polar vortex in the Southern Hemisphere from increasing greenhouse gases, although in summer stratospheric ozone recovery is expected to offset this intensification.

3.4.4.   Tropical Cyclones

Tropical cyclones occur in most tropical oceans and pose a significant threat to coastal populations and infrastructure, and marine interests such as shipping and offshore activities. Each year, about 90 tropical cyclones occur globally, and this number has remained roughly steady over the modern period of geostationary satellites (since around the mid-1970s). While the global frequency has remained steady, there can be substantial inter-annual to multi-decadal frequency variability within individual ocean basins (e.g., Webster et al., 2005). This regional variability, particularly when combined with substantial inter-annual to multi-decadal variability in tropical cyclone tracks (e.g., Kossin et al., 2010), presents a significant challenge for disaster planning and mitigation aimed at specific regions.

Tropical cyclones are perhaps most commonly associated with extreme wind, but storm-surge and freshwater flooding from extreme rainfall generally cause the great majority of damage and loss of life (e.g., Rappaport, 2000; Webster, 2008). Related indirect factors, such as the failure of the levee system in New Orleans during the passage of Hurricane Katrina (2005), or mudslides during the landfall of Hurricane Mitch (1998) in Central America, represent important related impacts (Case Study 9.2.5). Projected sea level rise will further compound tropical cyclone surge impacts. Tropical cyclones that track poleward can undergo a transition to become extratropical cyclones. While these storms have different characteristics than their tropical progenitors, they can still be accompanied by a storm surge that can impact regions well away from the tropics (e.g., Danard et al., 2004).

Tropical cyclones are typically classified in terms of their intensity, which is a measure of near-surface wind speed (sometimes categorized according to the Saffir-Simpson scale). The strongest storms (Saffir-Simpson category 3, 4, and 5) are comparatively rare but are generally responsible for the majority of damage (e.g., Landsea, 1993; Pielke Jr. et al., 2008). Additionally, there are marked differences in the characteristics of both

observed and projected tropical cyclone variability when comparing weaker and stronger tropical cyclones (e.g., Webster et al., 2005; Elsner et al., 2008; Bender et al., 2010), while records of the strongest storms are potentially less reliable than those of their weaker counterparts (Landsea et al., 2006).

In addition to intensity, the structure and areal extent of the wind field in tropical cyclones, which can be largely independent of intensity, also play an important role on potential impacts, particularly from storm surge (e.g., Irish and Resio, 2010), but measures of storm size are largely absent in historical data. Other relevant tropical cyclone measures include frequency, duration, and track. Forming robust physical links between all of the metrics briefly mentioned here and natural or human-induced changes in climate variability is a major challenge. Significant progress is being made, but substantial uncertainties still remain due largely to data quality issues (see Section 3.2.1 and below) and imperfect theoretical and modeling frameworks (see below).

Observed Changes

Detection of trends in tropical cyclone metrics such as frequency, intensity, and duration remains a significant challenge. Historical tropical cyclone records are known to be heterogeneous due to changing observing technology and reporting protocols (e.g., Landsea et al., 2004). Further heterogeneity is introduced when records from multiple ocean basins are combined to explore global trends because data quality and reporting protocols vary substantially between regions (Knapp and Kruk, 2010). Progress has been made toward a more homogeneous global record of tropical cyclone intensity using satellite data (Knapp and Kossin, 2007; Kossin et al., 2007), but these records are necessarily constrained to the satellite era and so only represent the past 30 to 40 years.

Natural variability combined with uncertainties in the historical data makes it difficult to detect trends in tropical cyclone activity. There have been no significant trends observed in global tropical cyclone frequency records, including over the present 40-year period of satellite observations (e.g., Webster et al., 2005). Regional trends in tropical cyclone frequency have been identified in the North Atlantic, but the fidelity of these trends is debated (Holland and Webster, 2007; Landsea, 2007; Mann et al., 2007a). Different methods for estimating undercounts in the earlier part of the North Atlantic tropical cyclone record provide mixed conclusions (Chang and Guo, 2007; Mann et al., 2007b; Kunkel et al., 2008; Vecchi and Knutson, 2008). Regional trends have not been detected in other oceans (Chan and Xu, 2009; Kubota and Chan, 2009; Callaghan and Power, 2011). It thus remains uncertain whether any observed increases in tropical cyclone frequency on time scales longer than about 40 years are robust, after accounting for past changes in observing capabilities (Knutson et al., 2010).

Frequency estimation requires only that a tropical cyclone be identified and reported at some point in its lifetime, whereas intensity estimation

requires a series of specifically targeted measurements over the entire duration of the tropical cyclone (e.g., Landsea et al., 2006). Consequently, intensity values in the historical records are especially sensitive to changing technology and improving methodology, which heightens the challenge of detecting trends within the backdrop of natural variability. Global reanalyses of tropical cyclone intensity using a homogenous satellite record have suggested that changing technology has introduced a non-stationary bias that inflates trends in measures of intensity (Kossin et al., 2007), but a significant upward trend in the intensity of the strongest tropical cyclones remains after this bias is accounted for (Elsner et al., 2008). While these analyses are suggestive of a link between observed global tropical cyclone intensity and climate change, they are necessarily confined to a roughly 30-year period of satellite observations, and cannot provide clear evidence for a longer-term trend.

Time series of power dissipation, an aggregate compound of tropical cyclone frequency, duration, and intensity that measures total energy consumption by tropical cyclones, show upward trends in the North Atlantic and weaker upward trends in the western North Pacific over the past 25 years (Emanuel, 2007), but interpretation of longer-term trends in this quantity is again constrained by data quality concerns. The variability and trend of power dissipation can be related to SST and other local factors such as tropopause temperature and vertical wind shear (Emanuel, 2007), but it is a current topic of debate whether local SST or the difference between local SST and mean tropical SST is the more physically relevant metric (Swanson, 2008). The distinction is an important one when making projections of changes in power dissipation based on projections of SST changes, particularly in the tropical Atlantic where SST has been increasing more rapidly than in the tropics as a whole (Vecchi et al., 2008). Accumulated cyclone energy, which is an integrated metric analogous to power dissipation, has been declining globally since reaching a high point in 2005, and is presently at a 40-year low point (Maue, 2009). The present period of quiescence, as well as the period of heightened activity leading up to the high point in 2005, does not clearly represent substantial departures from past variability (Maue, 2009).

Increases in tropical water vapor and rainfall (Trenberth et al., 2005; Lau and Wu, 2007) have been identified and there is some evidence for related changes in tropical cyclone-related rainfall (Lau et al., 2008a), but a robust and consistent trend in tropical cyclone rainfall has not yet been established due to a general lack of studies. Similarly, an increase in the length of the North Atlantic hurricane season has been noted (Kossin, 2008), but the uncertainty in the amplitude of the trends and the lack of additional studies limits the utility of these results for a meaningful assessment.

Estimates of tropical cyclone variability prior to the modern instrumental historical record have been constructed using archival documents (Chenoweth and Devine, 2008), coastal marsh sediment records, and isotope markers in coral, speleothems, and tree rings, among other methods (Frappier et al., 2007a). These estimates demonstrate centennial-to millennial-scale relationships between climate and tropical cyclone

activity (Donnelly and Woodruff, 2007; Frappier et al., 2007b; Nott et al., 2007; Nyberg et al., 2007; Scileppi and Donnelly, 2007; Neu, 2008; Woodruff et al., 2008a,b; Mann et al., 2009; Yu et al., 2009), but generally do not provide robust evidence that the observed post-industrial tropical cyclone activity is unprecedented.

The AR4 Summary for Policymakers concluded that it is *likely* that an increase had occurred in intense tropical cyclone activity since 1970 in some regions (IPCC, 2007b). The subsequent CCSP assessment report (Kunkel et al., 2008) concluded that it is *likely* that the frequency of tropical storms, hurricanes, and major hurricanes in the North Atlantic has increased over the past 100 years, a time in which Atlantic SSTs also increased. Kunkel et al. (2008) also concluded that the increase in Atlantic power dissipation is *likely* substantial since the 1950s. Based on research subsequent to the AR4 and Kunkel et al. (2008), which further elucidated the scope of uncertainties in the historical tropical cyclone data, the most recent assessment by the World Meteorological Organization (WMO) Expert Team on Climate Change Impacts on Tropical Cyclones (Knutson et al., 2010) concluded that it remains uncertain whether past changes in any tropical cyclone activity (frequency, intensity, rainfall) exceed the variability expected through natural causes, after accounting for changes over time in observing capabilities. The present assessment regarding observed trends in tropical cyclone activity is essentially identical to the WMO assessment (Knutson et al., 2010): there is *low confidence* that any observed long-term (i.e., 40 years or more) increases in tropical cyclone activity are robust, after accounting for past changes in observing capabilities.

Causes of the Observed Changes

In addition to the natural variability of tropical SSTs, several studies have concluded that there is a detectable tropical SST warming trend due to increasing greenhouse gases (Karoly and Wu, 2005; Knutson et al., 2006; Santer et al., 2006; Gillett et al., 2008a). The region where this anthropogenic warming has occurred encompasses tropical cyclogenesis regions, and Kunkel et al. (2008) stated that it is *very likely* that human-caused increases in greenhouse gases have contributed to the increase in SSTs in the North Atlantic and the Northwest Pacific hurricane formation regions over the 20th century.

Changes in the mean thermodynamic state of the tropics can be directly linked to tropical cyclone variability within the theoretical framework of potential intensity theory (Bister and Emanuel, 1998). In this framework, the expected response of tropical cyclone intensity to observed climate change is relatively straightforward: if climate change causes an increase in the ambient potential intensity that tropical cyclones move through, the distribution of intensities in a representative sample of storms is expected to shift toward greater intensities (Emanuel, 2000; Wing et al., 2007). The fractional changes associated with such a shift in the distribution would be largest in the upper quantiles of the distribution as the strongest tropical cyclones become stronger (Elsner et al., 2008).

Given the evidence that SST in the tropics has increased due to increasing greenhouse gases, and the theoretical expectation that increases in potential intensity will lead to stronger storms, it is essential to fully understand the relationship between SST and potential intensity. Observations demonstrate a strong positive correlation between SST and the potential intensity. This relationship suggests that SST increases will lead to increased potential intensity, which will then ultimately lead to stronger storms (Emanuel, 2000; Wing et al., 2007). However, there is a growing body of research suggesting that local potential intensity is controlled by the difference between local SST and spatially averaged SST in the tropics (Vecchi and Soden, 2007a; Xie et al., 2010; Ramsay and Sobel, 2011). Since increases in SST due to global warming are not expected to lead to continuously increasing SST gradients, this recent research suggests that increasing SST due to global warming, by itself, does not yet have a fully understood physical link to increasingly strong tropical cyclones.

The present period of heightened tropical cyclone activity in the North Atlantic, concurrent with comparative quiescence in other ocean basins (e.g., Maue, 2009), is apparently related to differences in the rate of SST increases, as global SST has been rising steadily but at a slower rate than has the Atlantic (Holland and Webster, 2007). The present period of relatively enhanced warming in the Atlantic has been proposed to be due to internal variability (Zhang and Delworth, 2009), anthropogenic tropospheric aerosols (Mann and Emanuel, 2006), and mineral (dust) aerosols (Evan et al., 2009). None of these proposed mechanisms provide a clear expectation that North Atlantic SST will continue to increase at a greater rate than the tropical mean SST.

Changes in tropical cyclone intensity, frequency, genesis location, duration, and track contribute to what is sometimes broadly defined as 'tropical cyclone activity.' Of these metrics, intensity has the most direct physically reconcilable link to climate variability within the framework of potential intensity theory, as described above (Kossin and Vimont, 2007). Statistical correlations between necessary ambient environmental conditions (e.g., low vertical wind shear and adequate atmospheric instability and moisture) and tropical cyclogenesis frequency have been well documented (DeMaria et al., 2001) but changes in these conditions due specifically to increasing greenhouse gas concentrations do not necessarily preserve the same statistical relationships. For example, the observed minimum SST threshold for tropical cyclogenesis is roughly 26°C. This relationship might lead to an expectation that anthropogenic warming of tropical SST and the resulting increase in the areal extent of the region of 26°C SST should lead to increases in tropical cyclone frequency. However, there is a growing body of evidence that the minimum SST threshold for tropical cyclogenesis increases at about the same rate as the SST increase due solely to greenhouse gas forcing (e.g., Ryan et al., 1992; Dutton et al., 2000; Yoshimura et al., 2006; Bengtsson et al., 2007; Knutson et al., 2008; Johnson and Xie, 2010). This is because the threshold conditions for tropical cyclogenesis are controlled not just by surface temperature but also by atmospheric stability (measured from the lower boundary to the tropopause), which responds to greenhouse gas forcing in a more complex way than SST

alone. That is, when the SST changes due to greenhouse warming are deconvolved from the background natural variability, that part of the SST variability, by itself, has no manifest effect on tropical cyclogenesis. In this case, the simple observed relationship between tropical cyclogenesis and SST, while robust, does not adequately capture the relevant physical mechanisms of tropical cyclogenesis in a warming world.

Another challenge to identifying causes behind observed changes in tropical cyclone activity is introduced by uncertainties in the reanalysis data used to identify environmental changes in regions where tropical cyclones develop and evolve (Bister and Emanuel, 2002; Emanuel, 2010). In particular, heterogeneity in upper-tropospheric kinematic and thermodynamic metrics complicates the interpretation of long-term changes in vertical wind shear and potential intensity, both of which are important environmental controls on tropical cyclones.

Based on a variety of model simulations, the expected long-term changes in global tropical cyclone characteristics under greenhouse warming is a decrease or little change in frequency concurrent with an increase in mean intensity. One of the challenges for identifying these changes in the existing data records is that the expected changes predicted by the models are generally small when compared with changes associated with observed short-term natural variability. Based on changes in tropical cyclone intensity predicted by idealized numerical simulations with CO_2-induced tropical SST warming, Knutson and Tuleya (2004) suggested that clearly detectable increases may not be manifest for decades to come. Their argument was based on a comparison of the amplitude of the modeled upward trend (i.e., the signal) in storm intensity with the amplitude of the interannual variability (i.e., the noise). The recent high-resolution dynamical downscaling study of Bender et al. (2010) supports this argument and suggests that the predicted increases in the frequency of the strongest Atlantic storms may not emerge as a clear statistically significant signal until the latter half of the 21st century under the SRES A1B warming scenario. Still, it should be noted that while these model projections suggest that a statistically significant signal may not emerge until some future time, the likelihood of more intense tropical cyclones is projected to continually increase throughout the 21st century.

With the exception of the North Atlantic, much of the global tropical cyclone data is confined to the period from the mid-20th century to present. In addition to the limited period of record, the uncertainties in the historical tropical cyclone data (Section 3.2.1 and this section) and the extent of tropical cyclone variability due to random processes and linkages with various climate modes such as El Niño, do not presently allow for the detection of any clear trends in tropical cyclone activity that can be attributed to greenhouse warming. As such, it remains unclear to what degree the causal phenomena described here have modulated post-industrial tropical cyclone activity.

The AR4 concluded that it is *more likely than not* that anthropogenic influence has contributed to increases in the frequency of the most intense tropical cyclones (Hegerl et al., 2007). Based on subsequent research that further elucidated the scope of uncertainties in both the historical tropical cyclone data as well as the physical mechanisms underpinning the observed relationships, no such attribution conclusion was drawn in the recent WMO assessment (Knutson et al., 2010). The present assessment regarding detection and attribution of trends in tropical cyclone activity is similar to the WMO assessment (Knutson et al., 2010): the uncertainties in the historical tropical cyclone records, the incomplete understanding of the physical mechanisms linking tropical cyclone metrics to climate change, and the degree of tropical cyclone variability – comprising random processes and linkages to various natural climate modes such as El Niño – provide only *low confidence* for the attribution of any detectable changes in tropical cyclone activity to anthropogenic influences.

Projected Changes and Uncertainties

The AR4 concluded (Meehl et al., 2007b) that a broad range of modeling studies project a *likely* increase in peak wind intensity and near-storm precipitation in future tropical cyclones. A reduction of the overall number of storms was also projected (but with lower confidence), with a greater reduction in weaker storms in most basins and an increase in the frequency of the most intense storms. Knutson et al. (2010) concluded that it is *likely* that the mean maximum wind speed and near-storm rainfall rates of tropical cyclones will increase with projected 21st-century warming, and it is *more likely than not* that the frequency of the most intense storms will increase substantially in some basins, but it is *likely* that overall global tropical cyclone frequency will decrease or remain essentially unchanged. The conclusions here are similar to those of the AR4 and Knutson et al. (2010).

The spatial resolution of some models such as the CMIP3 coupled ocean-atmosphere models used in the AR4 is generally not high enough to accurately resolve tropical cyclones, and especially to simulate their intensity (Randall et al., 2007). Higher-resolution global models have had some success in reproducing tropical cyclone-like vortices (e.g., Chauvin et al., 2006; Oouchi et al., 2006; Zhao et al., 2009), but only their coarse characteristics. Significant progress has been recently made, however, using downscaling techniques whereby high-resolution models capable of reproducing more realistic tropical cyclones are run using boundary conditions provided by either reanalysis data sets or output fields from lower-resolution climate models such as those used in the AR4 (e.g., Knutson et al., 2007; Emanuel et al., 2008; Knutson et al., 2008; Emanuel, 2010). A recent study by Bender et al. (2010) applies a cascading technique that downscales first from global to regional scale, and then uses the simulated storms from the regional model to initialize a very high-resolution hurricane forecasting model. These downscaling studies have been increasingly successful at reproducing observed tropical cyclone characteristics, which provides increased confidence in their projections, and it is expected that more progress will be made as computing resources improve. Still, awareness that limitations exist in the models used for tropical cyclone projections, particularly the ability to accurately reproduce natural climate phenomena

that are known to modulate storm behavior (e.g., ENSO and MJO), is important for context when interpreting model output (Sections 3.2.3.2 and 3.4.2).

While detection of long-term past increases in tropical cyclone activity is complicated by data quality and signal-to-noise issues (as stated above), theory (Emanuel, 1987) and idealized dynamical models (Knutson and Tuleya, 2004) both predict increases in tropical cyclone intensity under greenhouse warming. Recent simulations with high-resolution dynamical models (Oouchi et al., 2006; Bengtsson et al., 2007; Gualdi et al., 2008; Knutson et al., 2008; Sugi et al., 2009; Bender et al., 2010) and statistical-dynamical models (Emanuel, 2007) consistently find that greenhouse warming causes tropical cyclone intensity to shift toward stronger storms by the end of the 21st century (2 to 11% increase in mean maximum wind speed globally). These and other models also consistently project little change or a reduction in overall tropical cyclone frequency (e.g., Gualdi et al., 2008; Sugi et al., 2009; Murakami et al., 2011), but with an accompanying substantial fractional increase in the frequency of the strongest storms and increased precipitation rates (in the models for which these metrics were examined). Current models project changes in overall global frequency ranging from a decrease of 6 to 34% by the late 21st century (Knutson et al., 2010). The downscaling experiments of Bender et al. (2010) – which use an 18-model ensemble-mean of CMIP3 simulations to nudge a high-resolution dynamical model (Knutson et al., 2008) that is then used to initialize a very high-resolution dynamical model – project a 28% reduction in the overall frequency of Atlantic storms and an 80% increase in the frequency of Saffir-Simpson category 4 and 5 Atlantic hurricanes over the next 80 years (A1B scenario).

The projected decreases in global tropical cyclone frequency may be due to increases in vertical wind shear (Vecchi and Soden, 2007c; Zhao et al., 2009; Bender et al., 2010), a weakening of the tropical circulation (Sugi et al., 2002; Bengtsson et al., 2007) associated with a decrease in the upward mass flux accompanying deep convection (Held and Soden, 2006), or an increase in the saturation deficit of the middle troposphere (Emanuel et al., 2008). For individual basins, there is much more uncertainty in projections of tropical cyclone frequency, with changes of up to ±50% or more projected by various models (Knutson et al., 2010). When projected SST changes are considered in the absence of projected radiative forcing changes, Northern Hemisphere tropical cyclone frequency has been found to increase (Wehner et al., 2010), which is congruent with the hypothesis that SST changes alone do not capture the relevant physical mechanisms controlling tropical cyclogenesis (e.g., Emanuel, 2010).

As noted above, observed changes in rainfall associated with tropical cyclones have not been clearly established. However, as water vapor in the tropics increases (Trenberth et al., 2005) there is an expectation for increased heavy rainfall associated with tropical cyclones in response to associated moisture convergence increases (Held and Soden, 2006). This increase is expected to be compounded by increases in intensity as dynamical convergence under the storm is enhanced. Models in which

tropical cyclone precipitation rates have been examined are highly consistent in projecting increased rainfall within the area near the tropical cyclone center under 21st century warming, with increases of 3 to 37% (Knutson et al., 2010). Typical projected increases are near 20% within 100 km of storm centers.

Another type of projection that is sometimes inferred from the literature is based on extrapolation of an observed statistical relationship (see also Section 3.2.3). These relationships are typically constructed on past observed variability that represents a convolution of anthropogenically forced variability and natural variability across a broad range of time scales. In general, however, these relationships cannot be expected to represent all of the relevant physics that control the phenomena of interest, and their extrapolation beyond the range of the observed variability they are built on is not reliable. As an example, there is a strong observed correlation between local SST and tropical cyclone power dissipation (Emanuel, 2007). If 21st-century SST projections are applied to this relationship, power dissipation is projected to increase by about 300% in the next century (Vecchi et al., 2008; Knutson et al., 2010). Alternatively, there is a similarly strong relationship between power dissipation and relative SST, which represents the difference between local and tropical-mean SST and has been argued to serve as a proxy for local potential intensity (Vecchi and Soden, 2007a). When 21st-century projections of relative SST are considered, this latter relationship projects almost no change in power dissipation in the next century (Vecchi et al., 2006). Both of these statistical relationships can be reasonably defended based on physical arguments but it is not clear which, if either, is correct (Ramsay and Sobel, 2011).

When simulating 21st-century warming under the A1B emission scenario (or a close analog), the present models and downscaling techniques as a whole are consistent in projecting (1) decreases or no change in tropical cyclone frequency, (2) increases in intensity and fractional increases in number of most intense storms, and (3) increases in tropical cyclone-related rainfall rates. Differences in regional projections lead to lower confidence in basin-specific projections of intensity and rainfall, and confidence is particularly low for projections of frequency within individual basins. More specifically, while projections under 21st-century greenhouse warming indicate that it is *likely* that the global frequency of tropical cyclones will either decrease or remain essentially unchanged, an increase in mean tropical cyclone maximum wind speed is also *likely*, although increases may not occur in all tropical regions. This assessment is essentially identical with that of the recent WMO assessment (Knutson et al., 2010). Furthermore, while it is *likely* that overall global frequency will either decrease or remain essentially unchanged, it is *more likely than not* that the frequency of the most intense storms (e.g., Saffir-Simpson category 4 and 5) will increase substantially in some ocean basins, again agreeing with the recent WMO assessment (Knutson et al., 2010). Based on the level of consistency among models, and physical reasoning, it is *likely* that tropical cyclone-related rainfall rates will increase with greenhouse warming. Confidence in future projections for particular ocean basins is undermined by the inability of global models to reproduce accurate details at scales relevant to tropical cyclone

genesis, track, and intensity evolution. Of particular concern is the limited ability of global models to accurately simulate upper-tropospheric wind (Cordero and Forster, 2006; Bender et al., 2010), which modulates vertical wind shear and tropical cyclone genesis and intensity evolution. Thus there is *low confidence* in projections of changes in tropical cyclone genesis, location, tracks, duration, or areas of impact, and existing model projections do not show dramatic large-scale changes in these features.

In summary, there is *low confidence* that any observed long-term (i.e., 40 years or more) increases in tropical cyclone activity are robust, after accounting for past changes in observing capabilities. The uncertainties in the historical tropical cyclone records, the incomplete understanding of the physical mechanisms linking tropical cyclone metrics to climate change, and the degree of tropical cyclone variability provide only *low confidence* for the attribution of any detectable changes in tropical cyclone activity to anthropogenic influences. There is *low confidence* in projections of changes in tropical cyclone genesis, location, tracks, duration, or areas of impact. Based on the level of consistency among models, and physical reasoning, it is *likely* that tropical cyclone-related rainfall rates will increase with greenhouse warming. It is *likely* that the global frequency of tropical cyclones will either decrease or remain essentially unchanged. An increase in mean tropical cyclone maximum wind speed is *likely*, although increases may not occur in all tropical regions. While it is *likely* that overall global frequency will either decrease or remain essentially unchanged, it is *more likely than not* that the frequency of the most intense storms will increase substantially in some ocean basins.

3.4.5.   Extratropical Cyclones

Extratropical cyclones (synoptic-scale low-pressure systems) exist throughout the mid-latitudes in both hemispheres and mainly develop over the oceanic basins in the proximity of the upper-tropospheric jet streams, as a result of flow over mountains (lee cyclogenesis) or through conversions from tropical to extratropical systems. It should be noted that regionalized smaller-scale mid-latitude circulation phenomena such as polar lows and mesoscale cyclones are not treated in this section (but see Sections 3.3.3 and 3.4.3). Extratropical cyclones are the main poleward transporter of heat and moisture and may be accompanied by adverse weather conditions such as windstorms, the buildup of waves and storm surges, or extreme precipitation events. Thus, changes in the intensity of extratropical cyclones or a systematic shift in the geographical location of extratropical cyclone activity may have a great impact on a wide range of regional climate extremes as well as the long-term changes in temperature and precipitation. Extratropical cyclones mainly form and grow via atmospheric instabilities such as a disturbance along a zone of strong temperature contrast (baroclinic instabilities), which is a reservoir of available potential energy that can be converted into the kinetic energy associated with extratropical cyclones. Intensification of the cyclones may also take place due to processes such as release of energy due to

phase changes of water (latent heat release) (Gutowski et al., 1992; Wernli et al., 2002). Why should we expect climate change to influence extratropical cyclones? A simplified line of argument would be that both the large-scale low and high level pole to equator temperature gradients may change (possibly in opposite directions) in a climate change scenario leading to a change in the atmospheric instabilities responsible for cyclone formation and growth (baroclinicity). These changes may be induced by a variety of mechanisms operating in different parts of the atmospheric column ranging from changing surface conditions (Deser et al., 2007; Bader et al., 2011) to stratospheric changes (Son et al., 2010). In addition, changes in precipitation intensities within extratropical cyclones may change the latent heat release. According to theories on wave-mean flow interaction, changes in the extratropical storm tracks are also associated with changes in the large-scale flow (Robinson, 2000; Lorenz and Hartmann, 2003). A latitudinal shift of the upper tropospheric jet would be accompanied by a latitudinal shift in the extratropical storm track. It is, however, still unclear to what extent a latitudinal shift in the jet changes the total storm track activity rather than shifting it latitudinally (Wettstein and Wallace, 2010). Even within the very simplified outline above the possible impacts of climate change on extratropical cyclone development are many and clearly not trivial.

When validated using reanalyses with similar horizontal resolution, climate models are found to represent the general structure of the storm track pattern well (Bengtsson et al., 2006; Greeves et al., 2007; Ulbrich et al., 2008; Catto et al., 2010). However, using data from five different coupled models, the rate of transfer of zonal available potential energy to eddy available potential energy in synoptic systems was found to be too large, yielding too much energy and an overactive energy cycle (Marques et al., 2011). Models tend to have excessively zonal storm tracks and some show a poor extension of the storm tracks into Europe (Pinto et al., 2006; Greeves et al., 2007; Orsolini and Sorteberg, 2009). It has also been noted that representation of cyclone activity may depend on the physics formulations and the horizontal resolution of the model (Jung et al., 2006; Greeves et al., 2007).

Paleoclimatic proxies for extratropical cyclone variability are still few, but progress is being made in using coastal dune field development and sand grain content of peat bogs as proxies for storminess. Publications covering parts of western Europe indicate enhanced sand movement in European coastal areas during the Little Ice Age (Wilson et al., 2004; de Jong et al., 2006, 2007; Clemmensen et al., 2007; Clarke and Rendell, 2009; Sjogren, 2009). It should be noted that sand influx is also influenced by sediment availability, which is controlled mainly by the degree of vegetation cover and the moisture content of the sediment (Li et al., 2004; Wiggs et al., 2004). Intense cultivation, overgrazing, and forest disturbance make soils more prone to erosion, which can lead to increased sand transport even under less windy conditions. Thus the information gained from paleoclimatic proxies to put the last 100 years of extratropical cyclone variability in context is limited.

Century-long time-series of estimates of extremes in geostrophic wind deduced from triangles of pressure stations, pressure tendencies from

single stations (see Section 3.3.3 for details), or oceanic variables such as extremes in non-tide residuals are (if these are located in the vicinity of the main storm tracks) possible proxies for extratropical cyclone activity. Trend detection in extratropical cyclone variables such as number of cyclones, intensity, and activity (parameters integrating cyclone intensity, number, and possibly duration) became possible with the development of reanalyses, but remains challenging. Problems with reanalyses have been especially pronounced in the Southern Hemisphere (Hodges et al., 2003; Wang et al., 2006). Even though different reanalyses correspond well in the Northern Hemisphere (Hodges et al., 2003; Hanson et al., 2004), changes in the observing system giving artificial trends in integrated water vapor and kinetic energy (Bengtsson et al., 2004) may have influenced trends in both the number and intensity of cyclones. In addition, studies indicate that the magnitude and even the existence of the changes may depend on the choice of reanalysis (Trigo, 2006; Raible et al., 2008; Simmonds et al., 2008; Ulbrich et al., 2009) and cyclone tracking algorithm (Raible et al., 2008).

The AR4 noted a *likely* net increase in the frequency/intensity of Northern Hemisphere extreme extratropical cyclones and a poleward shift in the tracks since the 1950s (Trenberth et al., 2007; Table 3.8), and cited several papers showing increases in the number or strength of intense extratropical cyclones both over the North Pacific and the North Atlantic storm track (Trenberth et al., 2007, p. 312) during the last 50 years. Studies using reanalyses indicate a northward and eastward shift in the Atlantic cyclone activity during the last 60 years with both more frequent and more intense wintertime cyclones in the high-latitude Atlantic (Weisse et al., 2005; Wang et al., 2006; Schneidereit et al., 2007; Raible et al., 2008; Vilibic and Sepic, 2010) and fewer in the mid-latitude Atlantic (Wang et al., 2006; Raible et al., 2008). The increase in high-latitude cyclone activity was also reported in several studies of Arctic cyclone activity (X.D. Zhang et al., 2004; Sorteberg and Walsh, 2008; Sepp and Jaagus, 2011). Using ship-based trends in mean sea level pressure (MSLP) variance (which is tied to cyclone intensity), Chang (2007) found wintertime Atlantic trends to be consistent with National Centers for Environmental Prediction (NCEP) reanalysis trends in the Atlantic, but slightly weaker. There are inconsistencies among studies of extreme cyclones in reanalyses, since some studies show an increase in intensity and number of extreme Atlantic cyclones (Geng and Sugi, 2001; Paciorek et al., 2002; Lehmann et al., 2011) while others show a reduction (Gulev et al., 2001). These differences may in part be due to sensitivities of the identification schemes and different definitions of an extreme cyclone (Leckebusch et al., 2006; Pinto et al., 2006). New studies have confirmed that a positive NAM/NAO (see Section 3.4.3) corresponds to stronger Atlantic/European cyclone activity (e.g., Chang, 2009; Pinto et al., 2009; X.L. Wang et al., 2009b). However, studies using long historical records seem to suggest that some of these links may be statistically intermittent (Hanna et al., 2008; Matulla et al., 2008; Allan et al., 2009) due to interdecadal shifts in the location of the positions of the NAO pressure centers (Vicente-Serrano and Lopez-Moreno, 2008; X.D. Zhang et al., 2008). It is unclear to what extent the statistical intermittency implies that the underlying physical processes creating the connection act only intermittently. A possible influence of the Pacific North America (PNA)

pattern on the entrance of the North Atlantic storm track (over Newfoundland) has been reported by Pinto et al. (2011). It should be noted that there is some suggestion that the reanalyses cover a time period that starts with relatively low cyclonic activity in northern coastal Europe in the 1960s and reaches a maximum in the 1990s. Long-term European storminess proxies show no clear trends over the last century (Hanna et al., 2008; Allan et al., 2009; see Section 3.3.3 for details).

Studies using reanalyses and *in situ* data for the last 50 years have noted an increase in the number and intensity of north Pacific wintertime intense extratropical cyclone systems since the 1950s (Graham and Diaz, 2001; Simmonds and Keay, 2002; Raible et al., 2008) and cyclone activity (X.D. Zhang et al., 2004), but signs of some of the trends disagreed when different tracking algorithms or reanalysis products were used (Raible et al., 2008). A slight positive trend has been found in north Pacific extreme cyclones (Geng and Sugi, 2001; Gulev et al., 2001; Paciorek et al., 2002). Using ship measurements, Chang (2007) found intensity-related wintertime trends in the Pacific to be about 20 to 60% of that found in the reanalysis. Long-term *in situ* observations of north Pacific cyclones based on observed pressure data are considerably fewer than for coastal Europe. However, using hourly tide gauge records from the western coast of the United States as a proxy for storminess, an increasing trend in the extreme winter Non-Tide Residuals (NTR) has been observed in the last decades (Bromirski et al., 2003; Menendez et al., 2008). Years having high NTR were linked to a large-scale atmospheric circulation pattern, with intense storminess associated with a broad, south-easterly displaced, deep Aleutian low that directed storm tracks toward the US West Coast. North Pacific cyclonic activity has been linked to tropical SST anomalies (NINO3.4; see Section 3.4.2) and the PNA (Eichler and Higgins, 2006; Favre and Gershunov, 2006; Seierstad et al., 2007), showing that the PNA and NINO3.4 influence storminess, in particular over the eastern North Pacific with an equatorward shift in storm tracks in the North Pacific basin, as well as an increase in storm track activity along the US East Coast during El Niño events.

Based on reanalyses, North American cyclone numbers have increased over the last 50 years, with no statistically significant change in cyclone intensity (X.D. Zhang et al., 2004). Hourly MSLP data from Canadian stations showed that winter cyclones have become significantly more frequent, longer lasting, and stronger in the lower Canadian Arctic over the last 50 years (1953-2002), but less frequent and weaker in the south, especially along the southeast and southwest Canadian coasts (Wang et al., 2006). Further south, a tendency toward weaker low-pressure systems over the past few decades was found for US East Coast winter cyclones using reanalyses, but no statistically significant trends in the frequency of occurrence of systems (Hirsch et al., 2001).

Studies on extratropical cyclone activity in northern Asia are few. Using reanalyses, a decrease in extratropical cyclone activity (X.D. Zhang et al., 2004) and intensity (X.D. Zhang et al., 2004; X. Wang et al., 2009) over the last 50 years has been reported for northern Eurasia (60-40°N) with a possible northward shift with increased cyclone frequency in the higher latitudes (50-45°N) and decrease in the lower latitudes (south of 45°N),

based on a study with reanalyses. The low-latitude (south of 45°N) decrease was also noted by Zou et al. (2006), who reported a decrease in the number of severe storms for mainland China based on an analysis of extremes of observed 6-hourly pressure tendencies over the last 50 years.

Alexander and Power (2009) showed that the number of observed severe storms at Cape Otway (south-east Australia) has decreased since the mid-19th century, strengthening the evidence of a southward shift in Southern Hemisphere storm tracks previously noted using reanalyses (Fyfe, 2003; Hope et al., 2006; Wang et al., 2006). Frederiksen and Frederiksen (2007) linked the reduction in cyclogenesis at 30°S and southward shift to a decrease in the vertical mean meridional temperature gradient. Using reanalyses, both Pezza et al. (2007) and Lim and Simmonds (2009) have confirmed previous studies showing a trend toward more intense low-pressure systems. However, the trend of a decreasing number of cyclones seems to depend on the choice of reanalysis and pressure level (Lim and Simmonds, 2009), emphasizing the weaker consistency among reanalysis products for the Southern Hemisphere extratropical cyclones. Recent studies support the notion of more cyclones around Antarctica when the SAM (see Section 3.4.3) is in its positive phase and a shift of cyclones toward mid-latitudes when the SAM is in its negative phase (Pezza and Simmonds, 2008). Additionally, more intense (and fewer) cyclones seem to occur when the PDO (see Section 3.4.3) is strongly positive and vice versa (Pezza et al., 2007).

In conclusion, it is *likely* that there has been a poleward shift in the main northern and southern storm tracks during the last 50 years. There is strong agreement with respect to this change between several reanalysis products for a wide selection of cyclone parameters and cyclone identification methods and European and Australian pressure-based storminess proxies are consistent with a poleward shift over the last 50 years, which indicates that the evidence is robust. Advances have been made in documenting the observed decadal and multi-decadal variability of extratropical cyclones using proxies for storminess. So the recent poleward shift should be seen in light of new studies with longer time spans that indicate that the last 50 years coincide with relatively low cyclonic activity in northern coastal Europe in the beginning of the period. Several studies using reanalyses suggest an intensification of high-latitude cyclones, but there is still insufficient knowledge of how changes in the observational systems are influencing the cyclone intensification in reanalyses so even in cases of high agreement among the studies the evidence cannot be considered to be robust, thus we have only *low confidence* in these changes. Other regional changes in intensity and the number of cyclones have been reported. However, the level of agreement between different studies using different tracking algorithms, different reanalyses, or different cyclone parameters is still low. Thus, we have *low confidence* in the amplitude, and in some regions in the sign, of the regional changes.

Regarding possible causes of the observed poleward shift, the AR4 concluded that trends over recent decades in the Northern and Southern Annular Modes, which correspond to sea level pressure reductions over the poles, are *likely* related in part to human activity, but an anthropogenic influence on extratropical cyclones had not been formally detected, owing to large internal variability and problems due to changes in observing systems (Hegerl et al., 2007). Anthropogenic influences on these modes of variability are also discussed in Section 3.4.3.

Seasonal global sea level pressure changes have been shown to be inconsistent with simulated internal variability (Giannini et al., 2003; Gillett et al., 2005; Gillett and Stott, 2009; X.L. Wang et al., 2009a), but changes in sea level pressure in regions of extratropical cyclones (mid- and high latitudes) have not formally been attributed to anthropogenic forcings (Gillett and Stott, 2009). However, the trend pattern in atmospheric storminess as inferred from geostrophic wind energy and ocean wave heights has been found to contain a detectable response to anthropogenic and natural forcings with the effect of external forcings being strongest in the winter hemisphere (X.L. Wang et al., 2009a). Nevertheless, the models generally simulate smaller changes than observed and also appear to underestimate the internal variability, reducing the robustness of their detection results. New idealized studies have advanced the physical understanding of how storm tracks may respond to changes in the underlying surface conditions, indicating that a uniform SST increase weakens (reduced cyclone intensity or number of cyclones) and shifts the storm track poleward and strengthened SST gradients near the subtropical jet may lead to a meridional shift in the storm track either toward the poles or the equator depending on the location of the SST gradient change (Deser et al., 2007; Brayshaw et al., 2008; Semmler et al., 2008; Kodama and Iwasaki, 2009), but the average global cyclone activity is not expected to change much under moderate greenhouse gas forcing (O'Gorman and Schneider, 2008; Bengtsson et al., 2009). Studies have also emphasized the important role of stratospheric changes (induced by ozone or greenhouse gas changes) in explaining latitudinal shifts in storm tracks and several mechanisms have been proposed (Son et al., 2010). This has particularly strengthened the understanding of the Southern Hemisphere changes. According to Fogt et al. (2009) both coupled climate models and observed trends in the SAM were found to be outside the range of internal climate variability during the austral summer. This was mainly attributed to stratospheric ozone depletion (see Section 3.4.3).

In summary, there is *medium confidence* in an anthropogenic influence on the observed poleward shift in extratropical cyclone activity. It has not formally been attributed. However indirect evidence such as global anthropogenic influence on the sea level pressure distribution and trend patterns in atmospheric storminess inferred from geostrophic wind and ocean wave heights has been found. While physical understanding of how anthropogenic forcings may influence extratropical cyclone storm tracks has strengthened, the importance of the different mechanisms in the observed shifts is still unclear.

The AR4 reported that in a future warmer climate, a consistent projection from the majority of the coupled atmosphere-ocean GCMs is fewer

mid-latitude storms averaged over each hemisphere (Meehl et al., 2007b) and a poleward shift of storm tracks in both hemispheres (particularly evident in the Southern Hemisphere), with greater storm activity at higher latitudes (Meehl et al., 2007b).

A poleward shift in the upper level tropospheric storm track due to increased greenhouse gas forcing is supported by post-AR4 studies (Lorenz and DeWeaver, 2007; O'Gorman, 2010; Wu et al., 2011). It should be noted that other studies indicate that the poleward shift is less clear when models including a full stratosphere or ozone recovery are used (Huebener et al., 2007; Son et al., 2008; Morgenstern et al., 2010; Scaife et al., 2011) and the strength of the poleward shift is often seen more clearly in upper-level quantities than in low-level transient parameters (Ulbrich et al., 2008). Post-AR4 single model studies support the projection of a reduction in extratropical cyclones averaged over the Northern Hemisphere during future warming (Finnis et al., 2007; Bengtsson et al., 2009; Orsolini and Sorteberg, 2009). However, neither the global changes in storm frequency or intensity were found to be statistically significant by Bengtsson et al. (2009), although they were accompanied by significant increases in total and extreme precipitation.

Models tend to project a reduction of winter cyclone activity throughout the mid-latitude North Pacific and for some models a north-eastern movement of the North Pacific storm track (Loeptien et al., 2008; Ulbrich et al., 2008; Favre and Gershunov, 2009; McDonald, 2011). However, the exact geographical pattern of cyclone frequency anomalies exhibits large variations across models (Teng et al., 2008; Favre and Gershunov, 2009; Laine et al., 2009).

Using band-passed sea level pressure data from 16 CMIP3 coupled GCMs, Ulbrich et al. (2008) showed regional increases in the storm track activity over the Eastern North Atlantic/Western European area. This eastward or southeastward extension of the storm track is also found in other studies (Ulbrich et al., 2008; Laine et al., 2009; McDonald, 2011) and may be attributed to a local minimum in ocean warming in the central North Atlantic and subsequent local changes in baroclinicity (McDonald, 2011). In line with the eastward shift, Donat et al. (2010a) projected an increase in wind storm days for central Europe by the end of the 21st century. The increase varies according to the definition of storminess and one model projects a decrease. A common deficiency among many AR4 models is a coarsely resolved stratosphere and there are still concerns that this may lead to systematic biases in the Atlantic storm track response to increased anthropogenic forcing (Scaife et al., 2011). A reduction in cyclone frequency along the Canadian east coast has been reported (Bengtsson et al., 2006; Watterson, 2006; Pinto et al., 2007a; Teng et al., 2008; Long et al., 2009). New results for Southern Hemisphere cyclones confirm the previously projected poleward shift in storm tracks under increased greenhouse gases (Lim and Simmonds, 2009). That study projected a reduction of Southern Hemisphere extratropical cyclone frequency and intensity in mid-latitudes but a slight increase at high latitudes. The poleward shift due to increased greenhouse gases may be partly opposed by ozone recovery (Son et al., 2010).

Detailed analyses of changes in physical mechanisms related to cyclone changes in coupled climate models are still few. O'Gorman (2010) showed that changes in mean available potential energy of the atmosphere can account for much of the varied response in storm-track intensity to global warming, implying that changes in storm-track intensity are sensitive to competing effects of changes in temperature gradients and static stability in different atmospheric levels. Using two coupled climate models, Laine et al. (2009) indicate that the primary cause for synoptic activity changes at the western end of the Northern Hemisphere storm tracks is related to the baroclinic conversion processes linked to mean temperature gradient changes in localized regions of the western oceanic basins. They also found downstream changes in latent heat release during the developing and mature stages of the cyclone to be of importance and indicated that changes in diabatic process may be amplified by the upstream baroclinic changes [stronger (weaker) baroclinic activity in the west gives stronger (weaker) latent heat release downstream]. Pinto et al. (2009) found that regional increases in track density and intensity of extreme cyclones close to the British Isles using a single model was associated with an eastward shift of the jet stream into Europe, more frequent extreme values of baroclinicity, and stronger upper level divergence.

The modeled reduction in Southern Hemisphere extratropical cyclone frequency and intensity in the mid-latitudes has been attributed to the tropical upper tropospheric warming enhancing static stability and decreasing baroclinicity while an increased meridional temperature gradient in the high latitudes is suggested to be responsible for the increase in cyclone activity in this region (Lim and Simmonds, 2009). In addition to details in the modeled changes in local baroclinicity and diabatic changes, the geographical pattern of modeled response in cyclone activity has been reported to be influenced by the individual model's structure of intrinsic modes of variability (Branstator and Selten, 2009) and biases in the climatology (Kidston and Gerber, 2010).

In summary it is *likely* that there has been a poleward shift in the main Northern and Southern Hemisphere extratropical storm tracks during the last 50 years. There is *medium confidence* in an anthropogenic influence on this observed poleward shift. It has not formally been attributed. There is *low confidence* in past changes in regional intensity. There is *medium confidence* that an increased anthropogenic forcing will lead to a reduction in the number of mid-latitude cyclones averaged over each hemisphere, and there is also *medium confidence* in a poleward shift of the tropospheric storm tracks due to future anthropogenic forcings. Regional changes may be substantial and CMIP3 simulations show some regions with medium agreement. However, there are still uncertainties related to how the poorly resolved stratosphere in many CMIP3 models may influence the regional results. In addition, studies using different analysis techniques, different physical quantities, different thresholds, and different atmospheric vertical levels to represent cyclone activity and storm tracks result in different projections of regional changes. This leads to *low confidence* in region-specific projections.

3.5. Observed and Projected Impacts on the Natural Physical Environment

3.5.1. Droughts

Drought is generally "a period of abnormally dry weather long enough to cause a serious hydrological imbalance" (see the Glossary and Box 3-3). While lack of precipitation (i.e., meteorological drought; Box 3-3) is often the primary cause of drought, increased potential evapotranspiration induced by enhanced radiation, wind speed, or vapor pressure deficit (itself linked to temperature and relative humidity), as well as pre-conditioning (pre-event soil moisture; lake, snow, and/or groundwater storage) can contribute to the emergence of soil moisture and hydrological drought (Box 3-3). Actual evapotranspiration is additionally controlled by soil moisture, which constitutes a limiting factor for further drying under drought conditions, and other processes that impact vegetation development and phenology (e.g., temperature) are also relevant. As noted in the AR4 (Trenberth et al., 2007), there are few direct observations of drought-related variables, in particular of soil moisture, available for a global analysis (see also Section 3.2.1). Hence, proxies for drought ('drought indices') are often used to infer changes in drought conditions. Box 3-3 provides a discussion of the issue of drought definition and a description of commonly used drought indices. In order to understand the impact of droughts (e.g., on crop yields, general ecosystem functioning, water resources, and electricity production), their timing, duration, intensity, and spatial extent need to be characterized. Several weather elements may interact to increase the impact of droughts: enhanced air temperature can indirectly lead to enhanced evaporative demand (through enhanced vapor pressure deficit), although enhanced wind speed or increased incoming radiation are generally more important factors. Moreover, climate phenomena such as monsoons (Section 3.4.1) and ENSO (Section 3.4.2) affect changes in drought occurrence in some

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Box 3-3 | The Definition of Drought

Though a commonly used term, drought is defined in various ways, and these definitional issues make the analysis of changes in drought characteristics difficult. This explains why assessments of (past or projected) changes in drought can substantially differ between published studies or chosen indices (see Section 3.5.1). Some of these difficulties and their causes are highlighted in this box.

What is Drought or Dryness?
The Glossary defines drought as follows: "A period of abnormally dry weather long enough to cause a serious hydrological imbalance. Drought is a relative term, therefore any discussion in terms of precipitation deficit must refer to the particular precipitation-related activity that is under discussion. For example, shortage of precipitation during the growing season impinges on crop production or ecosystem function in general (due to soil moisture drought, also termed agricultural drought), and during the runoff and percolation season primarily affects water supplies (hydrological drought). Storage changes in soil moisture and groundwater are also affected by increases in actual evapotranspiration in addition to reductions in precipitation. A period with an abnormal precipitation deficit is defined as a meteorological drought. A megadrought is a very lengthy and pervasive drought, lasting much longer than normal, usually a decade or more."

As highlighted in the above definition, drought can be defined from different perspectives, depending on the stakeholders involved. The scientific literature commonly distinguishes *meteorological drought*, which refers to a deficit of precipitation, *soil moisture drought* (often called *agricultural drought*), which refers to a deficit of (mostly root zone) soil moisture, and *hydrological drought*, which refers to negative anomalies in streamflow, lake, and/or groundwater levels (e.g., Heim Jr., 2002). We use here the term 'soil moisture drought' instead of 'agricultural drought,' despite the widespread use of the latter term (e.g., Heim Jr., 2002; Wang, 2005), because soil moisture deficits have several additional effects beside those on agroecosystems, most importantly on other natural or managed ecosystems (including both forests and pastures), on building infrastructure through soil mechanical processes (e.g., Corti et al., 2009), and health through impacts on heat waves (Section 3.1.4). Water scarcity (linked to *socioeconomic drought*), which may be caused fully or in part by use from human activities, does not lie within the scope of this chapter (see Section 4.2.2); however, it should be noted that changing pressure on water resources by human uses may itself influence climate and possibly the drought conditions, for example, via declining groundwater levels, or enhanced local evapotranspiration and associated land-atmosphere feedbacks. Drought should not be confused with aridity, which describes the general characteristic of an arid climate (e.g., desert). Indeed, drought is considered a recurring feature of climate occurring in any region and is defined with respect to the average climate of the given region (e.g., Heim Jr., 2002; Dai, 2011). Nonetheless, the effects of droughts are not linear, given the existence of, for example, discrete soil moisture thresholds affecting vegetation and surface fluxes (e.g., Koster et al., 2004b; Seneviratne et al., 2010), which means that the same precipitation deficit or radiation excess relative to normal will not affect different regions equally (e.g., short-term lack of precipitation in a very humid region may not be critical for agriculture because of the ample soil moisture supply). In this chapter we often use the term 'dryness' instead of 'drought' as a more general term.

Drought Drivers

For soil moisture or hydrological droughts, the main drivers are reduced precipitation and/or increased evapotranspiration (Figure 3-9). Although the role of deficits in precipitation is generally considered more prominently in the literature, several drought indicators also explicitly or indirectly consider effects of evapotranspiration. In the context of climate projections, analyses suggest that changes in simulated soil moisture drought are mostly driven by changes in precipitation, with increased evapotranspiration from higher vapor pressure deficit (often linked to increased temperature) and available radiation modulating some of the changes (e.g., Burke and Brown, 2008; Sheffield and Wood, 2008a; Orlowsky and Seneviratne, 2011). It should nonetheless be noted that under strong drought conditions, soil moisture becomes limiting for evapotranspiration, thus limits further soil moisture depletion. Other important aspects for soil moisture and hydrological droughts are persistence and pre-conditioning. Because soil moisture, groundwater, and surface waters are associated with water storage, they have a characteristic memory (e.g., Vinnikov et al., 1996; Eltahir and Yeh, 1999; Koster and Suarez, 2001; Seneviratne et al., 2006b) and thus specific response times to drought forcing (e.g., Begueria et al., 2010; Fleig et al., 2011). The memory is also a function of the atmospheric forcing and system's feedbacks (Koster and Suarez, 2001; A.H. Wang et al., 2009), and the relevant storage is dependent on soil characteristics and rooting depth of the considered ecosystems. This means that drought has a different persistence depending on the affected system, and that it is also sensitive to pre-conditioning (Figure 3-9). Effects of pre-conditioning also explain the possible occurrence of multi-year droughts, whereby soil moisture anomalies can be carried over from one year to the next (e.g., Wang, 2005). However, other features can induce drought persistence, such as persistent circulation anomalies, possibly strengthened by land-atmosphere feedbacks (Schubert et al., 2004; Rowell and Jones, 2006). The choice of variable (e.g., precipitation, soil moisture, or streamflow) and time scale can strongly affect the ranking of drought events (Vidal et al., 2010).

Figure 3-9 | Simplified sketch of processes and drivers relevant for meteorological, soil moisture (agricultural), and hydrological droughts.

Drought Indices

Because of the complex definition of droughts, and the lack of soil moisture observations (Section 3.2.1), several indices have been developed to characterize (meteorological, soil moisture, and hydrological) drought (see, e.g., Heim Jr., 2002; Dai, 2011). These indicators include land surface, hydrological, or climate model simulations (providing estimates of, e.g., soil moisture or runoff) and indices based on measured meteorological or hydrological variables. We provide here a brief overview of the wide range of drought indices used in the literature for the analysis of recent and projected changes. Note that information on paleoclimate proxies such as tree rings, speleothems, lake sediments, or historical evidence (e.g., harvest dates) is not detailed here.

Some indices are based solely on precipitation data. A widely used index is the Standard Precipitation Index (SPI) (McKee et al., 1993; Lloyd-Hughes and Saunders, 2002), which consists of fitting and transforming a long-term precipitation record into a normal distribution that has zero mean and unit standard deviation. SPI values of -0.5 to -1 correspond to mild droughts, -1 to -1.5 to moderate droughts, -1.5 to -2 to severe droughts, and below -2 to extreme droughts. Similarly, values from 0 to 2 correspond to mildly wet to severely wet conditions, and values above 2 to extremely wet conditions (Lloyd-Hughes and Saunders, 2002). SPI can be computed over several time scales (e.g., 3, 6, 12, or more months) and thus indirectly considers effects of accumulating precipitation deficits, which are critical for soil moisture and hydrological droughts. Another index commonly used in the analysis of climate model simulations is the Consecutive Dry Days (CDD) index, which considers the maximum consecutive number of days without rain (i.e., below a given threshold, typically 1 mm day^{-1}) within a considered period (i.e., year in general; Frich et al., 2002; Alexander et al., 2006; Tebaldi et al., 2006). For seasonal time frames, the CDD periods can either be considered to be bound to the respective seasons (e.g., Figure 3-10) or considered in their entirety (across seasons) but assigned to a specific season. Though SPI and CDD are both only based on precipitation, they do not necessarily only consider the effects of meteorological drought, since periods without rain (thus less cloud cover) are bound to have higher daytime radiation forcing and generally higher temperatures, thus possibly positive evapotranspiration anomalies (unless soil moisture conditions are too dry and limit evapotranspiration).

Some indices reflect both precipitation and estimates of actual or potential evapotranspiration, in some cases also accounting for some temporal accumulation of the forcings or persistence of the drought anomalies. These include the Palmer Drought Severity Index (PDSI)

(Palmer, 1965), which measures the departure of moisture balance from normal conditions using a simple water balance model (e.g., Dai, 2011), as well as other indices such as the Precipitation Potential Evaporation Anomaly (PPEA, based on the cumulative difference between precipitation and potential evapotranspiration) used in Burke and Brown (2008) and the Standardized Precipitation-Evapotranspiration Index (SPEI, which considers cumulated anomalies of precipitation and potential evapotranspiration) described in Vicente-Serrano et al. (2010). PDSI has been widely used for decades (in particular in the United States), and also in climate change analyses (e.g., Dai et al., 2004; Burke and Brown, 2008; Dai, 2011); however, it has some shortcomings for climate change monitoring and projection. PDSI was originally calibrated for the central United States, which can impair the comparability of the index across regions (and also across time periods if drought mechanisms change over time). Thus it is often of advantage to renormalize the local PDSI (Dai, 2011), which can also be done using the self-calibrated PDSI (Wells et al., 2004), but several studies do not apply these steps. Moreover, the land surface model underlying the computation of the PDSI is essentially a simple bucket-type model, which is less sophisticated than more recent land surface and hydrological models and thus implies several limitations (e.g., Dai et al., 2004; Burke et al., 2006). Another important issue is that the parameterization of potential evapotranspiration as empirically (and solely) dependent on air temperature, which is often applied for these various indices (e.g., in the study of Dai et al., 2004) can lead to biased results (e.g., Donohue et al., 2010; Milly and Dunne, 2011; Shaw and Riha, 2011). Temperature is only an indirect driver of evapotranspiration, via its effect on vapor pressure deficit and via effects on vegetation phenology. Furthermore, approaches using potential evapotranspiration as a proxy for actual evapotranspiration do not consider soil moisture and vegetation control on evapotranspiration, which are important mechanisms limiting drought development.

For the assessment of soil moisture drought, simulated soil moisture anomalies also can be considered (Wang et al., 2005; Burke and Brown, 2008; Sheffield and Wood, 2008a; A.H. Wang et al., 2009; Dai, 2011; Orlowsky and Seneviratne, 2011). Simulated soil moisture anomalies integrate the effects of precipitation forcing, simulated actual evapotranspiration (resulting from atmospheric forcing and simulated soil moisture limitation on evapotranspiration), and simulated soil moisture persistence. Although the soil moisture simulated by (land-surface, hydrological, and climate) models often exhibits strong discrepancies in absolute terms, soil moisture anomalies can be compared with simple scaling and generally match reasonably well (e.g., Koster et al., 2009; A.H. Wang et al., 2009). Soil moisture persistence is found to be an important component in projected changes in soil moisture drought, with some regions displaying year-round dryness compared to reference (late 20th or pre-industrial) conditions due to the carry-over effect of soil moisture storage from season to season, leading to year-round soil moisture deficits compared to late 20th century climate (e.g., Wang et al., 2005, Figure 3-10). However, it should be noted that some land surface and hydrological models (used offline or coupled to climate models) suffer from similar shortcomings as noted above for PDSI – that is, they use simple bucket models or simplified representations of potential evapotranspiration. The latter issue has been suggested as being particularly critical for models used in offline mode (Milly and Dunne, 2011). Nonetheless, for the assessment of soil moisture drought, using simulated soil moisture anomalies seems less problematic than many other indices for the reasons highlighted in the above paragraphs.

The indices listed above have been used in various studies analyzing drought in the context of climate change, but with a few exceptions most available studies are based only on one index, which makes their comparison difficult. Nonetheless, these studies suggest that projections can be highly dependent on the choice of drought index. For instance, one study projected changes in drought area possibly varying between a negligible impact and a 5 to 45% increase depending on the drought index considered (Burke and Brown, 2008). Other drought indices are used to quantify hydrological drought (e.g., Heim Jr., 2002; Vidal et al., 2010; Dai, 2011), but are less commonly used in climate change studies. Further analyses or indices also consider the area affected by droughts (e.g., Burke et al., 2006; Sheffield and Wood, 2008a; Dai, 2011) or additional variables (such as snow or vegetation indices from satellite measurements, e.g., Heim Jr., 2002). As for the definition of other indices (Box 3-1), the determination of the reference period is critical for the assessment of changes in drought patterns independently of the chosen index. In general, late 20th-century conditions are used as reference (e.g., Figure 3-10).

In summary, drought indices often integrate precipitation, temperature, and other variables, but may emphasize different aspects of drought and should be carefully selected with respect to the drought characteristic in mind. In particular, some indices have specific shortcomings, especially in the context of climate change. For this reason, assessments of changes in drought characteristics with climate change should consider several indices including a specific evaluation of their relevance to the addressed question to support robust conclusions. In this assessment we focus on the following indices: consecutive dry days (CDD) and simulated soil moisture anomalies (SMA), although evidence based on other indices (e.g., PDSI for present climate) is also considered (Section 3.5.1; Tables 3-2 and 3-3).

regions. Hence, drought is a complex phenomenon that is strongly affected by other extremes considered in this chapter, but that is also affected by changes in mean climate features (Section 3.1.6). In addition, via land-atmosphere interactions, drought also has the potential to impact other weather and climate elements such as temperature and precipitation and associated extremes (Koster et al., 2004b; Seneviratne et al., 2006a; Hirschi et al., 2011; see also Section 3.1.4). Case Study 9.2.3 addresses aspects related to the management of adverse consequences of droughts; while Case Study 9.2.2 considers the possible impacts of high temperatures and drought on wildfire.

Observed Changes

There are still large uncertainties regarding observed global-scale trends in droughts. The AR4 reported based on analyses using PDSI (see Box 3-3) that very dry areas had more than doubled in extent since 1970 at the global scale (Trenberth et al., 2007). This assessment was, however, largely based on the study by Dai et al. (2004) only. These trends in the PDSI proxy were found to be largely affected by changes in temperature, not precipitation (Dai et al., 2004). On the other hand, based on soil moisture simulations with an observation-driven land surface model for the time period 1950-2000, Sheffield and Wood (2008a) have inferred trends in drought duration, intensity, and severity predominantly decreasing, but with strong regional variation and including increases in some regions. They concluded that there was an overall moistening trend over the considered time period, but also a switch since the 1970s to a drying trend, globally and in many regions, especially in high northern latitudes. Some regional studies are consistent with the results from Sheffield and Wood (2008a), regarding, for example, less widespread increase (or statistically insignificant changes or decreases) in some regions compared to the study of Dai et al. (2004) (e.g., in Europe, see below). More recently, Dai (2011) by extending the record did, however, find widespread increases in drought both based on various versions of PDSI (for 1950-2008) and soil moisture output from a land surface model (for 1948-2004). Hence there are still large uncertainties with respect to global assessments of past changes in droughts. Nonetheless, there is some agreement between studies over the different time frames (i.e., since 1950 versus 1970) and using different drought indicators regarding increasing drought occurrence in some regions (e.g., southern Europe, West Africa; see below and Table 3-2), although other regions also indicate opposite trends (e.g., central North America, northwestern Australia; see below and Table 3-2). As mentioned in Section 3.1.6, spatially coherent shifts in drought regimes are expected with changing global circulation patterns. Table 3-2 provides regional and continental-scale assessments of observed trends in dryness based on different indices (Box 3-3). The following paragraphs provide more details by continent.

From a paleoclimate perspective recent droughts are not unprecedented, with severe 'megadroughts' reported in the paleoclimatic record for Europe, North America, and Australia (Jansen et al., 2007). Recent studies extend this observation to African and Indian droughts (Sinha et al., 2007; Shanahan et al., 2009): much more severe and longer droughts

occurred in the past centuries with widespread ecological, political, and socioeconomic consequences. Overall, these studies confirm that in the last millennium several extreme droughts have occurred (Breda and Badeau, 2008; Kallis, 2008; Büntgen et al., 2010).

In North America, there is *medium confidence* that there has been an overall slight tendency toward less dryness (wetting trend with more soil moisture and runoff; Table 3-2), although analyses for some subregions also indicate tendencies toward increasing dryness. This assessment is based on several lines of evidence, including simulations with different hydrological models as well as PDSI and CDD estimates (Alexander et al., 2006; Andreadis and Lettenmaier, 2006; van der Schrier et al., 2006a; Kunkel et al., 2008; Sheffield and Wood, 2008a; Dai, 2011). The most severe droughts in the 20th century have occurred in the 1930s and 1950s, where the 1930s Dust Bowl was most intense and the 1950s drought most persistent (Andreadis et al., 2005) in the United States, while in Mexico the 1950s and late 1990s were the driest periods. Recent regional trends toward more severe drought conditions were identified over southern and western Canada, Alaska, and Mexico, with subregional exceptions (Dai, 2011).

In Europe, there is *medium confidence* regarding increases in dryness based on some indices in the southern part of the continent, but large inconsistencies between indices in this region, and inconsistent or statistically insignificant trends in the rest of the continent (Table 3-2). Although Dai et al. (2004) found an increase in dryness for most of the European continent based on PDSI, Lloyd-Hughes and Saunders (2002) and van der Schrier et al. (2006b) concluded, based on the analysis of SPI and self-calibrating PDSI for the 20th century (for 1901-1999 and 1901-2002, respectively), that no statistically significant changes were observed in extreme and moderate drought conditions in Europe [with the exception of the Mediterranean region in van der Schrier et al. (2006b)]. Sheffield and Wood (2008a) also found contrasting dryness trends in Europe, with increases in the southern and eastern part of the continent, but decreases elsewhere. Beniston (2009b) reported a strong increase in warm-dry conditions over all central-southern (including maritime) Europe via a quartile analysis from the middle to the end of the 20th century. Alexander et al. (2006) found trends toward increasing CDD mostly in the southern and central part of the continent. Trends of decreasing precipitation and discharge are consistent with increasing salinity in the Mediterranean Sea, indicating a trend toward freshwater deficits (Mariotti et al., 2008), but this could also be partly caused by increased human water use. In France, an analysis based on a variation of the PDSI model also reported a significant increasing trend in drought conditions, in particular from the 1990s onward (Corti et al., 2009). Stahl et al. (2010) investigated streamflow data across Europe and found negative trends (lower streamflow) in southern and eastern regions, and generally positive trends (higher streamflow) elsewhere (especially in northern latitudes). Low flows have decreased in most regions where the lowest mean monthly flow occurs in summer, but vary for catchments that have flow minima in winter and secondary low flows in summer. The exceptional 2003 summer heat wave on the European continent (see Section 3.3.1) was also associated with a

major soil moisture drought, as could be inferred from satellite measurements (Andersen et al., 2005), model simulations (Fischer et al., 2007a,b), and impacts on ecosystems (Ciais et al., 2005; Reichstein et al., 2007).

There is *low confidence* in dryness trends in South America (Table 3-2), partly due to lack of data and partly due to inconsistencies. For the Amazon, repeated intense droughts have been occurring in the last decades but no particular trend has been reported. The 2005 and 2010 droughts in Amazonia are, however, considered the strongest in the last century as inferred from integrating precipitation records and water storage estimates via satellite (measurements from the Gravity Recovery and Climate Experiment; Chen et al., 2009; Lewis et al., 2011). For other parts of South America, analyses of the return intervals between droughts in the instrumental and reconstructed precipitation series indicate that the probability of drought has increased during the late 19th and 20th centuries, consistent with selected long instrumental precipitation records and with a recession of glaciers in the Chilean and Argentinean Andean Cordillera (Le Quesne et al., 2006, 2009).

Changes in drought patterns have been reported for the monsoon regions of Asia and Africa with variations at the decadal time scale (e.g., Janicot, 2009). In Asia there is overall *low confidence* in trends in dryness both at the continental and regional scale, mostly due to spatially varying trends, except in East Asia where a range of studies, based on different indices, show increasing dryness in the second half of the 20th century, leading to *medium confidence* (Table 3-2).

In the Sahel, recent years have been characterized by greater interannual variability than the previous 40 years (Ali and Lebel, 2009; Greene et al., 2009), and by a contrast between the western Sahel remaining dry and the eastern Sahel returning to wetter conditions (Ali and Lebel, 2009). Giannini et al. (2008) report a drying of the African monsoon regions, related to warming of the tropical oceans, and variability related to ENSO. In the different subregions of Africa there is overall *low* to *medium confidence* regarding regional dryness trends (Table 3-2).

For Australia, Sheffield and Wood (2008a) found very limited increases in dryness from 1950 to 2000 based on soil moisture simulated using existing climate forcing (mostly in southeastern Australia) and some marked decreases in dryness in central Australia and the northwestern part of the continent. Dai (2011), for an extended period until 2008 and using different PDSI variants as well as soil moisture output from a land surface model, found a more extended drying trend in the eastern half of the continent, but also a decrease in dryness in most of the western half. Jung et al. (2010) inferred from a combination of remote sensing and quasi-globally distributed eddy covariance flux observations that in particular the decade after 1998 became drier in Australia (and parts of Africa and South America), leading to decreased evapotranspiration, but it is not clear if this is a trend or just decadal variation.

Following the assessment of observed changes in the AR4 (Chapter 3), which was largely based on one study (Dai et al., 2004), subsequent work has drawn a more differentiated picture both regionally and temporally. There is not enough evidence at present to suggest high confidence in observed trends in dryness due to lack of direct observations, some geographical inconsistencies in the trends, and some dependencies of inferred trends on the index choice. There is *medium confidence* that since the 1950s some regions of the world have experienced more intense and longer droughts (e.g., southern Europe, west Africa) but also opposite trends exist in other regions (e.g., central North America, northwestern Australia).

Causes of the Observed Changes

The AR4 (Hegerl et al., 2007) concluded that it is *more likely than not* that anthropogenic influence has contributed to the increase in the droughts observed in the second half of the 20th century. This assessment was based on several lines of evidence, including a detection study that identified an anthropogenic fingerprint in a global PDSI data set with high significance (Burke et al., 2006), although the model trend was weaker than observed and the relative contributions of natural external forcings and anthropogenic forcings were not assessed.

There is now a better understanding of the potential role of land-atmosphere feedbacks versus SST forcing for meteorological droughts (e.g., Schubert et al., 2008a,b), and some modeling studies have also addressed potential impacts of land use changes (e.g., Deo et al., 2009), but large uncertainties remain in the field of land surface modeling and land-atmosphere interactions, in part due to lack of observations (Seneviratne et al., 2010), inter-model discrepancies (Koster et al., 2004b; Dirmeyer et al., 2006; Pitman et al., 2009), and model resolution of orographic and other effects. Nonetheless, a new set of climate modeling studies show that US drought response to SST variability is consistent with observations (Schubert et al., 2009). Inferred trends in drought are also consistent with trends in global precipitation and temperature, and the latter two are consistent with expected responses to anthropogenic forcing (Hegerl et al., 2007; X. Zhang et al., 2007). The change in the pattern of global precipitation in the observations and in model simulations is also consistent with the theoretical understanding of hydrological response to global warming that wet regions become overall wetter and dry regions drier in a warming world (Held and Soden, 2006; see also Section 3.1.6), though some regions also display shifts in climate regimes (Section 3.1.6). Nonetheless, some single events have been reported as differing from projections (Seager et al., 2009), though this is not necessarily incompatible given the superimposition of anthropogenic climate change and natural climate variability (Section 3.1). For soil moisture and hydrological drought it has been suggested that the stomatal 'antitranspirant' responses of plants to rising atmospheric CO_2 may lead to a decrease in evapotranspiration (Gedney et al., 2006). This could mean that increasing CO_2 levels alleviate soil moisture and streamflow drought, but this result is still debated (e.g., Piao et al., 2007; Gerten et al., 2008), in particular due to the uncertainty in observed runoff trends used to infer these effects (e.g., Peel and McMahon, 2006; see also Section 3.2.1).

Overall, though new studies have furthered the understanding of the mechanisms leading to drought, there is still relatively limited evidence to provide an attribution of observed changes, in particular given the issues associated with the availability of observational data (Section 3.2.1) and the definition and computation of drought indicators (Box 3-3). This latter point was mostly identified in post-AR4 studies (Box 3-3). Moreover, regions where consistent increases in drought are identified (see 'Observed Changes') are only partly consistent with those where projections indicate an enhancement of drought conditions in coming decades (see next paragraphs). We thus assess that there is *medium confidence* (see also Section 3.1.5) that anthropogenic influence has contributed to some changes in the drought patterns observed in the second half of the 20th century, based on its attributed impact on precipitation and temperature changes (though temperature can only be indirectly related to drought trends; see Box 3-3). However there is *low confidence* in the attribution of changes in droughts at the level of individual regions.

Projected Changes and Uncertainties

The AR4 assessed that projections at the time indicated an increase in droughts, in particular in subtropical and mid-latitude areas (Christensen et al., 2007). An increase in dry spell length and frequency was considered *very likely* over the Mediterranean region, southern areas of Australia, and New Zealand and *likely* over most subtropical regions, with little change over northern Europe. Continental drying and the associated risk of drought were considered *likely* to increase in summer over many mid-latitude continental interiors (e.g., central and southern Europe, the Mediterranean region), in boreal spring, and dry periods of the annual cycle over Central America.

More recent global and regional climate simulations and hydrological models mostly support the projections from the AR4, as summarized in the following paragraphs (see also Table 3-3), although we assess the overall *confidence* in drought projections as *medium* given the definitional issues associated with dryness and the partial lack of agreement in model projections when based on different dryness indices (Box 3-3). Indeed, particular care is needed in inter-comparing 'drought' projections since very many different definitions are employed (corresponding to different types of droughts), from simple climatic indices such as CDD to more complex indices of soil moisture and hydrological drought (Box 3-3). A distinction also needs to be made between short-term and longer-term events. Blenkinsop and Fowler (2007a) and Burke et al. (2010), for example, show different trend strength, and sometimes sign (Blenkinsop and Fowler, 2007a), for changes in short- and long-term droughts with RCM ensembles applied to the United Kingdom (although uncertainties in the latter projections are large; see below). These various distinctions are generally not considered and most currently available studies only assess changes in very few (most commonly one or two) dryness indices.

On the global scale, Burke and Brown (2008) provided an analysis of projected changes in drought based on four indices (SPI, PDSI, PPEA,

and SMA; for definitions, see Box 3-3) using two model ensembles: one based on a GCM expressing uncertainty in parameter space, and a multi-model ensemble of 11 GCM simulations from CMIP3. Their analysis revealed that SPI, based solely on precipitation, showed little change in the proportion of the land surface in drought, and that all other indices, which include a measure of the atmospheric demand for moisture, showed a statistically significant increase with an additional 5 to 45% of the land surface in drought. This study also highlighted large uncertainties in regional changes in drought. For reasons highlighted in Box 3-3, using simulated soil moisture anomalies from the climate models avoids some shortcomings of other commonly used indices (although the quality of simulated soil moisture cannot be well evaluated due to lack of observations; Section 3.2 and Box 3-3). In the study of Burke and Brown (2008), this index showed weaker drying compared to PDSI and PPEA indices (but more pronounced drying than the SPI index). In this report, we display projected changes in soil moisture anomalies and CDD (Figure 3-10), this latter index being chosen for continuity with the AR4 (see Figure 10.18 of that report). It can be seen that the two indices partly agree on increased drought in some large regions (e.g., on the annual time scale, in Southern Europe and the Mediterranean region, central Europe, central North America, Central America and Mexico, northeast Brazil, and southern Africa), but some regions where the models show consistent increases in CDD (e.g., southeast Asia) do not show consistent decreases in soil moisture. Conversely, regions displaying a consistent decrease in CDD (e.g., in northeastern Asia) do not show a consistent increase in soil moisture. The substantial uncertainty of drought projections is particularly clear from the soil moisture projections, with, for example, no agreement among the models regarding the sign of changes in December to February over most of the globe. These results regarding changes in CDD and soil moisture are consistent with other published studies (Wang, 2005; Tebaldi et al., 2006; Burke and Brown, 2008; Sheffield and Wood, 2008b; Sillmann and Roeckner, 2008) and the areas that display consistent increasing drought tendencies for both indices have also been reported to display such tendencies for additional indices (e.g., Burke and Brown, 2008; Dai, 2011; Table 3-3). Sheffield and Wood (2008b) examined projections in drought frequency (for droughts of duration of 4 to 6 months and longer than 12 months, estimated from soil moisture anomalies) based on CMIP3 simulations with eight GCMs and the SRES scenarios A2, A1B, and B1. They concluded that drought was projected to increase in several regions under these three scenarios (mostly consistent with those displayed in Figure 3-10 for SMA), although the projections of drought intensification were stronger for the high CO_2 emissions scenarios (A2 and A1B) than for the more moderate scenario (B1). Regions showing statistically significant increases in drought frequency were found to be broadly similar for all three scenarios, despite the more moderate signal in the B1 scenario (their Figures 8 and 9). This study also highlighted the large uncertainty of scenarios for drought projections, as scenarios were found to span a large range of changes in drought frequency in most regions, from close to no change to two- to three-fold increases (their Figure 10).

Regional climate simulations and high-resolution global atmospheric model simulations over Europe also highlight the Mediterranean region
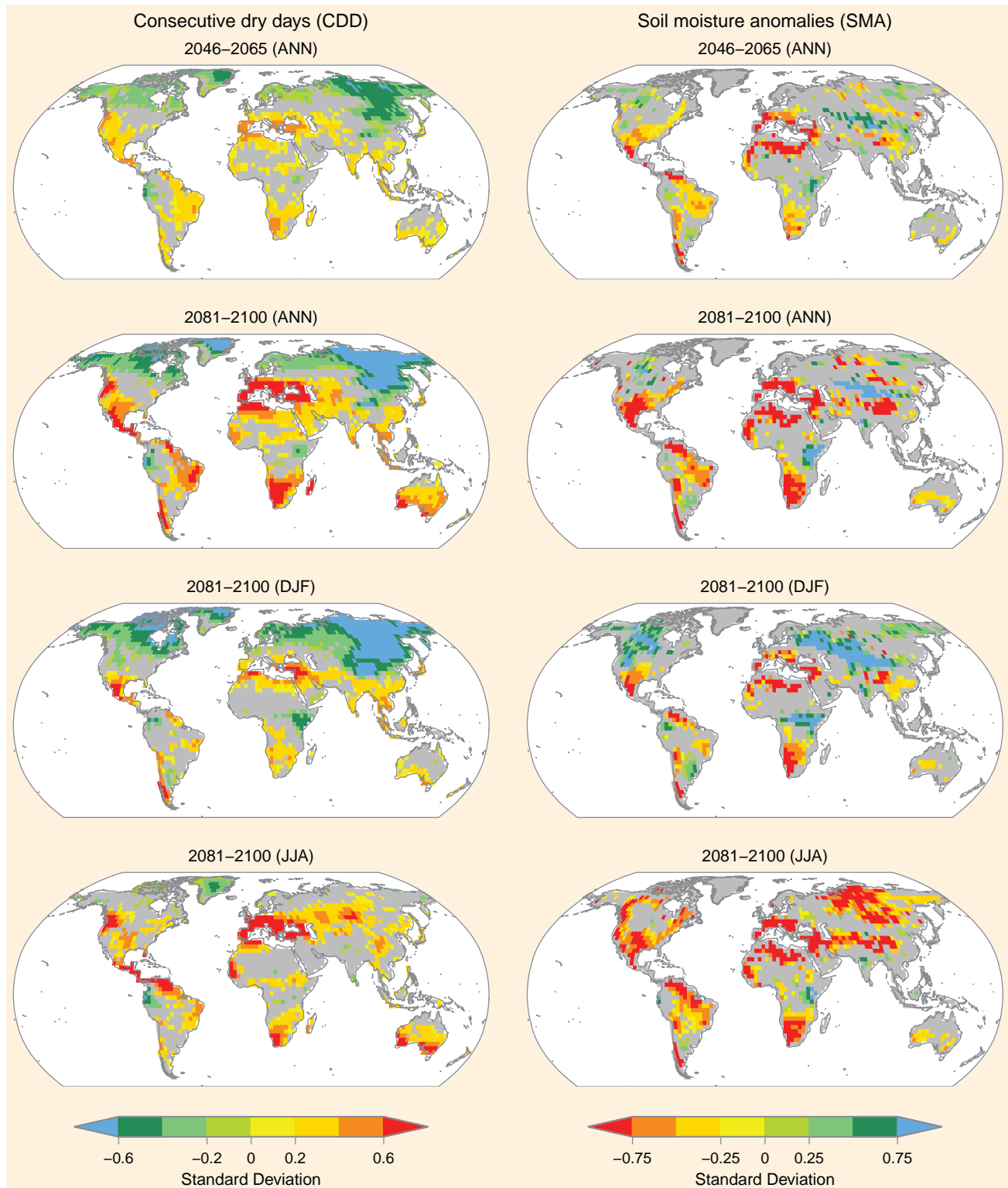
Figure 3-10 | Projected annual and seasonal changes in dryness assessed from two indices for 2081-2100 (bottom three rows, showing the annual time scale and two seasons, DJF and JJA) and 2046-2065 (top, annual time scale) with respect to 1980-1999. Left column: changes in the maximum number of CDD (days with precipitation <1 mm), based on 17 GCMs contributing to the CMIP3. Right column: changes in soil moisture (soil moisture anomalies, SMA), based on 15 GCMs contributing to the CMIP3. Increased dryness is indicated with warm colors (positive changes in CDD and negative SMA values). The maps show differences between the annual and seasonal averages over the respective 20-year periods, that is, the average of 2081-2100 or 2046-2065, respectively (based on simulations under emission scenario SRES A2), minus the average of 1980-1999 (from corresponding simulations for the 20th century). Differences are expressed in units of standard deviations, derived from detrended per year annual or seasonal estimates, respectively, from the three 20-year periods 1980-1999, 2046-2065, and 2081-2100 pooled together. Color shading is only applied for areas where at least 66% of the GCMs (12 out of 17 for CDD, 10 out of 15 for soil moisture) agree on the sign of the change; stippling is applied for regions where at least 90% of the GCMs (16 out of 17 for CDD, 14 out of 15 for soil moisture) agree on the sign of the change. Adapted from Orlowsky and Seneviratne (2011); updating Tebaldi et al. (2006) for SMA and for additional CMIP3 models, and including seasonal time frames. For more details, see Appendix 3.A.

as being affected by more severe droughts, consistent with available global projections (Table 3-3; see also Giorgi, 2006; Rowell and Jones, 2006; Beniston et al., 2007; Mariotti et al., 2008; Planton et al., 2008). Mediterranean (summer) droughts are projected to start earlier in the year and last longer. Also, increased variability during the dry and warm season is projected (Giorgi, 2006). One GCM-based study projected one to three weeks of additional dry days for the Mediterranean region by the end of the century (Giannakopoulos et al., 2009). For North America, intense and heavy episodic rainfall events with high runoff amounts are interspersed with longer relatively dry periods with increased evapotranspiration, particularly in the subtropics. There is a consensus of most climate model projections for a reduction in cool season precipitation across the US southwest and northwest Mexico (Christensen et al., 2007), with more frequent multi-year drought in the US southwest (Seager et al., 2007; Cayan et al., 2010). Reduced cool season precipitation promotes drier summer conditions by reducing the amount of soil water available for evapotranspiration in summer. For Australia, Alexander and Arblaster (2009) project increases in consecutive dry days, although consensus between models is only found in the interior of the continent. African studies indicate the possibility of relatively small-scale (500-km) heterogeneity of changes in precipitation and drought, based on climate model simulations (Funk et al., 2008; Shongwe et al., 2009). Regional climate simulations of South America project spatially coherent increases in CDD, particularly large over the Brazilian Plateau, and northern Chile and the Altiplano (Kitoh et al., 2011).

Available global and regional studies of hydrological drought (Hirabayashi et al., 2008b; Feyen and Dankers, 2009) project a higher likelihood of hydrological drought by the end of this century, with a substantial increase in the number of drought days (defined as streamflow below a specific threshold) during the last 30 years of the 21st century over North and South America, central and southern Africa, the Middle East, southern Asia from Indochina to southern China, and central and western Australia. Some regions, including eastern Europe to central Eurasia, inland China, and northern North America, project increases in drought. In contrast, wide areas over eastern Russia project a decrease in drought days. At least in Europe, hydrological drought is primarily projected to occur in the frost-free season.

Increased confidence in modeling drought stems from consistency between models and satisfactory simulation of drought indices during the past century (Sheffield and Wood, 2008a; Sillmann and Roeckner, 2008). Inter-model agreement is stronger for long-term droughts and larger spatial scales (in some regions, see above discussion), while local to regional and short-term precipitation deficits are highly spatially variable and much less consistent between models (Blenkinsop and Fowler, 2007b). Insufficient knowledge of the physical causes of meteorological droughts, and of the links to the large-scale atmospheric and ocean circulation, is still a source of uncertainty in drought simulations and projections. For example, plausible explanations have been proposed for projections of both a worsening drought and a substantial increase in rainfall in the Sahara (Biasutti et al., 2009; Burke et al., 2010). Another example is illustrated with the relationship of rainfall in southern

Australia with SSTs around northern Australia. On annual time scales, low rainfall is associated with cooler than normal SSTs. Yet the warming observed in SST over the past few decades has not been associated with increased rainfall, but with a trend toward more drought-like conditions (N. Nicholls, 2010).

There are still further sources of uncertainties affecting the projections of trends in meteorological drought for the coming century. The two most important may be uncertainties in the development of the ocean circulation and feedbacks between land surface and atmospheric processes. These latter processes are related to the effects of drought on vegetation physiology and dynamics (e.g., affecting canopy conductance, albedo, and roughness), with resulting (positive or negative) feedbacks to precipitation formation (Findell and Eltahir, 2003a,b; Koster et al., 2004b; Cook et al., 2006; Hohenegger et al., 2009; Seneviratne et al., 2010; van den Hurk and van Meijgaard, 2010), and possibly – as only recently highlighted – also feedbacks between droughts, fires, and aerosols (Bevan et al., 2009). Furthermore, the development of soil moisture that results from complex interactions among precipitation, water storage as soil moisture (and snow), and evapotranspiration by vegetation is still associated with large uncertainties, in particular because of lack of observations of soil moisture and evapotranspiration (Section 3.2.1), and issues in the representation of soil moisture-evapotranspiration coupling in current climate models (Dirmeyer et al., 2006; Seneviratne et al., 2010). Uncertainties regarding soil moisture-climate interactions are also due to uncertainties regarding the behavior of plant transpiration, growth, and water use efficiency under enhanced atmospheric CO_2 concentrations, which could potentially have impacts on the hydrological cycle (Betts et al., 2007), but are not well understood yet (Hungate et al., 2003; Piao et al., 2007; Bonan, 2008; Teuling et al., 2009; see also above discussion on the causes of observed changes). The space-time development of hydrological drought as a response to a meteorological drought and the associated soil moisture drought (drought propagation, e.g., Peters et al., 2003) also needs more attention. There is some understanding of these issues at the catchment scale (e.g., Tallaksen et al., 2009), but these need to be extended to the regional and continental scales. This would lead to better understanding of the projections of hydrological droughts, which would contribute to a better identification and attribution of droughts and help to improve global hydrological models and land surface models.

In summary, there is _medium confidence_ that since the 1950s some regions of the world have experienced trends toward more intense and longer droughts, in particular in southern Europe and West Africa, but in some regions droughts have become less frequent, less intense, or shorter, for example, central North America and northwestern Australia. There is _medium confidence_ that anthropogenic influence has contributed to some changes in the drought patterns observed in the second half of the 20th century, based on its attributed impact on precipitation and temperature changes (though temperature can only be indirectly related to drought trends; see Box 3-3). However there is _low confidence_ in the attribution of changes in droughts at the level

of single regions due to inconsistent or insufficient evidence. Post-AR4 studies indicate that there is *medium confidence* in a projected increase in duration and intensity of droughts in some regions of the world, including southern Europe and the Mediterranean region, central Europe, central North America, Central America and Mexico, northeast Brazil, and southern Africa. Elsewhere there is overall *low confidence* because of insufficient agreement of projections of drought changes (dependent both on model and dryness index). Definitional issues and lack of data preclude higher *confidence* than *medium* in observations of drought changes, while these issues plus the inability of models to include all the factors likely to influence droughts preclude stronger *confidence* than *medium* in the projections.

3.5.2.   Floods

A flood is "the overflowing of the normal confines of a stream or other body of water, or the accumulation of water over areas that are not normally submerged (some specific examples are discussed in Case Study 9.2.6). Floods include river (fluvial) floods, flash floods, urban floods, pluvial floods, sewer floods, coastal floods, and glacial lake outburst floods" (see Glossary). The main causes of floods are intense and/or long-lasting precipitation, snow/ice melt, a combination of these causes, dam break (e.g., glacial lakes), reduced conveyance due to ice jams or landslides, or by a local intense storm (Smith and Ward, 1998). Floods are affected by various characteristics of precipitation, such as intensity, duration, amount, timing, and phase (rain or snow). They are also affected by drainage basin conditions such as water levels in the rivers, the presence of snow and ice, soil character and status (frozen or not, soil moisture content and vertical distribution), rate and timing of snow/ice melt, urbanization, and the existence of dikes, dams, and reservoirs (Bates et al., 2008). Along coastal areas, flooding may be associated with storm surge events (Section 3.5.5). A change in the climate physically changes many of the factors affecting floods (e.g., precipitation, snow cover, soil moisture content, sea level, glacial lake conditions, vegetation) and thus may consequently change the characteristics of floods. Engineering developments such as dikes and reservoirs regulate flow, and land use may also affect floods. Therefore the assessment of causes of changes in floods is complex and difficult. The focus in this section is on changes in floods that might be related to changes in climate (i.e., referred to as 'climate-driven'), rather than changes in engineering developments or land use. However, because of partial lack of documentation, these can be difficult to distinguish in the instrumental record.

Literature on the impact of climate change on pluvial floods (e.g., flash floods and urban floods) is scarce, although the changes in heavy precipitation discussed in Section 3.3.2 may imply changes in pluvial floods in some regions. This chapter focuses on the spatial, temporal, and seasonal changes in high flows and peak discharge in rivers related to climate change, which cause changes in fluvial (river) floods. River discharge simulation under a changing climate scenario requires a set

of GCM or RCM outputs (e.g., precipitation and surface air temperature) and a hydrological model. A hydrological model may consist of a land surface model of a GCM or RCM and a river routing model. Different hydrological models may yield quantitatively different river discharge, but they may not yield different signs of the trend if the same GCM/ RCM outputs are used. So the ability of models to simulate floods, in particular regarding the signs of the past and future trends, depends on the ability of the GCM or RCM to simulate precipitation changes. The ability of a GCM or RCM to simulate temperature is important for river discharge simulation in snowmelt- and glacier-fed rivers. Downscaling and/or bias-correction are frequently applied to GCM/RCM outputs before hydrological simulations are conducted, which becomes a source of uncertainty. More details on the feasibility and uncertainties in hydrological projections are described later in this section. Coastal floods are discussed in Sections 3.5.3 and 3.5.5. Glacial lake outburst floods are discussed in Section 3.5.6. The impact of floods on human society and ecosystems and related changes are discussed in Chapter 4. Case Study 9.2.6 discusses the management of floods.

Worldwide instrumental records of floods at gauge stations are limited in spatial coverage and in time, and only a limited number of gauge stations have data that span more than 50 years, and even fewer more than 100 years (Rodier and Roche, 1984; see also Section 3.2.1). However, this can be overcome partly or substantially by using pre-instrumental flood data from documentary records (archival reports, in Europe continuous over the last 500 years) (Brázdil et al., 2005), and from geological indicators of paleofloods (sedimentary and biological records over centennial to millennial scales) (Kochel and Baker, 1982). Analysis of these pre-instrumental flood records suggest that (1) flood magnitude and frequency can be sensitive to modest alterations in atmospheric circulation, with greater sensitivity for 'rare' floods (e.g., 50-year flood and higher) than for smaller and more frequent floods (e.g., 2-year floods) (Knox, 2000; Redmond et al., 2002); (2) high interannual and interdecadal variability can be found in flood occurrences both in terms of frequency and magnitude although in most cases, cyclic or clusters of flood occurrence are observed in instrumental (Robson et al., 1998), historical (Vallve and Martin-Vide, 1998; Benito et al., 2003; Llasat et al., 2005), and paleoflood records (Ely et al., 1993; Benito et al., 2008); (3) past flood records may contain analogs of unusual large floods, similar to some recorded recently, sometimes considered to be the largest on record. For example, pre-instrumental flood data show that the 2002 summer flood in the Elbe did not reach the highest flood levels recorded in 1118 and 1845 although it was higher than other disastrous floods of 1432, 1805, etc. (Brázdil et al., 2006). However, the currently available pre-instrumental flood data is also limited, particularly in spatial coverage.

The AR4 and the IPCC Technical Paper VI based on the AR4 concluded that no gauge-based evidence had been found for a climate-driven globally widespread change in the magnitude/frequency of floods during the last decades (Rosenzweig et al., 2007; Bates et al., 2008). However, the AR4 also pointed to possible changes that may imply trends in flood occurrence with climate change. For instance, Trenberth et al. (2007) highlighted a catastrophic flood that occurred along several central

European rivers in 2002, although neither flood nor mean precipitation trends could be identified in this region; however, there was a trend toward increasing precipitation variability during the last century which itself could imply an enhanced probability of flood occurrence. Kundzewicz et al. (2007) argued that climate change (i.e., observed increase in precipitation intensity and other observed climate changes) might already have had an impact on floods. Regarding the spring peak flows, the AR4 concluded with *high confidence* that abundant evidence was found for an earlier occurrence in snowmelt- and glacier-fed rivers (Rosenzweig et al., 2007; Bates et al., 2008), though we expressly note here that a change in the timing of peak flows does not necessarily imply nor preclude changes in flood magnitude or frequency in the affected regions.

Although changes in flood magnitude/frequency might be expected in regions where temperature change affects precipitation type (i.e., rain/snow separation), snowmelt, or ice cover (in particular northern high-latitude and polar regions), widespread evidence of such climate-driven changes in floods is not available. For example, there is no evidence of widespread common trends in the magnitude of floods based on the daily river discharge of 139 Russian gauge stations for the last few to several decades, though a significant shift in spring discharge to earlier dates has been found (Shiklomanov et al., 2007). Lindström and Bergström (2004) noted that it is difficult to conclude that flood levels are increasing from an analysis of runoff trends in Sweden for 1807 to 2002.

In the United States and Canada during the 20th century and in the early 21st century, there is no compelling evidence for climate-driven changes in the magnitude or frequency of floods (Lins and Slack, 1999; Douglas et al., 2000; McCabe and Wolock, 2002; Cunderlik and Ouarda, 2009; Villarini et al., 2009). There are relatively abundant studies on the changes and trends for rivers in Europe such as rivers in Germany and its neighboring regions (Mudelsee et al., 2003; Tu et al., 2005; Yiou et al., 2006; Petrow and Merz, 2009), in the Swiss Alps (Allamano et al., 2009), in France (Renard et al., 2008), in Spain (Benito et al., 2005), and in the United Kingdom (Robson et al., 1998; Hannaford and Marsh, 2008), but a continental-scale assessment of climate-driven changes in the flood magnitude and frequency for Europe is difficult to provide because geographically organized patterns are not seen in the reported changes.

Available (limited) analyses for Asia suggest the following changes: the annual flood maxima of the lower Yangtze region show an upward trend over the last 40 years (Jiang et al., 2008), the likelihood for extreme floods in the Mekong River has increased during the second half of the 20th century although the probability of an average flood has decreased (Delgado et al., 2009), and both upward and downward trends are identified over the last four decades in four selected river basins of the northwestern Himalaya (Bhutiyani et al., 2008). In the Amazon region in South America, the 2009 flood set record highs in the 106 years of data for the Rio Negro at the Manaus gauge site in July 2009 (Marengo et al., 2011). Recent increases have also been reported

in flood frequency in some other river basins in South America (Camilloni and Barros, 2003; Barros et al., 2004). Conway et al. (2009) concluded that robust identification of hydrological change was severely limited by data limitations and other issues for sub-Saharan Africa. Di Baldassarre et al. (2010) found no evidence that the magnitude of African floods has increased during the 20th century. However, such analyses cover only limited parts of the world. Evidence in the scientific literature from the other parts of the world, and for other river basins, appears to be very limited.

Many river systems are not in their natural state anymore, making it difficult to separate changes in the streamflow data that are caused by the changes in climate from those caused by human regulation of the river systems. River engineering and land use may have altered flood probability. Many dams are designed to reduce flooding. Large dams have resulted in large-scale land use change and may have changed the effective rainfall in some regions (Hossain et al., 2009).

The above analysis indicates that research subsequent to the AR4 still does not show clear and widespread evidence of climate-driven observed changes in the magnitude or frequency of floods at the global level based on instrumental records, and there is thus *low confidence* regarding the magnitude and frequency and even the sign of these changes. The main reason for this lack of confidence is due to limited evidence in many regions, since available instrumental records of floods at gauge stations are limited in space and time, which limits the number of analyses. Moreover, the confounding effects of changes in land use and engineering mentioned above also make the identification of climate-driven trends difficult. There are limited regions with medium evidence, where no ubiquitous change is apparent (low agreement). Pre-instrumental flood data can provide information for longer periods, but current availability of these data is even scarcer particularly in spatial coverage. There is abundant evidence for an earlier occurrence of spring peak flows in snowmelt- and glacier-fed rivers (*high confidence*), though this feature may not necessarily be linked with changes in the magnitude of spring peak flows in the concerned regions.

The possible causes for changes in floods were discussed in the AR4 and Bates et al. (2008), but cause-and-effect between external forcing and changes in floods was not explicitly assessed. A rare example considered in Rosenzweig et al. (2007) and Bates et al. (2008) was a study by Milly et al. (2002) which, based on monthly river discharge, reported an impact of anthropogenic climate change on changes (mostly increases) in 'large' floods during the 20th century in selected extratropical river basins larger than 20,000 km[2], but they did not endorse the study because of the lack of widespread observed evidence of such trends in other studies. More recent literature has detected the influence of anthropogenically induced climate change in variables that affect floods, such as aspects of the hydrological cycle (see Section 3.2.2.2) including mean precipitation (X. Zhang et al., 2007), heavy precipitation (see Section 3.3.2), and snowpack (Barnett et al., 2008), though a direct statistical link between anthropogenic climate change and trends in the magnitude and frequency of floods is still not established.

In climates where seasonal snow storage and melting play a significant role in annual runoff, the hydrologic regime is affected by changes in temperature. In a warmer world, a smaller portion of precipitation falls as snow (Hirabayashi et al., 2008a) and the melting of winter snow occurs earlier in spring, resulting in a shift in peak river runoff to winter and early spring. This has been observed in the western United States (Regonda et al., 2005; Clow, 2010), in Canada (Zhang et al., 2001), and in other cold regions (Rosenzweig et al., 2007; Shiklomanov et al., 2007), along with an earlier breakup of river ice in Arctic rivers (Smith, 2000; Beltaos and Prowse, 2009). The observed trends toward earlier timing of snowmelt-driven streamflows in the western United States since 1950 are detectably different from natural variability (Barnett et al., 2008; Hidalgo et al., 2009). Thus, observed warming over several decades that is attributable to anthropogenic forcing has *likely* been linked to earlier spring peak flows in snowmelt- and glacier-fed rivers. It is unclear if observed warming over several decades has affected the magnitude of the snowmelt peak flows, but warming may result either in an increase in spring peak flows where winter snow depth increases (Meehl et al., 2007b) or a decrease in spring peak flows because of decreased snow cover and amounts (Hirabayashi et al., 2008b; Dankers and Feyen, 2009).

There is still a lack of studies identifying an influence of anthropogenic climate change over the past several decades on rain-generated peak streamflow trends because of availability and uncertainty in the observed streamflow data and low signal-to-noise ratio. Evidence has recently emerged that anthropogenic climate change could have increased the risk of rainfall-dominated flood occurrence in some river basins in the United Kingdom in autumn 2000 (Pall et al., 2011). Overall, there is *low confidence* (due to limited evidence) that anthropogenic climate change has affected the magnitude and frequency of floods, though it has detectably influenced several components of the hydrological cycle, such as precipitation and snowmelt, that may impact flood trends. The assessment of causes behind the changes in floods is inherently complex and difficult.

The number of studies that investigated projected flood changes in rivers especially at a regional or a continental scale was limited when the AR4 was published. Projections of flood changes at the catchment/river-basin scale were also not abundantly cited in the AR4. Nevertheless, Kundzewicz et al. (2007) and Bates et al. (2008) argued that more frequent heavy precipitation events projected over most regions would affect the risk of rain-generated floods (e.g., flash flooding and urban flooding).

The number of regional- or continental-scale studies of projected changes in floods is still limited. Recently, a few studies for Europe (Lehner et al., 2006; Dankers and Feyen, 2008, 2009) and a study for the globe (Hirabayashi et al., 2008b) have indicated changes in the frequency and/or magnitude of floods in the 21st century at large scale using daily river discharge calculated from RCM or GCM outputs and hydrological models. A notable change is projected to occur in northeastern Europe in the late 21st century because of a reduction in snow accumulation

(Dankers and Feyen, 2008, 2009; Hirabayashi et al., 2008b), that is, a decrease in the probability of floods, that generally corresponds to lower flood peaks. For other parts of the world, Hirabayashi et al. (2008b) show an increase in the risk of floods in most humid Asian monsoon regions, tropical Africa, and tropical South America with a decrease in the risk of floods in non-negligible areas of the world such as most parts of northern North America.

Projections of flood changes at the catchment/river-basin scale are also not abundant in the scientific literature. Several studies have been undertaken for UK catchments (Cameron, 2006; Kay et al., 2009; Prudhomme and Davies, 2009) and catchments in continental Europe and North America (Graham et al., 2007; Thodsen, 2007; Leander et al., 2008; Raff et al., 2009; van Pelt et al., 2009). However, projections for catchments in other regions such as Asia (Asokan and Dutta, 2008; Dairaku et al., 2008), the Middle East (Fujihara et al., 2008), South America (Nakaegawa and Vergara, 2010; Kitoh et al., 2011), and Africa (Taye et al., 2011) are rare.

Uncertainty is still large in the projected changes in the magnitude and frequency of floods. It has been recently recognized that the choice of GCMs is the largest source of uncertainties in hydrological projections at the catchment/river-basin scale, and that uncertainties from emission scenarios and downscaling methods are also relevant but less important (Graham et al., 2007; Leander et al., 2008; Kay et al., 2009; Prudhomme and Davies, 2009), although, in general, hydrological projections require downscaling and/or bias-correction of GCM outputs (e.g., precipitation and temperature). Also the choice of hydrological models was found to be relevant but less important (Kay et al., 2009; Taye et al., 2011). However, the relative importance of downscaling, bias-correction, and the choice of hydrological models may depend on the selected region/catchment, the selected downscaling and bias-correction methods, and the selected hydrological models (Wilby et al., 2008). For example, the sign of the above-mentioned flood changes in northeastern Europe is affected by differences in temporal downscaling and bias-correction methods applied in the different studies (Dankers and Feyen, 2009). Chen et al. (2011) demonstrated considerable uncertainty caused by several downscaling methods in a hydrological projection for a snowmelt-dominated Canadian catchment. Downscaling (see Section 3.2.3) and bias-correction are also a major source of uncertainty in rain-dominated catchments (van Pelt et al., 2009). We also note that bias-correction and statistical downscaling tend to ignore the energy closure of the climate system, which could be a non-negligible source of uncertainty in hydrological projections (Milly and Dunne, 2011).

The number of projections of flood magnitude and frequency changes is still limited at regional and continental scales. Projections at the catchment/river-basin scale are also not abundant in the peer-reviewed scientific literature, especially for regions outside Europe and North America. In addition, considerable uncertainty remains in the projections of flood changes, especially regarding their magnitude and frequency. Therefore, our assessment is that there is *low confidence* (due to limited evidence) in future changes in flood magnitude and frequency derived

from river discharge simulations. Nevertheless, as was argued by Kundzewicz et al. (2007) and Bates et al. (2008), physical reasoning suggests that projected increases in heavy rainfall in some catchments or regions would contribute to increases in rain-generated local floods (*medium confidence*). We note that heavy precipitation may be projected to increase despite a projected decrease of total precipitation depending on the regions considered (Section 3.3.2), and that changes in several variables (e.g., precipitation totals, frequency, and intensity, snow cover and snowmelt, soil moisture) are relevant for changes in floods. Confidence in change in one of these components alone may thus not be sufficient to confidently project changes in flood occurrence. Hence, *medium confidence* is attached to the above statement based on physical reasoning, although the link between increases in heavy rainfall and increases in local flooding seems apparent. The earlier shifts of spring peak flows in snowmelt- and glacier-fed rivers are robustly projected (Kundzewicz et al., 2007; Bates et al., 2008); so these are assessed as *very likely*, though this may not necessarily be relevant for flood occurrence. There is *low confidence* (due to limited evidence) in the projected magnitude of the earlier peak flows in snowmelt- and glacier-fed rivers.

In summary, there is limited to medium evidence available to assess climate-driven observed changes in the magnitude and frequency of floods at a regional scale because the available instrumental records of floods at gauge stations are limited in space and time, and because of confounding effects of changes in land use and engineering. Furthermore, there is *low agreement* in this evidence, and thus overall *low confidence* at the global scale regarding even the sign of these changes. There is *low confidence* (due to limited evidence) that anthropogenic climate change has affected the magnitude or frequency of floods, though it has detectably influenced several components of the hydrological cycle such as precipitation and snowmelt (*medium confidence* to *high confidence*), which may impact flood trends. Projected precipitation and temperature changes imply possible changes in floods, although overall there is *low confidence* in projections of changes in fluvial floods. *Confidence* is *low* due to limited evidence and because the causes of regional changes are complex, although there are exceptions to this statement. There is *medium confidence* (based on physical reasoning) that projected increases in heavy rainfall (Section 3.3.2) would contribute to increases in rain-generated local flooding, in some catchments or regions. Earlier spring peak flows in snowmelt- and glacier-fed rivers are *very likely*, but there is *low confidence* in their projected magnitude.

3.5.3.   Extreme Sea Levels

Transient sea level extremes and extreme coastal high water are caused by severe weather events or tectonic disturbances that cause tsunamis. Since tsunamis are not climate-related, they are not addressed here. The drop in atmospheric pressure and strong winds that accompany severe

weather events such as tropical or extratropical cyclones (Sections 3.4.4 and 3.4.5) can produce storm surges at the coast, which may be further elevated by wave setup caused by an onshore flux of momentum due to wave breaking in the surf zone. Various metrics are used to characterize extreme sea levels including storm-related highest values, annual maxima, or percentiles. Extreme sea levels may change in the future as a result of both changes in atmospheric storminess and mean sea level rise. However, neither contribution will be spatially uniform across the globe. For severe storm events such as tropical and extratropical cyclones, changes may occur in the frequency, intensity, or genesis regions of severe storms and such changes may vary between ocean basins (see Sections 3.4.4 and 3.4.5). Along some coastlines, land subsidence due to glacial isostatic adjustment (e.g., Lambeck et al., 2010) is causing a relative fall in sea levels. Variations in the rate of sea level rise can be large relative to mean sea level (Yin et al., 2010) and will occur as a result of variations in wind change (e.g., Timmermann et al., 2010), changes in atmospheric pressure and oceanic circulation (e.g., Tsimplis et al., 2008), and associated differences in water density and rates of thermal expansion (e.g., Bindoff et al., 2007; Church et al., 2010; Yin et al., 2010). In addition, if rapid melting of ice sheets occurs it would lead to non-uniform rates of sea level rise across the globe due to adjustments in the Earth's gravitational field (e.g., Mitrovica et al., 2010). On some coastlines, higher mean sea levels may alter the astronomical tidal range and the evolution of storm surges, and increase the wave height in the surf zones. As well as gradual increases in mean sea level that contribute to extreme impacts from transient extreme sea levels, rapid changes in sea level arising from, for example, collapse of ice shelves could be considered to be an extreme event with the potential to contribute to extreme impacts in the future. However, knowledge about the likelihood of such changes occurring is limited and so does not allow an assessment at this time.

Mean sea level has varied considerably over glacial time scales as the extent of ice caps and glaciers have fluctuated with global temperatures. Sea levels have risen around 120 to 130 m since the last glacial maximum 19 to 23 ka before present to around 7,000 years ago, and reached a level close to present at least 6,000 years ago (Lambeck et al., 2010). As well as the influence on sea level extremes caused by rapidly changing coastal bathymetries (Clarke and Rendell, 2009) and large-scale circulation patterns (Wanner et al., 2008), there is some evidence that changes in the behavior of severe tropical cyclones has changed on centennial time scales, which points to non-stationarity in extreme sea level events (Nott et al., 2009). Woodworth et al. (2011) use tide gauge records dating back to the 18th century, and salt marsh data, to show that sea level rise has accelerated over this time frame.

The AR4 reported that there was *high confidence* that the rate of observed sea level rise increased from the 19th to the 20th century (Bindoff et al., 2007). It also reported that the global mean sea level rose at an average rate of 1.7 (1.2 to 2.2) mm yr^{-1} over the 20th century, 1.8 (1.3 to 2.3) mm yr^{-1} over 1961 to 2003, and at a rate of 3.1 (2.4 to 3.8) mm yr^{-1} over 1993 to 2003. With updated satellite data to 2010, Church and White (2011) show that satellite-measured sea levels continue to rise at

a rate close to that of the upper range of the AR4 projections. Whether the faster rate of increase during the latter period reflects decadal variability or an increase in the longer-term trend is not clear. However, there is evidence that the contribution to sea level due to mass loss from Greenland and Antarctica is accelerating (Velicogna, 2009; Rignot et al., 2011; Sørensen et al., 2011). The AR4 also reported that the rise in mean sea level and variations in regional climate led to a *likely* increase in the trend of extreme high water worldwide in the late 20th century (Bindoff et al., 2007), it was *very likely* that humans contributed to sea level rise during the latter half of the 20th century (Hegerl et al., 2007), and therefore that it was *more likely than not* that humans contributed to the trend in extreme high sea levels (IPCC, 2007a). Since the AR4, Menendez and Woodworth (2010), using data from 258 tide gauges across the globe, have confirmed the earlier conclusions of Woodworth and Blackman (2004) that there was an increasing trend in extreme sea levels globally, more pronounced since the 1970s, and that this trend was consistent with trends in mean sea level (see also Lowe et al., 2010). Additional studies at particular locations support this finding (e.g., Marcos et al., 2009; Haigh et al., 2010).

Various studies also highlight the additional influence of climate variability on extreme sea level trends. Menendez and Woodworth (2010) report that ENSO (see Section 3.4.2) has a large influence on interannual variations in extreme sea levels in the Pacific Ocean and the monsoon regions based on sea level records since the 1970s. In southern Europe, Marcos et al. (2009) report that changes in extremes are also significantly negatively correlated with the NAO (see Section 3.4.3). Ullmann et al. (2007) concluded that maximum annual sea levels in the Camargue had risen twice as fast as mean sea level during the 20th century due to an increase in southerly winds associated with a general rise in sea level pressure over central Europe (Ullmann et al., 2008). Sea level trends from two tide gauges on the north coast of British Columbia from 1939 to 2003 were twice that of mean sea level rise, the additional contribution being due to the strong positive phase of the PDO (see Section 3.4.3), which has lasted since the mid-1970s (Abeysirigunawardena and Walker, 2008). Cayan et al. (2008) reported an increase of 20-fold at San Francisco since 1915 and 30-fold at La Jolla since 1933 in the frequency of exceedance of the 99.99th percentile sea level. They also noted that positive sea level anomalies of 10 to 20 cm that often persisted for several months during El Niño events produced an increase in storm surge peaks over this time. The spatial extent of these oscillations and their influence on extreme sea levels across the Pacific has been discussed by Merrifield et al. (2007). Church et al. (2006a) examined changes in extreme sea levels before and after 1950 in two tide gauge records of approximately 100 years on the east and west coasts of Australia, respectively. At both locations a stronger positive trend was found in the sea level exceeded by 0.01% of the observations than the median sea level, suggesting that in addition to mean sea level rise, other modes of variability or climate change are contributing to the extremes. At Mar del Plata, Argentina, Fiore et al. (2009) noted an increase in the number and duration of positive storm surges in the decade 1996 to 2005 compared to previous decades, which may be due to a combination of mean sea level rise and changes

in wind climatology resulting from a southward shift in the South Atlantic high.

Thus, studies since the AR4 conclude that trends in extreme sea level are generally consistent with changes in mean sea level (e.g., Marcos et al., 2009; Haigh et al., 2010; Menendez and Woodworth, 2010) although some studies note that the trends in extremes are larger than the observed trend in mean sea levels (e.g., Church et al., 2006a; Ullmann et al., 2007; Abeysirigunawardena and Walker, 2008) and may be influenced by modes of climate variability, such as the PDO on the Canadian west coast (e.g., Abeysirigunawardena and Walker, 2008). These studies are consistent with the conclusions from the AR4 that increases in extremes are related to trends in mean sea level and modes of variability in the regional climate.

The AR4 (Meehl et al., 2007b) projected sea level rise for 2090-2099 relative to 1980-1999 due to ocean thermal expansion, glaciers and ice caps, and modeled ice sheet contributions of 18 to 59 cm, which incorporates a 90% uncertainty range across all scenarios. An additional contribution to the sea level rise projections was taken into account for a possible rapid dynamic response of the Greenland and West Antarctic ice sheets, which could result in an accelerating contribution to sea level rise. This was estimated to be 10 to 20 cm of sea level rise by 2090-2099 using a simple linear relationship with projected temperature. Because of insufficient understanding of the dynamic response of ice sheets, Meehl et al. (2007b) also noted that a larger contribution could not be ruled out.

Several studies since the AR4 have developed statistical models that relate 20th-century (e.g., Rahmstorf, 2007; Horton et al., 2008) or longer (e.g., Vermeer and Rahmstorf, 2009; Grinsted et al., 2010) temperature and sea level rise to extrapolate future global mean sea level. These alternative approaches yield projections of sea level rise under a range of SRES scenarios by 2100 of 0.47 to 1.00 m (B1 to A2 scenarios; Horton et al., 2008), 0.50 to 1.40 m (B1 to A1FI scenarios; Rahmstorf, 2007), 0.75 to 1.90 m (B1 to A1FI scenarios; Vermeer and Rahmstorf, 2009), and 0.90 to 1.30 m (A1B scenario only; Grinsted et al., 2010). However, future rates of sea level rise may be less closely associated with global mean temperature if ice sheet dynamics play a larger role in the future (Cazenave and Llovel, 2010). Furthermore, Church et al. (2011) note that these models may overestimate future sea levels because non-climate related contributions to trends over the observational period such as groundwater depletion may not have been removed, and non-linear effects such as the reduction in glacier area as glaciers contract and the reduction in the efficiency of ocean heat uptake with global warming in the future are not accounted for. Pfeffer et al. (2008), using a dynamical model of glaciers, found that sea level rise of more than 2 m by 2100 is physically implausible. An estimate of 0.8 m by 2100 that included increased ice dynamics was considered most plausible.

New studies, whose focus is on quantifying the effect of storminess changes on storm surge, have been carried out over northern Europe since the AR4. Debernard and Roed (2008) used hydrodynamic models

to investigate storm surge changes over Europe in four regionally downscaled GCMs including two runs with B2, one with A2, and one with an A1B emission scenario. Despite large inter-model differences, statistically significant changes between 1961-1990 and 2071-2100 consisted of decreases in the 99th percentile surge heights south of Iceland, and an 8 to 10% increase along the coastlines of the eastern North Sea and the northwest British Isles, which occurred mainly in the winter season. Wang et al. (2008) projected a significant increase in wintertime storm surges around Ireland except the south Irish coast over 2031-2060 relative to 1961-1990 using a downscaled GCM under an A1B scenario. Sterl et al. (2009) joined the output from an ensemble of 17 GCM (CMIP3) simulations using the A1B emissions scenario over the model periods 1950-2000 and 2050-2100 into a single longer time series to estimate 10,000-year return values of surge heights along the Dutch coastline. No statistically significant change in this value was projected for the 21st century because projected wind speed changes were not associated with the surge-generating northerlies but rather non-surge generating south-westerlies.

Other studies have undertaken a sensitivity approach to compare the relative impact on extreme sea levels of severe weather changes and mean sea level rise. Over southeastern Australia, McInnes et al. (2009b) found that a 10% increase in wind speeds, consistent with the upper end of the range under an A1FI scenario from a multi-model ensemble for 2070 together with an A1FI sea level rise scenario, would produce extreme sea levels that were 12 to 15% higher than those including just the A1FI sea level rise projection alone. Brown et al. (2010) also investigated the relative impact of sea level rise and wind speed change on an extreme storm surge in the eastern Irish Sea. Both studies concluded that sea level rise rather than meteorological changes has the greater potential to increase extreme sea levels in these locations in the future.

The degree to which climate models (GCM or RCM) have sufficient resolution and/or internal physics to realistically capture the meteorological forcing responsible for storm surges is regionally dependent. For example current GCMs are unable to realistically represent tropical cyclones (see Section 3.4.4). This has led to the use of alternative approaches for investigating the impact of climate change on storm surges in tropical locations whereby large numbers of cyclones are generated using statistical models that govern the cyclones' characteristics over the observed period (e.g., McInnes et al., 2003). These models are then perturbed to represent projected future cyclone characteristics and used to force a hydrodynamic model. Recent studies on the tropical east coast of Australia reported in Harper et al. (2009) that employ these approaches show a relatively small impact of a 10% increase in tropical cyclone intensity on the 1-in-100 year storm tide (the combined sea level due to the storm surge and tide), and mean sea level rise being found to produce the larger contribution to changes in future 1-in-100 year sea level extremes. However, one study that has incorporated scenarios of sea level rise in the hydrodynamic modeling of hurricane-induced sea level extremes on the Louisiana coast found that increased coastal water depths had a large impact on surge propagation over land,

increasing storm surge heights by two to three times the sea level rise scenario, particularly in wetland-fronted areas (J.M. Smith et al., 2010).

To summarize, post-AR4 studies provide additional evidence that trends in extreme coastal high water across the globe reflect the increases in mean sea level, suggesting that mean sea level rise rather than changes in storminess are largely contributing to this increase (although data are sparse in many regions and this lowers the confidence in this assessment). It is therefore considered *likely* that sea level rise has led to a change in extreme coastal high water levels. It is *likely* that there has been an anthropogenic influence on increasing extreme coastal high water levels via mean sea level contributions. While changes in storminess may contribute to changes in sea level extremes, the limited geographical coverage of studies to date and the uncertainties associated with storminess changes overall (Sections 3.4.4 and 3.4.5) mean that a general assessment of the effects of storminess changes on storm surge is not possible at this time. On the basis of studies of observed trends in extreme coastal high water levels it is *very likely* that mean sea level rise will contribute to upward trends in the future.

3.5.4. Waves

Severe waves threaten the safety of coastal inhabitants and those involved in maritime activities and can damage and destroy coastal and marine infrastructure. Waves play a significant role in shaping a coastline by transporting energy from remote areas of the ocean to the coast. Energy dissipation via wave breaking contributes to beach erosion, longshore currents, and elevated coastal sea levels through wave set-up and wave run-up. Wave properties that influence these processes include wave height, the wave energy directional spectrum, and period. Studies of past and future changes in wave climate to date have tended to focus on wave height parameters such as 'Significant Wave Height' (SWH, the average height from trough to crest of the highest one-third of waves) and metrics of extreme waves, such as high percentiles or wave heights above particular thresholds, although one study (Dodet et al., 2010) also examines trends in mean wave direction and peak wave period. It should also be noted that waves may become an increasingly important factor along coastlines experiencing a decline in coastal protection afforded by sea ice (see Sections 3.5.5 and 3.5.7).

Wave climates have changed over paleoclimatic time scales. Wave modeling using paleobathymetries over the past 12,000 years indicates an increase in peak annual SWH of around 40% due to the increase in mean sea level, which redefines the location of the coastline, and hence progressively extends the fetch length in most of the shelf sea regions (Neill et al., 2009). Major circulation changes that result in changes in storminess and wind climate (see Section 3.3.3) have also affected wave climates. Evidence of enhanced storminess determined from sand drift and dune building along the western European coast indicates that enhanced storminess occurred over the period of the Little Ice Age

(1570-1900) and the mid Holocene (~8,200 years before present; Clarke and Rendell, 2009).

The AR4 reported statistically significant positive trends in SWH over the period 1950 to 2002 over most of the mid-latitudinal North Atlantic and North Pacific, as well as in the western subtropical South Atlantic, the eastern equatorial Indian Ocean and the East China and South China Sea, and declining trends around Australia, and parts of the Philippine, Coral, and Tasman Seas (Trenberth et al., 2007), based on voluntary observing ship data (e.g., Gulev and Grigorieva, 2004). Several studies that address trends in extreme wave conditions have been completed since the AR4 and the new studies generally provide more evidence for the previously reported positive trends in SWH and extreme waves in the north Atlantic and north Pacific. Global trends in 99th-percentile satellite-measured wave heights show a mostly significant positive trend of between 0.5 and 1.0% per year in the mid-latitude oceans but less clear trends over the tropical oceans from 1985 to 2008 (Young et al., 2011). X.L. Wang et al. (2009b) found that SWH increased in the boreal winter over the past half century in the high latitudes of the Northern Hemisphere (especially the northeast Atlantic), and decreased in more southerly northern latitudes based on the European Centre for Medium Range Weather Forecasts 40-year reanalysis (ERA-40). They also found that storminess around the 1880s was of similar magnitude to that in the 1990s. This is also found using the same data set by Le Cozannet et al. (2011), who relate the change in waves to the NAO pattern that is moderated by an east Atlantic pattern of climate variability during winter. A wave hindcast over the north-eastern Atlantic Ocean over the period 1953 to 2009 revealed a significant positive trend in SWH, as well as a counterclockwise shift in mean direction in the north and a slight but not significant increase in peak wave period in the northeast. In the south, no trend was found for SWH or wave period while a clockwise trend in mean direction was found (Dodet et al., 2010). In a regional North Sea hindcast, Weisse and Günther (2007) found a positive trend in 99th-percentile wave height from 1958 to the early 1990s followed by a declining trend to 2002 over the southern North Sea, except on the UK North Sea coast where negative trends occurred over much of the hindcast period.

On the North American Atlantic coast, Komar and Allan (2008) found a statistically significant trend of 0.059 m yr^{-1} in waves exceeding 3 m during the summer months over 30 years since the mid-1970s at Charleston, South Carolina, with weaker but statistically significant trends at wave buoys further north. These trends were associated with an increase in intensity and frequency of hurricanes over this period (see Section 3.4.4). In contrast, winter waves, generated by extratropical storms, were not found to have experienced a statistically significant change. In the eastern North Pacific, SWH is strongly correlated with El Niño (Section 3.4.2). However positive trends were also found in SWH and extreme wave height from the mid-1970s to 2006 in wave buoy data (Allan and Komar, 2006), for excesses of the 98th percentile SWH over 1985 to 2007 (Menendez et al., 2008) along the US west coast, and in hindcast SWH over 1948 to 1998 in the Southern Californian Bight (Adams et al., 2008). Positive though not statistically significant trends

in annual mean SWH were found over south-eastern South America for *in situ* wave data over the 1996-2006 period and in satellite wave data over 1993 to 2001, while simulated wave fields using reanalysis wind forcing over the period 1971 to 2005 produced statistically significant trends in SWH (Dragani et al., 2010). Trends at particular locations may be also influenced by local factors. For example, Suursaar and Kullas (2009) reported a slight decreasing trend in mean SWHs from 1966 to 2006 in the Gulf of Riga within the Baltic Sea, while the frequency and intensity of high wave events (i.e., the difference between the maximum and 99th-percentile wave height) showed rising trends. These changes were associated with a decrease in local average wind speed, but an intensification of westerly winds and storm events occurring further to the west.

In the Southern Ocean, SWH derived from satellite observations was found to be strongly positively correlated with the SAM, particularly from March to August (Hemer et al., 2010). However, the analysis of reliable long-term trends in the Southern Hemisphere remains challenging due to limited *in situ* data and problems of temporal homogeneity in reanalysis products (Wang et al., 2006). For example, Hemer et al. (2010) also found that trends in SWH derived from satellite data over 1998-2000 relative to 1993-1996 were positive only over the Southern Ocean south of 45°S whereas trends were positive across most of the Southern Hemisphere in the Corrected ERA-40 reanalysis (C-ERA-40; Hemer, 2010). Hemer (2010) found that the frequency of wave events exceeding the 98th percentile over the period 1985 to 2002 using data from a wave buoy situated on the west coast of Tasmania showed no statistically significant trend whereas a strong positive trend was found in equivalent fields of C-ERA-40 data.

New studies have demonstrated strong links between wave climate and natural modes of climate variability (Section 3.4.3). For example, along the US west coast and the western North Pacific, SWH was found to be strongly correlated with El Niño (Allan and Komar, 2006; Sasaki and Toshiyuki, 2007) and, in the Southern Ocean, SWH was positivity correlated with the SAM (Hemer et al., 2010). On the US east coast, positive trends in summer SWH were linked to increasing numbers of hurricanes (Komar and Allan, 2008). In the northeast Atlantic, trends in SWH exhibited significant positive (negative) correlations with the NAO in the north (south) and more generally, trends in SWH, mean wave direction, and peak wave period over the period 1953 to 2009 were related to the increase in the NAO index over this time (Dodet et al., 2010). One study (X.L. Wang et al., 2009a) reported a link between external forcing (i.e., anthropogenic forcing due to greenhouse gases and aerosols, and natural forcing due to solar and volcanic forcing) and an increase in SWH in the boreal winter in the high latitudes of the Northern Hemisphere (especially the northeast North Atlantic), and a decrease in more southerly northern latitudes over the past half century.

The AR4 projected an increase in extreme wave height in many regions of the mid-latitude oceans as a result of projected increases in wind speeds associated with more intense mid-latitude storms in these regions in a future warmer climate (Meehl et al., 2007b). At the regional scale,

increases in wave height were projected for most mid-latitude areas analyzed, including the North Atlantic, North Pacific, and Southern Ocean (Christensen et al., 2007) but with *low confidence* due to *low confidence* in projected changes in mid-latitude storm tracks and intensities (see Section 3.4.5). Several studies since then have developed wave climate projections that provide stronger evidence for future wave climate change. Global-scale projections of SWH were developed by Mori et al. (2010), using a 1.25° resolution wave model forced with projected winds from a 20-km global GCM, in which ensemble-averaged SST changes from the CMIP3 models provided the climate forcing. The spatial pattern of projected SWH change between 1979-2004 and 2075-2100 reflects the changes in the forcing winds, which are generally similar to the mean wind speed changes shown in Figure 3-8. Extreme waves (measured by a spatial and temporal average of the top 10 values over the 25-year period) were projected to exhibit large increases in the northern Pacific, particularly close to Japan due to an increase in strong tropical cyclones and also the Indian Ocean despite decreases in SWH.

A number of regional studies have also been completed since the AR4 in which forcing conditions were obtained for a few selected emission scenarios (typically B2 and A2, representing low-high ranges) from GCMs or RCMs. These studies provide additional evidence for positive projected trends in SWH and extreme waves along the western European coast (e.g., Debernard and Roed, 2008; Grabemann and Weisse, 2008) and the UK coast (Leake et al., 2007), declines in extreme wave height in the Mediterranean sea (Lionello et al., 2008) and the southeast coast of Australia (Hemer et al., 2010), and little change along the Portuguese coast (Andrade et al., 2007). However, considerable variation in projections can arise from the different climate models and scenarios used to force wave models, which lowers the confidence in the projections. For example, along the European North Sea coast, 99th-percentile wave height over the late 21st century relative to the late 20th century is projected to increase by 6 to 8% by Debernard and Roed (2008) based on wave model simulations with forcing from several GCMs under A2, B2, and A1B greenhouse gas scenarios, whereas they are projected to increase by up to 18% by Grabemann and Weisse (2008), who downscaled two GCMs under A2 and B2 emission scenarios. In one region, opposite trends in extreme waves were projected. Grabemann and Weisse (2008) project negative trends in 99th-percentile wave height along the UK North Sea coast, whereas Leake et al. (2007) downscaled the same GCM for the same emission scenarios, using a different RCM, and found positive changes in high percentile wave heights offshore of the East Anglia coastline. A wave projection study by Hemer et al. (2010) concluded that uncertainties arising from the method by which climate model winds were applied to wave model simulations (e.g., by applying bias-correction to winds or perturbing current climate winds with wind changes derived from climate models) made a larger contribution to the spread of RCM projections than the forcing from different GCMs or emission scenarios.

In summary, although post-AR4 studies are few and their regional coverage is limited, their findings generally support the evidence from earlier studies of wave climate trends. Most studies find a **link between variations in waves (both SWH and extremes) and internal climate variability. There is** *low confidence* **that there has been an anthropogenic influence on extreme wave heights (because of insufficient literature). Despite the existence of downscaling studies for some regions such as the eastern North Sea, there is overall** *low confidence* **in wave height projections because of the small number of studies, the lack of consistency of the wind projections between models, and limitations in their ability to simulate extreme winds. However, the strong linkages between wave height and winds and storminess means that it is** *likely* **that future negative or positive changes in SWH will reflect future changes in these parameters.**

3.5.5.   Coastal Impacts

Severe coastal hazards such as erosion and inundation are important in the context of disaster risk management and may be affected by climate change through rising sea levels and changes in extreme events. Increasing sea levels will also increase the potential for saltwater intrusion into coastal aquifers. Coastal inundation occurs during periods of extreme sea levels due to storm surges and high waves, particularly when combined with high tides. Although tropical and extratropical cyclones (Sections 3.4.4 and 3.4.5) are the most common causes of sea level extremes, other weather events that cause persistent winds such as anticyclones and fronts can also influence coastal sea levels (Green et al., 2009; McInnes et al., 2009b). In many parts of the world, sea levels are influenced by modes of large scale variability such as ENSO (Section 3.4.2). In the western equatorial Pacific, sea levels can fluctuate up to half a meter between ENSO phases (Church et al., 2006b) and in combination with extremes of the tidal cycle, can cause extensive inundation in low-lying atoll nations even in the absence of extreme weather events (Lowe et al., 2010).

Shoreline position can change from the combined effects of various factors such as:

1) Rising mean sea levels, which cause landward recession of coastlines made up of erodible materials (e.g., Ranasinghe and Stive, 2009)
2) Changes in coastal height due to isostatic rebound (Blewitt et al., 2010; Mitrovica et al., 2010), or sediment compaction from the removal of oil, gas, and water (Syvitski et al., 2009)
3) Changes in the frequency or severity of transient storm erosion events (K.Q. Zhang et al., 2004)
4) Changes in sediment supply to the coast (Stive et al., 2003; Nicholls et al., 2007; Tamura et al., 2010)
5) Changes in wave speed due to sea level rise, which alters wave refraction, or in wave direction, which can cause realignment of shorelines (Ranasinghe et al., 2004; Bryan et al., 2008; Tamura et al., 2010)
6) The loss of natural protective structures such as coral reefs (e.g., Sheppard et al., 2005; Gravelle and Mimura, 2008) due to increased ocean temperatures (Hoegh-Guldberg, 1999) and ocean acidification (Bongaerts et al., 2010) or the reduction in permafrost

or sea ice in mid- and high latitudes, which exposes soft shores to the effects of waves and severe storms (see Section 3.5.7; Manson and Solomon, 2007).

For example, permafrost degradation and sea ice retreat may contribute to coastal erosion in Arctic regions (see Section 3.5.7).

The susceptibility of coastal regions to erosion and inundation is related to various physical (e.g., shoreline slope), and geomorphological and ecosystem attributes, and therefore may be inferred to some extent from broad coastal characterizations. These include the presence of beaches, rocky shorelines, or coasts with cliffs; deltas; back-barrier environments such as estuaries and lagoons; the presence of mangroves, salt marshes, or sea grasses; and shorelines flanked by coral reefs (e.g., Nicholls et al., 2007) or by permafrost or seasonal sea ice, each of which are characterized by different vulnerability to climate change-driven hazards. For example, deltas are low-lying and hence generally prone to inundation, while beaches are comprised of loose particles and therefore erodible. However, the degree to which these systems are impacted by erosion and inundation will also be influenced by other factors affecting disaster responses. For example, reduced protection from high waves during severe storms could occur as a result of depleted mangrove forests or the degradation of coral reefs (e.g., Gravelle and Mimura, 2008), or loss of sea ice or permafrost (e.g., Manson and Solomon, 2007); there may be a loss of ecosystem services brought about by saltwater contamination of already limited freshwater reserves due to rising sea levels and these will amplify the risks brought about by climate change (McGranahan et al., 2007), and also reduce the resilience of coastal settlements to disasters. Dynamical processes such as vertical land movement also contribute to inundation potential (Haigh et al., 2009). Coastal regions may be rising or falling due to post-glacial rebound or slumping due to aquifer drawdown (Syvitski et al., 2009). Multiple contributions to coastal flooding such as heavy rainfall and flooding in coastal catchments that coincide with elevated sea levels may also be important. Ecosystems such as coral reefs also play an important role in providing material on which atolls are formed. Large-scale oceanic changes that are particularly relevant to both coral reefs and small island countries are discussed in Box 3-4.

As discussed in Section 3.5.3, mean sea level has risen by 120 to 130 m since the end of the last glacial maximum (Jansen et al., 2007), and this has had a profound effect on coastline position around the world. Coastlines have also evolved over this time frame due to changes in the action of the ocean on the coast through changes in wave climate (Neill et al., 2009) and tides (Gehrels et al., 1995), which arise from the changing geometries of coastlines over glacial time scales and changes in storminess (e.g., Clarke and Rendell, 2009).

The AR4 (Nicholls et al., 2007) reported that coasts are experiencing the adverse consequences of impacts such as increased coastal inundation, erosion, and ecosystem losses. However, attributing these changes to sea level rise is difficult due to the multiple drivers of change over the 20th century (R.J. Nicholls, 2010) and the scarcity and fragmentary nature of data sets that contribute to the problem of identifying and

attributing changes (e.g., Defeo et al., 2009). Since the AR4 there have been several new studies that examine coastline changes. In the Caribbean, the beach profiles at 200 sites across 113 beaches and eight islands were monitored on a three-monthly basis from 1985 to 2000, with most beaches found to be eroding and faster rates of erosion generally found on islands that had been impacted by a higher number of hurricanes (Cambers, 2009). However, the relative importance of anthropogenic factors, climate variability, and climate change on the eroding trends could not be separated quantitatively. In Australia, Church et al. (2008) report that despite the positive trend in sea levels during the 20th century, beaches have generally been free of chronic coastal erosion, and where it has been observed it has not been possible to unambiguously attribute it to sea level rise in the presence of other anthropogenic activities. Webb and Kench (2010) argue that the commonly held view of atoll nations being vulnerable to erosion must be reconsidered in the context of physical adjustments to the entire island shoreline, because erosion of some sectors may be balanced by progradation on other sectors. In their survey of 27 atoll islands across three central Pacific Nations (Tuvalu, Kiribati, and Federated States of Micronesia) over a 19- to 61-year period using photography and satellite imagery, they found that 43% of islands remained stable and 43% increased in area, with largest rates of increase in island area ranging from 0.1 to 5.6 ha per decade. Only 14% of islands studied exhibited a net reduction in area. On islands exhibiting either no net change or an increase in area, a larger redistribution of land area was evident in 65% of cases, consisting of mainly a shoreline recession on the ocean side and an elongation of the island or progradation of the shoreline on the lagoon side. Human settlements were present on 7 of the 27 atolls surveyed and the majority of those exhibited net accretion due in part to coastal protection works. For a coral reef island at the northern end of the Great Barrier Reef, Australia, Dawson and Smithers (2010) report a 6% increase in area and 4% increase in volume between 1967 and 2007 but with a net retreat on the east-southeast shoreline and advance on the western side. Chust et al. (2009) evaluated the relative contribution of local anthropogenic (non-climate change related) and sea level rise impacts on the coastal morphology and habitats in the Basque coast, northern Spain, for the period 1954 to 2004. They found that the impact from local anthropogenic influences was about an order of magnitude greater than that due to sea level rise over this period. Increased rates of coastal erosion have also been observed since 1935 in Canada's Gulf of St. Lawrence (Forbes et al., 2004).

The AR4 stated with *very high confidence* that the impact of climate change on coasts is exacerbated by increased pressures on the physical environment arising from human settlements in the coastal zone (Nicholls et al., 2007). The small number of studies that have been completed since the AR4 have been either unable to attribute coastline changes to specific causes in a quantitative way or else find strong evidence for non-climatic causes that are natural and/or anthropogenic.

The AR4 reported with *very high confidence* that coasts will be exposed to increasing impacts, including coastal erosion, over coming decades due to climate change and sea level rise, both of which will be

Box 3-4 | Small Island States

Small island states represent a distinct category of locations owing to their small size and highly maritime climates, which means that their concerns and information needs in relation to future climate change differ in many ways from those of the larger continental regions that are addressed in this chapter. Their small land area and often low elevation makes them particularly vulnerable to rising sea levels and impacts such as inundation, shoreline change, and saltwater intrusion into underground aquifers (Mimura, 1999). Their maritime environments lead to an additional emphasis on oceanic information to understand the impacts of climate change (see Case Study 9.2.9). Particular challenges exist for the assessment of past changes in climate given the sparse regional and temporal coverage of terrestrial-based observation networks and the limited *in situ* ocean observing network, although observations have improved somewhat in recent decades with the advent of satellite-based observations of meteorological and oceanic variables. However, the short length of these records hampers the investigation of long-term trends in the region. The resolution of GCMs is insufficient to represent small islands and few studies have been undertaken to provide projections for small islands using RCMs (Campbell et al., 2011). In regions such as the Pacific Ocean, large-scale climate features such as the South Pacific Convergence Zone ENSO (Section 3.4.2) have a substantial influence on the pattern and timing of precipitation, yet these features and processes are often poorly represented in GCMs (Collins et al., 2010). The purpose of this box is to present available information on observed trends and climate change projections that are not covered in the other sections of this chapter as well as discuss key aspects of the climate system that are particularly relevant for small islands. The *very likely* contribution of mean sea level rise to increased extreme sea levels (see Section 3.5.3), coupled with the *likely* increase in tropical cyclone maximum wind speed (see Section 3.4.4), is a specific issue for tropical small island states.

Although the underlying data sources are limited, some data for the Indian Ocean, South Pacific (Fiji), and Caribbean were available in the studies of Alexander et al. (2006) and Caesar et al. (2011). Problems of data availability and homogeneity for the Caribbean are discussed by T.S. Stephenson et al. (2008). Based on standard extremes indices, positive trends in warm days and warm nights and negative trends in cold days and cold nights[2] have occurred across the Indian Ocean and South Pacific region for the period 1971 to 2005 (Caesar et al., 2011) and the Caribbean for the period 1951 to 2003 (based on data from Alexander et al., 2006). Based on the same data sources, trends in average total wet-day precipitation were positive and statistically significant over the Indian Ocean region, negative over the South Pacific region, and weakly negative over the Caribbean. Trends in heavy and very heavy precipitation were positive over the Indian Ocean, negative over the South Pacific region, and close to zero over the Caribbean. We have *low confidence* in temperature trends over the Indian Ocean and South Pacific region due to the shorter record over which trends were assessed, whereas for the Caribbean, we have *medium confidence* in the temperature trends due to the longer records available for assessment. Because of the spatial heterogeneity exhibited in precipitation trends in general, there is insufficient evidence to assess observed rainfall trends. For the Caribbean, temperatures are projected to increase across the region by 1 to 4°C over 2071-2100 relative to 1961-1990 under the A2 and B2 scenarios and rainfall is mainly projected to decrease by 25 to 50% except in the north (Campbell et al., 2011). Based on this study and the evidence for projected temperature increases reported for other regions (see Table 3-3) we have *medium confidence* in the projected temperature increases for the Caribbean. However, due to the range of processes that contribute to rainfall change, some of which are poorly resolved by GCMs, there is insufficient evidence to assess projected rainfall changes on these small islands.

Given the low elevation of many small islands, sea level extremes are of particular relevance. Sea level extremes are strongly influenced by tidal extremes (Chowdhury et al., 2007; Merrifield et al., 2007). When the tide behavior is mostly semi-diurnal (two high and low tides per day), there will be a clustering of high spring tides around the time of the equinoxes whereas when the tide behavior is diurnal (one high and low tide per day), the clustering of high spring tides will occur around the time of the solstices. In addition, ENSO has a strong influence such that sea levels and their extremes are positively (negatively) correlated with the SOI in the tropical Pacific west (east) of 180° (Church et al., 2006b; Menendez et al., 2010). Tides and ENSO have contributed to the more frequent occurrence of sea level extremes and associated flooding experienced at some Pacific Islands such as Tuvalu in recent years, and make the task of determining the relative roles of these natural effects and mean sea level rise difficult (Lowe et al., 2010). Furthermore, the steep shelf margins that surround many islands and atolls in the Pacific support larger wave-induced contributions to sea level anomalies. Unfortunately, wave observations (including wave direction) that would facilitate more comprehensive studies of tide, surge, and wave extremes in the region are sparse, including those that are co-located with tide gauges (Lowe et al., 2010).

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[2] Termed "cool days" and "cool nights" in that study.

Coral reefs are a feature of many small islands and healthy reef systems mitigate against erosion and inundation by not only providing a buffer zone for the shoreline during extreme surge and wave events but also providing a source of carbonate sand and gravel, which are delivered to the shores by storms and swell to maintain the atoll (Woodroffe, 2008; Webb and Kench, 2010). Anthropogenic oceanic changes may indirectly contribute to extreme impacts for coral atolls by affecting the health of the surrounding reef system. Such changes include: (1) warming of the surface ocean, which slows or prevents growth in temperature-sensitive species and causes more frequent coral bleaching events (e.g., Hoegh-Guldberg, 1999; see also Chapter 4); (2) ocean acidification, caused by increases in atmospheric CO_2 being absorbed into the oceans, which lowers coral growth rates (Bongaerts et al., 2010); and (3) reduction in oxygen concentration in the ocean due to a combination of changes in temperature-driven gas solubility (Whitney et al., 2007), ocean ventilation due to circulation changes, and biological cycling of organic material (Keeling et al., 2010). Quantifying these changes and understanding their impact on coral reef health will be important to understanding the impact of anthropogenic climate change on atolls.

In summary, the small land area and often low elevation of small island states make them particularly vulnerable to rising sea levels and impacts such as inundation, shoreline change, and saltwater intrusion into underground aquifers. Short record lengths and the inadequate resolution of current climate models to represent small island states limit the assessment of changes in extremes. There is insufficient evidence to assess observed trends and future projections in rainfall across the small island regions considered here. The reported increases in warm days and nights and decreases in cold days and nights are of *medium confidence* over the Caribbean and of *low confidence* over the Pacific and Indian Oceans. There is *medium confidence* in the projected temperature increases across the Caribbean. The unique situation of small islands states and their maritime environments leads to an additional emphasis on oceanic information to understand the impacts of climate change. The *very likely* contribution of mean sea level rise to increased coastal high water levels, coupled with the *likely* increase in tropical cyclone maximum wind speed, is a specific issue for tropical small island states.

exacerbated by increasing human-induced pressures (Nicholls et al., 2007). However it was also noted that since coasts are dynamic systems, adaptation to climate change required understanding of processes operating on decadal to century time scales, yet this understanding was least developed.

Because of the diverse and complex nature of coastal impacts, assessments of the future impacts of climate change have focused on a wide range of questions and employed a diverse range of methods, making direct comparison of studies difficult (R.J. Nicholls, 2010). Two types of studies are examined here: the first are assessments, typically undertaken at the country or regional scale and which combine information on physical changes with the socioeconomic implications (e.g., Nicholls and de la Vega-Leinert, 2008); the second type are studies oriented around improved scientific understanding of the impacts of climate change. In terms of coastal assessments, Aunan and Romstad (2008) reported that Norway's generally steep and resistant coastlines contribute to a low physical susceptibility to accelerated sea level rise. Nicholls and de la Vega-Leinert (2008) reported that large parts of the coasts in Great Britain (including England, Wales, and Scotland) are already experiencing widespread sediment starvation and erosion, loss/degradation of coastal ecosystems, and significant exposure to coastal flooding. Lagoons, river deltas, and estuaries are assessed as being particularly vulnerable in Poland (Pruszak and Zawadzka, 2008). In Estonia, Kont et al. (2008) reported increased beach erosion, which is believed to be the result of increased storminess in the eastern Baltic Sea since 1954, combined with a decline in sea ice cover during the winter. Sterr (2008) reported that for Germany there is a high level of reliance on hard coastal protection against extreme sea level hazards, which will increase ecological vulnerability over time. In

France, the Atlantic coast Aquitaine region was considered more resilient to rising sea levels over the coming century because of the sediment storage in the extensive dune systems whereas the sandy coast regions of the Languedoc Roussillon region on the Mediterranean coast were considered more vulnerable because of narrow dune systems that are also highly urbanized (Vinchon et al., 2009). A coastal vulnerability assessment for Australia (Department of Climate Change, 2009) characterized future vulnerability in terms of coastal geomorphology, sediment type, and tide and wave characteristics, from which it concluded that the tropical northern coastline would be most sensitive to changes in tropical cyclone behavior while health of the coral reefs may also influence the tropical eastern coastline. The mid-latitude southern and eastern coastlines were expected to be most sensitive to changes in mean sea level, wave climate, and changes in storminess. A comparative study of the impact of sea level rise on coastal inundation across 84 developing countries showed that the greatest vulnerability to a 1 m sea level rise in terms of inundation of land area was located in East Asia and the Pacific, followed by South Asia, Latin America, and the Caribbean, the Middle East and North Africa, and finally sub-Saharan Africa (Dasgupta et al., 2009).

New models have been developed for the assessment of coastal vulnerability at the global to national level (Hinkel and Klein, 2009). At the local to regional scale, new techniques and approaches have also been developed to better quantify impacts from inundation due to future sea level rise. Bernier et al. (2007) evaluated spatial maps of extreme sea level for different return periods on a seasonal basis that were used to estimate seasonal risk of inundation under future sea level scenarios. McInnes et al. (2009a) developed spatial maps of storm tide

and using a simple inundation model with high-resolution Light Detection and Ranging (LIDAR) data and a land subdivision database, identified the impact of inundation on several coastal towns along the southeastern Australian coastline under future sea level and wind speed scenarios. Probabilistic approaches have also been used to evaluate extreme sea level exceedance under uncertain future sea level rise scenarios. Purvis et al. (2008) constructed a probability distribution around the range of future sea level rise estimates and used Monte Carlo sampling to apply the sea level change to a two-dimensional coastal inundation model. They showed that by evaluating the possible flood-related losses in this framework they were able to represent spatially the higher losses associated with the low-frequency but high-impact inundation events instead of considering only a single midrange scenario. Hunter (2010) combined sea level extremes evaluated from observations with projections of sea level rise to 2100 and showed, for example, that planning levels in Sydney, Australia, would need to be increased substantially to cope with increased risk of flooding. Along the Portuguese coast, Andrade et al. (2007) found that projected future climate in the HadCM3 model would not affect wave height along this coastline but the projected rotation in wave direction would increase the net littoral drift and the erosional response. Along a section of the southeast coast of the United Kingdom, the effect of sea level rise, surge, and wave climate change on the inshore wave climate was evaluated and the frequency and height of extreme waves was projected to increase in the north of the domain (Chini et al., 2010). On the basis of modeling the 25-year beach response along a stretch of the Portuguese coast to various climate change scenarios, Coelho et al. (2009) concluded that the projected stormier wave climate led to higher rates of beach erosion than mean sea level rise. Modeling of the evolution of soft rock shores with rising sea levels has revealed a relatively simple relationship between sea level rise and the equilibrium cliff profile (Walkden and Dickson, 2008).

To summarize, recent observational studies that identify trends and impacts at the coast are limited in regional coverage, which means there is *low confidence*, due to insufficient evidence, that anthropogenic climate change has been a major cause of any observed changes. However, recent coastal assessments at the national and regional scale and process-based studies have provided further evidence of the vulnerability of low-lying coastlines to rising sea levels and erosion, so that in the absence of adaptation there is *high confidence* that locations currently experiencing adverse impacts such as coastal erosion and inundation will continue to do so in the future due to increasing sea levels in the absence of changes in other contributing factors.

3.5.6.   Glacier, Geomorphological, and Geological Impacts

Mountains are prone to mass movements including landslides, avalanches, debris flows, and flooding that can lead to disasters. Changes in the cryosphere affect such extremes, but also water supply and hydropower generation. Many of the world's high mountain ranges are situated at the margins of tectonic plates, increasing the possibility of potentially hazardous interactions between climatic and geological processes. The principal drivers are glacier ice mass loss, mountain permafrost degradation, and possible increases in the intensity of precipitation (Liggins et al., 2010; McGuire, 2010). The possible consequences are changes in mass movement on short contemporary time scales, and modulations of seismicity and volcanic activity on longer, century to millennium time scales.

The AR4 assessed that "the late 20th century glacier wastage likely has been a response to post-1970 warming" (Lemke et al., 2007). However, the impacts of glacier retreat on the natural physical system in the context of changes in extreme events were not assessed in detail. Additionally, the AR4 did not assess geomorphological and geological impacts that might result from anthropogenic climate change. The most studied change in the high-mountain environment has been the retreat of glaciers (Paul et al., 2004; Kaser et al., 2006; Larsen et al., 2007; Rosenzweig et al., 2007). Alpine glaciers around the world were at maximum extent by the end of the Little Ice Age (~1850), and have retreated since then (Leclercq et al., 2011), with an accelerated decay during the past several decades (Zemp et al., 2007). Most glaciers have retreated since the mid-19th century (Francou et al., 2000; Cullen et al., 2006; Thompson et al., 2006; Larsen et al., 2007; Schiefer et al., 2007; Paul and Haeberli, 2008). Rates of retreat that exceed historical experience and internal (natural) variability have become apparent since the beginning of the 21st century (Reichert et al., 2002; Haeberli and Hohmann, 2008).

Outburst floods from lakes dammed by glaciers or unstable moraines [or 'glacial lake outburst floods' (GLOFs)] are commonly a result of glacier retreat and formation of lakes behind unstable natural dams (Clarke, 1982; Clague and Evans, 2000; Huggel et al., 2004; Dussaillant et al., 2010). In the past century, GLOFs have caused disasters in many high-mountain regions of the world (Rosenzweig et al., 2007), including the Andes (Reynolds et al., 1998; Carey, 2005; Hegglin and Huggel, 2008), the Caucasus and Central Asia (Narama et al., 2006; Aizen et al., 2007), the Himalayas (Vuichard and Zimmermann, 1987; Richardson and Reynolds, 2000; Xin et al., 2008; Bajracharya and Mool, 2009; Osti and Egashira, 2009), North America (Clague and Evans, 2000; Kershaw et al., 2005), and the European Alps (Haeberli, 1983; Haeberli et al., 2001; Vincent et al., 2010). However, because GLOFs are relatively rare, it is unclear whether their frequency of occurrence is changing at either the regional or global scale. Clague and Evans (2000) argue that outburst floods from moraine-dammed lakes in North America may have peaked due to a reduction in the number of the lakes since the end of the Little Ice Age. In contrast, a small but not statistically significant increase of GLOF events was observed in the Himalayas over the period 1940 to 2000 (Richardson and Reynolds, 2000), but the event documentation may not be complete. Over the past several decades, human mitigation measures at unstable glacier lakes in the Himalaya and European Alps may have prevented some potential GLOF events (Reynolds, 1998; Haeberli et al., 2001).

Evidence of degradation of mountain permafrost and attendant slope instability has emerged from recent studies in the European Alps

(Gruber and Haeberli, 2007; Huggel, 2009) and other mountain regions (Niu et al., 2005; Geertsema et al., 2006; Allen et al., 2011). This evidence includes several recent rock falls, rock slides, and rock avalanches in areas where permafrost thaw in steep bedrock is occurring. Landslides with volumes ranging up to a few million cubic meters have occurred in the Mont Blanc region (Barla et al., 2000), in Italy (Sosio et al., 2008; Huggel, 2009; Fischer et al., 2011), in Switzerland, and in British Columbia (Evans and Clague, 1998; Geertsema et al., 2006). Very large rock and ice avalanches with volumes of 30 to over 100 million m^3 include the 2002 Kolka avalanche in the Caucasus (Haeberli et al., 2004; Kotlyakov et al., 2004; Huggel et al., 2005), the 2005 Mt. Steller rock avalanche in the Alaska Range (Huggel et al., 2008), the 2007 Mt. Steele ice and rock avalanche in the St. Elias Mountains, Yukon (Lipovsky et al., 2008), and the 2010 Mt. Meager rock avalanche and debris flow in the Coast Mountains of British Columbia.

Quantification of possible trends in the frequency of landslides and ice avalanches in mountains is difficult due to incomplete documentation of past events, especially those that happened before regular satellite observations became available. Nevertheless, there has been an apparent increase in large rock slides during the past two decades, and especially during the first years of the 21st century in the European Alps (Ravanel and Deline, 2011), in the Southern Alps of New Zealand (Allen et al., 2011), and in northern British Columbia (Geertsema et al., 2006) in combination with temperature increases, glacier shrinkage, and permafrost degradation.

Research, however, has not yet provided any clear indication of a change in the frequency of debris flows due to recent deglaciation. Debris flow activity at a local site in the Swiss Alps was higher during the 19th century than today (Stoffel et al., 2005). In the French Alps no significant change in debris flow frequency has been observed since the 1950s in terrain above elevations of 2,200 m (Jomelli et al., 2004). Processes not, or not directly, driven by climate, such as sediment yield, can also be important for changes in the magnitude or frequency of alpine debris flows (Lugon and Stoffel, 2010).

Debris flows from both glaciated and unglaciated volcanoes, termed lahars, can be particularly large and hazardous. Lahars produced by volcanic eruptions on the glacier-clad Nevado del Huila volcano in Colombia in 2007 and 2008 were the largest rapid mass flows on Earth in recent years. Similarly, large mass flows occur on ice-covered active volcanoes in Iceland (Björnsson, 2003), including Eyjafjallajökull in 2010. Large rock and ice avalanches, with volumes up to 30 million m^3, have happened frequently (on average about one every four years) on the glaciated Alaskan volcano, Iliamna, and are thought to be related to elevated volcanic heat flow and possibly meteorological conditions (Huggel et al., 2007). Glacier decay on active volcanoes can lead to a reduction of lahar hazards due to less potential meltwater available for lahar generation, but it is difficult to make a general conclusion as local conditions also play important roles. In 1998, intense rainfall mobilized pyroclastic material on the flanks of Vesuvius and Campi Flegrei volcanoes, feeding approximately 150 debris flows that damaged nearby communities and resulted in 160 fatalities (Bondi and Salvatori, 2003).

In the same year, intense precipitation associated with Hurricane Mitch triggered a small flank collapse at Casita volcano in Nicaragua. This slope failure transformed into debris flows that destroyed two towns and claimed 2,500 lives (Scott et al., 2005). Following the 1991 Pinatubo eruption in the Philippines, heavy rains associated with tropical storms moved large volumes of volcanic sediment. The sediment dammed rivers, causing massive flooding across the region that continued for several years after the eruption ended (Newhall and Punongbayan, 1996).

A variety of climate and weather events can have geomorphological and geological impacts. Warming and degradation of mountain permafrost affect slope stability through a reduction in the shear strength of ice-filled rock discontinuities. For example, the 2003 European summer heat wave (Section 3.3.1) caused rapid thaw and thickening of the active layer, triggering a large number of mainly small rock falls (Gruber et al., 2004; Gruber and Haeberli, 2007). Permafrost thaw in sediment such as in talus slopes may increase both the frequency and magnitude of debris flows (Zimmermann et al., 1997; Rist and Phillips, 2005). The frost table at the base of the active layer is a barrier to groundwater infiltration and can cause the overlying non-frozen sediment to become saturated. Snow cover can also affect debris flow activity by supplying additional water to the soil, increasing pore water pressure and initiating slope failure (Kim et al., 2004). Many of the largest debris flows in the Alps in the past 20 years were triggered by intense rainfall in summer or fall when the snowline was elevated (Rickenmann and Zimmermann, 1993; Chiarle et al., 2007). Warming may increase the flow speed of frozen bodies of sediment (Kääb et al., 2007; Delaloye et al., 2008; Roer et al., 2008). Rock slopes can fail after they have been steepened by glacial erosion or unloaded (debuttressed) following glacier retreat (Augustinus, 1995). Although it may take centuries or even longer for a slope to fail following glacier retreat, recent landslides demonstrate that some slopes can respond to glacier down-wasting within a few decades or less (Oppikofer et al., 2008). Twentieth-century warming may have penetrated some decameters into thawing steep rock slopes in high mountains (Haeberli et al., 1997). Case studies indicate that both small and large slope failures can be triggered by exceptionally warm periods of weeks to months prior to the events (Gruber et al., 2004; Huggel, 2009; Fischer et al., 2011).

The spatial and temporal patterns of precipitation, the intensity and duration of rainfall, and antecedent rainfall are important factors in triggering shallow landslides (Iverson, 2000; Wieczorek et al., 2005; Sidle and Ochiai, 2006). In some regions antecedent rainfall is probably a more important factor than rainfall intensity (Kim et al., 1991; Glade, 1998), whereas in other regions rainfall duration and intensity are the critical factors (Jakob and Weatherly, 2003). Landslides in temperate and tropical mountains that have no seasonal snow cover are not temperature-sensitive and may be more strongly influenced by human activities such as poor land use practices, deforestation, and overgrazing (Sidle and Ochiai, 2006).

Rock and ice avalanches on glaciated volcanoes can be triggered by heat generated by volcanic activity. Their incidence may increase with

rising air and rock temperatures (Gruber and Haeberli, 2007) or during or following brief, anomalously warm events (Huggel et al., 2010) due to meltwater infiltration and shear strength reduction. Debuttressing effects due to glacier retreat can also destabilize or over-steepen slopes (Tuffen, 2010). Furthermore, on volcanoes, geothermal heat flow can enhance ice melting and thus create weak zones at the ice-bedrock interface; and hydrothermal alteration of rocks can decrease the slope stability (Huggel, 2009). On unglaciated high volcanoes in the Caribbean, Central America, Europe, Indonesia, the Philippines, and Japan, an increase in total rainfall or an increase in the frequency or magnitude of severe rainstorms (see Section 3.3.2) could cause more frequent debris flows by mobilizing unconsolidated, volcanic regolith and by raising pore-water pressures, which could lead to deep-seated slope failure. Heavy rainfall events could also influence the behavior of active volcanoes. For example, Mastin (1994) attributes the violent venting of volcanic gases at Mount St. Helens between 1989 and 1991 to slope instability or accelerated growth of cooling fractures within the lava dome following rainstorms, and Matthews et al. (2002) link episodes of intense tropical rainfall with collapses of the Soufriere Hills lava dome on Montserrat in the Caribbean. It is well established that ice mass wastage following the end of the last glaciations led to increased levels of seismicity associated with post-glacial rebound of the lithosphere (e.g., Muir-Wood, 2000; Stewart et al., 2000). There has been a large reduction in glacier cover in southern Alaska. Sauber and Molnia (2004) reported several hundred meters vertical reduction. This ice reduction may be responsible for an increase in seismicity in the region where earthquake faults are at the threshold of failure (Sauber and Molnia, 2004; Doser et al., 2007). An increase in the frequency of small earthquakes in the Icy Bay area, also in southeast Alaska, is interpreted to be a crustal response to glacier wastage between 2002 and 2006 (Sauber and Ruppert, 2008). Large-scale ice mass loss in glaciated volcanic terrain reduces the load on the crust and uppermost mantle, facilitating magma formation and its ascent into the crust (Jull and McKenzie, 1996) and allowing magma to reach the surface more easily (Sigmundsson et al., 2010). At the end of the last glaciation, this mechanism resulted in a more than 10-fold increase in the frequency of volcanic eruptions in Iceland (Sinton et al., 2005).

The AR4 projected that glaciers in mountains will lose additional mass over this century because more ice will be lost due to summer melting than is replenished by winter precipitation (Meehl et al., 2007b). The total area of glaciers in the European Alps may decrease by 20 to more than 50% by 2050 (Zemp et al., 2006; Huss et al., 2008). The projected glacier retreat in the 21st century may form new potentially unstable lakes. Probable sites of new lakes have been identified for some alpine glaciers (Frey et al., 2010). Rock slope and moraine failures may trigger damaging surge waves and outburst floods from these lakes. The temperature rise also will result in gradual degradation of mountain permafrost (Haeberli and Burn, 2002; Harris et al., 2009). The zone of warm permafrost (mean annual rock temperature approximately -2 to 0°C), which is more susceptible to slope failures than cold permafrost, may rise in elevation a few hundred meters during the next 100 years (Noetzli and Gruber, 2009). This in turn may shift the zone of enhanced

instability and landslide initiation toward higher-elevation slopes that in many regions are steeper, and therefore predisposed to failure. The response of bedrock temperatures to surface warming through thermal conduction will be slow, but warming will eventually penetrate to considerable depths in steep rock slopes (Noetzli et al., 2007). Other heat transport processes such as advection, however, may induce warming of bedrock at much faster rates (Gruber and Haeberli, 2007). The response of firn and ice temperatures to an increase in air temperature is faster and nonlinear (Haeberli and Funk, 1991; Suter et al., 2001; Vincent et al., 2007). Latent heat effects from refreezing meltwater can amplify the increase in air temperature in firn and ice (Huggel, 2009; Hoelzle et al., 2010). At higher temperatures, more ice melts and the strength of the remaining ice is lower; as a result, the frequency and perhaps size of ice avalanches may increase (Huggel et al., 2004; Caplan-Auerbach and Huggel, 2007). Warm extremes can trigger large rock and ice avalanches (Huggel et al., 2010).

Current low levels of seismicity in Antarctica and Greenland may be a consequence of ice-sheet loading, and isostatic rebound associated with accelerated deglaciation of these regions may result in an increase in earthquake activity, perhaps on time scales as short as 10 to 100 years (Turpeinen et al., 2008; Hampel et al., 2010). Future ice mass loss on glaciated volcanoes, notably in Iceland, Alaska, Kamchatka, the Cascade Range in the northwest United States, and the Andes, could lead to eruptions, either as a consequence of reduced load pressures on magma chambers or through increased magma-water interaction. Reduced ice load arising from future thinning of Iceland's Vatnajökull Ice Cap is projected to result in an additional 1.4 km^3 of magma produced in the underlying mantle every century (Pagli and Sigmundsson, 2008). Ice unloading may also promote failure of shallow magma reservoirs with a potential consequence of a small perturbation of the natural eruptive cycle (Sigmundsson et al., 2010). Initially, ice thinning of 100 m or more on volcanoes with glaciers more than 150-m thick, such as Sollipulli in Chile, may cause more explosive eruptions, with increased tephra hazards (Tuffen, 2010). Additionally, the potential for edifice lateral collapse could be enhanced by loss of support previously provided by ice (Tuffen, 2010) or to elevated pore-water pressures arising from meltwater (Capra, 2006; Deeming et al., 2010). Ultimately the loss of ice cover on glaciated volcanoes may reduce opportunities for explosions arising from magma-ice interaction. The incidence of ice-sourced lahars may also eventually fall, although exposure of new surfaces of volcanic debris due to ice wastage may provide the raw material for precipitation-related lahars. The likelihood of both volcanic and non-volcanic landslides may also increase due to greater availability of water, which could destabilize slopes. Many volcanoes provide a ready source of unconsolidated debris that can be rapidly transformed into potentially hazardous lahars by extreme precipitation events. Volcanoes in coastal, near-coastal, or island locations in the tropics are particularly susceptible to torrential rainfall associated with tropical cyclones, and the rainfall rate associated with tropical cyclones is projected to increase though the number of tropical cyclones is projected to decrease or stay essentially unchanged (see Section 3.4.4). The impact of future large explosive volcanic eruptions may also be exacerbated by an

increase in extreme precipitation events (see Section 3.3.2) that provide an effective means of transferring large volumes of unconsolidated ash and pyroclastic flow debris from the flanks of volcanoes into downstream areas.

Quantification of possible trends in the frequency of landslides and ice avalanches in mountains is difficult due to incomplete documentation of past events. There is *high confidence* that changes in heat waves, glacial retreat, and/or permafrost degradation will affect high mountain phenomena such as slope instabilities, mass movements, and glacial lake outburst floods, and *medium confidence* that temperature-related changes will influence bedrock stability. There is also *high confidence* that changes in heavy precipitation will affect landslides in some regions. There is *medium confidence* that high-mountain debris flows will begin earlier in the year because of earlier snowmelt, and that continued mountain permafrost degradation and glacier retreat will further decrease the stability of rock slopes. There is *low confidence* regarding future locations and timing of large rock avalanches, as these depend on local geological conditions and other non-climatic factors. There is *low confidence* in projections of an anthropogenic effect on phenomena such as shallow landslides in temperate and tropical regions, because these are strongly influenced by human activities such as poor land use practices, deforestation, and overgrazing. It is well established that ice mass wastage following the end of the last glaciations led to increased levels of seismicity, but there is *low confidence* in the nature of recent and projected future seismic responses to anthropogenic climate change.

3.5.7.   High-latitude Changes Including Permafrost

Permafrost is widespread in Arctic, in subarctic, in ice-free areas of Antarctica, and in high-mountain regions, and permafrost regions occupy approximately 23 million km^2 of land area in the Northern Hemisphere (Zhang et al., 1999). Melting of massive ground ice and thawing of ice-rich permafrost can lead to subsidence of the ground surface and to the formation of uneven topography known as thermokarst, having implications for ecosystems, landscape stability, and infrastructure performance (Walsh, 2005). See also Case Study 9.2.10 for discussion of the impacts of cold events in high latitudes. The active layer (near-surface layer that thaws and freezes seasonally over permafrost) plays an important role in cold regions because most ecological, hydrological, biogeochemical, and pedogenic (soil-forming) activity takes place within it (Hinzman et al., 2005).

Observations show that permafrost temperatures have increased since the 1980s (IPCC, 2007b). Temperatures in the colder permafrost of northern Alaska, the Canadian Arctic, and Russia have increased up to 3°C near the permafrost table and up to 1 to 2°C at depths of 10 to 20 m (Osterkamp, 2007; Romanovsky et al., 2010; S.L. Smith et al., 2010) since the late 1970s/early 1980s. Temperature increases have generally been less than

1°C in the warmer permafrost of the discontinuous permafrost zone of the polar regions (Osterkamp, 2007; Romanovsky et al., 2010; S.L. Smith et al., 2010), and also in the high-altitude permafrost of Mongolia and the Tibetan Plateau (Zhao et al., 2010). When the other conditions remain constant, active layer thickness is expected to increase in response to warming. Active layer thickness has increased by about 20 cm in the Russian Arctic between the early 1960s and 2000 (T. Zhang et al., 2005) and by up to 1.0 m over the Qinghai-Tibetan Plateau since the early 1980s (Wu and Zhang, 2010), with no significant trend in the North American Arctic since the early 1990s (Shiklomanov et al., 2010). However, over extreme warm summers, active layer thickness may increase substantially (Smith et al., 2009), potentially triggering active-layer detachment failures on slopes (Lewkowicz and Harris, 2005). Extensive thermokarst development has been found in Alaska (Jorgenson et al., 2006; Osterkamp et al., 2009), Canada (Vallée and Payette, 2007), and central Yakutia (Gavriliev and Efremov, 2003). Increased rates of retrogressive thaw slump activities have been reported on slopes over the Qinghai-Tibetan Plateau (Niu et al., 2005) and adjacent to tundra lakes over the Mackenzie Delta region of Canada (Lantz and Kokelj, 2008). Substantial expansion and deepening of thermokarst lakes was observed near Yakutsk with subsidence rates of 17 to 24 cm yr^{-1} from 1992 to 2001 (Fedorov and Konstantinov, 2003). Satellite remote sensing data show that thaw lake surface area has increased in continuous permafrost regions and decreased in discontinuous permafrost regions (Smith et al., 2005). Coasts with ice-bearing permafrost that are exposed to the Arctic Ocean are very sensitive to permafrost degradation. Some Arctic coasts are retreating at a rapid rate of 2 to 3 m yr^{-1} and the rate of erosion along Alaska's northeastern coastline has doubled over the past 50 years, related to declining sea ice extent, increasing sea surface temperature, rising sea level, thawing coastal permafrost, and possibly increases in storminess and waves (Jones et al., 2009; Karl et al., 2009)

Increases in air temperature are in part responsible for the observed increase in permafrost temperature over the Arctic and subarctic, but changes in snow cover also play a critical role (Osterkamp, 2005; Zhang, 2005; T. Zhang et al., 2005; S.L. Smith et al., 2010). Trends toward earlier snowfall in autumn and thicker snow cover during winter have resulted in a stronger snow insulation effect, and as a result a much warmer permafrost temperature than air temperature in the Arctic. On the other hand, permafrost temperature may decrease even if air temperature increases, if there is also a decrease in the duration and thickness of snow cover (Taylor et al., 2006). The lengthening of the thaw season and increases in summer air temperature have resulted in changes in active layer thickness. Model simulations have projected thickening of the active layer, a northward shift of the permafrost boundary, reductions in permafrost area, and an increase in permafrost temperature in the 21st century and beyond (Saito et al., 2007; Schaefer et al., 2011). The projected permafrost degradation may result in ancient carbon currently frozen in permafrost being released into the atmosphere, providing a positive feedback to the climate system (Schaefer et al., 2011). Expansion of lakes in the continuous permafrost zone may be due to thawing of ice-rich permafrost and melting of massive ground ice, while decreases in lake area in the discontinuous permafrost zone may be due to lake

bottom drainage (Smith et al., 2005). Overall, increased air temperature over high latitudes is primarily responsible for the development of thermokarst terrains and thaw lakes.

In summary, it is *likely* that there has been warming of permafrost in recent decades. There is *high confidence* that permafrost temperatures will continue to increase, and that there will be increases in active layer thickness and reductions in the area of permafrost in the Arctic and subarctic.

3.5.8.   Sand and Dust Storms

Sand and dust storms are widespread natural phenomena in many parts of the world. Heavy dust storms disrupt human activities. Dust aerosols in the atmosphere can cause a suite of health impacts including respiratory problems (Small et al., 2001). The long-range transport of dust can affect conditions at long distances from the dust sources, linking the biogeochemical cycles of land, atmosphere, and ocean (Martin and Gordon, 1988; Bergametti and Dulac, 1998; Kellogg and Griffin, 2006). For example, dust from the Saharan region and from Asia may reach North America and South America (McKendry et al., 2007). Some climate models have representations of dust aerosols (Textor et al., 2006). Climate variables that are most important to dust emission and transport such as soil moisture (see also Section 3.5.1), precipitation, wind, and vegetation cover are still subject to large uncertainties in climate model simulations. As a result, the sand and dust storm simulations have large uncertainties as well.

The Sahara (especially the Bodélé Depression in Chad) and east Asia have been recognized as the largest dust sources globally (Goudie, 2009). Over the few decades before the 1990s, the frequency of dust events increased in some regions such as the Sahel zone of Africa (Goudie and Middleton, 1992), and decreased in some other regions such as China (Zhang et al., 2003). There seems to be an increase in more recent years in China (Shao and Dong, 2006). Despite the importance of African dust, studies on long-term change in Sahel dust are limited. However, dust transported far away from the source region may provide some evidence of long-term changes in the Sahel region. The African dust transported to Barbados began to increase in the late 1960s and through the 1970s; transported dust reached a peak in the early 1980s but remains high into the present (Prospero and Lamb, 2003; Prospero et al., 2009).

Surface soil dust concentration during a sand and dust storm is controlled by a number of factors. The driving force for the production of dust storms is the surface wind associated with cold frontal systems sweeping across arid and semi-arid regions and lifting soil particles in the atmosphere. Dust emissions are also controlled by the surface conditions in source regions such as the desert coverage distributions, snow cover, and soil moisture. For example, in the Sahel region, the elevated high level of dust emission prior to the 1990s was related to the persistent drought during that time, and to long-term changes in the NAO (Ginoux et al., 2004; Chiapello et al., 2005; Engelstaedter et al., 2006), and perhaps to North Atlantic SST as well (Wong et al., 2008). Further evidence of the importance of climate on dust emission is that despite an increase of approximately 2 to 7% in desert areas in China over the four decades since 1960, dust storm frequency decreased in that period (Zhong, 1999). Studies on Asian soil dust production from 1960 to 2003 suggest that climatic variations have played a major role in the declining trends in dust emission and storm frequencies in China (Zhang et al., 2003; Zhou and Zhang, 2003; Zhao et al., 2004; Gong et al., 2006). Overall, changes in dust activity are affected by changes in the climate, such as wind and moisture conditions in the dust source regions. Changes in large-scale circulation play an additional role in the long-distance transport of dust. However, understanding of the physical mechanisms of the long-term trends in dust activity is not complete; for example, the relative importance of the various factors affecting dust frequency as outlined above is uncertain.

Future dust activity depends on two main factors: land use in the dust source regions, and climate both in the dust source region and large-scale circulation that affects long distance dust transport. Studies on projected future dust activity are very limited. It is difficult to project future land use. Precipitation, soil moisture, and runoff have been projected to decrease in major dust source regions (Figure 10.12 in Meehl et al., 2007b). Thomas et al. (2005) suggest that dune fields in southern Africa can become active again, and sand will become significantly exposed and move, as a consequence of 21st-century warming. A study based on simulations from two climate models also suggests increased desertification in arid and semi-arid China, especially in the second half of the 21st century (X.M. Wang et al., 2009). However, confident projected changes in wind are lacking (see Section 3.3.3).

In summary, there is *low confidence* in projecting future dust storm changes, although an increase could be expected where aridity increases. There is a lack of data and studies on past changes. There is also a lack of understanding of processes such as the relative importance of different climate variables affecting dust storms, as well as a high uncertainty in simulating important climate variables such as soil moisture, precipitation, and wind that affect dust storms.

Table 3-2 | Regional observed changes in temperature and precipitation extremes, including dryness, since 1950 unless indicated otherwise, and using late 20th-century values as reference (see Box 3-1), generally 1961-1990. See Figure 3-1 for definitions of regions. For assessments for small island states refer to Box 3-4.

A. North America and Central America

Regions	Tmax [WD = Warm Days; see Box 3-1] (using late 20th-century extreme values as reference, e.g., 90th/10th percentile)	Tmin [WN = Warm Nights; CN = Cold Nights; see Box 3-1] (using late 20th-century extreme values as reference, e.g., 90th/10th percentile)	Heat Waves (HW)/Warm Spells (WS) [WSDI = Warm Spell Duration Index, i.e., number or fraction of days belonging to spells of at least 6 days with Tmax >90th percentile] (using late 20th-century extreme values as reference)	Heavy Precipitation (HP) (using late 20th-century extreme values as reference, e.g., 90th percentile)	Dryness [CDD = Consecutive Dry Days; SMA = (Simulated) Soil Moisture Anomalies; PDSI = Palmer Drought Severity Index; see Box 3-3 for definitions]
All North America and Central America	*High confidence:* Likely overall increase in WD, decrease in CD (Aguilar et al., 2005; Alexander et al., 2006).	*High confidence:* Likely overall decrease in CN, increase in WN (Aguilar et al., 2006; Alexander et al., 2006).	*Medium confidence:* Overall increase since 1960 (Kunkel et al., 2008). Some areas with significant WSDI increase, others with insignificant WSDI increase or decrease (Alexander et al., 2006).	*High confidence:* Likely increase in many areas since 1950 (Aguilar et al., 2005; Alexander et al., 2006; Trenberth et al., 2007; Kunkel et al., 2008).	*Medium confidence:* Overall slight decrease in dryness (SMA, PDSI, CDD) since 1950; regional variability and 1930s drought dominate the signal (Aguilar et al., 2005; Alexander et al., 2006; Kunkel et al., 2008; Sheffield and Wood, 2008a; Dai, 2011).
W. North America (WNA, 3)	*High confidence:* Very likely large increases in WD, large decreases in CD (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*High confidence:* Very likely large decreases in CN, large increases in WN (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*Medium confidence:* Increase in WSDI (Alexander et al., 2006).	*Medium confidence:* Spatially varying trends. General increase, decrease in some areas (Alexander et al., 2006).	*Medium confidence:* No overall or slight decrease in dryness (SMA, PDSI, CDD) since 1950; large variability, large drought of 1930s dominates (Alexander et al., 2006; Kunkel et al., 2008; Sheffield and Wood, 2008a; Dai, 2011).
Central North America (CNA, 4)	*Medium confidence:* Spatially varying trends. Small increases in WD, decreases in CD in north CNA. Small decreases in CD in south CNA (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*Medium confidence:* Spatially varying trends. Small decreases in CN, increases in WN in north CNA. Small increases in CN, decreases in WN in south CNA (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*Medium confidence:* Spatially varying trends. Some areas with WSDI increase, others with WSDI decrease (Alexander et al., 2006).	*High confidence:* Very likely increase since 1950 (Alexander et al., 2006).	*Medium confidence:* Decrease in dryness (SMA, PDSI, CDD) and increase in mean precipitation since 1950; large variability, large drought of 1930s dominates (Alexander et al., 2006; Kunkel et al., 2008; Sheffield and Wood, 2008a; Dai, 2011).
E. North America (ENA, 5)	*Medium confidence:* Spatially varying trends. Overall increases in WD, decreases in CD; opposite or insignificant signal in a few areas (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*Medium confidence:* Weak and spatially varying trends. (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*Medium confidence:* Spatially varying trends. Many areas with WSDI increase, some areas with WSDI decrease (Alexander et al., 2006).	*High confidence:* Very likely increase since 1950 (Alexander et al., 2006).	*Medium confidence:* Slight decrease in dryness (SMA, PDSI, CDD) since 1950, large variability, large drought of 1930s dominates (Alexander et al., 2006; Kunkel et al., 2008; Sheffield and Wood, 2008a; Dai, 2011).
Alaska/N.W. Canada (ALA, 1)	*High confidence:* Very likely large increases in WD, large decreases in CD (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*High confidence:* Very likely large decreases in CN, large increases in WN (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*Low confidence:* Insufficient evidence.	*Medium confidence:* Slight tendency for increase, in southern Alaska; no significant trend (Kunkel et al., 2008).	*Medium confidence:* Inconsistent trends; increase in dryness (SMA, PDSI, CDD) since 1950 in part of the region. (Alexander et al., 2006; Kunkel et al., 2008; Sheffield and Wood, 2008a; Dai, 2011).
E. Canada, Greenland, Iceland (CGI, 2)	*High confidence:* Likely increases in WD in some areas, decrease in others. Decreases in CD in some areas, increase in others (Robeson, 2004; Alexander et al., 2006; Vincent and Mekis, 2006; Trenberth et al., 2007; Kunkel et al., 2008; Peterson et al., 2008a).	*Medium confidence:* Small increases in unusually cold nights, decreases in WN in northeastern Canada. Small decreases in CN, increases in WN in southeastern and south central Canada. (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*Medium confidence:* Some areas with WSDI increase, most others with WSDI decrease (Alexander et al., 2006).	*Medium confidence:* Increase in a few areas (Alexander et al., 2006).	*Low confidence:* Insufficient evidence.
Central America and Mexico (CAM, 6)	*Medium confidence:* Increases in WD, decreases in CD (Aguilar et al., 2005; Alexander et al., 2006).	*Medium confidence:* Decreases in CN, increases in WN (Aguilar et al., 2005; Alexander et al., 2006).	*Low confidence:* Spatially varying trends. A few areas increase, a few others decrease (Aguilar et al., 2005; Alexander et al., 2006).	*Medium confidence:* Spatially varying trends. Increase in many areas, decrease in a few areas, (Aguilar et al., 2005; Alexander et al., 2006).	*Low confidence:* Spatially varying trends, inconsistencies in trends in dryness (SMA, PDSI, CDD). (Aguilar et al., 2005; Sheffield and Wood, 2008a; Dai, 2011).

Table 3-2 (continued)

B. Europe and Mediterranean Region

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
All Europe and Mediterranean Region	*High confidence*: Overall *likely* increase in WD and *likely* decrease of CD over most of the continent since 1950. Strong increasing tendency in WD in most regions since 1976 onward; small or insignificant decrease in CD over same period (Alexander et al., 2006; see also entries for individual subregions).	*High confidence*: Overall *likely* increase in WN and *likely* decrease in CN over most of the continent since 1950. Strong increasing tendency in WN in most regions since 1976 onward; small or insignificant decrease in CN over same period (Klein Tank and Können, 2003; Alexander et al., 2006; see also entries for individual subregions).	*Medium confidence*: Increase of HW since 1950. Overall consistent positive trend of WSDI across Europe, but no coherent region with significant trends (Alexander et al., 2006). Availability of a few single studies for specific regions (see below).	*Medium confidence*: Increase in part of the region, mostly in winter, insignificant or inconsistent changes elsewhere, in particular in summer. Some inconsistencies in overall patterns between studies depending on considered indices. Most consistent signal over central W. Europe and European Russia (Klein Tank and Können, 2003; Haylock and Goodess, 2004; Alexander et al., 2006; Zolina et al., 2009).	*Medium confidence*: Inconsistent trends. Increase in dryness (SMA, PDSI, CDD) in part of the region; insignificant, inconsistent, or no changes elsewhere. Most consistent signal for increase in dryness in central and S. Europe since the 1950s. No signal in N. Europe (Kiktev et al., 2003; Haylock and Goodess, 2004; Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).
N. Europe (NEU, 11)	*Medium confidence*: Increase in WD and decrease in CD. Consistent signals for whole region, but generally not significant at the local scale (Alexander et al., 2006).	*Medium confidence*: Increase in WN and decrease in CN. Consistent signals over whole region but generally not significant at the local scale (Klein Tank and Können, 2003; Alexander et al., 2006).	*Medium confidence*: Increase in HW. Consistent tendency for increase in WSDI, but no significant trends (Alexander et al., 2006).	*Medium confidence*: Increase in winter in some areas, but often insignificant or inconsistent trends at subregional scale, in particular in summer (Fowler and Kilsby, 2003; Kiktev et al., 2003; Klein Tank and Können, 2003; Alexander et al., 2006; Maraun et al., 2008; Zolina et al., 2009).	*Medium confidence*: Spatially varying trends. Overall only slight or no increase in dryness (SMA, PDSI, CDD), slight decrease in dryness in part of the region (Kiktev et al., 2003; Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).
Central Europe (CEU, 12)	*High confidence*: *Likely* overall increase in WD and *likely* decrease in CD since 1950 in most regions. Some regional and temporal variations in significance of trends. *High confidence*: *Very likely* increase in WD since 1950, 1901 and 1880 and *likely* decrease in CD since 1950 and 1901 in west Central Europe (Alexander et al., 2006; Della-Marta et al., 2007a; Laurent and Parey, 2007). *Medium confidence*: Lower confidence in trends in east Central Europe due to lack of literature, partial lack of access to observations, overall weaker signals, and change point in trends at the end of the 1970s / beginning of 1980s. Strongest increase in WD since 1976 (Alexander et al., 2006; Bartholy and Pongracz, 2007; Hirschi et al., 2011).	*High confidence*: *Likely* overall increase in WN and *likely* overall decrease in CN at the yearly time scale. Some regional and seasonal variations in significance and in a few cases also the sign of the trends. *High confidence*: *Very likely* increase in WN and *very likely* decrease in CN since 1950 and 1901 in west Central Europe (Kiktev et al., 2003; Alexander et al., 2006). *Medium confidence*: Lower confidence in trends in east Central Europe due to lack of literature, partial lack of access to observations, overall weaker signals, and change point in trends at the end of the 1970s / beginning of 1980s. (Klein Tank and Können, 2003; Alexander et al., 2006; Bartholy and Pongracz, 2007).	*Medium confidence*: Increase in heat waves. Consistent tendency for WSDI increase but no significant trends (Alexander et al., 2006). Significant increase in max HW duration since 1880 in west Central Europe in summer (JJA) (Della-Marta et al., 2007a). Less significant signal in heat wave indices in east Central Europe due to presence of change point (Bartholy and Pongracz, 2007; Hirschi et al., 2011).	*Medium confidence*: Increase in part of the domain, in particular in central W. Europe and European Russia, especially in winter. Insignificant or inconsistent trends elsewhere, in particular in summer (Kiktev et al., 2003; Klein Tank and Können, 2003; Schmidli and Frei, 2005; Alexander et al., 2006; Barthoły and Pongracz, 2007; Kyselý, 2009; Tomassini and Jacob, 2009; Zolina et al., 2009).	*Medium confidence*: Spatially varying trends. Increase in dryness (SMA, PDSI, CDD) in part of the region but some regional variation in dryness trends and dependence of trends on considered studies (index, time period) (Kiktev et al., 2003; Alexander et al., 2006; Barthoły and Pongracz, 2007; Sheffield and Wood, 2008a; Brázdil et al., 2009; Dai, 2011).
S. Europe and Mediterranean (MED, 13)	*High confidence*: *Likely* increase in WD and *likely* decrease in CD in most of the region. Some regional and temporal variations in significance of trends. *Likely* strongest and most significant trends in the Iberian Peninsula and southern France (Alexander et al., 2006; Brunet et al., 2007; Della-Marta et al., 2007a; Bartolini et al., 2008; Kuglitsch et al., 2010; Rodríguez-Puebla et al., 2010; Hirschi et al., 2011). *Medium confidence*: Smaller or less significant trends in S.E. Europe and Italy due to change point in trends at the end of the 1970s / beginning of 1980s; sometimes linked with changes in sign of trends; strongest WD increase since 1976 (Barthoły and Pongracz, 2007; Bartolini et al., 2008; Toreti and Desiato, 2008; Kuglitsch et al., 2010; Hirschi et al., 2011).	*High confidence*: *Likely* increase in WN and *likely* decrease in CN in most of the region. Some regional variations in significance of trends. *Very likely* overall increase in WN and *very likely* overall decrease in CN in S.W. Europe and W. Mediterranean; *likely* strongest signals in Spain and southern France (Kiktev et al., 2003; Klein Tank and Können, 2003; Alexander et al., 2006; Brunet et al., 2007; Rodríguez-Puebla et al., 2010). *Likely* overall tendency for increase in WN and *likely* overall tendency for decrease in CN in S.E. Europe and E. Mediterranean (Kiktev et al., 2003; Klein Tank and Können, 2003; Alexander et al., 2006).	*High confidence*: *Likely* overall increase in HW in summer (JJA). Significant increase in max HW duration since 1880 in Iberian Peninsula and west Central Europe in JJA (Della-Marta et al., 2007a). Significant increase in max HW duration in Tuscany (Italy) (Bartolini et al., 2008). Significant increase in HW indices in Turkey and to a smaller extent in S.E. Europe and Turkey in JJA (Kuglitsch et al., 2010). Less significant signal in HW indices in S.E. Europe due to presence of change point in trends (Barthoły and Pongracz, 2007; Hirschi et al., 2011).	*Low confidence*: Inconsistent trends within domain and across studies (Kiktev et al., 2003; Klein Tank and Können, 2003; Alexander et al., 2006; García et al., 2007; Pavan et al., 2008; Zolina et al., 2009; Rodrigo, 2010).	*Medium confidence*: Overall increase in dryness (SMA, PDSI, CDD), but partial dependence on index and time period (Kiktev et al., 2003; Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).

Table 3-2 (continued)

C. Africa

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
All Africa	*Low confidence to medium confidence: Low confidence* due to insufficient evidence (lack of literature) in many regions. *Medium confidence* in increase in frequency of WD and decrease in frequency of CD in southern part of continent (Alexander et al., 2006). See also regional assessments.	*Low confidence to medium confidence* depending on region: *Low confidence* due to insufficient evidence (lack of literature) in many regions. *Medium confidence* in increase in frequency of WN in northern and southern part of continent (Alexander et al., 2006). *Medium confidence* in decrease in frequency of CN in southern part of continent (Alexander et al., 2006). See also regional assessments.	*Low confidence:* Insufficient evidence (lack of literature). Some analyses for localized regions (see regional assessments).	*Low confidence:* Partial lack of data and literature and inconsistent patterns in existing studies (New et al., 2006; Aguilar et al., 2009). See also regional assessments.	*Medium confidence:* Overall increase in dryness (SMA, PDSI); regional variability. 1970s prolonged Sahel drought dominates (Sheffield and Wood, 2008a; Dai, 2011). No apparent continent-wide trends in change in rainfall over the 20th century, although there was a continent-wide drought in 1983 and 1984 (Hulme et al., 2001). Wet season arrives 9–21 days later, large inter-annual variability of wet season start, local-scale geographical variability (Kniveton et al., 2009).
W. Africa (WAF, 15)	*Medium confidence:* Significant increase in temperature of warmest day and coldest day, significant increase in frequency of WD, and significant decrease of frequency of CD in western central Africa, Guinea Conakry, Nigeria, and Gambia (New et al., 2006; Aguilar et al., 2009). *Low confidence:* Lack of literature in other parts of the region.	*Medium confidence:* Decreases in frequency of CN in western central Africa, Nigeria, and Gambia; insignificant decreases in frequency of CN in Guinea Conakry (New et al., 2006; Aguilar et al., 2009). *Low confidence:* Lack of literature on changes in CN in other parts of the region. *Medium confidence:* Increases in frequency of WN (Alexander et al., 2006; New et al., 2006; Aguilar et al., 2009).	*Low confidence:* Insufficient evidence (lack of literature) for most of the region; increases in WSDI in Nigeria and Gambia (New et al., 2006).	*Medium confidence:* Precipitation from heavy events has decreased (western central Africa, Guinea Conakry) but low spatial coherence (Aguilar et al., 2009), rainfall intensity increased (New et al., 2006).	*Medium confidence:* 1970s prolonged Sahel drought dominates, conditions are still drier (SMA, PDSI, precipitation anomalies) than during the humid 1950s (L'Hôte et al., 2002; Dai et al., 2004; Sheffield and Wood, 2008a; Dai, 2011). Dry spell duration (CDD) overall increased from 1961 to 2000 (New et al., 2006). Recent years characterized by a greater interannual variability than previous 40 years, western Sahel remaining dry and the eastern Sahel returning to wetter conditions (Ali and Lebel, 2009).
E. Africa (EAF, 16)	*Low confidence:* Lack of evidence due to lack of literature and spatially non-uniform trends. Over time period 1939–1992 spatially non-uniform trends in daytime temperature, some areas with cooling (King'uyu et al., 2000). In southern tip of domain increases in WD, decreases in CD (Alexander et al., 2006).	*Medium confidence:* Over time period 1939–1992, spatially non-uniform trends, rise of nighttime temperature at several locations, but with many coastal areas and stations near large water bodies showing a significant decrease (King'uyu et al., 2000). In southern tip of domain, decreases in CN, increases in WN (Alexander et al., 2006).	*Low confidence:* Insufficient evidence (lack of literature) for most of region; increase in WSDI in southern tip of domain (New et al., 2006).	*Low confidence:* Insufficient evidence (lack of literature) to assess trends.	*Low confidence:* Spatially varying trends in dryness (SMA, PDSI) (Sheffield and Wood, 2008a; Dai, 2011).
S. Africa (SAF, 17)	*Medium confidence:* Increases in WD, decreases in CD (Alexander et al., 2006).	*Medium confidence:* Decreases in CN, increases in WN (King'uyu et al., 2000; Alexander et al., 2006).	*Medium confidence:* Increase in WSDI (New et al., 2006).	*Low confidence:* No spatially coherent patterns of trends in precipitation extremes (Kruger, 2006; New et al., 2006; Trenberth et al., 2007).	*Medium confidence:* Slight dry spell duration increase (Alexander et al., 2006; Kruger, 2006; New et al., 2006). General increase in dryness (SMA, PDSI) (Sheffield and Wood, 2008a; Dai, 2011).
Sahara (SAH, 14)	*Low confidence:* Lack of literature.	*Medium confidence:* Increases in WN (Alexander et al., 2006). *Low confidence:* Lack of literature on trends in CN.	*Low confidence:* Insufficient evidence (lack of literature).	*Low confidence:* Insufficient evidence.	*Low confidence:* Limited data, spatial variation in the trends (Dai, 2011).

Table 3-2 (continued)

D. South America

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
All South America	*Low confidence to medium confidence depending on region: Low confidence* in trends in WD and CD due to insufficient evidence in many regions, in particular in northern half of continent. *Medium confidence* in trends in southern half of continent but often spatially varying trends. See regional assessments for details and basis for continental assessment.	*Low confidence to medium confidence depending on region: Low confidence* in trends in WN and CN due to insufficient evidence in many regions, in particular in northern half of continent. *Medium confidence* in decrease in CN and increase in WN in southern half of continent. See regional assessments for details and basis for continental assessment.	*Low confidence:* Insufficient evidence (most of continent) or lack of coherent signal (Southeast South America). See regional assessments for details and basis for continental assessment.	*Low confidence:* Insufficient evidence or spatially varying trends. See regional assessments for details and basis for continental assessment.	*Low confidence:* Spatially varying trends, inconsistencies between studies (Sheffield and Wood, 2008a; Dai, 2011).
Amazon (AMZ, 7)	*Low confidence:* Insufficient, scattered, evidence (Alexander et al., 2006; Dufek et al., 2008).	*Low confidence:* Insufficient, scattered, evidence (Alexander et al., 2006; Dufek et al., 2008).	*Low confidence:* Insufficient evidence.	*Medium confidence:* Spatially varying trends. Increase in many areas, decrease in a few areas (Alexander et al., 2006; Haylock et al., 2006b).	*Low confidence:* Spatially varying trends, mixed results. Slight decrease in CDD (Dufek et al., 2008). Tendency to decreased dryness in much of region, but some opposite trends and inconsistencies between studies (Sheffield and Wood, 2008a; Dai, 2011).
N.E. Brazil (NEB, 8)	*Medium confidence:* Increases in WD (Silva and Azevedo, 2008).	*Medium confidence:* Increases in WN (Silva and Azevedo, 2008).	*Low confidence:* Insufficient evidence.	*Medium confidence:* Increase in many areas, decrease in a few areas, (Alexander et al., 2006; Haylock et al., 2006b; Santos and Brito, 2007; Dufek et al., 2008; Silva and Azevedo, 2008; Santos et al., 2009).	*Low confidence:* Spatially varying trends. Inconsistent trends in CDD (Santos and Brito, 2007; Dufek et al., 2008; Silva and Azevedo, 2008; Santos et al., 2009). Inconsistent trends in dryness (SMA, PDSI) between studies (Sheffield and Wood, 2008a; Dai, 2011).
S.E. South America (SSA, 10)	*Medium confidence:* Spatially varying trends. Increases in WD in some areas, decrease in others. Decreases in CD in some areas, increase in others (Rusticucci and Barrucand, 2004; Vincent et al., 2005; Alexander et al., 2006; Rusticucci and Renom, 2008; Marengo et al., 2009b).	*Medium confidence:* Decreases in CN, increases in WN (Rusticucci and Barrucand, 2004; Vincent et al., 2005; Alexander et al., 2006; Rusticucci and Renom, 2008; Marengo et al., 2009b).	*Low confidence:* Spatially varying trends. Some areas increase, others decrease (Alexander et al., 2006).	*Low confidence to medium confidence, depending on subregion: Medium confidence:* Increase in northern portion of domain (Alexander et al., 2006; Dufek et al., 2008; Sugahara et al., 2009; Penalba and Robledo, 2010). *Low confidence* in southern portion of domain: Insufficient evidence.	*Low confidence:* Slight increase in dryness, large variability (Haylock et al., 2006b; Dufek and Ambrizzi, 2008; Dufek et al., 2008; Llano and Penalba, 2011). Decrease in dryness (SMA, PDSI) in much of region (Sheffield and Wood, 2008a; Dai, 2011).
W. Coast South America (WSA, 9)	*Medium confidence:* Increases in WD in some areas, decrease in others. Decreases in CD in some areas, increase in others (Rosenbluth et al., 1997; Vincent et al., 2005; Alexander et al., 2006).	*Medium confidence:* Decreases in CN, increases in WN (Rosenbluth et al., 1997; Vincent et al., 2005; Alexander et al., 2006).	*Low confidence:* Insufficient evidence.	*Medium confidence:* Spatially varying trends. Decrease in many areas, increase in a few areas (Alexander et al., 2006; Haylock et al., 2006b).	*Low confidence:* Overall inconsistent and spatially varying signal (SMA, PDSI, CDD) (Dufek et al., 2008; Sheffield and Wood, 2008a; Dai, 2011).

E. Asia

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
All Asia	*Low confidence to high confidence* depending on region: On continental scale, *medium confidence* in overall increase in WD and decrease in CD (Alexander et al., 2006). See also individual regional entries.	*Low confidence to high confidence* depending on region: On continental scale, *medium confidence* in overall increase in WN and decrease in CN (Alexander et al., 2006). See also individual regional entries.	*Low confidence to medium confidence* depending on region: *Low confidence* due to insufficient evidence in several regions; *medium confidence* in trends in other regions (Alexander et al., 2006). See also individual regional entries.	*Low confidence to medium confidence* depending on region: *Low confidence* due to insufficient evidence or inconsistent trends in several regions; *medium confidence* in trends in HP in a few regions (Alexander et al., 2006). See also individual regional entries.	*Low confidence to medium confidence* depending on region: *Low confidence* in most regions due to spatially varying trends. Some areas have consistent increases, but others display decreases in dryness indicated by different measures (SMA, PDSI, CDD) (Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).
N. Asia (NAS, 18)	*High confidence: Likely* increases in WD, *likely* decreases in CD (Alexander et al., 2006).	*High confidence: Likely* decreases in CN, *likely* increases in WN (Alexander et al., 2006).	*Medium confidence:* Spatially varying trends. Overall WSDI increase, WSDI decrease in a few areas (Alexander et al., 2006).	*Low confidence:* Increase in some regions, but spatial variations (Alexander et al., 2006). Some increase in western Russia, especially in winter (DJF), 1950–2000 (Zolina et al., 2009).	*Low confidence:* Spatially varying trends. Tendency for increased dryness (SMA, PDSI, CDD) in central and northeastern N. Asia, other areas decreased dryness (Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).

Table 3-2 (continued)

E. Asia (Continued)

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
Central Asia (CAS, 20)	*High confidence: Likely* increases in WD, *likely* decreases in CD (Alexander et al., 2006).	*High confidence: Likely* decreases in CN, *likely* increases in WN (Alexander et al., 2006).	*Medium confidence:* WSDI increase in a few areas, insufficient evidence elsewhere (Alexander et al., 2006).	*Low confidence:* Spatially varying trends. Increase in a few areas, decrease in a few areas (Alexander et al., 2006).	*Low confidence:* Spatially varying trends in dryness (SMA, PDSI, CDD); partial lack of coverage in some studies (Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).
E. Asia (EAS, 22)	*High confidence: Likely* increases in WD, *likely* decreases in CD (Alexander et al., 2006; Ding et al., 2010).	*Medium confidence:* Decreases in CN, increases in WN (Alexander et al., 2006).	*Medium confidence:* Increase in warm season heat waves in China (Ding et al., 2010); increase in WSDI in northern China, but decline in southern China (Alexander et al., 2006).	*Low confidence:* Spatially varying trends. Increase in a few areas, decrease in a few areas (Alexander et al., 2006).	*Medium confidence:* Overall tendency for increased dryness (SMA, PDSI, CDD); few areas with opposite trends (Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).
S.E. Asia (SEA, 24)	*Medium confidence:* Increases in WD, decreases in CD in northern part of domain (Alexander et al., 2006). *Low confidence:* Insufficient evidence for Malay Archipelago.	*Medium confidence:* Decreases in CN, increases in WN in northern part of domain (Alexander et al., 2006). *Low confidence:* Insufficient evidence for Malay Archipelago.	*Low confidence:* Insufficient evidence.	*Low confidence:* Spatially varying trends and partial lack of evidence. Some areas increase, some areas decrease (Alexander et al., 2006).	*Low confidence:* Spatially varying trends, inconsistent trends in dryness (SMA, PDSI) between studies (Sheffield and Wood, 2008a; Dai, 2011).
S. Asia (SAS, 23)	*Medium confidence:* Increase in WD and decrease in CD (Alexander et al., 2006).	*Medium confidence:* Decreases in CN, increases in WN (Alexander et al., 2006).	*Low confidence:* Insufficient evidence.	*Low confidence:* Mixed signal in India (Alexander et al., 2006).	*Low confidence:* Inconsistent signal for different studies and indices. Decrease in CDD over India (Alexander et al., 2006). Increased dryness (SMA, PDSI) in central India (Sheffield and Wood, 2008a; Dai, 2011).
W. Asia (WAS, 19)	*High confidence: More likely than not* decrease in CD and *very likely* increase in WD (Rahimzadeh et al., 2009; Rehman, 2010).	*High confidence: Likely* decrease in CN and *likely* increase in WN (Rehman, 2010).	*Medium confidence:* WSDI increase (Alexander et al., 2006).	*Medium confidence:* Decrease in heavy precipitation events (Kwarteng et al., 2009; Rahimzadeh et al., 2009).	*Low confidence:* Lack of studies for part of the region; mixed results (Sheffield and Wood, 2008a; Rahimzadeh et al., 2009).
Tibetan Plateau (TIB, 21)	*High confidence: Likely* increase in WD and *likely* decrease in CD (Alexander et al., 2006).	*High confidence: Likely* decreases in CN, *likely* increases in WN (Alexander et al., 2006).	*Low confidence:* Spatially varying trends (Alexander et al., 2006).	*Low confidence:* Insufficient evidence.	*Low confidence:* Lack of studies. Tendency to decreased dryness (PDSI, SMA) in Dai (2011).

F. Australia/New Zealand

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
All Australia and New Zealand	*High confidence:* Overall *likely* increases in WD, *likely* decreases in CD. See individual regional entries for assessment basis and details.	*High confidence:* Overall *likely* decreases in CN, *likely* increases in WN. See individual regional entries for assessment basis and details.	*Low confidence to medium confidence* depending on region: See individual regional entries for assessment basis and details.	*Low confidence to high confidence* depending on region: Insufficient studies for assessment in N. Australia, *likely* decrease in HP in many areas in S. Australia. See individual regional entries for assessment basis and details.	*Medium confidence:* Some regions with dryness decreases, others with dryness increases. See individual regional entries for assessment basis and details.
N. Australia (NAU, 25)	*High confidence: Likely* increases in WD, *likely* decreases in CD. Weaker trends in northwest (Alexander et al., 2006).	*High confidence: Likely* decreases in CN, *likely* increases in WN (Alexander et al., 2006; Alexander and Arblaster, 2009).	*Medium confidence:* Increase in warm spells across southern Australia (Alexander and Arblaster, 2009).	*Low confidence:* Insufficient studies for assessment.	*Medium confidence:* Decrease in dryness (SMA, PDSI) in northwest since mid-20th century (Sheffield and Wood, 2008a; Dai, 2011).
S. Australia/ New Zealand (SAU, 26)	*High confidence: Very likely* increases in WD, *very likely* decreases in CD (Alexander et al., 2006). NZ positive trends vary across country, related to circulation changes (Chambers and Griffiths, 2008; Mullan et al., 2008).	*High confidence: Very likely* decreases in CN, *very likely* increases in WN (Alexander et al., 2006; Alexander and Arblaster, 2009). General decrease in frosts in NZ but trends vary across country, related to circulation changes (Chambers and Griffiths, 2008; Mullan et al., 2008).	*Medium confidence:* Increase in warm spells across southern Australia (Alexander and Arblaster, 2009).	*High confidence: Likely* decrease in heavy precipitation in many areas, especially where mean precipitation has decreased (CSIRO, 2007; Gallant et al., 2007; Alexander and Arblaster, 2009). NZ trends are positive in western N. and S. Islands and negative in eastern N. and S. Islands and are strongly correlated with changes in mean rainfall (Mullan et al., 2008).	*Medium confidence:* Increase in dryness (SMA, PDSI, CDD) in southeastern part and southwestern tip of Australia since mid-20th century. Decrease in dryness in central part of Australia (Alexander et al., 2006; Sheffield and Wood, 2008a; Dai, 2011).

Table 3-3 | Projected regional changes in temperature and precipitation (including dryness) extremes. See Figure 3-1 for definitions of regions (numbers indicated next to regions' names). For assessments for small island states, refer to Box 3-4. Projections are for the end of the 21st century versus the end of the 20th century (e.g., 1961-1990 or 1980-2000 versus 2071-2100 or 2080-2100) and for the A2/A1B emissions scenarios (except if noted otherwise). Late 20th-century extreme values (generally either 1961-1990 or ~1980-2000) are used as reference (see Box 3-1 for discussion). Codes for the source of modelling evidence: **G**: multiple GCMs; **R**: single RCM forced by single GCM; **R**: multiple RCMs forced by single GCM; **R**: multiple RCMs forced by multiple GCMs. T06 stands for Tebaldi et al. (2006), SW08b stands for Sheffield and Wood (2008b), and OS11 stands for Orlowsky and Seneviratne (2011).

A. North America and Central America

Regions	Tmax [WD = Warm Days, CD = Cold Days, RV20AHD: 20-year return value of annual maximum hottest day] (using late 20th-century extreme values as reference; see Box 3-1)	Tmin [WN = Warm Nights, CN = Cold Nights] (using late 20th-century extreme values as reference; see Box 3-1)	Heat Waves (HW)/Warm Spells (WS) (using late 20th-century extreme values as reference; see Box 3-1)	Heavy Precipitation (HP) [HPD = Heavy Precipitation Days, e.g., precipitation >95th percentile (p) %DP10 = Percentage of Days with Precipitation >10 mm HPC = Heavy Precipitation Contribution, generally fraction from precipitation >95th percentile RV20HP = 20-year return value of annual maximum daily precipitation rates] (using late 20th-century extreme values as reference; see Box 3-1)	Dryness [CDD = Consecutive Dry Days SMA = (Simulated) Soil Moisture Anomalies; see Box 3-3 for definitions]
All North America and Central America	*High confidence:* WD *very likely or likely to* increase and CD *very likely or likely to* decrease in all regions (Christensen et al., 2007; Meehl et al., 2007b; Karl et al., 2008; Fig. 3-3). *Very likely* increase in RV20AHD in all regions except CAM (Fig. 3-5). *Medium confidence:* Largest increases in WD in summer and fall particularly over the United States; largest decrease in CD in Canada in fall and winter (OS11).	*High confidence:* WN *very likely or likely to* increase and CN *very likely or likely to* decrease depending on subregion (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007b; Karl et al., 2008; Fig. 3-4). *Medium confidence:* Largest increase in WN and decrease in CN in summer, particularly in the United States (OS11).	*High confidence* to *medium confidence* depending on subregion. *High confidence: Likely* more frequent, longer, and/or more intense heat waves and warm spells over all North American subregions. *Medium confidence* in increase in warm spells in part of Central America, but lack of agreement in signal of change in heat waves (T06; Christensen et al., 2007; Karl et al., 2008; Clark et al., 2010; OS11).	*Low confidence* to *high confidence* depending on region and index: *Likely* increase in HP, including HPD, HPC and RV20HP, over Canada and Alaska; *Low confidence* to *medium confidence* in the south (particularly CAM) due to smaller and less consistent changes, and inconsistencies between %DP10 (decreases in winter and spring) and other indices (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007b; Karl et al., 2008; OS11; Fig. 3-6).	*Low confidence* to *medium confidence* depending on region. *Medium confidence* regarding increase in CDD, and SMA drought in Texas and N. Mexico (T06; SW08b; Fig. 3-10). *Low confidence:* Inconsistent change in other regions (SMA, CDD) (T06; SW08b; Fig. 3-10).
	G R	G	G	G	G
W. North America (WNA, 3)	*High confidence:* WD *very likely* to increase and CD *very likely* to decrease in all seasons (Christensen et al., 2007; Karl et al., 2008; Clark et al., 2010; Fig. 3-3). *Very likely* increase in RV20AHD (Fig. 3-5). *Medium confidence:* Overall weaker signal in spring and winter for both CD and WD (OS11). RCM simulations for 2030–2039 consistent with projected long-term increase in WD (Diffenbaugh and Ashfaq, 2010).	*High confidence:* WN *very likely* to increase and CN *very likely* to decrease (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007b; Karl et al., 2008; Fig. 3-4). *Medium confidence:* Largest WN increases and CN decreases in summer (OS11).	*High confidence: Likely* more frequent, longer, and/or more intense heat waves and warm spells (T06; Christensen et al., 2007; Meehl et al., 2007b; Karl et al., 2008; Clark et al., 2010; OS11). *Medium confidence:* RCM simulations for 2030–2039 and 2090–2099 consistent with projected long-term increase in frequency and/or intensity of HW (Diffenbaugh and Ashfaq, 2010; Kunkel et al., 2010).	*Low confidence* to *medium confidence* depending on subregion and index: *Medium confidence* in increase in HPD/HPC over northern part of domain (Canada); *low confidence* due to no signal or inconsistent signal in HPD/HPC changes over southern part of domain (T06; Fig. 3-6). *Medium confidence* in increase in RV20HP (Fig. 3-7).	*Low confidence:* Inconsistent signal in CDD and SMA changes (T06; SW08b; Fig. 3-10).
	G R	G	G R	G	G
Central North America (CNA, 4)	*High confidence:* WD *very likely* to increase and CD *very likely* to decrease in all seasons (Christensen et al., 2007; Karl et al., 2008; Clark et al., 2010; Fig. 3-3). *Very likely* increase in RV20AHD (Fig. 3-5). *Medium confidence:* Weaker signal for CD in spring and winter (OS11). RCM simulations for 2030–2039 consistent with projected long-term increase in WD (Diffenbaugh and Ashfaq, 2010).	*High confidence:* WN *very likely* to increase and CN *very likely* to decrease (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007b; Karl et al., 2008; Fig. 3-4).	*High confidence: Likely* more frequent, longer, and/or more intense heat waves and warm spells (T06; Christensen et al., 2007; Karl et al., 2008; Clark et al., 2010; OS11). *Medium confidence:* RCM simulations for 2030–2039 and 2090–2099 consistent with projected long-term increase in frequency and/or intensity of HW (Diffenbaugh and Ashfaq, 2010; Kunkel et al., 2010).	*Low confidence* to *medium confidence* depending on index: *Low confidence* in changes in HPD/HPC due to inconsistent or no signal (T06; Fig. 3-6). *Medium confidence* in increase in RV20HP (Fig. 3-7).	*Medium confidence:* Increase in CDD and decrease in SMA in southern part of the domain (SW08b; Fig. 3-10). *Low confidence:* inconsistent signal elsewhere (Fig. 3-10).
	G R	G	G R	G	G

Table 3-3 (continued)

A. North America and Central America (Continued)

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
E. North America (ENA, 5)	*High confidence:* WD *very likely* to increase and CD *very likely* to decrease in all seasons (Christensen et al., 2007; Karl et al., 2008; Clark et al., 2010; Fig. 3-3). *Very likely* increase in RV20AHD (Fig. 3-5). *Medium confidence:* Largest WD increase in summer and fall; weaker CD decrease in spring (OS11). RCM simulations for 2030–2039 consistent with projected long-term increase in WD (Diffenbaugh and Ashfaq, 2010). **G R**	*High confidence:* WN *very likely* to increase and CN *very likely* to decrease (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007b; Karl et al., 2008; Clark et al., 2010; OS11). *Medium confidence:* Largest WN increases and CN decreases in summer (OS11). **G**	*High confidence: Likely* more frequent, longer, and/or more intense heat waves and warm spells (T06; Christensen et al., 2007; Meehl et al., 2007b; Karl et al., 2008; Clark et al., 2010; OS11). *Medium confidence:* RCM simulations for 2030–2039 and 2090–2099 consistent with projected long-term increase in frequency of HW (Diffenbaugh and Ashfaq, 2010; Kunkel et al., 2010). **G R**	*Medium confidence:* Increase in HPD/HPC in northern part of domain but no signal or inconsistent signal in southern part (T06; Fig. 3-6). *Medium confidence* in increase in RV20HP (Fig. 3-7). **G**	*Low confidence:* Inconsistent signal in CDD, some consistent decrease in SMA (SW08b; Fig. 3-10). **G**
Alaska/ N.W. Canada (ALA, 1)	*High confidence:* WD *very likely* to increase and CD *very likely* to decrease (Christensen et al., 2007; Karl et al., 2008; Fig. 3-3). *Very likely* increase in RV20AHD (Fig. 3-5). *Medium confidence:* Strongest increase in WD in the fall (OS11). **G**	*High confidence:* WN *very likely* to increase and CN *very likely* to decrease (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007b; Karl et al., 2008; Fig. 3-4). **G**	*High confidence: Likely* more frequent and/or longer heat waves and warm spells (T06; Christensen et al., 2007; Meehl et al., 2007b; OS11). **G**	*High confidence: Likely* increase in HPD and HPC (T06; Fig. 3-6). *Likely* increase of RV20HP (Fig. 3-7). **G**	*Low confidence:* Inconsistent signal in change of CDD and SMA (T06; SW08b; Fig. 3-10). **G**
E. Canada, Greenland, Iceland (CGI, 2)	*High confidence:* WD *very likely* to increase and CD *very likely* to decrease (Christensen et al., 2007; Karl et al., 2008; Fig. 3-3). *Very likely* increase in RV20AHD (Fig. 3-5). *Medium confidence:* Strongest increase of WD in fall (in summer in Greenland), weakest in spring, weaker increase of CD in summer (OS11). **G**	*High confidence:* WN *very likely* to increase and CN *very likely* to decrease (T06; Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007b; Fig. 3-4). **G**	*High confidence: Likely* more frequent and/or longer heat waves and warm spells (T06; OS11). **G**	*High confidence: Likely* increase in HPD and HPC (T06; Fig. 3-6). *Likely* increase of RV20HP (Fig. 3-7). **G**	*Low confidence:* Inconsistent signal in CDD and/or SMA changes (T06; SW08b; Fig. 3-10). **G**
Central America and Mexico (CAM, 6)	*High confidence:* WD *likely* to increase and CD *likely* to decrease (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5). **G**	*High confidence:* WN *likely* to increase (T06; Fig. 3-4) and CN *likely* to decrease (Fig. 3-4). **G**	*Medium confidence to high confidence* depending on region: *Likely* more frequent, longer, and/or more intense heat waves and warm spells in most of the region; *medium confidence* in increase in warm spells in part of Central America, but lack of agreement in signal of change in heat waves (T06; OS11; Clark et al., 2010). **G**	*Low confidence:* Lack of agreement between models and indices regarding changes in %DP10, HPC, RV20HP, and other HP indicators (Kamiguchi et al., 2006; T06; Campbell et al., 2011; Figs. 3-6 and 3-7). **G R**	*Low confidence to medium confidence* depending on region. *Medium confidence:* Increased dryness (CDD, SMA) in Central America and Mexico; *low confidence* in change in dryness (CDD, SMA) in the extreme south of region due to inconsistent signal (Kamiguchi et al., 2006; T06; Campbell et al., 2011; Fig. 3-10). **G R**

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Table 3-3 (continued)

B. Europe and Mediterranean Region

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
All Europe and Mediterranean Region	*High confidence:* WD *very likely* to increase – largest increases in summer and Central/S. Europe and smallest in N. Europe (Scandinavia) (Goubanova and Li, 2007; Kjellström et al., 2007; Koffi and Koffi, 2008; Fischer and Schär, 2010; Fig. 3-3) and CD *very likely* to decrease (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5). *Medium confidence:* Changes in higher quantiles of Tmax generally greater than changes in lower quantiles of Tmax in summer in Central Europe and Mediterranean (Diffenbaugh et al., 2007; Kjellström et al., 2007; Fischer and Schär, 2009, 2010; OS11). G R	*High confidence:* CN *very likely* to decrease – largest decreases in winter in E. Europe and Scandinavia (Goubanova and Li, 2007; Kjellström et al., 2007; Sillmann and Roeckner, 2008); WN *very likely* to increase (T06; Fig. 3-4). G R	*High confidence: Likely* more frequent, longer, and/or more intense heat waves and warm spells but little change over Scandinavia (Beniston et al., 2007; Koffi and Koffi, 2008; Clark et al., 2010; OS11). G R	*Low confidence to high confidence,* depending on region: *Likely* overall increases in HPD, %DP10, and RV20HP and decreases in return periods of long (5-day) and short (1-day) events; strong signals in N. Europe particularly in winter, but lower confidence in changes in Central Europe and in particular the Mediterranean (T06; Beniston et al., 2007; Fowler et al., 2007a; Sillmann and Roeckner, 2008; Kendon et al., 2010; Figs. 3-6 and 3-7). *Likely* increase in HPC in some regions (Boberg et al., 2009b; Kendon et al., 2010). *Likely* greater changes in extremes than mean in many regions. Increase in HP intensity (and increase in HPC) despite decrease in summer mean in some regions – e.g., Central Europe (Beniston et al., 2007; Fowler et al., 2007a; Haugen and Iversen, 2008; May, 2008; Kyselý and Beranová, 2009). G R	*Medium confidence:* European area affected by stronger dryness (reduced SMA and CDD) with largest and most consistent changes in Mediterranean Europe (T06; Burke and Brown, 2008; May, 2008; SW08b; Sillmann and Roeckner, 2008; Fig. 3-10). G R
N. Europe (NEU, 11)	*High confidence: Very likely* increase in frequency of WD, but smaller than in Central and S. Europe (Fischer and Schär, 2010; Fig. 3-3). *Very likely* decrease in CD (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5). *Medium confidence:* Changes in lower quantiles of Tmax generally greater than for changes in higher quantiles of Tmax in fall, winter, and spring in Scandinavia and northeastern Europe (OS11). G R	*High confidence:* CN *very likely* to decrease (Kjellström et al., 2007; Sillmann and Roeckner, 2008; Fig. 3-4); WN *very likely* to increase (T06; Fig. 3-4). *Medium confidence:* Changes in lower quantiles of Tmin generally greater than changes in higher quantiles of Tmin in Scandinavia and northeastern Europe (Kjellström et al., 2007; OS11). G R	*High confidence: Likely* more frequent, longer and/or more intense heat waves and warm spells, but summer increases smaller than in S. Europe and little change over Scandinavia (Beniston et al., 2007; Koffi and Koffi, 2008; Fischer and Schär, 2010; OS11). *Medium confidence:* Some dependency of projections of changes in HW intensity on parameterization choice (Clark et al., 2010). G R	*High confidence: Very likely* increases in HP (intensity and frequency) and %DP10 north of 45°N in winter (Frei et al., 2006; T06; Beniston et al., 2007; Kendon et al., 2008; Fig. 3-6). *Likely* increase in RV20HP (Fig. 3-7). G R̲	*Medium confidence:* No major changes in dryness (CDD, SMA) in N. Europe (T06; SW08b; Sillmann and Roeckner, 2008; Fig. 3-10). G
Central Europe (CEU, 12)	*High confidence: Very likely* increase in frequency and intensity of WD (Fischer and Schär, 2010; Fig. 3-3) and decrease in frequency of CD (Fig. 3-3). *Very likely* increase in RV20AHD (Fig. 3-5). *Medium confidence:* Changes in higher quantiles of Tmax much larger than changes in lower quantiles of Tmax in summer; results in very large increase in Tmax variability (Diffenbaugh et al., 2007; Kjellström et al., 2007; Fischer and Schär, 2009, 2010; OS11). G R̲	*High confidence:* CN *very likely* to decrease (Goubanova and Li, 2007; Kjellström et al., 2007; Sillmann and Roeckner, 2008); WN *very likely* to increase (T06; Fig. 3-4). G R̲	*High confidence: Likely* more frequent, longer, and/or more intense heat waves and warm spells (Beniston et al., 2007; Koffi and Koffi, 2008; Clark et al., 2010; Fischer and Schär, 2010; OS11). *Medium confidence:* Some dependency of projections of changes in HW intensity on parameterization choice (Clark et al., 2010). G R̲	*High confidence: Likely* increases in HP (intensity and frequency) in large part of the region in winter (Frei et al., 2006; Beniston et al., 2007; Kendon et al., 2008; Kyselý and Beranová, 2009; Fig. 3-6). *Likely* increase in RV20HP (Fig. 3-7). *Medium confidence:* Inconsistent evidence for summer: increase in HP in summer evident in RCMs (Christensen and Christensen, 2003; Frei et al., 2006) versus no signal in GCMs (Fig. 3-6). G R̲	*Medium confidence:* Increase in dryness (CDD, SMA) in Central Europe (Seneviratne et al., 2006a; T06; Fig. 3-10). *Medium confidence:* Increase in short-term droughts (SW08b). G R

Table 3-3 (continued)

B. Europe and Mediterranean Region (Continued)

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
S. Europe and Mediterranean (MED, 13)	*High confidence: Very likely* increase in frequency and intensity of WD (Fischer and Schär, 2009, 2010; Giannakopoulos et al., 2009; Fig. 3-3) and decrease in CD (Fig. 3-4). *Very likely* increase in RV20AHD (Fig. 3-5). *High confidence:* Number of days with combined hot summer days (>35°C) and tropical nights (>20°C) *very likely* to increase (Fischer and Schär, 2010). *Medium confidence:* Changes in higher quantiles of Tmax greater than changes in lower quantiles of Tmax in summer (Diffenbaugh et al., 2007; Kjellström et al., 2007; Fischer and Schär, 2009, 2010; OS11). **G R**	*High confidence:* CN *very likely* to decrease (Goubanova and Li, 2007; Kjellström et al., 2007; Sillmann and Roeckner, 2008). *High confidence:* WN *very likely* to increase (T06; Sillmann and Roeckner, 2008; Giannakopoulos et al., 2009; Fig. 3-4). *High confidence:* Tropical nights *very likely* to increase (Sillmann and Roeckner, 2008). Number of days with combined hot summer days (>35°C) and tropical nights (>20°C) *very likely* to increase (Fischer and Schär, 2010). *Medium confidence:* Changes in higher quantiles of Tmin generally greater than changes in lower quantiles of Tmin in summer in the Mediterranean (Diffenbaugh et al., 2007; OS11). **G R**	*High confidence: Likely* more frequent and/or longer heat waves and warm spells (also increases in intensity); *likely* largest increases in S.W., S., and E. (Beniston et al., 2007; Diffenbaugh et al., 2007; Koffi and Koffi, 2008; Giannakopoulos et al., 2009; Clark et al., 2010; Fischer and Schär, 2010; OS11). **G R**	*Low confidence: Likely* Inconsistent change in HP intensity and %D10, depends on region and season; increase in HP intensity in all seasons except summer over parts of the region, but decrease in other parts, e.g., Iberian Peninsula (T06; Goubanova and Li, 2007; Giorgi and Lionello, 2008; Giannakopoulos et al., 2009; Fig. 3-6). *Low confidence* in changes in RV20HP (Fig. 3-7). **G R**	*Medium confidence:* Increase in dryness (CDD, SMA) in Mediterranean (T06; Beniston et al., 2007; SW08b; Sillmann and Roeckner, 2008; Giannakopoulos et al., 2009; Fig. 3-10). Consistent increase in area of drought (Burke and Brown, 2008). **G R**

C. Africa

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
All Africa	*High confidence:* WD *likely* to increase and CD *likely* to decrease in all regions (Fig. 3-3). *Likely* increase in RV20AHD in all regions (Fig. 3-5). *Medium confidence:* Increase in WD largest in summer and fall (OS11). **G**	*High confidence:* WN *likely* to increase (T06; Kharin et al., 2007; Fig. 3-4) and CN *likely* to decrease (Fig. 3-4). **G**	*High confidence: Likely* more frequent and/or longer heat waves and warm spells (T06; OS11). **G**	*Low confidence to high confidence depending on region:* Inconsistent change or no signal in HP indicators across much of continent (T06; Fig. 3-6, Fig. 3-7). Strongest and most consistent signal is *likely* increase in HP in E. Africa (T06; Figs. 3-6 and 3-7; Shongwe et al., 2011). **G**	*Low confidence to medium confidence depending on region: Low confidence* in most regions, *medium confidence* of increase in dryness (CDD, SMA) in southern Africa except eastern part (T06; SW08b; Fig. 3-10). **G**
W. Africa (WAF, 15)	*High confidence:* WD *likely* to increase and CD *likely* to decrease (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5). **G**	*High confidence:* WN *likely* to increase (T06; Fig. 3-4) and CN *likely* to decrease (Fig. 3-4). **G**	*High confidence: Likely* more frequent and/or longer heat waves and warm spells (T06; OS11). **G**	*Low confidence to medium confidence depending on subregion: Medium confidence* in slight or no change in HP indicators in most of region; *Low confidence* due to low model agreement in northern part of region (T06; Figs. 3-6 and 3-7). **G**	*Low confidence:* Inconsistent signal of change in CDD and SMA (T06; Fig. 3-10). **G**
E. Africa (EAF, 16)	*High confidence:* WD *likely* to increase and CD *likely* to decrease (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5). **G**	*High confidence:* WN *likely* to increase (T06; Fig. 3-4) and CN *likely* to decrease (Fig. 3-4). **G**	*High confidence: Likely* more frequent and/or longer heat waves and warm spells (T06; OS11). **G**	*High confidence: Likely* increase in HP indicators (T06; Fig. 3-6; Fig. 3-7; Shongwe et al., 2011). **G**	*Medium confidence:* Decreasing dryness in large part of region, especially based on change in SMA, and partly also in CDD (T06; Fig. 3-10; Shongwe et al., 2011). **G**
S. Africa (SAF, 17)	*High confidence:* WD *likely* to increase and CD *likely* to decrease (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5). **G**	*High confidence:* WN *likely* to increase (T06; Fig. 3-4) and CN *likely* to decrease (Fig. 3-4). **G**	*High confidence: Likely* more frequent and/or longer heat waves and warm spells (T06; OS11). **G**	*Low confidence:* Lack of agreement/signal in %DP10 and other HP indicators for the region as a whole (T06; Fig. 3-6). Some model agreement in increase in RV20HP (Fig. 3-7). Some evidence of increased HP intensity in the S.E. (Hewitson and Crane, 2006; Rocha et al., 2008; Shongwe et al., 2009). **G**	*Medium confidence:* Increase in dryness (CDD, SMA), except eastern part (T06; Shongwe et al., 2009; Fig. 3-10). Consistent increase in area of drought (Burke and Brown, 2008). **G**
Sahara (SAH, 14)	*High confidence:* WD *likely* to increase and CD *likely* to decrease (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5). **G**	*High confidence:* WN *likely* to increase (T06; Fig. 3-4) and CN *likely* to decrease (Fig. 3-4). **G**	*High confidence: Likely* more frequent and/or longer heat waves and warm spells (T06; OS11). **G**	*Low confidence:* Low agreement/no signal in %DP10, RV20HP, and other HP indicators (T06; Figs. 3-6 and 3-7). **G**	*Low confidence:* Inconsistent signal of change in CDD and SMA (T06; Fig. 3-10). **G**

Table 3-3 (continued)

D. South America

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
All South America	*High confidence:* WD *likely* to increase and CD *likely* to decrease in all regions (Fig. 3-3). *Likely* increase in RV20AHD in all regions (Fig. 3-5).	*High confidence:* WN *likely* to increase (T06; Kharin et al., 2007; Marengo et al., 2009a; Fig. 3-4) and CN *likely* to decrease (Marengo et al., 2009a; Fig. 3-4).	*Medium confidence to high confidence* depending on region: *Likely* more frequent and/or longer heat waves and warm spells on annual time scale; only *medium confidence* in increase in HW duration in southeastern South America (T06; OS11).	*Low confidence to medium confidence* depending on region: Inconsistent sign of change in HP indicators in some regions; some regions with model agreement (T06; Figs. 3-6 and 3-7).	*Low confidence to medium confidence* depending on region: Inconsistent signal except for dryness increase (CDD and SMA) in northeastern Brazil (SW08b; Fig. 3-10).
	G	G R	G	G	G
Amazon (AMZ, 7)	*High confidence:* WD *likely* to increase and CD *likely* to decrease (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5).	*High confidence: Very likely* increase in WN (T06; Marengo et al., 2009a; Fig. 3-4) and CN *likely* to decrease (Fig. 3-4).	*High confidence: Likely* more frequent and/or longer heat waves and warm spells based on multi-model assessments (T06, OS11); non-significant signal in single-model and single-index study by Uchiyama et al. (2006).	*Medium confidence:* Tendency to increase in precip >95th p and in RV20HP but less consistent increase for some other HP indicators and single studies (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Sorensson et al., 2010; Figs. 3-6 and 3-7).	*Low confidence:* Inconsistent signal in SMA (SW08b; Fig. 3-10) and inconsistent signal in CDD (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Sorensson et al., 2010; Fig. 3-10).
	G	G R	G	G R	G R
N.E. Brazil (NEB, 8)	*High confidence:* WD *likely* to increase and CD *likely* to decrease (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5).	*High confidence: Likely* increase of WN (T06; Marengo et al., 2009a; Fig. 3-4) and CN *likely* to decrease (Fig. 3-4).	*High confidence: Likely* more frequent and/or longer heat waves and warm spells (T06; OS11).	*Low confidence:* Slight or no change in %DP10 and other HP indicators (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Sorensson et al., 2010; Figs. 3-6 and 3-7).	*Medium confidence:* Dryness increase (CDD, SMA) (Kamiguchi et al., 2006; T06; SW08b; Marengo et al., 2009a; Sorensson et al., 2010; Fig. 3-10).
	G	G R	G	G R	G R
S.E. South America (SSA, 10)	*High confidence:* WD *likely* to increase and CD *likely* to decrease (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5).	*High confidence: Very likely* increase in WN (T06; Marengo et al., 2009a; Fig. 3-4) and CN *likely* to decrease (Fig. 3-4).	*Medium confidence:* Tendency for more frequent and/or longer heat waves and warm spells on annual time scale; weak signal compared to other regions but robust sign (increase) across majority of models (T06; OS11).	*Medium confidence:* Increase of %DP10, precipitation intensity, precip >95th p, and RV20HP in the northern portion. *Low confidence* in the southern portion: inconsistent change or no signal in HP indicators (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Nunez et al., 2009; Figs. 3-6 and 3-7).	*Low confidence:* Inconsistent signal in SMA (SW08b). *Low confidence:* Inconsistent signal in CDD (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Sorensson et al., 2010; Fig. 3-10).
	G	G R	G	G R	G R
W. Coast South America (WSA, 9)	*High confidence:* WD *likely* to increase and CD *likely* to decrease (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5).	*High confidence: Likely* increase in WN (T06; Marengo et al., 2009a; Fig. 3-4) and CN *likely* to decrease (Fig. 3-4).	*High confidence: Likely* more frequent and/or longer heat waves and warm spells on annual time scale (T06; OS11).	*Medium confidence:* Increase in %DP10, precip >95th p and other HP indicators in the tropics. *Low confidence* for the extratropics (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Sorensson et al., 2010; Fig. 3-6). *Low confidence* in changes in RV20HP (Fig. 3-7).	*Low confidence:* Inconsistent changes in SMA across domain (Fig. 3-10), CDD decrease in the tropics and increase in the extratropics (Kamiguchi et al., 2006; T06; Marengo et al., 2009a; Sorensson et al., 2010; Fig. 3-10). *Medium confidence:* CDD increase and SMA decrease in southwest SA (SW08b; Fig. 3-10).
	G	G R	G	G R	G R

Table 3-3 (continued)

E. Asia

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
All Asia	High confidence: WD likely to increase and CD likely to decrease in all regions (Fig. 3-3). Likely increase in RV20AHD in all regions (Fig. 3-5).	High confidence: WN likely to increase (T06; Kharin et al., 2007; Fig. 3-4) and CN likely to decrease (Fig. 3-4).	Low confidence to high confidence depending on region: Likely more frequent and/or longer heat waves and warm spells in most regions (continental) on the annual time scale; low confidence due to inconsistent signal in Indonesia, Philippines, Malaysia, Papua New Guinea, and neighboring islands for some HW indices (T06; OS11).	Low confidence to high confidence depending on region and index: High confidence regarding likely increase in HP in N. Asia; medium confidence regarding increase in HP in S.E. Asia, E. Asia, and Tibetan Plateau; low confidence regarding increase in HP in S. and W. Asia (T06; Figs. 3-6 and 3-7).	Low confidence: Inconsistent change in CDD and SMA between models in large part of domain (T06; Fig. 3-10).
	G	G	G	G	G
N. Asia (NAS, 18)	High confidence: WD likely to increase and CD likely to decrease (Fig. 3-3). Likely increase in RV20AHD (Fig. 3-5).	High confidence: WN likely to increase (T06; Fig. 3-4) and CN likely to decrease (Fig. 3-4).	High confidence: Likely more frequent and/or longer heat waves and warm spells (T06; OS11).	High confidence: Likely increase in HP (T06; Figs. 3-6 and 3-7) including more frequent and intense HPD over most regions (Emori and Brown, 2005; Kamiguchi et al., 2006).	Low confidence: Inconsistent change in dryness (CDD, SMA) between models in large part of domain (T06; SW08b; Fig. 3-10).
	G	G	G	G	G
Central Asia (CAS, 20)	High confidence: WD likely to increase and CD likely to decrease (Fig. 3-3). Likely increase in RV20AHD (Fig. 3-5).	High confidence: WN likely to increase (T06; Fig. 3-4) and CN likely to decrease (Fig. 3-4).	High confidence: Likely more frequent, longer and/or more intense heat waves and warm spells (T06; Clark et al., 2010; OS11).	Low confidence: Inconsistent signal in models regarding changes in HP (T06; Fig. 3-6). Some model consistency in projections of RV20HP (Fig. 3-7).	Low confidence: Inconsistent signals across indices (CDD, SMA) (T06; SW08b; Fig. 3-10).
	G	G	G	G	G
E. Asia (EAS, 22)	High confidence: WD likely to increase and CD likely to decrease across the region (Clark et al., 2010; Fig. 3-3), including in Korea (Boo et al., 2006; Im et al., 2008, 2011; Koo et al., 2009). Likely increase in RV20AHD (Fig. 3-5).	High confidence: WN likely to increase (T06; Fig. 3-4) and CN likely to decrease (Fig. 3-4), including in Korea (Boo et al., 2006; Koo et al., 2009; Im et al., 2011).	High confidence: Likely more frequent, longer, and/or more intense heat waves and warm spells (T06; Clark et al., 2010; OS11).	Medium confidence: Increases in HP (less consistent in %DP10 than other indicators) across the region (T06; Figs. 3-6 and 3-7), including increase in Japan and Korea (Emori and Brown, 2005; Kimoto et al., 2005; Boo et al., 2006; Kamiguchi et al., 2006; Kusunoki and Mizuta, 2008; Kitoh et al., 2009; Su et al., 2009; Kim et al., 2010; Im et al., 2011).	Low confidence: Inconsistent signal across indices (CDD, SMA) (T06; SW08b; Fig. 3-10).
	G R	G	G	G R	G
S.E. Asia (SEA, 24)	High confidence: WD likely to increase and CD likely to decrease (Fig. 3-3). Likely increase in RV20AHD (Fig. 3-5).	High confidence: WN likely to increase (T06; Fig. 3-4) and CN likely to decrease (Fig. 3-4).	Low confidence to high confidence depending on subregion and index: Likely more frequent and/or longer heat waves and warm spells on the annual time scale over continental areas; Low confidence in changes in some HW indices but model agreement of increases in a WS index in Indonesia, Philippines, Malaysia, Papua New Guinea, and neighboring islands (T06; OS11).	Medium confidence: Inconsistent signal in change of %DP10 across models (T06; Fig. 3-6) but more frequent and intense HPD, including increase in RV20HP, suggested by other indicators over most regions especially non-continental parts (Emori and Brown, 2005; Kamiguchi et al., 2006; Figs. 3-6 and 3-7).	Low confidence: Inconsistent signal of change in CDD and/or SMA (T06; SW08b; Fig. 3-10).
	G	G	G	G R	G
S. Asia (SAS, 23)	High confidence: WD likely to increase and CD likely to decrease (Kumar et al., 2006; Rajendran and Kitoh, 2008; Fig. 3-3). Likely increase in RV20AHD (Fig. 3-5).	High confidence: WN likely to increase (T06; Fig. 3-4) and CN likely to decrease (Fig. 3-4). Medium confidence: Extreme nighttime temperature warms faster than daytime (Kumar et al., 2006).	High confidence: Likely more frequent and/or longer heat waves and warm spells on annual time scale (T06; OS11). Medium confidence: Some dependency of magnitude of signal on index choice (OS11).	Low confidence: Slight or no increase in %DP10 (T06; Fig. 3-6). Some model consistency regarding increase in RV20HP (Fig. 3-7). Low confidence: More frequent and intense HPD over parts of S. Asia (Emori and Brown, 2005; Kamiguchi et al., 2006; Kharin et al., 2007; Rajendran and Kitoh, 2008).	Low confidence: Inconsistent signal of change in CDD and SMA (T06; SW08b; Fig. 3-10).
	G R	G R	G	G R	G

Table 3-3 (continued)

E. Asia (Continued)

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
W. Asia (WAS, 19)	*High confidence*: WD *likely* to increase and CD *likely* to decrease (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5).	*High confidence*: WN *likely* to increase (T06; Fig. 3-4) and CN *likely* to decrease (Fig. 3-4).	*High confidence*: *Likely* more frequent, longer and/or more intense heat waves and warm spells (T06; Clark et al., 2010; OS11). *Medium confidence*: Some dependency of magnitude of signal on index choice (OS11).	*Low confidence*: Inconsistent signal of change in HP (T06; Fig. 3-6; Fig. 3-7).	*Low confidence*: Inconsistent signal of change in CDD and SMA (T06; SW08b; Fig. 3-10).
	G	G	G	G	G
Tibetan Plateau (TIB, 21)	*High confidence*: WD *likely* to increase and CD *likely* to decrease (Fig. 3-3). *Likely* increase in RV20AHD (Fig. 3-5).	*High confidence*: WN *likely* to increase (T06; Fig. 3-4) and CN *likely* to decrease (Fig. 3-4).	*High confidence*: *Likely* more frequent, longer, and/or more intense heat waves and warm spells (T06; Clark et al., 2010; OS11).	*Medium confidence*: Increase in HP (T06; Figs. 3-6 and 3-7).	*Low confidence*: Inconsistent signal of change in CDD (T06; SW08b; Fig. 3-10).
	G	G	G	G	G

F. Australia/New Zealand

Regions	Tmax	Tmin	Heat Waves / Warm Spells	Heavy Precipitation	Dryness
All Australia and New Zealand	*High confidence*: WD *very likely* to increase and CD *very likely* to decrease in all regions (CSIRO, 2007; Mullan et al., 2008; Fig. 3-3). *Very likely* increase in RV20AHD (Fig. 3-5).	*High confidence*: WN *very likely* to increase everywhere (T06; Kharin et al., 2007; Alexander and Arblaster, 2009; Fig. 3-4). CN *very likely* to decrease (Fig. 3-4). *Medium confidence*: WN increase everywhere. Largest increases in WN in N. compared with S. and most consistent changes in inland regions (Alexander and Arblaster, 2009).	*High confidence*: *Likely* more frequent and/or longer heat waves and warm spells (T06; OS11; Alexander and Arblaster, 2009). *Medium confidence*: Strongest increases in HW duration in N.W. and most consistent increases inland (Alexander and Arblaster, 2009).	*Low confidence*: Lack of agreement regarding sign of change for different models and different indices, and spatial variations in signal (T06; Figs. 3-6 and 3-7). *Low confidence*: HPD tend to increase in E. and decrease in W. half of country – but considerable inter-model inconsistencies; HPC tends to increase everywhere – but considerable inter-model inconsistencies (Alexander and Arblaster, 2009).	*Low confidence to medium confidence* depending on region: Models agree on increase in CDD in S. Australia, but inconsistent signal over most of S. Australia in SMA; inconsistent signal in CDD and SMA in N. Australia (T06; SW08b; Fig. 3-10). Strongest CDD increases in W. half of Australia (Alexander and Arblaster, 2009). Inconsistent change in area of drought depending on index used (Burke and Brown, 2008).
	G	G	G	G	G
N. Australia (NAU, 25)	*High confidence*: WD *very likely* to increase and CD *very likely* to decrease (CSIRO, 2007; Fig. 3-3). *Very likely* increase in RV20AHD (Fig. 3-5).	*High confidence*: WN *very likely* to increase (T06; Alexander and Arblaster, 2009; Fig. 3-4) and CN *very likely* to decrease (Fig. 3-4). *Medium confidence*: Changes larger than in S. Australia (Alexander and Arblaster, 2009).	*High confidence*: *Likely* more frequent and/or longer heat waves and warm spells (T06; Alexander and Arblaster, 2009). *Medium confidence*: Strongest increases in N.W. and most consistent increases inland (Alexander and Arblaster, 2009).	*Low confidence*: Lack of agreement regarding sign of change for different models and different indices (T06; Fig. 3-6, Fig. 3-7).	*Low confidence*: Inconsistent signal in CDD and SMA (T06; SW08b; Fig. 3-10).
	G	G	G	G	G
S. Australia/ New Zealand (SAU, 26)	*High confidence*: WD *very likely* to increase and CD *very likely* to decrease (CSIRO, 2007; Fig. 3-3). *Very likely* increase in RV20AHD (Fig. 3-5). *Low confidence to medium confidence*: Strongest New Zealand increases in WD in North Island and largest decreases in frost days in South Island (Mullan et al., 2008).	*High confidence*: WN *very likely* to increase (T06; Alexander and Arblaster, 2009; Fig. 3-4) and CN *very likely* to decrease (Fig. 3-4). *Medium confidence*: Changes smaller than in N. Australia (Alexander and Arblaster, 2009).	*High confidence*: *Likely* more frequent and/or longer heat waves and warm spells (T06; Alexander and Arblaster, 2009; OS11). *Medium confidence*: Most consistent increases inland (Alexander and Arblaster, 2009).	*Low confidence*: Lack of agreement regarding sign of change for different models and different indices, and spatial variations in signal (T06; Fig. 3-6, Fig. 3-7). *Low confidence to medium confidence*: In New Zealand, increase in HP events at most locations (Mullan et al., 2008; Carey-Smith et al., 2010).	*Medium confidence*: Models agree on increase in CDD in southern Australia including S.W. (T06; Alexander and Arblaster, 2009; Fig. 3-10), but inconsistent signal in SMA over most of the region, slight decrease in S.W. (SW08b; Fig. 3-10).
	G R	G	G	G R	G

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A digital library of non-journal-based literature cited in this chapter that may not be readily available to the public has been compiled as part of the IPCC review and drafting process, and can be accessed via either the IPCC Secretariat or IPCC Working Group II web sites.

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