Chapter 3: Changes in Climate Extremes and their Impacts on the Natural Physical Environment

Coordinating Lead Authors:
Neville Nicholls (Australia), Sonia Seneviratne (Switzerland)

Lead Authors:
David Easterling (USA), Amadou Gaye (Senegal), Clare Goodess (United Kingdom), Shinjiro Kanae (Japan), James Kossin (USA), Yali Luo (China), Jose Marengo (Brazil), Kathleen McInnes (Australia), Mohamed Rahimi (Iran), Markus Reichstein (Germany), Asgeir Sorteberg (Norway), Carolina Vera (Argentina), Xuebin Zhang (Canada)

Contributing Authors:
Gerado Benito (Spain), Tereza Cavazos (Mexico), Paul Della-Marta (Switzerland), Sunling Gong (Canada), B. N. Goswami (India), Mark Hemer (Australia), Bart van den Hurk (Netherlands), Christian Huggel (Switzerland), Akio Kitoh (Japan), Simon Mason (USA), Wassila Thiaw (USA), Pascal Yiou (France), Tingjun Zhang (USA), Tianjun Zhou (China), Francis Zwiers (Canada)

Table of Contents

Executive Summary

The focus of this Chapter is on possible changes in the frequency and intensity of weather and climate extremes that may contribute to disasters. A changing climate may lead to changes in the frequency of occurrence of an extreme (rare event), or result in an unprecedented, previously unobserved, extreme. As well, a weather or climate event, although not drawn from the extreme tail of the distribution, still may be associated with disasters, possibly by leading to a crossing of a critical threshold in a social, ecological or physical system, or because it occurs simultaneously with another type of event combined with which it leads to extreme conditions (compound event). Some weather/climate events may increase the potential for disasters through their impacts on physical systems, such as floods and landslides after heavy rain. Phenomena such as El Niño are considered here because they can lead to climate extremes such as droughts and heavy rains in many regions simultaneously and this may be relevant to disaster management. As well, changes in phenomena such as El Niño or monsoons would also likely affect the frequency and intensity of extremes in several regions.

The events and phenomena examined in this Chapter are:

- Weather and climate elements (temperature; precipitation; winds)
- Weather and climate phenomena (monsoons; El Niño – Southern Oscillation and other modes of variability; tropical and extratropical cyclones)
- Impacts on the natural physical environment (droughts; floods; extreme sea level and coastal impacts; cryosphere and permafrost-related impacts; landslides; sand and dust storms)

For each of these events/phenomena/impacts, the evidence of whether or not they appear to be changing in frequency or intensity (and why) is summarised, as well as projections of future changes (and the confidence in these projections). The assessments herein are based on assessments in the IPCC AR4 modified, where appropriate, by post-AR4 research.

Research performed since the AR4 has reinforced the conclusion that for the period since 1950 it is *very likely* that there has been a decrease in the number of unusually cold days and nights, and an increase in the number of unusually warm days and nights on both a global and regional basis (where the respective extremes are defined with regard to the 1960-1990 base period). Furthermore, based on a limited number of regional analyses and implicit from the documented mean changes in daily temperatures, it is *likely* that warm spells, including heat waves, have increased since the middle of the 20th century. The few studies since the AR4 of annual maximum daily maximum and minimum temperatures suggest that human emission of greenhouse gases has *likely* had a detectable influence on extreme temperatures at the global and regional scales. Post-AR4 studies of temperature extremes have utilised larger model ensembles and generally reinforce the projections of changes in temperature extremes reached in AR4 as well as providing more regional detail (i.e., *virtually certain* warming trends in daily temperature extremes and *very likely* increases in heat waves over most land areas, the temperature extremes being defined with respect to the 1960-1990 base period). In some regions, the enhanced occurrence of hot extremes is projected to have particularly large impacts because they are associated with critical health thresholds.

Many studies conducted since the AR4 support its conclusion that increasing trends in precipitation extremes have *likely* occurred in many areas over the world. Overall, new studies since AR4 have substantially strengthened the AR4 assessment that it is *more likely than not* that anthropogenic influence has contributed to a global trend towards increases in the frequency of heavy precipitation events over the second half of the 20th century. The AR4 projected that it is *very likely* that the frequency of heavy precipitation (or proportion of total rainfall from heavy falls), will increase over most areas of the globe in the 21st century. In some regions, heavy daily precipitation events are projected to increase even if the annual total precipitation is projected to decrease. Post-AR4 analyses of climate model simulations generally confirm the AR4 assessment.

There is almost no literature on the attribution of the causes of any observed changes in strong winds, and thus no assessment can be provided at this time, as was the case in the AR4. Nonetheless, there have been several studies since the AR4 that have focussed on future changes in strong winds and the findings from these point to a decreased frequency of the strongest wind events in the tropics and increased frequency in the strongest wind events in the extratropics, although regional variations occur. However, the small number of studies of projected extreme winds, together with shortcomings in the simulation of these events, means that it is still difficult to credibly project changes in strong winds. Further complicating the projection of changes in tropical wind extremes is the projection of a *likely* increase in tropical cyclone winds.

The AR4 concluded that the current understanding of climate change in the monsoon regions remains one of considerable uncertainty with respect to circulation and precipitation. The AR4 projected that there “is a tendency for monsoonal circulations to result in increased precipitation mainly in the form of extremes due to enhanced moisture convergence, despite a tendency towards weakening of the monsoonal flows themselves. However, many aspects of tropical climatic responses remain uncertain.” Post-AR4 work has not substantially changed these conclusions. At regional scales, there is little consensus in climate models regarding the sign of future change in the monsoons. Land

use changes and aerosols from biomass burning have emerged as important forcings on the variability of monsoons, but are associated with large uncertainties.

Studies since the AR4 provide evidence of a tendency for recent El Niño episodes to be centred more in the central equatorial Pacific than in the east Pacific. In turn, this change in the location of the strongest sea surface temperature anomalies associated with El Niño may explain changes that have been noted in the remote (i.e, away from the equatorial Pacific) climate influences of the phenomenon. Apart from this, there is little evidence of trends in the temporal/seasonal nature of the El Niño–Southern Oscillation in recent decades. The possible role of increased greenhouse gases in affecting the behaviour of the El Niño – Southern Oscillation over the past 50-100 years is uncertain. 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. However, most models project a further increase in the relative frequency of central equatorial Pacific events.

Regarding other modes of climate variability, the AR4 noted that trends observed over recent decades in the North Atlantic Oscillation (NAO) and the Southern Annular Mode (SAM) were *likely* due in part to human activity. Recent studies also suggest that variability in the NAO is being affected by rising global temperatures and that projected warming may lead to a more positive NAO regime (although confidence in the ability of models to simulate the NAO is low). An increasing positive phase of the SAM in recent decades has been linked to stratospheric ozone depletion and to greenhouse gas increases. Models including both greenhouse gas and stratospheric ozone changes simulate a realistic trend in the SAM, although there is some concern that possible anthropogenic circulation changes are poorly characterized by trends in the annular modes. There is little consistency between model projections of these modes.

There have been no significant trends observed in the global annual number of tropical cyclones, including over the recent 40-year period of satellite observations. Regional trends in tropical cyclone frequency have been identified in the North Atlantic, but there is a lack of consensus regarding the fidelity of these trends. The uncertainties in the historical tropical cyclone records and the degree of tropical cyclone variability — comprising random processes and linkages to various natural climate modes such as El Niño — do not presently allow for the attribution of any observed changes in tropical cyclone activity to anthropogenic influences. It is *likely* that the global frequency of tropical cyclones will either decrease or remain essentially unchanged in future decades. An increase in mean tropical cyclone maximum wind speed is *likely*, although increases may not occur in all ocean basins. It is *more likely than not* that the increases in frequency of the most intense storms will vary substantially between ocean basins. It is *likely* that tropical cyclone-related rainfall rates will increase with greenhouse warming.

Research subsequent to the AR4 supports previous findings of a poleward shift in extratropical cyclones since the 1950s and an intensification of extratropical cyclones in high latitudes in the last 50 years. New evidence has strengthened the AR4 assessment that it is *likely* that anthropogenic forcing has contributed to the changes in extratropical storm tracks but a quantitative anthropogenic influence has not been detected formally, owing to large internal variability and problems due to changes in observing systems. It is *likely* that future anthropogenic climate change will influence regional cyclone activity. A reduction in mid-latitude storms averaged over each hemisphere is *likely* and it is *more likely than not* that high-latitude cyclone number and intensity will increase. There is little consistency among models regarding the detailed geographical pattern of projected cyclone activity changes.

The AR4 concluded from proxies based on precipitation data and estimates using the Palmer-drought severity index that it is *likely* that the intensity and duration of droughts have increased since the 1950s and that the area of drought-affected regions has increased since the 1970s. Anthropogenic influence on diagnosed drought trends was evaluated as *more likely than not* in the AR4. Research on regional drought since the AR4 further supports the above AR4 assessment. Lack of soil moisture observations partly prevents the analysis of trends in agricultural droughts in most regions, an issue also noted in the AR4. Post-AR4 studies have projected an increase in the global area affected by extreme drought over the 21st century being *likely*. However, the changes are dependent on the definition of the drought index, and on the region examined.

Research since the AR4 has not shown clear and widespread evidence of observed changes in floods at the global level except for the earlier spring flow in snow-dominated regions. After a period of frequent occurrence at the end of the Little Ice Age and a more stable period during the 20th century, glacial-lake outburst floods have increased in frequency in many regions. It is *more likely than not* that anthropogenic greenhouse gas emissions have affected floods because they have influenced components of the hydrological cycle, but the magnitude and even the sign of this anthropogenic influence is uncertain. The causes of regional changes in floods are complex. It is *likely* that anthropogenic influence has resulted in earlier spring flood peaks in snowmelt rivers. A few recent studies for Europe and one global study have projected changes in the frequency and/or magnitude of floods in the 21st century at a large scale. However, the sign of any projected trend varies regionally.

The AR4 reported that the rise in mean sea level and variations in regional climate led to a *likely* upward trend in extreme high water worldwide in the late 20th century. Subsequent to the AR4 a small number of additional studies of extreme sea levels have been undertaken, which support the AR4 conclusion, although some regional studies also note the relationship between extreme sea levels and modes of natural variability. It is *very likely* that mean sea level rise will contribute to upward trends in extreme sea levels in the future. The AR4 reported statistically significant positive trends in significant wave height in some parts of the globe for which data was available including most of the mid-latitudinal North Atlantic and North Pacific. Additional studies since the AR4 provide further evidence for positive trends in these and other locations. However, the small number of studies, and the different sources of wave data used in the studies, preclude a formal assessment at this time. Future changes to significant wave height are *likely* to reflect future changes in storminess and associated patterns of wind change. The AR4 concluded that hazards such as increased coastal inundation, erosion and ecosystem losses are adversely impacting coasts. New studies since the AR4 draw similar conclusions and also note the difficulty in apportioning the observed changes between natural climate variability, climate change and other anthropogenic causes.

Frequency of large landslides in cold regions and high mountains has *more likely than not* increased during the past two decades, and especially early into the 21st century. Earlier snow melt is *more likely than not* to result in earlier onset of high-mountain debris flows, and shallow landslides in lower mountain ranges are *more likely than not* to increase with the projected higher precipitation intensities. It is unclear if anthropogenic influence has contributed to any changes in temperate and tropical region landslides. New and potentially unstable lakes are *likely* to form during the 21st century following glacier retreat. Permafrost is *likely* thawing and has *likely* resulted in physical impacts in cold regions such as increased Arctic coastal erosion and development of thermokarst terrains and thaw lakes. The changes in permafrost and its associated physical impacts are *likely* due to anthropogenic influences because these changes are primarily caused by increase in air temperature and winter snow thickness. It is *likely* that permafrost will continue to thaw with an increase in its associated physical impacts. Due to projected sea ice retreat, and permafrost degradation, the frequency and magnitude of the rate of Arctic coastal erosion is *likely* to increase.

Over the past few decades, the frequency of dust events has increased in some regions such as the Sahel zone of Africa and decreased in some other regions such as northern China. There is high uncertainty in projected future changes in dust activity. Due to scarce evidence, assessments of the likelihood of past and projected changes in dust events, and the attribution of observed changes, cannot be provided at present.

In many cases changes in extremes closely follow changes in the average of a weather variable. However there are sufficient exceptions from this that one cannot assume that a change in an extreme will necessarily follow a change in the mean of the variable. This appears to be especially the case for short-duration heavy precipitation episodes, and temperature extremes at urban locations or in mid and high latitudes. For example, extreme precipitation is projected to increase even in some regions where total precipitation is projected to decrease.

This overall assessment highlights that our confidence in past and future changes including the direction and magnitude in 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. The different levels of confidence need to be taken into consideration in management strategies for disaster risk reduction involving climate and weather extremes.

3.1. Weather and Climate Events Related to Disasters

3.1.1. What Scientific Information is Needed to Inform Risk Management?

Extreme weather and climate events are important, albeit rare, aspects of the climate. The probability that a defined extreme event will occur at a given time and place is closely related to the statistical properties of climate at this location. This concept has found wide applications in engineering practice for many years: design values have been estimated from observed climate, and the likelihood for exceeding the design values has been assessed by assuming that climate within the expected life span of the engineering structure will remain the same as that from which the design values were derived. This presupposes, however, that climate is stationary (Milly et al., 2008). In a situation of transient climate changes, other approaches need to be developed to inform risk management.

When climate properties change at a location, the probability distribution functions (PDFs) for climate variables are modified, with attendant changes in the frequency and intensity of extreme events (as defined with respect to a past climatology). Different extremes and their related impacts may behave differently for a given change in mean climate, i.e., some may become more frequent, other less frequent. For example, in most locations, a globally warmer climate will probably make extreme high temperatures even warmer and thus more extreme, but it will also probably result in less extreme low temperatures. However, decreases in the occurrence or intensity of some extremes may not "compensate" for increases in other extremes, as a shift in the average climate means a change away from the range of climate to which natural and human systems have adapted, and in such circumstances it may thus be felt that the climate is becoming more extreme (FAQ 3.1). Moreover, changes in mean climate may also lead to the sudden occurrence of extremes that were not previously experienced at a given location, due to the crossing of critical thresholds, for instance in the case of heatwave-induced mortality, or the occurrence of droughts, floods, or storm surges (Box 3.1). Besides these changes in extremes associated with modifications of the mean climate, the statistical properties of the climate in some regions may also be modified in such a way that variability is changed, i.e., rare events become more (or less) distinct from the mean climate (Box 3.2). Finally, also contrasting extremes (both wet and dry extremes) or compound events may become more frequent in some regions (Section 3.1.4 and Box 3.4), and thus changes in given extremes or in the mean of some variables cannot be considered in isolation when assessing their resulting impacts on ecosystems and society.

Extreme weather and climate events occur on a wide range of space and time scales. A tornado may last for only a few minutes and cause damage only to a localized area. On the other hand, a drought may persist for years or even decades and may impact a region as large as a continent. In general, an extreme that occurs on a small time scale also tends to have a small space scale. The scale of extremes determines the data requirements for their analysis (e.g., hourly/daily versus monthly resolution, Section 3.2.1). It is also relevant to their understanding, as small-scale changes in a variable are often partially controlled by changes in other factors (topography, land-atmosphere exchanges) in addition to those induced by large-scale changes (large-scale circulation patterns, global temperature change). This consideration is further addressed in Section 3.2.2.

Preparedness for possible future changes in physical extremes requires several types of scientific information, including:

- Identification and definition of events that are relevant from a risk management perspective
- Observations and model experiments to analyse past and projected changes in identified extremes, and to identify the underlying mechanisms and causes
- Assessments of confidence in the likelihood of past and projected changes in extremes
- Prediction tools for early warning and forecasting of extremes, to allow adaptation to projected changes in identified extremes.

These various aspects are briefly addressed in the following subsections (3.1.1.1. to 3.1.1.4). The categories of weather and climate events that are considered in this chapter are discussed in Section 3.1.2, general characteristics of weather and climate events relevant to disasters are addressed in Section 3.1.3, and Section 3.1.4 and 3.1.5 briefly discuss issues associated with compound events, as well as the impacts of weather and climate extremes on the physical environment and associated feedbacks. Requirements and methods for investigating observed and projected changes, the underlying mechanisms and causes, and associated uncertainties are addressed in more detail in Sections 3.2.1 to 3.2.4. Finally, assessments on observed and projected changes in the considered weather and climate events are provided in Sections 3.3 to 3.5.

START BOX 3.1 HERE

Box 3.1: Extreme Impacts (of Non-Extreme Events) Versus Impacts of Extreme Events

As noted in Section 3.1.1.1, from a statistical perspective, extreme events are often defined as being equivalent to "rare" events, i.e., events from the extreme tails of the frequency distribution of a weather/climate variable (e.g., AR4 glossary definition). Many such extremes have a close association with disasters (e.g., heavy rainfalls with flood-related impacts, extreme temperatures with health impacts), though some rare events may not necessarily have extreme impacts in all climate regimes or regions.

In the context of a changing climate, an unprecedented extreme can arise when a trend in a weather/climate variable (e.g., temperature) contributes to a situation outside of the climatological frequency distribution for the variable (e.g., previously unobserved high temperature in this case). This may occur due to changes in mean, variability or shape (e.g., skewness) of the frequency distributions of the given weather/climate variables (Box 3.2). Note that because some of these changes may be slow (e.g., mean sea level rise), they might be considered as part of the climatological range within given time periods (i.e., 2nd half of 21st century), though extremes with respect to present-day climate. This aspect (abruptness of change) is also of strong relevance for adaptation, though we do not specifically address it in the present Chapter (e.g., temperature extremes are defined with respect to the 1960-1990 reference period in Tables 3.2 and 3.3. and Figures 3.1-3.4).

Events that may not be rare in a statistical sense (e.g., 80th percentile) may also be associated with extreme impacts, in particular if they are linked with the crossing of important thresholds: e.g., a medium deficit in precipitation in a region where mean evapotranspiration has significantly increased, moderately extreme ENSO events, or specific temperature thresholds for human health. Also the accumulation of several events which may each only be mildly extreme can lead to extreme impacts, as is the case for compound events or multiple clustered events (Section 3.1.4 and Box 3.4). Conversely, an extremely rare event may not necessarily lead to major impacts and disasters if it is not associated with some critical thresholds for the impacted systems (either by its nature or because of adaptation). Most global studies of changes in physical extremes do not consider how such extremes are related to actual impacts in the affected regions. While this aspect cannot be addressed in the present Chapter due to lack of corresponding literature, it should be noted that this gap could possibly be filled in the future if information on critical thresholds and their links to physical climate and weather events is more clearly inferred from impact studies (by conducting sensitivity experiments instead of driving impact models with single projections).

To illustrate how the resultant impacts may frame the definition of physical extremes and the identification of relevant changes in the context of global warming, Box 3.1, Figure 1 represents the relationship of the hypothetical frequency distribution of a weather/climate variable (top) with two impacts (bottom). In some cases, impacts may increase linearly with the intensity of the event. However, non-linear effects linked with discrete thresholds are common (e.g., Corti et al., 2009). The two hypothetical impact functions A and B in Box 3.1, Figure 1 (bottom) are assumed to be characterized by such critical thresholds. Note that these respective impact functions may be related either with physical (e.g., soil moisture content, slope instability) or social (health system, early warning systems, disaster risk management infrastructure) components, although we only consider the former of these two cases in the present Chapter given its scope. Threshold A lies within the present climate distribution and is not related to extreme conditions in a statistical sense, while threshold B lies outside the present climate distribution. From an impact perspective, both physical thresholds are relevant, even if only threshold B can be considered as a statistical extreme (AR4 definition) within the climate variable distribution.

INSERT BOX 3.1, FIGURE 1 HERE

Box 3.1, Figure 1: Link between climate/weather variable probability distribution function (PDF) and associated impacts (A and B), and implication for definition of "climate extremes" (see discussion in text). Note that the PDF of a climate variable is not necessarily Gaussian.

Box 3.1, Figure 2 illustrates possible changes in the frequency distribution of the weather/climate variable and the respective impact functions under climate change. The fact that a previously rare or extremely unlikely event occurs within the "new" climate is a function of the change in the PDF, i.e., in both mean and variability (examples C1 and C2), or even higher moments (e.g., skewness). However, the impact functions and related thresholds can also be modified (examples IA1 and IA2). This can be due to the adaptation of the society to the changed climate conditions (i.e., decreased impacts for the same threshold, and/or higher threshold, example IA1). Conversely, an increased vulnerability and susceptibility to damage for the same threshold may occur (example IA2), which may (or may not) be itself a consequence of climate change (e.g., modified land cover, compound events, increased overall vulnerability of society).

INSERT BOX 3.1, FIGURE 2 HERE

Box 3.1, Figure 2: Link between climate/weather variable PDF and associated impacts under climate change (see discussion in text). Note that the PDF of a climate variable is not necessarily Gaussian.

Hence, a comprehensive assessment of projected impacts of changes in climate extremes with enhanced greenhouse gas concentrations needs to consider how changes in atmospheric conditions (temperature, precipitation) translate to physical (e.g., droughts, floods, sea level rise), ecosystems (e.g., forest fires) and human systems (casualties, infrastructure damages) impacts. Links between climate events and physical impacts are addressed in the present chapter, while links to ecosystems and human systems impacts are addressed in Chapter 4. Note that these various impacts are related, since many impacts on human systems are themselves the results of impacts on physical systems or ecosystems.

An example of the complex links that can lead to physical impacts is illustrated in Figure 3.11 in Section 3.5.1. for the case of (meteorological, agricultural and hydrological) droughts. Similarly, Figure 3.12 in Section 3.5.5 illustrates the complex relationships between climate, weather phenomena and physical impacts in the coastal zone.

END BOX 3.1 HERE

START FAQ 3.1 HERE

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, as a whole, has become more or less extreme than in the past. 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 behaviour in the climate. Nevertheless, changes in integrative metrics of impacts, such as insurance payouts, could in principle provide a multi-sectoral indicator of whether the climate in a given region is becoming more extreme.

Widespread changes in some extremes (e.g., minimum temperatures) are being observed. Recent decades have also seen increasing weather and climate related insurance losses. As well, the media coverage of weather and climate disasters is becoming more global. With improving communication technology, news of a weather or climate disaster in one location can quickly spread to the whole world. As a result of all these factors, it is not surprising that the question of whether the global climate is becoming more extreme or more variable is often asked.

One possible approach for evaluating whether specific aspects of the climate are becoming more extreme would be to determine whether there have been changes in the habitual range of variation of certain 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 the temperature climate in that region had become more extreme. Temperature variations might therefore be considered as becoming more extreme if the difference between the highest and the lowest temperature observed in a year becomes increasingly larger. According to this approach, daily temperature over the globe may have become less extreme because there have generally been greater increases in annual minimum temperatures globally than in annual maximum temperatures. On the other hand, using such an approach, one might conclude that daily precipitation variations have become more extreme because observations suggest that the magnitude of the heaviest precipitation events has increased in many parts of the world.

Another approach, considering a somewhat different aspect of climate behaviour, 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 (Box 3.1). For example, an increase in the mean temperature alone usually results in an increase in hot extremes such as "unprecedented" heat waves 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. Note however, that both of these changes may have serious impacts. For example, increases in heat stress related mortality in humans and other organisms has been observed when very high daytime maximum temperature thresholds are repeatedly crossed as in a heat wave, and the winter mortality of pests such as the pine bark beetle, decreases when critical winter low temperature thresholds are crossed less frequently in temperate climates.

Many other approaches for assessing changes in the extremeness of climate, involving different aspects of climate behaviour and either individual or multiple climate elements, could be considered. Such approaches could use the internationally accepted indictors that are designed to monitor changes in simple extreme events, such as the extremes of daily precipitation accumulations, but would also have to consider indicators of change in complex extreme events resulting from a sequence of individual events, or the simultaneous occurrence of different types of extremes (Box 3.4). As the discussion above suggests implicitly, 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.

An inescapable fact of extremes is that their occurrence often has impacts that have 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 payoff to quantity if there has been an increase or decrease in that instrument. This instrument can be useful in risk management and disaster preparedness. But the development and use of such an instrument is always related to vulnerability and exposure and thus while it may, in principle, be possible to consider an instrument that is interpretable as a measure of climatic extremeness in a broad sense, it is difficult to disentangle changes in the instrument that reflect changes in vulnerability or exposure and that reflect 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 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.

It appears that there is no simple answer to the question if climate, as a whole, has become more or less extreme than in the past. For example, depending upon how “more extreme” and “less extreme” are defined, observed changes in temperature and precipitation could be interpreted as indicating that the climate has become either more or less extreme. It is difficult to devise a metric with a clear physical interpretation that encompasses multiple aspects of extremes or variability of weather and climate in a region or in the world that quantifies changes in the extremeness of climate in some overall sense, since there are very many different sorts of climate extremes and the relationships between various types of extremes and their impacts on human systems and ecosystems can be very complex. Economic instruments, such as insurance payouts, could, in principle, provide a means for determining whether climate is becoming more extreme in a comprehensive step, but the instrument would have to be carefully designed so that it could effectively separate the effects of non-climatic factors, such as changes in vulnerability and exposure, from purely climatic factors.

END FAQ 3.1 HERE

3.1.1.1. Identification and Definition of Events that are Relevant from a Risk Management Perspective

The identification and definition of weather and climate events that are relevant from a risk management perspective is complex and depends on the stakeholders involved. For instance, it is essential to distinguish between events that are extremes in a statistical sense (but may not necessarily have extreme impacts), and events that, without being located in the tails (extremes) of the statistical distribution of the specific variable, can lead to major impacts and disasters (e.g., critical thresholds, compound events). This distinction is addressed in Chapter 1 and discussed further in Box 3.1.

This perspective implies that we consider in Chapter 3 a wider range of climate and weather events, phenomena and impacts than those strictly defined as “extreme events” in the IPCC AR4. Indeed, the AR4 Glossary (IPCC, 2007a) provides the following definition for extreme events: “An extreme weather event is an event that is rare at a particular place and time of year. Definitions of *rare* vary, but an extreme weather event would normally be as rare as or rarer than the 10th or 90th *percentile* of the observed *probability density function*. By definition, the characteristics of what is called *extreme weather* may vary from place to place in an absolute sense. Single extreme events cannot be simply and directly attributed to *anthropogenic* climate change, as there is always a chance the event in question might have occurred naturally. When a pattern of extreme weather persists for some time, such as a season, it may be classed as an *extreme climate event*, especially if it yields an average or total that is itself extreme (e.g., *drought* or heavy rainfall over a season).”

In this chapter, beside *extreme events* corresponding to the above definition, we also consider *phenomena* that can influence the occurrence and intensity of extreme events and disasters, and *physical impacts* (see Section 3.1.2). Climate phenomena such as El Niño and tropical cyclones induce extreme (low or high) precipitation, wind, sea surface temperatures (SSTs) and sea levels, and thus contribute to “extreme events”. However, they are *per se* “normal” features of climate variability. Because of their links to extreme events, we need to consider all of their occurrences, and not only “extreme” El Niño events and tropical cyclones. Similarly, because physical impacts such as droughts, floods and landslides may occur as the result of the (extreme) combination of several non-extreme events (Section 3.1.4), some aspects linked to changes in mean climate (e.g., mean temperature changes or mean precipitation changes) need to be considered as well.

Regarding the use of the term “rare” in the AR4 definition quoted above, it is important to note that the “rarity” of a weather or climate event is not a well-defined concept. First, different percentiles may apply to the definition of extremes depending on the considered variable. A one in ten event as referred to in the above AR4 definition (10th or 90th percentile) is in many cases not sufficiently rare to qualify as “extreme”, and 5th/95th or 1st/99th percentiles are more appropriate; but in some cases, even more frequent events (one in five) may qualify as extremes, for instance because they are associated with specific thresholds and/or the given events, phenomena or impacts do themselves only occur occasionally (e.g., El Niño). Furthermore, rarity can only be determined for a given time period and region. A rare event in the present climate (100-year flood or 99%-percentile temperature or sea level) may become much more common under the future climate conditions, and, strictly speaking, may thus not be an "extreme event" any more. Depending on the ability of society to adapt to such changes (which also depends on the pace at which they occur), these may or may not lead to enhanced impacts and disasters. We address issues related to the definition of "extreme events" and of weather and climate events related to disasters in more detail in Box 3.1 and Section 3.1.3.

START BOX 3.2 HERE

Box 3.2: Do Changes in Extremes Scale with Changes in Mean Climate?

Changes in extremes can be caused by changes in the mean, or variability, or both. 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, or from a change in the variability or shape of the frequency distribution of daily maximum temperatures. Most climate change research, whether focussed on past, current or projected changes, has concentrated on documenting changes in mean quantities such as average temperatures, or total precipitation. If changes in the frequency of occurrence of hot days were mainly caused by 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 we may say that changes in the frequency of hot days "scaled" with changes in mean maximum temperature. If this was generally the case 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 extrapolation from changes in the mean maximum temperature to changes in the frequency of occurrence of hot days would be less credible. Is there evidence that the effect of changes in the mean temperature on the frequency of occurrence of an extreme event such as a hot day is sufficiently strong that we can ignore possible changes in the variability and shape of the frequency distribution?

Evidence regarding how strongly changes in extreme temperatures "scale" with changes in the mean temperature comes from empirical and modelling studies. Griffiths et al., (2005) examined trends (1961 – 2003) in daily maximum and minimum temperatures across the Asia – Pacific region. Significant decreases were observed in both maximum and minimum temperature standard deviation in China, Korea and some stations in Japan (probably reflecting urbanization effects), but also for some Thailand and coastal Australian sites. The South Pacific convergence zone (SPCZ) region between Fiji and the Solomon Islands showed a significant increase in maximum temperature variability. They concluded that for non-urban stations, the dominant distribution change for both maximum and minimum temperature involved a change in the mean, impacting on one or both extremes, with no change in standard deviation. This occurred from French Polynesia to Papua New Guinea (except for maximum temperature changes near the SPCZ), in Malaysia, the Philippines, and several outlying Japanese islands. For urbanized stations changes in *both* the mean and variance, impacting on one or both extremes, were found. This result was particularly evident for minimum temperature. These results suggest that changes in mean temperature may be used to predict changes in extreme temperatures, at least for non-urban tropical and maritime locations. But at urbanized or higher latitude locations, changes in variance should be considered as well. This is also illustrated by Figure 3.5 and analyses for the European continent (e.g., Klein Tank and Können, 2003; Brunet et al., 2006; Della-Marta et al., 2007b, see also Section 3.3.1.1)

An assessment of available studies of short-duration heavy precipitation in northern America (CCSP, 2008) found that in some regions there was an *increase* in heavy and/or very heavy precipitation even if there was no change or even a decrease in total (seasonal or annual) precipitation. So the assumption that extreme short-duration precipitation "scales" with changes in the total precipitation does not seem to be justified.

Models can also be used to examine the strength of the relationships between changes in mean or total quantities, and changes in extremes. Christensen and Christensen (2003) used a high-resolution climate model to examine the influence of greenhouse-gas-induced global warming upon heavy or extended precipitation episodes in Europe. Their results indicated that CO_2-induced warming might lead to a shift towards heavier intensive summertime precipitation, despite a projected mean decrease in summer precipitation, and suggested that this might be explained by the fact that the atmosphere will contain more water in a warmer climate (according to the Clausius–Clapeyron equation). Frei et al., (2006) analysed precipitation extremes simulated and projected by six European regional climate models (RCMs) and found that projected extremes increased more or decreased less than would be expected from the scaling of present day extremes. Also in Central Europe, climate-change projections suggest a stronger increase of temperature extremes compared to mean temperature (Fischer and Schär, 2010), in particular associated with soil moisture-temperature feedbacks (Seneviratne et al., 2006a).

Kharin et al., (2007) examined temperature and precipitation extremes and their potential future changes from an ensemble of global coupled climate models used in AR4. They found that changes in warm extremes were generally associated with changes in the mean summertime temperature. Cold extremes warmed faster than warm extremes by about 30%–40%, globally averaged, although this excessive warming was generally confined to regions where snow and sea ice retreat with global warming. With the exception of northern polar latitudes, relative changes in the intensity of precipitation extremes generally exceed relative changes in annual mean precipitation, particularly in tropical and subtropical regions.

The results of both empirical and model studies thus indicate that although in some situations extremes do scale closely with the mean, there are sufficient exceptions from this that changes in the variability and shape of probability distributions of weather variables need to be considered as well as changes in means, if we are to reach credible conclusions regarding possible future changes in extremes. This appears to be especially the case for short-duration precipitation, and temperatures at urban locations or in mid- and high-latitudes.

END BOX 3.2 HERE

3.1.1.2. Observations and Model Experiments to Analyse Past and Projected Changes in Identified Extremes, and the Underlying Mechanisms and Causes

The availability of observational data to analyse changes in identified events and to investigate the mechanisms of such changes is of central relevance for risk management. While observational data for variables such as temperature and precipitation are generally available in many parts of the world (despite a number of issues with these data), some other variables are almost unmonitored (soil moisture), or not monitored with sufficient temporal or spatial resolution to assess certain extremes (wind). Furthermore, changes in exposure and other problems can limit the availability of data to assess changes in monitored climate variables (which require long, homogenous series of observations). There has been progress regarding data issues in the past 15 years, partly in response to previous IPCC assessments that strongly highlighted these problems. These various aspects and their relevance to the analysis of trends in extremes are addressed in Section 3.2.1.

In order to produce credible projections of changes in identified climate events, phenomena or impacts, the relevant processes leading to these need to be reliably represented in global climate models (GCMs). This may not be feasible for all types of climate events, especially to the level of detail that is required for inferring associated impacts (Section 3.2.3). Indeed, some variables are not well or not at all simulated in current GCMs, and few observations are available to constrain or improve their representation. In some cases, dynamical or statistical downscaling can be used to compensate for these issues (for instance, by improving the representation of topography and land surface heterogeneity and their influence on extreme events), although downscaling has other inherent limitations (Section 3.2.3). 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 (see next subsection).

3.1.1.3. Assessments of Confidence in Estimates Regarding Likelihood of Changes in the Extremes

Key information necessary for risk management includes the confidence in estimates regarding the likelihood of changes in identified relevant (extreme) events and the credibility of climate models in capturing and projecting the underlying processes. Indeed, risk management requires these in order to assess the uncertainty in the given projections, the risk of occurrence of relevant events, and the cost of preventive measures.

Given the relevance of this aspect for risk management and adaptation, it is important to note that changes 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 3.3). For instance, recent studies have highlighted that observed trends tend to be better reproduced by climate models in the case of temperature extremes than for precipitation extremes. Similarly, projections of changes in temperature extremes tend to be more consistent between climate models than is the case for (wet and dry) precipitation extremes. Other, more complex extremes are even more difficult to simulate and project (e.g., agricultural/soil moisture drought, wind extremes, tropical and extra-tropical cyclones). These issues are addressed in more detail in the individual sections on the specific extremes, phenomena and physical impacts considered in this chapter (Sections 3.3. to 3.5), as well as in Box 3.3. Overall, we can infer that **our confidence in past and future changes in extremes depends on the type of extreme, as well as on the region and season, linked with the level of understanding and reliability of simulation of the underlying processes** (Box 3.3).

In this chapter, all assessments regarding past or projected changes in extremes are expressed using likelihood statements, as described in the AR4 Working Group I Technical Summary (Solomon et al., 2007). As pointed by Risbey and Kandlikar (2007), likelihood statements implicitly include confidence assessments of the tools and data basis (models, data, proxies) used to assess or project changes in a specific element, and the associated level of understanding. Thus, in the case of changes in extremes for which confidence in the "tools" or "data basis" is low, no likelihood assessment would be provided, even if the available climate projections display a high congruence. Examples of such cases for model projections are when models display a poor performance in simulating the specific extreme in the present climate, or when insufficient literature on model performance is available for the specific extreme, e.g., due to lack of observations. Similarly for observed changes, evidence may be based on scattered data (or publications) that are not sufficient to provide a robust assessment for a large region, or the observations may be of poor quality or only of indirect nature (proxies). In the case of changes in extremes for which confidence in the models and data is rated as "medium" (that is we have some confidence in the tools and evidence available to us, but there remain substantial doubts about the quality of these tools), likelihood assessments could be provided but would be weakened to take into account the level of confidence. In such cases the assessment would be that a specific change is "more likely than not" if enough evidence is available to at least indicate the direction of the change., however no stronger assessment would be provided. Note that this means that assessments such as "likely", "unlikely", "very likely", "very unlikely", "virtually certain" or "exceptionally unlikely" are only provided for changes in which confidence in the tools and data is high. In cases with low confidence regarding past or projected changes in some extremes, no likelihood statement is provided, but in such cases we specify whether the low confidence is due to lack of literature, lack of evidence (data, observations), or lack of understanding (Table 3.1).

START BOX 3.3 HERE

Box 3.3. How does the Credibility of Climate Change Projections of Extremes Differ Geographically and Between Variables?

Comparisons of observed and simulated climate demonstrated good agreement for many climate variables, especially at large horizontal scales (e.g., Räisänen*,* 2007). For instance, Box 3.3, Figure 1 and Box 3.3, Figure 2, which are reproduced from Figure 9.12 of the IPCC AR4 (Hegerl et al*.,* 2007) compare the ability of 14 climate models to simulate the decadal variations of temperature through the 20th century. When the models included both natural and anthropogenic forcings, they consistently reproduced the decadal variations in global mean temperature (see panel at bottom left-hand corner of Box 3.3, Figure 1). Without the anthropogenic influences the models consistently failed to reproduce the decadal temperature variations. However, when the same models' abilities to simulate the temperature variations on smaller domains are assessed, although the mean temperature produced by the ensemble generally tracked the observed temperature changes, the consistency between the models was poorer than was the case for the global mean. 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.

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 (i.e., less consistency between models) 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 signal expected from anthropogenic factors is large relative to the model uncertainty and the natural variability, compared with higher latitudes. Box 3.3, Figure 1 and Box 3.3, Figure 2 also indicate 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.

INSERT BOX 3.3, FIGURE 1 HERE

Box 3.3, Figure 1: Comparison for the Americas of multi-model data set of model simulations containing all forcings (red shaded regions) and containing natural forcings only (blue shaded regions) with observed decadal mean temperature changes (°C) from 1906 to 2005 from the Hadley Centre/Climatic Research Unit gridded surface temperature data set (HadCRUT3, Brohan et al., 2006). The panel labelled GLO shows comparison for global mean; LAN, global land; and OCE, global ocean data. Remaining panels display results for 22 sub-continental scale regions. Shaded bands represent the middle 90% range estimated from the multi-model ensemble. Note that the model simulations have not been scaled in any way. The same simulations are used as in Figure 9.5 of AR4 (58 simulations using all forcings from 14 models, and 19 simulations using natural forcings only from 5 models) (Hegerl et al*.,* 2007). Each simulation was sampled so that coverage corresponds to that of the observations, and was centred relative to the 1901 to 1950 mean obtained by that simulation in the region of interest. Observations in each region were centred relative to the same period. The observations in each region are generally consistent with model simulations that include anthropogenic and natural forcings, whereas in many regions the observations are inconsistent with model simulations that include natural forcings only. Lines are dashed where spatial coverage is less than 50%. From Hegerl et al., (2007).

INSERT BOX 3.3, FIGURE 2 HERE

Box 3.3, Figure 2: Same as Box 3.3, Figure 1 for Europe, Africa, Asia and Oceania. From Hegerl et al., (2007).

Uncertainty in projections of extremes also depend on the considered variables, phenomena or impacts. There is more model uncertainty for variables other than temperature, especially precipitation (Räisänen, 2007; Hawkins and Sutton, 2010, see also Section 3.2.3). And the situation is more difficult again for extremes. Thus climate models simulate changes in extreme temperatures quite well, but the frequency, distribution and intensity of heavy precipitation is less well simulated (Randall et al., 2007) as are 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 than for (wet and dry) precipitation extremes (Tebaldi et al., 2006) and significant inconsistencies are also found for projections of agricultural (soil moisture) droughts (Wang, 2005). 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.

In summary, confidence in climate change projections is greatest for temperature, especially on global scales, and decreases when other variables are considered, and as we focus on smaller spatial domains. Confidence in projections for extremes is lower than for projections of long-term averages.

END BOX 3.3 HERE

3.1.1.4. Prediction Tools for Early Warning and Forecasting of Extremes

In the context of global warming, climate models can be used not only for long-term climate change projections, but also for short-term and, in particular, subseasonal and seasonal-to-interannual predictions (with some differences in the level of complexity of the represented physical processes). In this respect, they can also be viewed as tools potentially helping adaptation to climate change, despite limitations for some climate extremes (Sections 3.1.1.3 and 3.1.1.4). There have been significant advances in this research field in recent years and new developments are currently taking place (Case study9.x, Chapter 9). These applications, which provide a direct testing of model algorithms, might also help improve the quality of long-term projections for currently less well simulated climate extremes.

3.1.2. Categories of Weather and Climate Events to be Discussed in this Chapter

In this Chapter, we focus on changes in weather and climate relevant to extreme events and disasters, grouped into the following categories (Table 3.1):

- Weather and climate elements (temperature, precipitation, wind)
- Phenomena influencing the occurrence of weather and climate extremes (monsoons, El Niño and other modes of variability, tropical and extratropical cyclones)
- Impacts on the natural physical environment (droughts, floods, extreme sea level, waves, and coastal impacts, as well as other physical impacts, including cryosphere and permafrost-related impacts, landslides, and sand and dust storms)

The possible relevance of these elements, phenomena, and impacts to disaster risk management is discussed in Sections 3.3 to 3.5, along with observed and projected changes and the apparent causes and uncertainties. Table 3.1 summarises our overall (global) assessments of observed and projected changes, and of the attribution of the observed changes, for each category or phenomenon. Note that impacts on ecosystems (e.g., bushfires) and human systems (e.g., urban flooding) are addressed in Chapter 4. Tables 3.2 and 3.3 (and Figures 3.1 to 3.4) provide more regional detail of observed and projected changes in temperature and precipitation extremes, for which there is more detailed information available than for some of the other events and phenomena listed in Table 3.1.

INSERT TABLE 3.1 HERE

Table 3.1: Overview of considered extremes and summary of observed and projected changes on global scale. Regional details on observed and projected changes in temperature and precipitation extremes are provided in Tables 3.2 and 3.3.

It is noteworthy that the distinction between the three categories outlined above and in Table 3.1 is somewhat arbitrary, and many categories are related. In the case of the third category, "impacts on the natural physical environment", a specific distinction between these events and those considered under "weather and climate elements" is that they are not induced by changes in only one of the considered weather and climate elements, but are generally the results of specific conditions in several elements, 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 meteorological and surface conditions (and are thus often better viewed as compound events, see, e.g., Section 3.1.4 and Box 3.4). Indeed, floods will more likely occur over saturated soils (Section 3.5.2), even in the case of moderate precipitation events, and droughts can be linked to precipitation deficits but may in many cases also be impacted by enhanced temperature and radiation and associated evapotranspiration excess as well as by pre-event soil moisture conditions (Section 3.5.1). Similar considerations apply to the other types of extremes included in this category.

Another arbitrary choice made here is the separate category for phenomena 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 see droughts in some regions with, simultaneously, heavy rains and floods occurring elsewhere. Similarly, impacts of monsoons in terms of weather and climate extremes are generally related to either drought or flood conditions induced by the monsoon conditions. It could, of course, be feasible simply to examine such changes under the respective headings of "droughts" and or "heavy precipitations", or "floods". However, a change in the frequency or nature of El Niño – Southern Oscillation episodes would affect extremes in many locations simultaneously (also linked with modifications of the relationships between these episodes and precipitation in specific regions). Similarly, changes in monsoon patterns would affect large regions and often several countries. This is especially important from an international disaster perspective because coping with disasters in several regions simultaneously may be challenging (see also Box 3.4).

3.1.3. Characteristics of Weather and Climate Events Relevant to Disasters (Duration, Timing, Magnitude)

Several physical characteristics of climate and weather events are relevant to disasters. One important characteristic is the rarity of the given event, i.e., whether it is located in one of the extreme tails of the distribution of the weather/climate variable ("extreme event" following the definition provided in the IPCC AR4 glossary, see Section 3.1.1.1). Other relevant aspects include the event's duration, intensity, spatial area affected, timing, frequency, onset date, continuity (i.e., whether there are "breaks" within a spell), and pre-conditioning (e.g., rapid transition from a slowly developing meteorological drought into an agricultural drought). Those aspects most relevant for resulting impacts are determined by potential critical thresholds within the affected physical systems, ecosystems or human systems (Box 3.1).

The very nature of extremes, their differing spatial and temporal scales, and their dependency on the climate state and context (i.e., season and region, such as summer versus winter droughts in extratropical regions), means that it is not practical nor useful to define extremes precisely (see also Sections 3.1.1.1 and Box 3.1). In the climate literature, the term "extreme events" has been used broadly to describe a range of phenomena. Different definitions and thresholds have been used in different analyses.

One way to examine changes in short duration extreme events (or weather extremes) is to analyze descriptive extremes "indices" based on daily data. These indices may summarize complex events, such as the frequency of tropical cyclones of a given intensity within an ocean basin, but more often they involve basic weather elements such as temperature or precipitation. Such indices often involve the calculation of the number of days in a year exceeding specific thresholds defined relative to the climate such as 90th percentile or as a fixed value. Examples of such "day-count" indices (Alexander et al., 2006) are the number of days with minimum temperature below the long-term 10th percentile in the 1961–1990 base period (relative thresholds), or the number of days with rainfall amount higher than 25mm (absolute thresholds). These extreme events are of moderately extreme nature and typically occur a few times every year. Indices that are based on the frequency of exceedance of absolute thresholds (e.g., a daily rainfall exceeding 25mm or the number of frost days) may not reflect extremes in all locations. Day-count indices based on relative thresholds such as percentiles partially allow for spatial comparisons, because they sample the same part of the PDFs of the given variables at each location. Averaging such indices across large regions may enhance signal to noise ratios and thus improve the chances of detecting the responses of extremes to external forcing. Nonetheless, the comparability can be hindered by the fact that the PDFs may actually look very different in the tail beyond the indicated percentile values. Depending on their definition (for instance whether they consider seasonal changes), it may also be difficult to link the changes in these indices with underlying physical processes or changes in impacts. For example, a decrease in the annual number of days with minimum temperature below the long-term 10th percentile may be due to an increase in winter temperature and/or an increase in summer temperature, and thus may not be induced by the same processes nor correspond well with changes in low-temperature related impacts in all regions. For this reason, mechanistic and impact studies generally need to be performed on the regional scale and have to consider the timing of extremes. In the case of impact studies, regional vulnerability also needs to be taken into account: indeed, similar percentiles may not be associated with the same impacts in different regions (Box 3.1).

Such "extremes indices" can also be expanded to include events that are not necessarily of short duration (e.g., longest drought period within a 10-year period) or even quantities that are not extreme events per se (e.g., growing season length) but can be related to impacts and disasters induced by climate and weather events. They may also be based in some cases on monthly values rather than daily data. It is important to note that the processes that produce short duration extreme weather events and long duration extreme climate events (e.g., multi-year drought) can be very different. As a result, daily weather occurring during an extreme climate event may not always be extreme. For example, if a summer is extremely wet (i.e, the **seasonal rainfall total** is well above average), this does not necessarily mean that extreme **daily** precipitation amounts will be observed during that summer, and in fact, there can still be days without any precipitation. This is why sets of extreme indices need to be optimally chosen so as to consider the range of characteristic time scales of impacts and disasters induced both by weather and climate events.

Several lists of extreme indices have been established within international projects and initiatives (sometimes including more than 100 indices). The usefulness of such lists of indices is that they allow comparability across studies, as well as between observational and modelling studies. Moreover, in the case of observations, derived indices may be easier to get access to than raw data, which are generally not freely distributed by meteorological services. Examples of studies based on the analysis of such extreme indices are provided in e.g., Jones et al., (1999), Haylock and Nicholls (2000), Frich et al., (2002), Klein Tank et al., (2002), Schmidli and Frei (2005), Alexander et al., (2006), Tebaldi et al., (2006), Perkins et al., (2009).

An alternative approach to the use of extreme indices is statistical Extreme Value Theory (EVT), which is generally used to describe the frequency and intensity of rare events that typically occur less than once per year or period of interest. One approach, called the block maximum approach, predicts that the most extreme value in a block (of time) will tend to have the Generalized Extreme Value distribution (GEV; e.g., Coles, 2001) as the block lengthens. Applications in climatology typically consider blocks to be of length one season, one year or in some cases, multiple years. Empirical evidence suggests that for weather elements such as temperature, precipitation, and wind speed, the GEV distribution does indeed provide a good description of the behaviour of block maxima for blocks of a season or longer. An alternative formulation of the problem, in which exceedances above a very high threshold are studied, leads to the Generalized Pareto distribution (Coles, 2001). While used less frequently, this approach is also generally found to provide satisfactory descriptions of the frequency and intensity of rare extreme events. An advantage of the GEV approach is that it is possible to account for non-stationarity, from for example, external forcing (Zwiers et al., 2010), in a relatively straightforward manner. Examples of the types of block maxima considered in the application of EVT include the annual maximum amount of precipitation collected in a day or in 5-day periods (pentads), the annual peak flow in a river, or the highest annual temperature. In engineering practice, EVT is typically used to estimate design values from such series of extreme values. It is possible to estimate the magnitude of events that are unprecedented in the available record, say events that might be expected to occur once in a hundred or thousand years, though estimation for rarer events is associated with substantially higher uncertainty. Future changes in the intensity or frequency of extreme weather and climate events can also be evaluated this way. Studies based on EVT include, e.g., those by Zwiers and Kharin (1998), Kharin and Zwiers (2000), Frei et al., (2006), Laurent and Parey (2007), Della-Marta et al., (2007a), Kharin et al., (2007), Brown et al., (2008).

The complexity of the investigation of extremes and the requirement for high-quality data to diagnose changes in extreme events (see also Section 3.2.1) means that in practice one or other of the "extreme indices" or EVT approaches may be more appropriate depending on the data availability and research question being addressed. For many issues, they can be considered as complementary.

3.1.4. Compound (Multiple) Events

Much of the analysis of changes of extremes has, up to now, focused on individual extremes of a variable. However, the simultaneous or near-simultaneous occurrence of two or more extremes of several variables (e.g., high sea level coinciding with tropical cyclone landfall) or of the same variable (also referred to as clustered multiple events) can exacerbate the impact that would be suffered from the extreme events if they occurred in isolation (Box 3.4). Examples of clustered multiple events are for instance tropical cyclones or extratropical cyclones generated a few days apart with the same path and/or intensities, which may occur when there is persistence in atmospheric circulation and genesis conditions.

Compound events may also refer to the combination of two or more climate/weather events, which, individually, may not be considered extreme, but lead together to an extreme impact. An example is an above-average (but not extreme) rainfall event falling on above-average saturated soil, and thus leading to floods. Note that this may also be the result of a series of wet days resulting in saturated soils, followed by a further (possibly even average) event that, because of soil preconditioning, may lead to a disaster such as landslide, flooding, or even dam failure. Similarly, drought and heat extremes can lead in combination to changes in the possibility or intensity of forest and bush fires (see Chapter 4). As well, the near-simultaneous occurrence of two or more weather/climate events (e.g tropical cyclones) may also be considered “extreme”, if such an occurrence is very rare.

In some cases, there might also be positive feedbacks between two types of extremes, which means that their simultaneous occurrence is not due to chance but to reinforcing mechanisms linking the two extremes (Section 3.1.5 and Box 3.4). In addition, it is also possible that the same region may be affected at the same time by different types of (unrelated) extremes (“contrasting events”), e.g., enhanced drought conditions and more frequent heavy rainfall (Box 3.4), which in combination lead to a much higher vulnerability of the region because it has to adapt simultaneously to changes in two opposite extremes. A more detailed discussion of compound events and how they may change with global warming is provided in Box 3.4. Despite their importance, neither the climate sciences nor the statistical sciences have yet developed adequate frameworks for characterizing such events and assessing whether their frequency and intensity is changing.

3.1.5. Impacts of Weather and Climate Events on the Physical Environment & Associated Feedbacks

Most atmospheric weather/climate events lead to the potential for disasters through their impacts on physical systems (soil moisture content, slope instability, erosion, sea level height) rather than their direct effects on humans or ecosystems. Examples include landslides or avalanches after heavy rains or snow, or forest fire after drought and heat waves. Thus it is important to consider how these different types of impacts are related to weather and climate (as also highlighted in Box 3.1, and in Section 3.5 for the individual considered impacts on the physical environment).

In addition, any changes in the physical environment may feed back into the weather/climate system. For instance, impacts on soil moisture availability are known to play a major part in controlling air temperature, boundary-layer development, precipitation formation and land carbon uptake (e.g., Betts, 2004; Koster et al., 2004b; Ciais et al., 2005; Seneviratne et al., 2006a; Reichstein et al., 2007; Seneviratne et al., 2010). They have also been suggested to impact monsoons in some regions (Grimm et al., 2007; Collini et al., 2008, see Section 3.4.1.2). Also, fires arising from drought might locally lead to pyrocumulus and heavy rain (Tryhorn et al., 2008). An example of a positive feedback between two types of extremes can be given for the case of droughts and heat waves in transitional climate regions, with heat waves leading to enhanced drought via enhanced evaporation, and drought conditions leading to enhanced temperature anomalies via decreased evaporative cooling (see also Box 3.4, and Sections 3.3.1 and 3.5.1). Despite these examples, there is still little literature on the role of feedbacks for the occurrence of extreme events and the interactions between different types of extremes.

Finally, it is important to note that impacts to ecosystems (Chapter 4) can also induce major feedbacks to the climate system, for instance through their modulation of soil moisture-climate feedbacks or through resulting impacts to the carbon cycle. Also socio-economic impacts (e.g., land use changes) can lead to (more indirect) feedbacks to the climate system.

START BOX 3.4 HERE

Box 3.4: Is it more Likely that in the Future Compound or Contrasting Extremes will Occur in the Same Region?

The close proximity in time of a drought followed by a flood in a specific region can have even more devastating impacts than would either extreme by itself. Most of this Chapter is devoted to assessing the literature regarding possible changes in the probability of occurrence of single extremes. The question of whether climate change may lead to changes in the probability of occurrence of pairs or groups of extremes occurring together, or at least close in time, is discussed in this Box.

Quantitative estimates of the probability that in the future more compound extremes will take place requires the determination of the degree to which the probability of occurrence of the separate events or their impacts are correlated or not, and whether this correlation may change in the future. Various causes for correlation between events and their impacts can be identified:

1. a common external forcing factor for changing the probability of the two events (e.g., regional warming)
2. mutual reinforcement of one event by the other and vice versa due to system feedbacks
3. dependence of the impact of one event on the occurrence of another one.

While relationships between events are obvious in the case of some related types of extremes (e.g., “wet extremes”, i.e., heavy precipitation and floods), it is important to also consider the probability of mutual correlation between contrasting events (e.g., increased probability of both droughts and floods in the same region) and whether it may increase (or decrease) in the context of climate change. Indeed, it may be more difficult for society to adapt simultaneously to contrasting extremes, which may require more coping capacities than in the case of related extremes.

The erratic occurrence of extreme events, related to the inherent chaotic fluctuation of the climate system, usually limits our ability to assess the mutual correlation quantitatively. However, for each of the above categories some examples can be given to illustrate the conceptual picture.

Common external forcing

Many areas in the world are exposed to climate extremes of various origins, such as droughts, heat waves, intense precipitation, storm surges or hurricanes. Quantitative estimates of changes in the likelihood of simultaneous extreme events within a given region in response to global warming require a solid and common attributed link between the occurrence of the events and the anthropogenic effect on climate. However, apart from a few regional studies, a systematic assessment of regions where multiple climate extremes are subject to change in response to global warming has not been carried out.

At a regional level, Alexander and Arblaster (2009) explored projected changes of temperature and precipitation extremes in Australia using a multi-model approach. Although in their study model results show apparent deficiencies in reproducing many of the observed trends of climate extreme indices, consensus existed in projected increases in heat wave duration and warm nights, and consecutive number of dry days and heavy precipitation contribution. For example, the models projected increases in both heat wave duration and the consecutive number of dry days.

A more anecdotic example is reported by Lenderink et al., (2009), discussing the causal link between a strong temperature anomaly in the Netherlands and surroundings in July 2006, followed by record breaking heavy rainfall in the coastal area in August upon a sudden change of the regional atmospheric circulation picking up large amounts of moisture from the North Sea. Although rapid changes of weather regimes are common to most areas in the world, this case illustrates how a common external forcing (large scale heating due to a persistent atmospheric circulation) affected both the intensity of the heat wave in July and that of the extreme precipitation in the Dutch coastal region in response to high North Sea temperatures.

Another important dimension of a common external forcing is the change of the risk of extreme events in different regions that are unrelated with respect to their climate, but related with respect to their vulnerability. For instance, the dependence of agricultural production on El Niño in several countries of the world may give rise to a widespread (global) reduction of crop yield during El Niño/La Nina events. Thus, a change in the frequency or intensity of El Niño events could, simultaneously, affect the frequency of occurrence of droughts and floods in many parts of the world. A similar pattern of change could be caused by a change in the strength of the global or regional monsoons. It is for this reason that this Chapter examines the literature related to how a changing climate might affect the El Niño – Southern Oscillation and monsoons.

Mutual reinforcement due to feedbacks

Several studies have pointed out the various land-atmosphere feedback pathways that can give rise to regional low precipitation and drought conditions (Schubert et al., 2004; Schubert et al., 2008b; van Heerwaarden et al., 2009). In addition, the risk of extremely high temperatures and heat waves can increase during drought conditions due to lack of evaporative cooling, while the hot conditions also lead to a strengthening of the drought (Seneviratne et al., 2006a; Fischer et al., 2007a; Jaeger and Seneviratne, 2010). Persistence associated with soil moisture may also affect the persistence of heat waves, though this effect appears to be small (Lorenz et al., 2010). Due to the mutual feedbacks between temperature, evaporation, soil moisture and precipitation, the probability of droughts and heat waves to occur simultaneously is thus larger than for every individual event in regions where soil moisture can become a limiting factor for evapotranspiration (Koster et al., 2004b; Seneviratne et al., 2010).

Conditional occurrence or impact of individual events

Van den Brink et al., (2005) explored the simultaneous occurrence of sea level surges and high river discharge in the Netherlands using an archive of seasonal predictions from a recent episode. The closure of a dynamic storm surge barrier depends on the water level in the harbour behind the barrier, which in turn depends both on the sea level (including tidal and surge waves) and the discharge from the Rhine River. At high discharge rates the barrier needs to close at lower sea levels than for normal discharge conditions, to avoid flooding of the harbour. Both extreme events (storm surges and extreme river discharges) can be considered to be uncorrelated, but the common impact on the inland water level introduces an effective mutual dependence. The projected increased frequency of closure of the storm surge barrier is still mainly dependent on the mean sea level rise, and quantitative estimates of the effect of changes in the river discharge regime have yet to be made.

Another example of reinforcing extremes or their impacts is the impact of the severe southeast Australian bushfires of 7 February 2009, which occurred during a very prolonged drought. The drought led to drier fuels in forests, thus making conditions more conducive for bushfires.

Contrasting extremes

The factors discussed above may also apply to contrasting extremes. For instance, Christensen and Christensen (2003) and Trenberth et al., (2005) point out that a warmer climate may lead to an increased likelihood of extreme precipitation under warmer conditions but at the same time also be associated with increased risk of drought. A projected warming leads to an increase in potential evaporation, with increased risks of both agricultural drought (due to enhanced actual evapotranspiration and possible decrease in precipitation) and meteorological droughts (due to decreased relative humidity, seasonality of precipitation in, e.g., monsoon areas, or soil moisture-atmosphere feedbacks). Simultaneously, the frequency and intensity of heavy precipitation events may increase with temperature at a rate proportional to the Clausius Clapeyron relationship or higher, as verified by using observations by Lenderink and Van Meijgaard, (2008), due to latent heat release in the showers or other feedbacks. Thus, higher temperature gives rise to both an enhanced drought risk and a higher likelihood of intense precipitation. This apparent paradox is captured in many national climate change scenarios (e.g., van den Hurk et al., 2007). These cases can be seen as examples for mutual correlation due to a common forcing (enhanced greenhouse gas concentrations).

In some cases, contrasting extremes may also lead to mutual reinforcement, or at least a mutual dependence (points 2. and 3. above). Thus intense precipitation events can be triggered in response to strong convection of air that is heated and/or moistened near the surface. For instance, there is evidence that thunderstorms caused by bushfires (pyrocumulus) can lead to flash flooding (e.g., Tryhorn et al., 2008), due to induced heavy rainfall. Moreover the fires can cause modifications of soil characteristics, thereby increasing the possibility of flooding from the heavy rain. So a warming climate may, in such cases, lead to enhanced risk of these combinations of events.

Summary

In summary, it is difficult to give a definitive answer to the question of whether compound or contrasting events may be more likely in the future. The above anecdotic evidence does, however, suggest that new, surprising combinations of events are likely to occur. It should be noted as well that enhanced impacts from compound events can also occur due to increased vulnerability of a system to a given event due to the impact of another event, or because of increased exposure to a given event due to the impact of another event (and climate change may lead to such changes in vulnerability or exposure).

END BOX 3.4 HERE

3.2. Requirements and Methods for Analysing 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. Summaries of these assessments are provided in Table 3.1. Overviews of observed regional changes in temperature and precipitation extremes are provided in Figures 3.1. and 3.2., as well as in Table 3.2. In this sub-section issues are discussed related to the data and observations used to examine observed changes in extremes. This will allow the reader to place the results in later sections and their uncertainties in context with the data used to derive these results.

Issues with data availability are especially critical when searching for 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 (Zhang et al., 2004b; Trömel and Schönwiese, 2005). Trend analyses of extreme events may require data transformations for non-normally distributed data, and accounting for serial autocorrelation in climate time series (Smith, 2008). To avoid excessive statistical limitations, trend analyses of extremes have traditionally focused on standard and robust statistics that describe moderately extreme events that occur a few times a year (see also Section 3.1.3).

Another important criterion constraining data availability for the analysis of extremes is the respective time scale on which they occur (Sections 3.1.1.1. and 3.1.3), 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 and unusually wet periods on the order of a month or longer. Most meteorological records before the 17th century consist of testimonies of extreme events that affected society and hence stuck in people's memories (Le Roy Ladurie, 1971; Heino et al., 1999). To examine changes in extremes occurring on short time scales, particularly of climate elements such as temperature and precipitation, 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.

Where data are available, several problems can still limit the analysis of observations. First, although the situation is changing, 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. The last two items are addressed in more detail in the following paragraphs.

Regarding the quality of measurements, long-term observations of climate are often available only at weather stations, such as at airports, that were designed to take observations in support of developing weather forecasts, and not for climate purposes, and this can result in lower quality data. Another problem affecting precipitation measurements is the undercatch of rain gauges, especially in winter (e.g., Sevruk, 1996; Yang et al., 2005). Furthermore, there are a number of data problems that can affect values that exceed thresholds, and are thus most relevant to the analysis of extremes. Quality control procedures designed to flag a value suspected of being erroneous can impact the research results by flagging extreme values that are truly correct, or by not flagging a truly incorrect value. This can happen in particular in the case of large daily precipitation totals associated with convective storms, or in the case of an isolated extreme temperature event. Quality assurance checks are typically implemented to examine the data on a station-by-station basis. These employ both internal checks, such as climatological bounds checks (e.g., is the value reasonable for the location and season), and spatial checks using comparison with nearby climate stations. An isolated but intense thunderstorm may result in an extreme daily precipitation total at one station, but not impact any surrounding stations and thus result, incorrectly, in a flagged value. In recent years particular care has been given to develop automated quality assurance procedures that minimize the flagging of valid observations (false positives), but do remove the truly incorrect values (Durre et al., 2008).

Whether or not climate data are homogeneous can also significantly impact the results of an analysis of extremes. 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. Inhomogeneities occur in a climate time series due to a variety of reasons. These include changes in the location of an observing station (Trewin, 2010), changes in instrumentation (e.g., the introduction of the Stevenson Screen) (e.g., Nicholls et al., 1996), the installation or removal of a wind shield on a precipitation gauge, land use/land cover changes, or changes in the daily observing time. Some meteorological elements are especially vulnerable to uncertainties caused by even small changes in the exposure of the measuring equipment. For instance, erection of buildings or changes in vegetative cover can produce a bias in wind measurements. When a change occurs it can result in either a discontinuity in the time series (slight jump) 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 homogeneity detection and adjustments for daily and sub-daily data are only now being developed (e.g., Vincent et al., 2002; Della-Marta and Wanner, 2006), and have not been widely implemented.

With respect to temperature and precipitation measurements, the above mentioned issues have been partly addressed in the past 15 years. However, they still 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, extra-tropical and tropical cyclones, Section 3.4), as well as impacts on the physical environment (e.g., droughts, floods, cryosphere impacts, Section 3.5).

Thunderstorms and tornadoes are not well observed in many parts of the world. Tornado occurrence since 1950 in the USA., 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 extra-tropical cyclones (ETCs), 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 efforts in reanalysis have attempted to produce more homogeneous reanalyses that show promise for examining changes in ETCs and other climate features (Compo et al., 2006).

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, 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 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 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 (agricultural as well as hydrological) drought mechanisms in climate, land surface and hydrological models, and the monitoring of on-going changes in regional terrestrial water storage. As a consequence, these need to be inferred from simple climate indices or model-based approaches (e.g., Heim Jr, 2002; Dai et al., 2004; Sheffield and Wood, 2008). 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. Additionally, ground observations of snow, which are lacking in several regions, are important for the investigation of several physical impacts, in particular those related to the cryosphere and runoff generation (e.g., Essery et al., 2009; Rott et al., 2010).

All of the mentioned issues lead to uncertainties in observed trends in extremes. In many instances, great care has been taken to develop procedures to improve the data which in turn helps to reduce uncertainty. Progress has been in particular achieved in the last 15 years, partly in response to previous IPCC assessments that strongly highlighted these problems. 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 data bases of daily temperature and precipitation covering up to 70% of the global land area, has allowed robust analyses of extremes (c.f., Alexander et al., 2006). These global analyses of temperature and precipitation extremes (e.g., Alexander et al., 2006) are consistent with what would be expected from analyses of mean values using homogeneity-adjusted data (e.g., Vose et al., 2005), which provides more confidence in the results (although such consistency may not necessarily be expected for all extremes at all locations, see Box 3.3.). 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 on both a global (Alexander et al., 2006) and regional basis (Sections 3.3.1 and 3.3.2). Nonetheless, as highlighted above, for many extremes, data remain sparse and problematic resulting in less ability to establish changes particularly on a global basis.

The AR4 (Trenberth et al., 2007) cited a lack of data sets available to determine long-term trends in many climate extremes. In many instances this is still the case for some variables such as wind, for small-scale phenomena such as tornadoes or hail, and also for diagnosing changes in agricultural droughts (soil moisture). For some extremes more and improved data sets have become available or have been more thoroughly analysed, since the AR4. Changes in unusually warm nights and days and in unusually cold nights and days, and heat waves since the middle of the 20th century have been now documented in more regions than were possible for the AR4. The same is true for changes in heavy and extreme precipitation events and for meteorological drought. There is more evidence for shifts in extra-tropical cyclone storm tracks and changes in intensity of these storms. However, recent developments in tropical cyclone research have led to increased uncertainty regarding past changes in tropical cyclone activity, particularly in the period before widespread satellite observations.

INSERT FIGURE 3.1 AND FIGURE 3.2 HERE

Figure 3.1: Regional observed changes in temperature and precipitation extremes (Americas)

Figure 3.2: Regional observed changes in temperature and precipitation extremes (Europe, Africa, Asia and Oceania). See Figure 3.1 for definition of symbols

INSERT TABLE 3.2 HERE

Table 3.2: Regional observed changes in temperature and precipitation extremes. Assessments for which no likelihood statements are available yet are displayed in grey in the Table (empty arrows on Figures 3.1 and 3.2).

3.2.2. The Causes Behind the Changes

3.2.2.1. Why Extremes Change and What are the Possible Causes

This section addresses the question of the attribution of causes for observed or projected changes in extremes. In Sections 3.3. to 3.5, the causes for observed changes in the evaluated extremes are assessed. A summary of these assessments is provided in Table 3.1.

Climate variations and change are induced both by the chaotic nature of the climate system (natural internal variability), and by changes in external forcings, which include natural external forcings such as changes in solar irradiance and volcanism, and anthropogenic forcings such as increased greenhouse gas emissions principally due to the burning of fossil fuels (but also air pollution and land use changes). At the global scale, it has been established by the AR4 that most of the observed increase in global mean surface temperatures since the mid-20th century is *very likely* due to the increase in greenhouse gas concentrations (Hegerl et al., 2007). On regional scales, variability internal to the climate system may play a larger role than on global scales. However, there is evidence of human influence on regional temperatures as well, at least in some regions (Stott et al., 2004; Zhang et al., 2006; Zwiers et al., 2010). Since the AR4, the effects of external forcing on changes on the hydrological cycle (Stott et al., 2010, see also Section 3.2.2.2), and on the cryosphere (Min et al., 2008b) have also been detected. The warming is expected to continue in the foreseeable future even if there is no additional increases to greenhouse gases in the atmosphere ("committed warming"), due both to the long atmospheric half-life of CO_2 and the thermal inertia of the oceans (IPCC, 2007b) although the rate of warming will reduce rapidly if atmospheric CO_2 concentrations are reduced (Matthews and Weaver, 2010). Given the impact of enhanced greenhouse gas concentrations on the climate as a whole, extremes are expected to change as well.

A diagnosed trend or change in extremes can either be the result of changes in external forcing, or a manifestation of the natural internal variability of the climate system, or a combination of the two. With scarce data (as is the case for extremes, see Section 3.2.1) and for relatively short-term trends, it can be challenging to distinguish between these two alternative explanations, which is clearly of key relevance for climate change attribution. For this latter question, one also needs to distinguish between the effects of anthropogenic and natural forcings.

When addressing the causes of diagnosed changes in climate mean and extremes at various locations, an additional dimension to be considered is the role of feedbacks and interactions between processes for these resulting changes (Section 3.1.5), and their links to external forcings. There are still many uncertainties in modelling these interactions. As well, there is still a lack of data on regional external forcing such as land use changes, and the mechanisms that cause observed changes may not be fully represented in model simulations. These factors can further complicate attribution of regional climate changes, and especially of extremes.

Since it is impossible to experiment with the real atmosphere to determine the roles of different external forcings on the climate system, our main source of information on the climate response to external forcings in the past and for the future is climate model simulations. Our understanding of how the extremes have responded to external forcings in the past, and how they will respond in the future, also needs to come from climate model simulations, either directly or indirectly (using e.g., sensitivity experiments, see also Section 3.2.2.3). Therefore, we need to use climate model simulations, together with observations, to understand the causes behind the changes in extremes.

3.2.2.2. 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. Some relevant large-scale impacts of global warming affecting extremes include the overall changes in temperature induced by enhanced radiation forcing, the enhanced humidity content of the atmosphere (linked with the Clausius-Clapeyron relationship), the increased land-sea contrast in temperatures, which can, e.g., affect circulation patterns and in particular monsoons. On the regional and local scales, other processes can contribute to modulate the overall changes in extremes, in particular land-atmosphere interactions (e.g., Seneviratne et al., 2006a). A detectable change in the mean climate can be a strong indication of a change in extremes in some circumstances (Gutowski et al., 2008b) (Box 3.3).This section briefly reviews the current understanding of the causes 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 since the mid-20th century, the AR4 concluded that they are *very likely* due for the most part to observed increase in anthropogenic greenhouse gas concentrations. Anthropogenic warming was also detected in the troposphere and in the global oceans. Greenhouse gas forcing alone during the past half century 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 during the past half century can be explained without external forcing, and *very unlikely* that it is due to known natural external causes alone. The warming took place at a time when natural external forcing factors such as solar output would likely have produced cooling. At sub-global scale, anthropogenically-forced warming over the past 50 years has also been detected in all continents (Hegerl et al., 2007; Gillett et al., 2008b).

Overall, attribution at scales smaller than continental, with limited exceptions (e.g., Barnett et al., 2008), 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 forcings at these scales. Averaging over smaller regions reduces the natural variability less than does averaging over large regions, making it more difficult to distinguish between changes expected from different external forcings, or between external forcing and natural variability. Temperature changes associated with some modes of variability are poorly simulated by models in some regions and seasons. In addition, the small-scale details of external forcing, and the response simulated by models are less credible than large-scale features. Furthermore, the inclusion of additional forcing factors, such as land-use change and aerosols that are likely more important at regional scales, remains a challenge (Lohmann and Feichter, 2007; Pitman et al., 2009; Rotstayn et al., 2009). Because of these, regional scale detection is still hard to achieve.

Nonetheless, recent work has expanded the literature in addressing the detection and attribution of changes in climate at smaller spatial scales and for seasonal averages (Stott et al., 2010). For instance, Min and Hense (2007) assessed the consistency between observed changes in surface temperature over six populated continents and several alternative proposed explanations for those changes, including influence of anthropogenic and natural external forcing, and internal variability of the climate system, based on a Bayesian decision theory. They found that anthropogenic forcing was required for most continent-season cases to best match the observed changes. Jones et al., (2008) examined summer (June-August) mean temperatures over the past century over a set of sub-continental regions of the Northern Hemisphere. When signals were regressed individually against the observations, an anthropogenic signal was detected in each of 14 regions except for one, central North America, although the results were more uncertain when anthropogenic and natural signals were considered together. Burkholder and Karoly (2007) detected an anthropogenic signal in multi-decadal trends of a U.S. climate extreme index and Dean and Stott (2009) detected a signal in New Zealand temperatures. While these new studies provide more evidence of anthropogenic influence at increasingly smaller spatial scales, they have not significantly changed the AR4 assessment on attributing regional temperature change to causes (Hegerl et al., 2007).

One of the significant advances since AR4 is the emerging evidence of human influence on global atmospheric moisture content and precipitation. According to the Clausius-Clapeyron relationship, the saturation vapor pressure increases exponentially with temperature. Since moisture condenses out of supersaturated air, it is physically plausible that the distribution of relative humidity would remain roughly constant under climate change. Observations also seem to suggest relatively constant relative humidity on climatological time scales (Peixoto and Oort, 1992). This means that specific humidity increases about 7% for a one degree increase in temperature. Indeed, observations indicate significant increases between 1973 and 2003 in global surface specific humidity but not in relative humidity (Willett et al., 2008), consistent with the Clausius-Clapeyron relationship. Anthropogenic influence has been detected 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). A comparison of observed precipitation trends over two periods during the 20th century averaged over latitudinal bands over land with those simulated by fourteen climate models forced by the combined effects of anthropogenic and natural external forcing, and by four climate models forced by natural forcing alone detected the influence of anthropogenic forcing (Zhang et al., 2007a). While these changes cannot be explained by internal climate variability or natural forcing, the magnitude of change in the observations is greater than those simulated. Furthermore, evidence from measurements in the Netherlands suggest that hourly precipitation extremes may in some cases increase more strongly with temperature (twice as fast) than would be assumed from the Clausius-Clapeyron relationship alone (Lenderink and Van Meijgaard, 2008). The influence of anthropogenic greenhouse gases and sulphate aerosols on changes in precipitation over high-latitude land areas north of 55°N has also been detected (Min et al., 2008a). Detection is possible here, despite limited data coverage, in part because the response to forcing is relatively strong in the region, and because internal variability is low in this region.

3.2.2.3. How to Attribute Causes to a Change in Extreme

The causes of climate change have been assessed based on climate change detection and attribution approaches (Santer et al., 1996; Mitchell et al., 2001; Hegerl et al., 2007; Hegerl et al., 2010). The attribution of causes to change in extremes may be assessed similarly. Recent discussion during the joint Expert Meeting of IPCC WGI/WGII has resulted in a set of definitions and terminologies on detection and attribution for both Working Groups. The resulting guidance paper on detection and attribution (Hegerl et al., 2010) has the following definitions on detection and attribution. 'Detection' of change is defined as the process demonstrating that climate or a system affected by climate has changed in some defined statistical sense without providing a reason for that change. 'Attribution' is the process of evaluating the relative contributions of multiple causal factors to a change or event with an assignment of confidence. Attribution involves careful assessment of observed changes in relation to those that are expected to have occurred in response to external forcing, typically as simulated by climate models.

There are different approaches to attribution problems but single-step attribution and multi-step attributions are most often used in climate literature. Single-step attribution to external forcings 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. Modelling can involve a single comprehensive model or a sequence of models. Multi-step attribution to external forcings comprises assessments that attribute an observed change in a variable of interest to a change in climate, plus separate assessments that attribute the change in climate to external forcings. In this case, confidence in the attribution cannot be higher than the lower confidence in the two assessment steps.

Attribution of changes in climate extremes has been a considerable challenge due to several factors. Observed data are limited in both quantity and quality (Section 3.2.1), resulting in uncertainty in the estimate 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. Global climate models may not simulate some extremes such as tropical cyclones with reasonable fidelity or may not simulate some other extremes such as small spatial scale floods at all. For some extremes (e.g., agricultural drought), too little observational data may be available to assess the model performance. In addition, differences in the spatial scale of extremes from the observations and from the model simulations also make it difficult to compare observations with model simulations. For example, climate models operate on model grids much larger than an area typically represented by an in-situ observation site. On the one hand, models are not able to produce point estimate of extremes such as the annual maximum amount of daily precipitation at an observational site; on the other hand, the limited availability of observation stations in many parts of the world makes it impossible to produce accurate estimates of area-averaged daily precipitation at model resolutions; furthermore, the scale of resolved motions may not allow a model to simulate the circulation features that produce intense precipitation in the real world.

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 alleviate this problem, and make it possible to conduct single-step detection and attribution assessment. For example, model-simulated sea level pressure 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 (Wang et al., 2009c). Barnett et al., (2008) downscaled GCM-simulated precipitation and temperature data as input to hydrological and snow depth models to infer past and future changes in temperature, timing of the peak flow, and snow water equivalent for the western U.S., and then conducted a detection and attribution analysis on human-induced changes in these variables.

A single-step attribution of cause and effect on extremes or physical impacts of extremes may not always be possible. When this is the case, multiple-step attribution may still be feasible. The assessment would then need to be based on indirect evidence, physical understanding and expert judgement, or a combination of these. For instance, in the northern high latitude regions, spring temperature has increased, and the timing of spring peak floods of snowmelt rivers has shifted towards earlier dates (Zhang et al., 2001; Regonda et al., 2005). The change in streamflow may be attributable to anthropogenic 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. In such a case, it may not be possible to quantify the magnitude of the effect of external forcing on flow regime change because a direct link between the two has not been established, so the confidence in the overall assessment would be similar to or weaker than the lower confidence in the two steps in the assessment. The physical understanding that snow melts earlier as spring temperature increases, enhances our confidence in the assessments. 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 can play an important role in such cases (e.g., Findell and Delworth, 2005; Seneviratne et al., 2006a; Haarsma et al., 2009). These can allow the distinction of the influence on extremes from different drivers that may, in turn, be influenced by external forcings.

Extreme events are by definition rare, which means that there are also few data available to make an assessment (Section 3.2.1). When a rare and catastrophic 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 the occurrence of such events usually involves multiple factors, it is very difficult to attribute an individual event to specific causes (Allen, 2003; Hegerl et al., 2007, 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. For example, Stott et al., (2004) detected anthropogenic influence on mean summer temperature in southern Europe; they then estimated the effect of anthropogenic forcing on the likelihood of a warm summer, and finally inferred an anthropogenic influence on the likelihood of the 2003 European heat wave. A similar approach has been applied to estimate the contribution of anthropogenic greenhouse gas emissions to the England & Wales autumn 2000 flood probability (Pall et al., 2010).

START FAQ 3.2 HERE

FAQ 3.2: Can we Attribute Individual Extreme Events to Climate Change?

Changes in climate extremes are expected as the climate warms in response to increasing atmospheric greenhouse gases resulting from human activities, such as the use of fossil fuels. However, determining whether a specific, single extreme event is due to increasing greenhouse gases, is difficult, if not impossible, for two reasons: 1) a wide range of extreme events occur normally even in an unchanging climate, and 2) extreme events are usually caused by a combination of factors, most of which would not be directly related to changing atmospheric composition. Nevertheless, analysis of the warming observed over the past century suggests that the likelihood of some extreme events, such as heat waves, has increased due to greenhouse warming, and that the likelihood of others, such as frost or extremely cold nights, has decreased. 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.

People affected by an extreme weather events often ask whether human influences on the climate could be held to some extent responsible. Recent years have seen many extreme events that some commentators have linked to increasing greenhouse gases. These include the prolonged drought in Australia, the extremely hot summer in Europe in 2003, the intense North Atlantic hurricane seasons of 2004 and 2005 and the extreme rainfall events in Mumbai, India in July 2005, and the historically warmest January and February in Vancouver, Canada that affected 2010 Winter Olympic Games. Could a human influence such as increased concentrations of greenhouse gases in the atmosphere have 'caused' any of these events?

FAQ 3.2, Figure 1 shows the distribution of monthly mean November temperatures averaged across the State of New South Wales in Australia, using data from 1950-2009. The mean temperature for November 2009 (the bar on the far right hand end of the Figure) lies about 3.5 standard deviations above the 1950-2008 mean. A simple statistical calculation suggests that there is perhaps less than one chance in a thousand that such a temperature would be observed in the 1950-2008 climate, and the 2009 temperature certainly looks unusual in the Figure, relative to the other years plotted there. Is this rare occurrence an indication of changing climate? In the CRUTEM3V global land surface temperature data set, about one in every 1000 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 2009 November temperature in New South Wales could have occurred in the past during a period when the effects of greenhouse gas increases were much less pronounced. A similar calculation shows that a warm month as extreme as June, 2003 in Switzerland was also not without precedent during the first half of the 20th century, although in that case, only about one in every 13000 monthly means was as extreme.

A second complicating factor 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. For example, several factors contributed to the extremely hot European summer of 2003, including a persistent high-pressure system that was associated with very clear skies and dry soil, which left more solar energy available to heat the land because less energy was consumed to evaporate moisture from the soil. Similarly, the formation of a hurricane requires warm SSTs and specific atmospheric circulation conditions. Because some factors may be strongly affected by human activities, such as SSTs, but others may not, it is not simple to isolate a human influence on a single, specific extreme event.

Nevertheless, it may be possible to use climate models to determine whether human influences have changed the likelihood of certain types of extreme events. For example, in the case of the 2003 European heat wave, a climate model was run including only historical changes in natural factors that affect the climate, such as volcanic activity and changes in solar output. Next, the model was run again including both human and natural factors, which produced a simulation of the evolution of the European climate that was much closer to that which had actually occurred. Based on

[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 deviation, and compare it with the total number of observations for the grid and month in that period. The ratio between these two numbers is 0.00107.

these experiments, it was estimated that over the 20th century, human influences more than doubled the likelihood of having a summer in Europe as hot as that of 2003, and that in the absence of human influences, the probability would probably have been one in many hundred years. More detailed modelling work will be required to estimate the change in likelihood for specific high-impact events, such as the occurrence of a series of very warm nights in an urban area such as Paris.

INSERT FAQ 3.2, FIGURE 1 HERE

FAQ 3.2, Figure 1: The distribution of monthly mean November temperatures averaged across the State of New South Wales in Australia, using data from 1950–2009. Data from Australian Bureau of Meteorology. The mean temperature for November 2009 (the bar on the far right hand end of the Figure) was more than three standard deviations from the long-term mean (calculated from 1950–2008 data).

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. Nevertheless, careful statistical analyses are required, since the likelihood of individual extremes, such as a late-spring frost, could change due to changes in climate variability as well as changes in average climate conditions. Such analyses rely on climate-model based estimates of climate variability, and thus the climate models used should adequately represent that variability. The same likelihood-based approach has been used to examine anthropogenic greenhouse gas contribution to flood probability.

Finally, it should be remembered that 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). However, a long-term trend in an extreme (e.g, heatwave occurrences), especially if observed at many locations, is a different matter. It is certainly feasible, in these circumstances, 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.

END FAQ 3.2 HERE

3.2.3. Projected Long-Term Changes and Uncertainties

In this sub-section we discuss the requirements and methods used for preparing climate change projections, with a clear focus on projections of extremes and the associated uncertainties. Much of the discussion is based closely on AR4 (Christensen et al., 2007) with consideration of some additional issues relevant to projections of extremes in the context of risk and disaster management. More detailed assessment of projections for specific extremes is 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 and precipitation extremes are provided in Figures 3.3. and 3.4. as well as in Table 3.3.

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, typically draws on information from four sources: Atmosphere-Ocean General Circulation Model (AOGCM) simulations; downscaling of AOGCM-simulated data using techniques to enhance regional detail; physical understanding of the processes governing regional responses; and recent historical climate change. At the time of the AR4, AOGCMs were the main source of globally-available regional information on the range of possible future climates including extremes (Christensen et al., 2007). A clearer picture of the more robust aspects of regional climate change was, however, emerging at that time, due to improvements in model resolution, more credible simulations of processes of importance for regional change, the availability of more and better historical climate data, and the availability of an expanding set of global simulations.

State-of-the-art AOGCMs are based on physical laws and processes expressed as equations, which the model represents on a grid and integrates forward in time. Processes with scales too small to be resolved on the spatial scale of the model grid are represented through modules based on observations and physical theory called parameterizations. This is partly due to limitations in computing power, but also results from limitations in scientific understanding or in the availability of detailed observations of some physical processes and parameters (relevant for e.g., cloud-aerosol interactions or land-atmosphere exchanges). AOGCMs show significant and improving skill in representing many important average climate features, and even essential aspects of many of the patterns of climate variability observed across a range of time scales. This makes them 'fit for purpose' for many applications. However, when we wish to project climate and weather extremes, not all atmospheric phenomena potentially of relevance can be realistically simulated using these global models and the development of projections of extreme events has provided one of the motivations for the development of regionalisation or downscaling techniques (Carter et al., 2007).

Downscaling techniques have been specifically developed for the study of regional- and local-scale climate change. Downscaling is the use of high-resolution dynamical models or statistical techniques to simulate weather and climate at finer spatial resolutions than is possible with AOGCMs – a step which is particularly relevant for many extremes given their spatial scale (e.g., convective events and wind gusts, see also Section 3.2.1). All downscaling approaches are, nonetheless, constrained by the reliability of large-scale information coming from the AOGCMs. Recent advances in downscaling for extremes are discussed below. However, as global models continue to develop, and their spatial resolution continues to improve, they are becoming increasingly useful for investigating important smaller-scale features, including changes in extreme weather events, and further improvements in regional-scale representation are expected with increased computing power (though it should not be assumed that greater resolution necessarily translates into greater credibility of projections).

There are two main downscaling approaches, dynamical and statistical (Christensen et al., 2007). The most common approach to dynamical downscaling uses high-resolution regional climate models (RCMs), currently at scales of 20km-50km, but in some cases down to 10-15km (e.g., Dankers et al., 2007), to represent regional sub-domains, using either observed (reanalysis) or lower-resolution AOGCM data to provide their boundary conditions (i.e., the atmospheric behaviour on the boundaries of the sub-domain). Using non-hydrostatic mesoscale models, applications at 1-2km resolution are also possible for shorter periods (typically a few months, a few years at most) – a scale at which clouds and convection can be resolved (e.g., Grell et al., 2000; Hay et al., 2006; Hohenegger et al., 2008). For the higher-resolution simulations (i.e., < 10-20km), double-nesting may be required (i.e., embedding of very high-resolution simulations within coarser-scale RCM simulations). 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; CCSP, 2008). The main advantage of dynamical downscaling is its potential for capturing mesoscale nonlinear effects and providing information for many climate variables while ensuring that such information is internally consistent within the physical constraints of the model. As in the case of AOGCMs, RCMs are formulated using physical principles and they can credibly reproduce a broad range of climates around the world, which increases confidence in their ability to realistically downscale future climates. For many users, the main drawbacks of dynamical models are their computational cost and that they do not provide information at the point (i.e., weather station) scale (a scale at which the RCM parameterizations would not work).

Statistical downscaling methods use cross-spatial-scale relationships that have been derived from observed data, and apply these to climate model data (Christensen et al., 2007). They also include weather generators which 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 Africa (e.g., Hewitson and Crane, 2006), Australia (e.g., Timbal et al., 2008; Timbal et al., 2009), South America (e.g., D'Onofrio et al., 2010) and Canada (e.g., Dibike et al., 2008). Statistical downscaling methods have the advantage to users of being computationally inexpensive, potentially able to access finer spatial scales than dynamical methods and applicable to parameters that cannot be directly obtained from the RCM outputs. Seasonal indices of extremes can, for example, be simulated directly without having to first produce daily time series (Haylock et al., 2006a). Although based on statistical relationships rather than physical laws, the reliability of statistical downscaling methods can be explored by assessing their ability to reproduce shifts in the observed climate (i.e., to reproduce non-stationary climates). Statistical models can, for example, reproduce the observed rainfall decline in the late 1960s in the southwest of Australia (Timbal, 2004) and in the mid-1990s in the southeast of Australia (Timbal and Jones, 2008). However, they 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 (thus minimising problems of stationarity), and in some methods, can lack coherency among multiple climate variables and/or multiple sites. In the case of downscaling extremes, one specific disadvantage of some statistical methods is that they cannot produce events greater in magnitude than have been observed before (Timbal et al., 2009). In addition, both present-day performance and the projected climate change can be very sensitive to the choice of predictors.

There have been rather few systematic inter-comparisons (in terms of both their ability to simulate present-day climate and their projected changes) of dynamical and statistical downscaling approaches, particularly inter-comparisons focusing on extremes (Fowler et al., 2007a). Two examples focus on extreme precipitation for the UK (Haylock et al., 2006a) and the Alps (Schmidli et al., 2007), respectively. The latter study indicates that the best statistical methods can reproduce the magnitude of the observed extremes with similar skill to the RCMs, but underestimate interannual variability. For users of downscaled information, the identification and selection of appropriate methods for impact assessment and adaptation planning may depend on factors such as ease of accessibility, resource requirements and type of output, as much as performance (Fowler et al., 2007a; Wilby et al., 2009).

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 emissions, concentrations of radiatively active species, radiative forcing, and climate response including downscaling. At each step, 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. Despite this, there is considerable confidence that climate models provide credible quantitative estimates of future climate change, particularly at continental scales and above (Randall et al., 2007).

The AR4 concluded (Randall et al., 2007) that one source of confidence in climate models comes from the fact that AOGCMs are based on established physical laws, while a second source of confidence comes from their ability to simulate important aspects of the current climate. Current global model ability to represent many important features of observed climate variability increases confidence that they simulate the essential physical processes relevant for the simulation of future climate change. However, the skill of global and regional climate models in representing key processes depends on the underlying processes themselves – particularly those involving feedbacks, and this is especially the case for climate extremes and associated impacts. Some processes are still poorly represented and/or understood despite major improvements in the simulations of others (see Box 3.3. and below).

A third source of confidence comes from the ability of models to reproduce features of past climates and climate changes. The AR4 demonstrated that global statistics of extreme events for present day climate are surprisingly well simulated by current AOGCMs considering their resolution and large-scale systematic errors (Randall et al., 2007). However, the assessment of climate model performance with respect to extremes, particularly at the regional or local scale, is still limited by the fact that the very rarity of extreme events makes statistical evaluation of model performance less robust than is the case for average climate. Also, evaluation is still hampered by incomplete data on the historical frequency and severity of extremes, particularly for variables other than temperature and precipitation (Trenberth et al., 2007) .

Most shortcomings in AOGCMs and in many 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). Limitations in computing power and in the scientific understanding of some physical processes, including the complexity of the feedbacks involved (Section 3.1.5), currently restrict further global and regional model improvements. 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 at approximately 20 km horizontal resolution are still not resolved sufficiently finely to simulate the high wind speeds and low pressure centres of the most intense hurricanes (Gutowski et al., 2008a). 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 AOGCMs and of most current RCMs. Extremes may also be impacted by mesoscale circulations that AOGCMs 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 e.g., with soil moisture and vegetation processes (Section 3.2.1.) and associated parameters (e.g., maps of soil types, c.f. Seneviratne et al., 2006b; Anders and Rockel, 2009).

Since many extreme events 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 in projections of changes in extremes varies with extreme type, season, and geographical region (Box 3.3). Confidence and credibility in projected changes in extremes increase when the physical mechanisms producing extremes in models are considered reliable (Kendon et al., 2009). 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 (van Oldenborgh et al., 2005; Alexander and Arblaster, 2009). It should, however, be noted that detection of trends is a signal-to-noise problem and that the noise is greater at regional and smaller scales so perhaps models should not be expected to simulate such trends well (Alexander and Arblaster, 2009). It should also be stressed that physical consistency of simulations with observed behaviour provides only necessary and not sufficient evidence for credible projections (Gutowski et al., 2008a). Knowledge on the sufficient conditions for accurate projections is limited by the fact that we do not yet know how to properly evaluate climate models for the sake of increasing credibility of projections (Glecker et al., 2008).

While downscaling techniques can improve the AOGCM information at fine scales by accounting for the effects of regional forcing, they are all still affected by systematic errors in the driving AOGCMs. Uncertainty due to structural or parameter errors in AOGCMs propagates directly from global model simulations as input to downscaling models and thus to downscaled information. Additionally, forcing factors such as land-use changes at local scales are not generally incorporated in either dynamical or statistical downscaling. Moreover, most downscaling approaches do not allow for the diagnosed fine-scale processes to feedback onto the larger scales – exceptions are approaches such as two-way nesting of RCMs (e.g., Lorenz and Jacob, 2005) or variable-resolution AOGCMS (e.g., Déqué et al., 1998). In many cases, regional downscaling has been rather ad-hoc and driven by specific and localised applications; this is especially true for statistical downscaling. As a result, there has been rather little coordinated evaluation or application of various downscaling techniques. Although some examples of intercomparisons are noted above, these are not complete end-to-end assessments fully exploring the effects of downscaling on the projected impacts, including impacts modelling uncertainty. In general, downscaled information has been rather underused in impact and adaptation assessments (Giorgi et al., 2009; Wilby et al., 2009). For example, much of the regional climate change material assessed in the AR4 WGI report was based on relatively coarse resolution AOGCM simulations (e.g., Christensen et al., 2007), although the growing use of higher resolution, downscaled scenarios for impacts assessment was acknowledged by WGII (Carter et al., 2007).

3.2.3.3. Ways of Exploring and Quantifying Uncertainties

Uncertainties can be explored, and quantified to some extent, through a combined use of observations, process understanding, a hierarchy of climate models, and ensemble simulations. Ensembles of model simulations represent a fundamental resource for studying the possible range of plausible climate responses to a given forcing (Meehl et al., 2007b; Randall et al., 2007). Such ensembles can be generated either by collecting results from a range of models from different modelling centres (multi-model ensembles) or by generating simulations with different initial conditions (intra-model ensembles) or varying multiple internal model parameters within plausible ranges (perturbed and stochastic physics ensembles).

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 AR4 AOGCM 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 simulations (referred to henceforth as the AR4 MME - multi-model ensemble) were made available in a central archive. 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.

It is important to distinguish between the uncertainty due to lack of agreement in the model projections (termed *insufficient congruence* in Tables 3.1-3.3), the uncertainty due to *insufficient evidence* (insufficient observational data to constrain the model projections or insufficient number of simulations to infer projections), and the uncertainty induced by *insufficient literature*, which refers to the lack of published analyses of projections (the terms in italic referring to assessments provided in Tables 3.1-3.3). 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 which may not be available from all modelling groups). *Insufficient congruence* of model projections, may itself be induced by different factors, most importantly the uncertainty in the initialization of climate projections, the uncertainty in emission scenarios, and the inter-model uncertainty (e.g., Hawkins and Sutton, 2009). Hawkins and Sutton (2009) examined how the influence of these three important sources of uncertainty impact the overall uncertainty of regional climate predictions of *mean temperature* changes as the forecast lead-time increases, based on global climate simulations. At short lead-times (a decade or so) natural internal variability is very important because good projections of some aspects of the internal variability (e.g., the El Niño – Southern Oscillation) rely on good initialisations of models and this may not be possible in current climate change models. At longer lead-times (50-100 years) uncertainty in future emissions of greenhouse gases ("scenario uncertainty") becomes the dominant source of uncertainty for projections of mean temperature, even on a regional scale. Inter-model uncertainty (e.g., differences between the ways models treat important aspects of the climate system such as clouds and land surface processes) is important at all lead-times for mean temperature, although it is overwhelmed by scenario uncertainty at long lead-times. Whereas the results discussed above are for mean temperature, a similar analysis for mean precipitation reveals somewhat different results regarding the impact of inter-model uncertainty, which is found to dominate the overall uncertainty at all lead times (Hawkins and Sutton, 2010). This is consistent with the analysis of e.g., Tebaldi et al., (2006), where inter-model uncertainty was found to still overlap or even be larger than scenario uncertainty for certain extremes (e.g., consecutive number of dry days) at long lead times. Hence, the respective impacts of model versus emission scenario uncertainty are expected to strongly depend on the considered variable and extreme (see also Box 3.3).

Uncertainty analysis of the 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). Where the ensemble mean projected climate change is larger than the standard deviation, the signal is generally considered to be 'robust'. 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, since the ensemble was strictly an "ensemble of opportunity", without sampling protocol, the spread of models did not span the full possible range of uncertainty. Also, the possible dependence of different models on one another (e.g., due to shared parameterizations) was not assessed.

Furthermore, this particular metric, that assesses sign agreement only can provide misleading conclusions in cases, for example, where the projected changes are near zero.

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), and new techniques have been developed for exploiting the full ensemble information, in some cases using observational constraints to construct PDFs (Tebaldi and Knutti, 2007; Tebaldi and Sanso, 2009). Perturbed-physics ensembles have also become available (e.g., Collins et al., 2006; Murphy et al., 2007), and subsequently, advances had been made in developing probabilistic information at regional scales from the AOGCM simulations, although these methods still remain in the exploratory phase and focused on variables such as mean temperature. There has been less development extending this to downscaled regional information and to extremes (Fowler et al., 2007b; Fowler and Ekstrom, 2009) although downscaling methods are maturing and being more widely applied (despite being still restricted in terms of geographical coverage).

Both statistical and dynamical downscaling methods are affected by the uncertainties which affect the global models. A further level of uncertainty associated with the downscaling step also needs to be taken into consideration. The extent to which particular GCM and RCM biases may interact with each other or cancel out (Laprise et al., 2008) has not been extensively studied – although, for example, Kjellström and Lind (2009) conclude that the wet bias over the Baltic Sea in their chosen driving GCM is reinforced by the particular RCM used. As well as structural differences in RCMs, the choice of regional domain may introduce uncertainty, and the choice of large-scale predictors is one source of uncertainty in statistical downscaling. While downscaling provides more spatial detail, the added value of this step needs to be assessed (Laprise et al., 2008). One test of this is whether or not the downscaled outputs agree better with observations than the GCM outputs for the same variable. However this is only one test of model credibility – an over-fitted statistical model, for example, may not be credible for future projections. Spatial inhomogeneity of both land-use and land-cover change, and aerosol forcing, add to regional uncertainty. This means that the factors inducing uncertainty in the projections of extremes in different regions may differ considerably.

The increasing availability of co-ordinated RCM simulations for different regions permits more systematic exploration of 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 co-ordinated simulations covering the globe is planned (Giorgi et al., 2009). Increasingly, RCM output from these co-ordinated simulations is made available at the daily timescale, facilitating the analysis of extreme events.

Attempts have been made to quantify the relative importance of the different sources of uncertainty in downscaled simulations – focusing largely on mean temperature and precipitation rather than extremes. Based on an analysis of a large European RCM ensemble, Déqué et al., (2007) concluded that for the end of the 21st century, the uncertainty in mean changes related to the choice of driving GCM is generally larger than that due to choice of RCM or emissions scenario and natural variability. However, the choice of RCM was found to be as important as choice of GCM for summer precipitation – a finding confirmed by other studies (e.g., de Elía et al., 2008). Ensuring adequate sampling of RCMs may be more important for extremes than for changes in mean values (Frei et al., 2006; Fowler et al., 2007b). Natural variability, for example, has been shown to make a significant contribution on at least multi-annual timescales and potentially up to multidecadal timescales in the case of European projections of precipitation extremes (Kendon et al., 2008).

Many weather/climate extremes have impacts on physical systems such as soil moisture and streamflow, landslides or avalanches (after heavy rains or snow, for instance), dust storms, forest fire (after drought and heat waves), and glacier mass balance. In turn, changes in the physical environment can feedback onto the weather/climate system (Section 3.1.5). The degree to which uncertainties in these feedbacks influence the regional projections of different climate variables has not been systematically studied but is not expected to be uniform.

Roe and Baker (2007) pointed out that uncertainties in projections of future climate change have not lessened substantially in the last decades. They show that the breadth of the probability distribution, and in particular, the probability of large temperature increases, is relatively insensitive to decreases in uncertainties associated with the underlying climate processes. Since then, the sources of uncertainty have in general been more widely sampled – with most studies, for example, now using multiple models and emissions scenarios rather than relying on a single model or emissions scenario. Perturbed-physics ensembles have been extended from only considering atmospheric parameters to those involved in other model components, such as carbon cycle models (Huntingford et al., 2009) – which tends to increase the upper range of the projected mean temperature change (and hence the extremes). Much of the work on uncertainty has focused on the AOGCM scale (assisted by the availability of the AR4 MME), but more work is now possible at the regional and local scale using the emerging RCM ensembles.

3.2.3.4. Specific User Needs Regarding Climate Projections of Extremes

Alongside these scientific and technical developments in climate modelling and downscaling, there has been a growing recognition post-AR4 of the need to provide appropriate projections and related documentation and guidance for decision making, particularly with respect to adaptation (although rather little consideration has been given to risk and disaster management). It is important that the most appropriate method for constructing projections is matched to the particular application with respect to factors such as spatial and temporal resolution and complexity (Wilby et al., 2009). An essential aspect of this is the improved linkage between climate and impacts modelling – it is not always possible or recommended to use raw climate model output to directly drive impacts models. The needs of hydrological modelling (both for streamflow and soil moisture), for example, impose very specific demands, including high spatial resolution and consistency, which are not yet fully met (e.g., Koster et al., 2004a; Seneviratne et al., 2010) particularly with respect to extremes (Fowler et al., 2007a; Fowler and Wilby, 2007; Maraun et al., 2010).

User needs with respect to extreme events (see Chapters 1 and 2 and 3.1) tend to be more complex than for mean climate. Thus while the former needs are reflected in the various extremes discussed in Sections 3.3 to 3.5, there are some major gaps in what can currently be covered. In particular, there is a lack of peer-reviewed work on compound (multiple) events (Section 3.1.4 and Box 3.4), although Bayesian approaches have been used to construct joint PDFs of temperature and precipitation changes (Murphy et al., 2007; Tebaldi and Sanso, 2009). Systematic changes in the exceedances of joint extremes of temperature and precipitation quantiles (cool/dry, cool/wet, warm/dry and warm/wet modes) have been found for a number of European sites in an analysis of an RCM ensemble (Beniston, 2009). By the end of the century, the 'cool' modes are almost absent, while the 'warm' modes continue the increase observed in the 20th century: with the warm/dry mode dominating for Lugano in southern Europe and the warm/wet mode dominating for Copenhagen in northern Europe.

For some extremes and applications, sub-daily information is requested by users – for analysis of urban drainage, for example. While AOGCMS and RCMs operate at sub-daily timesteps, output is rarely archived at six-hourly or shorter temporal resolutions. Where limited studies have been undertaken of RCMs, there is evidence that at the typically used spatial resolutions they do not well represent sub-daily precipitation and the diurnal cycle of convection (Gutowski et al., 2003; Brockhaus et al., 2008; Lenderink and Van Meijgaard, 2008). The use of higher spatial resolutions sufficient to resolve convection and clouds (i.e., 1-2 km) has been suggested to give improved representation of the diurnal cycle (Hohenegger et al., 2008), although higher resolution does not necessarily guarantee improved simulation of precipitation (Hay et al., 2006). 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.

High-spatial resolution is a common request from many users particularly with respect to precipitation – although detailed high-resolution climate change projections may not be critical nor essential for all aspects of adaptation planning (Dessai et al., 2009; Wilby et al., 2009). AOGCMs and RCMs provide area-averaged or spatially-aggregated precipitation (Osborn and Hulme, 1997; Chen and Knutson, 2008), while statistical downscaling has the potential to provide point or station-scale output. Area-averaging means that model grid boxes tend to have more days of light precipitation (Frei et al., 2003; Barring et al., 2006), and also reduces the magnitude of extremes, compared with point values. These scaling effects are expected because: (1) models sample area means; (2) observations sample points; (3) precipitation is not continuous over space; (4) therefore an extreme occurring at one location on a day does not mean that it will occur at other locations on the same days; and, (5), therefore it is expected that areal extremes will be smaller than point extremes. Haylock et al., (2008), for example, explored this 'areal reduction' or scaling issue in observed European temperature and precipitation extremes, comparing 25 km gridded values with station values. There is a clear reduction in the magnitude of all extremes higher than the annual 75th percentile of precipitation and the 90th percentile for temperature. The reduction factors also increase with return period – the median reduction for the 10-year return period is 0.66 for precipitation (exceeding 0.5 for some stations). Reductions of return period estimates have also been demonstrated at coarser aggregations for the U.S. (Chen and Knutson, 2008). These effects are relevant both to impacts studies and the inter-comparison of dynamical and statistical downscaling approaches (Schmidli et al., 2007; Timbal et al., 2008). The handling of these scaling issues may also have an effect on the magnitude of projected changes (Chen and Knutson, 2008).

While the spatial resolution of both global and regional models is increasing, the added value of this increased resolution should not be assumed and a balance may have to be made between spatial detail and robustness of the climate change signal (Hay et al., 2006) – the latter can be improved by spatial pooling and averaging (Fowler et al., 2007b; Coelho et al., 2008; Kendon et al., 2008). An issue with higher-resolution simulations is the fact that the resolved processes may still be insufficiently constrained with observational data, because observations are not available with the required spatial detail and comprehensiveness.

Different users and decision makers tend to be interested in projections over different future time periods. Information about changes at the end of the 21st century is more relevant where major infrastructure planning is involved, for example, while for many businesses including the insurance sector the next 20 or 30 years is considered long-term. For adaptation and development planning, the 2020s (i.e., 2011-2040) is considered important for climate risk information (Wilby et al., 2009). The focus in this chapter is on what the IPCC defines as 'long-term' projections out to the end of the century – as distinct from 'near-term' seasonal-to-decadal predictions. In the latter case, there is an attempt to produce an estimate of the actual evolution of the climate in the future, whereas long-term projections depend upon the underlying emissions scenario and the associated assumptions – developments that may or may not be realised. In the case of seasonal prediction prescribing initial conditions adequately is an important concern and the predictions themselves can be directly verified (Doblas-Reyes et al., 2009). The move towards fully initialized decadal prediction is very recent (Meehl et al., 2009b) – with the first co-ordinated simulations being developed in advance of AR5.

The AR4 MME provides output through the historical period (1850), to the present day and out to 2100 – giving flexibility in the periods for which projections can be constructed from global model output. Transient output is not yet so widely available from the more computationally expensive RCMs, but is available for Europe and North America, for example. At the time of AR4, RCMs were conventionally run for two snapshot periods – a present-day period (typically 1961–1990) and a scenario period (typically 2071–2100) (Christensen et al., 2007). Since then, emphasis has shifted more towards the middle of the 21st century to address requirements from stakeholders. Co-ordinated simulations for North America, for example, focus on 2041-2070 (Mearns et al., 2009), while a large ensemble of transient RCM runs for Europe for the period 1950–2050 has recently been completed, with many of the runs extending out to 2100 (van der Linden and Mitchell, 2009). While the signal-to-noise ratio of change is greatest at the end of the century, projections for the middle of the century or earlier are more relevant for many impacts applications. The balance of uncertainties is somewhat different for earlier compared with later future periods (see also Section 3.2.3.3). For some variables (mean temperature, temperature extremes), the choice of emission scenario becomes more critical than model uncertainty for the later future periods (Tebaldi et al., 2006; Hawkins and Sutton, 2009), though this does not apply for mean precipitation and some precipitation-related extremes (Tebaldi et al., 2006; Hawkins and Sutton, 2009), and has in particular not been evaluated in detail for a wide range of extremes.

3.2.3.5. Projections of Specific Extremes and their Confidence

In Sections 3.3 to 3.5, projections of the various extremes identified as being of interest in Section 3.1, are assessed. The AR4 projected changes for each of these extremes are first outlined, and then post-AR4 research is assessed to determine if any change from the AR4 assessment is justified for any of the extremes. The studies reported and assessed inevitably use a variety of different base-line and future scenario periods to calculate projected changes, together with different underlying climate model runs and emissions scenarios. Even where common data sets are used, such as the AR4 MME, different studies tend to use a different number of ensemble members. Thus care is needed in inter-comparing the magnitude of projected changes in extremes from different studies. This is not generally done here, therefore, with the focus more on the direction of change with some indication of the general magnitude of change rather than providing quantified change and ranges for all assessed studies.

The likelihood language developed for AR4 is used to describe the projected changes for each type of extreme wherever possible, i.e., "more likely than not/less likely than not", "likely/unlikely", "very likely/very unlikely" and "virtually certain/exceptionally unlikely". These terms are used both in the Sections 3.3 to 3.5 text and in the summary Tables 3.1 and 3.3. Table 3.1 provides an overview of all considered extremes (including both observed and projected changes, as well as the attribution of observed changes), while Table 3.3 focuses on projected changes in temperature and precipitation extremes. As highlighted in Section 3.2.3.3., the Tables use the term '*insufficient evidence*' where observations or the number or available projections are too limited to provide a robust assessment of projected changes, '*insufficient literature*' where there is not sufficient published literature on climate projections to make an assessment, and '*insufficient congruence*' where projections from different studies are divergent. Changes which are robust across models and studies and which are supported by an understanding of the processes are given higher confidence (see also Section 3.1.1.3). The regions included in Table 3.3 are rather fewer and in some cases sub-regions differ from those used for the observed changes in Table 3.2 since the availability of projections is generally less than for observations. Spatial scale issues and lack of literature mean that no information is provided for 'Small Islands' in either Table.

INSERT FIGURE 3.3 HERE

Figure 3.3: Regional projected changes in temperature and precipitation extremes (Americas)

INSERT FIGURE 3.4. HERE

Figure 3.4: Regional projected changes in temperature and precipitation extremes (Europe, Africa, Asia, and Oceania). See Figure 3.3. for definition of symbols.

INSERT TABLE 3.3 HERE

Table 3.3: Projected regional changes in temperature and precipitation extremes. The key for the employed abbreviations is found below the Table. Assessments for which no likelihood statements are available yet are displayed in grey in the Table (empty arrows on Figures 3.3 and 3.4).

3.3. Observed and Projected Changes of Weather and Climate

3.3.1. Temperature

Temperature is associated with several types of extremes, e.g., heat waves and cold snaps, and related impacts, e.g., on human health, ecosystems, and energy consumption (Chapter 4). Observed changes reported on in this section are based primarily on instrumental records. Temperature extremes often occur on weather timescales which require daily or higher timescale resolution data to accurately assess possible changes (Section 3.2.1). However, paleoclimatic temperature reconstructions can offer further insight to long-term changes in the occurrence of temperature extremes and their impacts. Where instrumental data is used, it is important to distinguish between mean, maximum, and minimum temperature, as well as between cold and warm extremes, due to their differing impacts. The difference between the daily maximum and minimum temperature defines the diurnal temperature range (DTR). Spell lengths (e.g., duration of heat waves) are relevant for a number of impacts.

Techniques to homogenize monthly and annual means of temperature data have a long history and have been well vetted over the past 25 years. However, homogenizing daily and even hourly temperature data has only received attention in the past decade (Section 3.2.1). Furthermore, the robustness of these methods is still an area of active research, though recent developments are showing promise. Some methods simply take adjustment factors calculated at the monthly or annual time scale and apply them to daily data (Vincent et al., 2002) while others are more sophisticated in their approach (Section 3.2.1), but there is not yet a global data set of adjusted daily temperature data as there is with monthly data. For example the changes in extreme temperature in Australia shown in Collins et al., (2000) are based on daily homogenised series of using a method which corrects series at all percentiles (Trewin and Trevitt, 1996). Della-Marta and Wanner (2006) generalized the Trewin and Trevitt (1996) method to homogenise skewness inhomogeneities and to be applicable in circumstances where overlapping daily data are not available. Della-Marta et al., (2007b) applied this method to 25 European daily temperature records which were previously unhomogenised (Wijngaard et al., 2003).

3.3.1.1. Observed Changes

The latest IPCC report (AR4) provides an extensive assessment of observed changes in temperature extremes (Trenberth et al., 2007). The following paragraphs provide a summary of the main results of this assessment. Wherever relevant, results from more recent investigations are included. We first discuss changes in mean temperatures, since changes in some temperature extremes are related to these (see also Box 3.2.). Moreover, they are also relevant for change in other extremes, such as precipitation (associated with changes in relative humidity, Section 3.3.2) or droughts (associated with changes in evaporative demand, Section 3.5.1).

Global mean surface temperatures rose by 0.74°C ± 0.18°C over the 100-year period 1906–2005. The rate of warming over the 50-year period 1956–2005 is almost double that over the last 100 years (0.13°C ± 0.03°C vs. 0.07°C ± 0.02°C per decade). Moreover, trends over land are stronger than over the oceans. For the globe as a whole, surface air temperatures over land rose at about double the ocean rate after 1979 (more than 0.27°C per decade vs. 0.13°C per decade), with the greatest warming during winter (December to February) and spring (March to May) in the Northern Hemisphere (Trenberth et al., 2007).

Regarding changes in temperature extremes on a global scale, the AR4 reports an increase in the number of warm extremes and a reduction in the number of daily cold extremes in 70 to 75% of the land regions where data are available. The most marked changes are for cold nights (below the 10th percentile threshold, based on 1961–1990), which have become rarer over the 1951 to 2003 period, whilst warm nights (above the 90th percentile threshold) have become more frequent (Trenberth et al., 2007). 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, with a trend in diurnal temperature range (DTR) of –0.07°C per decade. For 1979 to 2004, the corresponding linear trends for the land areas where data are available were 0.29°C per decade for both maximum and minimum temperature with no trend for DTR (Vose et al., 2005).

On the regional and daily time scale, the AR4 (Trenberth et al., 2007) reports a decrease in the number of very cold days and nights and an increase in the number of extremely hot days and warm nights in most regions since the 1950s. Since 1979, daily minimum temperature has increased in most areas except western Australia and southern Argentina, and parts of the western Pacific Ocean, while daily maximum temperature also increased in most regions except northern Peru, northern Argentina, northwestern Australia, and parts of the North Pacific Ocean. In southern South America, significant increasing trends were found in the frequency of occurrence of warm nights and decreasing trends in the occurrence of cold nights, but no consistent changes in the indices based on daily maximum temperature. In Central America and northern South America, high extremes of both minimum and maximum temperature have increased.

Additional regions studied since the AR4 include central and eastern Europe (Bartholy and Pongracz, 2007; Kurbis et al., 2009), the Tibetan Plateau (You et al., 2008) and China (You et al., 2010), and North America (Peterson et al., 2008a), all of which document increases in unusually warm nights and days and a reduction in unusually cold nights and days. A study for the eastern Mediterranean also reports an increase in heat wave intensity, heat wave number and heat wave length in summer (Kuglitsch et al., 2010). A study for Uruguay (Rusticucci and Renom, 2008) suggests more complex trends in this region, with a reduction of cold nights, a positive but not significant trend in warm nights, non-significant decreases in cold days at most investigated stations, and inconsistent trends in warm days. This is consistent with reported trends in southern South America (Trenberth et al., 2007).

As reported in the AR4 (Trenberth et al., 2007), Alexander et al., (2006) and Caesar et al., (2006) have brought many regional results together, gridding the common indices or data for the period since 1951. According to Alexander et al., (2006) over 70% of the global land area sampled shows a significant decrease in the annual occurrence of cold nights; a significant increase in the annual occurrence of warm nights took place over 73% of the area. This implies a positive shift in the distribution of daily minimum temperature (Tmin) throughout the globe. Changes in the occurrence of cold and warm days show warming as well, but generally less marked. This is consistent with Tmin increasing more than maximum temperature (Tmax), leading to a reduction in DTR since 1951 in many regions. More recently, Meehl et al., (2009a) found the ratio of the number of record daily maximum temperatures to record daily minimum temperatures averaged across the USA is now about 2 to 1, whereas in the 1960s the ratio was approximately 1 to 1. However, some regions experienced an increase in DTR at least in some seasons: for instance, over the 1979-2005 time period, a DTR increase was reported in Europe for the spring and summer seasons, while DTR generally decreased in autumn and winter (Klok and Klein Tank, 2009). Recently, Makowski et al., (2009) have shown that trends in DTR in Europe appear to follow closely trends in surface solar radiation (induced by changes in aerosols and cloud cover), (Wild, 2009), and display a decreasing trend over the so-called “dimming period” and positive trend over the “brightening period” in summer and autumn. Links between trends in surface solar radiation and DTR may also be enhanced by soil moisture feedbacks (Jaeger and Seneviratne, 2010).

At the time of the AR4 (Trenberth et al., 2007), only a few studies had examined changes in both the high and low tail of the same daily (minimum, maximum or mean) temperature distribution. Results suggest that these do not warm uniformly in several regions (e.g., Alexander et al., 2006). For instance, Klein Tank and Können (2003) analysed such changes over Europe using standard indices, and found that the annual number of warm extremes (days above the 90th percentile for 1961 to 1990) of the daily minimum and maximum temperature distributions increased twice as fast during the last 25 years than expected from the corresponding decrease in the number of cold extremes (days below the 10th percentile). Brunet et al., (2006) examined Spanish stations for the period 1894 to 2003 and found greater reductions in the number of cold days than increases in hot days. Since 1973, however, warm days have been rising dramatically, particularly near the Mediterranean coast. On the other hand, Griffith et al (2005) report consistent trends in the low and high tails of the temperature distributions and no significant changes in standard deviation for stations in the Asia-Pacific region in the 1961-2003 period, with the exception of urbanized locations (see also Box 3.2).

INSERT FIGURE 3.5 HERE

Figure 3.5: Annual PDFs for temperature indices for 202 global stations with at least 80% complete data between 1901-2003 for three time periods: 1901-1950 (black), 1951-1978 (blue), and 1979-2003 (red). The x-axis represents the percentage of time during the year when the indicators were below the 10th percentile for cold nights (left) and above the 90th percentile for warm nights (right). From Alexander et al., (2006).

Using the recently homogenized time series noted above, Della-Marta et al., (2007a) found that previous estimates of European summer temperature increases in mean and extreme temperatures over the period 1880-2005 are conservative (Klein Tank et al., 2002). Mean summer maximum temperature change over the region is reported to be +1.6 ± 0.4°C whereas previous estimates were around +1.3 ± 0.2°C. Similarly the frequency of hot days has almost tripled and the maximum length heat wave, defined as maximum number of consecutive days the summer daily maximum temperature is above the 95th percentile, has doubled over the 1880-2005 period. Della-Marta et al., (2007a) also showed that European daily maximum summer temperature variability has increased since 1880 by +6 ± 2% and in central western Europe +11 ± 2%. The increase in the variability of summer temperature accounts for up to 40% of the changes in hot days showing that small changes in the variance of a PDFs lead to large increases in the response of extreme events (Katz and Brown, 1992). Modelling results suggest that observed and projected changes in variability in this region could be related to changes in soil moisture and land-atmosphere-precipitation feedbacks (Seneviratne et al., 2006a; Diffenbaugh et al., 2007; Fischer et al., 2007a; see also FAQ 3.2). Kuglitsch et al., (2009; 2010) homogenised and analysed over 250 daily maximum and minimum temperature series in the Mediterranean region since 1960. They used a variety of methods to detect and correct the daily temperature series and found that after homogenisation the positive trends in the frequency of hot days and heat waves in this area are higher than previously derived. This is due to the correction of many warm biased temperature records in the region during the 1960s and 1970s.

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; Schaer and Jendritzky, 2004). That summer (June to August) was the hottest since comparable instrumental records began around 1780 (1.4°C above the previous warmest in 1807) and evidence suggests it was 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) and the 2009 heat wave in southeastern Australia. A few studies have found significant changes in heat wave occurrences. A study for the eastern Mediterranean reports an increase in heat wave intensity, heat wave number and heat wave length in summer over the 1960-2006 time period (Kuglitsch et al., 2010). Ding et al., (2009) found increasing numbers of heat waves over most of China for the 1961-2007 period and Kunkel et al., (2008) found that the USA has experienced a strong increase in heat waves since 1960, however the heat waves of the 1930s associated with the extreme drought conditions still dominate the 1895-2005 time series. 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 significant drought conditions. Drought conditions have been shown to be an important factor, potentially enhancing temperature anomalies during heat waves due to suppressed evaporative cooling (see also Section 3.3.1.2 and Box 3.4).

Regional paleoclimatic temperature reconstruction can help place the recent instrumentally observed temperature extremes in the context of a much longer period. For example Dobrovolny et al., (2010) reconstruct monthly and seasonal temperature over central Europe back to 1500 using a variety of temperature proxy records. They conclude that only two recent temperature extremes, the summer 2003 heatwave and the July heatwave of 2006 exceed the +2 standard error (associated with the reconstruction method) of previous monthly temperature extremes since 1500. Whereas the coldest periods within the last five centuries have occurred in the winter and spring of 1690.

In summary, regional and global analyses of temperature extremes nearly all show patterns consistent with a warming climate. Only a very few regions show changes in temperature extremes consistent with cooling, most notably the southeastern U.S. which has a documented decrease in mean annual temperatures over the 20th century (Trenberth et al., 2007). Regional observed changes in temperature extremes are detailed in Table 3.2. Research performed since the AR4 reinforces the conclusions that for the period since 1950 it is *very likely* that there has been a decrease in the number of both unusually cold days and nights, and an increase in the number of unusually warm days and nights on both a global and regional basis. Furthermore, based on a limited number of regional analyses and implicit from the documented changes in daily temperatures, it appears that warm spells, including heat waves defined in various ways, have *likely* increased in frequency since the middle of the 20th century in many regions.

3.3.1.2. Causes Behind the Changes

There is already an extensive body of literature on past and future changes in temperature extremes, and the underlying causes and mechanisms for these changes (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 (Durre et al., 2000; Brabson et al., 2005; Seneviratne et al., 2006a; Diffenbaugh et al., 2007; Fischer et al., 2007a; Vautard et al., 2007), whereby the latter can act as an amplifying mechanism through impacts on evaporative cooling (e.g., Jaeger and Seneviratne, 2010) but also induce enhanced persistence due to soil moisture memory (Lorenz et al., 2010). These latter effects are mostly relevant in transitional climate regions between dry and wet climates (Koster et al., 2004b; Seneviratne et al., 2010). When considering impacts of heat waves, e.g., on human health, changes in other climate variables such as relative humidity (e.g., Diffenbaugh et al., 2007; Fischer and Schär, 2010) are also of relevance.

Compared with studies on mean temperature, studies of the attribution of extreme temperature changes are limited. Regarding possible human influences on these changes in temperature extremes, the AR4 (Hegerl et al., 2007) concludes that surface temperature extremes have *likely* been affected by anthropogenic forcing. This assessment is based on multiple lines of evidence of temperature extremes at the global scale including an increase in the number of warm extremes, and a reduction in the number of cold extremes. There is also evidence that anthropogenic forcing may have significantly increased the likelihood of regional heat waves (Alexander et al., 2006).

Post-AR4 studies tend to confirm the assessment of Hegerl et al., (2007). For example, Shiogama et al., (2006) used an optimal detection method to compare changes in daily extreme temperatures including annual maximum daily maximum and daily minimum temperatures and annual minimum daily maximum and daily minimum temperatures from observations with those simulated by a GCM at the global scale. They found evidence of anthropogenic warming in the annual warmest night, and the coldest day and night from 1950-1990 over the globe.

Detection studies of external influences on extreme temperature changes at regional scale are also very limited. Regional trends in temperature extremes could be related to regional processes and forcings that have been a challenge for climate model simulations. For example, Portmann et al., (2009) demonstrated that the rate of increase in the number of hot days per year in late spring in the southeastern U.S. over recent decades has a statistically significant inverse relationship to climatological precipitation. They speculate that changes in biogenic aerosols resulting from land use changes could be responsible. However, anthropogenic influence has been detected in temperature extremes in some regions. Meehl et al., (2007b) showed that most of the observed changes in temperature extremes for the second half of the 20th century over the U.S. are due 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 of frost days, an increase in growing season length, and an increase in heat wave intensity all show similar changes over the U.S. 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., (2010) 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 (ANT) forcing or anthropogenic and natural external forcings combined (ALL) by multiple GCMs. They fitted generalized extreme value (GEV) distributions to the observed extreme temperatures with a time-evolving pattern of location parameters as obtained from the model simulation, and found that both ANT and ALL influence can be detected in all the extreme temperature variables at the global scale over the land, and also regionally over many large land areas. They concluded that the influence of anthropogenic forcing has had a detectable influence on extreme temperatures that have impacts on human society and natural systems at global and regional scales. External influence is estimated to have resulted in large changes in the likelihood of extreme annual maximum and minimum daily temperatures. Globally, waiting times for events that were expected to recurred 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, and to have decreased to less than 10 or 15 years for annual maximum daily minimum and daily maximum temperatures respectively (Figure 3.6).

INSERT FIGURE 3.6 HERE

Figure 3.6: Estimated waiting time (years) and their 5% and 95% uncertainty limits for 1960s 20-yr return values of annual extreme daily temperatures in the 1990s climate (see text for more details). From Zwiers et al., (2010). Red, green, blue, pink error bars are for annual minimum daily minimum temperature (TNn), annual maximum daily minimum temperature (TNx), annual minimum daily maximum temperature (TXn), and annual maximum daily maximum temperature (TXx), respectively. Grey areas indicate insufficient data.

The new studies that attribute observed changes in temperature extremes at global and continental and sometimes regional scales to external forcing add support to the AR4 assessment that surface temperature extremes are *likely* affected by anthropogenic forcing.

3.3.1.3. Projected Changes and Uncertainties

Regarding projections of extreme temperatures, the AR4 (Meehl et al., 2007b) states that is *very likely* that heat waves will be more intense, more frequent and longer lasting in a future warmer climate (Figure 3.7). Cold episodes are projected to decrease significantly in a future warmer climate. Almost everywhere, daily minimum temperatures are projected to increase faster than daily maximum temperatures, leading to a decrease in diurnal temperature range. Decreases in frost days are projected to occur almost everywhere in the middle and high latitudes. Regional projected changes in temperature extremes are detailed in Table 3.3.

The AR4 (Meehl et al., 2007b) reports several studies explicitly addressing possible future changes in heat waves (using a number of different definitions), which found an increased risk of more intense, longer-lasting and more frequent heat waves in a future climate (Meehl and Tebaldi, 2004; Schär et al., 2004; Clark et al., 2006). A multi-model ensemble simulated the observed increase in heat waves over the latter part of the 20th century, and heat waves were projected to increase globally and over most regions (Tebaldi et al., 2006), although different model parameters influenced the magnitude of this projection (Clark et al., 2006). Meehl and Tebaldi (2004) showed that the pattern of future changes in heat waves, with greatest intensity increases over western Europe, the Mediterranean and the southeast and western USA, is related in part to circulation changes due to the increase in greenhouse gases. An additional factor leading to extreme heat is drier soils in a future warmer climate (Section 3.3.1.2), mostly in regions shifting from a wet to transitional climate regime (Seneviratne et al., 2006a). Schär et al., (2004), Stott et al., (2004) and Beniston (2004) used the European 2003 heat wave as an example of the types of heat waves that are likely to become more common in a future warmer climate (i.e., as a temporal analogue). There is also evidence that human influence has at least doubled the risk of the European exceptionally warm summer of 2003 and it is possible that by the 2040s, summers over southern Europe will be as warm or warmer 50% of the time (Jones et al., 2008).

A projected decrease in diurnal temperature range (DTR) in most regions in a future warmer climate was noted in the AR4 (e.g., Stone et al., 2001). The AR4 also reported on possible future cold air outbreaks. Vavrus et al., (2006) analysed seven AOGCMs run with the A1B scenario, and define a cold air outbreak as two or more consecutive days when daily temperatures are at least two standard deviations below the present-day winter mean. They found a 50 to 100% decline in the frequency of cold air outbreaks in Northern Hemisphere winter in most areas compared to the present, with the smallest reductions occurring in western North America, the North Atlantic and southern Europe and Asia, due to atmospheric circulation changes associated with the increase in greenhouse gases.

Post-AR4 studies of temperature extremes have utilised larger model ensembles (Kharin et al., 2007; Sterl et al., 2008) and generally reinforce the conclusions of AR4 as well as providing more regional detail. The U.S. Climate Change Science Program (CCSP) assessed changes in extremes over northern America. They reached the following conclusions regarding projected changes in temperature extremes (Gutowski et al., 2008a):

- Abnormally hot days and nights and heat waves are very likely to become more frequent.
- Cold days and cold nights are very likely to become much less frequent.
- Climate models indicate that currently rare extreme warm events will become more commonplace. For example, for a mid-range scenario of future greenhouse gas emissions, a day so hot that it is currently experienced only once every 20 years would occur every three years by the middle of the century over much of the continental U.S. and every five years over most of Canada. By the end of the century, it would occur every other year or more.

Analysis of the AR4 MME for Australia also indicates 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°C or 40°C are indicated (Suppiah et al., 2007). A study with a single RCM projects more frequent warm nights in the entire tropical South American region and fewer cold nights (Marengo et al., 2009a).

Analyses of both global and regional model outputs show major increases in warm temperature extremes across the Mediterranean including events such as hot days (Tmax >30°C) and tropical nights (Tmax >20°C) (Giannakopoulos et al., 2009; Tolika et al., 2009). Comparison of RCM projections 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 PDF for 2007 lies entirely within the PDF for 2071–2100 - thus 2007 might be considered a 'normal' summer of the future (Founda and Giannakopoulos, 2009; Tolika et al., 2009). This is consistent with earlier analyses of the 2003 European hot summer (see above). In contrast to this 'temporal analogue' approach, Beniston et al., (2007) take a 'spatial analogue' approach, concluding from an analysis of RCM output that regions such as France and Hungary, for example, may experience as many days per year above 30°C as currently experienced in Spain and Sicily. In this RCM ensemble, France is the area with the largest warming in the uppermost percentiles of daily summer temperatures although the mean warming is greatest in the Mediterranean (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).

Temperature extremes were the type of extremes projected to change with most confidence in the AR4 (IPCC, 2007a). If changes in temperature extremes scale with changes in mean temperature (i.e., simple shifts of the PDF), we can infer that it is *virtually certain* that hot (cold) extremes will increase (decrease) in the coming decades (if these extremes are defined 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 (Box 3.2. and hereafter), though in most such reported cases hot extremes tend to increase more than mean temperature, and thus the above statement for hot extremes (*virtually certain* increase) still applies. Central and eastern Europe is a region for which it is now established that projected changes in temperature extremes result from both changes in the mean as well as by changes in the shape of the PDFs (Schär et al., 2004). The main mechanism for the widening of the distribution is linked to the drying of the soil in this region (Seneviratne et al., 2006a, see also Section 3.3.1.2). The role of land-atmosphere interactions for projected changes in temperature distribution functions, in particular through feedbacks with soil moisture or snow content, is also discussed in other studies (Brabson et al., 2005; Kharin and Zwiers, 2005; Clarke and Rendell, 2006; Jaeger and Seneviratne, 2010).

Furthermore, remote surface heating may induce circulation changes that modify the temperature distribution (Haarsma et al., 2009).

INSERT FIGURE 3.7 HERE

Figure 3.7: (a) Globally averaged changes in heat waves (defined as the longest period in the year of at least five consecutive days with maximum temperature at least 5°C higher than the 1961-1990 climatology of the same calendar day) based on multi-model simulations from nine global coupled climate models, adapted from Tebaldi et al., (2006). (b) Changes in spatial patterns of simulated heat waves between two 20-year means (2080–2099 minus 1980–1999) for the A1B scenario. Solid lines in (a) show the 10-year smoothed multi-model ensemble means; the envelope indicates the ensemble mean standard deviation. Stippling in (b) denotes areas where at least five of the nine models concur in determining that the change is statistically significant. Extreme indices are calculated only over land. Extremes indices are calculated following Frich et al., (2002). Each model's time series is centred around its 1980 to 1999 average and normalised (rescaled) by its standard deviation computed (after de-trending) over the period 1960 to 2099. The models are then aggregated into an ensemble average, both at the global and at the grid-box level. Thus, changes are given in units of standard deviations. From Meehl et al., (2007b).

Local, mesoscale and regional feedback mechanisms, in particular with land surface conditions (e.g., soil moisture, vegetation, snow) may significantly impact projections of temperature extremes at the local and regional scale, and induce uncertainties in these projections. Indeed, these processes occur on a small scale, not resolved by the models. In addition, lack of observational data (e.g., for soil moisture and snow cover, see Section 3.2.1) implies additional uncertainties induced by the lack of evidence to constrain current climate models (e.g., Roesch, 2006; Boe and Terray, 2008; Hall et al., 2008; Brown and Mote, 2009). Regarding mesoscale processes, lack of information may also affect confidence in projections. One example is changes in Mediterranean heat waves which are suggested to have the largest impact in coastal areas, due to the role of enhanced relative humidity for 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).

Ganguly et al., (2009) statistically validated global warming trends across ensembles and at regional scales and found that observed heat wave intensities in the current decade are larger than worst-case projections. They also showed that model projections are relatively insensitive to initial conditions, while uncertainty bounds obtained by comparison with recent observations are wider than ensemble ranges.

In summary, climate change projections suggest a *virtual certain* increase in hot extremes and *virtual certain* decrease in cold extremes in most regions. This is mostly linked with mean changes in temperatures, although changes in temperature variability can play an important role in some regions.

3.3.2. Precipitation

Because climates are so diverse across different parts of the world, it is difficult to provide a single definition of "extreme precipitation". In general, three different methods have been used to define extreme precipitation (see also Sections 3.1.3 and 3.2.1), either based on 1) relative thresholds, i.e, percentiles, 2) absolute thresholds, or 3) return values. As an example of the first case, a daily precipitation event with an amount greater than the 95th percentile of daily precipitation for all wet days within a 30-year period can be considered as extreme. Regarding the second type of definition, a precipitation amount that exceeds predetermined thresholds above which damage may occur can also be considered as an extreme. For example, 2 inches/day of rain in the U.S., and 50mm/day or 100mm/day of rain in China have been considered as extremes. A drawback of this definition is that such an event may not occur everywhere, and the damage for the same amount of rain in different regions may be quite different depending partly on climatology. The third type of definition is common in engineering practice: engineers often use return values associated with a pre-determined level of probability for exceedance as design values, estimated from annual maximum one day or multi-day precipitation amounts over many years. Return values, similarly to relative thresholds, are defined for a given time period and region and may change over time.

The occurrence of hail associated with severe thunderstorms represents a significant hazard and can cause serious damage to automobiles, houses, and crops, and is therefore here considered separately to other extreme precipitation types. Hail occurs in most mid-latitude regions and is common in India, China, North America, and Europe. Hail damage accounts for 50% of the 20 highest insurance payouts in Australia (Hennessy et al., 2007). However, increases in public awareness and changes in reporting practices lead 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).

Climate models are important tools for understanding past changes in precipitation and projecting future changes. However, for various reasons, the direct use of output from climate models is often inadequate for studies of attribution changes of precipitation in general and of precipitation extremes in particular. The most important reason is related to the fact that precipitation is often localized, with very high variability across space, and extreme precipitation events are often of very short duration. The spatial and temporal resolutions of climate models are not fine enough to well represent processes and phenomena that are relevant to precipitation, such as mesoscale atmospheric processes, topography, and land-sea distribution. Simulating extremes may be more challenging for shorter time periods because the relevant processes are likely to occur on smaller scales and thus be less well resolved. Thus, the time scale of extremes that a model can simulate well is probably tied to its spatial resolution (e.g., Gutowski et al., 2003). Some of these modelling shortcomings can be partly addressed with (dynamical and/or statistical) downscaling approaches (Section 3.2.3). In addition, in some parts of the world, precipitation extremes are poorly monitored by very sparse network systems (Section 3.2.1), resulting in high uncertainty in precipitation estimates, especially for extreme precipitation that is localized in space and of short duration, and thus limited possibilities to thoroughly validate modelling and downscaling approaches.

3.3.2.1. Observed Changes

Recent studies on past and current changes of heavy precipitation extremes in North America, some of which are included in the recent assessment of the U.S. Climate Change Science Program (CCSP) report (Kunkel et al., 2008), indicate an increasing trend over the last half century. Based on station data from Canada, the U.S., and Mexico, Peterson et al., (2008a) suggest 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 U.S., DeGaetano (2009) shows 20% reduction in the return period for extreme precipitation of different return levels over 1950–2007; Gleason et al., (2008) find an increasing trend in the area experiencing a much above-normal proportion of heavy daily precipitation from 1950 to 2006; Pryor et al., (2009) show 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 towards increased annual total precipitation, number of rainy days and intense precipitation (e.g., fraction of precipitation derived from events in excess of the 90th percentile value) are focused on the central plains/northwestern Midwest (Pryor et al., 2009). In the core of the North American monsoon region in northwest Mexico, 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. These precipitation events appear to have been derived from tropical cyclones. However, no significant changes were found in coastal stations (Cavazos et al., 2008).

Positive trends in extreme rainfall events are evident in southeastern South America, north central Argentina, northwest Peru and Ecuador (Marengo et al., 2009b; Re and Ricardo Barros, 2009). In the State of Sao Paulo, Brazil, there is evidence for an increase in magnitude and frequency of extreme precipitation events over 1950-1999 (Dufek and Ambrizzi, 2008) and 1933-2005 (Sugahara et al., 2009). Penalba and Robeldo (2010) report increases in the annual frequencies in spatially coherent areas over the La Plata Basin for both heavy and all (>0.1mm) precipitation events during summer, autumn and spring of 1950–2000. Winter is the exception, with negative trends, some of which are significant in the lower and middle Uruguay and Paraná Rivers.

A number of recent regional studies have been completed for European countries (Moberg et al., 2006; Bartholy and Pongracz, 2007; Maraun et al., 2008; Pavan et al., 2008; Zolina et al., 2008; Costa and Soares, 2009; Durão et al., 2009; Kysely, 2009; Rodda et al., 2009). According to Moberg et al., (2006), averaged over 121 European stations north of 40°N, trends in 90th, 95th and 98th percentiles of daily winter precipitation, as well as winter precipitation totals, have increased significantly over the 1901–2000 period. No overall long-term trend was observed in summer precipitation totals. For the United Kingdom (UK), Maraun et al., (2008) show widespread shifts towards greater contribution from heavier precipitation categories in winter, spring and (to a lesser extent) autumn, and towards light and moderate categories in summer during 1961–2006. Extreme rainfalls during 1961–2006 have increased up to 20% relative to the 1911–1960 period in the north-west of the UK and in parts of East Anglia, although there have also been changes in other areas, including decreases of the same magnitude over central England (Rodda et al., 2009). Zolina et al., (2008) present similar seasonally dependent changes of precipitation extremes over Germany during 1950–2004. Bartholy and Pongracz (2007) identify increases in the intensity and frequency of extreme precipitation in the Carpathian basin on an annual basis. Kysely (2009) identify spatially coherent increasing trends for heavy precipitation intensity in winter in the western region of the Czech Republic during 1961–2005. Increasing but insignificant and spatially less coherent trends in heavy precipitation prevail also in summer. Opposite trends occur in spring and the changes are spatially least coherent and insignificant in autumn. In contrast, at Emilia-Romagna, a region of northern Italy, the frequency of intense to extreme events decreases during winter, but increases during summer over the central mountains, while the number of rainy days decreases in summer during 1951–2004 (Pavan et al., 2008). In southern Portugal, spatial coherence of extreme precipitation events has increased and spatial variability decreased during 1961–2000 (Durão et al., 2009); short-term precipitation intensity tend to increase over the region, although the trend signals of the four wetness indices are insignificant at the majority of stations during the last three decades of the twentieth century (Costa and Soares, 2009).

Several recent studies have focused on Africa and, in general, have not found significant trends in extreme precipitation (Kruger, 2006; New et al., 2006; Seleshi and Camberlin, 2006; Aguilar et al., 2009). There has been no real evidence of changes in precipitation in most of south and west Africa over 1910 to 2004 (Kruger, 2006) and 1961 to 2000 (New et al., 2006). Central Africa showed a decrease in heavy precipitation over the last half century (Aguilar et al., 2009). However, data coverage for large parts of central Africa was poor. There were decreasing trends in heavy precipitation over eastern, southwestern and southern parts of Ethiopia, whereas no trends are found in the remaining part of Ethiopia over the period 1965-2002 (Seleshi and Camberlin, 2006).

Observations at 143 weather stations in ten Asia-Pacific Network countries during 1955–2007 did not indicate systematic, regional trends in the frequency and duration of extreme precipitation events (Choi et al., 2009). However, other studies have suggested significant trends in extreme precipitation at sub-regional scales in the Asia-Pacific region. Significant rising trends in extreme rainfall over the Indian region have been noted (Rajeevan et al., 2008; Krishnamurthy et al., 2009), especially during the monsoon seasons (Pattanaik and Rajeevan, 2009; Sen Roy, 2009). Zhai et al., (2005) found significant increases over the period 1951-2000 in extreme precipitation in western China, in the mid–lower reaches of the Yangtze River, and in parts of the southwest and south China coastal area, but a significant decrease in extremes is observed in north China and the Sichuan Basin. For most precipitation indices, such as heavy precipitation days and maximum one–day precipitation, You et al., (2008) observe increasing trends in the southern and northern Tibetan Plateau and decreasing trends in the central Tibetan Plateau during 1961–2005. However, the precipitation indices show insignificant increases on average. Bhutiyani et al., (2010) indicate no trend in the winter precipitation but significant decreasing trend in the monsoon precipitation in the northwestern Himalaya during 1866-2006. During the summer of 1978–2002, positive trends for heavy (25-50 mm per day) and extreme (>50mm per day) precipitation near the east coasts of east Asia and southeast Asia are observed, while negative trends are seen over southwest Asia, central China, and northeast Asia (Yao et al., 2008). Summer extreme precipitation over south China increased significantly since the early 1990s (Ning and Qian, 2009). In Peninsular Malaysia during 1971–2005 intensity of extreme precipitation increased and frequency decreased, while the trend detected for the proportion of extreme rainfall over total rainfall amount was insignificant (Zin et al., 2009). Only a few recent studies have been completed for Australia (Aryal et al., 2009). 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.

There have been few studies of recent trends of hailstorm frequency. Changes in hail occurrence are generally considered either directly, through analysis of actual hail measurements, or indirectly, through analysis of environmental conditions associated with hail events. The environmental conditions are typically taken from reanalysis data. Both approaches have their associated caveats; data homogeneity issues pose challenges in identifying trends in spatially and temporally rare events such as hail storms, while reanalysis data is based partly on models whose physical approximations may not be optimal for simulating conditions conducive for hail production. Kunz et al., (2009) find that hail days significantly increased during 1974-2003 in a state in southwest Germany. Cao (2008) suggests an increasing frequency of severe hail events in Ontario, Canada during 1979-2002. Xie et al., (2008) find no trend in the mean annual hail days in China from 1960 to early 1980s but a significant decreasing trend afterwards, with mostly flat or decreasing trends in mean annual hail days over their entire record length of 46 years. Brooks and Dotzek (2008) used environmental conditions derived from reanalysis data to count the frequency of favorable environments for significant severe thunderstorms in the region east of the Rocky Mountains in the U.S., and found significant variability but no clear trend in the past 50 years. Cao (2008) analyzed direct measurements of hail frequency over Ontario, Canada, and identified a robust upward trend in association with changes in atmospheric changes in convective instability and available precipitable water.

In summary, while many studies conducted since the AR4 confirm its conclusion that, despite spatial and seasonal variations in the changes of precipitation extremes, increasing trends in precipitation extremes are observed in many areas over the world (e.g., Trenberth et al., 2007), some studies have found a decreasing trend and/or no clear change in precipitation extremes over Africa (e.g., Aguilar et al., 2009) and the Asia-Pacific (Choi et al., 2009). Regional observed changes in precipitation extremes are detailed in Table 3.2 and a geographical overview is provided in Figures 3.1 and 3.2.

3.3.2.2. Causes Behind these Changes

As atmospheric moisture content increases with increases in global mean temperature, extreme precipitation is expected to increase as well and at a rate faster than changes in mean precipitation content (Allen and Ingram, 2002). In some regions, extreme precipitation is projected to increase, even if mean precipitation is projected to decrease (Christensen and Christensen, 2003; Kharin et al., 2007, see also Box 3.3). The observed change in heavy precipitation appears to be consistent with the expected response to anthropogenic forcing 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 concludes that it is *more likely than not* that anthropogenic influence has 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 3.2.2.2), 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 3.2.2.2). Additionally, one observational study also suggests a strong influence of moisture on short duration extreme precipitation. Wang and Zhang (2008) show that winter season maximum daily precipitation in North America appears to be significantly influenced by atmospheric moisture content, with an increase in moisture corresponding to an increase in maximum daily precipitation. This behaviour has also been seen in model projections of extreme winter precipitation under global warming (Gutowski et al., 2008b). The thermodynamic constraint based on the Clausius-Clapeyron relation is a good predictor for extreme precipitation changes in a warmer world at regions where the nature of the ambient flows change little (Pall et al., 2007). This may support the judgment that the observed increase in extreme precipitation may, in part, be attributable to anthropogenic influence. However, the thermodynamic constraint may not be a good predictor in regions with circulation changes such as mid- to higher-latitudes where advective effects associated with changes in atmospheric circulation produce changes in mean and extreme precipitation (Meehl et al., 2005), and in the tropics (Emori and Brown, 2005) if the thermal equator shifts northward with a relatively large increase in precipitation at 0–20°N, with a concurrent decrease at 0–20°S (Pall et al., 2007). Additionally, changes of 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 be part of the reason why observations do not show increases in precipitation extremes everywhere, although a low signal to noise ratio may also play a role. However, even in regions where the Clausius-Clapeyron constraint is not closely followed, it is still appears to be a better predictor for future changes in extreme precipitation than the change in mean precipitation (Pall et al., 2007). An observational study seems also to support this thermodynamical theory. Analysis of daily precipitation from the Special Sensor Microwave Imager (SSM/I) over the tropical oceans shows a direct link between rainfall extremes and temperature: heavy rainfall events increase during warm periods (El Niño) and decreases 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 also suggest that hourly precipitation extremes may in some cases increase more strongly with temperature (twice as fast) than would be assumed from the Clausius-Clapeyron relationship alone (Lenderink and Van Meijgaard, 2008).

Perfect model studies (e.g., Min et al., 2009) indicate that changes in precipitation extremes should be detectable at least on large scales. However, a quantitative comparison between model-simulated extreme precipitation and in-situ observations is more difficult because of a low signal to noise ratio and high uncertainty in both observed and model simulated extreme precipitation. There is a mismatch between spatial scales represented by area-mean extremes simulated by climate models and point estimations from station observations (Osborn and Hulme, 1997; Kharin and Zwiers, 2005). Because the number of observation stations is limited (Section 3.2.1), it is also not possible to produce reliable area estimates of daily precipitation based on station observations. The fact that most current GCMs do not simulate smaller-scale (<100 km) variations in precipitation intensity (e.g., Sections 3.1.1.2 and 3.2.3) associated with convective storms, also complicates the problem. It may still be a decade away before the influence of external forcing on daily extreme precipitation at regional scales can be detected (Fowler and Wilby, 2010). However, this conclusion may be seasonally dependent. For example, by now there is about a 50% chance of detecting anthropogenic influence on UK extreme precipitation in winter, but the likelihood of the detection in other seasons is very small (Fowler and Wilby, 2010).

Overall, new studies since AR4 have provided further evidence to support the AR4 assessment that it is *more likely than not* that anthropogenic influence has contributed to a trend towards increases in the frequency of heavy precipitation events over the 2nd half of the 20th century in many regions. However, there is still not enough evidence to make a more confident assessment regarding the causes of observed changes in extreme precipitation than that provided in the AR4 report. There is almost no literature on the attribution of changes in hail extremes, and thus no assessment can be provided for these at this point in time.

3.3.2.3. Projected Changes and Uncertainties

Regarding projected changes in extreme precipitation, the AR4 concluded that it is *very likely* that heavy precipitation events, i.e., the frequency (or proportion of total rainfall from heavy falls) of heavy precipitation, will increase over most areas of the globe in the 21st century (IPCC, 2007a) – see Figure 3.8. The tendency for an increase in heavy daily precipitation events in many regions was found to include some regions in which the mean precipitation is projected to decrease (see also Section 3.3.2.2 and Box 3.2). Post-AR4 analyses of climate model simulations generally confirm this assessment although uncertainties and model biases remain greater for precipitation than for temperature (e.g., Hawkins and Sutton, 2010, see also Section 3.2.3 and Box 3.3). More GCM and RCM ensembles have now been analysed for some regions, leading to increased robustness of the projected changes.

Kharin et al., (2007) analyzed changes in annual maxima of 24-hour precipitation in the AR4 MME. Between the time periods 2046–2065 and 1981–2000, the median MME response in extreme precipitation shows increases in the tropics and in mid- and high latitudes, and decreases in small regions in the subtropics. Decreases in extreme precipitation occur over much smaller regions compared to those for mean precipitation, and are generally not statistically significant. There are extensive subtropical areas where the models project an increase in the intensity of precipitation extremes, even though mean precipitation decreases. Except for a few small subtropical regions where the amplitude of extreme precipitation decreases, 20-year return period values for late-twentieth-century extreme precipitation events are reduced almost everywhere over the globe. Roughly speaking, the return times are reduced by a factor of two with a 10% increase in the amplitude of the 20-year return value (Figure 3.9). Return times decrease almost everywhere over landmasses, except for north Africa where they tend to increase. The greatest reductions in waiting time occur in tropical regions and high latitudes.

INSERT FIGURE 3.8 HERE

Figure 3.8: Changes in extremes based on multi-model simulations from nine global coupled climate models, adapted from Tebaldi et al., (2006). (a) Globally averaged changes in precipitation intensity (defined as the annual total precipitation divided by the number of wet days) for a low (SRES B1), middle (SRES A1B) and high (SRES A2) scenario. (b) Changes in spatial patterns of simulated precipitation intensity between two 20-year means (2080–2099 minus 1980–1999) for the A1B scenario. Solid lines in (a) are the 10-year smoothed multi-model ensemble means; the envelope indicates the ensemble mean standard deviation. Stippling in (b) denotes areas where at least five of the nine models concur in determining that the change is statistically significant. Extreme indices are calculated only over land following Frich et al., (2002). Each model's time series was centred on its 1980 to 1999 average and normalised (rescaled) by its standard deviation computed (after de-trending) over the period 1960 to 2099. The models were then aggregated into an ensemble average, both at the global and at the grid-box level. Thus, changes are given in units of standard deviations. From Meehl et al., (2007b).

INSERT FIGURE 3.9 HERE

Figure 3.9: Projected waiting times for late-twentieth-century 20-year return values of annual maximum 24-hour precipitation rates in the mid-21st century (left) and in late-21st century (right) by 14 GCMs that contributed to the IPCC AR4, under three different emission scenarios SRES B1, A1B and A2 (adapted from Kharin et al., 2007). The vertical extent of the whiskers in both directions describes the range of projected changes by all 14 climate models used in the study. The boxes indicate the central 50% of model projected changes, and the horizontal bar in the middle of the box indicates the median projection amongst the 14 models (that is, 7 models project waiting times longer than the median and 7 models project waiting times shorter than the median). Although the uncertainty range of the projected change in extreme precipitation is large, almost all models suggest that the waiting time for a late 20th century 20-year extreme 24-hour precipitation event will be reduced to substantially less than 20 years by mid-21st and much more by late-21st century, indicating an increase in frequency of the extreme precipitation at continental and sub-continental scales under all three forcing scenarios. Three global domains are: the entire globe (GLB), the global land areas (LND), the global ocean areas (OCN). Five zonal bands are: Northern Hemisphere Extratropics (NHE, 35°-90°N), Southern Hemisphere Extratropics (SHE, 35°-90°S), Tropics (TRO, 10°S-10°N), Northern subtropics (NTR, 10°-35°N), and Southern subtropics (35°-10°S). The nine continental/sub-continental land-only regions are: Africa (AFR, 20°W-60°E and 40°S-30°N), Central Asia (ASI, 45°-180°E and 30°-65°N), Australia (AUS, 105°E–180° and 45°–10°S), Europe (EUR, 20°W–45°E and 30°–65°N), North America (NAM, 165°–30°W and 25°–65°N), South America (SAM, 115°–30°W and 55°S–25°N), South Asia (SAS, 60°–160°E and 10°S–30°N), Arctic (ARC, 180° to 180° and 65°–90°N), Antarctica (ANT, 180° to 180° and 90°–65°S).

Several other post-AR4 studies use different collections of models, or different analysis techniques, or focus on different regions, but in general they confirm the findings of Kharin et al., (2007). Regional projected changes in precipitation extremes are detailed in Table 3.3 and on Figures 3.3 and 3.4. Shongwe et al., (2009a, b) analyzed change in mean and extreme precipitation in southern Africa and east Africa as simulated by twelve GCMs. Unlike Kharin et al., (2007) where every model is treated as equally credible, they assign different weights to each model according to model performance in simulating observed precipitation change. They project an increase in intensity in both heavy rainfall events and in mean precipitation rates and less severe droughts in east Africa, more severe precipitation deficits in the southwest of southern Africa and enhanced precipitation farther north in Zambia, Malawi, and northern Mozambique. Rocha et al., (2008) evaluated differences in the precipitation regime over southeastern Africa simulated by two GCMs. The intensity of all episode categories of precipitation events is increased during 2071–2100 relative to

1961–1990 practically over the whole region, whereas the number of episodes is decreased in most of the region and for most episode categories. By analyzing simulations with a single GCM, Khon et al., (2007) find a general increase in extreme precipitation for the different regions in northern Eurasia especially for winter. Su et al., (2009) find that for the Yangtze River Basin region in 2001–2050, the 50-year heavy precipitation and drought events during 1951–2000 become more frequent, with return periods falling to below 25 years. Extreme precipitation is projected to increase over Australia in 2080–2099 relative to 1980–1999 as indicated by analysis of the AR4 MME. However, there is very little model agreement between the AR4 MME that this change is significant (Alexander and Arblaster, 2009). For the Indian region, the Hadley Centre coupled model HadCM3 projects increases in the magnitude of the heaviest rainfall with CO_2 doubling (Turner and Slingo, 2009).

Future changes in extreme precipitation indices were projected with the high-resolution Meteorological Research Institute and Japan Meteorological Agency 20-km horizontal grid AGCM by Kamiguchi et al., (2006). At the end of the 21st century, heavy precipitation was projected to increase substantially in south Asia, the Amazon, and west Africa, with increased dry spell persistence in South Africa, south Australia, and the Amazon. 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.

High-spatial resolution is important for studies of extreme precipitation, particularly in regions of complex orography. Many post-AR4 studies have employed statistical and dynamical downscaling (Section 3.2.3) to project precipitation extremes. Wang and Zhang (2008) investigated possible changes in North American extreme precipitation probability during winter from 1949–1999 to 2050–2099, using statistical downscaling. Downscaled results suggest a strong increase in extreme precipitation over the south and central U.S. but decreases over the Canadian prairies. This spatial pattern is similar to that of the underlying GCM simulations, with more-detailed structure and much smaller amplitude over regions where the downscaling procedure has skill. These differences are perhaps due to a spatial-scale mismatch between the statistical downscaling results and those estimated using the GCM simulations. Results from the downscaling procedure represent small scales corresponding to station locations, while those from model simulations represent areas of tens of thousands of square kilometres (Section 3.2.3).

Many post-AR4 studies have employed the dynamical downscaling approach to investigate future changes in climate extremes using RCMs, sometimes combined with statistical downscaling. These RCM-based results are broadly consistent with those obtained from GCM simulations, although RCM studies generally present more detailed information (the added-value of which needs to be assessed – Section 3.2.3). Projected European precipitation extremes tend to increase in northern Europe (Frei et al., 2006; Beniston et al., 2007; Schmidli et al., 2007), especially during winter (Haugen and Iverson, 2008; May, 2008), and decrease in southern Europe (Beniston et al., 2007). Fowler and Ekström (2009) project increases in both short-duration (1-day) and longer-duration (10-day) precipitation extremes across the UK during winter, spring and autumn. In summer, model projections for the UK span the zero change line, although there is low confidence due to poor model performance in this season (see Section 3.2.3). Using daily statistics from various models Boberg et al., (2009a, b) report 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 sub-regions.

In double-nested model simulations with a horizontal grid spacing of 10 km, Tomassini and Jacob (2009) find positive trends over Germany (as in the observations), although they are relatively small compared with the uncertainties except for the higher emissions A2 scenario. For the Upper Mississippi River Basin region during October–March, the intensity of extreme precipitation is projected to increase (Gutowski et al., 2008b). Simulations with a single RCM indicate 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 Western Port region of Australia, consistent with results based on the AR4 MME (Alexander and Arblaster, 2009), and the size of this increase is greater in 2070 than in 2030 (Abbs and Rafter, 2008). Tropical and northern Africa are projected to suffer less severe rainfall events by 2025 during most seasons except for autumn when both future land use changes and increasing greenhouse-gas concentrations are considered in the simulations (Paeth and Thamm, 2007).

Kysely and Beranova (2009) examined scenarios of change in extreme precipitation events in 24 future climate runs of 10 RCMs, focusing on a specific area of central Europe with complex orography. They show 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. Frei et al., (2006) analyzed the simulations of 6 European RCMs and they found that RCMs are capable of representing mesoscale spatial patterns in precipitation extremes, which are not resolved by current GCMs. However, large differences in summer extreme events are found when RCM formulation contributes significantly to projection uncertainty. Déqué et al., (2007) explored the uncertainty sources in seasonal precipitation projections from 10 RCM over Europe (Christensen et al., 2002). They found that the GCM-associated uncertainties are generally larger than those associated with the RCM simulations, but that the choice of the RCM is a source of uncertainty for summer precipitation projections, which can add the same uncertainty level as the choice of the GCM. However in this study, only two different AOGCMS were involved, and one of these was actually a relatively high resolution time slice (HadAM3). The conclusion by Déqué et al., (2007) is probably affected by these limitations. The downscaling of the RCMs off the time slice were going from a 100 km resolution to a 50 km resolution, which is a fairly small step. This may be one of the reasons why the RCM choice has such a small effect. Using the results from the same multimodel ensemble to assess changes to precipitation extremes over Europe by 2070–2100, Fowler et al., (2007b) found that the magnitude of change is strongly influenced by the driving GCM but moderated by the RCM, which also influences spatial pattern. May (2008) and Frei et al., (2006) described a significant overestimate in projected changes of precipitation over the Baltic Sea in models, related to unrealistic increase in summer SSTs in this area. Kendon et al., (2009) estimated the confidence in projected changes in daily precipitation across Europe using the Hadley Centre HadAM3P model. They found that 'other large scale changes' play a minor role in driving projected precipitation changes over much of Europe in winter and southern Europe in summer, at least on large spatial scales, allowing them to make confident statements about future changes.

Schmidli et al., (2007) compared 6 statistical downscaling models (SDMs) and 3 RCMs in their ability to downscale daily precipitation statistics in a region of complex topography (European Alps). In winter, over complex terrain, the better RCMs achieve significantly higher skills than the SDMs, while over flat terrain and in summer, the differences are small. Overall, downscaling does significantly contribute to the uncertainty in regional climate projections especially in summer because of stochastic processes appearing at the mesoscale and of the stronger role of local feedbacks (land surface processes, convection) during that season (Section 3.1.5). In exploring the ability of 2 SDMs in reproducing the direction of the projected changes in indices of precipitation extremes, Hundecha and Bardossy (2008) for instance, concluded that statistical downscaling seems to be more reliable during seasons when local climate is determined by large-scale circulation than by local convective processes.

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 significant 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 do not detect significant similarity between simulated and actual patterns for the indices of precipitation extremes in most cases. Wehner et al., (2010) show that at high resolution (approximately 60 km at the equator) an AGCM can reproduce the precipitation return values of comparable magnitude as those from high-quality observations. However, at the resolutions typical of the coupled GCMs used in the IPCC AR4, the precipitation return values are severely underestimated. Also, 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. These observations reveal a link between rainfall extremes and temperature, with heavy rain events increasing during warm periods and decreasing during cold periods. Furthermore, the observed amplification of rainfall extremes is found to be larger than that predicted by models, 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 a monotonic 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 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.

3.3.3. Wind

Extreme wind speeds pose a threat to human safety, maritime and aviation activities and the integrity of infrastructure. As well, other attributes of wind can cause extreme impacts. Trends in average wind speed can influence evaporation which in turn may influence water availability and droughts (e.g., McVicar et al.*,* 2008). Rapid transition in wind direction can affect forest fires, causing fires burning on the flank of the fire to flare up and become the new fire front (see Section 4.2.2.2, Mills*,* 2005). 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 Section 3.5.4 and 3.5.5). Therefore, general changes in a range of wind parameters at the global and regional scale are of interest, but these changes are not clearly delineated within the context of extremes. For example, extreme winds in Europe are most often associated with intense winter cyclones (e.g., Knippertz et al., 2000), but these events are only indirectly related to the average atmospheric circulation (Christensen et al., 2007).

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. Changes in wind extremes may arise from changes in the intensity or location of their associated phenomena or from other changes to the climate system (e.g., a change in local convective activity). 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).

3.3.3.1. Observed Changes

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. It concluded that mid-latitude westerlies have increased in both hemispheres (Trenberth et al., 2007).

Studies conducted since the AR4 are still too few to enable a comprehensive assessment of extreme wind changes. Long-term high-quality wind measurements from terrestrial anemometers are sparse in many parts of the globe due to the influence of changing instrumentation, station location, and surrounding land use (e.g., Cherry, 1988; Pryor et al., 2007; Jakob, 2010) and these issues have hampered the direct investigation of wind climatology changes. Nevertheless there have been a small number of new studies that have analysed wind speed trends from wind observations along with earlier studies for different parts of the world, some of which have also examined trends in extremes. These studies tend to point to declining trends in extremes in mid-latitudes and increasing trends in high latitudes. Several studies have compared the trends from anemometers with reanalysis products and in some cases find considerable differences (Hundecha et al., 2008) including differences in the sign of the trends (e.g., Smits et al., 2005; McVicar et al., 2008). New studies using wind proxies in the North Atlantic and Europe generally support earlier studies and indicate that there is a tendency for increased storminess around 1900 and in the 1990s, while the 1960s and 1970s were periods of low storm activity; but there are no consistent long term trends in the different available studies.

In the Northern Hemisphere, Pirazolli and Tomasin (2003) report a generally declining trend in winds from 1951 to the mid-1970s and an increasing trend since then, based on central Mediterranean records. The trends apply to both annual mean and annual maximum winds. Over the Netherlands, Smits et al., (2005) report declining trends in winds, including strong winds over 1962-2002. Significant declining trends in both summer and winter wind speeds were reported over China by Xu et al., (2006) and over the Tibetan plateau by Zhang et al., (2007b). In North America, Pryor et al., (2007) reported declining trends in wind over much of the USA over the 1973-2005 period. Lynch et al., (2004) report increasing trends in Alaska from 1921-2001. Hundecha et al., (2008) examined trends in extreme winds using non-stationary extreme value analysis over the Gulf of St Lawrence over the period 1979–2004, and found little change in wind extremes over this period.

In the Southern Hemisphere, McVicar et al., (2008) reports a statistically significant decline in wind speed over 57% of Australia over the 1975-2006 period. Positive (though not necessarily significant) trends are found over about 12% of the country including Tasmania, the interior of the mainland and coastal regions in the southeast and the far east. In Antarctica, Turner et al., (2005) reported increasing trends in mean wind speeds over the second half of the 20th century.

Some of these studies also compared anemometer-based trends to those from reanalysis products and reported differing or even opposite trends in the reanalysis data. Hundecha et al., (2008) compared anemometer trends with North American Regional Reanalysis data and found similarity in the directions of the change in the annual extremes at the selected stations but different magnitudes. Smits et al., (2005) compared in-situ trends with NCEP reanalysis over the Netherlands, and McVicar et al., (2008) with both NCEP and ERA40 over Australia, and found largely opposite trends. On the other hand, declining trends reported by Xu et al., (2006) over China were generally consistent with trends in NCEP reanalyses. Note, however, that the accuracy of trends from reanalysis data is still debated, since data assimilation can induce artificial trends in the products (e.g., Bengtsson et al., 2004).

Proxies for wind that use pressure tendencies and geostropic winds calculated from triangles of pressure observations from which storminess can be inferred have also been employed in a number of studies over Europe and the Atlantic (see 3.4.5). These studies suggest that there is a tendency for increased storminess around 1900 and in the 1990s, while the 1960s and 1970s were periods of low storm activity; but there are no long-term trends consistent between different available studies. More recent studies confirm these findings and illustrate that storminess in this region exhibits strong inter-decadal variability (Alexandersson et al., 2000; Allan et al., 2009; Wang et al., 2009b). The later half of the 20th century was punctuated by a peak in storminess around 1990 which according to Wang et al., (2009b) is unprecedented since 1874. However, no long-term trends were detected in storminess over this time period (Barring and von Storch, 2004; Barring and Fortuniak, 2009) or the period for which reanalysis data exist (Raible, 2007; Della-Marta et al., 2009).

3.3.3.2. Causes Behind the Changes

There is very little literature on the attribution of changes in winds including extremes, and so no assessment can be provided for this element at this point in time. Only one study, Wang et al., (2009c), formally detects a link between external forcing and positive trends in the high northern latitudes and negative trends in the northern midlatitudes using a proxy for wind (geostrophic wind energy) in the boreal winter.

Other studies have examined the likely causes for changes in winds including extreme winds. For example, Pirazolli and Tomasin (2003) report declining trends between 1951 and the mid-1970s and increasing trends since in the central Mediterranean. They find that the changes are positively correlated with temperature but not with the NAO index. For the British Columbian coast, Abeysirigunawardena et al., (2009) found that higher extreme winds tend to occur during the negative (i.e., cold) ENSO phase, consistent with an earlier study by Bromirski et al., (2005) who found a northward displacement of storminess in the northeast Pacific during La Niña episodes. Turner et al., (2005) note that the generally increasing trend of mean wind speeds over recent decades in Antarctica is consistent with the change in the nature of the SAM towards its high index state.

Declining wind speeds over China were reported by Xu et al., (2006) for both the winter and summer seasons. The winter declines, which result in a weakened winter monsoon circulation, were associated with greater warming over high-latitude land areas consistent with changes expected from anthropogenic warming. However the declines in summer wind speeds, resulting in a weakened summer monsoon circulation, were attributed to a cooling in central China associated with increased air pollution.

3.3.3.3. Projected Changes and Uncertainties

Projections of wind speed changes in general and wind extremes in particular were not specifically addressed in the AR4 although references are made to wind speed in relation to other variables and phenomena such as mid-latitude storm tracks, tropical cyclones and ocean waves (Christensen et al., 2007; Meehl et al., 2007b). The AR4 (IPCC, 2007a) reports that it is likely that future tropical cyclones (typhoons and hurricanes) will become more intense, with larger peak wind speeds associated with ongoing increases of tropical SSTs. It also reports that there is higher confidence in the projected poleward shift of the storm tracks and associated changes in wind patterns.

The small number of studies of projected extreme winds, together with shortcomings in the simulation of these events, means that it is still difficult to credibly project their future changes. Confidence in projections of wind and in particular extreme wind remains low because of the general low level of confidence in projected circulation changes in GCMs. The inability of GCMs at their present resolution to capture small scale meteorological phenomena that are often associated with extreme winds also contributes to the low confidence, although RCMs may help to address this problem as the number and regional extent of studies increases.

New studies since the AR4 provide more evidence for an increase in extreme winds in the high northern latitudes in the boreal winter and the southern ocean in the austral winter. McInnes et al., (2010) analysed global changes in average and 99th percentile wind speed (defined as the threshold dividing the highest 1% of daily winds from the remaining wind values) using daily wind speeds from nineteen models from the AR4 MMD. The top panels of Figure 3.10, reproduced from that study, show changes in average 10-m wind speeds for December to February (DJF) and June to August (JJA) for 2081-2100 relative to 1981-2000. In DJF in the Northern Hemisphere, increases in wind speed averaging 10% or more occur across northern Europe, the Arctic, northern North America and the northern Pacific between 40 and 50ºN. The wind speed increases at around 50-60ºN combined with declines to the southeast of this in the northern Pacific and Atlantic Oceans reflect the poleward movement of the storm tracks (see 3.4.5.3). Declines in wind also occur in the Mediterranean and Arabian Seas, much of Asia, and the eastern Equatorial Pacific. In JJA, consistent wind speed increases of 10% or more occur across much of the eastern U.S. through to central South America and northern and central Europe while large parts of the northeastern and equatorial Pacific undergo wind speed decrease. In the Southern Hemisphere in both seasons, a consistent strengthening of winds of over 10% occurs in the circumpolar trough between about 45 and 60ºS accompanied by a weakening of winds between about 30 and 40ºS which is associated with the poleward movement of the storm track (see 3.4.5.3). Wind speed increase also occurs in the southern Pacific Ocean between about 10 and 25°S.

INSERT FIGURE 3.10 HERE

Figure 3.10: The average of the multi-model 10 m mean wind speeds (top) and 99th percentile daily wind speeds (bottom) for the period 2080 to 2099 relative to 1980 to 1999 (% 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. Fine black stippling indicates where more than 90% of the models agree on the sign of the change and bold grey stippling (in white or light coloured areas) indicates where 66% of models agree on a small change between ±2 %. From McInnes et al., (2010).

Extreme wind speeds (bottom panels of Figure 3.10) show consistency between models over a larger portion of the globe in the direction of change, but the changes are generally less than ±5%. Increases in extremes occur in the high latitudes of both hemispheres and decreases across the lower latitudes as reported by Gasteneau and Soden (2009), but regional differences are apparent. For example, in DJF, consistent increases in extremes are seen across much of northern Europe, north Africa and west, east and southeast Asia and eastern North America. At least 90% of models agree on an increase in extreme winds of between 5 and 10% across the Arctic. In JJA, consistent increases in extremes of up to 5% occur in eastern South America, northern Australia and the south Pacific between 10 and 20 °S, and parts of Africa particularly in the northeast. Consistent increases are seen over much of the Southern Ocean, while consistent decreases are seen across large parts of the Atlantic and Indian Oceans and the northeast Pacific and northern North America.

Since the AR4 there have been several studies which have focussed on future changes to extreme winds. Gastineau and Soden (2009) used a 17-model ensemble to explore global changes in percentiles of 850 hPa wind speed. Zonally averaged changes presented in that study indicated agreement between models of a decreased frequency of the strongest wind events in the tropics and increased frequency in the strongest wind events in the extratropics.

Several regional studies have also been undertaken over Europe. Debernard and Roed (2008) reported projected statistically significant increases in 99th percentile winds across much of northern Europe, the British Isles and the ocean to the west and decreases to the south of Iceland in a variety of models under various emission scenarios (A2, B2, A1B). Increases in extreme (98th percentile) wind speeds in winter over large parts of Central Europe are also found in studies of both global and regional climate model output by Donat et al., (2009; 2010). A GCM ensemble indicates an increase of about 5% in wind speeds associated with storm events, although the changes are not statistically significant in all models.

Studies of extreme wind speed from eight RCMs have also been undertaken. Rockel and Woth (2007) reported a future increase in mean daily wind speed during winter months, and a decrease during autumn in areas influenced by North Atlantic extra-tropical cyclones. Further support for increases in extreme wind speeds over large areas of northern Europe is provided by Haugen and Iverson (2008). They report that extreme wind events become more frequent over large parts of northern Europe but note that the model responses are related to the representation of the Scandinavian pattern. Beniston et al., (2007) report that extreme wind speeds increase between 45° and 55°N, except over and south of the Alps, and become more north-westerly, but the magnitude of the increase depends on the specific RCM used. These changes were attributed to reductions in mean sea-level pressure and the generation of more North Sea storms. Given the level of agreement across models on mean and 99th percentile wind speeds illustrated in Figure 3.10, the degree to which the findings of these studies are robust across a larger set of GCMs or are a function of the particular selection of GCMs that were downscaled is not clear.

Sailor et al., (2008) statistically downscaled winds from several different climate models, to develop projections of winds over five airports in the northwest U.S. and results for 2050 suggest that summertime wind speeds may decrease by 5–10%, while changes to wintertime wind speeds were less certain.

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 still not well understood, but a variety of extremes such as floods, drought or even heat waves may occur more or less frequently in the monsoon regions as a consequence of climate change. Generally, however, precipitation is the most important variable for inhabitants of monsoon regions, but it is also a variable associated with larger uncertainties in climate simulations (Wang et al., 2005; Kang and Shukla, 2006). Changes in extremes in the monsoon regions can be characterized more broadly than via precipitation only. Thus monsoon changes could be better depicted by large-scale dynamics, circulation or moisture convergence. However, few studies have focused on observed changes in the large-scale and in the regional monsoon circulations. Hence, in this section we focus on monsoon-induced changes in rainfall, but when literature is available we also provide assessments on associated circulation changes.

3.4.1.1. Observed Changes

Considering precipitation as perhaps the most important aspect of monsoon, several studies have focused on changes in this variable as an indicator of changes in monsoon induced by climate change. The delineation of the global monsoon has been mostly performed using rainfall data or outgoing longwave radiation (OLR) fields (Kim et al., 2008). The metrics based on rainfall have been used in various studies on global and regional monsoons (see IPCC, 2007a). Lau and Wu (2007) reveal two opposite time evolutions in the occurrence of rainfall events in the tropics, in overall agreement with the Climate Research Unit's (CRU) gauge-only rainfall data over land: a negative trend in moderate rain events and a positive trend in heavy and light rain events. Positive trends in intense rain located in deep convective cores of the Intertropical Convergence Zone (ITCZ), South Pacific Convergence Zone, Indian Ocean and monsoon regions. Studies based on observations for 1951-2003 (Alexander et al., 2006) suggest an increase in heavy precipitation in all the monsoon regions of the planet.

The American monsoon regions are vulnerable to climate change and especially to extreme climate events such as intense droughts and floods (e.g., Cavazos et al., 2008; Kunkel et al., 2008; Marengo et al., 2009a; Marengo et al., 2009b; Soares and Marengo, 2009; Arriaga-Ramírez and Cavazos, 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 (Gan et al., 2006; da Silva and de Carvalho, 2007; Raia and Cavalcanti, 2008; Nieto-Ferreira and Rickenbach, 2010). Increase in heavy precipitation during 1960-2000 in the South American monsoon have been documented by Marengo et al., (2009a; 2009b), and Rusticucci et al., (2009). For the North American monsoon region, Cavazos et al., (2008) find increases in the intensity of precipitation in the mountain sites of northwestern Mexico over the 1961-1998 period, which appear to be related to an increased contribution from heavy precipitation derived from tropical cyclones (TCs). The authors also find that TC-related extreme precipitation events are associated with SST anomalies similar to weak La Niña conditions in the eastern Equatorial Pacific and a strong land-sea thermal contrast over northwest Mexico and the U.S. southwest two weeks prior to their onset. Arriaga-Ramirez and Cavazos (2010) find that total and extreme rainfall in the monsoon region of western Mexico and the U.S. southwest have significantly increased during 1961–1998, mainly by an important contribution from the winter season. Groisman and Knight (2008) find that consecutive dry days with periods longer than one month, have significantly increased in the U.S. southwest.

Zhou et al., (2008b) focused on large- or regional-scale dynamic fluctuations rather than on the regional-scale precipitation variations for the southeast Asian-Australian monsoon. In the Indo Pacific region, covering the southeast Asian and north Australian monsoon, Caesar et al., (2010) identify less 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 have been identified, there is generally a trend towards wetter conditions in common with the global results of Alexander et al., (2006). Some of the extreme precipitation appears to be positively correlated with a La Niña-like SST pattern. Guo et al., (2010) analyze near-surface wind speed change in China and its monsoon regions from 1969 to 2005 and show a significant weakening in annual and seasonal mean wind. These changes indicate reduced fluctuations in wind and wind storms in recent decades, contributing to decreased frequency and magnitude of dust storms (though an increase has been reported in more recent years, see also Section 3.5.8). The trivial changes in summer winds in east and southeast China suggest fairly steady monsoon winds over the decades. A main cause of the weakening wind is shown to be the weakening in the lower-tropospheric pressure-gradient force, a result pointing to climate variation as the primary source of the wind speed change. Superimposed on the climate effect is the urban effect. Liu et al., (2010) shows a decline in recorded precipitation events over in China 1960–2000, which is mainly accounted for by the decrease of light precipitation events, with intensities of 0.1–0.3 mm/day.

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. Previously, Goswami et al., (2006) found that for 1950-2003, both the frequency of occurrence and intensity of extreme rain events over central India exhibited a significant increasing trend, while that of the weak and moderate rain events showed a significant decreasing trend. Sen Roy (2009) investigated changes in extreme hourly rainfall in India, and found widespread increases in extreme 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. Goswami et al., (2006) explain the higher intensity of extreme rain events over central India have reflected a significant decreasing trend in light precipitation, which has also being detected in China (Liu et al., 2010).

In the African monsoon region, Fontaine et al., (2010) investigated recent observed trends using high-resolution gridded precipitation from the Climatic Research Unit (period 1979–2002), OLR and the NCEP reanalyses. The results show a rainfall increase in north Africa since the mid-90s with significant northward migrations of rainfall amounts, i.e., +1.5°C for the 400 mm July to September isohyets, whereas deep convection has significantly increased and shifted northward. After 1993–1994, the migration of the Saharan heat low towards northwest has been more marked.

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 few exceptions in some monsoon regions, this has not changed since.

3.4.1.2. Causes Behind the Changes

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). The trend is strongly linked to the warming trend over the central eastern Pacific and the western tropical Indian Ocean (Zhou et al., 2008b). The decrease in global land monsoon rainfall mainly occurred in the north African and south Asian monsoons. The long-term changes of the other monsoon subsystems are not significant in the context of regional averages (Zhou et al., 2008a). For the west African monsoon, Joly and Voldore (2009) explore the role of Gulf of Guinea SSTs in its interannual variability. In most of the studied CMIP3 simulations, the inter-annual 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 trend in north African 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). 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., 2010). The link between tropical cyclones as well as the role of SST anomalies in the Eastern Equatorial Pacific, and observed increases in rainfall extremes in the North American monsoon has been investigated by Cavazos et al., (2008).

An important aspect for global monsoon patterns is the seasonal reversal of the prevailing winds. The significant weakening in annual and seasonal mean wind over China (Guo et al., 2010) indicates reduced fluctuations in wind and wind storms in recent decades, contributing to decreased frequency and magnitude of dust storms. A main cause of the weakening wind is a weakening of the lower-tropospheric pressure-gradient force. The observed changes in the African monsoon region (Fontaine et al., 2010) are associated with significant reinforcements of the southwesterly low-level winds and Tropical Easterly jet and with a northward shift of the African Easterly jet.

The CMIP3 models are able to capture the major monsoon rainfall regions around the globe, however, in regional aspects of monsoon rainfall climatology, simulations show remarkable differences depending on the horizontal resolution of the respective atmospheric models, with higher resolution models producing more realistic regional details of precipitation climatology because of better representation of surface topography. It is useful to examine changes in monsoon in a global perspective as a regional monsoon system interacts with other monsoon(s) to some extent (Meehl and Arblaster, 2002; Biasutti et al., 2003), and there is a potential improvement in the signal/noise ratio in the global monsoon system when compared with that in regional monsoons (see also Section 3.2.2 for the attribution of regional vs global changes). Observations show a negative trend in global monsoon rainfall over land during 1948–2003, primarily due to the weakening of the summer monsoon rainfall in the Northern Hemisphere (Wang and Ding, 2006). 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, through 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 trend in the Northern Hemisphere monsoons detected in the CMIP3 models is generally consistent with the observations, albeit with much weaker magnitude (Kim et al., 2008). The global oceanic monsoon precipitation has increased since 1980, and this positive trend is reproduced by 20 of the 21 CMIP3 models, though the models that do not include natural forcing (with MRI CGCM2.3.2a an exception) produce a more significant positive trend. The model resolution does not exhibit a considerable influence on trend simulation (Kim et al., 2008).

In summary, the CMIP3 models are able to simulate the global monsoon characteristics reasonably well. However, models do not agree in the sign of the trend of large-scale changes in the monsoon circulation, and models of finer resolution do not provide better representations of tropical monsoon circulation trend (Kim et al., 2008). As well, models with finer resolution do not show a significant east Asian summer monsoon response to external forcing (Kripalani et al., 2007a). AGCM studies suggest that several dynamic monsoon indices representing Asian-Australian monsoon circulation are forced primarily by tropical SST changes (Zhou et al., 2009) in association with El Niño activity.

Changes in regional monsoons are strongly influenced by the changes in the states of dominant patterns of climate variability such as the El Niño – Southern Oscillation (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). However, it is not always clear how those modes may have changed in response to external forcing (Shiogama et al., 2005). Additionally, model-based evidence has suggested that land surface processes and land use changes could in some instances significantly impact regional monsoon. Tropical land cover change in Africa and southeast Asia appears to have weaker local climatic impacts than Amazonia does, in large part due to influences of the Asian and African monsoon circulation systems in those regions (Voldoire and Royer, 2004; Mabuchi et al., 2005a, b). Grimm et al., (2007) suggest that in the South American monsoon region, precipitation anomalies, remotely forced in the spring, produce soil moisture and near surface temperature anomalies, which alter the surface pressure and wind divergence. In this regard, 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 monsoon is difficult to quantify, due both to the poor documentation of land use and difficulties in simulating monsoon at fine scales. Moreover, there are still large uncertainties and a strong model dependency in the representation of the relevant land surface processes, associated parameters, and resulting interactions (Pitman et al., 2009).

3.4.1.3. Projected Changes and Uncertainties

The AR4 concluded (Christensen et al., 2007) that 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." Post-AR4 work has not substantially changed these conclusions.

As global warming is projected to lead to faster warming over land than over the oceans (Sutton et al., 2007), the continental-scale land-sea thermal contrast, a major factor affecting monsoon circulations, may become stronger in summer and weaker in winter. Based on this hypothesis, a simple scenario is that the summer monsoon will be stronger and the winter monsoon will be weaker in the future than the present. However, model results are not as straightforward as 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). Complicating this picture further, however, is the fact that observations and models suggest that changes in monsoons are related at least in part to changes in observed SSTs (see 3.4.1.2). Changes in global SSTs are expected to be affected by anthropogenic forcing, so this may lead to changes in monsoon circulations. Furthermore, changes in rainfall depend not just upon SSTs but also upon changes in the spatial and temporal SST patterns and regional changes in atmospheric circulation.

At regional scales, there is little consensus in GCM projections regarding the sign of future change in the monsoons characteristics, mainly 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), in the Australian monsoon in southern summer, and in the southern part of the west African monsoon, but with some decreases in the Sahel in northern summer. Heavy precipitation is projected to increase in all monsoon regions by the end of the 21st century as derived from the CMIP3 model (Tebaldi et al., 2006).

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 of 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 scenario (Marengo et al., 2009a; Soares and Marengo, 2009).

The south Asian summer monsoon could be weakened and its onset delayed due to rising temperatures in the future, according to a recent modeling study. Asfaq et al., (2009) suggest weakening of the large-scale monsoon flow and suppression of the dominant intraseasonal oscillatory modes with overall weakening of the south Asian summer monsoon by the end of the 21st century. Such changes in monsoon dynamics could have substantial impacts by decreasing summer precipitation in key areas of south Asia. In contrast, an earlier study of the AR4 MME indicates a significant increase in mean south Asian summer monsoon precipitation of 8% and a possible extension of the monsoon period, together with intensification of extreme excess and deficient monsoons (Kripalani et al., 2007b).

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 band at the end of the 21st century. They found a delay in early summer rain withdrawal over the region extending from Taiwan, 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. 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 at the mid-troposphere from the Asian continent to the northwest Japan provides favorable conditions for intense precipitation in the Baiu season in Japan (Kanada et al., 2009). 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 AR4 MME mean SST changes. The frequency of heavy precipitation is projected to increase at the end of the 21st century for hourly as well as daily precipitation. Further, extreme hourly precipitation is projected to increase even in the near future (2030s) when the temperature increase is still modest. Much remains to be learned about the mechanisms that produce such inter-decadal changes in the east Asian summer monsoon, and the response of the east Asian monsoon to global warming in at least some models is not significant (Kripalani et al., 2007a).

Some of the uncertainty on global and regional climate change projections in the monsoon regions results from 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 on the global and regional scale (e.g., changes in land-use/cover). Kharin and Zwiers (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 a very large disagreement in the Tropics suggesting that some physical processes associated with extreme precipitation are not well represented by the models. This reduces confidence in the projected changes in extreme precipitation over the monsoon regions.

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. Recently, 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, often exceeding 50% of the climatological values. Many of the biases are pervasive, being common to most simulations. Three-member ensembles of baseline simulations (1961–1990) from an RCM at 50 km resolution have confirmed that significant improvements in the representation of regional processes over south Asia can be achieved by models with higher spatial resolution (Kumar et al., 2006). Moreover, confirming the importance of resolution, RCMs simulate more realistic climatic characteristics over east Asia than GCMs, whether driven by re-analyses or by GCMs (Christensen et al., 2007). Subseasonal extremes of precipitation and active-break cycles of the Indian summer monsoon in climate-change projections have been analyzed by Turner and Slingo (2009). They found that the chance of reaching particular thresholds of heavy rainfall approximately doubled over northern India. The local distribution of such projections is uncertain, however, given the large spread in mean monsoon rainfall change and associated extremes amongst the GCMs. According to AR4, monsoon rainfall simulations and projections vary substantially from model to model in northern Australia, thus there is little confidence in model precipitation projections over that particular region (Christensen et al., 2007).

Many of the important climatic effects of the Madden Julian Oscillation (MJO), 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 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), although 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 tropical region highly uncertain. Lin et al., (2008a) show that the CGCMs have significant problems and display a wide range of skill in simulating the North American monsoon and associated intraseasonal 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). The rainy season of the semi-arid African Sahel is projected by twenty-first simulations to start later and become shorter (Biasutti and Sobel, 2009). However, the robust agreement across models on the seasonal distribution of Sahel rainfall changes stands in contrast with large uncertainty for summertime rainfall totals there.

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 the emissions into concentrations of radiatively active species (i.e., via atmospheric chemistry and carbon-cycle models) and especially those derived from aerosols product of biomass burning. 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.

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). 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 cool ocean temperatures in this region. Both extremes 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. 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.

3.4.2.1. Observed Changes

The AR4 noted that the nature of the El Niño – Southern Oscillation has varied substantially over time, with strong events from the late 19th century through the first quarter of the 20th century and again after 1950. A climate shift around 1976–1977 was associated with a shift to generally above-normal SSTs in the central and eastern Pacific and a tendency towards more prolonged and stronger El Niño episodes (Trenberth et al., 2007). Paleoclimatic evidence suggested that the phenomenon was quite weak up to a few thousand years ago.

Research subsequent to the AR4 has provided evidence from fossil corals that the El Niño – Southern Oscillation 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 timescales, there is evidence that the El Niño – Southern Oscillation 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 or LGM (An et al., 2004). Fossil coral evidence does indicate that the phenomenon did continue to operate during the LGM (Tudhope et al., 2001).

Instrumental data (SST and surface atmospheric pressure measurements) allow us a more detailed study of changes in the behaviour of the phenomenon over the past century or so. Ocean temperatures in the central equatorial Pacific (the so-called NINO3 index) suggest that the phenomenon was particularly active during the 1970s and less active in the 1950s and 1960s, with perhaps a trend toward more frequent or stronger El Niño episodes over the past 50–100 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 or SOI (another index of the phenomenon – 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 behaviour 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 towards what might be considered more ''El Niño-like'' behaviour in the SOI (and more weakly in NINO3.4), but only through the period March–September and not in November–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 the El Niño–Southern Oscillation 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). In turn, this change in the location of the strongest SST anomalies associated with El Niño 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) show that episodes with the warming centred in the central Pacific exhibit different patterns of Australian rainfall variations than do other varieties of El Niño events.

3.4.2.2. Causes Behind the Changes

Regarding possible causes of changes in the El Niño – Southern Oscillation phenomenon, the AR4 concluded that "as yet there is no detectable change in ENSO variability in the observations, and no consistent picture of how it might be expected to change in response to anthropogenic forcing" (Hegerl et al., 2007). However, models did suggest that orbital variations could affect the ENSO behaviour by, for instance, reproducing an apparent increase in event frequency and amplitude throughout the Holocene (Jansen et al., 2007).

Post-AR4 studies have not changed the AR4 assessment that orbital variations could affect the ENSO activity and that there is still no clear indication of possible role of anthropogenic influence on ENSO activity. Vecchi and Wittenberg (2010) note that the "tropical Pacific could generate variations in ENSO frequency and intensity on its own (via chaotic behaviour), respond to external radiative forcings (e.g., changes in greenhouse gases, volcanic eruptions, atmospheric aerosols, etc), or both". The paleoevidence indicates that the El Niño – Southern Oscillation can continue to operate, although altered perhaps in intensity, through quite anomalous climate periods, but that it does fluctuate in response to changes in radiative forcing caused by orbital variations (Vecchi and Wittenberg, 2010). Cane (2005) noted that a relatively simple coupled model suggested that systematic changes in the El Niño could be stimulated by seasonal changes in insolation. However, a more comprehensive model simulation (Wittenberg, 2009) has suggested that long-term changes in the behaviour of the phenomenon might occur even without forcing from radiative changes.

The possible role of increased greenhouse gases in affecting the behaviour of the El Niño – Southern Oscillation over the past 50-100 years is uncertain. Some studies (e.g., Zhang et al., 2008a) 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., Powers and Smith, 2007) suggest that changes in the phenomenon are still within the range of natural variability (ie, that no change has yet been detected, let alone attributed). Yeh et al., (2009) suggested that changes in the background temperature associated with increases in greenhouse gases should affect the behaviour of the El Niño, such as the location of the strongest SST anomalies, because El Niño behaviour is strongly related to the average ocean temperature gradients in the equatorial Pacific.

A caveat regarding all projections of future behaviour of the El Niño – Southern Oscillation arises from systematic biases in the depiction of El Niño – Southern Oscillation behaviour through the 20th century by models. 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 the El Niño – Southern Oscillation 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 behaviour of the El Niño – Southern Oscillation, 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 of the El Niño – Southern Oscillation.

3.4.2.3. Projected Changes and Uncertainties

AR4 established that all models exhibited continued El Niño – Southern Oscillation (ENSO) interannual variability in projections through the 21st century, but the projected behaviour 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).

Global warming is expected to lead to a mean reduction of the zonal winds across the equatorial Pacific (Vecchi and Soden, 2007b). This change may be described as an "El Niño – like" average change because during an El Niño episode these winds generally weaken. However, there is only limited correspondence between these changes in mean state of the equatorial Pacific and an El Niño episode. For instance, climate models project that the Indonesian region would become wetter, and this is distinctly different to a typical El Niño event.

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 an increase in ENSO activity from doubling or quadrupling CO_2, but a considerable decrease in activity when CO_2 was increased by a substantial factor of 16 times (Cherchi et al., 2008).

The remote impacts, on rainfall for instance, of ENSO may also change as CO_2 increases, even if the equatorial Pacific aspect of the El Niño – Southern Oscillation 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 centred 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).

The position at the time of the AR4 was that there was no consistency of projections of changes in ENSO variability or frequency in the future (Meehl et al., 2007b). This position has not been changed as a result of post-AR4 studies. The evidence is that the nature of the El Niño – Southern Oscillation 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 the El Niño – Southern Oscillation will as well. However, Vecchi and Wittenberg (2010) conclude “the ENSO variations we see in decades to come may be different than those we’ve seen in recent decades – yet we are not currently at a state to confidently project what those changes will be”. However, they also observe that we are confident that El Niño and La Niña events will likely continue to occur and influence the climate but that there will continue to be variations in the phenomenon and its impacts, on a variety of timescales.

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. As well, coupled models still have difficulty simulating the El Niño – Southern Oscillation convincingly. Moreover, most of the models are not able to reproduce the typical wavetrains observed in the circulation anomalies associated with ENSO in the Southern Hemisphere (Vera and Silvestri, 2009) and the Northern Hemisphere (Joseph and Nigam, 2006). Such model limitations somewhat undermine our confidence in the projected changes by the majority of the models. Further research is required to analyse differences between the model simulations and projections of El Niño behaviour, to determine what causes these differences.

3.4.3. Other Modes of Variability

Other natural modes of variability 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 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 Europe and cold and dry winters in northern Canada and Greenland. Scaife et al., (2008) discuss the relationship between the NAO and European extremes. 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 refers to north-south shifts in atmospheric mass between the Southern Hemisphere middle and high latitudes and is the most important pattern of 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. Also it has been related to anomalously dry conditions over southern South America, New Zealand, and Tasmania and with 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, and brings heavy rainfall over the east Africa and severe droughts/forest fires over the Indonesian region. 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 weather extremes (Zhang et al., 2010). As is the case with ENSO, the simultaneous occurrence of climate extremes such as droughts associated with any of these various modes of variability may have consequences for disaster management.

3.4.3.1. Observed Changes

The AR4 noted that both the NAO and the SAM have exhibited trends towards their positive phase (strengthened midlatitude westerlies) over the last three to four decades, although both have returned to near their long-term mean state in the last five years (Trenberth et al., 2007). In the Northern Hemisphere, this trend has been associated with the observed winter change in storm tracks, precipitation and temperature patterns. A paleoclimate study (Goodkin et al., 2008) reported enhanced multidecadal variability in the NAO during the late 20th century, compared with the period 1800-1850. SAM has an influence on the interannual variability of precipitation in southeastern South America (Silvestri and Vera, 2003) and New Zealand (Ummenhofer and England, 2007). The SAM influence on temperature extremes in southern South America has also been reported (Barrucand et al., 2008). The SAM trends are related to contrasting trends of strong warming in the Antarctic Peninsula and a cooling over most of interior Antarctica (e.g., Marshall et al., 2006). Complicating these trends, Silvestri and Vera (2009) reported changes in the typical hemispheric circulation pattern related to SAM and its associated impact on both temperature and precipitation anomalies, particularly over South America and Australia, between the 1960s–70s and 1980s–90s. Recent work (Ummenhofer et al., 2008; Ummenhofer et al., 2009a; Ummenhofer et al., 2009b) has implicated the IOD as a cause of droughts in Australia, and heavy rainfall in east Africa (Ummenhofer et al., 2009c). The time scales of the multi-decadal modes of variability in these modes are so long that it is difficult to diagnose any change in their behaviour in modern data.

3.4.3.2. Causes Behind the Changes

The AR4 (Hegerl et al., 2007) noted that trends over recent decades in the NAO and SAM are *likely* related in part to human activity. The increasing positive phase of the SAM has been linked to stratospheric ozone depletion and to greenhouse gas increases. Models including both greenhouse gas and stratospheric ozone changes simulate a realistic trend in the SAM. However model simulations can show positive trends in the annular modes at the surface, but negative trends higher in the atmosphere, and it has been argued that anthropogenic circulation changes are poorly characterized by trends in the annular modes (Woollings et al., 2008). Goodkin et al., (2008) conclude that the variability in the NAO is linked with changes in the mean temperature of the Northern Hemisphere.

3.4.3.3. Projected Changes and Uncertainties

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, limitations in coupled model ability in reproducing the observed SAM impact on climate variability in the Southern Hemisphere has been reported (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).

The AR4 noted that 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. During the 21st century, although stratospheric ozone concentrations are expected to stabilise or recover, tending to lead to a weakening of the SAM, polar vortex intensification is likely to continue due to the increases in greenhouse gases. A very recent study (Woolings et al., 2010) found a tendency towards a more positive NAO under anthropogenic forcing through the 21st century, 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 multidecadal 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.

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 been remarkably steady over the modern period of geostationary satellites (since around the mid-1970's). 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 fresh-water flooding from extreme rainfall generally cause the great majority of damage and loss of life. 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, are also important impacts. Projected sea level rise will further compound tropical cyclone surge impacts. Tropical cyclones that track northward 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 northern waters 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. While there is a relationship between intensity and storm surge, the structure and areal extent of the wind field also play an important role. Other relevant tropical cyclone measures include frequency, duration, and track. Forming robust physical links between all of these metrics and natural or human-induced climate variability is a major challenge. Significant progress is being made, but substantial uncertainties still remain due largely to data quality issues (see 3.2.1, and below) and imperfect theoretical and modeling frameworks (see below).

3.4.4.1. Observed Changes

Detection of trends in tropical cyclone metrics such as frequency, intensity, and duration remains a significant challenge. Historical tropical cyclone records, which begin in 1851 in the North Atlantic and typically in the mid-20th century in other regions, 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 30-40 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., 2007b). Landsea et al., (2009) showed that a large contribution of the observed long-term trend in the record of North Atlantic tropical cyclone frequency is due to a trend in the frequency of short-lived storms, a subset of storms that may be particularly sensitive to changes in technology and reporting protocols. However, Emanuel (2010) demonstrates that the changes in short-duration storms may also have physical causes, and Kossin et al., (2010) find that much of the changes in the frequency of short-duration storms in the Atlantic have occurred in the Gulf of Mexico in close proximity to land and thus largely avoids the data-quality issues with pre-satellite storm undercounts.

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., 2007a; 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). It thus remains uncertain whether any reported long-term increases in tropical cyclone frequency are robust, after accounting for past changes in observing capabilities (Knutson et al., 2010).

Whereas frequency estimation requires only that a tropical cyclone be identified and reported at some point in its lifetime, 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 tropical cyclone intensity and climate change, they are necessarily confined to a 30+ year period of satellite observations, and do not 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 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, but it is a present point of debate whether local SST or SST relative to mean tropical SST is the more physically relevant metric (Swanson, 2008). The distinction is an important one when making projections of power dissipation based on projections of SST, particularly in the Atlantic where SST has been increasing more rapidly than the tropics as a whole (Vecchi et al., 2008).

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 clear trend in tropical cyclone rainfall has not yet been established due to a general lack of studies.

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; Woodruff et al., 2008b; 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 Policy Makers concluded that it is likely that a trend had occurred in intense tropical cyclone activity since 1970 in some regions (IPCC, 2007b). In somewhat more detail, it was further stated that "*there is observational evidence for an increase in intense tropical cyclone activity in the North Atlantic since about 1970, correlated with increases of tropical SSTs. There are also suggestions of increased intense tropical cyclone activity in some other regions where concerns over data quality are greater. Multi-decadal variability and the quality of the tropical cyclone records prior to routine satellite observations in about 1970 complicate the detection of long-term trends in tropical cyclone activity. There is no clear trend in the annual numbers of tropical cyclones*." The subsequent U.S. CCSP SAP 3.3 (Kunkel et al., 2008) concluded that "*Atlantic tropical storm and hurricane destructive potential as measured by the Power Dissipation Index (which combines storm intensity, duration, and frequency) has increased*". *The report concludes that "the power dissipation increase is substantial since about 1970, and is likely substantial since the 1950s and 60s, in association with warming Atlantic SSTs*", and that "*it is likely that the annual numbers of tropical storms, hurricanes and major hurricanes in the North Atlantic have increased over the past 100 years, a time in which Atlantic SSTs also increased*", but that "*the evidence is not compelling for significant trends beginning in the late 1800s*". Based on research subsequent to the IPCC AR4 and CCSP SAP3.3,which further elucidated the scope of uncertainties in the historical tropical cyclone data, the most recent assessment by the World Meteorological Organization Expert Team on Climate Change Impacts on Tropical Cyclones (Knutson et al., 2010) does not assign a likely confidence level to the reported increases in annual numbers of tropical storms, hurricanes and major hurricanes counts over the past 100 years in the North Atlantic basin, nor does it conclude that the Atlantic Power Dissipation Index increase is likely substantial since the 1950s and 60s.

Our assessment regarding observed trends in tropical cyclone activity are unchanged from the WMO report (Knutson et al., 2010):

1. It is uncertain whether any reported long-term increases in tropical cyclone frequency are robust, after accounting for past changes in observing capabilities.
2. An increase globally since 1983 in the intensities of the strongest tropical cyclones has been reported (Elsner et al., 2008); however, the short time period of the data does not allow for a convincing detection and attribution of an anthropogenic signal compared with variability from natural causes.
3. A detectable change in tropical cyclone-related rainfall has not been established by existing studies.
4. There is no conclusive evidence that any observed changes in tropical cyclone genesis, tracks, duration, or surge flooding exceed the variability expected from natural causes.

3.4.4.2. Causes Behind the 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 the CCSP SAP 3.3 report (CCSP, 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). Such a shift in the distribution would be most evident at the upper quantiles of the distribution as the strongest tropical cyclones become stronger (Elsner et al., 2008).

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. Statistical correlations between necessary ambient environmental conditions and tropical cyclogenesis frequency have been well documented (DeMaria et al., 2001). For example, there is an apparent minimum SST threshold for genesis. However, these relationships are less formally based on physical arguments and may be neither stationary in time nor independent of other factors (Nolan et al., 2007; Knutson et al., 2008). Similarly, the pathways through which climate variability can affect tropical cyclone genesis position, duration, and tracks are not well understood, and guidance from dynamical models is still limited, although statistical correlations have been identified.

A further complication in determining cause and effect arises from the strong relationship between intensity and duration (Kossin and Vimont, 2007). Since tropical cyclones moving through a favourable environment intensify at an average rate of about 12 m s^{-1} per day (Emanuel, 2000), the lifetime maximum intensity of a storm depends on its duration, which can depend on its genesis location. There are then three distinct, but not mutually exclusive pathways inducing an upward shift in a distribution of tropical cyclone intensities: increasing mean ambient potential intensity, increasing mean intensification rate, or increasing the mean duration of the intensification periods. The first is more easily linked to climate and tested in a numerical or theoretical framework, but the mechanistic links to relate the latter two to climate variability are significantly more difficult to uncover.

Based on a variety of model simulations, the expected long-term changes in tropical cyclone characteristics under greenhouse warming is a decrease 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 an informal comparison of the amplitude of the modelled 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 SRES A1B warming scenarios.

With the exception of the North Atlantic, global tropical cyclone data is generally 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 above) 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 the historical tropical cyclone data, no such attribution conclusion was drawn in the recent WMO report (Knutson et al., 2010), which states on p. 14 of their Supplementary Information "we do not draw such an attribution conclusion in this assessment. Specifically we do not conclude that there has been a detectable change in tropical cyclone metrics relative to expected variability from natural causes, particularly owing to concerns about limitations of available observations and limited understanding of the possible role of natural climate variability in producing low frequency changes in the tropical cyclone *metrics examined*."

The conclusions of the present report are similar to the WMO report (Knutson et al., 2010): the uncertainties in the historical tropical cyclone records and the degree of tropical cyclone variability — comprising random processes and linkages to various natural climate modes such as El Niño — do not presently allow for the attribution of any observed changes in tropical cyclone activity to anthropogenic influences.

3.4.4.3. Projected Changes and Uncertainties

The AR4 concluded (Meehl et al., 2007b) that "*results from embedded high-resolution models and global models, ranging in grid spacing from 100 km to 9 km, project a likely increase of peak wind intensities and notably, where analysed, increased near-storm precipitation in future tropical cyclones. Most recent published modelling studies investigating tropical storm frequency simulate a decrease in the overall number of storms, though there is less confidence in these projections and in the projected decrease of relatively weak storms in most basins, with an increase in the numbers of the most intense tropical cyclones*." The conclusions here are similar to those in the AR4, but somewhat more detail is now possible.

The spatial resolution of models such as the CMIP 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.

While it remains uncertain whether long-term past changes in global tropical cyclone activity have exceeded the variability expected through natural causes (Knutson et al., 2010), theory (Emanuel, 1987) and idealized dynamical models (Knutson and Tuleya, 2004) predict increases in tropical cyclone intensity under greenhouse warming. The 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. These models also consistently project little change or a reduction in overall tropical cyclone frequency, but with an accompanying substantial fractional increase in the frequency of the strongest storms and increased precipitation rates. Mean 21st century global cyclone intensity changes under conditions roughly equivalent to A1B emissions scenarios are projected between 3 and 11%, and a decrease of –6 to –34% is projected in global tropical cyclone frequency. The downscaling experiments of Bender et al., (2010), which, as described above, use an ensemble of AR4 MMD 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 a 75% increase in the frequency of Saffir-Simpson category 4 and 5 hurricanes. In addition to a decrease in frequency and an increase in intensity, higher resolution models also consistently project increased precipitation rates (~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. 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 timescales. 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. Alternatively, there is a similarly strong relationship between power dissipation and relative SST, which represents the difference between local and globally-averaged 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 of 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.

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 (+3 to +11% globally) is likely, although increases may not occur in all tropical regions (Knutson et al., 2010). It is more likely than not that the frequency of the most intense storms will increase by more than 11% in some ocean basins. As noted above in 3.4.4.1, observed changes in tropical cyclone-related rainfall have not been clearly established. However, as water vapour in the tropics increases (Trenberth et al., 2005) there is an expectation for increased tropical cyclone-related rainfall in response to associated moisture convergence increases (Held and Soden, 2006; see also Section 3.3.1.2). This increase is expected to be compounded by increases in intensity as dynamical convergence under the storm is enhanced. Models 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%. 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.

When simulating 21st century warming under the A1B emission scenario (or a close analogue), 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, rainfall, and confidence is particularly low for projections of frequency within individual basins. Current models project frequency changes ranging from -6 to -34% globally, and up to ± 50% or more in individual basins by the late 21st century. 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.

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 or as a result of flow over mountains (lee cyclogenesis). They may be accompanied by adverse weather conditions such as windstorms, the build up of waves and storm surges or extreme precipitation events. In addition, they are the main poleward transporter of heat and moisture. 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 baroclinic instability such as a disturbance along a zone of strong temperature contrast, which is a reservoir of potential energy that can be converted into the kinetic energy associated with extratropical cyclones. In addition, intensification of the system may also take place due to latent heat release or other diabatic processes (Gutowski et al., 1992).

3.4.5.1. Observed Changes

The AR4 noted a likely net increase in 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 report on several papers showing increases in the number or strength of intense extratropical cyclone 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 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., 2006a; Schneidereit et al., 2007; Raible et al., 2008; Vilibic and Sepic, 2010) and fewer (Wang et al., 2006a; Raible et al., 2008) in the mid latitude Atlantic. The increase in high latitude cyclone activity is also reported in several studies of Arctic cyclone activity (Zhang et al., 2004c; Sorteberg and Walsh, 2008), but the magnitude and even the existence of the changes may depend on the choice of reanalysis (Simmonds et al., 2008).

Since the AR4 several studies of historical coastal European storminess based on the 99th and 95th percentiles of pressure tendencies or geostrophic wind deduced from triangles of pressure stations have documented large decadal variability in the storminess (Andrade et al., 2008; Hanna et al., 2008; Matulla et al., 2008; Wang et al., 2008; Allan et al., 2009; Barring and Fortuniak, 2009). Periods with peak storminess vary for different regions and there are no long-term trends over the century that are consistent among the different studies. There is however a tendency for increased storminess around 1900 and in the 1990s while the 1960s and 1970s were periods of low storm activity.

Long term in situ observations of north Pacific extreme cyclones are considerably fewer than for the Atlantic cyclones. Bromiski et al., (2003) provided an estimate of the variation in ''storminess'' from 1858 to 2000 using an hourly tide gauge record from San Francisco (West Coast, U.S.). They noted no substantial change in the monthly non-tide residuals (NTR), but a significant increasing trend in the highest 2% of extreme winter NTR since about 1950. The increasing trend in the extreme NTR was also noted by Menendez et al., (2008) using significant wave height from 26 buoys between 30–45°N near the western coast of the U.S. covering the period 1985–2007. 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 western U.S. coast. This is in line with the study of Graham and Diaz (2001) using reanalysis and in situ data for the last 50 years which noted a significant increase in the number and intensity of north Pacific wintertime intense extratropical cyclone systems since the 1950s. This trend was accompanied by an eastward shift and an intensification of the Aleutian Low from the mid-1970s when a generally anticyclonic period gave way to more intense cyclonic activity (Favre and Gershunov, 2006). The study of Raible et al., (2008) points in the same direction as the above-mentioned studies showing increased intensity of Pacific extratropical cyclones in all seasons during the 1958–2001 period. It should be noted that by using MSLP observations made by ships, Chang (2007) found trends in the Pacific to be much smaller than that found in the NCEP reanalysis.

Using hourly mean sea level pressure data observed at 83 Canadian stations for up to 50 years (1953–2002), Wang et al., (2006a) showed that winter cyclones have become significantly more frequent, longer lasting, and stronger in the lower Canadian Arctic, but less frequent and weaker in the south, especially along the southeast and southwest coasts. Winter cyclone deepening rates were reported to increase in the zone around 60°N but decreased in the Great Lakes area and southern Prairies–British Columbia. Using a longer time period (1900 to 1990), Angel and Isard (1998) reported a significant annually and cold season increase in the number of strong cyclones across the Great Lakes. This seems to contradict the findings of Wang et al., (2006a), but also the Angel and Isard study finds a slight decrease since the 1950s. Studying U.S. East Coast winter cyclones using reanalyses, Hirsch et al., (2001) found a tendency toward weaker low-pressure systems over the past few decades and no statistically significant trends in their frequency.

Studies on extratropical cyclone activity in northern Asia are few. Zhang et al., (2004c) noted a decrease in cyclone activity (a parameter integrating cyclone intensity, number and duration) over Eurasia (60–40°N) over the period 1948-2002, while Wang et al., (2008) reported on deceasing trends in intensity (1958–2001) of seasonal and annual extratropical cyclones in the eastern part of Eurasia (80-140°E and 60–40°N). Wang et al., (2008) also noted a 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) which reported a decrease in the number of severe storms for mainland China using the 95th and 99th percentiles of observed 6-hourly pressure changes (1954 – 2004).

Using reanalyses, Pezza et al., (2007) confirms previous studies showing a trend towards fewer and more intense systems in the Southern Hemisphere. A new study (Lim and Simmonds, 2009) using the ERA-40 reanalysis instead of the NCEP reanalysis used in previous studies, confirms the trend towards more intense systems, but does not support the decrease in cyclone density seen in previous studies. This emphasises the weaker consistency among reanalysis products for the Southern Hemisphere extratropical cyclones and the possibility of some of the trends being biased by data inhomogeneities (Wang et al., 2006a). Wang et al., (2006a) noted a poleward shift in storm tracks in the Southern Hemisphere, confirming previous studies (Fyfe, 2003; Hope et al., 2006) and Alexander and Power (2009) show that the number of observed severe storms at Cape Otway (south-east Australia) has decreased significantly since the mid-19th century, strengthening the evidence of a southward shift in Southern Hemisphere storm tracks. Fredriksen and Fredriksen (2007) linked the reduction in cyclogenesis at 30°S and southward shift to a decrease in the vertical mean meridional temperature gradient.

In summary, research subsequent to the AR4 supports previous findings of a poleward shift in the tracks, but do not provide sufficient information to increase the degree of confidence in the assessment. There are few post AR4 studies on global changes in the intensity of extreme cyclones, but there is growing evidence of a intensification of extratropical cyclones in high-latitudes. Trends in the total number of cyclones are less clear and seem more sensitive to tracking scheme, choice of physical quantity to represent the cyclone and choice of reanalysis data set. New insight into the regional variability and trends in extratropical cyclones has emerged since AR4. In the Atlantic, studies using reanalysis points toward a northward shift in the cyclone activity during the last 60 years with both more and more intense wintertime cyclones in the high-latitude Atlantic, but there is no clear overall increase in number or intensification if the whole Atlantic is considered. The Atlantic trends should be seen in light of new studies with longer time spans indicating that the reanalysis cover a time period which starts with relatively low cyclonic activity in northern coastal Europe in the 1960s and reaches a maximum in the 1990s. For the Pacific, new studies indicate a increase in intensity and there are indications that this is accompanied by an eastward shift in the Aleutian Low. New studies on Southern Hemisphere extratropical cyclones confirm previous studies reporting a poleward shift and a possible intensification of the Southern Hemisphere cyclones. However, the latter conclusion relies on reanalysis products that may contain inhomogeneities affecting the Southern Hemisphere trend estimates. Advances have been made in documenting the observed decadal and multidecadal variability of cyclones (Andrade et al., 2008; Hanna et al., 2008; Matulla et al., 2008; Allan et al., 2009; Barring and Fortuniak, 2009), but insufficient knowledge of the observed decadal and multidecadal variability and how the influence of reanalysis inhomogeneities are influencing cyclone number and intensity trends over the last 50 years is still limiting our confidence in understanding historical extratropical cyclone changes.

3.4.5.2. Causes Behind the Changes

Regarding possible causes of trends, 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, affecting storm tracks, winds and temperature patterns in both hemispheres. Simulated and observed changes in extratropical cyclones are broadly consistent, but an anthropogenic influence has not yet been detected, owing to large internal variability and problems due to changes in observing systems (Hegerl et al., 2007).

New studies have advanced the physical understanding of how stormtracks may respond to changes in the underlying surface condition and external forcing and seem to support the notion that average global cyclone activity may not be expected to change much under moderate greenhouse gas forcing. Idealized model simulations indicate that a uniform SST increase weakens (reduced cyclone intensity or density) and shifts the stormtrack poleward (Kodama and Iwasaki, 2009), and strengthened SST gradients near the subtropical jet may lead to a meridional shift in the stormtrack either towards the poles or the equator depending on the location of the SST gradient change (Brayshaw et al., 2008). By varying the longwave optical thickness as a proxy for changes in greenhouse gasses, O'Gorman and Schneider (2008) found that eddy kinetic energy is fairly insensitive to changes in radiative forcing near the present climate. These idealized experiments are consistent with the single model study of Bengtsson, et al., (2009) using a higher resolution AGCM.

Large-scale circulation anomalies and cyclone activity are closely connected. Several new studies confirmed that positive (negative) NAM/NAO corresponds to stronger (weaker) Atlantic/European cyclone activity (e.g., Chang, 2009; Pinto et al., 2009). However, studies using long historical records also seem to suggest that some of these links are intermittent (Hanna et al., 2008; Matulla et al., 2008; Allan et al., 2009). This possible nonstationary relationship between cyclone activity and NAO has been linked to interdecadal shifts in the location of the positions of the NAO pressure centers (Vicente-Serrano and Lopez-Moreno, 2008; Zhang et al., 2008b). Cyclone activity in Canada was found to closely co-vary with the states of NAO, the PDO, and the ENSO (Wang et al., 2006a). North Pacific cyclonic activity has been linked to tropical SST anomalies (NINO3.4) and PNA (Eichler and Higgins, 2006; Favre and Gershunov, 2006; Seierstad et al., 2007), showing that the PNA and NINO3.4 influence storminess and in particular over the eastern north Pacific. During El Niño events, there is an equatorward shift in storm tracks in the North Pacific basin, as well as an increase of storm track activity along the U.S. East Coast. Seierstad et al., (2007) noted that the relationship between NAO and storminess may to a large extent be accounted for by a basic relation between storminess and the local mean sea level pressure, indicating that the cause and effect of the association between the NAO and cyclonic activity is unclear. On the other hand they identified the PNA to be an important non-local factor for storminess north of the Aleutian Low. In the Southern Hemisphere, cyclone activity is related to the SAM with more cyclones around Antarctica when the SAM is in its positive phase, but more cyclones toward midlatitudes when the SAM is in its negative phase. More recent studies support this notion (Pezza and Simmonds, 2008). Additionally, more intense (and fewer) cyclones seem to occur when the PDO is strongly positive and vice versa (Pezza et al., 2007).

In summary, some changes in extratropical cyclones are related to variations in the modes of variability discussed in Sections 3.4.2 and 3.4.3. AR4 noted that observed changes in NAM and SAM are inconsistent with simulated internal variability (Hegerl et al., 2007). Anthropogenic influence on the sea level pressure distribution has also been detected in individual seasons (Giannini et al., 2003; Gillett et al., 2005; Wang et al., 2009c). Thus changes in these modes of variability may be affecting changes in extratropical cyclone occurrence. Some evidence has been found for changes in atmospheric storminess. 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 (Wang et al., 2009c). However, they note that climate models generally simulate smaller changes than observed and also appear to under-estimate the internal variability, reducing the robustness of their detection results.

Improved physical understanding of how stormtracks may respond to changes in SSTs and increased greenhouse gases (Deser et al., 2007; Brayshaw et al., 2008; Semmler et al., 2008; Kodama and Iwasaki, 2009) strengthen the notion that anthropogenic forcing may cause regional changes in both number of extratropical cyclones and intensity. Though the trend pattern in atmospheric storminess and ocean wave height contains a detectable response to anthropogenic forcing, it is still not possible to separately detect the effects of different external forcings. This new evidence has strengthened but does not alter the AR4 assessment that it is *likely* that anthropogenic forcing has contributed to the changes in extratropical storm tracks, because simulated and observed changes in extratropical cyclones are broadly consistent, but that a quantitative anthropogenic influence has not yet been detected formally, owing to large internal variability and problems due to changes in observing systems.

3.4.5.3. Projected Changes and Uncertainties

The AR4 reports that for 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), 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). Idealized studies (e.g., Deser et al., 2007; Lorenz and DeWeaver, 2007; Brayshaw et al., 2008; O'Gorman and Schneider, 2008; Kodama and Iwasaki, 2009) and diagnostic studies (Laine et al., 2009; Lim and Simmonds, 2009) on the response of extratropical cyclone changes to changes in radiative forcing or surface characteristics has provided new insight that can be used to understand the different model responses, but in depth analysis of changes in physical mechanisms related to cyclone changes in coupled climate models is still limited, and the inter-model differences are not well understood. This is complicated by the fact that studies use different analysis techniques, different physical quantities, different thresholds and different atmospheric vertical levels to represent cyclone activity and storm tracks (Raible et al., 2008). This diversity highlights different aspects of the cyclones, but makes it difficult to combine the results into a common view of future extratropical cyclone changes.

The Northern Hemisphere poleward shift in the stormtrack is supported by post-AR4 studies (Lorenz and DeWeaver, 2007). However, the strength of the poleward shift is often seen more clearly in upper-level mean quantities such as monthly zonal winds in 300hPa than in low-level transient parameters. Using bandpassed mean sea level pressures from 16 AR4 coupled GCMs, Ulbrich et al., (2008) show a wintertime poleward shift of stormtrack activity in some regions. It should be noted that other studies indicate that the poleward shift is less clear when models including a full stratosphere (Huebener et al., 2007) and ozone recovery (Son et al., 2008) are used. Post AR4 single model studies support the projection of a reduction in mid-latitude cyclones averaged over each 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 are found to be statistically significant by Bengsston et al., (2009), although they are accompanied by significant increases in total and extreme precipitation.

Models tend to show a northern movement of the North Pacific storm track (Loeptien et al., 2008; Ulbrich et al., 2008; Favre and Gershunov, 2009). However, the exact geographical pattern of cyclone frequency anomalies exhibits large variations across models. Some show indications of increased frequency along the U.S. west coast (Teng et al., 2008; Laine et al., 2009) while others show opposite results (Favre and Gershunov, 2009).

The large-scale response of cyclones in the North Atlantic is less clear than over the North Pacific. While some models exhibit a northward movement of the stormtracks (Pinto et al., 2007; Teng et al., 2008; Long et al., 2009; Orsolini and Sorteberg, 2009) others show more of an eastward extension (Ulbrich et al., 2008; Laine et al., 2009). In contrast, Huebner et al., (2007) report a southward shift in the North Atlantic stormtrack using a coupled model with a full stratosphere. Models showing a northward movement of the stormtrack tend to report a reduction in cyclone frequency along the Canadian east coast (Bengtsson et al., 2006; Watterson, 2006; Pinto et al., 2007; Teng et al., 2008; Long et al., 2009) consistent with changes observed during 1958–2001, reported by Wang et al., (2006a). A more detailed analysis of the AR4 MME for Europe, indicates an increase of between 18 and 62% in the number of storm days (the increase varies according to the definition of storminess and one model shows a decrease) associated with increased frequency of westerly flow (Donat et al., 2009). The mean intensity of storm cyclones increases by about 10% in the Eastern Atlantic, close to the British Isles and into the North Sea – increases which are also reflected in wind speed changes in these regions (Section 3.3.3).

In depth analysis of mechanisms responsible for projected regional changes in cyclone density and intensity are few. Using two coupled climate models, Laine et al., (2009) indicate that the primary cause for synoptic activity changes at the western end of the storm tracks is related to the baroclinic conversion processes linked to mean temperature gradient changes in localized regions of the western oceanic basins. Further downstream changes in latent heat release during the developing and mature stages of eddy are also important. They indicate that changes in diabatic process may be amplified by the upstream synoptic changes (stronger (weaker) baroclinic activity in the west gives stronger (weaker) latent heat release downstream).

New results on Southern Hemisphere cyclones confirm the previously projected poleward shift in stormtracks under increased greenhouse gases (Lim and Simmonds, 2009). They report a reduction of Southern Hemisphere extratropical cyclone frequency and intensity in midlatitudes but a slight increase at high latitude. The midlatitude changes were attributed to the tropical upper tropospheric warming enhancing static stability which decreases baroclinicity while an increased meridional temperature gradient in the high latitudes may be responsible for the increase of cyclone activity in this region (Lim and Simmonds, 2009).

In summary, it is *likely* that future anthropogenic climate change may influence cyclone activity through its impact on upper and lower level baroclinity and diabatic heating. A reduction in the number of mid-latitude cyclone averaged over each hemisphere is likely and it is *more likely than not* that high-latitude cyclone number and intensity will increase. It should be noted that the projected changes are fairly modest compared to interannual variability.

Regional changes may be substantial, but there is little consistency between models on the geographical pattern of cyclone activity changes. This leads to lower confidence in region-specific projections. The geographical pattern of modelled response in cyclone activity to various forcing is likely to be influenced by the individual model's structure of intrinsic modes of variability (Branstator and Selten, 2009) as well as details in the modelled changes in local baroclinicity and diabatic changes. However, models tend to show a poleward shift over the Southern Hemisphere, and a poleward and eastward shift of the North Pacific extratropical cyclones. Changes in low-level cyclone activity over the North Atlantic are less consistent, with some models showing an eastward extension while others have a poleward shift. New diagnostic studies (Laine et al., 2009; Lim and Simmonds, 2009) on the response of extratropical cyclone changes to changes in radiative forcing or surface characteristics has provided new insight that can be used to understand the different model responses, but in depth analysis of changes in physical mechanisms related to cyclone changes in coupled climate models is still limited, and the inter-model differences are not well understood. This is further complicated by the fact that studies use different analysis techniques, different physical quantities, different thresholds and different atmospheric vertical levels to represent cyclone activity and storm tracks (Raible et al., 2008). This diversity highlights different aspects of the cyclones, but makes it difficult to combine the results into a common view of future extratropical cyclone changes.

3.5. Observed and Projected Impacts on the Natural Physical Environment

3.5.1. Droughts

Drought is generally caused by 'a period of abnormally dry weather sufficiently prolonged for the lack of precipitation to cause a serious hydrological imbalance' (Heim Jr, 2002; IPCC, 2007a, glossary) and has been defined from different perspectives, e.g., meteorological drought related to deficit of precipitation, agricultural drought related to root zone soil water balance, or hydrological drought related to streamflow, lake and groundwater levels (e.g., Heim Jr, 2002). While lack of precipitation (i.e., meteorological drought) is often the primary precondition (see above definition), increased evapotranspiration (e.g., Easterling et al., 2007; Corti et al., 2009) as well as preconditioning (pre-event soil moisture and/or groundwater storage) are critical factors that can contribute to the emergence of agricultural and hydrological drought (Figure 3.11). 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 are often used to infer changes in drought conditions. These proxies include indices such as the Palmer Drought Severity Index (PDSI) (Palmer, 1965) or the Standard Precipitation Index (SPI) (McKee et al., 1993; Lloyd-Hughes and Saunders, 2002), land-surface model simulations (e.g., Sheffield and Wood, 2008), and paleoclimate proxies such as tree rings. Hence, drought indices often integrate temperature, precipitation and other variables, but may be problematic when not integrating all necessary information (Nicholls and Alexander, 2007). In order to understand the impact of droughts (e.g., on crop yields, general ecosystem functioning, etc.), the timing, the duration and intensity need to be characterized. The maximum number of consecutive dry days is often used as an overall drought index for a whole year, while other indices such as the PDSI characterize specific situations within a year. Other weather elements may interact to increase the impact of droughts (see also Figure 3.11): Enhanced air temperature leads to enhanced evaporative demand, as does enhanced wind speed. Moreover, climate phenomena such as monsoons (Section 3.4.1) and ENSO (Section 3.4.2) affect changes in drought occurrence in some regions. Hence, drought is a complex phenomenon that is strongly affected by other extremes considered in this Chapter. Moreover, via land-atmosphere interactions, drought also has the potential to feedback and exacerbate other weather and climate elements such as temperature and precipitation (Koster et al., 2004b; Seneviratne et al., 2006a) (see also Section 3.1.5 and Box 3.4).

3.5.1.1. Observed Changes

The AR4 reports that very dry areas (PDSI < -3) more than doubled in extent since 1970 on the global scale (Trenberth et al., 2007). However from a paleoclimate perspective recent droughts are not unprecedented with severe “mega droughts” reported in the paleoclimatic record for Europe, North America and Australia. 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 (often associated with very warm air temperature) have occurred (Breda and Badeau, 2008; Kallis, 2008); hence the current situation is not unprecedented.

INSERT FIGURE 3.11 HERE

Figure 3.11: Processes and interactions involved in meteorological, agricultural, and hydrological droughts (red: positive impacts; blue: negative impacts). Dashed lines denote indirect feedbacks of soil moisture on temperature and precipitation. For simplicity, the role of interactions with other variables of the Figure (e.g., evapotranspiration, relative humidity) in these feedbacks, and feedbacks of soil moisture to other meteorological variables (e.g., circulation anomalies) are not highlighted.

Globally, 2–3 fold increases of area affected by extreme or severe droughts have been inferred by a modelling study which reproduced the global drying trend (PDSI) since the 1950s (Burke et al., 2006). This trend in the PDSI proxy is largely affected by the changes in temperature, not precipitation. Beniston (2009) found a strong increase in warm-dry mode over all central-southern (incl. maritime) Europe via a quartile-analysis from mid- to the end of the 20th century. Trends of decreasing precipitation and discharge are consistent with increasing salinity in the Mediterranean, indicating a trend towards fresh water 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). The exceptional 2003 summer heat wave on the European continent (see Section 3.3.1) was also associated with a major drought, as could be inferred from satellite measurements (Andersen et al., 2005), model simulations (Fischer et al., 2007a; Fischer et al., 2007b), and impacts on ecosystems (Ciais et al., 2005; Reichstein et al., 2007). In the U.S., droughts are becoming more severe in some regions, but there are no clear trends for North America as a whole (Kunkel et al., 2008; Wang et al., 2009b), with an observational record dating back to 1895. The most severe droughts have occurred in the 1930s in the U.S. and Canada, while in Mexico the 1950s and late 1990s were the driest periods. Recent regional trends towards more severe drought conditions are observed over southern and western Canada, Alaska and Mexico. Furthermore, Easterling et al., (2007) showed that the increase in precipitation in the continental USA has masked an increasing tendency for more droughts due to increasing temperatures. For the Amazon, repeated strong droughts have been occurring in the last decades but no particular trend has been reported. The 2005 drought in Amazonia is however considered the strongest in the last century both from precipitation records and water storage estimates via satellite (measurements from the Gravity Recovery and Climate Experiment (GRACE)), (Chen et al., 2009). 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 Argentinian Andean Cordillera (Le Quesne et al., 2006; Le Quesne et al., 2009). Changes in drought patterns have been reported for the monsoon regions of Asia and Africa with variations at the decadal timescale (e.g., Janicot, 2009). In the Sahel, recent years are characterized by a 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 monsoon regions, related to warming of the tropical oceans, and variability related to the El Niño–Southern Oscillation.

In conclusion, the assessment of the AR4 that since the 1950s and in particular the 1970s it is *likely* that more intense and longer droughts have occurred over larger areas and generally in the Northern Hemisphere (Trenberth et al., 2007) has been supported by post-AR4 research analyzing regional drought.

3.5.1.2. Causes Behind the Changes

AR4 (Hegerl et al., 2007) also concludes 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 multiple lines of evidence: a detection study identified an anthropogenic fingerprint in a global PDSI data set with high significance (Burke et al., 2006), and studies of some regions indicate that droughts in those regions are linked either to SST changes that, in some instances, may be linked to anthropogenic aerosol forcing (e.g., Sahel) or to a circulation response to anthropogenic forcing (e.g., southwest Australia).

There is now a better understanding of the potential role of land-atmosphere feedbacks versus SST forcing for droughts (e.g., Schubert et al., 2008a; Schubert et al., 2008b) as well as of potential impacts of land use changes (Deo et al., 2009), but large uncertainties remain in the field of land surface modelling and land-atmosphere interactions, in part due to lack of observations (Seneviratne et al., 2010) and inter-model discrepancies (Koster et al., 2004b; Dirmeyer et al., 2006; Pitman et al., 2009). Nonetheless, a new set of climate modelling studies show that U.S. drought response to SST variability is consistent with observations (Schubert et al., 2009). 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), but this result is still debated. Additionally, model-dependent results regarding past trends, which could point to deficiencies in the relevant parameterizations, cannot be credibly compared with observations, due to the lack of reliable globally-available runoff and evapotranspiration observations (e.g., Peel and McMahon, 2006; Teuling 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; Zhang et al., 2007a). The change in the pattern of global precipitation in the observations and in model simulations are also consistent with theoretical understanding of hydrological response to global warming that wet regions become wetter and dry regions drier in a warming world (Held and Soden, 2006). However, the recent U.S. drought that began in the 2005/2006 winter in the southeastern U.S. is different from what would be expected from model projected anthropogenic climate change in this region: The drought was caused by a reduction in precipitation (with simultaneous reduction in evaporation), but models project an increase in precipitation minus evaporation (Seager et al., 2009). Though these new studies have improved the understanding of the mechanisms leading to drought, there is still not enough evidence to alter the AR4 assessment, in particular given the associated observational data issues (Section 3.2.1).

3.5.1.3. Projected Changes and Uncertainties

AR4 model projections indicate an increase in droughts in particular in subtropical and mid-latitude areas (Christensen et al., 2007). An increase in dry spell length and frequency is considered very likely over the Mediterranean area, 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 are considered likely to increase in summer over many mid-latitude continental interiors (e.g., central and southern Europe, the Mediterranean), in boreal spring and dry periods of the annual cycle over Central America. More recent global and regional climate simulations support the projections from AR4, as summarized in the following paragraphs.

Particular care is needed in intercomparing 'drought' projections since very many different definitions are employed (corresponding to different types of droughts), from simple climatic indices such as maximum consecutive dry days to more complex indices of hydrological and agricultural drought (see above). A distinction also needs to be made between short-term and longer-term events. Blenkinsop and Fowler (2007), for example, demonstrate that while an RCM ensemble indicate an increase in short-term summer drought over most of the UK, the longer (multi-season) droughts are projected to become shorter and less severe (although uncertainties in the latter projections are large – see below).

Burke and Brown (2008) project an increase in the global area affected by extreme drought from 1% to 21% over the 21st century. However, the changes are dependent on the definition of the drought index. Areas where drought is indicated to increase across all indices examined include the Mediterranean, Amazonia and southern Africa. These results are consistent with findings by Sillmann and Rockner (2008) who show increasing dry spells in regions which are already affected by drought today. The consecutive dry days index increases significantly around the Mediterranean Sea, Australia and southern Africa, as well in the north-eastern part of South America and the Pacific coast of Central and South America. One GCM-based study suggests one to three weeks of additional dry days for the Mediterranean by the end of the century (Giannakopoulos et al., 2009).

Regional climate simulations over Europe also highlight the Mediterranean region as being affected by more severe droughts (Giorgi, 2006; Beniston et al., 2007; Mariotti et al., 2008; Planton et al., 2008). Mediterranean droughts are likely to start earlier in the year and last longer. Also increased variability during the dry and warm season is projected (Giorgi, 2006). 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 consensus of most climate-model projections regarding a reduction of cool season precipitation across the U.S. southwest and northwest Mexico (Christensen et al., 2007) with more frequent multi-year drought in the American southwest (Seager et al., 2007). 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) find increases in consecutive dry days, although consensus among the models is only found in the interior.

Increased confidence in modelling drought stems from consistency between models and satisfactory simulation of drought indices during the past century (Sheffield and Wood, 2008; Sillmann and Roeckner, 2008). Inter-model agreement is stronger for long-term droughts and larger spatial scales, while local to regional and short-term precipitation deficits are highly spatially variable and much less consistent between models (Blenkinsop and Fowler, 2007). Lack of complete knowledge of the physical causes of meteorological droughts, and links to the large-scale atmospheric and ocean circulation are 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 and Sobel, 2009). 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 to more drought-like conditions (Nicholls, 2009).

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), and possibly - as only recently highlighted – also feedbacks between droughts, fires and aerosols (Bevan et al., 2009).

Furthermore, the development of "agricultural drought" that results from complex interactions of 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 behaviour of plants' transpiration, growth and water-use efficiency under enhanced atmospheric CO_2 concentrations, which could potentially have major impacts on the hydrological cycle (Betts et al., 2007), but are not well established yet (Hungate et al., 2003; Piao et al., 2007; Bonan, 2008; Teuling et al., 2009).

3.5.2. Floods

Floods are natural physical impacts produced by a transient high water level along a river channel, lake or on a sea coast. When humans are impacted, floods can become "natural disasters." Floods include river floods, flash floods, urban floods, sewer floods, coastal floods, and glacial lake outburst floods (GLOFs). The main causes of floods are intense and/or long-lasting precipitation, snow/ice melt, a combination of previous types, dam break (e.g., glacial lakes), reduced conveyance due to ice jams or landslides, or by a local intense storm (Smith and Ward, 1998). Climate-related floods depend on precipitation intensity, volume, duration, timing, phase (rain or snow), antecedent conditions of rivers and their drainage basins (e.g., presence of snow and ice, soil character and status, wetness, rate and timing of snow/ice melt, urbanisation, existence of dikes, dams, and/or reservoirs) (Bates et al., 2008), while along coastal areas flooding may be associated with storm surge events. This chapter focuses on the spatial, temporal and seasonal changes in high flows and peak discharge in rivers related to climate change, while the impact of floods on human society and ecosystems and related changes are discussed in Chapter 4. Coastal floods are described as a part of the section on extreme sea level and coastal impacts (Section 3.5.5). GLOFs are discussed in Section 3.5.6.

3.5.2.1. Observed Changes

The AR4 concluded that no gauge-based evidence had been found for climate-related trend in the magnitude/frequency of floods during the last decades (Rosenzweig et al., 2007), while it noted that flood damages were increasing (Kundzewicz et al., 2007) and that an increase in heavy precipitation events was already *"likely"* in the late 20thcentury trend (Trenberth et al., 2007). The AR4 also highlighted a catastrophic flood that occurred along several central European rivers in 2002 in a similar context; no significant trend in flood occurrences was found but the trend in precipitation variability was indicative of an enhancement of flood occurrence (Trenberth et al., 2007). On the other hand, the AR4 concluded that abundant evidence was found for an earlier occurrence of spring peak river flows in snow-dominated regions (Rosenzweig et al., 2007). Research subsequent to the AR4 still does not show clear and widespread evidence of observed changes in flooding at the global level based on instrumental records, except for the earlier spring flow in snow-dominated regions.

Worldwide instrumental records of floods at gauge stations are limited in spatial coverage and in time, and only a limited number of gauge stations spans more than 50 years, and even fewer over 100 years (Rodier and Roche, 1984, see also Section 3.1.1.2). Pre-instrumental flood data sources can be obtained from documentary records (archival reports, in Europe continuous over the last 500 yrs) (Brazdil et al., 2005), and from geological indicators known as paleofloods (sedimentary and biological records over centuries to millennia scales) (Kochel and Baker, 1982). Analysis of these centennial past flood records have revealed that (1) flood magnitude and frequency are very sensitive to subtle alterations in atmospheric circulation, with greater sensitivity on largest “rare” floods (50-year flood and higher) than on smaller frequent floods (2-year floods) (Knox, 2000; Redmond et al., 2002); (2) high interannual and interdecadal variability is 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 analogues of unusual large floods, as the ones recorded recently, sometimes claimed to be the largest on record. For example, pre-instrumental flood data shows 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.

Although flood trends might be seen in the north polar region and in northern regions where temperature change affects snowmelt or ice cover, widespread evidence of this (except for earlier spring flow) is not found. For example, Cunderlik and Ouarda (2009) reported that snowmelt spring floods come significantly earlier in the southern part of Canada, and one fifth of all the analyzed stations show significant negative trends in the magnitude of snowmelt floods over the last three decades. On the other hand, there is no evidence of widespread common trends in the magnitude of extreme floods based on the daily river discharge of 139 Russian gauge stations for the last few to several decades, while a significant shift to earlier spring discharge is found as well (Shiklomanov et al., 2007).

In Europe, significant upward trends in the magnitude and frequency of floods were detected in a considerable fraction of river basins in Germany for the period 1951–2002, particularly in western, southern, and central Germany and particularly for winter floods, although there is no ubiquitous increase of floods all over Germany (Petrow and Merz, 2009). This is apparently in agreement with an upward trend in annual and winter flood discharges since 1984 in the Meuse river (northwest Germany, The Netherlands, and Belgium) and its tributaries (except Geul River) (Tu et al., 2005). Similar results are found by Allamano et al., (2009) for the Swiss Alps where they found a significant increase of flood peaks during the last century. In contrast, a slight decrease in winter floods and no change in summer maximum flow were reported in east and northeast Germany and in the Czech Republic (Elbe and Oder rivers) (Mudelsee et al., 2003). In France there is no evidence of a widespread trend in annual flow maxima over the last four decades, although there is evidence of a decreasing flood frequency trend in the Pyrenees, and increasing annual flow maxima in the northeast region (Renard et al., 2008). In Spain, southern Atlantic catchments showed a downward trend in flood magnitude and frequency, whereas in central and northern Atlantic basins no significant trend in frequency and magnitude of large floods is observed (Benito et al., 2005). Flood records from a network of catchments in the UK showed significant positive trends over the past four decades in high-flow indicators primarily in maritime-influenced, upland catchments in the north and west of the UK (Hannaford and Marsh, 2008), although in previous studies such changes were not so obvious (Robson et al., 1998). Although there are relatively abundant studies for rivers in Europe as described above, a continental scale assessment for Europe is difficult to obtain because geographically organized patterns are not seen.

The number of analyses for rivers in the other parts of the world based on the stream gauge records is limited. The limited examples in Asia are as follows; annual flood maxima of the lower Yangtze region shows an upward trend over the last 40 years (Jiang et al., 2008), an increasing likelihood of extreme floods during the last half of the century is found for the Mekong river (Delgado et al., 2009), and both upward and downward trends were detected 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, 2010). 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.

In summary, except for the abundant evidence for an earlier occurrence of spring peak river flows in snow-dominated regions (*likely*), no clear and widespread observed evidence is found in the AR4 and research subsequent to the AR4. Besides, instrumental records of floods at gauge stations are limited in spatial coverage and in time, which limits the number of analyses. Pre-instrumental flood data can provide information for a longer period, but these data are also limited.

3.5.2.2. Causes Behind the Changes

Floods are affected by various characteristics of precipitation, such as intensity, duration, amount, timing, phase (rain or snow). They are also affected by drainage basin conditions such as water levels in the rivers, presence of snow and ice, soil character and status (frozen or not, saturated or unsaturated), wetness (soil moisture), rate and timing of snow/ice melt, urbanisation, existence of dikes, dams, and reservoirs (Bates et al., 2008). A change in the climate physically changes many of these factors affecting floods 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 complicated and difficult.

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 and from those caused by human regulation of the river systems. River engineering and land use may have altered flood probability. Many dams have a function to reduce flood. However, the largest and most pervasive contributors to increased flooding on the Mississippi River system over the past 100-150 years were wing dikes and related navigational structures, followed by progressive levee construction (Pinter et al., 2008). 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 possible causes for changes in floods were assessed in the AR4 report. Cause-and-effect between external forcing and changes in floods has not been established. However, anthropogenic influence has been detected in the environments that affect floods, such as aspects of the hydrological cycle (e.g., Zhang et al., 2007a; see also Section 3.3.2) including precipitation and atmospheric moisture. Anthropogenic influence is also clearly detected in streamflow regimes in the western USA (Barnett et al., 2008; Hidalgo et al., 2009).

In climates where seasonal snow storage and melting plays a significant role in annual runoff, the hydrologic regime is affected by changes in temperature. In a warmer world, a smaller portion of precipitation will fall 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 U.S. (Regonda et al., 2005; Clow, 2010) and in Canada (Zhang et al., 2001), along with an earlier breakup of river ice in Russian Arctic rivers (Smith, 2000). The observed trends toward earlier timing of snowmelt-driven streamflows in the western U.S. since 1950 are detectably different from natural variability (Barnett et al., 2008; Hidalgo et al., 2009). It is unclear if greenhouse gas emissions have affected the magnitude of the snowmelt flood peak, but projected warming may result in an increase in the spring river discharge where winter snow depth increases (Meehl et al., 2007b) or a decrease in spring flood peak (Hirabayashi et al., 2008b; Dankers and Feyen, 2009).

There is still a lack of studies identifying an influence of anthropogenic warming on peak streamflow for regions with little or no snowfall because of uncertainty in the observed streamflow data and low signal to noise ratio. However, evidence has emerged that anthropogenic forcing may have influenced the likelihood of a rainfall-dominated flood event in the UK (Pall et al., 2010). Additionally, it has been projected for many rain-dominated catchments that flow seasonality will increase, with higher flows in the peak flow season but little change in the timing of the peak or low flows (Kundzewicz et al., 2007). More recent hydrological simulation studies also show an increase in the probability of flooding due to a projected rainfall increase in rain-dominated catchments (e.g., humid Asia) where short-term extreme precipitation and long-term precipitation are both projected to increase (e.g., Asokan and Dutta, 2008; Dairaku et al., 2008; Hirabayashi et al., 2008b).

In summary it is *more likely than not* that anthropogenic forcing leading to enhanced greenhouse gas concentrations has affected floods because they have detectably influenced components of the hydrological cycle such as mean precipitation (Zhang et al., 2007a), heavy precipitation (see Section 3.3.2), and snowpack (Barnett et al., 2008). Floods are also projected to change in the future due to anthropogenic warming (see Section 3.5.2.3), but the magnitude and even the sign of this anthropogenic influence have yet not been detected/attributed in scientific literature, and the exact causes for regional changes in floods cannot be clearly ascertained. It is *likely* that anthropogenic influence has resulted in earlier spring flood peaks in snow-melting rivers; the observed earlier spring runoff is consistent with expected change under anthropogenic forcing. It should be noted that these two assessments are based on expert judgement rather than a formal model-based attribution study, although Pall et al., (2010) do provide more direct evidence of an anthropogenic influence on a specific extreme flood event.

3.5.2.3. Projected Changes and Uncertainties

The number of studies that showed the projection of flood changes in rivers especially at a regional or a continental scale was limited when AR4 was published. A rare example was Milly et al., (2002) who, using monthly river discharge calculated from climate model outputs, demonstrated the changes (mostly increases) in 'large' floods at selected extratropical river basins larger than 20,000km^2.

The number of studies 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 demonstrated changes in the frequency and/or magnitude of floods in the 21st century at a large scale using daily river discharge calculated from RCM or GCM outputs and hydrological models at a regional or a continental scale. For Europe, most notable changes are projected to occur in northern and northeastern Europe in the late 21st century, but the results are varied. Three studies (Dankers and Feyen, 2008; Hirabayashi et al., 2008b; Dankers and Feyen, 2009) show a decrease in the probability of extreme floods, that generally corresponds to lower flood peaks, in northern and northeastern Europe because of a shorter snow season, while one study (Lehner et al., 2006) shows an increase in floods in the same region. Changes in floods in central and western Europe are less prominent and with not much consistency seen between the four studies. 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, which were implied in an earlier study (Manabe et al., 2004) that used annual mean runoff changes obtained from a coarse resolution GCM. This projected change was also implied in earlier studies by the changes in precipitation in monsoon seasons (e.g., Palmer and Räisänen, 2002).

Lehner et al., (2006) and Hirabayashi et al., (2008b) both showed the geographical distribution of changes in hydrological drought in a future warmer climate as well as the changes in floods. From this it is possible to identify regions which are projected to experience changes in hydrological floods and droughts. However, the results for Europe are not consistent between these two studies. Most of south and southeast Asia, tropical South America and Sahel are projected to suffer both from hydrological floods and droughts, but this result does not have high reliability because only one model was used (Hirabayashi et al., 2008b).

Projections of flood changes at a 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 like Asia (Asokan and Dutta, 2008; Dairaku et al., 2008), the Middle East (Fujihara et al., 2008), Africa and South America are very rare. Most projections for rain-dominated catchments are carried out because rainfall intensification, which is anticipated to cause more or more severe floods, is projected by climate models in regions where those catchments are located. Flood probability is generally projected to increase in such catchments, but uncertainty is still large in the changes in the magnitude and frequency of floods (Cameron, 2006; Kay et al., 2009). Earlier spring flooding is projected in snow-dominated catchments, but the change in the magnitude of spring flood also varies between projections.

It has been recently recognized that the choice of GCMs is the largest source of uncertainties in hydrological projections, and uncertainties from downscaling methods are of secondary importance (Graham et al., 2007; Leander et al., 2008; Kay et al., 2009; Prudhomme and Davies, 2009), although, in general, hydrological-model projections require downscaling and bias-correction of GCM outputs (e.g., precipitation and temperature). The choice of hydrological models is also of secondary importance (Kay et al., 2009). Nevertheless, uncertainty analysis in the hydrological projections is still in its infancy, and the results 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 above mentioned inconsistency between the projections of flood changes in snow-dominated regions in Europe (Lehner et al., 2006; Dankers and Feyen, 2008; Hirabayashi et al., 2008b; Dankers and Feyen, 2009) has been considered to be primarily due to differences in the downscaling and bias-correction methods applied in the different studies (Dankers and Feyen, 2009). Downscaling and bias-correction are also a major source of uncertainty in rain-dominated catchments (van Pelt et al., 2009).

In summary, the number of projections on flood changes is still limited at a regional and continental scale, and those projections often show some degree of uncertainty. Projections at a catchment/river-basin scale are also not abundant in the peer-reviewed scientific literature. In particular, projections for catchments except for Europe and North America are very rare. In addition, considerable uncertainty has remained in the projections of flood changes, especially regarding their magnitude and frequency. The exception is the robust projection of the earlier shift of spring peak discharge in snow-dominated regions. Therefore, it is currently difficult to make a statement on the confidence/likelihood of flood change projections due to anthropogenically induced climate change, except for the robustly projected earlier shift of spring floods (*likely*), because of insufficient reliability of climate models and downscaling methods.

3.5.3. Extreme Sea Levels

Extreme sea levels are caused by severe storms such as tropical or extratropical cyclones. The associated falling atmospheric pressure and strong winds can produce storm surges at the coast, which may be further elevated by coastal wave breaking which causes an onshore flux of momentum known as wave setup. Changes in extreme sea level may arise from changes in atmospheric storminess, (see sections 3.4.4 and 3.4.5) and will also occur as a result of mean sea level rise.

3.5.3.1. Observed Changes

The AR4 reported with 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 1.8 [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. Whether the faster rate of increase during the latter period reflected decadal variability or an increase in the longer term trend was not clear. However there is increasing evidence that the contribution to sea level due to mass loss from Greenland and Antarctica is accelerating (Velicogna, 2009). The total 20th-century rise was estimated to be 0.17 [0.12 to 0.22] m (Bindoff et al., 2007).

The AR4 reported that the rise in mean sea level and variations in regional climate led to a likely increase in trend of extreme high water worldwide in the late 20th century (Bindoff et al., 2007) and that it was *more likely than not* that humans contributed to the trend in extreme high sea levels (IPCC, 2007a). This conclusion was based on a number of studies of sea level extremes, the most geographically comprehensive being that of Woodworth and Blackman (2004) who found that increases in 99th percentile sea levels at 141 tide gauges across the globe since 1975 were mostly attributable to the trend in mean sea level. Since the AR4, several new studies have been undertaken. These studies provide further evidence that changes in extremes are related to trends in mean sea level and modes of variability in the regional climate. The overall assessment of these studies confirms but does not change the AR4 assessment.

Several studies since the AR4 report that trends in extreme sea level are broadly consistent with changes in mean sea level. Menendez and Woodworth (2010), using sea level records from 258 tide gauges across the globe, confirms the earlier conclusions of Woodworth and Blackman (2004) that there has been a trend in extreme sea levels globally, which has been more pronounced since the 1970's, and this trend is consistent with trends in mean sea level. Marcos et al., (2009) found changes in extreme sea levels in 73 tide gauges in the Mediterranean and the southern Atlantic Ocean since 1940 were consistent with mean sea level changes. Haigh et al., (2010), using an expanded and spatially more comprehensive sea level data set for the English Channel, concluded that extreme sea levels increased at all of the 18 sites, but at rates not statistically different from mean sea level rise.

A number of studies also highlight the additional influence of climate variability on extreme sea level trends. Menendez and Woodworth (2010) report that ENSO has a large influence on interannual variations in extreme sea levels since the 1970s throughout the Pacific Ocean and the monsoon regions. In southern Europe, Marcos et al., (2009) find that in addition to mean sea level changes, changes in extremes are also significantly negatively correlated with the NAO. A more localised study in the Camargue (Rhone Delta) region of southern France by Ullmann et al., (2007) concluded that maximum annual sea levels had risen twice as fast as mean sea level during the 20th century. Subsequent studies that have examined the role of changes in weather conditions in extreme sea level trends in this region find that while most extremes occur during particular weather patterns that are associated with the negative NAO phase (Ullmann and Moron, 2008) the increased frequency of sea surges in this region in the latter part of the 20th Century is due to an increase in southerly winds associated with a general rise in sea level pressure over central Europe over this period (Ullmann et al., 2008).

Abeysirigunawardena and Walker (2008) report that sea level trends from two tide gauge records over the period from 1939 to 2003 in Prince Rupert Sound on the north coast of British Columbia were twice that of mean sea level rise, the additional contribution being due to the strong positive PDO phase which has lasted since the mid-1970s. Cayan et al., (2008) reported increases in the frequency of exceedance of the 99.99th percentile sea level of 20-fold at San Francisco since 1915 and 30-fold at La Jolla since 1933 and also note that positive sea level anomalies of 10 to 20 cm often persisted for several months during El Niño events, which causes an increase in storm surge peaks.

In the Southern Hemisphere, Church et al., (2006b) examined changes in extreme sea levels before and after 1950 in two tide gauge records of approximately 100 years at Fort Denison and Fremantle on the east and west coasts of Australia respectively. At both locations a stronger positive trend is found in the 99.99 percentile sea level (the sea level which is exceeded by 0.01 per cent 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) note an increase in the number and duration of positive storm surges in the decade 1996 to 2005 compared to previous decades. However the relative contributions of mean sea level rise and changes in wind climatology due to a southward shift in the South Atlantic high are not quantified.

3.5.3.2. Causes Behind the Changes

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., 2006b; Ullmann et al., 2007; Abeysirigunawardena and Walker, 2008). Several studies also find that extreme sea levels are influenced by modes of climate variability (e.g., Abeysirigunawardena and Walker, 2008; Marcos et al., 2009; Menendez and Woodworth, 2010). These studies support 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.

3.5.3.3. Projected Changes and Uncertainties

The AR4 (Meehl et al., 2007b) projected sea level rise for 2090–2099 relative to 1980–1999. The estimated rise from ocean thermal expansion, glaciers and ice caps, and modelled ice sheet contributions is projected to be 18–59 cm with a 90% confidence range. An additional allowance to the sea level rise projections was made 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 10–20 cm of sea level rise 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.

The AR4 (Christensen et al., 2007) suggests that the dynamical downscaling step in providing forcing for regional surge (and correspondingly wave) models is robust (i.e., does not add to the uncertainty), but that the general low level of confidence in projected circulation changes from GCMs implies a substantial uncertainty in surge (and ocean wave) projections.

New studies carried out over the northern European region since the AR4, whose focus is on changes to storminess, have attempted to address uncertainties in extreme sea level changes using a large ensemble of simulations (e.g., Sterl et al., 2009) or address uncertainties due to scale issues by downscaling in RCM simulations (e.g., Wang et al., 2008), or both (Debernard and Roed, 2008). These studies project increases in storm surge height along the eastern North Sea coast, the Irish west coast, and the Irish Sea, consistent with earlier studies. However, the small number of studies and limited regional coverage of such studies do not provide a basis to change the AR4 assessment of projected extreme sea level changes as the ensemble of model simulations is still small and the results show considerable regional variations. Other studies have focused more on an exploration of scenarios of future changes in mean sea level in relation to changes in meteorological forcing and conclude that mean sea level rise will be the main factor in extreme sea level changes in the future (e.g., Harper et al., 2009; McInnes et al., 2009b; Brown et al., 2010).

Debernard and Roed (2008) investigated the effect of changing meteorological conditions on storm surges over Europe in several models under A2 and B2 greenhouse gas scenarios. Despite large inter-model differences, statistically significant differences between 2071-2100 and 1961-1990 include decreases in storm surge south of Iceland, and an 8-10% increase in the 99th percentile storm surge heights along the coastlines of the eastern North Sea and northwest of the British Isles. The changes relate mainly to changes in the winter season in the climate models.

Wang et al., (2008) examined storm surges in Irish coastal seas in 30-year time slices for the periods 1961-1990 and 2031-2060 from an A1B simulation downscaled by the Rossby Centre Regional Atmosphere model. The results show an increase in storm surge events around Irish coastal areas in the future time-slice, except along the south Irish coast. There is also a significant increase in the height of the extreme surges along the west and east Irish coasts, with most of the extreme surges occurring in wintertime.

Sterl et al., (2009) used a 17-member ensemble of A1B simulations from 1950 to 2100 to examine future changes to the 10000-year sea level height along the Dutch coastline. By concatenating the output from the 17 ensemble members over the model periods 1950-2000 and 2050-2100 into a single longer time series for each time slice, return periods were estimated with narrower uncertainties and no statistically significant change in the 10000 year return values of surge heights along the Dutch coastline were found during the 21st century. This was attributed to the fact that wind speed changes in the future climate were not associated with the surge-generating northerlies but rather southwesterlies. However, they stress that the result is based only on output from one climate model.

Other studies have undertaken a sensitivity approach and compared the relative impact on extreme sea levels of meteorological changes and mean sea level rise by perturbing the meteorological conditions which caused current climate storm surges. 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 (note that the lower end of this range was for wind decrease) produced an increase in sea levels that were 20 to 35% of that due to the upper end of the A1FI sea level rise scenario for 2070. Brown et al., (2010) investigated the relative impact of sea level rise and wind speed change on an extreme storm surge in the eastern Irish Sea. Both studies conclude that sea level rise has the greater potential to increase extreme sea levels 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 will be regionally dependant. For example current

GCMs are unable to realistically represent tropical cyclones. This has led to the use of alternative approaches for investigating the impact of climate change on storm surges in tropical Australia. For example, methods have been used that rely on the generation of synthetic cyclones whose characteristics are perturbed to represent projected future cyclone characteristics in this region (e.g., McInnes et al., 2003). 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, with mean sea level rise producing the larger contribution to changes in future sea level extremes.

3.5.4. Waves

Severe waves can damage and destroy coastal infrastructure and threaten the safety of coastal inhabitants. 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. Properties of waves that influence these processes include wave height, direction, and period although to date studies of past and future wave climate changes have tended to focus on wave height parameters such as 'significant wave height' (SWH), which is the height from trough to crest of the highest one third of waves.

3.5.4.1. Observed Changes

The AR4 reported statistically significant positive trends in SWH 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 (Trenberth et al., 2007), based on trends in SWH from voluntary observing ship data (VOS) (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 trends in the north Atlantic and north Pacific (Weisse and Günther, 2007; Wang et al., 2009b). Positive trends in wave height are also found along the U.S. east and west coasts (Allan and Komar, 2006; Komar and Allan, 2008; Menendez et al., 2008), and the southern ocean (Hemer et al., 2010). Wave climate studies on the U.S. west coast have found a positive correlation between wave height and El Niño (Allan and Komar, 2006; Adams et al., 2008; Menendez et al., 2008). However, the different sources of wave information (i.e., direct measurements, satellite observations and reanalysis products) and the focus of the studies on different geographical regions contribute to uncertainties for observed wave climate changes. Until more studies are completed and the relationship between different wave data products are better understood, a stronger assessment will not be possible.

Generally confirming previously reported regional trends, Wang et al., (2009b) found that wave heights 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 ERA-40 reanalysis products. Weisse and Günther (2007) analysed extreme wave conditions from a regional North Sea hindcast (1958–2002) and found a positive trend in severe wave heights in the southern North Sea from 1958 to the early 1990s, followed by a declining trend since. Along the UK North Sea coast, a reduction in severe wave conditions was observed over much of the hindcast period.

However trends at particular locations may be influenced by local factors. For example, Suursaar and Kullas (2009) reported a slight decreasing trend in mean SWHs from 1966–2006, while the frequency and intensity of high wave events showed rising trends. These changes were associated with a decrease in local average wind speed, but an intensification of the westerly winds and storm events.

On the North American Atlantic coast, Komar and Allan (2008) found a statistically significant increasing trend in wave heights of 0.059 m/yr at Charleston, South Carolina during the summer months since the 1970s with lower but statistically significant trends at wave buoys further north. The positive trends are associated with an increase in intensity and frequency of hurricanes over the period. In contrast, the waves measured during the winter, generated by extratropical storms, were not found to have experienced a statistically significant change.

Positive trends in wave height were also found by Allan and Komar (2006) and Menendez et al., (2008) along the U.S. west coast based on 25 and 22 years of wave records respectively. Both studies find a strong relationship between wave height and El Niño which is also found by Adams et al., (2008) further south over the Southern California Bight using a 50 year wave hindcast. Similarly, over the western north Pacific, Sasaki & Toshiyuki (2007) find that the 90th percentile of the summertime SWH which is associated with typhoons in eastern Asia was strongly correlated with cyclonic circulation in the western North Pacific and warm SST anomalies in the Nino 3.4 region.

Hemer et al., (2010) find a positive trend in wave height mainly confined to the region south of 45°S over the period 1998–2000 relative to 1993–1996 based on satellite data whereas extensive positive trends are seen over much of the Southern Hemisphere in the ERA-40 waves reanalysis over the same period.

3.5.4.2. Causes Behind the Changes

Wave climate studies point to strong links in wave climate and natural modes of climate variability (e.g., Allan and Komar, 2006; Adams et al., 2008). However, only one study (Wang et al., 2009c) detects 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 wave heights in the boreal winter over the past half century in the high-latitudes of the Northern Hemisphere (especially the northeast North Atlantic), and a decrease in more southerly northern latitudes.

3.5.4.3. Projected Changes and Uncertainties

The AR4 projected a general tendency for more intense but fewer storms outside the tropics, with a tendency towards higher ocean waves in several regions (Meehl et al., 2007b), and increases in wave height were projected for most of mid-latitude areas analysed, including the north seas (Christensen et al., 2007) but with low confidence due to the low confidence in projected changes in mid-latitude storm tracks and intensities.

Since the AR4, there have been several studies that have developed regional (Andrade et al., 2007; Leake et al., 2007; Debernard and Roed, 2008; Grabemann and Weisse, 2008; Lionello et al., 2008; Hemer et al., 2009) and global (Mori et al., 2009) wave climate projections. Forcing conditions are typically obtained for a few selected emission scenarios (typically B2 and A2, representing low-high ranges) from a single or at most three coarse resolution GCMs. While these additional downscaling studies in more climate model simulations provide further evidence for projected increases in wave height in some regions such as the eastern North Sea coast, they do not change the low level of confidence in the findings due to the small number of climate models upon which the studies are based.

Wang et al., (2009a) compared dynamical and statistical downscaling methods for estimating seasonal statistics of SWH. They found that dynamical downscaling approaches, which have been common practice over the past few years, have not adequately resolved the issue of model variability biases. They found that the dynamical approach for downscaling was poorer than the statistical approach in terms of reproducing the observed climate and interannual variability of the wave heights. They also reported a better reproduction of the interannual variability of seasonal statistics (including extremes) when using high temporal resolution forcing data, stressing the importance of higher resolution data from climate model outputs.

Mori et al., (2009) forced a global wave model with the 20km high-resolution atmospheric MRI/JMA GCM, for three time slices (1979-2004, 2015-2031, and 2075-2100) following the A1B scenario. They project higher maximum wave heights in mid and high-latitudes. Lower mean wave heights are projected for mid-latitudes. The projected changes are qualitatively consistent with global wave projections carried out by Wang and Swail (2006b) and are also consistent with patterns of extreme wind change reported in Gasteneau and Soden (2009) and Figure 3.10.

Debernard and Roed (2008) examined wave climate changes around Europe in several models under A2, B2 and A1B greenhouse gas scenarios. They project a 6% decrease in 99th percentile SWH from 1960-1990 to 2070-2100 southwest of Iceland. A 6-8% increase in the annual 99th percentile SWH is projected along the eastern coast of the North Sea and the Skagerrak. An increase in the annual 99th percentile SWH is also projected along the west coast of the British Isles, was found to be associated to a change in the winter storm track.

Grabemann and Weisse (2008) used a regional wave model to downscale two GCMs under A2 and B2 emission scenarios. An increase of up to 18% from the ensemble mean long-term 99th percentile SWH is projected for 2071-2100 compared to 1961-1990 in the North Sea, except for off the English coast. This is in contrast to Leake et al., (2007) who 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. Lionello et al., (2008) project mostly decreases in extreme SWH for 2071-2100 over the Mediterranean Sea with larger decreases for the A2 scenario using winds downscaled from a GCM.

3.5.5. Coastal Impacts

Two classes of coastal hazard that are particularly significant in the context of disaster management are coastal inundation and shoreline stability. The frequency and severity of such events will be affected by climate change through rising sea levels and changes in extreme events. Figure 3.12 illustrates the interactions between various forms of climate forcing and coastal impacts. Several additional contributions to coastal impacts are also acknowledged such as extreme rainfall and runoff in coastal catchments which may contribute to coastal flooding. Multiple effects may occur on some coastlines as increasing ocean temperatures reduce natural barriers that protect against the erosive forces of waves. Examples include the melting of sea ice and permafrost in high latitudes (see Section 3.5.6) and the degradation of coral reefs through increased coral bleaching in the tropics. Wind can also have a direct erosive effect on coastlines, and coastal exposure to this influence is exacerbated during periods of extreme low coastal sea levels such as negative surges.

INSERT FIGURE 3.12 HERE

Figure 3.12: Relationships between climate, weather phenomena and physical impacts in the coastal zone.

Coastal inundation occurs during periods of extreme sea levels due to storm surges and high waves, particularly when combined with high tides. While tropical and extra-tropical cyclones are the most common causes of sea level extremes, other weather events can cause sea level extremes. For example, Green et al., (2009) reports an example of extreme sea levels and inundation affecting the low-lying Torres Strait Islands between the Cape Yorke Peninsula of Australia and Papua New Guinea as a result of persistent southeasterly winds from an anti-cyclone to the south. On the southeastern coast of Australia, frontal systems are a major cause of storm surges (McInnes et al., 2009b). In many parts of the world sea levels are also influenced by modes of variability such as ENSO. In the western equatorial Pacific, sea levels can fluctuate up to half a metre from one phase of ENSO to the other (Church et al., 2006a) and in combination with extremes of the tidal cycle, can cause extensive inundation in low-lying atoll nations in the absence of extreme weather events.

Extreme sea levels and high waves may lead to significant erosion of the coastline. In general, changes in shoreline position can arise from the combined effects of various factors such as:

1. A gradual rise in mean sea level, which causes a landward recession of coastlines that are made up of erodible materials.
2. Changes in the frequency or severity of transient storm erosion events (Zhang et al., 2004a).
3. Changes in sediment supply to the coast (Stive et al., 2003; Nicholls et al., 2007).
4. Changes in wave direction or period through sea level rise which alters wave refraction or climate variability which can cause realignment of shorelines (Ranasinghe et al., 2004; Bryan et al., 2008).
5. The loss of natural protective structures such as, coral reefs (e.g., Sheppard et al., 2005; Gravelle and Mimura, 2008) or in polar regions the melting of permafrost or sea ice which exposes soft shores to the buffering effects of waves and severe storms (Manson and Solomon, 2007).

The degree to which the processes described above will impact the coast are also a function of the coastal attributes themselves. For example, coastal elevation relative to sea level determines the severity and frequency of coastal inundation. In this regard, vertical movement of the land adjacent to the coast is also an important consideration (Haigh et al., 2009). Some coastal regions may be rising due to post-glacial rebound or slumping due to aquifer drawdown, the latter of which has anthropogenic origins. Similarly, the erodability of the coast is dependent on its particular physical (e.g., shoreline slope) and geomorphological attributes.

The susceptibility of a coastal region to erosion and inundation may be inferred from the following broad coastal characteristics, e.g., Nicholls et al., (2007):

- Beaches, rocky shorelines and cliffed coasts
- Deltas
- Estuaries and lagoons
- Mangroves, saltmarshes and sea grasses
- Coral reefs

Deltas are low-lying and hence generally prone to inundation, beaches are comprised of loose particles and therefore erodible. However, the degree to which these systems may be impacted by erosion and inundation may also be influenced by other factors which may affect disaster responses. For example, depleted mangrove forests or the degradation of coral reefs may reduce the buffering effect from high waves during severe storms, (e.g., Gravelle and Mimura, 2008); 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 amplify the risks of climate change (McGranahan et al., 2007), and also reduce the resilience of coastal settlements to disasters.

3.5.5.1. Observed Changes

The AR4 (Nicholls et al., 2007) reported that coasts are experiencing the adverse consequences of hazards such as increased coastal inundation, erosion and ecosystem losses. Since the AR4 a small number of additional studies that address shoreline evolution have been completed which do not change the AR4 assessment. The studies highlight the difficult task of clearly identifying a response due to climate change against a background of often large change brought about by other anthropogenic drivers, and of natural ongoing evolution and changes that occur due to natural climate variability such as ENSO (e.g., Ranasinghe et al., 2004; Allan and Komar, 2006). The scarcity and fragmentary nature of data sets as noted in Defeo et al., (2009) contributes to this problem.

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 (Cambers, 2009). Most beaches surveyed were found to be eroding, with faster rates of erosion generally found on islands that had been impacted by a higher number of hurricanes. The relative importance of anthropogenic factors, climate variability and climate change on the eroding trends could not be separated quantitatively.

Church et al., (2008) report that despite the positive trend in sea levels during the 20th century, Australia has generally been free of chronic coastal erosion problems. Where coastal erosion has been observed, it has not been possible to unambiguously attribute an erosion signal to sea level rise, in the presence of other anthropogenic activities.

A quantitative analysis of physical changes in 27 atoll islands across three central Pacific islands (Tuvalu, Kiribati and Federated States of Micronesia) over a 19 to 61 year period found 86% of islands remained stable or increased in area (43%) over the timeframe of analysis (Webb and Kench, 2010). Largest decadal rates of increase in island area range between 0.1 to 5.6 hectares. Only 14% of study islands exhibited a net reduction in island area. Despite small net changes in area, islands exhibited larger gross changes which represented a net lagoonward migration of islands in 65% of cases.

Chust et al., (2009) evaluate the relative contribution of local anthropogenic (non-climate change related) and sea level rise impacts on the coastal morphology and habitats in the Gipuzkoan littoral zone (Basque coast, northern Spain) for the period 1954–2004. 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.

3.5.5.2. Causes Behind the Changes

Assessments of coastal erosion that have been undertaken since the AR4 in the Caribbean (Cambers, 2009), Pacific (Webb and Kench, 2010), Australia (Church et al., 2008) and northern Spain (Chust et al., 2009) have tended to highlight the large natural and/or non-climatic anthropogenic contribution to current shoreline trends which prevent the identification of a climate change signal. The small number of studies that have been completed since the AR4 are either unable to attribute the coastline changes seen to different causes in a quantitative way or else find strong evidence for non-climatic causes that are natural and/or anthropogenic. This is consistent with the AR4, which stated with very high confidence that the impact of climate change on coasts is exacerbated by increasing human-induced pressures.

3.5.5.3. Projected Changes

The AR4 reported with very high confidence that coasts will be exposed to increasing risks, including coastal erosion, over coming decades due to climate change and sea level rise both of which will be exacerbated by increasing human-induced pressures (Nicholls et al., 2007). However it was also noted that since coasts are dynamic systems, adapting to climate change required insight into processes at decadal to century scales, at which understanding is least developed.

Since the AR4 several new studies have been completed that build understanding of how climate change will impact the coastlines in the future. These include new nationwide coastal assessments in several European countries and Australia that qualitatively assess coastal vulnerability based on the physical and geomorphological attributes of the coast and known existing vulnerabilities (e.g., Nicholls and de la Vega-Leinert, 2008). There have been several studies that model and map inundation from future scenarios of extreme sea level (e.g., Bernier et al., 2007; McInnes et al., 2009a), new studies that employ probabilistic frameworks to incorporate future climate uncertainty in impact studies which show promise for managing the large uncertainties in climate change projections (e.g., Purvis et al., 2008; Hunter, 2010) and studies that investigate the relative impact of wave climate and mean sea level changes on shoreline stability (Andrade et al., 2008; Coelho et al., 2009).

SURVAS (Synthesis and Upscaling of sea level Rise Vulnerability Assessment Studies) provides a qualitative assessment of vulnerability to climate change across Europe (Nicholls and de la Vega-Leinert, 2008). Aunan and Romstad (2008) report 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) report for Great Britain that large parts of the coasts (including England, Wales, and Scotland) already experience problems, including 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) report increased beach erosion, which is believed to be the result of recent increased storminess in the eastern Baltic Sea, combined with a decline in sea-ice cover during the winter. Sterr (2008) reports for Germany that there is a high level of reliance on hard coastal protection against extreme sea level hazards which will increase ecological vulnerability over time. A coastal vulnerability assessment for Australia (Department of Climate Change, 2009), identifies four broad coastal regions based on geomorphology, sediment type and tide and wave characteristics. The tropical northwestern coastline is expected to be most sensitive to changes in tropical cyclone behaviour while health of the coral reefs may also influence the tropical eastern coastline. The midlatitude southern and eastern coastlines are expected to be most sensitive to changes in mean sea level, wave climate and changes in storminess.

There have also been several studies that have developed methods for investigating the impact of inundation on the natural environment. Bernier et al., (2007) evaluated species vulnerability to inundation from future sea level rise using seasonal return periods of high water. McInnes et al., (2009a) developed spatial maps of stormtide and used high resolution LiDAR data to investigate exposure of coastal land to inundation under future sea level and wind speed scenarios along the Victorian coastline of southeast Australia. Probabilistic approaches have also been used to evaluate extreme sea level exceedance under uncertain future sea level rise scenarios. In the approach described in Purvis et al., (2008), a plausible probability distribution is applied to the range of future sea level rise estimates and Monte-Carlo sampling used to apply the sea level change to a 2D coastal inundation model. It is shown that evaluating the possible flood related losses (in monetary terms) in this framework is able to represent spatially the higher losses associated with the low frequency but high impact events compared with considering only a single midrange scenario. Hunter (2010) presents a method of combining sea-level extremes evaluated from observations with projections of sea level rise to 2100 to evaluate the probabilities of extreme events being exceeded over different future time horizons.

For the Portuguese coast, two studies report that projected changes in wave climate are likely to cause increased erosion in the future. Andrade et al., (2008) find that projected future climate in the HadCM3 model will not affect wave height along this coastline but the rotation in wave direction will increase the net littoral drift and the erosional response. On the basis of modelling various climate change scenarios for the next 25 years, Coelho et al., (2009) also find that the effects of sea level rise are less important than changes in wave action along a stretch of the Portuguese coast.

There have also been further developments in coastal erosion modelling within probabilistic frameworks that can take into account storm duration and sequencing (i.e., the compound effects on beach erosion that result from storms that occur in short succession) (Callaghan et al., 2008). Such methods have not as yet been applied in a climate change context.

3.5.6. Glaciers and Mountain Impacts

The steep topography of high-mountains is prone to gravity-driven mass movements such as landslides, avalanches, and floods that can lead to disasters.

3.5.6.1. Observed Changes

High-mountain environments are characterized by fast changes especially in recent decades with unprecedented retreat of glaciers all over the world (Paul et al., 2004; Kaser et al., 2006; Larsen et al., 2007; Rosenzweig et al., 2007). Conditions beyond historical experience have arisen at the beginning of the 21st century (Haeberli and Hohmann, 2008).

Most of the observed changes in glacier, permafrost, and snow related events are caused by temperature increases (Lemke et al., 2007). While an increase in air temperature can result in an increase of firn and ice temperature, the more visible effect of warming is the impact on glacier geometry (thickness, length, area, volume). Glacier geometry changes are controlled by the mass balance and dynamics of a glacier.

Since their last maximum at the end of the Little Ice Age (~1850) glaciers are predominantly retreating, interrupted by short periods of advance during the 20th century (Oerlemans, 2005). The mass loss of glaciers has clearly been increased towards the more recent years, with thickness losses in water equivalent ranging from 0.14 m from 1976 to 1985, to 0.25 m from 1986 to 1995, to 0.58 m during the period 1996–2005 (Zemp et al., 2007). Glacier length is also decreasing. The magnitude of downwasting at glacier terminal areas has been reported as up to 4–5 m/yr between 1985 and 2000 for the Swiss Alps (Paul and Haeberli, 2008), and up to 5–10 m/yr in southeast Alaska and British Columbia for about the last two decades of the 20th century (Larsen et al., 2007; Schiefer et al., 2007).

Evidence of mountain permafrost degradation and slope destabilization comes from a number of recent slope failures in permafrost areas, including a magnitude scale from block and rock fall to rock avalanches (volumes of ~10^2 to 10^7 m^3), observed in the European Alps (Gruber and Haeberli, 2007; Huggel, 2009) and also in other mountain regions (Niu et al., 2005; Allen et al., 2010). Examples are the 1997 Brenva rock avalanche in the Mont Blanc region (Barla et al., 2000), the 2004 Thurwieser rock avalanche, Italy (Sosio et al., 2008), rock slides from Dents du Midi and Dents Blanches, Switzerland, in 2006, or from Monte Rosa, Italy, in 2007 (Huggel, 2009; Fischer et al., 2010a), with volumes of a few millions of cubic meters. Very large rock and ice avalanches with volumes of 50 to over 100 million m^3 have occurred in the 2002 Caucasus Kolka avalanche (Haeberli et al., 2004; Kotlyakov et al., 2004; Huggel et al., 2005) and in 2005, Mt. Steller, south-central Alaska (Huggel et al., 2008).

Quantification of trends in occurrence of such events is difficult due to uncertainty in documentation, despite a generally increasing level of documentation in recent years. Nevertheless, there is an apparent increase of large rock slides during the past two decades, and especially during the first years of the 21st century the frequency has increased in the European Alps and the Southern Alps of New Zealand (Allen et al., 2010; Fischer et al., 2010b), in parallel with strong temperature increases, glacier shrinkage, and permafrost degradation.

3.5.6.2. Causes Behind the Changes

Hazards and extreme events in high mountains occur due to cumulative changes in glacier and permafrost, or are of a stochastic nature. Glacier lake outburst floods (GLOFs) are typically a result of cumulative developments, and occur (i) only once (e.g., full-breach failure of moraine-dammed lakes), (ii) for the first time (e.g., new formation and outburst of glacial lakes), and/or (iii) repeatedly (e.g., ice-dammed lakes with drainage cycles, or ice fall) (Clarke, 1982; Clague and Evans, 2000; Huggel et al., 2004; Dussaillant et al., 2010). In the past decades GLOFs have caused severe 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). Due to the relatively rare occurrence of GLOFs, clear information on possible changes of occurrence of such extreme events on the regional or global level is lacking. For the Himalayas a small but not statistically significant increase of GLOF events was observed over the period 1940 to 2000 (Richardson and Reynolds, 2000).

Degradation of permafrost due to warming affects slope stability. However, monitoring of mountain permafrost temperatures has a short history with only about 20 years of data (Vonder Mühll et al., 1998; Niu et al., 2005; Harris et al., 2009) for gently sloped terrain, and less than 10 years for steep rock slopes (Gruber et al., 2004b). Any significant warming trend of bedrock permafrost cannot yet be derived from the small number of monitoring years, but the 2003 European summer heat wave (Section 3.3.1.1) has been associated with rapid thaw and extension of the active layer, and an increased number of predominantly small-scale rock fall events (Gruber et al., 2004a; Gruber and Haeberli, 2007).

Shallow landslides and debris flows generally follow a stochastic pattern as they are primarily triggered by precipitation. The spatial and temporal patterns of precipitation, the intensity, the duration of rainfall and the antecedent rainfall are all important for shallow landslides (Iverson, 2000; Wieczorek et al., 2005; Sidle and Ochiai, 2006). In some regions the influence of antecedent rainfall on landslide triggering is likely to dominate over rainfall intensity (Kim et al., 1991; Glade, 1998), although some uncertainty may be involved from temporally insufficient resolution of rainfall records. Landslides in temperate and tropical mountains usually are not temperature sensitive and can be more strongly influenced by human activities such as poor land-use practise, deforestation, overgrazing, etc.

For shallow landsliding and debris flows in high mountains, observations indicate that the initiation zones move upwards as glaciers retreat and new poorly consolidated sediment becomes exposed (Rickenmann and Zimmermann, 1993; Zimmermann and Haeberli, 1993; Haeberli and Beniston, 1998). Research has so far not provided any clear indications of change in the frequency of debris flows. In the Swiss Alps it was found that debris flow activity on a local site was higher during the 19th century than today (Stoffel et al., 2005) while in the French Alps no significant variation of debris flow frequency could be observed since the 1950s in high-mountain terrain above 2200 m a.s.l (Jomelli et al., 2004). Indirect climate effects such as increase of available sediment or changing seasonal snow patterns can also influence debris flow activity (Rebetez et al., 1997; Beniston, 2006). Statistics are not completely clear but there could be an increase of debris flow activity in alpine regions during the past decades due to extreme rainfall events, in combination with a snow fall line located at high elevation, contributing to enhanced liquid precipitation. The elevated activity of high-mountain landslide activity during the recent warming is consistent with findings on the occurrence of large events during the post-Ice Age and early Holocene (Holm et al., 2004; Prager et al., 2009).

Several events in the past decades have shown that particularly severe physical impacts can result from interacting and cascading processes. Typical processes are outburst floods of glacier lakes due to impact waves generated by failure of moraine slopes (Hubbard et al., 2005; Vilimek et al., 2005) or ice and rock avalanches (Clague and Evans, 2000). Very large rock-ice avalanches and debris flows, triggered by initial rock or ice failures (Huggel et al., 2005; Evans et al., 2009), or volcanic eruptions (Pierson et al., 1990) have killed hundreds to thousands of people during the 20th century. There is no indication so far as to whether such large events with cascading processes have increased during the past decades.

The initiation of shallow landslides and debris flows in cold regions and high mountains can be influenced by the thermal state (frozen vs. unfrozen conditions), and related hydraulic effects of scree slopes (Haeberli et al., 1990; Rickenmann and Zimmermann, 1993). Permafrost thawing and related depth increase of the active layer, together with incomplete thaw consolidation after melt, may increase both frequency and magnitude (higher potential erosion depth) of debris flows (Zimmermann et al., 1997; Rist and Phillips, 2005). On the other hand, permafrost at the base of the active layer also acts as a hydraulic barrier to groundwater percolation and can imply local saturation within the non-frozen debris. Snow cover distribution and melt can also have an important effect on debris flow activity by supplying additional liquid water for soil saturation favoring slope instabilities (Kim et al., 2004). The large debris flow events in the past 20 years, triggered by intensive rainfall and affecting extensive areas of the Alps, occurred in summer or fall and were typically characterized by a high elevation of the snow fall limit (Rickenmann and Zimmermann, 1993; Chiarle et al., 2007). Warming may directly influence the flow speed of frozen bodies of debris and rock such as rock glaciers. In recent years ground and remote sensing based monitoring has revealed remarkable acceleration of rock glaciers surface flow-speed of up to 4 m y^{-1}(Kääb et al., 2007; Roer et al., 2008). At some specific sites in the Alps, flow-speeds have occasionally reached up to 15 m y^{-1}, associated with slope instabilities (Delaloye et al., 2008). These phenomena have only recently been identified and could lead to large single events such as debris avalanches or alter the frequency and magnitude of debris flows.

Rock slope failure is often a result of slope steepening by glacial erosion and unloading or debuttressing due to glacier retreat (Augustinus, 1995), though it may take decades for a slope failure to occur due to glacier retreat. Recent rock slope failures, including the one in Grindelwald, Swiss Alps (Oppikofer et al., 2008), have confirmed the short response to glacier downwasting within a few decades or event shorter. 20th century warming may have reached some decameters depth on high steep slopes (Haeberli et al., 1997), and will continue to reach increasingly greater depths with future warming. Case studies of exceptionally warm periods of weeks to months duration indicate that both small-scale and large-scale slope failures can be triggered (Gruber et al., 2004a; Huggel, 2009; Fischer et al., 2010a).

Observed changes in physical impacts such as landslides, avalanches, and GLOFs that are primarily temperature driven and occur in cold and high mountain regions have *likely* been influenced by the anthropogenic greenhouse gas increase, since there is a direct physical link between warming and those changes, and the warming in those regions over the second half of the 20th century has been observed and attributed to anthropogenic influence. There is however a lack of evidence to assess any influence or lack of influence from anthropogenic warming on other observed physical impacts such as shallow landslides in lower latitude regions that are primarily precipitation driven, since it is difficult to determine the causes of precipitation change in those regions while poor land-use practices also may have contributed to landslide activity (e.g., Sidle and Ochiai, 2006).

3.5.6.3. Projected Changes

Given the projected rise of air temperature during the 21st century, it is *very likely* that mountain glacier areas will further reduce. European Alp glaciers are projected to decrease on the order of 20% to >50% (of the 2000 glacier area reference state) by about 2050 (Zemp et al., 2006; Huss et al., 2008) for a 2–3°C temperature increase over the 1961-1990 mean state. The warming climate favors rapid and sometimes unexpected developments of glacier decay and related mass movements (Huggel et al., 2010a), and as a result glaciers are increasingly in an imbalance. Projected glacier retreat in the 21st century will *likely* form new and potentially unstable lakes. Probable sites of new lakes have already been identified for some alpine glaciers (Frey et al., 2010). Of special concern in combination with existing and new natural and artificial lakes are rock slope and moraine instabilities that can result in impact waves and outburst floods. For rock slopes, the ongoing temperature rise will result in gradual permafrost degradation to increasing depths (Haeberli and Burn, 2002; Harris et al., 2009). At near-surface bedrock, the temperature rise is faster than at depth and warm permafrost areas (~-2 to 0°C), considered to be more susceptible to slope failures, may rise a few hundred meters during the next 100 years, depending on air temperature increase and the climate scenario applied (Noetzli and Gruber, 2009). The climate signal then penetrates to greater depth where the response of bedrock temperatures to ambient warming is delayed by decades or centuries (Noetzli et al., 2007). The response of firn and ice temperature to an increase in air temperature is typically faster and non-linear (Haeberli and Funk, 1991; Suter et al., 2001; Vincent et al., 2007). Latent heat effects from refreezing melt water can amplify the increase in air temperature in firn and ice (Huggel, 2009). At higher temperatures, there is more melting water and the strength of ice is lower, as a result, ice avalanches increase (Huggel et al., 2004; Caplan-Auerbach and Huggel, 2007).

Future extreme climatic events such as heat waves can result in rapid near-surface thawing and reach greater depth along advection corridors. Recent studies indicate that warm extremes can have a triggering effect for large landslides (rock and ice avalanches) but the physical processes are not yet well understood (Huggel et al., 2010b). For warm extremes with a potential to trigger slope instabilities (5-, 10- and 30-day warm events), based on the assessment of several RCMs it is projected that such high-temperature events for the period 2001-2050 compared to a 1951-2000 reference period increase about 1.5 to 4 times by 2050, and in some models up to 10 times (Huggel et al., 2010b).

Generally speaking, it is *likely* that continued permafrost degradation will lead to a general decrease of rock slope stability (Gruber and Haeberli, 2007). Future locations and timing of large rock avalanches are extremely difficult to predict, as they depend on a multitude of factors, including local geological conditions and failure mechanisms are not known in detail. There is some concern that the probability of large, combined events, such as landslides impacting lakes and generating large outburst floods, will increase (Haeberli and Hohmann, 2008; Huggel et al., 2010a).

It is *more likely than not* that the magnitude of shallow landslides and debris flows from recently deglaciated terrain will increase because of higher availability of unconsolidated sediment (Haeberli and Beniston, 1998), though future changes in rainfall amount and intensities will affect this projection. Changes in frequency of debris flows are difficult to project as they depend on the future frequency of debris flow triggering rainstorms. It is *more likely than not* that high-mountain debris flows will have earlier onset because of earlier snow melt. As extreme precipitation is *very likely* to increase in the future in many places of the world (Beniston et al., 2007; Christensen et al., 2007; Meehl et al., 2007b), shallow landslides in lower mountain ranges are *more likely than not* to increase.

Future changes in the magnitude and frequency of shallow landslides in temperate and tropical regions chiefly depend on frequency and intensities of rainfall events and anthropogenic land-use. Landslides can be triggered both by long-lasting (days to weeks) rainfall periods and short-term high-intensity rainfall events. In some regions social-economic pressure is likely to lead to land use practices that increase the frequency of landslides.

It is *very likely* that glacier retreat will continue and accelerate given that air temperature will continue to increase (Lemke et al., 2007). It is *likely* that new lakes will form in some regions in the 21st century due to glacier retreat, however, uncertainty on the projection of the location and timing of future glacier lake outburst floods is high (Frey et al., 2010). Projected changes in shallow landslide and debris flows in temperate regions are uncertain because of high uncertainty in projected changes of precipitation and in the land-use practices (Sidle and Ochiai, 2006).

3.5.7. Permafrost and High-Latitude Impacts

Permafrost is widespread in Arctic, Subarctic, and high-mountains regions, and in ice-free areas of Antarctica. Permafrost regions occupy approximately 23 million km^2 of land areas in the Northern Hemisphere (Zhang et al., 1999). The permafrost temperature regime is a sensitive indicator of climatic variability and change (Lachenbruch and Marshall, 1986; Osterkamp, 2005). Melting of massive ground ice and thawing of ice-rich permafrost can lead to subsidence of ground surface and to the formation of uneven topography known as thermokarst, generating dramatic changes in ecosystems, landscapes, and infrastructure performance (Nelson et al., 2001; Walsh, 2005). The active layer (the layer over the permafrost that thaws and freezes seasonally) 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). Creation and drainage of thaw lakes and changes in lake surface area as a whole due to permafrost degradation would present challenges for ecosystems, natural resources, and the people who depend upon them (Hinzman et al., 2005; Smith et al., 2005a). Rapid Arctic coastal erosion increases threats to villages and industries.

3.5.7.1. Observed Changes

Observed evidence shows that temperatures at the top of the permafrost have increased by up to 3°C since the early 1980s (Lemke et al., 2007; Harris et al., 2009). Over the high Arctic such as in northern Alaska (Osterkamp, 2005, 2007) and Russia (Obserman and Mazhitova, 2001), permafrost temperatures have increased by about 2 to 3°C. The magnitude of permafrost temperature increase is up to 1.0ºC in the Interior of Alaska (Osterkamp, 2005, 2007), much of the Canadian Arctic (Smith et al., 2005b), Mongolia (Sharkhuu, 2003), and on the Tibetan Plateau (Cheng and Wu, 2007). Generally speaking, the magnitude of permafrost temperature increase in continuous permafrost regions is greater than in discontinuous permafrost regions. Increases in snow insulation effect may contribute significantly to the greater permafrost temperature increase in the high Arctic, and contribute to local and regional variability of permafrost temperature increase (Zhang et al., 2005). When the other conditions remain constant, active layer thickness is expected to increase in response to climate warming, especially in summer. Observed evidence shows that active layer thickness has increased about 20cm in the Russian Arctic from the early 1960s to 2000 (Zhang et al., 2005; Wu and Zhang, 2008), no significant trend in North American Arctic since the early 1990s (Brown et al., 2000), and up to 1.0 m from since the early 1980s over the Qinghai-Tibetan Plateau (Wu and Zhang, 2010). Extensive thermokarst development has been found in Alaska (Yoshikawa and Hinzman, 2003; Osterkamp et al., 2009), in the central Yakutia (Gavriliev and Efremov, 2003), and on the Qinghai-Tibetan Plateau (Niu et al., 2005). Significant expansion and deepening of thermokarst lakes were observed near Yakutsk with subsidence rates of 17 to 24 cm yr^{-1} from 1992–2001 (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., 2005a).

The most sensitive regions of permafrost degradation are coasts with ice-bearing permafrost that are exposed to the Arctic Ocean. Due to the increased storm activity, long sea ice free seasons, and thawing permafrost, the Arctic coasts are retreating in a rapid rate of 2 to 3 m yr^{-1}(Rachold et al., 2003; Jorgenson and Brown, 2005) with an extreme of about 34 m yr^{-1} at Newtok, Alaska in 2003 (Karl et al., 2009). The rate of erosion along Alaska's northeastern coastline has doubled over the past 50 years (Karl et al., 2009).

3.5.7.2. Causes Behind the Changes

Increases in air temperature are in part responsible for the observed increase in permafrost temperature over the Arctic and Subarctic, and changes in snow cover also play a critical role (Osterkamp, 2005; Zhang et al., 2005). Earlier snowfall in autumn and thicker snow cover during winter provides a strong insulation effect, resulting in an increase of permafrost temperature much higher than that of air temperature in the Arctic. Changes in active layer thickness are primarily controlled by changes in length of thaw season and summer air temperature. The combination of Arctic sea ice retreat, storm activity increase, and permafrost degradation is responsible for rapid Arctic coast erosion in recent decades (Atkinson et al., 2006). Expansion of lake areas 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., 2005a). Overall, increased air temperature over high latitudes is primarily responsible for development of thermokarst terrains and thaw lakes.

3.5.7.3. Projected Changes

Widespread increases in active layer thickness in the Arctic and Subarctic are expected in response to global warming over the 21st century (IPCC, 2007a). Due to sea ice retreat, and permafrost degradation, with possibly a contribution from more storminess, the frequency and magnitude of the rate of Arctic coastal erosion will *likely* increase. For example, it has been projected that the coastal erosion rate at Newtok, Alaska will range from 11 to 25 m yr^{-1} in the next 20 years (Karl et al., 2009).

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, 1998; Kellogg and Griffin, 2006). For example, dust from the Saharan region and from Asia may reach North America (McKendry et al., 2007).

3.5.8.1. Observed Changes

The Sahara (especially Bodélé Depression in Chad) and east Asia have been recognized as the strongest dust sources globally (Goudie, 2009). Over the past few decades, the frequency of dust events has 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), but there seems to also be an increase in more recent years (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 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 in to the present (Prospero and Lamb, 2003; Prospero et al., 2009). The dust frequency in Asia has decreased since the late 1970s

3.5.8.2. Causes Behind the Changes

Surface soil dust concentration during a sand and dust storm is controlled by a number of factors in a specific region. The driving force for the production of dust storms is the surface wind associated with cold frontal systems sweeping across the dry desert areas and lifting soil particles in the atmosphere. Dust emissions are also controlled by the surface conditions such as the desert coverage distributions, snow cover and soil moisture. In the Sahel region, the elevated high level of dust emission is related to the persistent drought since the 1970s, and to long-term changes in the North Atlantic Oscillation (Ginoux et al., 2004; Chiapello et al., 2005; Engelstaedter et al., 2006), and perhaps to North Atlantic SST as well (Wong et al., 2008). The long-term change in China dust storm frequency is influenced by climate variations, rather than desertification processes. The desert areas increased by ~2 to ~7% (Zhong, 1999) in China during 1960-2000, when the dust storm frequency decreased. A 44-year simulation study of Asian soil dust production with a dynamic desert distribution from 1960 to 2003 suggests that climatic variations play a major role in the declining trends in dust emission and storm frequencies (Zhang et al., 2003; Zhou and Zhang, 2003; Zhao et al., 2004) in China. Changes in wind (Wang et al., 2006c), meridional temperature gradients and cyclone frequencies (Qian et al., 2002), large-scale circulations such as the Asian polar vortex (Gong et al., 2006), the Siberia high (Ding et al., 2009), rainfall and vegetation (Zhou and Zhang, 2003) all contributed to the decrease in the observed dust frequency in China. Overall, the observed changes in dust activity are mainly the result of long-term 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, there are a large number of potential causes affecting dust frequency in China, but their relative importance is uncertain.

3.5.8.3. Projected Changes

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, Meehl et al., 2007b).Thomas et al., (2005) suggest that dune fields in southern Africa can be reactivated, and sand will become significantly exposed and move, as a consequence of 21st century climate 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 (Wang et al., 2009d). However, projected changes in wind are lacking.

References

Abbs, D., and T. Rafter, 2008: *The effect of climate change on extreme rainfall events in the Western Port Region*. CSIRO Marine and Atmospheric Research, Australia, 23 pp.

Abeysirigunawardena, D.S., and I.J. Walker, 2008: Sea Level Responses to Climatic Variability and Change in Northern British Columbia. *Atmosphere-Ocean,* **46(3)**, 277-296

Abeysirigunawardena, D.S., E. Gilleland, D. Bronaugh, and W. Wong, 2009: Extreme Wind Regime Responses to Climate Variability and Change in the Inner South Coast of British Columbia, Canada. *Atmosphere-Ocean,* **47(1)**, 41-62

Adams, P.N., D.L. Imnan, and N.E. Graham, 2008: Southern California deep-water wave climate: Characterization and application to coastal processes. *Journal of Coastal Research,* **24(4)**, 1022-1035

Aguilar, E., et al., 2009: Changes in temperature and precipitation extremes in western central Africa, Guinea Conakry, and Zimbabwe, 1955-2006. *Journal of Geophysical Research,* **114(D02115)**

Aguilar, E., et al., 2005: Changes in precipitation and temperature extremes in Central America and northern South America, 1961-2003. *Journal of Geophysical Research,* **110(D23107)**

Aizen, V.B., V.A. Kuzmichenok, A.B. Surazakov, and E.M. Aizen, 2007: Glacier changes in the Tien Shan as determined from topographic and remotely sensed data. *Global and Planetary Change,* **56(3-4)**, 328-340

Alexander, L.V., and J.M. Arblaster, 2009: Assessing trends in observed and modelled climate extremes over Australia in relation to future projections. *International Journal of Climatology,* **29(3)**, 417-435

Alexander, L.V., and S. Power, 2009: Severe storms inferred from 150 years of sub-daily pressure observations along Victoria's "Shipwreck Coast". *Australian Meteorological and Oceanographic Journal,* **58(2)**, 129-133

Alexander, L.V., P. Uotila, and N. Nicholls, 2009: Influence of sea surface temperature variability on global temperature and precipitation extremes. *Journal of Geophysical Research,* **114(D18116)**

Alexander, L.V., et al., 2006: Global observed changes in daily climate extremes of temperature and precipitation. *Journal of Geophysical Research-Atmospheres,* **111(D05109)**

Alexandersson, H., H. Tuomenvirta, T. Schmith, and K. Iden, 2000: Trends of storms in NW Europe derived from an updated pressure data set. *Climate Research,* **14(1)**, 71-73

Ali, A., and T. Lebel, 2009: The Sahelian standardized rainfall index revisited. *International Journal of Climatology,* **29(12)**, 1705-1714

Allamano, P., P. Claps, and F. Laio, 2009: Global warming increases flood risk in mountainous areas. *Geophysical Research Letters,* **36(L24404)**

Allan, J.C., and P.D. Komar, 2006: Climate controls on US West Coast erosion processes. *Journal of Coastal Research,* **22(3)**, 511-529

Allan, R., S. Tett, and L. Alexander, 2009: Fluctuations in autumn-winter severe storms over the British Isles: 1920 to present. *International Journal of Climatology,* **29(3)**, 357-371

Allan, R.P., and B.J. Soden, 2008: Atmospheric warming and the amplification of precipitation extremes. *Science,* **321**, 1481-1484

Allen, M., 2003: Liability for climate change: will it ever be possible to sue anyone for damaging the climate? *Nature,* **421(6926)**, 891-892

Allen, M.R., and W.J. Ingram, 2002: Constraints on future changes in climate and the hydrologic cycle. *Nature,* **419**, 224-232

Allen, S.K., S.C. Cox, and I.F. Owens, 2010: Rock-avalanches and other landslides in the central Southern Alps of New Zealand: A regional assessment of possible climate change impacts. *Landslides*, doi:10.1007/s10346-010-0222-z

An, S.I., A. Timmermann, L. Bejarano, F.F. Jin, F. Justino, Z. Liu, and A.W. Tudhope, 2004: Modeling evidence for enhanced El Niño-Southern Oscillation amplitude during the Last Glacial Maximum. *Paleoceanography,* **19(PA4009)**

Anders, I., and B. Rockel, 2009: The influence of prescribed soil type distribution on the representation of present climate in a regional climate model. *Climate Dynamics,* **33(2-3)**, 177-186

Andersen, O.B., S.I. Seneviratne, J. Hinderer, and P. Viterbo, 2005: GRACE-derived terrestrial water storage depletion associated with the 2003 European heat wave. *Geophysical Research Letters,* **32(L18405)**

Anderson, C.J., et al., 2003: Hydrological processes in regional climate model simulations of the central United States flood of June-July 1993. *Journal of Hydrometeorology,* **4(3)**, 584-598

Andrade, C., H.O. Pires, R. Taborda, and M.C. Freitas, 2007: Projecting future changes in wave climate and coastal response in Portugal by the end of the 21st Century. *Journal of Coastal Research,* **50**, 263-257

Andrade, C., R.M. Trigo, M.C. Freitas, M.C. Gallego, P. Borges, and A.M. Ramos, 2008: Comparing historic records of storm frequency and the North Atlantic Oscillation (NAO) chronology for the Azores region. *Holocene,* **18(5)**, 745-754

Angel, J.R., and S.A. Isard, 1998: The frequency and intensity of Great Lake cyclones. *Journal of Climate,* **11(1)**, 61-71

Arriaga-Ramírez, S., and T. Cavazos, 2010: Regional trends of daily precipitation indices in Northwest Mexico and the Southwest United States. *Journal of Geophysical Research - Atmospheres*, doi:10.1029/2009JD013248
Aryal, S.K., B.C. Bates, E.D. Campbell, Y. Li, M.J. Palmer, and N.R. Viney, 2009: Characterizing and Modeling Temporal and Spatial Trends in Rainfall Extremes. *Journal of Hydrometeorology,* **10(1)**, 241-253
Ashfaq, M., Y. Shi, W.W. Tung, R.J. Trapp, X. Gao, S.J. Pal, and N.S. Diffenbaugh, 2009: Suppression of south Asian summer monsoon precipitation in the 21st century. *Journal of Geophysical Research,* **36(L01704)**
Asokan, S.M., and D. Dutta, 2008: Analysis of water resources in the Mahanadi River Basin, India under projected climate conditions. *Hydrological Processes,* **22(18)**, 3589-3603
Atkinson, D.E., et al., 2006: Canadian cryospheric response to an anomalous warm summer: A synthesis of the climate change action fund project "The state of the arctic cryosphere during the extreme warm summer of 1998". *Atmosphere-Ocean,* **44(4)**, 347-375
Augustinus, P., 1995: Rock mass strength and the stability of some glacial valley slopes. *Zeitschrift Für Geomorphologie,* **39(1)**, 55-68
Aunan, K., and B. Romstad, 2008: Strong coasts and vulnerable communities: Potential implications of accelerated sea-level rise for Norway. *Journal of Coastal Research,* **24(2)**, 403-409
Bajracharya, S.R., and P. Mool, 2009: Glaciers, glacial lakes and glacial lake outburst floods in the Mount Everest region, Nepal. *Annals of Glaciology,* **50(53)**, 81-86
Barla, G., F. Dutto, and G. Mortara, 2000: Brenva glacier rock avalanche of 18 January 1997 on the Mount Blanc range, northwest Italy. *Landslide News,* **13**, 2-5
Barnett, T.P., et al., 2008: Human-induced changes in the hydrology of the western United States. *Science,* **319(5866)**, 1080-1083
Barring, L., and H. von Storch, 2004: Scandinavian storminess since about 1800. *Geophys. Res. Lett.,* **31(20)**, doi:10.1029/2004GL020441
Barring, L., and K. Fortuniak, 2009: Multi-indices analysis of southern Scandinavian storminess 1780-2005 and links to interdecadal variations in the NW Europe-North Sea region. *International Journal of Climatology,* **29(3)**, 373-384
Barring, L., T. Holt, M.L. Linderson, M. Radziejewski, M. Moriondo, and J.P. Palutikof, 2006: Defining dry/wet spells for point observations, observed area averages, and regional climate model gridboxes in Europe. *Climate Research,* **31(1)**, 35-49
Barrucand, M., M. Rusticucci, and W. Vargas, 2008: Temperature extremes in the south of South America in relation to Atlantic Ocean surface temperature and Southern Hemisphere circulation. *Journal of Geophysical Research-Atmospheres,* **113(D20111)**, doi:10.1029/2007D009026
Bartholy, J., and R. Pongracz, 2007: Regional analysis of extreme temperature and precipitation indices for the Carpathian Basin from 1946 to 2001. *Global and Planetary Change,* **57**, 83-95
Bates, B.C., Z.W. Kundzewics, S. Wu, and J.P. Palutikof, 2008: *Climate Change and Water. Technical Paper of the Intergovernmental Panel on Climate Change*. IPCC Secretariat, Geneva, 210 pp. http://www.ipcc.ch/pdf/technical-papers/climate-change-water-en.pdf
Bender, M.A., T.R. Knutson, R.E. Tuleya, J.J. Sirutis, G.A. Vecchi, S.T. Garner, and I.M. Held, 2010: Modeled impact of anthropogenic warming on the frequency of intense Atlantic hurricanes. *Science,* **327(5964)**, 454-458
Bengtsson, L., S. Hagemann, and K.I. Hodges, 2004: Can climate trends be calculated from reanalysis data? *Journal of Geophysical Research-Atmospheres,* **109(D11111)**
Bengtsson, L., K.I. Hodges, and E. Roeckner, 2006: Storm tracks and climate change. *Journal of Climate,* **19(15)**, 3518-3543
Bengtsson, L., K.I. Hodges, and N. Keenlyside, 2009: Will Extratropical Storms Intensify in a Warmer Climate? *Journal of Climate,* **22(9)**, 2276-2301
Bengtsson, L., K.I. Hodges, M. Esch, N. Keenlyside, L. Kornblueh, J.J. Luo, and T. Yamagata, 2007: How may tropical cyclones change in a warmer climate? *Tellus.Series A: Dynamic Meteorology and Oceanography,* **59(4)**, 539-561
Beniston, M., 2004: The 2003 heat wave in Europe: A shape of things to come? An analysis based on Swiss climatological data and model simulations. *Geophysical Research Letters,* **31(L02202)**
Beniston, M., 2006: August 2005 intense rainfall event in Switzerland: Not necessarily an analog for strong convective events in a greenhouse climate. *Geophysical Research Letters,* **33(L05701)**
Beniston, M., 2009: Trends in joint quantiles of temperature and precipitation in Europe since 1901 and projected for 2100. *Geophysical Research Letters,* **36(L07707)**
Beniston, M., et al., 2007: Future extreme events in European climate: an exploration of regional climate model projections. *Climatic Change,* **81**, 71-95
Benito, G., A. Diez-Herrero, and M.F. de Villalta, 2003: Magnitude and frequency of flooding in the Tagus basin (Central Spain) over the last millennium. *Climatic Change,* **58(1-2)**, 171-192

Benito, G., M. Barriendos, C. Llasat, M. Machado, and V. Thorndycraft, 2005: Impacts on natural hazards of climatic origin. Flood risk. In: *A Preliminary General Assessment of the Impacts in Spain Due to the Effects of Climate Change* [J.M. Moreno (ed.)]. Ministry of Environment, Spain, pp. 507-527

Benito, G., V.R. Thorndycraft, M. Rico, Y. Sanchez-Moya, and A. Sopena, 2008: Palaeoflood and floodplain records from Spain: Evidence for long-term climate variability and environmental changes. *Geomorphology,* **101(1-2)**, 68-77

Bergametti, G., 1998: Mineral aerosols: Renewed interest for climate forcing and tropospheric chemistry studies. *IGACtivities Newsletter,* **1**, 13-17

Bernier, N.B., K.R. Thompson, J. Ou, and H. Ritchie, 2007: Mapping the return periods of extreme sea levels: Allowing for short sea level records, seasonality, and climate change. *Global and Planetary Change,* **57(1-2)**, 139-150

Betts, A.K., 2004: Understanding hydrometeorology using global models. *Bulletin of the American Meteorological Society,* **85(11)**, 1673-1688

Betts, R.A., et al., 2007: Projected increase in continental runoff due to plant responses to increasing carbon dioxide. *Nature,* **448(7157)**, 1037-1041

Bevan, S.L.N., P. R. J. , W.M.F. Grey, S.O. Los, and S.E. Plummer, 2009: Impact of atmospheric aerosol from biomass burning on Amazon dry-season drought. *Journal of Geophysical Research-Atmospheres,* **114(D09204)**

Bhutiyani, M.R., V.S. Kale, and N.J. Pawar, 2008: Changing streamflow patterns in the rivers of northwestern Himalaya: Implications of global warming in the 20th century. *Current science,* **95(5)**, 618-626

Bhutiyani, M.R., V.S. Kale, and N.J. Pawar, 2010: Climate change and the precipitation variations in the northwestern Himalaya: 1866-2006. *International Journal of Climatology,* **30(4)**, 535-548

Biasutti, M., and A.H. Sobel, 2009: Delayed Sahel rainfall and global seasonal cycle in a warmer climate. *Geophysical Research Letters,* **36(L23707)**

Biasutti, M., D.S. Battisti, and E.S. Sarachik, 2003: The annual cycle over the tropical Atlantic, South America, and Africa. *Journal of Climate,* **16(15)**, 2491-2508

Bindoff, N.L., et al., 2007: Observations: Oceanic Climate Change and Sea Level. In: *The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [S. Solomon, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA

Bister, M., and K.A. Emanuel, 1998: Dissipative heating and hurricane intensity. *Meteorology and Atmospheric Physics, Vienna, Austria,* **65(3-4)**, 233-240

Blenkinsop, S., and H.J. Fowler, 2007: Changes in European drought characteristics projected by the PRUDENCE regional climate models. *International Journal of Climatology,* **27(12)**, 1595-1610

Boberg, F., P. Berg, P. Thejll, W.J. Gutowski, and J.H. Christensen, 2009a: Improved confidence in climate change projections of precipitation evaluated using daily statistics from the PRUDENCE ensemble. *Climate Dynamics,* **32(7-8)**, 1097-1106

Boberg, F., P. Berg, P. Thejll, W.J. Gutowski, and J.H. Christensen, 2009b: Improved confidence in climate change projections of precipitation further evaluated using daily statistics from the PRUDENCE ensemble. *Climate Dynamics*, doi:10.1007/s00382-009-0683-8

Boe, J., and L. Terray, 2008: Uncertainties in summer evapotranspiration changes over Europe and implications for regional climate change. *Geophysical Research Letters,* **35(L05702)**

Boer, G.J., 2009: Changes in Interannual Variability and Decadal Potential Predictability under Global Warming. *Journal of Climate,* **22**, 3098-3109

Bollasina, M., and S. Nigam, 2009: Indian Ocean SST, evaporation, and precipitation during the South Asian summer monsoon in IPCC-AR4 coupled simulations. *Climate Dynamics,* **33(7-8)**, 1017-1032

Bonan, G.B., 2008: Forests and climate change: Forcing, feedbacks, and the climate benefit of forests. *Science,* **320**, 1444-1449

Boo, K.O., W.T. Kwon, and H.J. Baek, 2006: Change of extreme events of temperature and precipitation over Korea using regional projection of future climate change. *Geophysical Research Letters,* **33(L01701)**

Brabson, B.B., D.H. Lister, P.D. Jones, and J.P. Palutikof, 2005: Soil moisture and predicted spells of extreme temperatures in Britain. *Journal of Geophysical Research,* **110(D05104)**

Branstator, G., and F. Selten, 2009: "Modes of Variability" and Climate Change. *Journal of Climate,* **22(10)**, 2639-2658

Brayshaw, D.J., B. Hoskins, and M. Blackburn, 2008: The storm-track response to idealized SST perturbations in an aquaplanet GCM. *Journal of the Atmospheric Sciences,* **65(9)**, 2842-2860

Brazdil, R., C. Pfister, H. Wanner, H. Von Storch, and J. Luterbacher, 2005: Historical climatology in Europe - The state of the art. *Clim. Change,* **70(3)**, 363-430

Brázdil, R., P. Dobrovolny, V. Kakos, and O. Kotyza, 2006: Historical and recent floods in the Czech Republic: causes, seasonality, trends, impacts. In: *Flood Risk Management: Hazards, Vulnerability and Mitigation Measures* [J. Schanze, et al. (eds.)]. Springer, Netherlands, pp. 247-259

Breda, N., and V. Badeau, 2008: Forest tree responses to extreme drought and some biotic events: Towards a selection according to hazard tolerance? *Comptes Rendus Geoscience,* **340(9-10)**, 651-662

Brockhaus, P., D. Luthi, and C. Schar, 2008: Aspects of the diurnal cycle in a regional climate model. *Meteorologische Zeitschrift,* **17(4)**, 433-443

Brohan, P., J.J. Kennedy, I. Harris, S.F.B. Tett, and P.D. Jones, 2006: Uncertainty estimates in regional and global observed temperature changes: a new dataset from 1850. *Journal of Geophysical Research,* **111(D12106)**

Bromirski, P.D., R.E. Flick, and D.R. Cayan, 2003: Storminess variability along the California coast: 1858-2000. *Journal of Climate,* **16(6)**, 982

Bromirski, P.D., D.R. Cayan, and R.E. Flick, 2005: Wave spectral energy variability in the northeast Pacific. *Journal of Geophysical Research-Oceans,* **110(C03005)**

Brooks, H.E., and N. Dotzek, 2008: The spatial distribution of severe convective storms and an analysis of their secular changes. In: *Climate Extremes and Society* [H.F. Diaz, and R. Murnane (eds.)]. Cambridge University Press, Cambridge and New York, pp. 35-53

Brown, J., M. Collins, and A. Tudhope, 2006: Coupled model simulations of mid-Holocene ENSO and comparisons with coral oxygen isotope record. *Advances in Geosciences,* **6**, 29-33

Brown, J.K.M., M. Hinkel, and F.E. Nelson, 2000: The circumpolar Active Layer Monitoring (CALM) program: research design and initial results. *Polar Geography,* **24(3)**, 166-258

Brown, J.M., A.J. Souza, and J. Wolf, 2010: Surge modelling in the eastern Irish Sea: present and future storm impact. *Ocean Dynamics,* **60(2)**, 227-236

Brown, R.D., and P.W. Mote, 2009: The Response of Northern Hemisphere Snow Cover to a Changing Climate. *Journal of Climate,* **22(8)**, 2124-2145

Brown, S.J., J. Caesar, and C.A.T. Ferro, 2008: Global changes in extreme daily temperature since 1950. *Journal of Geophysical Research-Atmospheres,* **113(D05115)**

Brunet, M., et al., 2006: The development of a new dataset of Spanish daily adjusted temperature series (SDATS) (1850-2003). *International Journal of Climatology,* **26**, 1777-1802

Bryan, K.R., P.S. Kench, and D.E. Hart, 2008: Multi-decadal coastal change in New Zealand: Evidence, mechanisms and implications. *New Zealand Geographer,* **64(2)**, 117-128

Burke, E.J., and S.J. Brown, 2008: Evaluating uncertainties in the projection of future drought. *Journal of Hydrometeorology,* **9(2)**, 292-299

Burke, E.J., S.J. Brown, and N. Christidis, 2006: Modeling the recent evolution of global drought and projections for the twenty-first century with the Hadley Centre climate model. *Journal of Hydrometeorology,* **7(5)**, 1113-1125

Burkholder, B.A., and D.J. Karoly, 2007: *An assessment of US climate variability using the Climate Extremes Index - Paper 2B.9*. In Nineteenth Conference on Climate Variability and Change, 15-18 January, 2007, San Antonio, TX, American Meteorological Society, Boston. ams.confex.com/ams/pdfpapers/117942.pdf

Burton, A., C.G. Kilsby, H.J. Fowler, P.S.P. Cowpertwait, and P.E. O'Connell, 2008: RainSim: A spatial-temporal stochastic rainfall modelling system. *Environmental Modelling & Software,* **23(12)**, 1356-1369

Caesar, J., L. Alexander, and R. Vose, 2006: Large-scale changes in observed daily maximum and minimum temperatures: Creation and analysis of a new gridded data set. *Journal of Geophysical Research,* **111(D05101)**

Caesar, J., et al., 2010: Changes in temperature and precipitation extremes over the Indo-Pacific region from 1971 to 2005. *International Journal of Climatology*, doi:10.1002/joc.2118

Callaghan, D.P., P. Nielsen, A. Short, and R. Ranasinghe, 2008: Statistical simulation of wave climate and extreme beach erosion. *Coastal Engineering,* **55(5)**, 375-390

Cambers, G., 2009: Caribbean beach changes and climate change adaptation. *Aquatic Ecosystem Health & Management,* **12(2)**, 168-176

Cameron, D., 2006: An application of the UKCIP02 climate change scenarios to flood estimation by continuous simulation for a gauged catchment in the northeast of Scotland, UK (with uncertainty). *Journal of Hydrology,* **328(1-2)**, 212-226

Cane, M.A., 2005: The evolution of El Niño, past and future. *Earth and Planetary Science Letters,* **230(3-4)**, 227-240

Cao, Z., 2008: Severe hail frequency over Ontario, Canada: Recent trend and variability. *Geophysical Research Letters,* **35(L14803)**

Caplan-Auerbach, J., and C. Huggel, 2007: Precursory seismicity associated with frequent, large ice avalanches on Iliamna volcano, Alaska, USA. *Journal of Glaciology,* **53(180)**, 128-140

Carey, M., 2005: Living and dying with glaciers: people's historical vulnerability to avalanches and outburst floods in Peru. *Global and Planetary Change,* **47(2-4)**, 122-134

Carter, T.R., R.N. Jones, X. Lu, S. Bhadwal, C. Conde, L.O. Mearns, B.C. O'Neill, M.D.A. Rounsevell, and M.B. Zurek, 2007: New assessment methods and the characterisation of future conditions. In: *Impacts, Adaptation and Vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [M.L. Parry, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp. 133-171

Cassou, C., L. Terray, and A.S. Phillips, 2005: Tropical Atlantic influence on European heat waves. *Journal of Climate,* **18(15)**, 2805-2811
Cavazos, T., C. Turrent, and D.P. Lettenmaier, 2008: Extreme precipitation trends associated with tropical cyclones in the core of the North American monsoon. *Geophysical Research Letters,* **35(L21703)**
Cayan, D.R., P.D. Bromirski, K. Hayhoe, M. Tyree, M.D. Dettinger, and R.E. Flick, 2008: Climate change projections of sea level extremes along the California coast. *Climatic Change,* **87**, S57-S73
CCSP, 2008: Weather and Climate Extremes in a Changing Climate. Regions of Focus: North America, Hawaii, Caribbean, and U.S. Pacific Islands. A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research. [T.R. Karl, et al. (eds.)]. Department of Commerce, NOAA's National Climatic Data Center, Washington, DC., pp. 164
Chan, J.C., and M. Xu, 2009: Interannual and interdecadal variations of tropical cyclone landfalling in East Asia. Part I: Time series analysis. *2008 Western Pacific Geophysics Meeting (AGU), Cairns (Australia), 29 Jul-Aug 1 2008,* **29**, 1285-1293
Chang, E.K.M., 2009: Are band-pass variance statistics useful measures of storm track activity? Re-examining storm track variability associated with the NAO using multiple storm track measures. *Climate Dynamics,* **33(2-3)**, 277-296
Chang, E.K.M., and Y. Guo, 2007: Is the number of North Atlantic tropical cyclones significantly underestimated prior to the availability of satellite observations? *Geophysical Research Letters,* **34(L14801)**
Chauvin, F., J.F. Royer, and M. Deque, 2006: Response of hurricane-type vortices to global warming as simulated by ARPEGE-Climat at high resolution. *Climate Dynamics,* **27(4)**, 377-399
Chen, C.-T., and T. Knutson, 2008: On the verification and comparison of extreme rainfall indices from climate models. *Journal of Climate,* **21**, 1605-1621
Chen, J.L., C.R. Wilson, B.D. Tapley, Z.L. Yang, and G.Y. Niu, 2009: 2005 drought event in the Amazon River basin as measured by GRACE and estimated by climate models. *Journal of Geophysical Research - Solid Earth,* **114(B05404)**
Cheng, G.D., and T.H. Wu, 2007: Responses of permafrost to climate change and their environmental significance, Qinghai-Tibet Plateau. *Journal of Geophysical Research-Earth Surface,* **112(F02S03)**
Chenoweth, M., and M. Devine, 2008: A document-based 318-year record of tropical cyclones in the Lesser Antilles, 1690-2007. *Geochemistry Geophysics Geosystems,* **9(Q08013)**
Cherchi, A., S. Masina, and A. Navarra, 2008: Impact of extreme CO2 levels on tropical climate: a CGCM study. *Climate Dynamics,* **31**, 743-758
Cherry, N.J., 1988: Comment on long-term associations between wind speeds and the urban heat-island of Phoenix, Arizona. *Journal of Applied Meteorology,* **27(7)**, 878-880
Chiapello, I., C. Moulin, and J.M. Prospero, 2005: Understanding the long-term variability of African dust transport across the Atlantic as recorded in both Barbados surface concentrations and large-scale Total Ozone Mapping Spectrometer (TOMS) optical thickness. *Journal of Geophysical Research-Atmospheres,* **110(D18S10)**
Chiarle, M., S. Iannotti, G. Mortara, and P. Deline, 2007: Recent debris flow occurrences associated with glaciers in the Alps. *Global and Planetary Change,* **56**, 123-136
Choi, G., et al., 2009: Changes in means and extreme events of temperature and precipitation in the Asia-Pacific Network region, 1955-2007. *International Journal of Climatology,* **29(13)**, 1956-1975
Christensen, J., et al., 2007: Regional Climate Projections. In: *The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [S. Solomon, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA
Christensen, J.H., and O.B. Christensen, 2003: Climate modeling: Severe summertime flooding in Europe. *Nature,* **421**, 805-806
Christensen, J.H., and O.B. Christensen, 2007: A summary of the PRUDENCE model projections of changes in European climate by the end of this century. *Climatic Change,* **81**, 7-30
Christensen, J.H., T.R. Carter, and F. Giorgi, 2002: PRUDENCE employs new methods to assess European climate change. *Eos Transactions (AGU),* **83**, 147
Church, J.A., N.J. White, and J.R. Hunter, 2006a: Sea-level rise at tropical Pacific and Indian Ocean Islands. *Global and Planetary Change,* **53**, 155-168
Church, J.A., J.R. Hunter, K.L. McInnes, and N.J. White, 2006b: Sea-level rise around the Australian coastline and the changing frequency of extreme sea-level events. *Australian Meteorological Magazine,* **55(4)**, 253-260
Church, J.A., N.J. White, J.R. Hunter, K.L. McInnes, C.P. J., and S.P. O'Farrell, 2008: Sea-level Rise. In: *Transitions: Pathways Towards Sustainable Urban Development in Australia* [E.P. Newton (ed.)]. CSIRO Publishing and Springer Science, Collingwood, Australia, and Dordrecht, The Netherlands, pp. 691
Chust, G., A. Borja, P. Liria, I. Galparsoro, M. Marcos, A. Caballero, and R. Castro, 2009: Human impacts overwhelm the effects of sea-level rise on Basque coastal habitats (N Spain) between 1954 and 2004. *Estuarine Coastal and Shelf Science,* **84(4)**, 453-462

Ciais, P., et al., 2005: Europe-wide reduction in primary productivity caused by the heat and drought in 2003. *Nature,* **437(7058)**, 529-533
Clague, J.J., and S.G. Evans, 2000: A review of catastrophic drainage of moraine-dammed lakes in British Columbia. *Quaternary Science Reviews,* **19(17-18)**, 1763-1783
Clark, R.T., S.J. Brown, and J.M. Murphy, 2006: Modeling northern hemisphere summer heat extreme changes and their uncertainties using a physics ensemble of climate sensitivity experiments. *Journal of Climate,* **19(17)**, 4418-4435
Clarke, G.K.C., 1982: Glacier Outburst Floods from' Hazard Lake', Yukon Territory, and the Problem of Flood Magnitude Prediction. *Journal of Glaciology,* **28(98)**, 3-21
Clarke, M.L., and H.M. Rendell, 2006: Effects of storminess, sand supply and the North Atlantic Oscillation on sand invasion and coastal dune accretion in western Portugal. *Holocene,* **16(3)**, 341-355
Clow, D.W., 2010: Changes in the timing of snowmelt and streamflow in Colorado: a response to recent warming. *Journal of Climate,* **23(9)**, 2293-2306
Cobb, K.M., C.D. Charles, H. Cheng, and R.L. Edwards, 2003: El Niño/Southern Oscillation and tropical Pacific climate during the last millennium. *Nature,* **424**, 271-276
Coelho, C., R. Silva, F. Veloso-Gomes, and F. Taveira-Pinto, 2009: Potential effects of climate change on northwest Portuguese coastal zones. *ICES Journal of Marine Science,* **66(7)**, 1497-1507
Coelho, C.A.S., C.A.T. Ferro, D.B. Stephenson, and D.J. Steinskog, 2008: Methods for exploring spatial and temporal variability of extreme events in climate data. *Journal of Climate,* **21(10)**, 2072-2092
Coles, S., 2001: *An Introduction to Statistical Modeling of Extreme Values*. Springer-Verlag, Heidelberg, 208 pp.
Collini, E.A., E.H. Berbery, V. Barros, and M. Pyle, 2008: How Does Soil Moisture Influence the Early Stages of the South American Monsoon? *Journal of Climate,* **21**, 195-213
Collins, D.A., P.M. Della-Marta, N. Plummer, and B.C. Trewin, 2000: Trends in annual frequencies of extreme temperature events in Australia. *Australian Meteorological Magazine,* **49(4)**, 277-292
Collins, M., B.B.B. Booth, G.R. Harris, J.M. Murphy, D.M.H. Sexton, and M.J. Webb, 2006: Towards quantifying uncertainty in transient climate change. *Climate Dynamics,* **27(2-3)**, 127-147
Compo, G.P., J.S. Whitaker, and P.D. Sardeshmukh, 2006: Feasibility of a 100-year reanalysis using only surface pressure data. *Bulletin of the American Meteorological Society,* **87**, 175-190
Conroy, J.L., A. Restrepo, J.T. Overpeck, M. Seinitz-Kannan, J.E. Cole, M.B. Bush, and P.A. Colinvaux, 2009: Unprecedented recent warming of surface temperatures in the eastern tropical Pacific Ocean. *Nature Geoscience,* **2**, 46-50
Cook, B.I., G.B. Bonan, and S. Levis, 2006: Soil moisture feedbacks to precipitation in southern Africa. *Journal of Climate,* **19(17)**, 4198-4206
Cook, K.H., and E.K. Vizy, 2006: Coupled model simulations of the west African monsoon system: Twentieth- and Twenty-First-century simulations. *Journal of Climate,* **19(15)**, 3681-3703
Coppola, E., and F. Giorgi, 2005: Climate change in tropical regions from high-resolution time-slice AGCM experiments. *Quarterly Journal of the Royal Meteorological Society,* **131(612)**, 3123-3145
Corti, T., V. Muccione, P. Köllner-Heck, D. Bresch, and S.I. Seneviratne, 2009: Simulating past droughts and associated building damages in France. *Hydrology and Earth System Sciences,* **13**, 1739-1747
Costa, A.C., and A. Soares, 2009: Trends in extreme precipitation indices derived from a daily rainfall database for the South of Portugal. *International Journal of Climatology,* **29**, 1956-1975
Cubasch, U., J. Waszkewitz, G. Hegerl, and J. Perlwitz, 1995: Regional climate changes as simulated in time-slice experiments. *Climatic Change,* **31(2-4)**, 273-304
Cunderlik, J.M., and T.B.M.J. Ouarda, 2009: Trends in the timing and magnitude of floods in Canada. *Journal of Hydrology,* **375(3-4)**, 471-480
D'Onofrio, A., J.P. Boulanger, and E.C. Segura, 2010: CHAC: a weather pattern classification system for regional climate downscaling of daily precipitation. *Climatic Change,* **98(3-4)**, 405-427
da Silva, A.E., and L.M.V. de Carvalho, 2007: Large-scale index for South America Monsoon (LISAM). *Atmospheric Science Letters,* **8(2)**, 51-57
Dai, A., T.T. Qian, K.E. Trenbeth, and J.D. Milliman, 2009: Changes in Continental Freshwater Discharge from 1948 to 2004. *Journal of Climate,* **22(10)**, 2773-2792
Dai, A.G., K.E. Trenberth, and T.T. Qian, 2004: A global dataset of Palmer Drought Severity Index for 1870-2002: Relationship with soil moisture and effects of surface warming. *Journal of Hydrometeorology,* **5(6)**, 1117-1130
Dairaku, K., S. Emori, and H. Higashi, 2008: Potential Changes in Extreme Events Under Global Climate Change. *Journal of Disaster Research,* **3(1)**, 39-50
Danard, M.B., S.K. Dube, G. Gönnert, A. Munroe, T.S. Murty, P. Chittibabu, A.D. Rao, and P.C. Sinha, 2004: Storm Surges from Extra-Tropical Cyclones. *Natural Hazards,* **32(2)**, 177-190
Dankers, R., and L. Feyen, 2008: Climate change impact on flood hazard in Europe: An assessment based on high-resolution climate simulations. *Journal of Geophysical Research-Atmospheres,* **113(D19105)**

Dankers, R., and L. Feyen, 2009: Flood hazard in Europe in an ensemble of regional climate scenarios. *Journal of Geophysical Research-Atmospheres,* **114(D16108)**
Dankers, R., O.B. Christensen, L. Feyen, M. Kalas, and A. de Roo, 2007: Evaluation of very high-resolution climate model data for simulating flood hazards in the Upper Danube Basin. *Journal of Hydrology,* **347(3-4)**, 319-331
de Elía, R., D. Caya, H. Côté, A. Frigon, S. Biner, M. Giguère, D. Paquin, R. Harvey, and D. Plummer, 2008: Evaluation of uncertainties in the CRCM-simulated North American climate. *Climate Dynamics,* **30(2-3)**, 113-132
Dean, S.M., and P.A. Stott, 2009: The Effect of Local Circulation Variability on the Detection and Attribution of New Zealand Temperature Trends. *Journal of Climate,* **22(23)**, 6217-6229
Debernard, J.B., and L.P. Roed, 2008: Future wind, wave and storm surge climate in the Northern Seas: a revisit. *Tellus Series A-Dynamic Meteorology and Oceanography,* **60(3)**, 427-438
Defeo, O., A. McLachlan, D.S. Schoeman, T.A. Schlacher, J. Dugan, A. Jones, M. Lastra, and F. Scapini, 2009: Threats to sandy beach ecosystems: A review. *Estuarine Coastal and Shelf Science,* **81(1)**, 1-12
DeGaetano, A.T., 2009: Time-dependent changes in extreme-precipitation return-period amounts in the Continental United States. *Journal of Applied Meteorology and Climatology,* **48**, 2086-2099
Delaloye, R., T. Strozzi, C. Lambiel, E. Perruchoud, and H. Raetzo, 2008: Landslide-like development of rockglaciers detected with ERS-1/2 SAR interferometry. In: *Proceedings ESA FRINGE Symposium 2007*
Delgado, J.M., H. Apel, and B. Merz, 2009: Flood trends and variability in the Mekong river. *Hydrology and Earth System Sciences,* **6**, 6691-6719
Della-Marta, P., 2006: A method of homogenizing the extremes and mean of daily temperature measurements. *Journal of Climate,* **19(17)**, 4179-4197
Della-Marta, P.M., and H. Wanner, 2006: A method of homogenizing the extremes and mean of daily temperature measurements. *Journal of Climate,* **19**, 4179-4197
Della-Marta, P.M., M.R. Haylock, J. Luterbacher, and H. Wanner, 2007a: Doubled length of western European summer heat waves since 1880. *Journal of Geophysical Research-Atmospheres,* **112(D15103)**
Della-Marta, P.M., J. Luterbacher, H. von Weissenfluh, E. Xoplaki, M. Brunet, and H. Wanner, 2007b: Summer heat waves over western Europe 1880-2003, their relationship to large-scale forcings and predictability. *Climate Dynamics,* **29(2-3)**, 251-275
Della-Marta, P.M., H. Mathis, C. Frei, M.A. Liniger, J. Kleinn, and C. Appenzeller, 2009: The return period of wind storms over Europe. *International Journal of Climatology,* **29(3)**, 437-459
DeMaria, M., J.A. Knaff, and B.H. Connell, 2001: A tropical cyclone genesis parameter for the tropical Atlantic. *Weather and Forecasting,* **16**, 219-233
Deo, R.C., J.I. Syktus, C.A. McAlpine, P.J. Lawrence, H.A. McGowan, and S.R. Phinn, 2009: Impact of historical land cover change on daily indices of climate extremes including droughts in eastern Australia. *Geophysical Research Letters,* **36(L08705)**
Department of Climate Change, 2009: *Climate Change Risks to Australia's Coast. A First Pass National Assessment*. Report published by the Australian Government's Department of Climate Change, 172 pp. http://www.climatechange.gov.au/publications (15/07/10)
Déqué, M., P. Marquet, and R.G. Jones, 1998: Simulation of climate change over Europe using a global variable resolution general circulation model. *Climate Dynamics,* **14(3)**, 173-189
Déqué, M., et al., 2007: An intercomparison of regional climate simulations for Europe: assessing uncertainties in model projections. *Climatic Change,* **81**, 53-70
Deser, C., R.A. Tomas, and S. Peng, 2007: The transient atmospheric circulation response to North Atlantic SST and sea ice anomalies. *Journal of Climate,* **20(18)**, 4751-4767
Dessai, S., M. Hulme, R. Lempert, and R. Pielke Jr, 2009: Climate prediction: a limit to adaption? In: *Adapting to Climate Change: Thresholds, Values, Goverenance* [W.N. Adger, et al. (eds.)]. Cambridge University Press, Cambridge, UK, pp. 64-78
Dibike, Y.B., P. Gachon, A. St-Hilaire, T. Ouarda, and V.T.V. Nguyen, 2008: Uncertainty analysis of statistically downscaled temperature and precipitation regimes in Northern Canada. *Theoretical and Applied Climatology,* **91(1-4)**, 149-170
Diffenbaugh, N.S., J.S. Pal, F. Giorgi, and X. Gao, 2007: Heat stress intensification in the Mediterranean climate change hotspot. *Geophysical Research Letters,* **34(L11706)**
Ding, T., W. Qian, and Z. Yan, 2009: Changes in hot days and heat waves in China during 1961-2007. *International Journal of Climatology*, doi:10.1002/joc.1989
Dirmeyer, P.A., R.D. Koster, and Z. Guo, 2006: Do Global Models Properly Represent the Feedback between Land and Atmosphere? *Journal of Hydrometeorology,* **7**, 1177-1198
Doblas-Reyes, F.J., A. Weisheimer, M. Deque, N. Keenlyside, M. McVean, J.M. Murphy, P. Rogel, D. Smith, and T.N. Palmer, 2009: Addressing model uncertainty in seasonal and annual dynamical ensemble forecasts. *Quarterly Journal of the Royal Meteorological Society,* **135(643)**, 1538-1559

Dobrovolny, P., et al., 2010: Monthly, seasonal and annual temperature reconstructions for Central Europe derived from documentary evidence and instrumental records since AD 1500. *Climatic Change,* **101**, 69-107
Donat, M.G., G.C. Leckebusch, J.G. Pinto, and U. Ulbrich, 2009: European storminess and associated atmospheric circulation: future changes in a multi-model ensemble of GCM simulation. *International Journal of Climatology,* **submitted**
Donat, M.G., G.C. Leckebusch, S. Wild, and U. Ulbrich, 2010: Future changes of European winter storm losses and extreme wind speeds in multi-model GCM and RCM simulations. *Natural Hazards and Earth System Sciences,* **submitted**
Donnelly, J.P., and J.D. Woodruff, 2007: Intense hurricane activity over the past 5,000 years controlled by El Nino and the West African monsoon. *Nature,* **447**, 465-468
Dufek, A.S., and T. Ambrizzi, 2008: Precipitation variability in Sao Paulo State, Brazil. *Theoretical and Applied Climatology,* **93(3-4)**, 167-178
Dufek, A.S., T. Ambrizzi, and R.P. da Rocha, 2008: Are Reanalysis Data Useful for Calculating Climate Indices over South America? In: *Trends and Directions in Climate Research* [L. Gimeno, et al. (eds.)]. Vol. 1146. Annals of the New York Academy of Sciences, New York, NY, USA, pp. 87-104
Durão, R.M., M.J. Pereira, A.C. Costa, J. Delgado, G. del Barrio, and A. Soares, 2009: Spatial-temporal dynamics of precipitation extremes in southern Portugal: a geostatistical assessment study. *International Journal of Climatology*, doi:10.1002/joc.1999
Durre, I., J.M. Wallace, and D.P. Lettenmaier, 2000: Dependence of extreme daily maximum temperatures on antecedent soil moisture in the contiguous United States during summer. *Journal of Climate,* **13(14)**, 2641-2651
Durre, I., M.J. Menne, and R.S. Vose, 2008: Strategies for evaluating quality assurance procedures. *Journal of Applied Meteorology and Climatology,* **47(6)**, 1785-1791
Dussaillant, A., G. Benito, W. Buytaert, P. Carling, C. Meier, and F. Espinoza, 2010: Repeated glacial-lake outburst floods in Patagonia: an increasing hazard? *Natural Hazards*, doi:10.1007/s11069-009-9479-8
Easterling, D.R., T.W.R. Wallis, J.H. Lawrimore, and R.R. Heim, 2007: Effects of precipitation and temperature trends on U.S. drought. *Geophysical Research Letters,* **34(L20709)**
Easterling, D.R., et al., 2008: Measures to Improve Our Understanding of Weather and Climate Extremes. In: *Weather and Climate Extremes in a Changing Climate. Regions of Focus: North America, Hawaii, Caribbean, and U.S. Pacific Islands*. [T.R. Karl, et al. (eds.)]. A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research, Washington, DC.
Eichler, T., and W. Higgins, 2006: Climatology and ENSO-related variability of North American extratropical cyclone activity. *Journal of Climate,* **19(10)**, 2076-2093
Elsner, J.B., J.P. Kossin, and T.H. Jagger, 2008: The increasing intensity of the strongest tropical cyclones. *Nature,* **455(7209)**, 92-95
Ely, L.L., Y. Enzel, V.R. Baker, and D.R. Cayan, 1993: A 5000-Year Record of Extreme Floods and Climate-Change in the Southwestern United-States. *Science,* **262(5132)**, 410-412
Emanuel, K., 2000: A statistical analysis of tropical cyclone intensity. *Monthly Weather Review, Boston, MA,* **128(4)**, 1139-1152
Emanuel, K., 2010: Tropical Cyclone Activity Downscaled from NOAA-CIRES Reanalysis. *Journal of Advances in Modeling Earth Systems,* **submitted**
Emanuel, K., R. Sundararajan, and J. Williams, 2008: Hurricanes and Global Warming: Results from Downscaling IPCC AR4 Simulations. *Bulletin of the American Meteorological Society,* **89(3)**, 347-367
Emanuel, K.A., 1987: Dependence of hurricane intensity on climate. *Nature,* **326(61112)**, 483-485
Emanuel, K.A., 2007: Environmental factors affecting tropical cyclone power dissipation. *Journal of Climate,* **20**, 5497-5509
Emori, S., and S.J. Brown, 2005: Dynamic and thermodynamic changes in mean and extreme precipitation under changed climate. *Geophysical Research Letters,* **32(L17706)**
Engelstaedter, S., I. Tegen, and R. Washington, 2006: North African dust emissions and transport. *Earth-Science Reviews,* **79(1-2)**, 73-100
Essery, R., N. Rutter, J. Pomeroy, R. Baxter, M. Stahli, D. Gustafsson, A. Barr, P. Bartlett, and K. Elder, 2009: An Evaluation of Forest Snow Process Simulations. *Bulletin of the American Meteorological Society,* **90(8)**, 1120-1135
Evans, S.G., N.F. Bishop, L.F. Smoll, P.V. Murillo, K.B. Delaney, and A. Oliver-Smith, 2009: A re-examination of the mechanism and human impact of catastrophic mass flows originating on Nevado Huascarán, Cordillera Blanca, Peru in 1962 and 1970. *Engineering Geology,* **108(1-2)**, 96-118
Favre, A., and A. Gershunov, 2006: Extra-tropical cyclonic/anticyclonic activity in North-Eastern Pacific and air temperature extremes in Western North America. *Climate Dynamics,* **26(6)**, 617-629

Favre, A., and A. Gershunov, 2009: North Pacific cyclonic and anticyclonic transients in a global warming context: possible consequences for Western North American daily precipitation and temperature extremes. *Climate Dynamics,* **32(7-8)**, 969-987

Fedorov, A., and P. Konstantinov, 2003: Observations of surface dynamics with thermokarst initiation, Yukechi site, central Yukutia. In: *Proceedings of the 8th International Conference on Permafrost, 21-25, July 2003, Zurich, Switzerland* [M. Philips, et al. (eds.)]. Balkema, Netherlands, pp. 239-243

Fekete, B.M., C.J. Vörösmarty, J.O. Roads, and C.J. Willmott, 2004: Uncertainties in precipitation and their impacts on runoff estimates. *Journal of Climate,* **17**, 294-304

Findell, K.L., and E.A.B. Eltahir, 2003a: Atmospheric controls on soil moisture-boundary layer interactions. Part I: Framework development. *Journal of Hydrometeorology,* **4(3)**, 552-569

Findell, K.L., and E.A.B. Eltahir, 2003b: Atmospheric controls on soil moisture-boundary layer interactions. Part II: Feedbacks within the continental United States. *Journal of Hydrometeorology,* **4(3)**, 570-583

Findell, K.L., and T.L. Delworth, 2005: A modeling study of dynamic and thermodynamic mechanisms for summer drying in response to global warming. *Geophysical Research Letters,* **32(L16702)**

Finnis, J., M.M. Holland, M.C. Serreze, and J.J. Cassano, 2007: Response of Northern Hemisphere extratropical cyclone activity and associated precipitation to climate change, as represented by the Community Climate System Model. *Journal of Geophysical Research-Biogeosciences,* **112(G04S42)**

Fiore, M.M.E., E.E. D'Onofrio, J.L. Pousa, E.J. Schnack, and G.R. Bertola, 2009: Storm surges and coastal impacts at Mar del Plata, Argentina. *Continental Shelf Research,* **29(14)**, 1643-1649

Fischer, E.M., and C. Schär, 2009: Future changes in daily summer temperature variability: driving processes and role for temperature extremes. *Climate Dynamics,* **33(7-8)**, 917-935

Fischer, E.M., and C. Schär, 2010: Consistent geographical patterns of changes in high-impact European heatwaves. *Nature Geoscience*, doi:10.1007/s00382-010-0780-8

Fischer, E.M., S.I. Seneviratne, D. Lüthi, and C. Schär, 2007a: The contribution of land-atmosphere coupling to recent European summer heatwaves. *Geophysical Research Letters,* **34(L06707)**

Fischer, E.M., S.I. Seneviratne, P.L. Vidale, D. Lüthi, and C. Schär, 2007b: Soil moisture - atmosphere interactions during the 2003 European summer heatwave. *Journal of Climate,* **20**, 5081-5099

Fischer, L., H. Eisenbeiss, A. Kääb, C. Huggel, and W. Haeberli, 2010a: Detecting topographic changes in steep high-mountain flanks using combined repeat airborne LiDAR and aerial optical imagery: a case study on climate-induced hazards at Monte Rosa east face, Italian Alps. *Permafrost and Periglacial Processes,* **submitted**

Fischer, L., R.S. Purves, C. Huggel, J. Noetzli, and W. Haeberli, 2010b: On the influence of geological, topographic and glaciological factors on slope instabilities: analyses of recent Alpine rock avalanches. *Natural Hazards and Earth System Sciences,* **submitted**

Fontaine, B., P. Roucou, M. Gaetani, and R. Marteau, 2010: Recent changes in precipitation, ITCZ convection and northern tropical circulation over North Africa (1979–2007). *International Journal of Climatology*, doi:10.1002/joc.2108

Founda, D., and C. Giannakopoulos, 2009: The exceptionally hot summer of 2007 in Athens, Greece - A typical summer in the future climate? *Global and Planetary Change,* **67(3-4)**, 227-236

Fowler, H.J., and R.L. Wilby, 2007: Beyond the downscaling comparison study. *International Journal of Climatology,* **27(12)**, 1543-1545

Fowler, H.J., and M. Ekstrom, 2009: Multi-model ensemble estimates of climate change impacts on UK seasonal precipitation extremes. *International Journal of Climatology,* **29**, 385-416

Fowler, H.J., and R.L. Wilby, 2010: Detecting changes in seasonal precipitation extremes using regional climate model projections: Implications for managing fluvial flood risk. *Water Resources Research,* **46(W0525)**

Fowler, H.J., S. Blenkinsop, and C. Tebaldi, 2007a: Linking climate change modelling to impacts studies: recent advances in downscaling techniques for hydrological modelling. *International Journal of Climatology,* **27(12)**, 1547-1578

Fowler, H.J., M. Ekstrom, S. Blenkinsop, and A.P. Smith, 2007b: Estimating change in extreme European precipitation using a multimodel ensemble. *Journal of Geophysical Research-Atmospheres,* **112(D18104)**

Frappier, A.B., T. Knutson, K.B. Liu, and K. Emanual, 2007a: Perspective: Coordinating paleoclimate research on tropical cyclones with hurricane-climate theory and modelling. *Tellus.Series A: Dynamic Meteorology and Oceanography,* **59(4)**, 529-537

Frappier, A.B., D. Sahagian, S.J. Carpenter, L.A. Gonzalez, and B.R. Frappier, 2007b: Stalagmite stable isotope record of recent tropical cyclone events. *Geology,* **35(2)**, 111-114

Frederiksen, J.S., and C.S. Frederiksen, 2007: Interdecadal changes in southern hemisphere winter storm track modes. *Tellus Series A-Dynamic Meteorology and Oceanography,* **59(5)**, 599-617

Frei, C., and C. Schär, 2001: Detection of probability of trends in rare events: Theory and application to heavy precipitation in the Alpine region. *Journal of Climate,* **14**, 1568-1584

Frei, C., R. Schöll, S. Fukutome, J. Schmidli, and P.L. Vidale, 2006: Future change of precipitation extremes in Europe: Intercomparison of scenarios from regional climate models. *Journal of Geophysical Research - Atmospheres,* **111(D06105)**
Frei, C., J.H. Christensen, M. Deque, D. Jacob, R.G. Jones, and P.L. Vidale, 2003: Daily precipitation statistics in regional climate models: Evaluation and intercomparison for the European Alps. *Journal of Geophysical Research-Atmospheres,* **108(D3)**
Frey, H., W. Haeberli, A. Linsbauer, C. Huggel, and F. Paul, 2010: A multi-level strategy for anticipating future glacier lake formation and associated hazard potentials. *Natural Hazards and Earth System Sciences,* **10(2)**, 339-352
Frich, P., L.V. Alexander, P.M. Della-Marta, B. Gleason, M. Haylock, A.M.G. Klein Tank, and T. Peterson, 2002: Observed coherent changes in climatic extremes during the second half of the twentieth century. *Climate Research,* **19**, 193-212
Fu, C.B., et al., 2005: Regional climate model intercomparison project for Asia. *Bulletin of the American Meteorological Society,* **86(2)**, 257-266
Fujihara, Y., K. Tanaka, T. Watanabe, T. Nagano, and T. Kojiri, 2008: Assessing the impacts of climate change on the water resources of the Seyhan River Basin in Turkey: Use of dynamically downscaled data for hydrologic simulations. *Journal of Hydrology,* **353(1-2)**, 33-48
Fyfe, J.C., 2003: Extratropical southern hemisphere cyclones: Harbingers of climate change? *Journal of Climate,* **16(17)**, 2802-2805
Gan, M.A., V.B. Rao, and M.C.L. Moscati, 2006: South American monsoon indices. *Atmospheric Science Letters,* **6**, 219-223
Ganguly, A.R., K. Steinhaeuser, D.J. Erickson, M. Branstetter, E.S. Parish, N. Singh, J.B. Drake, and L. Buja, 2009: Higher trends but larger uncertainty and geographic variability in 21st century temperature and heat waves. *Proceedings of the National Academy of Sciences of the United States of America,* **106(37)**, 15555-15559
Gastineau, G., and B.J. Soden, 2009: Model projected changes of extreme wind events in response to global warming. *Geophysical Research Letters,* **36(L10810)**
Gavriliev, P.P., and P.V. Efremov, 2003: Effects of cryogenic processes on Yakuita landscapes under climate warming. In: *Proceedings of the 8th International Conference on Permafrost, 21-25, July 2003, Zurich, Switzerland* [M. Philips, et al. (eds.)]. Balkema, Netherlands, pp. 277-282
Gedney, N., P.M. Cox, R.A. Betts, O. Boucher, C. Huntingford, and P.A. Stott, 2006: Detection of a direct carbon dioxide effect in continental river runoff records. *Nature,* **439(7078)**, 835-838
Giannakopoulos, C., P. Le Sager, M. Bindi, M. Moriondo, E. Kostopoulou, and C.M. Goodess, 2009: Climatic changes and associated impacts in the Mediterranean resulting from a 2°C global warming. *Global and Planetary Change,* **68(3)**, 209-224
Giannini, A., R. Saravanan, and P. Chang, 2003: Oceanic forcing of Sahel rainfall on interannual to interdecadal time scales. *Science,* **302(5647)**, 1027-1030
Giannini, A., M. Biasutti, and M.M. Verstraete, 2008: A climate model-based review of drought in the Sahel: Desertification, the re-greening and climate change. *Global and Planetary Change,* **64(3-4)**, 119-128
Gibelin, A.L., and M. Deque, 2003: Anthropogenic climate change over the Mediterranean region simulated by a global variable resolution model. *Climate Dynamics,* **20(4)**, 327-339
Gillett, N.P., R.J. Allan, and T.J. Ansell, 2005: Detection of external influence on sea level pressure with a multi-model ensemble. *Geophysical Research Letters,* **32(L19714)**
Gillett, N.P., P.A. Stott, and B.D. Santer, 2008a: Attribution of cyclogenesis region sea surface temperature change to anthropogenic influence. *Geophysical Research Letters,* **35(L09707)**
Gillett, N.P., D.A. Stone, P.A. Stott, T. Nozawa, A.Y. Karpechko, G.C. Hegerl, M.F. Wehner, and P.D. Jones, 2008b: Attribution of polar warming to human influence. *Nature Geoscience,* **1**, 750-754
Ginoux, P., J.M. Prospero, O. Torres, and M. Chin, 2004: Long-term simulation of global dust distribution with the GOCART model: correlation with North Atlantic Oscillation. *Environmental Modelling & Software,* **19(2)**, 113-128
Giorgi, F., 2006: Climate change hot-spots. *Geophysical Research Letters,* **33(L08707)**
Giorgi, F., and P. Lionello, 2008: Climate change projections for the Mediterranean region. *Global and Planetary Change,* **63(2-3)**, 90-104
Giorgi, F., C. Jones, and G.R. Asra, 2009: Addressing climate information needs at the regional level: the CORDEX framework. *WMO Bulletin,* **58(3)**, 175-183
Glade, T., 1998: Establishing the frequency and magnitude of landslide-triggering rainstorm events in New Zealand. *Environmental Geology,* **55(2)**, 160-174
Gleason, K.L., J.H. Lawrimore, D.H. Levinson, and T.R. Karl, 2008: A Revised U.S. Climate Extremes Index. *Journal of Climate,* **21**, 2124-2137
Glecker, P.J., K.E. Taylor, and C. Doutriaux, 2008: Performance metrics for climate models. *Journal of Geophysical Research,* **113(D06104)**

Gong, S.L., X.Y. Zhang, T.L. Zhao, X.B. Zhang, L.A. Barrie, I.G. McKendry, and C.S. Zhao, 2006: A simulated climatology of Asian dust aerosol and its trans-Pacific transport. Part II: Interannual variability and climate connections. *Journal of Climate,* **19(1)**, 104-122

Goodkin, N.F., K.A. Hughen, S.C. Doney, and W.B. Curry, 2008: Increased multidecadal variability of the North Atlantic Oscillation since 1781. *Nature Geoscience,* **1**, 844-848

Goswami, B.N., V. Venugopal, D. Sengupta, M.S. Madhusoodanan, and P.K. Xavier, 2006: Increasing trend of extreme rain events over India in a warming environment. *Science,* **314(5804)**, 1442-1445

Goubanova, K., and L. Li, 2007: Extremes in temperature and precipitation around the Mediterranean basin in an ensemble of future climate scenario simulations. *Global and Planetary Change,* **57(1-2)**, 27-42

Goudie, A.S., 2009: Dust storms: Recent developments. *Journal of Environmental Management,* **90(1)**, 89-94

Goudie, A.S., and N.J. Middleton, 1992: The Changing Frequency of Dust Storms through Time. *Climatic Change,* **20(3)**, 197-225

Grabemann, I., and R. Weisse, 2008: Climate change impact on extreme wave conditions in the North Sea: an ensemble study. *Ocean Dynamics,* **58(3-4)**, 199-212

Graham, L.P., J. Andreasson, and B. Carlsson, 2007: Assessing climate change impacts on hydrology from an ensemble of regional climate models, model scales and linking methods - a case study on the Lule River basin. *Climatic Change,* **81**, 293-307

Graham, N.E., and H.F. Diaz, 2001: Evidence for intensification of North Pacific winter cyclones since 1948. *Bulletin of the American Meteorological Society,* **82(9)**, 1869-1893

Gravelle, G., and N. Mimura, 2008: Vulnerability assessment of sea-level rise in Viti Levu, Fiji Islands. *Sustainability Science,* **3(2)**, 171-180

Green, D., L. Alexander, K.L. McInnes, J. Church, N. Nicholls, and W. White, 2009: An assessment of climate change impacts and adaptation for the Torres Strait Islands, Australia. *Climatic Change*, doi:10.1007/s10584-009-9756-2

Greene, A.M., A. Giannini, and S.E. Zebiak, 2009: Drought return times in the Sahel: A question of attribution. *Geophysical Research Letters,* **36(L12701)**

Grell, G.A., L. Schade, R. Knoche, A. Pfeiffer, and J. Egger, 2000: Nonhydrostatic climate simulations of precipitation over complex terrain. *Journal of Geophysical Research-Atmospheres,* **105(D24)**, 29595-29608

Griffiths, G.M., et al., 2005: Change in mean temperature as a predictor of extreme temperature change in the Asia-Pacific region. *International Journal of Climatology,* **25**, 1301-1330

Grimm, A., J. Pal, and F. Giorgi, 2007: Connection between Spring Conditions and Peak Summer Monsoon Rainfall in South America: Role of Soil Moisture, Surface Temperature, and Topography in Eastern Brazil. *Journal of Climate,* **20**, 5929-5945

Grimm, A.M., and R.G. Tedeschi, 2009: ENSO and Extreme Rainfall Events in South America. *Journal of Climate,* **22(7)**, 1589-1609

Groisman, P.Y., and R.W. Knight, 2008: Prolonged dry episodes over the conterminous United States: New tendencies emerging during the last 40 years. *Journal of Climate,* **21(9)**, 1850-1862

Gruber, S., and W. Haeberli, 2007: Permafrost in steep bedrock slopes and its temperature-related destabilization following climate change. *Journal of Geophysical Research,* **112(F02S18)**

Gruber, S., M. Hoelzle, and W. Haeberli, 2004a: Permafrost thaw and destabilization of Alpine rock walls in the hot summer of 2003. *Geophysical Research Letters,* **31(L13504)**

Gruber, S., M. Hoelzle, and W. Haeberli, 2004b: Rock-wall temperatures in the Alps: modelling their topographic distribution and regional differences. *Permafrost and Periglacial Processes,* **15(3)**, 299-307

Gualdi, S., E. Scoccimarro, and A. Navarra, 2008: Changes in Tropical Cyclone Activity due to Global Warming: Results from a High-Resolution Coupled General Circulation Model. *Journal of Climate,* **21(20)**, 5204-5228

Gulev, S.K., and V. Grigorieva, 2004: Last century changes in ocean wind wave height from global visual wave data. *Geophysical Research Letters,* **31(L24302)**

Guo, H., M. Xu, and Q. Hub, 2010: Changes in near-surface wind speed in China: 1969–2005. *International Journal of Climatology*, doi:10.1002/joc.2091

Gutowski, W.J., L.E. Branscome, and D.A. Stewart, 1992: Life-cycles of moist baroclinic eddies. *Journal of the Atmospheric Sciences,* **49(4)**, 306-319

Gutowski, W.J., S.G. Decker, R.A. Donavon, Z.T. Pan, R.W. Arritt, and E.S. Takle, 2003: Temporal-spatial scales of observed and simulated precipitation in central U.S. climate. *Journal of Climate,* **16(22)**, 3841-3847

Gutowski, W.J., G.C. Hegerl, G.J. Holland, T.R. Knutson, L.O. Mearns, R.J. Stouffer, P.J. Webster, M.F. Wehner, and F.W. Zwiers, 2008a: Causes of Observed Changes in Extremes and Projections of Future Changes. In: *Weather and Climate Extremes in a Changing Climate. Regions of Focus: North America, Hawaii, Caribbean, and U.S. Pacific Islands*. [T.R. Karl, et al. (eds.)]. A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research, Washington, DC.

Gutowski, W.J., Jr., S.S. Willis, J.C. Patton, B.R.J. Schwedler, R.W. Arritt, and E.S. Takle, 2008b: Changes in extreme, cold-season synoptic precipitation events under global warming. *Geophysical Research Letters,* **35(L20710)**

Haarsma, R.J., F. Selten, B.V. Hurk, W. Hazeleger, and X.L. Wang, 2009: Drier Mediterranean soils due to greenhouse warming bring easterly winds over summertime central Europe. *Geophysical Research Letters,* **36(L04705)**
Haeberli, W., 1983: Frequency and characteristics of glacier floods in the Swiss Alps. *Annals of Glaciology,* **4**, 85-90
Haeberli, W., and M. Funk, 1991: Borehole temperatures at the Colle Gnifetti core-drilling site(Monte Rosa, Swiss Alps). *Journal of Glaciology,* **37(125)**, 37-46
Haeberli, W., and M. Beniston, 1998: Climate change and its impacts on glaciers and permafrost in the Alps. *Ambio*, 258-265
Haeberli, W., and C.R. Burn, 2002: Natural hazards in forests: glacier and permafrost effects as related to climate change. In: *Environmental Change and Geomorphic Hazards in Forests, IUFRO Research Series 9* [R.C. Sidle (ed.)]. CABI Publishing, Wallingford/New York, pp. 167-202
Haeberli, W., and R. Hohmann, 2008: Climate, glaciers and permafrost in the Swiss Alps 2050: scenarios, consequences and recommendations. In: *Ninth International Conference on Permafrost* [D.L. Kane, and K.M. Hinkel (eds.)]. Vol. 1. Institue of Northern Engineering, University of Alaska, Fairbanks, pp. 607-612
Haeberli, W., D. Rickenmann, and M. Zimmerman, 1990: Investigation of 1987 debris flows in the Swiss Alps: general concept and geophysical soundings. *Hydrology of Mountainous [Regions-II:] Artificial Reservoirs, Water and Slopes. [IAHS] Publication,* **194**
Haeberli, W., M. Wegmann, and D. Vonder Mühll, 1997: Slope stability problems related to glacier shrinkage and permafrost degradation in the Alps. *Eclogae Geologicae Helvetiae,* **90(3)**, 407-414
Haeberli, W., A. Kääb, D. Vonder Mühll, and P. Teysseire, 2001: Prevention of outburst floods from periglacial lakes at Grubengletscher, Valais, Swiss Alps. *Journal of Glaciology,* **47(156)**, 111-122
Haeberli, W., C. Huggel, A. Kääb, S. Zgraggen-Oswald, A. Polkvoj, I. Galushkin, I. Zotikov, and N. Osokin, 2004: The Kolka-Karmadon rock/ice slide of 20 September 2002: an extraordinary event of historical dimensions in North Ossetia, Russian Caucasus. *Journal of Glaciology,* **50(171)**, 533-546
Haigh, I., R. Nicholls, and N. Wells, 2009: Mean sea level trends around the English Channel over the 20th century and their wider context. *Continental Shelf Research,* **29(17)**, 2083-2098
Haigh, I., R. Nicholls, and N. Wells, 2010: Assessing changes in extreme sea levels: Application to the English Channel, 1900–2006. *Continental Shelf Research*, doi:10.1016/j.csr.2010.02.002
Hall, A., X. Qu, and J.D. Neelin, 2008: Improving predictions of summer climate change in the United States. *Geophysical Research Letters,* **35(L01702)**
Hanna, E., J. Cappelen, R. Allan, T. Jonsson, F. Le Blancq, T. Lillington, and K. Hickey, 2008: New Insights into North European and North Atlantic Surface Pressure Variability, Storminess, and Related Climatic Change since 1830. *Journal of Climate,* **21(24)**, 6739-6766
Hannaford, J., and T.J. Marsh, 2008: High-flow and flood trends in a network of undisturbed catchments in the UK. *International Journal of Climatology,* **28(10)**, 1325-1338
Harper, B., T. Hardy, L. Mason, and R. Fryar, 2009: Developments in storm tide modelling and risk assessment in the Australian region. *Natural Hazards,* **51(1)**, 225-238
Harris, C., et al., 2009: Permafrost and climate in Europe: Monitoring and modelling thermal, geomorphological and geotechnical responses. *Earth-Science Reviews,* **92(3-4)**, 117-171
Haugen, J.E., and T. Iverson, 2008: Response in extremes of daily precipitation and wind from a downscaled multi-model ensemble of anthropogenic global climate change scenarios. *Tellus (A Dyn.Meteorol.Oceanogr.),* **60**, 411-426
Hawkins, E., and R. Sutton, 2009: The Potential to Narrow Uncertainty in Regional Climate Predictions. *Bulletin of the American Meteorological Society,* **90(8)**, 1095-1107
Hawkins, E., and R. Sutton, 2010: The potential to narrow uncertainty in projections of regional precipitation change. *Climate Dynamics*, doi:10.1007/s00382-010-0810-6
Hay, L.E., M.P. Clark, M. Pagowski, G.H. Leavesley, and W.J. Gutowski, 2006: One-way coupling of an atmospheric and a hydrologic model in Colorado. *Journal of Hydrometeorology,* **7(4)**, 569-589
Haylock, M., and N. Nicholls, 2000: Trends in extreme rainfall indices for an updated high quality data set for Australia, 1910-1998. *International Journal of Climatology,* **20(13)**, 1533-1541
Haylock, M.R., G.C. Cawley, C. Harpham, R.L. Wilby, and C.M. Goodess, 2006a: Downscaling heavy precipitation over the United Kingdom: A comparison of dynamical and statistical methods and their future scenarios. *International Journal of Climatology,* **26(10)**, 1397-1415
Haylock, M.R., N. Hofstra, A. Tank, E.J. Klok, P.D. Jones, and M. New, 2008: A European daily high-resolution gridded data set of surface temperature and precipitation for 1950-2006. *Journal of Geophysical Research-Atmospheres,* **113(D20119)**
Haylock, M.R., et al., 2006b: Trends in total and extreme South American rainfall in 1960-2000 and links with sea surface temperature. *Journal of Climate,* **19(8)**, 1490-1512
Hegerl, G.C., O. Hoegh-Guldberg, G. Casassa, M.P. Hoerling, R.S. Kovats, C. Parmesan, D.W. Pierce, and P.A. Stott, 2010: Good Practice Guidance Paper on Detection and Attribution Related to Anthropogenic Climate Change. In: *Meeting Report of the Intergovernmental Panel on Climate Change Expert Meeting on Detection and*

Attribution of Anthropogenic Climate Change [T.F. Stocker, et al. (eds.)]. IPCC Working Group I Technical Support Unit, University of Bern, Bern, Switzerland

Hegerl, G.C., F.W. Zwiers, P. Braconnot, N.P. Gillett, Y. Luo, J.A. Marengo Orsini, N. Nicholls, J.E. Penner, and P.A. Stott, 2007: Understanding and Attributing Climate Change. In: *The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [S. Solomon, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA

Hegglin, E., and C. Huggel, 2008: An Integrated Assessment of Vulnerability to Glacial Hazards. *Mountain Research and Development,* **28(3)**, 299-309

Heim Jr, R.R., 2002: A Review of Twentieth-Century Drought Indices Used in the United States. *Bulletin of the American Meteorological Society,* **83**, 1149-1165

Heino, R., et al., 1999: Progress in the study of climatic extremes in northern and central Europe. *Climatic Change,* **42(1)**, 151-181

Held, I.M., and B.J. Soden, 2006: Robust responses of the hydrological cycle to global warming. *Journal of Climate,* **19(21)**, 5686-5699

Hemer, M.A., K. McInnes, and R. Ranasinghe, 2009: *Future Projections of the East Australian wave climate*. In Proceedings of the 11th International waves conference, Halifax.

Hemer, M.A., J.A. Church, and J.R. Hunter, 2010: Variability and trends in the directional wave climate of the Southern Hemisphere. *International Journal of Climatology,* **30(4)**, 475-491

Hendon, H.H., D.W.J. Thompson, and M.C. Wheeler, 2007: Australian rainfall and surface temperature variations associated with the Southern Hemisphere annular mode. *Journal of Climate,* **20(11)**, 2452-2467

Hennessy, K., B. Fitzharris, B.C. Bates, N. Harvey, S.M. Howden, L. Hughes, J. Salinger, and R. Warrick, 2007: Australia and New Zealand. In: *Impacts, Adaptation and Vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [M.L. Parry, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp. 507-540

Hewitson, B.C., and R.G. Crane, 2006: Consensus between GCM climate change projections with empirical downscaling: Precipitation downscaling over South Africa. *International Journal of Climatology,* **26(10)**, 1315-1337

Hidalgo, H.G., et al., 2009: Detection and Attribution of Streamflow Timing Changes to Climate Change in the Western United States. *Journal of Climate,* **22(13)**, 3838-3855

Hinzman, L.D., et al., 2005: Evidence and implications of recent climate change in northern Alaska and other arctic regions. *Climatic Change,* **72(3)**, 251-298

Hirabayashi, Y., S. Kanae, K. Motoya, K. Masuda, and P. Doll, 2008a: A 59-year (1948-2006) global meteorological forcing data set for land surface models. Part II: Global snowfall estimation. *Hydrological Research Letters,* **2**, 65-69

Hirabayashi, Y., S. Kanae, S. Emori, T. Oki, and M. Kimoto, 2008b: Global projections of changing risks of floods and droughts in a changing climate. *Hydrological Sciences Journal,* **53(4)**, 754-772

Hirsch, M.E., A.T. DeGaetano, and S.J. Colucci, 2001: An East Coast winter storm climatology. *Journal of Climate,* **14(5)**, 882-899

Hoerling, M., J. Hurrell, J. Eischeid, and A. Phillips, 2006: Detection and attribution of twentieth-century northern and southern African rainfall change. *Journal of Climate,* **19(16)**, 3989-4008

Hohenegger, C., P. Brockhaus, and C. Schär, 2008: Towards climate simulations at cloud-resolving scales. *Meteorology Zeitschrift,* **17(4)**, 383-394

Hohenegger, C., P. Brockhaus, C.S. Bretherton, and C. Schar, 2009: The Soil Moisture-Precipitation Feedback in Simulations with Explicit and Parameterized Convection. *Journal of Climate,* **22(19)**, 5003-5020

Holland, G.J., and P.J. Webster, 2007: Heightened tropical cyclone activity in the North Atlantic: Natural variability or climate trend? *AMS Forum: Climate Change Manifested by Changes in Weather (87. AMS Meeting), San Antonio, TX (USA), 13-18 Jan 2007,* **365**, 2695-2716

Holm, K., M. Bovis, and M. Jakob, 2004: The landslide response of alpine basins to post-Little Ice Age glacial thinning and retreat in southwestern British Columbia. *Geomorphology,* **57(3-4)**, 201-216

Hope, P.K., W. Drosdowsky, and N. Nicholls, 2006: Shifts in the synoptic systems influencing southwest Western Australia. *Climate Dynamics,* **26(7-8)**, 751-764

Hossain, F., I. Jeyachandran, and R. Pielke Jr, 2009: Have large dams altered extreme precipitation patterns. *Eos Transactions (AGU),* **90(48)**

Hourdin, F., et al., 2010: Amma-model intercomparison project. *Bulletin of the American Meteorological Society,* **91(1)**, 95-104

Hubbard, B., A. Heald, J.M. Reynolds, D. Quincey, S.D. Richardson, M.Z. Luyo, N.S. Portilla, and M.J. Hambrey, 2005: Impact of a rock avalanche on a moraine-dammed proglacial lake: Laguna Safuna Alta, Cordillera Blanca, Peru. *Earth Surface Processes and Landforms,* **30(10)**

Huebener, H., U. Cubasch, U. Langematz, T. Spangehl, F. Niehoerster, I. Fast, and M. Kunze, 2007: Ensemble climate simulations using a fully coupled ocean-troposphere-stratosphere general circulation model. *Philosophical Transactions of the Royal Society A-Mathematical Physical and Engineering Sciences,* **365(1857)**, 2089-2101
Huggel, C., 2009: Recent extreme slope failures in glacial environments: effects of thermal perturbation. *Quaternary Science Reviews,* **28(11-12)**, 1119-1130
Huggel, C., J. Caplan-Auerbach, and R. Wessels, 2008: Recent Extreme Avalanches: Triggered by Climate Change. *Eos Transactions (AGU),* **89(47)**, 469-470
Huggel, C., O. Korup, and S. Gruber, 2010a: *Landslide hazards and climate change in high mountains*. Treatise on Geomophology. Elsevier Science (in press)
Huggel, C., W. Haeberli, A. Kääb, D. Bieri, and S. Richardson, 2004: An assessment procedure for glacial hazards in the Swiss Alps. *Canadian Geotechnical Journal,* **41(6)**, 1068-1083
Huggel, C., S. Zgraggen-Oswald, W. Haeberli, A. Kääb, A. Polkvoj, I. Galushkin, and S.G. Evans, 2005: The 2002 rock/ice avalanche at Kolka/Karmadon, Russian Caucasus: assessment of extraordinary avalanche formation and mobility, and application of QuickBird satellite imagery. *Natural Hazards and Earth System Sciences,* **5(2)**, 173-187
Huggel, C., N. Salzmann, S. Allen, J. Caplan-Auerbach, L. Fischer, W. Haeberli, C. Larsen, D. Schneider, and R. Wessels, 2010b: Recent and future warm extreme events and high-mountain slope failures. *Philosophical Transactions of the Royal Society A,* **368**, 2435-2459
Hundecha, Y., and A. Bardossy, 2008: Statistical downscaling of extreme of daily precipitation and temperature and construction of their future scenarios. *International Journal of Climatology,* **28**, 589-610
Hundecha, Y., A. St-Hilaire, T.B.M.J. Ouarda, S. El Adlouni, and P. Gachon, 2008: A Nonstationary Extreme Value Analysis for the Assessment of Changes in Extreme Annual Wind Speed over the Gulf of St. Lawrence, Canada. *Journal of Applied Meteorology and Climatology,* **47(11)**, 2745-2759
Hungate, B.A., J.S. Dukes, M.R.L.Y. Shaw, and C.B. Field, 2003: Nitrogen and climate change. *Science,* **302**, 1512-1513
Hunter, J., 2010: Estimating sea-level extremes under conditions of uncertain sea-level rise. *Climatic Change,* **99**, 331-350
Huntingford, C., J.A. Lowe, B.B.B. Booth, C.D. Jones, G.R. Harris, L.K. Gohar, and P. Meir, 2009: Contributions of carbon cycle uncertainty to future climate projection spread. *Tellus Series B-Chemical and Physical Meteorology,* **61(2)**, 355-360
Huss, M., D. Farinotti, A. Bauder, and M. Funk, 2008: Modelling runoff from highly glacierized alpine drainage basins in a changing climate. *Hydrological Processes,* **22(19)**, 3888-3902
Im, E.S., and W.-T. Kwon, 2007: Characteristics of Extreme Climate Sequences over Korea Using a Regional Climate Change Scenario. *SOLA,* **3**, 17-20
Im, E.S., I.W. Jung, and D.H. Bae, 2010: The temporal and spatial structures of recent and future trends in extreme indices over Korea from a regional climate projection. *International Journal of Climatology*, doi:10.1002/joc.2063
Im, E.S., J.B. Ahn, W.T. Kwon, and F. Giorgi, 2008: Multi-decadal scenario simulation over Korea using a one-way double-nested regional climate model system. Part 2: future climate projection (2021-2050). *Climate Dynamics,* **30(2-3)**, 239-254
Inoue, J., J.P. Liu, J.O. Pinto, and J.A. Curry, 2006: Intercomparison of Arctic Regional Climate Models: Modeling clouds and radiation for SHEBA in May 1998. *Journal of Climate,* **19(17)**, 4167-4178
IPCC, 2007a: *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor and H.L. Miller (eds.)]*. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 996 pp.
IPCC, 2007b: Summary for Policymakers. In: *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [S. Solomon, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp. 1-18
Iverson, R.M., 2000: Landslide triggering by rain infiltration. *Water Resources Research,* **36(7)**, 1897-1910
Jaeger, E.B., and S.I. Seneviratne, 2010: Impact of soil moisture-atmosphere coupling on European climate extremes and trends in a regional climate model. *Climate Dynamics*, doi:10.1007/s00382-010-0780-8
Jakob, D., 2010: Challenges in developing a high-quality surface wind-speed data set for Australia. *Geophysical Research Letters,* **submitted**
Janicot, S., 2009: A comparison of Indian and African monsoon variability at different time scales. (Une comparaison de la variabilité des moussons indienne et africaine à différentes échelles de temps). *Comptes Rendus Geoscience,* **341(8-9)**, 575-590

Jansen, E., et al., 2007: Palaeoclimate. In: *The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [S. Solomon, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA
Jiang, T., Z.W. Kundzewicz, and B. Su, 2008: Changes in monthly precipitation and flood hazard in the Yangtze River Basin, China. *International Journal of Climatology,* **28(11)**, 1471-1481
Joly, M., and A. Voldoire, 2009: Role of the Gulf of Guinea in the inter-annual variability of the West African monsoon: what do we learn from CMIP3 coupled simulations? *International Journal of Climatology*, doi:10.1002/joc.2026
Jomelli, V., V.P. Pech, C. Chochillon, and D. Brunstein, 2004: Geomorphic variations of debris flows and recent climatic change in the French Alps. *Climatic Change,* **64(1)**, 77-102
Jones, G.S., P.A. Stott, and N. Christidis, 2008: Human contribution to rapidly increasing frequency of very warm Northern Hemisphere summers. *Journal of Geophysical Research,* **113(D02109)**
Jones, P.D., E.M. Horton, C.K. Folland, M. Hulme, D.E. Parker, and T.A. Basnett, 1999: The use of indices to identify changes in changes in climatic extremes. *Climate Change,* **42(1)**, 131-149
Jorgenson, M.T., and J. Brown, 2005: Classification of the Alaskan Beaufort Sea Coast and estimation of carbon and sediment inputs from coastal erosion. *Geo-Marine Letters,* **25(2-3)**, 69-80
Joseph, R., and S. Nigam, 2006: ENSO evolution and teleconnections in IPCC's twentieth-century climate simulations: realistic representation? *Journal of Climate,* **19**, 4360-4377
Kääb, A., R. Frauenfelder, and I. Roer, 2007: On the response of rockglacier creep to surface temperature increase. *Global and Planetary Change,* **56(1-2)**, 172-187
Kallis, G., 2008: Droughts. *Annual Review of Environment and Resources,* **33**, 85-118
Kamiguchi, K., A. Kitoh, T. Uchiyama, R. Mizuta, and A. Noda, 2006: Changes in Precipitation-based Extremes Indices Due to Global Warming Projected by a Global 20-km-mesh Atmospheric Model. *SOLA,* **2**, 64-67
Kanada, S., M. Nakano, and T. Kato, 2009: Mean atmospheric structures around Japan during July in global warming experiments using a semi-cloud resolving model. *Hydrological Research Letters,* **submitted**
Kang, I.S., and J. Shukla, 2006: Dynamic seasonal prediction and predictability of the monsoon. In: *The Asian Monsoon* [B. Wang (ed.)]. Springer/Praxis, New York, pp. 585-612
Karl, T.R., J.M. Melillo, and T.C. Peterson (eds.), 2009: *Global Climate Change Impacts in the United States*. Cambridge University Press, Cambridge and New York, 189 pp. http://downloads.globalchange.gov/usimpacts/pdfs/climate-impacts-report.pdf (15/07/10)
Karl, T.R., G.A. Meehl, D.M. Christopher, S.J. Hassol, A.M. Waple, and W.L. Murray, 2008: *Weather and Climate Extremes in a Changing Climate. Regions of Focus: North America, Hawaii, Caribbean, and U.S. Pacific Islands*. A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research, Washington, DC., 164 pp. http://www.climatescience.gov/Library/sap/sap3-3/final-report/sap3-3-final-all.pdf (15/07/10)
Karoly, D.J., and Q.G. Wu, 2005: Detection of regional surface temperature trends. *Journal of Climate,* **18(21)**, 4337-4343
Kaser, G., J.G. Cogley, M.B. Dyurgerov, M.F. Meier, and A. Ohmura, 2006: Mass balance of glaciers and ice caps: Consensus estimates for 1961-2004. *Geophysical Research Letters,* **33(L19501)**
Katz, R.W., and B.G. Brown, 1992: Extreme events in a changing climate: Variability is more important than averages. *Climatic Change,* **21(3)**, 289-302
Kay, A.L., H.N. Davies, V.A. Bell, and R.G. Jones, 2009: Comparison of uncertainty sources for climate change impacts: flood frequency in England. *Climatic Change,* **92(1-2)**, 41-63
Keenlyside, N.S., M. Latif, J. Jungclaus, L. Kornblueh, and E. Roeckner, 2008: Advancing decadal-scale climate prediction in the North Atlantic sector. *Nature,* **453**, 84-88
Kellogg, C.A., and D.W. Griffin, 2006: Aerobiology and the global transport of desert dust. *Trends in Ecology & Evolution,* **21(11)**, 638-644
Kendon, E.J., D.P. Rowell, and R.G. Jones, 2009: Mechanisms and reliability of future projected changes in daily precipitation. *Climate Dynamics*, doi:10.1007/s00382-009-0639-z
Kendon, E.J., D.P. Rowell, R.G. Jones, and E. Buonomo, 2008: Robustness of Future Changes in Local Precipitation Extremes. *Journal of Climate,* **21(17)**, 4280-4297
Kenyon, J., and G.C. Hegerl, 2008: Influence of modes of climate variability on global temperature extremes. *Journal of Climate,* **21**, 3872-3889
Kershaw, J.A., J.J. Clague, and S.G. Evans, 2005: Geomorphic and sedimentological signature of a two-phase outburst flood from moraine-dammed Queen Bess Lake, British Columbia, Canada. *Earth Surface Processes and Landforms,* **30(1)**, 1-25
Kharin, V., F.W. Zwiers, X. Zhang, and G.C. Hegerl, 2007: Changes in temperature and precipitation extremes in the IPCC ensemble of global coupled model simulations. *Journal of Climate,* **20**, 1419-1444
Kharin, V.V., and F.W. Zwiers, 2000: Changes in the extremes in an ensemble of transient climate simulation with a coupled atmosphere-ocean GCM. *Journal of Climate,* **13**, 3760-3788

Kharin, V.V., and F.W. Zwiers, 2005: Estimating extremes in transient climate change simulations. *Journal of Climate,* **18**, 1156-1173
Khon, V.C., I.I. Mokhov, E. Roeckner, and V.A. Semenov, 2007: Regional changes of precipitation characteristics in Northern Eurasia from simulations with global climate model. *Global and Planetary Change,* **57(1-2)**, 118-123
Kiktev, D., D.M.H. Sexton, L. Alexander, and C.K. Folland, 2003: Comparison of modeled and observed trends in indices of daily climate extremes. *Journal of Climate,* **16(22)**, 3560-3571
Kiktev, D.M., J. Caesar, and L. Alexander, 2009: Temperature and precipitation extremes in the second half of the twentieth century from numerical modeling results and observational data. *Izvestiya Atmospheric and Oceanic Physics,* **45**, 284-293
Kilsby, C., P. Jones, A. Burton, A. Ford, H. Fowler, C. Harpham, P. James, A. Smith, and R. Wilby, 2007: A daily weather generator for use in climate change studies. *Environmental Modelling and Software,* **22**, 1705-1719
Kim, D.W., and H.R. Byun, 2009: Future pattern of Asian drought under global warming scenario. *Theoretical and Applied Climatology,* **98(1-2)**, 137-150
Kim, H.J., B. Wang, and Q.H. Ding, 2008: The Global Monsoon Variability Simulated by CMIP3 Coupled Climate Models. *Journal of Climate,* **21(20)**, 5271-5294
Kim, H.J., R.C. Sidle, R.D. Moore, and R. Hudson, 2004: Throughflow variability during snowmelt in a forested mountain catchment, coastal British Columbia, Canada. *Hydrological Processes,* **18(7)**, 1219-1236
Kim, S.K., W.P. Hong, and Y.M. Kim, 1991: Prediction of rainfall-triggered landslides in Korea. In: *Proceedings of the 6th International Symposium on Landslides, Christchurch, New Zealand* [D.H. Bell (ed.)]. Balkema, Rotterdam, pp. 989-994
Kimoto, M., N. Yasutomi, C. Yokoyama, and S. Emori, 2005: Projected Changes in Precipitation Characteristics around Japan under the Global Warming. *SOLA,* **1**, 85-88
Kitoh, A., and T. Uchiyama, 2006: Changes in onset and withdrawal of the East Asian summer rainy season by multi-model global warming experiments. *Journal of the Meteorological Society of Japan,* **84(2)**, 247-258
Kitoh, A., T. Ose, K. Kurihara, S. Kusunoki, M. Sugi, and G. Kakushin Team-3 Modeling, 2009: Projection of changes in future weather extremes using super-high-resolution global and regional atmospheric models in the KAKUSHIN Program: Results of preliminary experiments. *Hydrological Research Letters,* **3**, 49-53
Kjellstrom, E., and P. Lind, 2009: Changes in the water budget in the Baltic Sea drainage basin in future warmer climates as simulated by the regional climate model RCA3. *Boreal Environment Research,* **14(1)**, 114-124
Kjellstrom, E., L. Barring, D. Jacob, R. Jones, G. Lenderink, and C. Schar, 2007: Modelling daily temperature extremes: recent climate and future changes over Europe. *Climatic Change,* **81**, 249-265
Klein Tank, A.M.G., and G.P. Können, 2003: Trends in indices of daily temperature and precipitation extremes in Europe, 1946-1999. *Journal of Climate,* **16**, 3665-3680
Klein Tank, A.M.G., et al., 2002: Daily dataset of 20th century surface air temperature and precipitation series for the European Climate Assessment. *International Journal of Climatology,* **22**, 1441-1453
Klok, E.J., and A.M.G. Klein Tank, 2009: Updated and extended European dataset of daily climate observations. *International Journal of Climatology,* **29(8)**, 1182-1191
Knapp, K.R., and J.P. Kossin, 2007: A new global tropical cyclone data set from ISCCP B1 geostationary satellite observations. *Journal of Applied Remote Sensing,* **1(013505)**
Knapp, K.R., and M.C. Kruk, 2010: Quantifying inter-agency differences in tropical cyclone best track wind speed estimates. *Monthly Weather Review,* **138(4)**, 1459-1473
Knippertz, P., U. Ulbrich, and P. Speth, 2000: Changing cyclones and surface wind speeds over the North Atlantic and Europe in a transient GHG experiment. *Climate Research,* **15(2)**, 109-122
Knox, J.C., 2000: Sensitivity of modern and Holocene floods to climate change. *Quaternary Science Reviews,* **19(1-5)**, 439-457
Knutson, T.R., and R.E. Tuleya, 2004: Impact of CO2-Induced Warming on Simulated Hurricane Intensity and Precipitation: Sensitivity to the Choice of Climate Model and Convective Parameterization. *Journal of Climate,* **17(18)**, 3477-3495
Knutson, T.R., J.J. Sirutis, S.T. Garner, I.M. Held, and R.E. Tuleya, 2007: Simulation of the recent multidecadal increase of Atlantic hurricane activity using an 18-km-grid regional model. *Bulletin of the American Meteorological Society,* **88**, 1549-1565
Knutson, T.R., J.J. Sirutis, S.T. Garner, G.A. Vecchi, and I.M. Held, 2008: Simulated reduction in Atlantic hurricane frequency under twenty-first- century warming conditions. *Nature Geoscience,* **1(6)**, 359-364
Knutson, T.R., T.L. Delworth, K.W. Dixon, I.M. Held, J. Lu, V. Ramaswamy, M.D. Schwarzkopf, G. Stenchikov, and R.J. Stouffer, 2006: Assessment of twentieth-century regional surface temperature trends using the GFDL CM2 coupled models. *Journal of Climate,* **19(9)**, 1624-1651
Knutson, T.R., et al., 2010: Tropical cyclones and climate change. *Nature Geoscience,* **3**, 157-163
Kochel, R.C., and V.R. Baker, 1982: Paleoflood Hydrology. *Science,* **215(4531)**, 353-361

Kodama, C., and T. Iwasaki, 2009: Influence of the SST Rise on Baroclinic Instability Wave Activity under an Aquaplanet Condition. *Journal of the Atmospheric Sciences,* **66(8)**, 2272-2287
Koffi, B., and E. Koffi, 2008: Heat waves across Europe by the end of the 21st century: multiregional climate simulations. *Climate Research,* **36**, 153-168
Komar, P.D., and J.C. Allan, 2008: Increasing hurricane-generated wave heights along the US East Coast and their climate controls. *Journal of Coastal Research,* **24(2)**, 479-488
Kont, A., J. Jaagus, R. Aunap, U. Ratas, and R. Rivis, 2008: Implications of sea-level rise for Estonia. *Journal of Coastal Research,* **24(2)**, 423-431
Koo, G.S., K.O. Boo, and W.T. Kwon, 2009: Projection of temperature over Korea using an MM5 regional climate simulation. *Climate Research,* **40(2-3)**, 241-248
Kossin, J.P., and D.J. Vimont, 2007: A more general framework for understanding Atlantic hurricane variability and trends. *Bulletin of the American Meteorology Society,* **88(11)**, 1767-1781
Kossin, J.P., S.J. Camargo, and M. Sitkowski, 2010: Climate modulation of North Atlantic hurricane tracks. *Journal of Climate,* **submitted**
Kossin, J.P., K.R. Knapp, D.J. Vimont, R.J. Murnane, and B.A. Harper, 2007: A globally consistent reanalysis of hurricane variability and trends. *Geophysical Research Letters,* **34(L04815)**
Koster, R.D., M.J. Suarez, P. Liu, U. Jambor, A. Berg, M. Kistler, R. Reichle, M. Rodell, and J. Famiglietti, 2004a: Realistic initialization of land surface states: Impacts on subseasonal forecast skill. *Journal of Hydrometeorology,* **5(6)**, 1049-1063
Koster, R.D., et al., 2004b: Regions of strong coupling between soil moisture and precipitation. *Science,* **305**, 1138-1140
Kotlyakov, V.M., O.V. Rototaeva, and G.A. Nosenko, 2004: The September 2002 Kolka glacier catastrophe in North Ossetia, Russian Federation: evidence and analysis. *Mountain Research and Development,* **24(1)**, 78-83
Kripalani, R.H., J.H. Oh, and H.S. Chaudhari, 2007a: Response of the East Asian summer monsoon to doubled atmospheric CO2: Coupled climate model simulations and projections under IPCC AR4. *Theoretical and Applied Climatology,* **87(1-4)**, 1-28
Kripalani, R.H., J.H. Oh, A. Kulkarni, S.S. Sabade, and H.S. Chaudhari, 2007b: South Asian summer monsoon precipitation variability: Coupled climate model simulations and projections under IPCC AR4. *Theoretical and Applied Climatology,* **90(3-4)**, 133-159
Krishnamurthy, C.K.B., U. Lall, and H.H. Kwon, 2009: Changing frequency and intensity of rainfall extremes over India from 1951 to 2003. *Journal of Climate,* **22**, 4737-4746
Kruger, A.C., 2006: Observed trends in daily precipitation indices in South Africa: 1910-2004. *International Journal of Climatology,* **26**, 2275-2285
Kubota, H., and J.C.L. Chan, 2009: Interdecadal variability of tropical cyclone landfall in the Philippines from 1902 to 2005. *Geophysical Research Letters,* **36(L12802)**
Kuglitsch, F.G., A. Toreti, E. Xoplaki, P.M. Della-Marta, J. Luterbacher, and H. Wanner, 2009: Homogenization of daily maximum temperature series in the Mediterranean. *Journal of Geophysical Research-Atmospheres,* **114(D15108)**
Kuglitsch, F.G., A. Toreti, E. Xoplaki, P.M. Della-Marta, C. Zerefos, M. Türkes, and J. Luterbacher, 2010: Heat wave changes in the eastern mediterranean since 1960. *Geophysical Research Letters,* **37(L04802)**
Kuleshov, Y., L. Qi, R. Fawcett, and D. Jones, 2008: On tropical cyclone activity in the Southern Hemisphere: Trends and the ENSO connection. *Geophysical Research Letters,* **35(L14S08)**
Kumar, K.R., A.K. Sahai, K.K. Kumar, S.K. Patwardhan, P.K. Mishra, J.V. Revadekar, K. Kamala, and G.B. Pant, 2006: High-resolution climate change scenarios for India for the 21st century. *Current science,* **90(3)**, 334-345
Kundzewicz, Z.W., et al., 2007: Freshwater resources and their management. In: *Impacts, Adaptation and Vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [M.L. Parry, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, USA, pp. 173-210
Kunkel, K.E., et al., 2008: Observed Changes in Weather and Climate Extremes. In: *Weather and Climate Extremes in a Changing Climate. Regions of Focus: North America, Hawaii, Caribbean, and U.S. Pacific Islands.* [T.R. Karl, et al. (eds.)]. A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research, Washington, DC., pp. 222
Kunz, M., J. Sander, and C. Kottmeier, 2009: Recent trends of thunderstorm and hailstorm frequency and their relation to atmospheric characteristics in southwest Germany. *International Journal of Climatology,* **29(15)**, 2283-2297
Kurbis, K., M. Mudelsee, G. Tetzlaff, and R. Brazdil, 2009: Trends in extremes of temperature, dewpoint, and precipitation from long instrumental series from central Europe. *Theoretical and Applied Climatology,* **98**, 187-195
Kusunoki, S., and R. Mizuta, 2008: Future Changes in the Baiu Rain Band Projected by a 20-km Mesh Global Atmospheric Model: Sea Surface Temperature Dependence. *SOLA,* **4**, 85-88

Kwarteng, A.Y., A.S. Dorvlo, and G.T.V. Kumar, 2009: Analysis of a 27-year rainfall data (1977-2003) in the Sultanate of Oman. *International Journal of Climatology,* **29(4)**, 605-617
Kysely, J., 2009: Trends in heavy precipitation in the Czech Republic over 1961-2005. *International Journal of Climatology,* **29**, 1745-1758
Kysely, J., and R. Beranova, 2009: Climate-change effects on extreme precipitation in central Europe: uncertainties of scenarios based on regional climate models. *Theoretical and Applied Climatology,* **95**, 361-374
Lachenbruch, A.H., and B.V. Marshall, 1986: Changing climate: geothermal evidence from permafrost in the Alaskan Arctic. *Science,* **234**, 689-696
Laine, A., M. Kageyama, D. Salas-Melia, G. Ramstein, S. Planton, S. Denvil, and S. Tyteca, 2009: An Energetics Study of Wintertime Northern Hemisphere Storm Tracks under 4 x CO2 Conditions in Two Ocean-Atmosphere Coupled Models. *Journal of Climate,* **22(3)**, 819-839
Landsea, C., G.A. Vecchi, L. Bengtsson, and T.R. Knutson, 2009: Impact of duration thresholds on Atlantic tropical cyclone counts. *Journal of Climate*, doi:10.1175/2009JCLI3034.1
Landsea, C.W., 2007: Counting Atlantic tropical cyclones back to 1900. *Eos Transactions (AGU),* **88(18)**, 197-202
Landsea, C.W., B.A. Harper, K. Hoarau, and J.A. Knaff, 2006: Can we detect trends in extreme tropical cyclones? *Science,* **313**, 452-454
Landsea, C.W., C. Anderson, N. Charles, G. Clark, J. Dunion, J. Fernandez-Partagas, P. Hungerford, C. Neumann, and M. Zimmer, 2004: The Atlantic hurricane database re-analysis project: Documentation for the 1851-1910 alterations and additions to the HURDAT database. In: *Hurricanes and Typhoons: Past, Present and Future* [R.J. Murnane, and K.B. Liu (eds.)]. Columbia University Press, New York, pp. 177-221
Laprise, R., et al., 2008: Challenging some tenets of Regional Climate Modelling. *Meteorology and Atmospheric Physics,* **100(1-4)**, 3-22
Larsen, C.F., R.J. Motyka, A.A. Arendt, K.A. Echelmeyer, and P.E. Geissler, 2007: Glacier changes in southeast Alaska and northwest British Columbia and contribution to sea level rise. *Journal of Geophysical Research,* **112(F01007)**
Lau, K.M., and H.T. Wu, 2007: Detecting trends in tropical rainfall characteristics, 1973-2003. *International Journal of Climatology,* **27**, 979-988
Lau, K.M., Y.P. Zhou, and H.T. Wu, 2008a: Have tropical cyclones been feeding more extreme rainfall? *Journal of Geophysical Research,* **113(D23113)**
Lau, N.C., A. Leetma, and M.J. Nath, 2008b: Interactions between the responses of North American climate to El Niño -La Niña and to the secular warming trend in the Indian-Western Pacific Oceans. *Journal of Climate,* **21(3)**, 476-494
Laurent, C., and S. Parey, 2007: Estimation of 100-year return-period temperatures in France in a non-stationary climate: Results from observations and IPCC scenarios. *Global and Planetary Change,* **57**, 177-188
Le Quesne, C., C. Acuna, J.A. Boninsegna, A. Rivera, and J. Barichivich, 2009: Long-term glacier variations in the Central Andes of Argentina and Chile, inferred from historical records and tree-ring reconstructed precipitation. *Palaeogeography, Palaeoclimatology, Palaeoecology,* **281**, 334-344
Le Quesne, C., D.W. Stahle, M.K. Cleaveland, M.D. Therrell, J.C. Aravena, and J. Barichivich, 2006: Ancient *Austrocedrus* tree-ring chronologies used to reconstruct central Chile precipitation variability from A.D. 1200 to 2000. *Journal of Climate,* **19(22)**, 5731-5744
Le Roy Ladurie, E., 1971: *Times of feast, times of famine: a history of climate since the year 1000*. Doubleday, Garden City, New York, 426 pp.
Leake, J., et al., 2007: *Predicted Wave Climate for the UK: towards an Integrated Model of Coastal Impacts of Climate Change*. In Estuarine and Coastal Modeling Congress 2007. Proceeding of the Tenth International Conference on Estuarine and Coastal Modeling Congress 2007.
Leander, R., T.A. Buishand, B.J.J.M. van den Hurk, and M.J.M. de Wit, 2008: Estimated changes in flood quantiles of the river Meuse from resampling of regional climate model output. *Journal of Hydrology,* **351(3-4)**, 331-343
Legates, D.R., H.F. Lins, and G.J. McCabe, 2005: Comments on "Evidence for global runoff increase related to climate warming" by Labat et al. *Advances in Water Resources,* **28(12)**, 1310-1315
Lehner, B., P. Doll, J. Alcamo, T. Henrichs, and F. Kaspar, 2006: Estimating the impact of global change on flood and drought risks in europe: A continental, integrated analysis. *Climatic Change,* **75(3)**, 273-299
Leloup, J., M. Lengaigne, and J.-P. Boulanger, 2008: Twentieth century ENSO characteristics in the IPCC database. *Climate Dynamics,* **30**, 227-291
Lemke, P., et al., 2007: Observations: Changes in Snow, Ice and Frozen Ground. In: *The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [S. Solomon, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp. 337-383
Lenderink, G., and E. Van Meijgaard, 2008: Increase in hourly precipitation extremes beyond expectations from temperature changes. *Nature Geoscience,* **1(8)**, 511-514

Lenderink, G., E. van Meijgaard, and F. Selten, 2009: Intense coastal rainfall in the Netherlands in response to high sea surface temperatures: analysis of the event of August 2006 from the perspective of a changing climate. *Climate Dynamics,* **32(1)**, 19-33

Leslie, L.M., M. Leplastrier, and B.W. Buckley, 2008: Estimating future trends in severe hailstorms over the Sydney Basin: A climate modelling study. *Atmospheric Research,* **87(1)**, 37-57

Li, H., A. Dai, T. Zhou, and J. Lu, 2008: Responses of East Asian summer monsoon to historical SST and atmospheric forcing during 1950-2000. *Climate Dynamics,* **34(4)**, 501-514

Lim, E.-P., and I. Simmonds, 2009: Effect of tropospheric temperature change on the zonal mean circulation and SH winter extratropical cyclones. *Climate Dynamics,* **33(1)**, 19-32

Lin, J.L., B.E. Mapes, K.M. Weickmann, G.N. Kiladis, S.D. Schubert, M.J. Suarez, J.T. Bacmeister, and M.I. Lee, 2008a: North American monsoon and convectively coupled equatorial waves simulated by IPCC AR4 coupled GCMs. *Journal of Climate,* **21(12)**, 2919-2937

Lin, J.L., K.M. Weickman, G.N. Kiladis, B.E. Mapes, S.D. Schubert, M.J. Suarez, J.T. Bacmeister, and M.I. Lee, 2008b: Subseasonal variability associated with Asian summer monsoon simulated by 14 IPCC AR4 coupled GCMs. *Journal of Climate,* **21(18)**, 4541-4567

Lionello, P., S. Cogo, M.B. Galati, and A. Sanna, 2008: The Mediterranean surface wave climate inferred from future scenario simulations. *Global and Planetary Change,* **63(2-3)**, 152-162

Liu, B., M. Xu, and M. Henderson, 2010: Where have all the showers gone? Regional declines in light precipitation events in China, 1960–2000. *International Journal of Climatology*, doi:10.1002/joc.2144

Llano, M.P., and O. Penalba, 2010: A climatic analysis of dry sequences in Argentina. *International Journal of Climatology*

Llasat, M.C., M. Barriendos, A. Barrera, and T. Rigo, 2005: Floods in Catalonia (NE Spain) since the 14th century. Climatological and meteorological aspects from historical documentary sources and old instrumental records. *Journal of Hydrology,* **313(1-2)**, 32-47

Lloyd-Hughes, B., and M.D. Saunders, 2002: A drought climatology for Europe. *International Journal of Climatology,* **22**, 1571-1592

Loeptien, U., O. Zolina, S. Gulev, M. Latif, and V. Soloviov, 2008: Cyclone life cycle characteristics over the Northern Hemisphere in coupled GCMs. *Climate Dynamics,* **31(5)**, 507-532

Lohmann, U., and J. Feichter, 2007: Global indirect aerosol effects: a review. *Atmospheric Chemistry and Physics,* **5**, 715-737

Long, Z., W. Perrie, J. Gyakum, R. Laprise, and D. Caya, 2009: Scenario changes in the climatology of winter midlatitude cyclone activity over eastern North America and the Northwest Atlantic. *Journal of Geophysical Research-Atmospheres,* **114(D12111)**

Lorenz, D.J., and E.T. DeWeaver, 2007: Tropopause height and zonal wind response to global warming in the IPCC scenario integrations. *Journal of Geophysical Research-Atmospheres,* **112(D10119)**

Lorenz, P., and D. Jacob, 2005: Influence of regional scale information on the global circulation: A two-way nesting climate simulation. *Geophysical Research Letters,* **32(L18706)**

Lorenz, R., E.B. Jaeger, and S.I. Seneviratne, 2010: Persistence of heat waves and its link to soil moisture memory. *Geophysical Research Letters,* **37(L09703)**

Lu, J., G.A. Vecchi, and T. Reichler, 2007: Expansion of the Hadley cell under global warming. *Geophysical Research Letters,* **34(L06805)**

Luterbacher, J., D. Dietrich, E. Xoplaki, M. Grosjean, and H. Wanner, 2004: European seasonal and annual temperature variability, trends, and extremes since 1500. *Science,* **303**, 1499-1503

Lynch, A.H., J.A. Curry, R.D. Brunner, and J.A. Maslanik, 2004: Toward an integrated assessment of the impacts of extreme wind events on Barrow, Alaska. *Bulletin of the American Meteorological Society,* **85(2)**, 209-221

Mabuchi, K., Y. Sato, and H. Kida, 2005a: Climatic Impact of Vegetation Change in the Asian Tropical Region. Part I: Case of the Northern Hemisphere Summer. *Journal of Climate,* **18**, 410-428

Mabuchi, K., Y. Sato, and H. Kida, 2005b: Climatic Impact of Vegetation Change in the Asian Tropical Region. Part II: Case of the Northern Hemisphere Winter and Impact on the Extratropical Circulation. *Journal of Climate,* **18**, 429-446

Makowski, K., E.B. Jaeger, M. Chiacchio, M. Wild, T. Ewen, and A. Ohmura, 2009: On the relationship between diurnal temperature range and surface solar radiation in Europe. *Journal of Geophysical Research,* **114(D00D07)**

Manabe, S., P.C.D. Milly, and R. Wetherald, 2004: Simulated long-term changes in river discharge and soil moisture due to global warming. *Hydrological Sciences Journal-Journal Des Sciences Hydrologiques,* **49(4)**, 625-642

Mann, M.E., T.A. Sabbatelli, and U. Neu, 2007a: Evidence for a modest undercount bias in early historical Atlantic tropical cyclone counts. *Geophysical Research Letters,* **34(L22707)**

Mann, M.E., K.A. Emanual, G.J. Holland, and P.J. Webster, 2007b: Atlantic tropical cyclones revisited. *Eos Transactions (AGU),* **88**, 349-350

Mann, M.E., J.D. Woodruff, J.P. Donnelly, and Z. Zhang, 2009: Atlantic hurricanes and climate over the past 1,500 years. *Nature,* **460**, 880-883

Manson, G.K., and S.M. Solomon, 2007: Past and future forcing of Beaufort sea coastal change. *Atmosphere-Ocean,* **45(2)**, 107-122

Maraun, D., T.J. Osborn, and N.P. Gillet, 2008: United Kingdom daily precipitation intensity: improved early data, error estimates and an update from 2000 to 2006. *International Journal of Climatology,* **28**, 833-842

Maraun, D., et al., 2010: Precipitation downscaling under climate change. Recent developments to bridge the gap between dynamical models and the end user. *Reviews of Geophysics,* **in press**

Marcos, M., M.N. Tsimplis, and A.G.P. Shaw, 2009: Sea level extremes in southern Europe. *Journal of Geophysical Research-Oceans,* **114(C01007)**

Marengo, J.A., 2010: Extreme rainfall and the flood of the Century in Amazonia 2009 - State of the climate in 2009. *Bulletin of the American Meteorological Society,* **in review**, S149

Marengo, J.A., R. Jones, L.M. Alves, and M.C. Valverde, 2009a: Future change of temperature and precipitation extremes in South America as derived from the PRECIS regional climate modeling system. *International Journal of Climatology,* **15**, 2241-2255

Marengo, J.A., M. Rusticucci, O. Penalba, and M. Renom, 2009b: An intercomparison of observed and simulated extreme rainfall and temperature events during the last half of the twentieth century: part 2: historical trends. *Climatic Change*, doi:10.1007/s10584-009-9743-7

Mariotti, A., N. Zeng, J.H. Yoon, V. Artale, A. Navarra, P. Alpert, and L.Z.X. Li, 2008: Mediterranean water cycle changes: transition to drier 21st century conditions in observations and CMIP3 simulations. *Environmental Research Letters,* **3(4)**, Art. 044001 (8 pp.)

Marshall, G.J., A. Orr, N.P.M. van Lipzig, and J.C. King, 2006: The impact of a changing Southern Hemisphere annular mode on Antarctic Peninsula summer temperatures. *Journal of Climate,* **19(20)**, 5388-5404

Martin, J.H., and R.M. Gordon, 1988: Northeast Pacific Iron Distributions in Relation to Phytoplankton Productivity. *Deep-Sea Research Part A -Oceanographic Research Papers,* **35(2)**, 177-196

Matthews, H.D., and A.J. Weaver, 2010: Committed climate warming. *Nature Geoscience,* **3(3)**, 142-143

Matulla, C., W. Schoener, H. Alexandersson, H. von Storch, and X.L. Wang, 2008: European storminess: late nineteenth century to present. *Climate Dynamics,* **31(2-3)**, 125-130

May, W., 2008: Potential future changes in the characteristics of daily precipitation in Europe simulated by the HIRHAM regional climate model. *Climate Dynamics,* **30**, 581-603

McGranahan, G., D. Balk, and B. Anderson, 2007: The rising tide: assessing the risks of climate change and human settlements in low elevation coastal zones. *Environment and Urbanization,* **19(1)**, 17-37

McInnes, K.L., I. Macadam, and J.G. O'Grady, 2009a: *The Effect of Climate Change on Extreme Sea Levels along Victoria's Coast. A Project Undertaken for the Department of Sustainability and Environment, Victoria as part of the 'Future Coasts' Program*. CSIRO, Victoria, Australia www.climatechange.vic.gov.au (15/07/10)

McInnes, K.L., T.A. Erwin, and J.M. Bathols, 2010: Model projected changes in wind due to global warming. *Geophysical Research Letters,* **submitted**

McInnes, K.L., K.J.E. Walsh, G.D. Hubbert, and T. Beer, 2003: Impact of sea-level rise and storm surges on a coastal community. *Natural Hazards,* **30(2)**, 187-207

McInnes, K.L., I. Macadam, G.D. Hubbert, and J.G. O'Grady, 2009b: A modelling approach for estimating the frequency of sea level extremes and the impact of climate change in southeast Australia. *Natural Hazards,* **51(1)**, 115-137

McKee, T.B., N.J. Doesken, and J. Kleist, 1993: The relationship of drought frequency and duration to time scales. . In: *Preprints, 8th Conference on Applied Climatology, Anaheim, California*, pp. 179-184

McKendry, I.G., et al., 2007: Trans-Pacific transport of Saharan dust to western North America: A case study. *Journal of Geophysical Research-Atmospheres,* **112(D01103)**

McVicar, T.R., T.G. Van Niel, L.T. Li, M.L. Roderick, D.P. Rayner, L. Ricciardulli, and R.J. Donohue, 2008: Wind speed climatology and trends for Australia, 1975-2006: Capturing the stilling phenomenon and comparison with near-surface reanalysis output. *Geophysical Research Letters,* **35(L20403)**

Mearns, L.O., W. Gutowski, R. Jones, R. Leung, S. McGinnis, A. Nunes, and Y. Qian, 2009: A Regional Climate Change Assessment Program for North America. *Eos Transactions (AGU),* **90(36)**, 311

Meehl, G.A., and J.M. Arblaster, 2002: The tropospheric biennial oscillation and Asian-Australian monsoon rainfall. *Journal of Climate,* **15(7)**, 722-744

Meehl, G.A., and C. Tebaldi, 2004: More intense, more frequent, and longer lasting heat waves in the 21st century. *Science,* **305**, 994-997

Meehl, G.A., J.M. Arblaster, and C. Tebaldi, 2005: Understanding future patterns of increased precipitation intensity in climate model simulations. *Geophysical Research Letters,* **32(L18719)**

Meehl, G.A., C. Tebaldi, G. Walton, D. Easterling, and L. McDaniel, 2009a: The relative increase of record high maximum temperatures compared to record low minimum temperatures in the U.S. *Geophysical Research Letters,* **36(L23701)**

Meehl, G.A., C. Covey, T. Delworth, M. Latif, B. McAvaney, J.F.B. Mitchell, R.J. Stouffer, and K.E. Taylor, 2007a: The WCRP CMIP3 multimodel dataset - A new era in climate change research. *Bulletin of the American Meteorological Society,* **88(9)**, 1383-1394
Meehl, G.A., et al., 2007b: Global Climate Projections. In: *The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [S. Solomon, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA
Meehl, G.A., et al., 2009b: Decadal Prediction: Can It Be Skillful? *Bulletin of the American Meteorological Society,* **90(10)**, 1467-1485
Menendez, C.G., et al., 2010: Downscaling extreme month-long anomalies in southern South America. *Climatic Change,* **98(3-4)**, 379-403
Menendez, M., and P.L. Woodworth, 2010: Changes in extreme high water levels based on a quasi-global tide-gauge dataset. *Journal of Geophysical Research,* **in press**
Menendez, M., F.J. Mendez, I.J. Losada, and N.E. Graham, 2008: Variability of extreme wave heights in the northeast Pacific Ocean based on buoy measurements. *Geophysical Research Letters,* **35(L22607)**
Menne, M.J., and C.N. Williams Jr., 2009: Homogenization of temperature series via pairwise comparisons. *Journal of Climate,* **22(7)**, 1700-1717
Miller, R., G. Schmidt, and D. Shindell, 2006: Forced annular variations in the 20th century intergovernmental panel on climate change fourth assessment report models. *Journal of Geophysical Research,* **111(D18101)**
Mills, G.A., 2005: A re-examination of the synoptic and mesoscale meteorology of Ash Wednesday 1983. *Australian Meteorological Magazine,* **54(1)**, 35-55
Milly, P.C.D., R.T. Wetherald, K.A. Dunne, and T.L. Delworth, 2002: Increasing risk of great floods in a changing climate. *Nature,* **415(6871)**, 514-517
Milly, P.C.D., J. Betancourt, M. Falkenmark, R.M. Hirsch, Z.W. Kundzewicz, D.P. Lettenmaier, and R.J. Stouffer, 2008: Stationarity is dead: Whither water management? *Science,* **319(5863)**, 573-574
Min, S.-K., X. Zhang, and F. Zwiers, 2008a: Human-induced arctic moistening. *Science,* **320(5875)**, 518-520
Min, S.-K., X. Zhang, F.W. Zwiers, and T. Agnew, 2008b: Human influence on Arctic sea ice detectable from early 1990s onwards. *Geophysical Research Letters,* **35(L21701)**
Min, S.K., and A. Hense, 2007: A Bayesian assessment of climate change using multimodel ensembles. Part II: Regional and seasonal mean surface temperatures. *Journal of Climate,* **20(12)**, 2769-2790
Min, S.K., X. Zhang, F.W. Zwiers, P. Friederichs, and A. Hense, 2009: Signal detectability in extreme precipitation changes assessed from twentieth century climate simulations. *Climate Dynamics,* **32(1)**, 95-111
Mitchell, J.F.B., D.J. Karoly, G.C. Hegerl, F.W. Zwiers, M.R. Allen, and J. Marengo, 2001: Detection of climate change and attribution of causes. In: *Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change* [J.T. Houghton, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp. 695-738
Moberg, A., et al., 2006: Indices for daily temperature and precipitation extremes in Europe analyzed for the period 1901-2000. *Journal of Geophysical Research-Atmospheres,* **111(D22106)**
Mori, N., R. Iwashima, T. Yasuda, H. Mase, T. Tom, and Y. Oku, 2009: *Impact of global climate change on wave climate*. In Proceedings of Coastal Dynamics 2009.
Mudelsee, M., M. Borngen, G. Tetzlaff, and U. Grunewald, 2003: No upward trends in the occurrence of extreme floods in central Europe. *Nature,* **425(6954)**, 166-169
Müller, W.A., and E. Roeckner, 2008: ENSO teleconnections in projections of future climate in ECHAM5/MPI-OM. *Climate Dynamics,* **31(5)**, 533-549
Murphy, J.M., B.B.B. Booth, M. Collins, G.R. Harris, D.M.H. Sexton, and M.J. Webb, 2007: A methodology for probabilistic predictions of regional climate change from perturbed physics ensembles. *Philosophical Transactions of the Royal Society A-Mathematical Physical and Engineering Sciences,* **365(1857)**, 1993-2028
Narama, C., Y. Shimamura, D. Nakayama, and K. Abdrakhmatov, 2006: Recent changes of glacier coverage in the western Terskey-Alatoo range, Kyrgyz Republic, using Corona and Landsat. *Annals of Glaciology,* **43(1)**, 223-229
Nelson, F.E., O.A. Anisimov, and N.I. Shiklomanov, 2001: Subsidence risk from thawing. *Nature,* **410**, 889-890
Neu, U., 2008: Is recent major hurricane activity normal? *Nature,* **451(7181)**, E5
New, M., et al., 2006: Evidence of trends in daily climate extremes over southern and west Africa. *Journal of Geophysical Research,* **111(D14102)**
Niall, S., and K. Walsh, 2005: The impact of climate change on hailstorms in southeastern Australia. *International Journal of Climatology,* **25**, 1933-1952
Nicholls, N., 1995: Long-term climate monitoring and extreme events. *Climatic Change,* **31**, 231-245
Nicholls, N., 2008: Recent trends in the seasonal and temporal behaviour of the El Niño Southern Oscillation. *Geophysical Research Letters,* **35(L19703)**
Nicholls, N., 2009: Local and remote causes of the southern Australian autumn-winter rainfall decline, 1958-2007. *Climate Dynamics*, doi:10.1007/s00382-009-0527-6.

Nicholls, N., and L. Alexander, 2007: Has the climate become more variable or extreme? Progress 1992-2006. *Progress in Physical Geography,* **31**, 77-87
Nicholls, N., R. Tapp, K. Burrows, and D. Richards, 1996: Historical thermometer exposures in Australia. *International Journal of Climatology,* **16**, 705-710
Nicholls, R.J., and A.C. de la Vega-Leinert, 2008: Implications of sea-level rise for Europe's coasts: An introduction. *Journal of Coastal Research,* **24(2)**, 285-287
Nicholls, R.J., P.P. Wong, V.R. Burkett, J.O. Codignotto, J.E. Hay, R.F. McLean, S. Ragoonaden, and C.D. Woodroffe, 2007: Coastal systems and low-lying areas. In: *Impacts, Adaptation and Vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [M.L. Parry, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp. 315-356
Nieto-Ferreira, R., and T. Rickenbach, 2010: Regionality of monsoon onset in South America: a threestage conceptual model. *International Journal of Climatology*, doi:10.1002/joc.2161
Ning, L., and Y. Qian, 2009: Interdecadal Change in Extreme Precipitation over South China and Its Mechanism. *Advances in Atmospheric Sciences,* **26(1)**, 109-118
Niu, F., G. Cheng, W. Ni, and D. Jin, 2005: Engineering-related slope failure in permafrost regions of the Qinghai-Tibet Plateau. *Cold Regions Science and Technology,* **42(3)**, 215-225
Noetzli, J., and S. Gruber, 2009: Transient thermal effects in Alpine permafrost. *The Cryosphere,* **3**, 85-99
Noetzli, J., S. Gruber, T. Kohl, N. Salzmann, and W. Haeberli, 2007: Three-dimensional distribution and evolution of permafrost temperatures in idealized high-mountain topography. *Journal of Geophysical Research,* **112(F02S13)**
Nolan, D.S., E.D. Rappin, and K.A. Emanuel, 2007: Tropical cyclogenesis sensitivity to environmental parameters in radiative-convective equilibrium. *Quarterly Journal of the Royal Meteorological Society,* **133(629)**, 2085-2107
Nott, J., J. Haig, H. Neil, and D. Gillieson, 2007: Greater frequency variability of landfalling tropical cyclones at centennial compared to seasonal and decadal scales. *Earth and Planetary Science Letters,* **255**, 367-372
Nunez, M.N., S.A. Solman, and M.F. Cabre, 2009: Regional climate change experiments over southern South America. II: Climate change scenarios in the late twenty-first century. *Climate Dynamics,* **32(7-8)**, 1081-1095
Nyberg, J., B.A. Malmgren, A. Winter, M.R. Jury, K. Halimeda Kilbourne, and T.M. Quinn, 2007: Low Atlantic hurricane activity in the 1970s and 1980s compared to the past 270 years. *Nature,* **447(7145)**, 698-701
O'Gorman, P.A., and T. Schneider, 2008: Energy of Midlatitude Transient Eddies in Idealized Simulations of Changed Climates. *Journal of Climate,* **21(22)**, 5797-5806
O'Gorman, P.A., and T. Schneider, 2009a: The physical basis for increases in precipitation extremes in simulations of 21st-century climate change. *Proceedings of the National Academy of Sciences of the United States of America,* **106(35)**, 14773-14777
O'Gorman, P.A., and T. Schneider, 2009b: Scaling of Precipitation Extremes over a Wide Range of Climates Simulated with an Idealized GCM. *Journal of Climate,* **22(21)**, 5676-5685
Obserman, N.G., and G.G. Mazhitova, 2001: Permafrost dynamics in the northeast of European Russia at the end of 20th century. *Norwegian Journal of Geography,* **55**, 241-244
Oerlemans, J., 2005: Extracting a climate signal from 169 glacier records. *Science,* **308(5722)**, 675-677
Oki, T., T. Nishimura, and P. Dirmeyer, 1999: Assessment of annual runoff from land surface models using Total Runoff Integrating Pathways (TRIP). *Journal of the Meteorological Society of Japan,* **77**, 235-255
Oouchi, K., J. Yoshimura, H. Yoshimura, R. Mizuta, S. Kusunoki, and A. Noda, 2006: Tropical Cyclone Climatology in a Global-Warming Climate as Simulated in a 20 km-Mesh Global Atmospheric Model: Frequency and Wind Intensity Analyses. *Journal of the Meteorological Society of Japan,* **84(2)**, 259-276
Oppikofer, T., M. Jaboyedoff, and H.R. Keusen, 2008: Collapse at the eastern Eiger flank in the Swiss Alps. *Nature Geoscience,* **1(8)**, 531-535
Orsolini, Y., and A. Sorteberg, 2009: Projected Changes in Eurasian and Arctic Summer Cyclones under Global Warming in the Bergen Climate Model. *Atmospheric and Oceanic Science Letters,* **2(1)**, 62
Osborn, T.J., and M. Hulme, 1997: Development of a relationship between station and grid-box rainday frequencies for climate model evaluation. *Journal of Climate,* **10(8)**, 1885-1908
Osterkamp, T.E., 2005: The recent warming of permafrost in Alaska. *Global and Planetary Change,* **49**, 187-202
Osterkamp, T.E., 2007: Characteristics of the recent warming of permafrost in Alaska. *Journal of Geophysical Research,* **112(F02S02)**
Osterkamp, T.E., M.T. Jorgenson, E.A.G. Schuur, Y.L. Shur, M.Z. Kanevskiy, J.G. Vogel, and V.E. Tumskoy, 2009: Physical and ecological changes associated with warming permafrost and thermokarst in Interior Alaska. *Permafrost and Periglacial Processes,* **20**, 235-256
Osti, R., and S. Egashira, 2009: Hydrodynamic characteristics of the Tam Pokhari glacial lake outburst flood in the Mt. Everest region, Nepal. *Hydrological Processes,* **23(20)**, 2943-2955

Otto-Bliesner, B.L., et al., 2009: A comparison of PMIP2 model simulations and the MARGO proxy reconstruction for tropical sea surface temperatures at last glacial maximum. *Climate Dynamics,* **32(6)**, 799-815
Paeth, H., and H.-P. Thamm, 2007: Regional modelling of future African climate north of 15 degrees S including greenhouse warming and land degradation. *Climatic Change,* **83(3)**, 401-427
Pall, P., M.R. Allen, and D.A. Stone, 2007: Testing the Clausius-Clapeyron constraint on changes in extreme precipitation under CO2 warming. *Climate Dynamics,* **28**, 351-363
Pall, P., T. Aina, D.A. Stone, P.A. Stott, T. Nozawa, A.G.J. Hilbberts, D. Lohmann, and M.R. Allen, 2010: Anthropogenic greenhouse gas contribution to UK autumn flood risk. *Nature,* **submitted**
Palmer, T.N., and J. Räisänen, 2002: Quantifying the risk of extreme seasonal precipitation events in a changing climate. *Nature,* **415(6871)**, 514-517
Palmer, W.C., 1965: *Meteorological drought*. Report 45, US Weather Bureau, Washington DC
Pattanaik, D.R., and M. Rajeevan, 2009: Variability of extreme rainfall events over India during southwest monsoon season. *Meteorological Applications*, doi:10.1002/met.164
Paul, F., and W. Haeberli, 2008: Spatial variability of glacier elevation changes in the Swiss Alps obtained from two digital elevation models. *Geophysical Research Letters,* **35(L21502)**
Paul, F., A. Kääb, M. Maisch, T. Kellenberger, and W. Haeberli, 2004: Rapid disintegration of Alpine glaciers observed with satellite data. *Geophysical Research Letters,* **31(L21402)**, 33-44
Pavan, V., R. Tomozeiu, C. Cacciamani, and M. Di Lorenzo, 2008: Daily precipitation observations over Emilia-Romagna: mean values and extremes. *International Journal of Climatology,* **28(15)**, 2065-2079
Peel, M.C., and T.A. McMahon, 2006: Continental runoff: A quality-controlled global runoff data set. *Nature,* **444(E14)**
Peixoto, J.P., and A.H. Oort, 1992: *Physics of climate*. American Institute of Physics, Springer-Verlag, New York, 520 pp.
Penalba, O.C., and F.A. Robeldo, 2010: Spatial and temporal variability of the frequency of extreme daily rainfall regime in the La Plata Basin during the 20th century. *Climate Change,* **98(3-4)**, 531-550
Perkins, S.E., A.J. Pitman, and S.A. Sisson, 2009: Smaller projected increases in 20-year temperature returns over Australia in skill-selected climate models. *Geophysical Research Letters,* **36(L06710)**
Peterson, T.C., and M.J. Manton, 2008: Monitoring changes in climate extremes - A tale of international collaboration. *Bulletin of the American Meteorological Society,* **89(9)**, 1266-1271
Peterson, T.C., X. Zhang, M. Brunet-India, and J.L. Vazquez-Aguirre, 2008a: Changes in North American extremes derived from daily weather data. *Journal of Geophysical Research,* **113(DO7113)**
Peterson, T.C., D. Anderson, S.J. Cohen, M. Cortez, R. Murname, C. Parmesan, D. Phillips, R. Pulwarty, and J. Stone, 2008b: Why weather and climate extremes matter. In: *Weather and Climate Extremes in a Changing Climate. Regions of Focus: North America, Hawaii, Caribbean, and U.S. Pacific Islands*. [T.R. Karl, et al. (eds.)]. A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research, Washington, DC., pp. 11-33
Petrow, T., and B. Merz, 2009: Trends in flood magnitude, frequency and seasonality in Germany in the period 1951-2002. *Journal of Hydrology,* **371(1-4)**, 129-141
Pezza, A.B., and I. Simmonds, 2008: Large-scale Factors in Tropical and Extratropical Cyclone Transition and Extreme Weather Events. *Trends and Directions in Climate Research: Annals of the New York Academy of Sciences,* **1146**, 189-211
Pezza, A.B., I. Simmonds, and J.A. Renwick, 2007: Southern Hemisphere cyclones and anticyclones: Recent trends and links with decadal variability in the Pacific Ocean. *International Journal of Climatology,* **27(11)**, 1403-1419
Philander, S.G., 1990: *El Niño, La Niña and the Southern Oscillation*. Academic Press, San Diego, California, 293 pp.
Piao, S., P. Friedlingstein, P. Ciais, N. de Noblet-Ducoudré, D. Labat, and S. Zaehle, 2007: Changes in climate and land use have a larger direct impact than rising CO2 on global river runoff trends. *Proceedings of the National Academy of Sciences,* **104**, 15.242-15.247
Pierson, T.C., R.J. Janda, J.C. Thouret, and C.A. Borrero, 1990: Perturbation and melting of snow and ice by the 13 November 1985 eruption of Nevado del Ruiz, Colombia, and consequent mobilization, flow and deposition of lahars. *Journal of Volcanology and Geothermal Research,* **41(1-4)**, 17-66
Pinto, J.G., S. Zacharias, A.H. Fink, G.C. Leckebusch, and U. Ulbrich, 2009: Factors contributing to the development of extreme North Atlantic cyclones and their relationship with the NAO. *Climate Dynamics,* **32(5)**, 711-737
Pinto, J.G., U. Ulbrich, G.C. Leckebusch, T. Spangehl, M. Reyers, and S. Zacharias, 2007: Changes in storm track and cyclone activity in three SRES ensemble experiments with the ECHAM5/MPI-OM1 GCM. *Climate Dynamics,* **29(2-3)**, 195-210
Pirazzoli, P.A., and A. Tomasin, 2003: Recent near-surface wind changes in the central Mediterranean and Adriatic areas. *International Journal of Climatology,* **23(8)**, 963-973
Pitman, A.J., et al., 2009: Uncertainties in climate responses to past land cover change: first results from the LUCID intercomparison study. *Geophysical Research Letters,* **36(L14814)**

Planton, S., M. Deque, F. Chauvin, and L. Terray, 2008: Expected impacts of climate change on extreme climate events. *Comptes Rendus Geoscience,* **340(9-10)**, 564-574
Portmann, R.W., S. Solomon, and G.C. Hegerl, 2009: Linkages between climate change, extreme temperature and precipitation across the United States. *Proceedings of the National Academy of Sciences,* **106**, 7324-7329
Powers, S.B., and I.N. Smith, 2007: Weakening of the Walker Circulation and apparent dominance of El Niño both reach record levels, but has ENSO really changed? *Geophysical Research Letters,* **34(L18702)**
Prager, C., S. Ivy-Ochs, M. Ostermann, H. Synal, and G. Patzelt, 2009: Geology and radiometric 14C-, 36Cl-and Th-/U-dating of the Fernpass rockslide (Tyrol, Austria). *Geomorphology,* **103(1)**, 93-103
Prospero, J.M., and P.J. Lamb, 2003: African droughts and dust transport to the Caribbean: Climate change implications. *Science,* **302(5647)**, 1024-1027
Prospero, J.M., E. Blades, R. Naidu, and M.C. Lavoie, 2009: Reply to: African dust and asthma in the Caribbean-medical and statistical perspectives by M A Monteil and R Antoine. *International Journal of Biometeorology,* **53(5)**, 383-385
Prudhomme, C., and H. Davies, 2009: Assessing uncertainties in climate change impact analyses on the river flow regimes in the UK. Part 2: future climate. *Climatic Change,* **93(1-2)**, 197-222
Pruszak, Z., and E. Zawadzka, 2008: Potential implications of sea-level rise for Poland. *Journal of Coastal Research,* **24(2)**, 410-422
Pryor, S.C., R.J. Barthelmie, and E.S. Riley, 2007: Historical evolution of wind climates in the USA. *Journal of Physics: Conference Series,* **75(012065)**
Pryor, S.C., J.A. Howe, and K.E. Kunkel, 2009: How spatially coherent and statistically robust are temporal changes in extreme precipitation in the contiguous USA? *International Journal of Climatology,* **29(1)**, 31-45
Purvis, M.J., P.D. Bates, and C.M. Hayes, 2008: A probabilistic methodology to estimate future coastal flood risk due to sea level rise. *Coastal Engineering,* **55(12)**, 1062-1073
Qian, B., S. Gameda, and H. Hayhoe, 2008: Performance of stochastic weather generators LARS-WG and AAFC-WG for reproducing daily extremes of diverse Canadian climates. *Climate Research,* **37(1)**, 17-33
Qian, W.H., L.S. Quan, and S.Y. Shi, 2002: Variations of the dust storm in China and its climatic control. *Journal of Climate,* **15(10)**, 1216-1229
Rachold, V., M. Lack, and M.N. Grigoriev, 2003: A Geo Information System (GIS) for Circum-Arctic Coastal Dynamics. In: *Proceedings of the 8th International Conference on Permafrost, 21-25, July 2003, Zurich, Switzerland* [M. Philips, et al. (eds.)]. Balkema, Netherlands, pp. 923-927
Raff, D.A., T. Pruitt, and L.D. Brekke, 2009: A framework for assessing flood frequency based on climate projection information. *Hydrology and Earth System Sciences,* **13(11)**, 2119-2136
Rahimzadeh, F., A. Asgari, and E. Fattahi, 2009: Variability of extreme temperature and precipitation in Iran during recent decades. *International Journal of Climatology,* **29(3)**, 329-343
Raia, A., and I.F.D. Cavalcanti, 2008: The Life Cycle of the South American Monsoon System. *Journal of Climate,* **21(23)**, 6227-6246
Raible, C.C., 2007: On the relation between extremes of midlatitude cyclones and the atmospheric circulation using ERA40. *Geophysical Research Letters,* **34(L07703)**
Raible, C.C., P.M. Della-Marta, C. Schwierz, H. Wernli, and R. Blender, 2008: Northern hemisphere extratropical cyclones: A comparison of detection and tracking methods and different reanalyses. *Monthly Weather Review,* **136(3)**, 880-897
Räisänen, J., 2007: How reliable are climate models? *Tellus,* **59A**, 2-29
Rajeevan, M., J. Bhate, and A.K. Jaswal, 2008: Analysis of variability and trends of extreme rainfall events over India using 104 years of gridded daily rainfall data. *Geophysical Research Letters,* **35(L18707)**
Ranasinghe, R., R. McLoughlin, A. Short, and G. Symonds, 2004: The Southern Oscillation Index, wave climate, and beach rotation. *Marine Geology,* **204(3-4)**, 273-287
Randall, D.A., et al., 2007: Climate Models and their Evaluation. In: *The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [S. Solomon, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA
Re, M., and V. Ricardo Barros, 2009: Extreme rainfalls in SE South America. *Climatic Change,* **96(1-2)**, 119-136
Rebetez, M., R. Lugon, and P. Baeriswyl, 1997: Climatic change and debris flows in high mountain regions: The case study of the Ritigraben torrent (Swiss Alps). *Climatic Change,* **36(3)**, 371-389
Rebetez, M., O. Dupont, and M. Giroud, 2008: An analysis of the July 2006 heatwave extent in Europe compared to the record year of 2003. *Theoretical and Applied Climatology,* **95(1-2)**, 1-7
Redmond, K.T., Y. Enzel, P.K. House, and F. Biondi, 2002: Climate impact on flood frequency at the decadal to millennial time scales. In: *Ancient Floods, Modern Hazards: Principles and Applications of Paleoflood Hydrology* [P.K. House, et al. (eds.)]. Vol. 5. American Geophysical Union Water Science and Applications, pp. 21-46
Regonda, S.K., B. Rajagopalan, M. Clark, and J. Pitlick, 2005: Seasonal cycle shifts in hydroclimatology over the western United States. *Journal of Climate,* **18(2)**, 372-384

Rehman, S., 2010: Temperature and rainfall variation over Dhahran, Saudi Arabia, (1970-2006). *International Journal of Climatology,* **30(3)**, 445-449

Reichstein, M., et al., 2007: Reduction of ecosystem productivity and respiration during the European summer 2003 climate anomaly: a joint flux tower, remote sensing and modeling analysis. *Global Change Biology,* **13(3)**, 634-651

Rein, B., A. Lückge, L. Reinhardt, F. Sirocko, A. Wolf, and W.C. Dullo, 2005: El Niño variability off Peru during the last 20,000 years. *Paleoceanography,* **20(PA4003)**

Renard, B., et al., 2008: Regional methods for trend detection: Assessing field significance and regional consistency. *Water Resources Research,* **44(W08419)**

Reynolds, J.M., A. Dolecki, and C. Portocarrero, 1998: The construction of a drainage tunnel as part of glacial lake hazard mitigation at Hualcan, Cordillera Blanca, Peru. *Engineering Geology Special Publications,* **15(1)**, 41-48

Richardson, S.D., and J.M. Reynolds, 2000: An overview of glacial hazards in the Himalayas. *Quaternary International,* **65**, 31-47

Rickenmann, D., and M. Zimmermann, 1993: The 1987 debris flows in Switzerland: documentation and analysis. *Geomorphology (Amsterdam),* **8(2-3)**, 175-189

Risbey, J.S., and M. Kandlikar, 2007: Expressions of likelihood and confidence in the IPCC uncertainty assessment process. *Climatic Change,* **85**, 19-31

Rist, A., and M. Phillips, 2005: First results of investigations on hydrothermal processes within the active layer above alpine permafrost in steep terrain. *Norsk Geografisk Tidsskrift-Norwegian Journal of Geography,* **59(2)**, 177-183

Robeson, S., 2004: Trends in time-varying percentiles of daily minimum and maximum temperature over North America. *Geophysical Research Letters,* **31(L04203)**

Robock, A., K.Y. Vinnikov, G. Srinivasan, J.K. Entin, S.E. Hollinger, N.A. Speranskaya, S. Liu, and A. Namkhai, 2000: The global soil moisture data bank. *Bulletin of the American Meteorological Society,* **81(6)**, 1281-1299

Robson, A.J., T.K. Jones, D.W. Reed, and A.C. Bayliss, 1998: A study of national trend and variation in UK floods. *International Journal of Climatology,* **18(2)**, 165-182

Rocha, A., P. Melo-Goncalves, C. Marques, J. Ferreira, and J.M. Castanheira, 2008: High-frequency precipitation changes in southeastern Africa due to anthropogenic forcing. *International Journal of Climatology,* **28(9)**, 1239-1253

Rockel, B., and K. Woth, 2007: Extremes of near-surface wind speed over Europe and their future changes as estimated from an ensemble of RCM simulations *Climatic Change,* **81**, 267-280

Rodda, J.C., M.A. Little, H.J.E. Rodda, and P.E. McSharry, 2009: A Comparative Study of the Magnitude, Frequency and Distribution of Intense Rainfall in the United Kingdom. *International Journal of Climatology*, doi:10.1002/joc.2024

Rodier, J.A., and M. Roche, 1984: *World Catalogue of Maximum observed floods. IAHS Pub. No. 143*. IAHS Press, Wallingford, UK

Roe, G.H., and M.B. Baker, 2007: Why is climate sensitivity so unpredictable? *Science,* **318(5850)**, 629-632

Roer, I., W. Haeberli, and M. Avian, 2008: Observations and considerations on destabilizing active rock glaciers in the European Alps. In: *Ninth International Conference on Permafrost* [D.L. Kane, and K.M. Hinkel (eds.)]. Institute of Northern Engineering, University of Alaska, Fairbanks

Roesch, A., 2006: Evaluation of surface albedo and snow cover in AR4 coupled climate models. *Journal of Geophysical Research,* **111(D15111)**

Ropelewski, C.F., and M.A. Bell, 2008: Shifts in the statistics of daily rainfall in South America conditional on ENSO phase. *Journal of Climate,* **21(5)**, 849-865

Rosenbluth, B., H.A. Fuenzalida, and P. Aceituno, 1997: Recent temperature variations in southern South America. *International Journal of Climatology,* **17(1)**, 67-85

Rosenzweig, C., et al., 2007: Assessment of Observed Changes and Responses in Natural and Managed Systems. In: *Impacts, Adaptation and Vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [M.L. Parry, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, USA, pp. 79-131

Rotstayn, L.D., et al., 2009: Possible impacts of anthropogenic and natural aerosols on Australian climate: A review. *International Journal of Climatology,* **29**, 461-479

Rott, H., et al., 2010: Cold Regions Hydrology High-Resolution Observatory for Snow and Cold Land Processes. *IEEE Proceedings,* **98(5)**, 752-765

Rusticucci, M., and M. Barrucand, 2004: Observed trends and changes in temperature extremes over Argentina. *Journal of Climate,* **17**, 4099-4107

Rusticucci, M., and M. Renom, 2008: Variability and trends in indices of quality-controlled daily temperature extremes in Uruguay. *International Journal of Climatology,* **28**, 1083-1095

Rusticucci, M., J. Marengo, O. Penalba, and M. Renom, 2009: An intercomparison of model-simulated in extreme rainfall and temperature events during the last half of the twentieth century. Part 1: mean values and variability. *Climate Change*, doi:10.1007/s10584-009-9742-8

Sailor, D.J., M. Smith, and M. Hart, 2008: Climate change implications for wind power resources in the Northwest United States. *Renewable Energy,* **33(11)**, 2393-2406

Santer, B.D., T.M.L. Wigley, T. Barnett, and E. Anyamba, 1996: Detection of climate change and attribution of causes. In: *Contribution of Working Group I to the Second Assessment Report of the Intergovernmental Panel on Climate Change* [J.T. Houghton, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp. 407-444

Santer, B.D., et al., 2007: Identification of human-induced changes in atmospheric moisture content. *Proceedings of the National Academy of Sciences of the United States of America,* **104(39)**, 15248-15253

Santer, B.D., et al., 2006: Forced and unforced ocean temperature changes in Atlantic and Pacific tropical cyclogenesis regions. *Proceedings of the National Academy of Sciences,* **103(38)**, 13905-13910

Santos, C.A., and J.I.B. Brito, 2007: Análise dos índices de extremos para o semi-árido do Brasil e suas relações com TSM e IVDN. *Revista Brasileira de Meteorologia,* **22(3)**, 303-312

Santos, C.A.C., J. Brito, I, B., T.V.R. Rao, and E.A. Meneses, 2009: Tendências dos Índices de precipitação no Estado do Ceará. *Revista Brasileira de Meteorologia,* **24(1)**, 39-47

Sasaki, W., and H. Toshiyuki, 2007: Interannual variability and predictability of summertime significant wave heights in the western North Pacific. *Journal of Oceanography,* **63(2)**, 203-213

Scaife, A.A., C.K. Folland, L.V. Alexander, A. Moberg, and J.R. Knight, 2008: European climate extremes and the North Atlantic Oscillation. *Journal of Climate,* **21**, 72-83

Scaife, A.A., et al., 2009: The CLIVAR C20C project: selected twentieth century climate events. *Climate Dynamics,* **33(5)**, 603-614

Schaer, C., and G. Jendritzky, 2004: Climate change: Hot news from summer 2003. *Nature,* **432(7017)**, 559-560

Schär, C., P.L. Vidale, D. Lüthi, C. Frei, C. Häberli, M.A. Liniger, and C. Appenzeller, 2004: The role of increasing temperature variability in European summer heatwaves. *Nature,* **427(322)**, 332-336

Schiefer, E., B. Menounos, and R. Wheate, 2007: Recent volume loss of British Columbian glaciers, Canada. *Geophysical Research Letters,* **34(L16503)**

Schmidli, J., and C. Frei, 2005: Trends of heavy precipitation and wet and dry spells in Switzerland during the 20th century. *International Journal of Climatology,* **25**, 753-771

Schmidli, J., C.M. Goodess, C. Frei, M.R. Haylock, Y. Hundecha, J. Ribalaygua, and T. Schmith, 2007: Statistical and dynamical downscaling of precipitation: An evaluation and comparison of scenarios for the European Alps. *Journal of Geophysical Research-Atmospheres,* **112(D04105)**

Schneidereit, A., R. Blender, K. Fraedrich, and F. Lunkeit, 2007: Icelandic climate and north Atlantic cyclones in ERA-40 reanalyses. *Meteorologische Zeitschrift,* **16(1)**, 17-23

Schubert, S., et al., 2009: A US CLIVAR Project to Assess and Compare the Responses of Global Climate Models to Drought-Related SST Forcing Patterns: Overview and Results. *Journal of Climate,* **22(19)**, 5251-5272

Schubert, S.D., Y. Chang, M.J. Suarez, and P.J. Pegion, 2008a: ENSO and wintertime extreme precipitation events over the contiguous united states. *Journal of Climate,* **21(1)**, 22-39

Schubert, S.D., M.J. Suarez, P.J. Pegion, R.D. Koster, and J.T. Bacmeister, 2004: On the cause of the 1930s Dust Bowl. *Science,* **303(5665)**, 1855-1859

Schubert, S.D., M.J. Suarez, P.J. Pegion, R.D. Koster, and J.T. Bacmeister, 2008b: Potential predictability of long-term drought and pluvial conditions in the US Great Plains. *Journal of Climate,* **21(4)**, 802-816

Scileppi, E., and J.P. Donnelly, 2007: Sedimentary evidence of hurricane strikes in western Long Island, New York. *Geochemistry Geophysics Geosystems,* **8**, 1-25

Seager, R., A. Tzanova, and J. Nakamura, 2009: Drought in the southeastern United States: Causes, variability over the last millennium, and the potential for future hydroclimate change. *Journal of Climate,* **22(19)**, 5021-5045

Seager, R., et al., 2007: Model projections of an imminent transition to a more arid climate in southwestern North America. *Science,* **316(5828)**, 1181-1184

Seierstad, I.A., D.B. Stephenson, and N.G. Kvamsto, 2007: How useful are teleconnection patterns for explaining variability in extratropical storminess ? *Tellus Series A-Dynamic Meteorology and Oceanography,* **59(2)**, 170-181

Seleshi, Y., and P. Camberlin, 2006: Recent changes in dry spell and extreme rainfall events in Ethiopia. *Theoretical and Applied Climatology,* **83(1-4)**, 181-191

Semenov, M.A., 2008: Simulation of extreme weather events by a stochastic weather generator. *Climate Research,* **35(3)**, 203-212

Semmler, T., S. Varghese, R. McGrath, P. Nolan, S. Wang, P. Lynch, and C. O'Dowd, 2008: Regional climate model simulations of North Atlantic cyclones: frequency and intensity changes. *Climate Research,* **36(1)**, 1-16

Sen Roy, S., 2009: A spatial analysis of extreme hourly precipitation patterns in India. *International Journal of Climatology,* **29**, 345-355

Seneviratne, S.I., D. Lüthi, M. Litschi, and C. Schär, 2006a: Land-atmosphere coupling and climate change in Europe. *Nature,* **443(7108)**, 205-209
Seneviratne, S.I., T. Corti, E.L. Davin, M. Hirschi, E. Jaeger, I. Lehner, B. Orlowsky, and A.J. Teuling, 2010: Investigating soil moisture-climate interactions in a changing climate: A review. *Earth Science Reviews,* **99(3-4)**, 125-161
Seneviratne, S.I., et al., 2006b: Soil moisture memory in AGCM simulations: Analysis of global land-atmosphere coupling experiment (GLACE) data. *Journal of Hydrometeorology,* **7(5)**, 1090-1112
Sevruk, B., 1996: Adjustment of tipping-bucket precipitation gauge measurements. *Atmospheric Research,* **42**, 237-246
Shanahan, T.M., J.T. Overpeck, K.J. Anchukaitis, J.W. Beck, J.E. Cole, D.L. Dettman, J.A. Peck, C.A. Scholz, and J.W. King, 2009: Atlantic forcing of persistent drought in West Africa. *Science,* **324(5925)**, 377-380
Shao, Y., and C.H. Dong, 2006: A review on East Asian dust storm climate, modelling and monitoring. *Global and Planetary Change,* **52(1-4)**, 1-22
Sharkhuu, N., 2003: Recent changes in permafrost in Mongolia. In: *Proceedings of the 8th International Conference on Permafrost, 21-25, July 2003, Zurich, Switzerland* [M. Philips, et al. (eds.)]. Balkema, Netherlands, pp. 1029-1034
Sheffield, J., and E.F. Wood, 2008: Global trends and variability in soil moisture and drought characteristics, 1950-2000, from observation-driven simulations of the terrestrial hydrologic cycle. *Journal of Climate,* **21(3)**, 432-458
Sheppard, C., D.J. Dixon, M. Gourlay, A. Sheppard, and R. Payet, 2005: Coral mortality increases wave energy reaching shores protected by reef flats: Examples from the Seychelles. *Estuarine Coastal and Shelf Science,* **64(2-3)**, 223-234
Shiklomanov, A.I., R.B. Lammers, M.A. Rawlins, L.C. Smith, and T.M. Pavelsky, 2007: Temporal and spatial variations in maximum river discharge from a new Russian data set. *Journal of Geophysical Research-Biogeosciences,* **112(G04S53)**
Shiogama, H., M. Watanabe, M. Kimoto, and T. Nozawa, 2005: Anthropogenic and natural forcing impacts on ENSO-like decadal variability during the second half of the 20th century. *Geophysical Research Letters,* **32(L21714)**
Shiogama, H., N. Christidis, J. Caesar, T. Yokohata, T. Nozawa, and S. Emori, 2006: Detection of greenhouse gas and aerosol influences in changes in temperature extremes. *SOLA,* **2**, 152-155
Shongwe, M.E., G.J. van Oldenborgh, B.J.J.M. van den Hurk, B. de Boer, C.A.S. Coelho, and M.K. van Aalst, 2009a: Projected Changes in Mean and Extreme Precipitation in Africa under Global Warming. Part I: East Africa. *Journal of Climate,* **under review**
Shongwe, M.E., G.J. van Oldenborgh, B.J.J.M. van den Hurk, B. de Boer, C.A.S. Coelho, and M.K. van Aalst, 2009b: Projected Changes in Mean and Extreme Precipitation in Africa under Global Warming. Part I: Southern Africa. *Journal of Climate,* **22(13)**, 3819-3837
Sidle, R.C., and H. Ochiai, 2006: *Landslides: processes, prediction, and land use*. American Geophysical Union
Sillmann, J., and E. Roeckner, 2008: Indices for extreme events in projections of anthropogenic climate change. *Climatic Change,* **86(1-2)**, 83-104
Silva, A.G., and P. Azevedo, 2008: Índices de tendências de Mudanças Climáticas no Estado da Bahia. *Engenheiria Ambiental,* **5**, 141-151
Silvestri, G.E., and C.S. Vera, 2003: Antarctic Oscillation signal on precipitation anomalies over southeastern South America. *Geophysical Research Letters,* **30(2115)**
Silvestri, G.E., and C.S. Vera, 2009: Nonstationary Impacts of the Southern Annular Mode on Southern Hemisphere Climate. *Journal of Climate,* **22(22)**, 6142-6148
Simmonds, I., C. Burke, and K. Keay, 2008: Arctic Climate Change as Manifest in Cyclone Behavior. *Journal of Climate,* **21(22)**, 5777-5796
Sinha, A., K.G. Cannariato, L.D. Stott, H. Cheng, R.L. Edwards, M.G. Yadava, R. Ramesh, and I.B. Singh, 2007: A 900-year (600 to 1500 A. D.) record of the Indian summer monsoon precipitation from the core monsoon zone of India. *Geophysical Research Letters,* **34(L16707)**
Small, I., J. van der Meer, and R.E.G. Upshur, 2001: Acting on an environmental health disaster: The case of the Aral Sea. *Environmental Health Perspectives,* **109(6)**, 547-549
Smith, K., and R. Ward, 1998: *Floods. Physical processes and human impacts*. John Wiley, Chichester, 382 pp.
Smith, L.C., 2000: Trends in Russian Arctic river-ice formation and breakup, 1917 to 1994. *Physical Geography,* **21**, 46-56
Smith, L.C., Y. Sheng, G.M. MacDonald, and L.D. Hinzman, 2005a: Disappearing Arctic Lakes. *Science,* **308(5727)**, 1429
Smith, R.L., 2008: Statistical trend analysis. In: *Weather and Climate Extremes in a Changing Climate. Regions of Focus: North America, Hawaii, Caribbean, and U.S. Pacific Islands.* [T.R. Karl, et al. (eds.)]. A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research, Washington, DC.
Smith, S.L., M.M. Burgess, D. Riseborough, and F.M. Nixon, 2005b: Recent trends from Canadian permafrost thermal monitoring network sites. *Permafrost and Periglacial Processes,* **16**, 19-30

Smits, A., A.M.G.K. Tank, and G.P. Konnen, 2005: Trends in storminess over the Netherlands, 1962-2002. *International Journal of Climatology,* **25(10)**, 1331-1344
Soares, W.R., and J.A. Marengo, 2009: Assessments of moisture fluxes east of the Andes in South America in a global warming scenario. *International Journal of Climatology,* **29(10)**, 1395-1414
Solomon, S., et al., 2007: Technical Summary. In: *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [S. Solomon, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA
Son, S.W., et al., 2008: The impact of stratospheric ozone recovery on the Southern Hemisphere westerly jet. *Science,* **320(5882)**, 1486-1489
Sorteberg, A., and J.E. Walsh, 2008: Seasonal cyclone variability at 70 degrees N and its impact on moisture transport into the Arctic. *Tellus Series A-Dynamic Meteorology and Oceanography,* **60(3)**, 570-586
Sosio, R., G.B. Crosta, and O. Hungr, 2008: Complete dynamic modeling calibration for the Thurwieser rock avalanche (Italian Central Alps) *Engineering Geology,* **100(1-2)**, 11-26
Sterl, A., H. van den Brink, H. de Vries, R. Haarsma, and E. van Meijgaard, 2009: An ensemble study of extreme storm surge related water levels in the North Sea in a changing climate. *Ocean Science,* **5(3)**, 369-378
Sterl, A., et al., 2008: When can we expect extremely high surface temperatures? *Geophysical Research Letters,* **35(L14703)**
Sterr, H., 2008: Assessment of vulnerability and adaptation to sea-level rise for the coastal zone of Germany. *Journal of Coastal Research,* **24(2)**, 380-393
Stive, M.J.F., Z.B. Wang, and C. Lakhan, 2003: Morphodynamic modeling of tidal basins and coastal inlets. In: *Advances in Coastal Modeling* [V.C. Lakhan (ed.)]. Elsevier, Amsterdam, pp. 367-392
Stoffel, M., I. Lièvre, D. Conus, M.A. Grichting, H. Raetzo, H.W. Gärtner, and M. Monbaron, 2005: 400 years of debris-flow activity and triggering weather conditions: Ritigraben, Valais, Switzerland. *Arctic, Antarctic, and Alpine Research,* **37(3)**, 387-395
Stone, D.A., A.J. Weaver, and R.J. Stouffer, 2001: Projection of climate change onto modes of atmospheric variability. *Journal of Climate,* **14**, 3551-3565
Stott, P.A., D.A. Stone, and M.R. Allen, 2004: Human contribution to the European heatwave of 2003. *Nature,* **432(7017)**, 610-614
Stott, P.A., N.P. Gillett, G.C. Hegerl, D. Karoly, D. Stone, X. Zhang, and F.W. Zwiers, 2010: Detection and attribution of climate change: a regional perspective. *Wiley Interdisciplinary Reviews: Climate Change,* **1(2)**, 192-211
Su, B., Z.W. Kundzewicz, and T. Jiang, 2009: Simulation of extreme precipitation over the Yangtze River Basin using Wakeby distribution. *Theoretical and Applied Climatology,* **96(3-4)**, 209-219
Sugahara, S., R.P. da Rocha, and R. Silveira, 2009: Non-stationary frequency analysis of extreme daily rainfall in Sao Paulo, Brazil. *International Journal of Climatology,* **29(9)**, 1339-1349
Sugi, M., H. Murakami, and J. Yoshimura, 2009: A reduction in global tropical cyclone frequency due to global warming. *SOLA,* **5**, 164-167
Sugiyama, M., H. Shiogama, and S. Emori, 2010: Precipitation extreme changes exceeding moisture content increases in MIROC and IPCC climate models. *Proceedings of the National Academy of Sciences of the United States of America,* **107(2)**, 571-575
Suppiah, R., K. Hennessy, P.H. Whetton, K. McInnes, I. Macadam, J. Bathols, J. Ricketts, and C.M. Page, 2007: Australian climate change projections derived from simulations performed for the IPCC 4th Assessment Report. *Australian Meteorological Magazine,* **56(3)**, 131-152
Suter, S., M. Laternser, W. Haeberli, R. Frauenfelder, and M. Hoelzle, 2001: Cold firn and ice of high-altitude glaciers in the Alps: measurements and distribution modelling. *Journal of Glaciology,* **47(156)**, 85-96
Sutton, R.T., B. Dong, and J.M. Gregory, 2007: Land/sea warming ratio in response to climate change: IPCC AR4 model results and comparison with observations. *Geophysical Research Letters,* **34(L02701)**
Suursaar, U., and T. Kullas, 2009: Decadal variations in wave heights off Cape Kelba, Saaremaa Island, and their relationships with changes in wind climate. *Oceanologia,* **51(1)**, 39-61
Swanson, K.L., 2008: Non-locality of Atlantic tropical cyclone intensitites. *Geochemistry Geophysics Geosystems,* **9(Q04V01)**
Tanaka, H.L., N. Ishizaki, and D. Nohara, 2005: Intercomparison of the Intensities and Trends of Hadley, Walker and Monsoon Circulations in the Global Warming Projections. *SOLA,* **1**, 77-80
Taschetto, A.S., C.C. Ummenhofer, A. Sen Gupta, and M.H. England, 2009: The effect of anomalous warming in the central Pacific on the Australian monsoon. *Geophysical Research Letters,* **36(L12704)**
Tebaldi, C., and R. Knutti, 2007: The use of the multi-model ensemble in probabilistic climate projections. *Philosophical Transactions of the Royal Society A-Mathematical Physical and Engineering Sciences,* **365(1857)**, 2053-2075
Tebaldi, C., and B. Sanso, 2009: Joint projections of temperature and precipitation change from multiple climate models: a hierarchical Bayesian approach. *Journal of the Royal Statistical Society Series a-Statistics in Society,* **172**, 83-106

Tebaldi, C., K. Hayhoe, J.M. Arblaster, and G.A. Meehl, 2006: Going to the extremes. An intercomparison of model-simulated historical and future changes in extreme events. *Climatic Change,* **79(3-4)**, 185-211
Teng, H., W.M. Washington, and G.A. Meehl, 2008: Interannual variations and future change of wintertime extratropical cyclone activity over North America in CCSM3. *Climate Dynamics,* **30(7-8)**, 673-686
Teuling, A.J., et al., 2009: A regional perspective on trends in continental evaporation. *Geophysical Research Letters,* **36(L02404)**
Thodsen, H., 2007: The influence of climate change on stream flow in Danish rivers. *Journal of Hydrology,* **333(2-4)**, 226-238
Thomas, D.S.G., M. Knight, and G.F.S. Wiggs, 2005: Remobilization of southern African desert dune systems by twenty-first century global warming. *Nature,* **435(7046)**, 1218-1221
Timbal, B., 2004: Southwest Australia past and future rainfall trends. *Climate Research,* **26**, 233-249
Timbal, B., and D. Jones, 2008: Future projections of winter rainfall in southeast Australia using a statistical downscaling technique. *Climate Change,* **86**, 165-187
Timbal, B., P. Hope, and S. Charles, 2008: Evaluating the Consistency between Statistically Downscaled and Global Dynamical Model Climate Change Projections. *Journal of Climate,* **21(22)**, 6052-6059
Timbal, B., E. Fernandez, and Z. Li, 2009: Generalization of a statistical downscaling model to provide local climate change projections for Australia. *Environmental Modelling & Software,* **24**, 341-358
Tolika, K., P. Maheras, and I. Tegoulias, 2009: Extreme temperatures in Greece during 2007: Could this be a "return to the future"? *Geophysical Research Letters,* **36(L10813)**
Tomassini, L., and D. Jacob, 2009: Spatial analysis of trends in extreme precipitation events in high-resolution climate model results and observations for Germany. *Journal of Geophysical Research-Atmospheres,* **114(D12113)**
Trenberth, K.E., J. Fasullo, and L. Smith, 2005: Trends and variability in column-integrated atmospheric water vapor. *Climate Dynamics,* **24**, 741-758
Trenberth, K.E., D. Stepaniak, J. Hurrell, and M. Fiorino, 2001: Quality of reanalyses in the tropics. *Journal of Climate,* **15**, 1499-1510
Trenberth, K.E., et al., 2007: Observations: Surface and Atmospheric Climate Change. In: *The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [S. Solomon, et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA
Trewin, B., 2010: Exposure, instrumentation and observing practice effects on land temperature measurements. *Wiley Interdisciplinary Reviews: Climate Change*, doi:10.1002/wcc.46
Trewin, B.C., and A.C.F. Trevitt, 1996: The development of composite temperature records. *International Journal of Climatology,* **16(11)**, 1227-1242
Trömel, S., and C.D. Schönwiese, 2005: A generalized method of time series decomposition into significant components including probability assessments of extreme events and application to observed German precipitation data. *Meteorologische Zeitschrift,* **14**, 417-427
Tryhorn, L., A. Lynch, and R. Abramson, 2008: On the meteorological mechanisms driving postfire flash floods: A case study. *Monthly Weather Review,* **136**, 1778-1791
Tu, M., M.J. Hall, P.J.M. de Laat, and M.J.M. de Wit, 2005: Extreme floods in the Meuse river over the past century: aggravated by land-use changes? *Physics and Chemistry of the Earth,* **30(4-5)**, 267-276
Tudhope, A.W., C.P. Chilcott, M.T. McCulloch, E.R. Cook, J. Chappell, R.M. Ellam, D.W. Lea, J.M. Lough, and G.B. Shimmield, 2001: Variability in the El Niño-Southern Oscillation Through a Glacial-Interglacial Cycle. *Science,* **291(5508)**, 1511-1517
Turner, A.G., and J.M. Slingo, 2009: Subseasonal extremes of precipitation and active-break cycles of the Indian summer monsoon in a climate-change scenario. *Quarterly Journal of the Royal Meteorological Society,* **135(640)**, 549-567
Turner, J., S.R. Colwell, G.J. Marshall, T.A. Lachlan-Cope, A.M. Carleton, P.D. Jones, V. Lagun, P.A. Reid, and S. Iagovkina, 2005: Antarctic climate change during the last 50 years. *International Journal of Climatology,* **25(3)**, 279-294
Ulbrich, U., J.G. Pinto, H. Kupfer, G.C. Leckebusch, T. Spangehl, and M. Reyers, 2008: Changing northern hemisphere storm tracks in an ensemble of IPCC climate change simulations. *Journal of Climate,* **21(8)**, 1669-1679
Ullmann, A., and V. Moron, 2008: Weather regimes and sea surge variations over the Gulf of Lions (French Mediterranean coast) during the 20th century. *International Journal of Climatology,* **28(2)**, 159-171
Ullmann, A., P.A. Pirazzoli, and A. Tomasin, 2007: Sea surges in Camargue: Trends over the 20th century. *Continental Shelf Research,* **27(7)**, 922-934
Ullmann, A., P.A. Pirazzoli, and V. Moron, 2008: Sea surges around the Gulf of Lions and atmospheric conditions. *Global and Planetary Change,* **63(2-3)**, 203-214
Ummenhofer, C.C., and M.H. England, 2007: Interannual extremes in New Zealand precipitation linked to modes of Southern Hemisphere climate variability. *Journal of Climate,* **20(21)**, 5418-5440

Ummenhofer, C.C., A. Sen Gupta, M.J. Pook, and M.H. England, 2008: Anomalous rainfall over southwest Western Australia forced by Indian Ocean sea surface temperatures. *Journal of Climate,* **21**, 5113-5134

Ummenhofer, C.C., A. Sen Gupta, M.H. England, and C.J.C. Reason, 2009a: Contributions of Indian Ocean sea surface temperatures to enhanced East African rainfall. *Journal of Climate,* **22**, 993-1013

Ummenhofer, C.C., A. Sen Gupta, A.S. Taschetto, and M.H. England, 2009b: Modulation of Australian Precipitation by Meridional Gradients in East Indian Ocean Sea Surface Temperature. *Journal of Climate,* **22**, 5597-5610

Ummenhofer, C.C., M.H. England, G.A. Meyers, P.C. McIntosh, M.J. Pook, J.S. Risbey, A. Sen Gupta, and A.S. Taschetto, 2009c: What causes Southeast Australia's worst droughts? *Geophysical Research Letters,* **36(L04706)**

Vallve, M.B., and J. Martin-Vide, 1998: Secular climatic oscillations as indicated by catastrophic floods in the Spanish Mediterranean coastal area (14th-19th centuries). *Climatic Change,* **38(4)**, 473-491

Van den Brink, H.W., G.P. Konnen, J.D. Opsteegh, G.J. Van Oldenborgh, and G. Burgers, 2005: Estimating return periods of extreme events from ECMWF seasonal forecast ensembles. *International Journal of Climatology,* **25(10)**, 1345-1354

van den Hurk, B., et al., 2007: New climate change scenarios for the Netherlands. *Water Science and Technology,* **56(4)**, 27-33

van der Linden, P., and J.F.B. Mitchell (eds.), 2009: *ENSEMBLES: Climate Change and its Impacts: Summary of research and results from the ENSEMBLES project*. Met Office Hadley Centre, FitzRoy Road, Exeter EX1 3PB, UK, 160 pp.

van Heerwaarden, C.C., J.V.G. de Arellano, A.F. Moene, and A.A.M. Holtslag, 2009: Interactions between dry-air entrainment, surface evaporation and convective boundary-layer development. *Quarterly Journal of the Royal Meteorological Society,* **135(642)**, 1277-1291

van Oldenborgh, G.J., S. Philip, and M. Collins, 2005: El Niño in a changing climate: a multi-model study. *Ocean Science,* **2**, 267-298

van Pelt, S.C., P. Kabat, H.W. ter Maat, B.J.J.M. van den Hurk, and A.H. Weerts, 2009: Discharge simulations performed with a hydrological model using bias corrected regional climate model input. *Hydrology and Earth System Sciences,* **13**, 2387-2397

Vautard, R., et al., 2007: Summertime European heat and drought waves induced by wintertime Mediterranean rainfall deficit. *Geophysical Research Letters,* **34(L07711)**

Vavrus, S.J., J.E. Walsh, W.L. Chapman, and D. Portis, 2006: The behavior of extreme cold air outbreaks under greenhouse warming. *International Journal of Climatology,* **26**, 1133-1147

Vecchi, G.A., and B.J. Soden, 2007a: Effect of remote sea surface temperature change on tropical cyclone potential intensity. *Nature,* **450(7172)**, 1066-1070

Vecchi, G.A., and B.J. Soden, 2007b: Global Warming and the Weakening of the Tropical Circulation. *Journal of Climate,* **20(17)**, 4316-4340

Vecchi, G.A., and T.R. Knutson, 2008: On Estimates of Historical North Atlantic Tropical Cyclone Activity. *Journal of Climate,* **21(14)**, 3580-3600

Vecchi, G.A., and A.T. Wittenberg, 2010: El Niño and our future climate: where do we stand? *WIRES: Climate Change*, doi:10.1002/wcc.33

Vecchi, G.A., K.L. Swanson, and B.J. Soden, 2008: Whither hurricane activity. *Science,* **322(5902)**

Vecchi, G.A., B.J. Soden, A.T. Wittenberg, I.M. Held, A. Leetmaa, and M.J. Harrison, 2006: Weakening of tropical Pacific atmospheric circulation due to anthropogenic forcing. *Nature,* **441**, 73-76

Velicogna, I., 2009: Increasing rates of ice mass loss from the Greenland and Antarctic ice sheets revealed by GRACE. *Geophysical Research Letters,* **36(L19503)**

Vera, C., and G. Silvestri, 2009: Precipitation interannual variability in South America from the WCRP-CMIP3 multi-model dataset. *Climate Dynamics,* **32(7-8)**, 1003-1014

Vera, C., G. Silvestri, B. Liebmann, and P. Gonzalez, 2006: Climate change scenarios for seasonal precipitation in South America from IPCC-AR4 models. *Geophysical Research Letters,* **33(L13707)**

Vicente-Serrano, S.M., and J.I. Lopez-Moreno, 2008: Nonstationary influence of the North Atlantic Oscillation on European precipitation. *Journal of Geophysical Research-Atmospheres,* **113(D20120)**

Vilibic, I., and J. Sepic, 2010: Long-term variability and trends of sea level storminess and extremes in European Seas. *Global and Planetary Change,* **71(1-2)**, 1-12

Vilimek, V., M.L. Zapata, J. Klimes, Z. Patzelt, and N. Santillan, 2005: Influence of glacial retreat on natural hazards of the Palcacocha Lake area, Peru. *Landslides,* **2(2)**, 107-115

Vincent, C., E. Le Meur, D. Six, M. Funk, M. Hoelzle, and S. Preunkert, 2007: Very high-elevation Mont Blanc glaciated areas not affected by the 20th century climate change. *Journal of Geophysical Research,* **112(D09120)**

Vincent, L.A., and E. Mekis, 2006: Changes in daily and extreme temperature and precipitation indices for Canada over the 20th century. *Atmosphere-Ocean,* **44(2)**, 177-193

Vincent, L.A., X. Zhang, B.R. Bonsal, and W.D. Hogg, 2002: Homogenization of daily temperatures over Canada. *Journal of Climate,* **15**, 1322-1334
Vincent, L.A., et al., 2005: Observed trends in indices of daily temperature extremes in South America 1960-2000. *Journal of Climate,* **18**, 5011-5023
Voldoire, A., and J.F. Royer, 2004: Tropical deforestation and climate variability. *Climate Dynamics,* **22(8)**, 857-874
Vonder Mühll, D., T. Stucki, and W. Haeberli, 1998: *Borehole temperatures in alpine permafrost: a ten year series*. In Proceedings of the Seventh International Conference on Permafrost, pp. 1089-1095.
Vose, R.S., D.R. Easterling, and B. Gleason, 2005: Maximum and minimum temperature trends for the globe: an update through 2004. *Geophysical Research Letters,* **32(L23822)**
Vuichard, D., and M. Zimmermann, 1987: The 1985 catastrophic drainage of a moraine-dammed lake, Khumbu Himal, Nepal: cause and consequences. *Mountain Research and Development,* **7(2)**, 91-110
Wakazuki, Y., M. Nakamura, S. Kanada, and C. Muroi, 2008: Climatological Reproducibility Evaluation and Future Climate Projection of Extreme Precipitation Events in the Baiu Season Using a High-Resolution Non-Hydrostatic RCM in Comparison with an AGCM. *Journal of the Meteorological Society of Japan,* **86(6)**, 951-967
Walsh, J., 2005: Cryosphere and Hydrology. In: *Arctic Climate Impact Assessment*. Cambridge Univesity Press, Cambridge and New York, pp. 183-242
Wang, B., and Q.H. Ding, 2006: Changes in global monsoon precipitation over the past 56 years. *Geophysical Research Letters,* **33(L06711)**
Wang, B., Q.H. Ding, X.H. Fu, I.S. Kang, K. Jin, J. Shukla, and F. Doblas-Reyes, 2005: Fundamental challenge in simulation and prediction of summer monsoon rainfall. *Geophysical Research Letters,* **32(L15711)**
Wang, G., and H.H. Hendon, 2007: Sensitivity of Australian rainfall to inter-El Niño variations. *Journal of Climate,* **20**, 4211-4226
Wang, G.L., 2005: Agricultural drought in a future climate: results from 15 global climate models participating in the IPCC 4th assessment. *Climate Dynamics,* **25**, 739-753
Wang, J., and X. Zhang, 2008: Downscaling and projection of winter extreme daily precipitation over North America. *Journal of Climate,* **21(5)**, 923-937
Wang, S., R. McGrath, J. Hanafin, P. Lynch, T. Semmler, and P. Nolan, 2008: The impact of climate change on storm surges over Irish waters. *Ocean Modelling,* **25(1-2)**, 83-94
Wang, X.L., V.R. Swail, and F.W. Zwiers, 2006a: Climatology and changes of extratropical cyclone activity: Comparison of ERA-40 with NCEP-NCAR reanalysis for 1958-2001. *Journal of Climate,* **19(13)**, 3145-3166
Wang, X.L., V.R. Swail, and W. Perrie, 2006b: Historical and possible future changes of wave heights in northern hemisphere ocean. In: *Atmosphere-Ocean Interactions*. Vol. 2. Wessex Institute of Technology Press, Southhampton, pp. 240-240
Wang, X.L., V.R. Swail, and A. Cox, 2009a: Dynamical versus Statistical Downscaling Methods for Ocean Wave Heights. *International Journal of Climatology,* **30(3)**, 317-332
Wang, X.L., F.W. Zwiers, V.R. Swail, and Y. Feng, 2009b: Trends and variability of storminess in the Northeast Atlantic region, 1874-2007. *Climate Dynamics,* **33(7-8)**, 1179-1195
Wang, X.L., V.R. Swail, F.W. Zwiers, X. Zhang, and Y. Feng, 2009c: Detection of external influence on trends of atmospheric storminess and northern oceans wave heights. *Climate Dynamics,* **32(2-3)**, 189-203
Wang, X.M., Z.J. Zhou, and Z.B. Dong, 2006c: Control of dust emissions by geomorphic conditions, wind environments and land use in northern China: An examination based on dust storm frequency from 1960 to 2003. *Geomorphology,* **81(3-4)**, 292-308
Wang, X.M., Y. Yang, Z.B. Dong, and C.X. Zhang, 2009d: Responses of dune activity and desertification in China to global warming in the twenty-first century. *Global and Planetary Change,* **67(3-4)**, 167-185
Watterson, I.G., 2006: The intensity of precipitation during extratropical cyclones in global warming simulations: a link to cyclone intensity? *Tellus Series A-Dynamic Meteorology and Oceanography,* **58(1)**, 82-97
Webb, A.P., and P. Kench, 2010: The dynamic response of reef islands to sea level rise: Evidence from multi-decadal analysis of island change in the central Pacific. *Global and Planetary Change*, doi:10.1016/j.gloplacha.2010.05.003
Webster, P.J., G.J. Holland, J.A. Curry, and H.R. Chang, 2005: Changes in tropical cyclone number, duration, and intensity in a warming environment. *Science,* **309(5742)**, 1844-1846
Wehner, M.F., R.L. Smith, G. Bala, and P. Duffy, 2010: The effect of horizontal resolution on simulation of very extreme US precipitation events in a global atmosphere model. *Climate Dynamics,* **34(2-3)**, 241-247
Weisse, R., and H. Günther, 2007: Wave climate and long-term changes for the Southern North Sea obtained from a high-resolution hindcast 1958-2002. *Ocean Dynamics,* **57(3)**, 161-172
Weisse, R., H. Von Storch, and F. Feser, 2005: Northeast Atlantic and North Sea storminess as simulated by a regional climate model during 1958-2001 and comparison with observations. *Journal of Climate,* **18(3)**, 465-479
Weng, H., S.K. Behera, and T. Yamagata, 2009: Anomalous winter climate conditions in the Pacific Rim during recent El Niño Modoki and El Niño events. *Climate Dynamics,* **32**, 663-674

Wieczorek, G.F., T. Glade, M. Jakob, and O. Hungr, 2005: Climatic factors influencing occurrence of debris flows. In: *Debris-flow hazards and related phenomena* [M. Jakob, and O. Hungr (eds.)]. Springer, Berlin, Heidelberg, pp. 325-362
Wijngaard, J.B., A.M.G.K. Tank, and G.P. Konnen, 2003: Homogeneity of 20th century European daily temperature and precipitation series. *International Journal of Climatology,* **23(6)**, 679-692
Wilby, R.L., K.J. Beven, and N.S. Reynard, 2008: Climate change and fluvial flood risk in the UK: more of the same? *Hydrological Processes,* **22(14)**, 2511-2523
Wilby, R.L., J. Troni, Y. Biot, L. Tedd, B.C. Hewitson, D.M. Smith, and R.T. Sutton, 2009: A review of climate risk information for adaptation and development planning. *International Journal of Climatology,* **29(9)**, 1193-1215
Wild, M., 2009: Global dimming and brightening: A review. *Journal of Geophysical Research,* **114(D00D16)**
Willett, K.M., N.P. Gillett, P.D. Jones, and P.W. Thorne, 2007: Attribution of observed surface humidity changes to human influence. *Nature,* **449(7163)**, 710-712
Willett, K.M., P.D. Jones, N.P. Gillett, and P.W. Thorne, 2008: Recent Changes in Surface Humidity: Development of the HadCRUH Dataset. *Journal of Climate,* **21(20)**, 5364-5383
Wing, A.A., A.H. Sobel, and S.J. Camargo, 2007: Relationship between the potential and actual intensities of tropical cyclones on interannual time scales. *Geophysical Research Letters,* **34(L08810)**
Wittenberg, A.T., 2009: Are historical records sufficient to constrain ENSO simulations? *Geophysical Research Letters,* **36(L12702)**
Wong, S., A.E. Dessler, N.M. Mahowald, P.R. Colarco, and A. da Silva, 2008: Long-term variability in Saharan dust transport and its link to North Atlantic sea surface temperature. *Geophysical Research Letters,* **35(L07812)**
Woodruff, J.D., J.P. Donnelly, K.A. Emanual, and P. Lane, 2008a: Assessing sedimentary records of paleohurricane activity using modeled hurricane climatology. *Geochemistry Geophysics Geosystems,* **9(Q09V10)**
Woodruff, J.D., J.P. Donnelly, D. Mohrig, and W.R. Geyer, 2008b: Reconstructing relative flooding intensities responsible for hurricane-induced deposits from Laguna Playa Grande, Vieques, Puerto Rico. *Geology,* **36(5)**, 391-394
Woodworth, P.L., and D.L. Blackman, 2004: Evidence for systematic changes in extreme high waters since the mid-1970s. *Journal of Climate,* **17(6)**, 1190-1197
Woolings, T., A. Hannachi, B. Hoskins, and A. Turner, 2010: A regime view of the North Atlantic Oscillation and its response to anthropogenic forcing. *Journal of Climate,* **23**, 1291-1307
Woollings, T.J., B.J. Hoskins, M. Blackburn, and P. Berrisford, 2008: A new Rossby wave-breaking interpretation of the North Atlantic Oscillation. *Journal of Atmospheric Science,* **65**, 609-626
Wu, Q., and T. Zhang, 2008: Recent permafrost warming on the Qinghai-Tibetan Plateau. *Journal of Geophysical Research,* **113(D13108)**
Wu, Q., and T. Zhang, 2010: Changes in Active Layer Thickness over the Qinghai-Tibetan Plateau from 1995-2007. *Journal of Geophysical Research,* **115(D09107)**
Xie, B., Q. Zhang, and Y. Wang, 2008: Trends in hail in China during 1960–2005. *Geophysical Research Letters,* **35(L13801)**
Xin, W., L. Shiyin, G. Wanqin, and X. Junli, 2008: Assessment and Simulation of Glacier Lake Outburst Floods for Longbasaba and Pida Lakes, China. *Mountain Research and Development,* **28(3)**, 310-317
Xoplaki, E., J.F. Gonzalez-Rouco, J. Luterbacher, and H. Wanner, 2003: Mediterranean summer air temperature variability and its connection to the large-scale atmospheric circulation and SSTs. *Climate Dynamics,* **20(7-8)**, 723-739
Xu, M., C.P. Chang, C.B. Fu, Y. Qi, A. Robock, D. Robinson, and H.M. Zhang, 2006: Steady decline of east Asian monsoon winds, 1969-2000: Evidence from direct ground measurements of wind speed. *Journal of Geophysical Research-Atmospheres,* **111(D24111)**
Yang, D., D. Kane, Z. Zhang, D. Legates, and B. Goodison, 2005: Bias corrections of long-term (1973-2004) daily precipitation data over the northern regions. *Geophysical Research Letters,* **32(L19501)**
Yao, C., S. Yang, W. Qian, Z. Lin, and M. Wen, 2008: Regional summer precipitation events in Asia and their changes in the past decades. *Journal of Geophysical Research-Atmospheres,* **113(D17107)**
Yeh, S.W., J.S. Kug, B. Dewitte, M.H. Kwon, B.P. Kirtman, and F.F. Jin, 2009: El Niño in a changing climate. *Nature,* **461**, 511-514
Yoshikawa, K., and L.D. Hinzman, 2003: Shrinking thermokarst ponds and groundwater dynamics in discontinuous permafrost. *Permafrost and Periglacial Processes,* **14(2)**, 151-160
You, Q., S. Kang, E. Aguilar, and Y. Yan, 2008: Changes in daily climate extremes in the eastern and central Tibetan Plateau during 1961-2005. *Journal of Geophysical Research-Atmospheres,* **113(D07101)**
You, Q., S. Kang, E. Aguilar, N. Pepin, W.A. Flügel, Y. Yan, Y. Xu, Y. Zhang, and J. Huang, 2010: Changes in daily climate extremes in China and their connection to the large scale atmospheric circulation during 1961-2003. *Climate Dynamics*, doi:10.1007/s00382-009-0735-0
Yu, J.Y., and H.Y. Kao, 2007: Decadal changes of ENSO persistence barrier in SST and ocean heat content indices: 1958-2001. *Journal of Geophysical Research-Atmospheres,* **112(D13106)**

Yu, K.F., J.X. Zhao, Q. Shi, and Q.S. Meng, 2009: Reconstruction of storm/tsunami records over the last 4000 years using transported coral blocks and lagoon sediments in the southern South China Sea. *Quaternary International,* **195(1-2)**, 128-137
Zemp, M., W. Haeberli, and J. Eamer, 2007: Glaciers and ice caps. In: *Global Outlook for Ice and Snow*. UNEP/GRID, Arendal, Norway, pp. 115-152 http://www.unep.org/geo/geo_ice/ (15/07/10)
Zemp, M., W. Haeberli, M. Hoelzle, and F. Paul, 2006: Alpine glaciers to disappear within decades? *Geophysical Research Letters,* **33(L13504)**
Zhai, P.M., X. Zhang, H. Wan, and X.H. Pan, 2005: Trends in total precipitation and frequency of daily precipitation extremes over China. *Journal of Climate,* **18(7)**, 1096-1108
Zhang, K.Q., B.C. Douglas, and S.P. Leatherman, 2004a: Global warming and coastal erosion. *Climatic Change,* **64(1-2)**, 41-58
Zhang, Q., Y. Guan, and H. Yang, 2008a: ENSO amplitude change in observation and coupled models. *Advances in Atmospheric Sciences,* **25(3)**, 361-366
Zhang, T., R.G. Barry, K. Knowles, J.A. Heginbottom, and J. Brown, 1999: Statistics and characteristics of permafrost and ground ice distribution in the Northern Hemisphere. *Polar Geography,* **23(2)**, 147-169
Zhang, T., et al., 2005: Spatial and temporal variability of active layer thickness over the Russian Arctic drainage basin. *Journal of Geophysical Research,* **110(D16101)**
Zhang, X., F.W. Zwiers, and G. Li, 2004b: Monte Carlo experiments on the detection of trends in extreme values. *Journal of Climate,* **17**, 1945-1952
Zhang, X., F.W. Zwiers, and P.A. Stott, 2006: Multimodel multisignal climate change detection at regional scale. *Journal of Climate,* **19(17)**, 4294-4307
Zhang, X., K.D. Harvey, W.D. Hogg, and T.R. Yuzyk, 2001: Trends in Canadian streamflow. *Water Resources Research,* **37(4)**, 987-998
Zhang, X., J.E. Walsh, J. Zhang, U.S. Bhatt, and M. Ikeda, 2004c: Climatology and interannual variability of arctic cyclone activity: 1948-2002. *Journal of Climate,* **17(12)**, 2300-2317
Zhang, X., A. Sorteberg, J. Zhang, R. Gerdes, and J.C. Comiso, 2008b: Recent radical shifts of atmospheric circulations and rapid changes in Arctic climate system. *Geophysical Research Letters,* **35(L22701)**
Zhang, X., J. Wang, F.W. Zwiers, P. Ya, and P.Y. Groisman, 2010: The influence of large scale climate variability on winter maximum daily precipitation over North America. *Journal of Climate*, doi:10.1175/2010JCLI3249.1
Zhang, X., F.W. Zwiers, G.C. Hegerl, F.H. Lambert, N.P. Gillett, S. Solomon, P.A. Stott, and T. Nozawa, 2007a: Detection of human influence on twentieth-century precipitation trends. *Nature,* **448(7152)**, 461-465
Zhang, X.Y., S.L. Gong, T.L. Zhao, R. Arimoto, Y.Q. Wang, and Z.J. Zhou, 2003: Sources of Asian dust and role of climate change versus desertification in Asian dust emission. *Geophysical Research Letters,* **30(2272)**
Zhang, Y.Q., C.M. Liu, Y.H. Tang, and Y.H. Yang, 2007b: Trends in pan evaporation and reference and actual evapotranspiration across the Tibetan Plateau. *Journal of Geophysical Research-Atmospheres,* **112(D12110)**
Zhao, C.S., X. Dabu, and Y. Li, 2004: Relationship between climatic factors and dust storm frequency in Inner Mongolia of China. *Geophysical Research Letters,* **31(L01103)**
Zhao, M., I.M. Held, S.J. Lin, and G.A. Vecchi, 2009: Simulations of Global Hurricane Climatology, Interannual Variability, and Response to Global Warming Using a 50-km Resolution GCM. *Journal of Climate,* **22(24)**, 6653-6678
Zhong, D.C., 1999: The dynamic changes and trends of modern desert in China. *Advance in Earth Sciences,* **14(3)**, 229-234 (in Chinese)
Zhou, T.J., L.X. Zhang, and H.M. Li, 2008a: Changes in global land monsoon area and total rainfall accumulation over the last half century. *Geophysical Research Letters,* **35(L16707)**
Zhou, T.J., R.C. Yu, H.M. Li, and B. Wang, 2008b: Ocean forcing to changes in global monsoon precipitation over the recent half-century. *Journal of Climate,* **21(15)**, 3833-3852
Zhou, T.J., et al., 2009: The CLIVAR C20C project: which components of the Asian-Australian monsoon circulation variations are forced and reproducible? *Climate Dynamics,* **33(7-8)**, 1051-1068
Zhou, Z.J., and G.C. Zhang, 2003: Typical severe dust storms in northern China during 1954-2002. *Chinese Science Bulletin,* **48(21)**, 2366-2370
Zimmermann, M., and W. Haeberli, 1993: Climatic change and debris flow activity in high-mountain areas-a case study in the Swiss Alps. *Catena Supplement,* **22**, 59-59
Zimmermann, M., P. Mani, and H. Romang, 1997: Magnitude-frequency aspects of alpine debris flows. *Eclogae Geologicae Helvetiae,* **90(3)**, 415-420
Zin, W.Z.W., S. Jamaludin, S.M. Deni, and A.A. Jemain, 2009: Recent changes in extreme rainfall events in Peninsular Malaysia: 1971-2005. *Theoretical and Applied Climatology,* **99(3-4)**, 303-314
Zolina, O., C. Simmer, A. Kapala, S. Bachner, S. Gulev, and H. Maechel, 2008: Seasonally dependent changes of precipitation extremes over Germany since 1950 from a very dense observational network. *Journal of Geophysical Research - Atmospheres,* **113(D06110)**

Zou, X., L.V. Alexander, D. Parker, and J. Caesar, 2006: Variations in severe storms over China. *Geophysical Research Letters,* **33(L17701)**
Zwiers, F.W., and V.V. Kharin, 1998: Changes in the extremes of the climate simulated by CCC GCM2 under CO2 doubling. *Journal of Climate,* **11**, 2200-2222
Zwiers, F.W., X. Zhang, and Y. Feng, 2010: Anthropogenic influence on long return period daily temperature extremes at regional scales. *Journal of Climate,* **submitted**

Chapter 3: Changes in Climate Extremes and their Impacts on the Natural Physical Environment

Coordinating Lead Authors:
Neville Nicholls (Australia), Sonia Seneviratne (Switzerland)

Lead Authors:
David Easterling (USA), Amadou Gaye (Senegal), Clare Goodess (United Kingdom), Shinjiro Kanae (Japan), James Kossin (USA), Yali Luo (China), Jose Marengo (Brazil), Kathleen McInnes (Australia), Mohamed Rahimi (Iran), Markus Reichstein (Germany), Asgeir Sorteberg (Norway), Carolina Vera (Argentina), Xuebin Zhang (Canada)

Contributing Authors:
Gerado Benito (Spain), Tereza Cavazos (Mexico), Paul Della-Marta (Switzerland), Sunling Gong (Canada), B. N. Goswami (India), Mark Hemer (Australia), Bart van den Hurk (Netherlands), Christian Huggel (Switzerland), Akio Kitoh (Japan), Simon Mason (USA), Wassila Thiaw (USA), Pascal Yiou (France), Tingjun Zhang (USA), Tianjun Zhou (China), Francis Zwiers (Canada)

Tables and Figures

Table 3.1: Overview of considered extremes and summary of observed and projected changes on global scale. Regional details on observed and projected changes in temperature and precipitation extremes are provided in Tables 3.2 and 3.3.

		Observed Changes (since 1950)	Attribution of Observed Changes	Projected Changes (until 2100)
Weather and climate elements	Temperature	*Very likely* decrease in number of unusually cold days and nights. *Very likely* increase in number of unusually warm days and nights, both on global and regional basis. *Likely* increase in warm spells, including heat waves.	*Likely* anthropogenic influence on trends in extreme temperature.	*Virtually certain* decrease in number of unusually cold days and nights (as defined with regard to 1960-1990 climate). *Virtually certain* increase in number of unusually warm days and nights (with regard to 1970-1990 climate). *Very likely increase in* warm spells, including heat waves (with regard to 1960-1990 climate).
	Precipitation	*Likely* global trend towards increases in the frequency of heavy precipitation events.	*More likely than not* anthropogenic influence on global trend towards increases in frequency of heavy precipitation events since 1950.	*Very likely* increase in frequency of heavy precipitation events over most areas of the globe.
	Winds	*Insufficient evidence*	*Insufficient evidence*	Increases *likely* in mid- to high-latitudes.
Phenomena related to weather and climate extremes	Monsoons	*More likely than not* changes in monsoon characteristics (precipitation amounts, patterns) but *insufficient evidence* for more specific statements.	*Insufficient evidence*	*Insufficient reliability of climate models*
	El Niño and other modes of variability	*More likely than not* change in center of El Niño Southern Oscillation (ENSO) episodes (more frequent central equatorial Pacific events). *Insufficient evidence* for more specific statements on ENSO trends. *Likely* trends in North Atlantic Oscillation (NAO) and Southern Annular Mode (SAM).	*Likely* anthropogenic influence on identified trends in NAO and SAM.	*Insufficient congruence of climate scenarios* for detailed statements on ENSO and other modes of variability.
	Tropical cyclones	*No trend* in global annual frequency of tropical cyclones over period of satellite observations (1970-present). *Lack of consensus* regarding fidelity and significance of non-zero frequency trends in individual ocean basins. *Significant increasing global trend* in intensity since 1983.	*Insufficient evidence*	*Likely* decrease or unchanged global frequency of tropical cyclones. *Likely* increase in mean maximum wind speed, but possibly not in all basins. *Likely* increase in tropical cyclone-related rainfall rates.
	Extra-tropical cyclones	*Likely* poleward shift in extratropical cyclones. *More likely than not* intensification of extratropical cyclones in high latitudes.	*Likely* anthropogenic influence on poleward shift.	*Likely* impacts on regional cyclone activity. *Likely* reduction of mid-latitude storms. *More likely than not* increase in high-latitude cyclone number and intensity.

Impacts on physical environment	Droughts	*Likely* increase in area affected by meteorological drought (precipitation deficit). *Likely* increase in total area affected by agricultural drought based on precipitation and temperature trends, as well as PDSI-based analyses, but *insufficient direct evidence* from actual observations (soil moisture deficits).	*More likely than not* influence on increase in area affected by droughts.	*Likely* increase in area affected by droughts.
	Floods	*Insufficient observations* of change in the magnitude and frequency in floods at the global level. *Likely* earlier spring peak in snow-dominated regions.	*More likely than not* anthropogenic influence on floods. *Likely* anthropogenic influence on earlier spring peak in snow-dominated regions.	*Insufficient literature* except for the earlier spring peak in snow-dominated regions. *Likely* earlier spring peak in snow-dominated regions.
	Extreme sea level and coastal impacts	*Likely* increase in extreme high water worldwide related to trends 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 trends in extreme sea levels.
	Other impacts	*More likely than not* increase in large landslides in some regions. *Likely* thawing of permafrost with *likely* resultant physical impacts.	*Likely* anthropogenic influence on thawing of permafrost. *Insufficient evidence* for trends in other physical impacts in cold regions.	New and potentially unstable lakes are *likely* to form during the 21st century following glacier retreat. Earlier snow melt is *more likely than not* to result in earlier onset of high-mountain debris flows, and shallow landslides in lower mountain ranges are *more likely than not* to increase with the projected higher precipitation intensities. Arctic coastal erosion is *likely* to increase.

Table 3.2: Regional observed changes in temperature and precipitation extremes. Assessments for which no likelihood statements are available yet are displayed in grey in the Table (empty arrows on Figures 3.1 and 3.2).

Regions	Sub-Region	Tmax	Tmin	Warm Spells (Heat Waves)	Heavy Precipitation	Drought
		Observations	Observations	Observations	Observations	Observations
North America	All North America	*Likely* overall increase in unusually warm days, decrease in unusually cold days (Alexander et al., 2006).	*Likely* overall decrease in unusually cold nights, increase in unusually warm nights (Alexander et al., 2006).	Increase since 1960 (Kunkel et al., 2008).	*Likely* increase in many areas since 1950, (Trenberth et al., 2007; Kunkel et al., 2008).	No overall change, regional variability, 1930s drought dominates (Kunkel et al., 2008).
	W. North America	*Very likely* large increases in unusually warm days, large decreases in unusually cold days (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*Very likely* large decreases in unusually cold nights, large increases in unusually warm nights (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	Increase in warm spells (Alexander et al., 2006).	General increase, decrease in some areas, (Alexander et al., 2006).	Slight increase since 1950, large variability, large drought of 1930s dominates (Kunkel et al., 2008).
	Central North America	*Very likely* small increases in unusually warm days, decreases in unusually cold days (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*Likely* small decreases in unusually cold nights, increases in unusually warm nights (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	Some areas increase, others decrease (Alexander et al., 2006).	*Very likely* increase since 1950, (Alexander et al., 2006).	Slight decrease since 1950, large variability, large drought of 1930s dominates (Kunkel et al., 2008).
	E. North America	*Very likely* increases in unusually warm days, decreases in unusually cold days (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*Very likely* decreases in unusually cold nights, increases in unusually warm nights (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	Some areas increase, others decrease (Alexander et al., 2006).	*Very likely* increase since 1950 (Alexander et al., 2006).	Slight decrease since 1950, large variability, large drought of 1930s dominates (Kunkel et al., 2008).
	Alaska	*Very likely* large increases in unusually warm days, large decreases in unusually cold days (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	*Very likely* large decreases in unusually cold nights, large increases in unusually warm nights (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).		Suggestion of increase, no significant trend (Kunkel et al., 2008).	*More likely than not* slight increase since the 1950s (Kunkel et al., 2008).
	E. Canada,	*Likely* increases in unusually	*Likely* decreases in unusually	Some areas increase, most	Increase in a few areas	

	Greenland, Iceland	warm days in some areas, decrease in others. Decreases in unusually cold days 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).	cold nights, increases in unusually warm nights (Robeson, 2004; Vincent and Mekis, 2006; Kunkel et al., 2008; Peterson et al., 2008a).	others decrease (Alexander et al., 2006).	(Alexander et al., 2006).	
Europe	All Europe	*Very likely* increases in unusually warm days, decreases in unusually cold days (Kiktev et al., 2003; Bartholy and Pongracz, 2007; Della-Marta et al., 2007a; Trenberth et al., 2007; Kurbis et al., 2009).	*Very likely* decreases in unusually cold nights, increases in unusually warm nights (Kiktev et al., 2003; Bartholy and Pongracz, 2007; Trenberth et al., 2007; Kurbis et al., 2009).		*More likely than not* increase in most areas, decrease in a few areas, (Alexander et al., 2006; Bartholy and Pongracz, 2007).	
	N. Europe	*Very likely* increases in unusually warm days, decreases in unusually cold days (Kiktev et al., 2003; Bartholy and Pongracz, 2007; Della-Marta et al., 2007a; Trenberth et al., 2007; Kurbis et al., 2009).	*Very likely* decreases in unusually cold nights, increases in unusually warm nights (Kiktev et al., 2003; Bartholy and Pongracz, 2007; Trenberth et al., 2007; Kurbis et al., 2009).		*More likely than not* increase in most areas, decrease in a few areas, (Alexander et al., 2006; Bartholy and Pongracz, 2007).	
	S. Europe and Mediterranean	*Likely* large increases in unusually warm days, *likely* decreases in unusually cold days (Della-Marta et al., 2007a; Trenberth et al., 2007).	*Likely* decrease in unusually cold nights, increases in unusually warm nights (Trenberth et al., 2007).	*Likely* large increase in heatwave length in the Iberian Peninsula (Della-Marta et al., 2007a). *Likely* large increase in heat wave intensity, heat wave number and heat wave length in summer in the Eastern Mediterranean (Kuglitsch et al., 2010).	*More likely than not* increase in most areas, decrease in a few areas, (Alexander et al., 2006).	
Africa	All Africa					
	W. Africa	*Likely* increases in unusually	*Likely* decreases in unusually		Increase in many areas	

		warm days, decreases in unusually cold days (Trenberth et al., 2007).	cold nights, increases in unusually warm nights (Trenberth et al., 2007).		(Trenberth et al., 2007).	
	E. Africa	Increases in unusually warm days, decreases in unusually cold days (Trenberth et al., 2007).	Decreases in unusually cold nights, increases in unusually warm nights (Trenberth et al., 2007).		Decrease (Trenberth et al., 2007).	
	S. Africa	Increases in unusually warm days, decreases in unusually cold days (Trenberth et al., 2007).	Decreases in unusually cold nights, increases in unusually warm nights (Trenberth et al., 2007).		Increase (Trenberth et al., 2007).	
	Sahara	Increases in unusually warm days, decreases in unusually cold days (Trenberth et al., 2007).	Decreases in unusually cold nights, increases in unusually warm nights (Trenberth et al., 2007).			
Central and South America	All South America					
	Central America and northern South America	Increases in unusually warm days, decreases in unusually cold days (Aguilar et al., 2005; Alexander et al., 2006; Brown et al., 2008).	Decreases in unusually cold nights, increases in unusually warm nights (Aguilar et al., 2005; Alexander et al., 2006; Brown et al., 2008).	A few areas increase, a few others decrease (Aguilar et al., 2005; Alexander et al., 2006).	Increase in many areas, decrease in a few areas, (Aguilar et al., 2005; Alexander et al., 2006).	Increase of CDD in some areas, others decrease (Aguilar et al., 2005).
	Amazon	Increases in unusually warm days, decreases in unusually cold days (river mouth (Alexander et al., 2006).	Decreases in unusually cold nights, increases in unusually warm nights (Alexander et al., 2006; Dufek et al., 2008).		Increase in many areas, decrease in a few areas (Alexander et al., 2006; Haylock et al., 2006b).	Slight decrease of CDD (Dufek et al., 2008).
	Northeastern Brazil	Increases in unusually warm days (Silva and Azevedo, 2008).	Increases in unusually warm nights (Silva and Azevedo, 2008).		Increase in many areas, decrease in a few areas, (Alexander et al., 2006; Haylock et al., 2006b; Santos and Brito, 2007; Silva and Azevedo, 2008; Santos et al., 2009).	Increase of CDD in many areas, decrease in a few areas (Santos and Brito, 2007; Silva and Azevedo, 2008; Santos et al., 2009).
	Southeastern South America	Increases in unusually warm days in some areas, decrease in others. Decreases in unusually cold days in some areas, increase in others, (Rusticucci and Barrucand, 2004; Vincent	Decreases in unusually cold nights, increases in unusually warm nights (Rusticucci and Barrucand, 2004; Vincent et al., 2005; Alexander et al., 2006; Brown et al., 2008;	Some areas increase, others decrease (Alexander et al., 2006).	Increase (Alexander et al., 2006; Dufek et al., 2008; Sugahara et al., 2009; Penalba and Robeldo, 2010).	Slight increase, large variability, (Haylock et al., 2006b; Dufek and Ambrizzi, 2008; Dufek et al., 2008; Llano and Penalba, 2010; Penalba

		et al., 2005; Alexander et al., 2006; Brown et al., 2008; Rusticucci and Renom, 2008; Marengo et al., 2009b).	Rusticucci and Renom, 2008; Marengo et al., 2009b).			and Robeldo, 2010).
	W. Coast South America	Increases in unusually warm days in some areas, decrease in others. Decreases in unusually cold days in some areas, increase in others, (Rosenbluth et al., 1997; Vincent et al., 2005; Alexander et al., 2006).	Decreases in unusually cold nights, increases in unusually warm nights (Rosenbluth et al., 1997; Vincent et al., 2005; Alexander et al., 2006).		Decrease in many areas, increase in a few areas (Alexander et al., 2006; Haylock et al., 2006b).	Slight increase in some areas, (Dufek et al., 2008).
Asia	All Asia					
	N. Asia	*Likely* increases in unusually warm days, decreases in unusually cold days (Alexander et al., 2006).	Decreases in unusually cold nights, increases in unusually warm nights (Alexander et al., 2006; Trenberth et al., 2007).		Increase, (Trenberth et al., 2007).	
	Central Asia	*Likely* increases in unusually warm days, decreases in unusually cold days (Alexander et al., 2006; Trenberth et al., 2007).	Decreases in unusually cold nights, increases in unusually warm nights (Alexander et al., 2006; Trenberth et al., 2007).			
	East Asia	*Likely* increases in unusually warm days, decreases in unusually cold days (Trenberth et al., 2007; Ding et al., 2009).	Decreases in unusually cold nights, increases in unusually warm nights (Trenberth et al., 2007; Ding et al., 2009).	Increase in warm season heat waves in China (Ding et al., 2009), but decline in all warm spells (Alexander et al., 2006).		
	S.E. Asia	Increases in unusually warm days, decreases in unusually cold days, northern part (Alexander et al., 2006).	Decreases in unusually cold nights, increases in unusually warm nights, northern part. (Alexander et al., 2006).			
	S. Central Asia	Decrease in unusually warm and cold days, (Alexander et al., 2006).	Decreases in unusually cold nights, increases in unusually warm nights (Alexander et al., 2006).			
	Tibetan Plateau	Decrease in unusually warm and cold days (Alexander et al., 2006).	Decreases in unusually cold nights, increases in unusually warm nights (Alexander et al., 2006).			

	W. Asia	*More likely than not* decrease in unusually cold days and very likely increase in unusually warm days (Choi et al., 2009; Rahimzadeh et al., 2009; Rehman, 2010).	*Likely* decrease in unusually cold nights and likely increase in unusually warm nights (Choi et al., 2009; Rehman, 2010).		*More likely than not* decrease in heavy precipitation events. (Kwarteng et al., 2009; Rahimzadeh et al., 2009).	*Likely* increase in drought (Kwarteng et al., 2009; Rahimzadeh et al., 2009).
Australia/ New Zealand	N. Australia/NZ	Increases in unusually warm days, decreases in unusually cold days (Alexander et al., 2006; Trenberth et al., 2007).	Decreases in unusually cold nights, increases in unusually warm nights (Alexander et al., 2006).			
	S. Australia/NZ	Increases in unusually warm days, decreases in unusually cold days (Alexander et al., 2006; Trenberth et al., 2007).	Decreases in unusually cold nights, increases in unusually warm nights (Alexander et al., 2006).			

Table 3.3: Projected regional changes in temperature and precipitation extremes. The key for the employed abbreviations is found below the Table. Assessments for which no likelihood statements are available yet are displayed in grey in the Table (empty arrows on Figures 3.3 and 3.4).

Regions	**Sub-Region**	Tmax [HD: hot days CD: cold days]		Tmin [WN: warm nights CN: cold nights]		Warm Spells [HWD: heat wave duration]		Heavy precipitation [HPD: heavy precipitation days HPC: heavy precipitation contribution]		Dry spells [CDD: consecutive dry days] [EDI: effective dry days]	
		Projections		Projections		Projections		Projections		Projections	
North America	All North America										
	Canada	HD *very likely* to increase & CD *very likely* to decrease over all regions (Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007b).	A	WN *very likely* to increase & CN *very likely* to decrease over all regions (Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007b).	**A**	*Very likely* more frequent heat waves & warm spells over all regions (Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007b).	A	*Very likely* more frequent & intense HPD over most regions (Christensen et al., 2007; Kharin et al., 2007; Meehl et al., 2007b).	A		
	USA & N Mexico	HD *very likely* to increase & CD *very likely* to decrease over all regions (Christensen et al., 2007; Karl et al., 2008).	A	WN *very likely* to increase & CN *very likely* to decrease over all regions (Christensen et al., 2007; Karl et al., 2008).	A	*Very likely* more frequent heat waves & warm spells over all regions (Christensen et al., 2007; Karl et al., 2008).	A	*Very likely* more frequent & intense HPD over most regions except SW US and N Mexico (Christensen et al., 2007; Karl et al., 2008).	A	*Likely* increase in drought area in SW US & N Mexico. No change or possible decline in other regions (Christensen et al., 2007; Karl et al., 2008).	A
	W. North America										
	Central North America										
	E. North America										
	Alaska										
	E. Canada, Greenland, Iceland										
	All Europe	HD *very likely* to	A	CN *very likely* to	A	HWD *very likely* to	A	*Very likely* increases in	A	Little/no change in	A

		increase – largest increases in summer and C/S Europe & smallest in N Europe (Scandinavia) (Goubanova and Li, 2007; Kjellstrom et al., 2007; Koffi and Koffi, 2008). Tmax changes generally > mean changes (Diffenbaugh et al., 2007; Kjellstrom et al., 2007; Fischer and Schär, 2009; Fischer and Schär, 2010).	**R**	decrease – largest decreases in winter & E Europe & Scandinavia (Goubanova and Li, 2007; Kjellstrom et al., 2007; Sillmann and Roeckner, 2008). Tmin changes generally > mean changes (Diffenbaugh et al., 2007; Goubanova and Li, 2007; Kjellstrom et al., 2007). WN *very likely* to increase – largest increases in Mediterranean (Sillmann and Roeckner, 2008).	G **R**	increase (also increases in intensity & frequency) – likely by a factor of at least (Beniston et al., 2007; Christensen et al., 2007; Kysely and Beranova, 2009)	**R**	HPD and decreases in return periods of long (e.g., 5-day) and short (1-day) HP across most of Europe, but uncertainty in magnitude of changes (Beniston et al., 2007; Fowler et al., 2007b; Sillmann and Roeckner, 2008; Kendon et al., 2009). *Likely* increase in HPC in some regions (Boberg et al., 2009a; Kendon et al., 2009). *Likely* greater changes in extremes and rarer events than mean. *Very likely* increase in HP intensity (& increase in HPC) despite decrease in summer mean in some regions – e.g. C Europe (Beniston et al., 2007; Fowler et al., 2007b; Haugen and Iverson, 2008; May, 2008; Kysely and Beranova, 2009).	**G** **R**	CDD in N Europe, increase in C Europe and largest increases in S Europe. 21 Frequency/length of CDD increases over much of the continent – length increases in the S (May, 2008).	G R
Europe	N. Europe	See all Europe		See all Europe		HWD *very likely* to increase, but summer increases < than in S Europe (Beniston et al., 2007; Kysely and Beranova, 2009).	A **R**	*Very likely* increases in HP (intensity and frequency) north of 45N in winter (Frei et al., 2006; Beniston et al., 2007; Kendon et al., 2008).	A R	Little/no change in CDD (Sillmann and Roeckner, 2008).	A G
	S. Europe and Mediterranean	*Very likely* large increase in HD (Fischer and Schär, 2009; Giannakopoulos et al.,	A B G R	WN *very likely* to increase –*likely* largest changes in E Mediterranean (Sillmann	A B **G**	*Very likely* large increase in HWD (also increases in intensity and frequency) - *likely*	A B G **R**	About *as likely as not* increase in HP intensity in all seasons except summer over parts of the	A G **R**	CDD *very likely* to increase by a month or more, especially in S Iberian Peninsula,	A B **G** R

		2009; Fischer and Schär, 2010).		and Roeckner, 2008; Giannakopoulos et al., 2009).		largest increases in SW, S & E (Beniston et al., 2007; Diffenbaugh et al., 2007; Koffi and Koffi, 2008; Giannakopoulos et al., 2009; Fischer and Schär, 2010).		region, but decrease in some parts, e.g., Iberian Peninsula (Goubanova and Li, 2007; Giorgi and Lionello, 2008; Giannakopoulos et al., 2009).		E Adriatic and S Greece (Beniston et al., 2007; Sillmann and Roeckner, 2008; Giannakopoulos et al., 2009).	
Africa	All Africa										
	W. Africa										
	E. Africa										
	S. Africa										
	Sahara										
Central and South America	All South America										
	Central America and northern South America										
	Amazon	*Lack of evidence*		*Very likely* increase of warm nights (Tebaldi et al., 2006; Marengo et al., 2009a).	A **G** R	*Very likely* to increase (Tebaldi et al., 2006).	A **G**	*Insufficient evidence* (Tebaldi et al., 2006; Marengo et al., 2009a).	A **G** R	*Insufficient evidence* (Tebaldi et al., 2006; Marengo et al., 2009a).	A **G** R
	Northeastern Brazil	*Lack of evidence*		*Likely* increase of warm nights (Tebaldi et al., 2006; Marengo et al., 2009a).	A **G** R	*Likely* to increase (Tebaldi et al., 2006).	A **G**	*Insufficient evidence* (Tebaldi et al., 2006; Marengo et al., 2009a).	A **G** R	*Likely* to increase (Tebaldi et al., 2006; Marengo et al., 2009a).	A **G** R
	Southeastern South America	*Lack of evidence*		*Very likely* increase of warm nights (Tebaldi et al., 2006; Marengo et al., 2009a).	A **G** R	*Likely* to increase (Tebaldi et al., 2006).	A **G**	*Very likely* to increase (Tebaldi et al., 2006; Marengo et al., 2009a; Nunez et al., 2009).	A **G** R	*Likely* to increase (Tebaldi et al., 2006; Marengo et al., 2009a).	A **G** R
	Western Coast of South America	*Lack of evidence*		*Likely* increase of warm nights (Tebaldi et al., 2006; Marengo et al., 2009a).	A **G** R	*Very likely* to increase (Tebaldi et al., 2006).	A **G**	*Likely* to increase in the tropics and likely to decrease in the extratropics (Tebaldi et al., 2006; Marengo et al., 2009a).	A **G** R	*Insufficient evidence* (Tebaldi et al., 2006; Marengo et al., 2009a)	A **G** R

	All Asia										
Asia	N. Asia									General drought (EDI<-1) less frequent & shorter duration. (Kim and Byun, 2009)	A **G**
	Central Asia										
	East Asia	Tmax (95th percentile) increases (by up to 4-5°C) in Korea. HD increases (by up to 26 days) in Korea (Boo et al., 2006; Im and Kwon, 2007; Im et al., 2008; Koo et al., 2009; Im et al., 2010).	A B R	Tmin (5th percentile) increases (by up to 7-9°C) in Korea (Boo et al., 2006; Koo et al., 2009; Im et al., 2010).	A B R			HP & HPD increases in Korea (Boo et al., 2006; Im et al., 2010). HPD increase in Yangtze and Japan (Kimoto et al., 2005; Kusunoki and Mizuta, 2008). HP frequency (hourly & daily) increases (Kitoh et al., 2009). 50-year HP events increase in mid-lower Yangtze (Su et al., 2009).	A B G R	Extreme drought (EDI<-2) intensifies in the southern area & Asian monsoon region (Kim and Byun, 2009).	A **G**
	S.E. Asia									Extreme drought (EDI<-2) intensifies in the Asian monsoon region (Kim and Byun, 2009).	A **G**
	S. Asia	Tmax increases by 2°C in most areas of India (Kumar et al., 2006).	A R	Tmin increases by 5°C in India (Kumar et al., 2006).	A R			Increases in maximum 1-day & 5-day precipitation in India – especially in western Ghats & NW peninsular India (Wakazuki et al., 2008).	A R	Extreme drought (EDI<-2) intensifies in the Asian monsoon region (Kim and Byun, 2009).	A **G**
	W. Asia									Extreme drought (EDI<-2) more frequent & intensifies (especially in Syria & vicinity). (Kim and	A **G**

										Byun, 2009)	
	Tibetan Plateau										
Australia/New Zealand	Australia			WN increase everywhere. Largest increases in N (~60%) compared with S (~30%). Most consistent changes in inland regions (Alexander and Arblaster, 2009).	A **G**	HWD increases everywhere. Strongest increases in NW & most consistent increases inland (Alexander and Arblaster, 2009).	A **G**	HPD tend to increase in E & 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).	A **G**		

Notes:
Codes for projection period & emissions scenarios:
A: Projections for end of century (2071-2100 minus 1961-1990 or 2080-2099 minus 1980-1999) and A2 or A1B emissions scenarios.
B: Prior to 2050, any SRES.

Codes for downscaling method:
G: Based on GCM simulations. Bold: multi-GCM.
R: Based on RCM simulations. Bold: multi-GCM. Underlined: multi-RCM.

EDI: Effective Drought Index. Calculated from precipitation only – considers Effective Precipitation: the summed value of daily precipitation with a time-dependent reduction function.

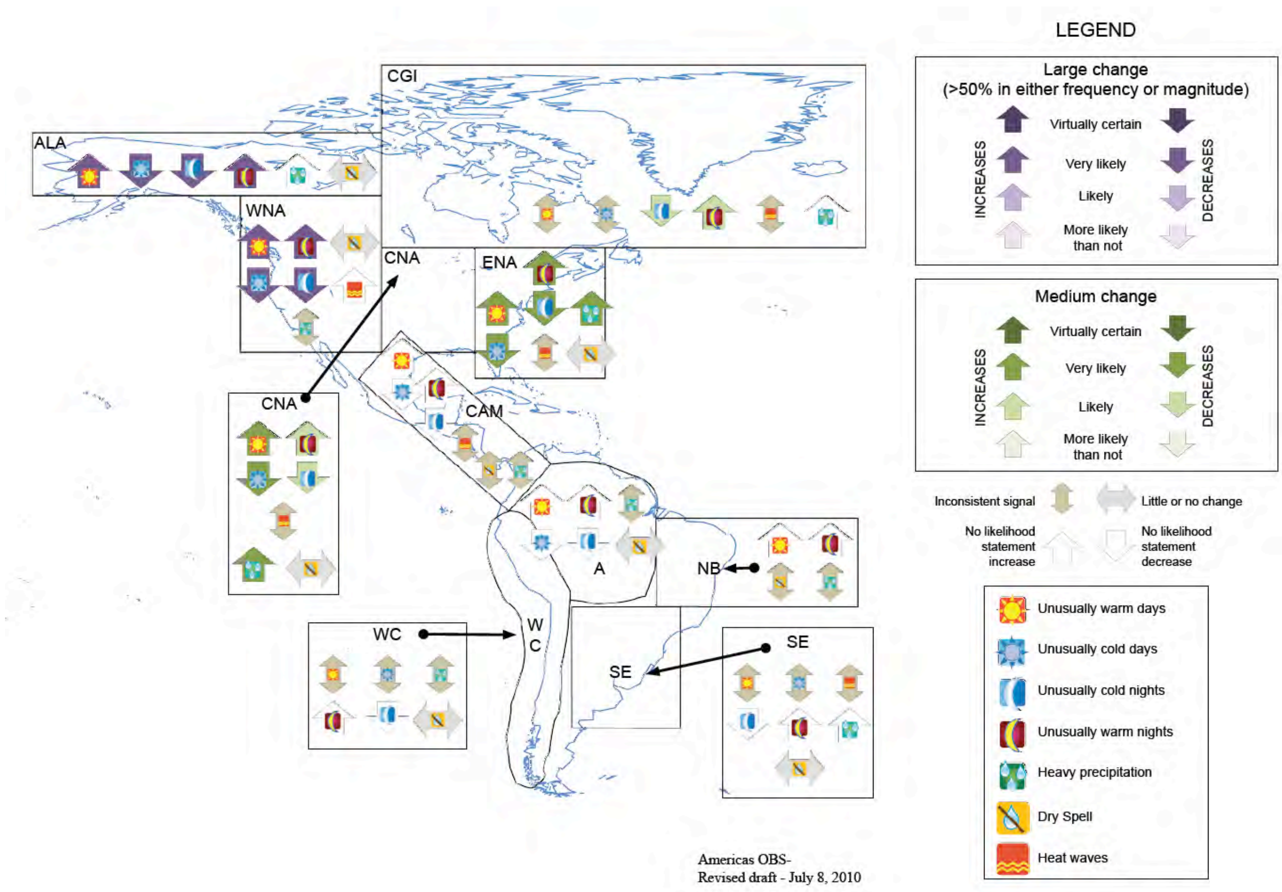


Figure 3.1: Regional observed changes in temperature and precipitation extremes (Americas)

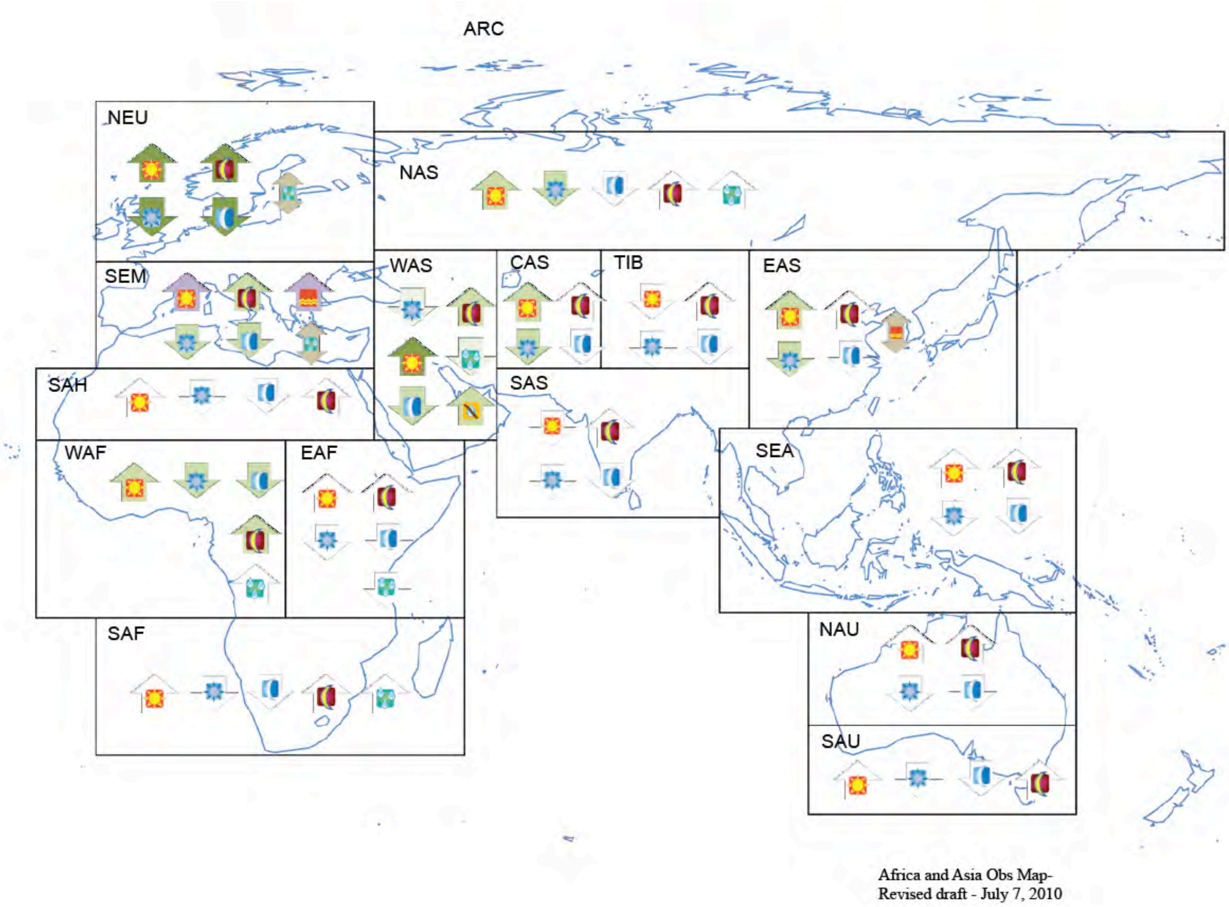


Figure 3.2: Regional observed changes in temperature and precipitation extremes (Europe, Africa, Asia and Oceania). See Figure 3.1 for definition of symbols

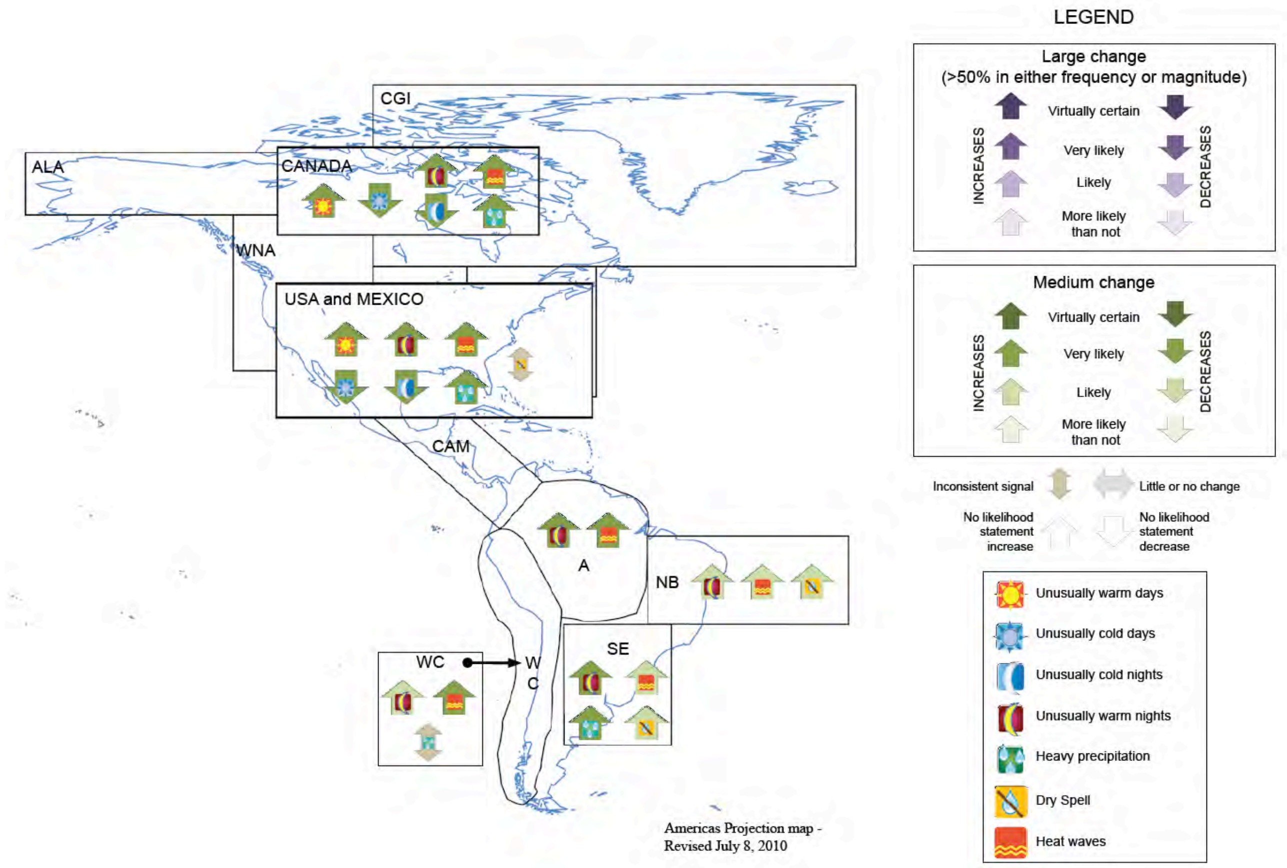


Figure 3.3: Regional projected changes in temperature and precipitation extremes (Americas)

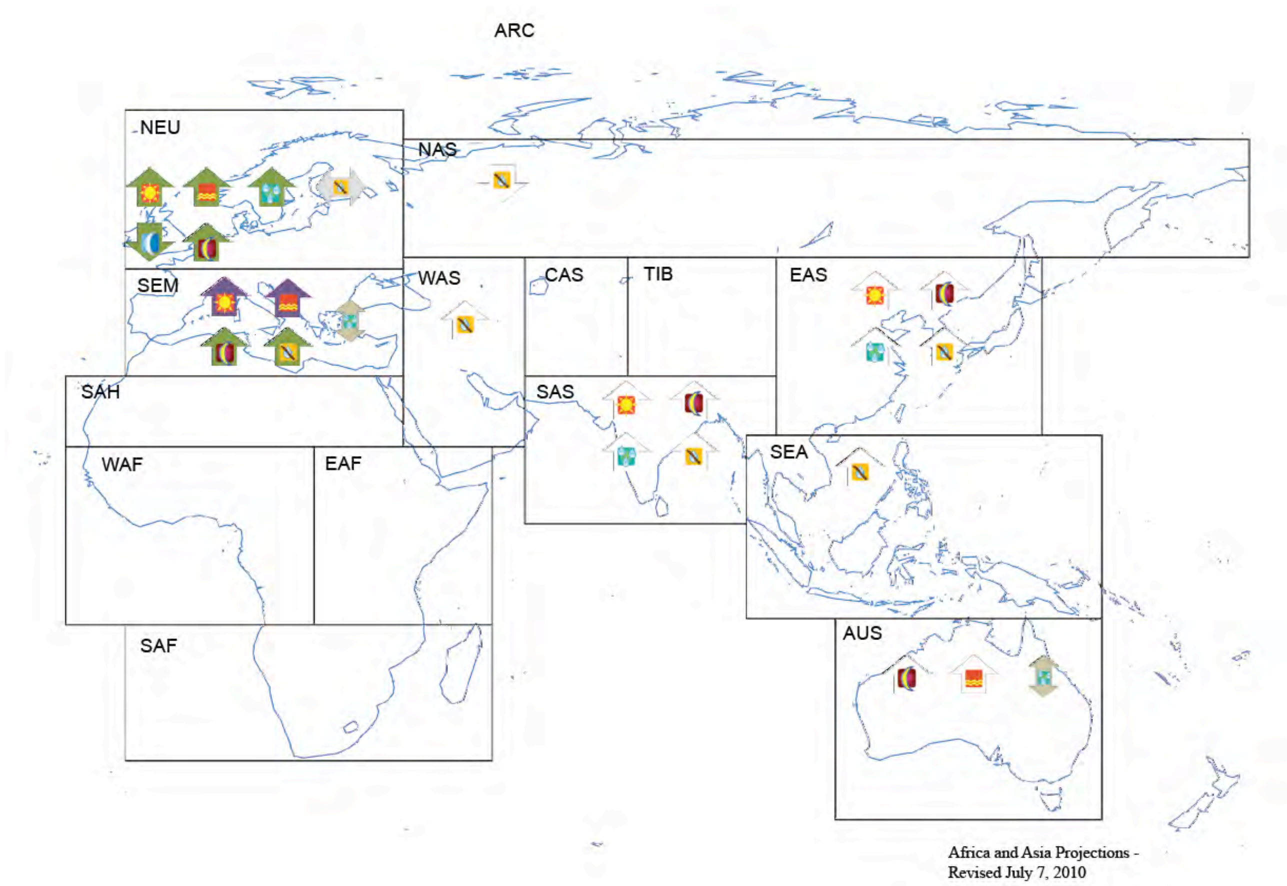


Figure 3.4: Regional projected changes in temperature and precipitation extremes (Europe, Africa, Asia, and Oceania). See Figure 3.3. for definition of symbols.

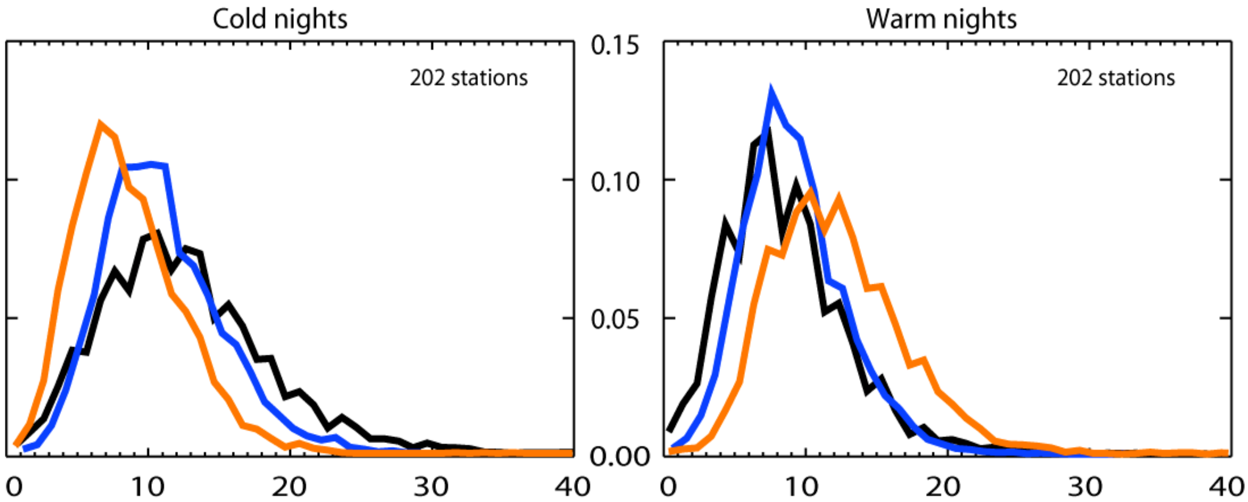


Figure 3.5: Annual PDFs for temperature indices for 202 global stations with at least 80% complete data between 1901-2003 for three time periods: 1901-1950 (black), 1951-1978 (blue), and 1979-2003 (red). The x-axis represents the percentage of time during the year when the indicators were below the 10th percentile for cold nights (left) and above the 90th percentile for warm nights (right). From Alexander et al., (2006).

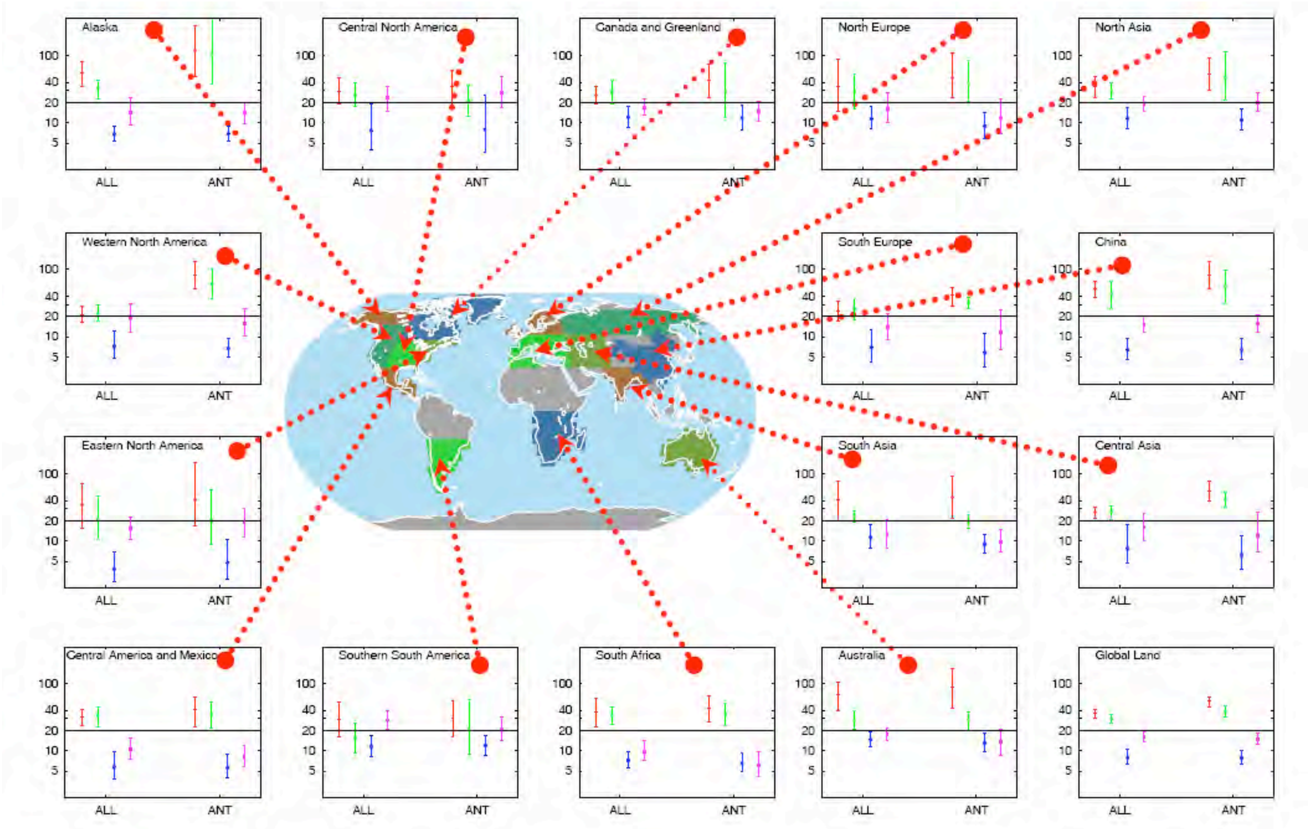


Figure 3.6: Estimated waiting time (years) and their 5% and 95% uncertainty limits for 1960s 20-yr return values of annual extreme daily temperatures in the 1990s climate (see text for more details). From Zwiers et al., (2010). Red, green, blue, pink error bars are for annual minimum daily minimum temperature (TNn), annual maximum daily minimum temperature (TNx), annual minimum daily maximum temperature (TXn), and annual maximum daily maximum temperature (TXx), respectively. Grey areas indicate insufficient data

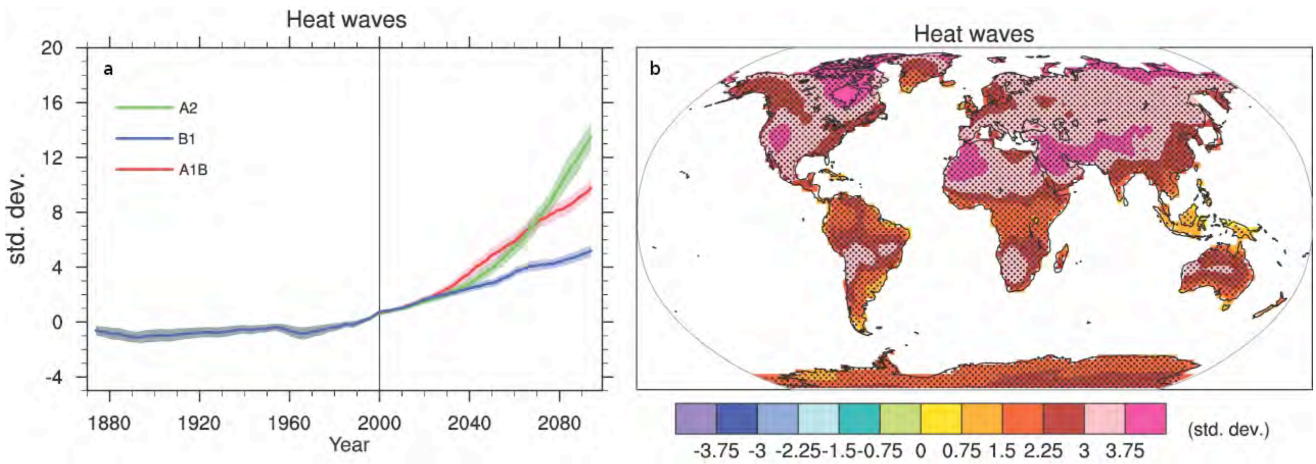


Figure 3.7: (a) Globally averaged changes in heat waves (defined as the longest period in the year of at least five consecutive days with maximum temperature at least 5°C higher than the 1961-1990 climatology of the same calendar day) based on multi-model simulations from nine global coupled climate models, adapted from Tebaldi et al., (2006). (b) Changes in spatial patterns of simulated heat waves between two 20-year means (2080–2099 minus 1980–1999) for the A1B scenario. Solid lines in (a) show the 10-year smoothed multi-model ensemble means; the envelope indicates the ensemble mean standard deviation. Stippling in (b) denotes areas where at least five of the nine models concur in determining that the change is statistically significant. Extreme indices are calculated only over land. Extremes indices are calculated following Frich et al., (2002). Each model's time series is centred around its 1980 to 1999 average and normalised (rescaled) by its standard deviation computed (after de-trending) over the period 1960 to 2099. The models are then aggregated into an ensemble average, both at the global and at the grid-box level. Thus, changes are given in units of standard deviations. From Meehl et al., (2007b).

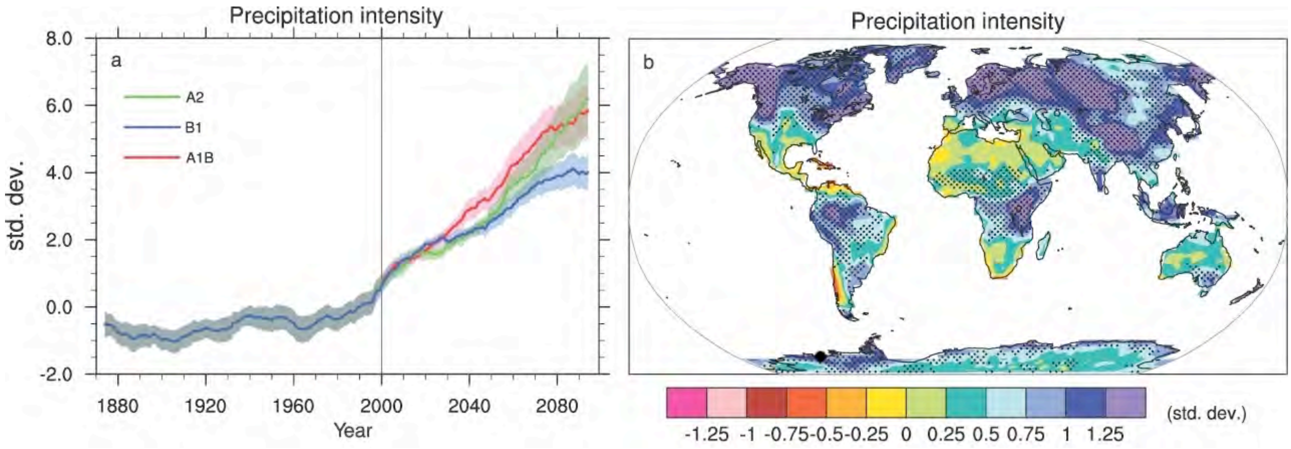


Figure 3.8: Changes in extremes based on multi-model simulations from nine global coupled climate models, adapted from Tebaldi et al., (2006). (a) Globally averaged changes in precipitation intensity (defined as the annual total precipitation divided by the number of wet days) for a low (SRES B1), middle (SRES A1B) and high (SRES A2) scenario. (b) Changes in spatial patterns of simulated precipitation intensity between two 20-year means (2080–2099 minus 1980–1999) for the A1B scenario. Solid lines in (a) are the 10-year smoothed multi-model ensemble means; the envelope indicates the ensemble mean standard deviation. Stippling in (b) denotes areas where at least five of the nine models concur in determining that the change is statistically significant. Extreme indices are calculated only over land following Frich et al., (2002). Each model's time series was centred on its 1980 to 1999 average and normalised (rescaled) by its standard deviation computed (after de-trending) over the period 1960 to 2099. The models were then aggregated into an ensemble average, both at the global and at the grid-box level. Thus, changes are given in units of standard deviations. From Meehl et al., (2007b).

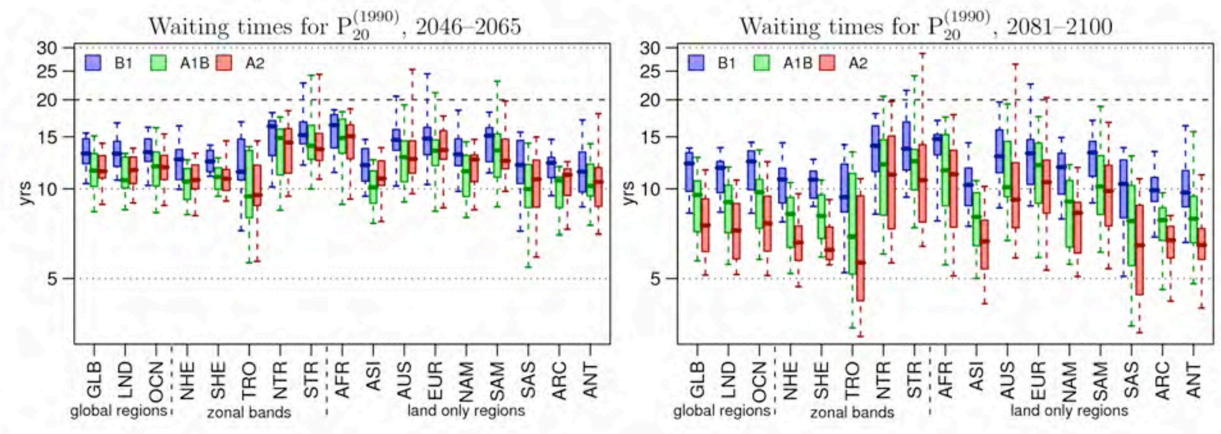


Figure 3.9: Projected waiting times for late-twentieth-century 20-year return values of annual maximum 24-hour precipitation rates in the mid-21st century (left) and in late-21st century (right) by 14 GCMs that contributed to the IPCC AR4, under three different emission scenarios SRES B1, A1B and A2 (adapted from Kharin et al., 2007). The vertical extent of the whiskers in both directions describes the range of projected changes by all 14 climate models used in the study. The boxes indicate the central 50% of model projected changes, and the horizontal bar in the middle of the box indicates the median projection amongst the 14 models (that is, 7 models project waiting times longer than the median and 7 models project waiting times shorter than the median). Although the uncertainty range of the projected change in extreme precipitation is large, almost all models suggest that the waiting time for a late 20th century 20-year extreme 24-hour precipitation event will be reduced to substantially less than 20 years by mid-21st and much more by late-21st century, indicating an increase in frequency of the extreme precipitation at continental and sub-continental scales under all three forcing scenarios. Three global domains are: the entire globe (GLB), the global land areas (LND), the global ocean areas (OCN). Five zonal bands are: Northern Hemisphere Extratropics (NHE, 35°-90°N), Southern Hemisphere Extratropics (SHE, 35°-90°S), Tropics (TRO, 10°S-10°N), Northern subtropics (NTR, 10°-35°N), and Southern subtropics (35°-10°S). The nine continental/sub-continental land-only regions are: Africa (AFR, 20°W-60°E and 40°S-30°N), Central Asia (ASI, 45°-180°E and 30°-65°N), Australia (AUS, 105°E–180° and 45°–10°S), Europe (EUR, 20°W–45°E and 30°–65°N), North America (NAM, 165°–30°W and 25°–65°N), South America (SAM, 115°–30°W and 55°S–25°N), South Asia (SAS, 60°–160°E and 10°S–30°N), Arctic (ARC, 180° to 180° and 65°–90°N), Antarctica (ANT, 180° to 180° and 90°–65°S).

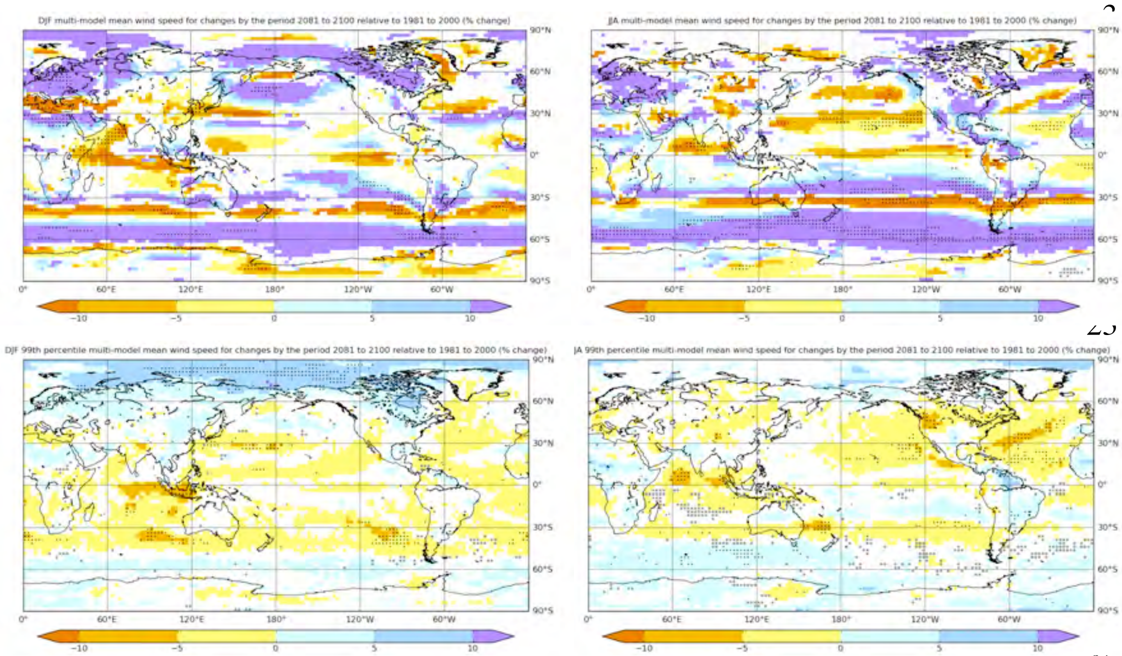


Figure 3.10: The average of the multi-model 10 m mean wind speeds (top) and 99th percentile daily wind speeds (bottom) for the period 2080 to 2099 relative to 1980 to 1999 (% change) for December to February (left) and June to August (right) plotted only where more than 66% of the models agree on the sign of the change. Fine black stippling indicates where more than 90% of the models agree on the sign of the change and bold grey stippling (in white or light coloured areas) indicates where 66% of models agree on a small change between ±2 %. From McInnes et al., (2010).

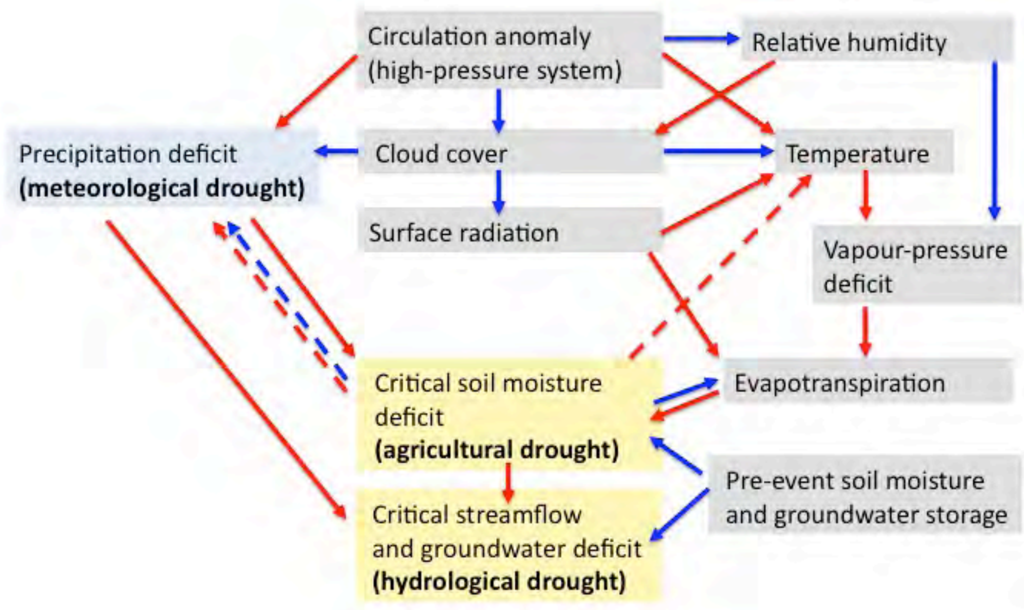


Figure 3.11: Processes and interactions involved in meteorological, agricultural, and hydrological droughts (red: positive impacts; blue: negative impacts). Dashed lines denote indirect feedbacks of soil moisture on temperature and precipitation. For simplicity, the role of interactions with other variables of the Figure (e.g., evapotranspiration, relative humidity) in these feedbacks, and feedbacks of soil moisture to other meteorological variables (e.g., circulation anomalies) are not highlighted.

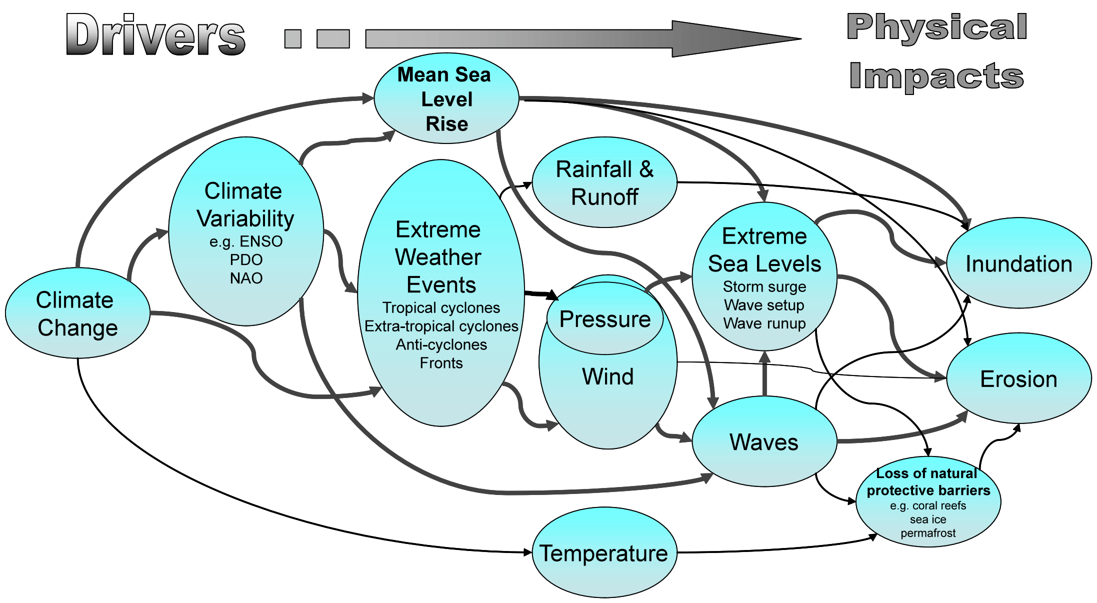
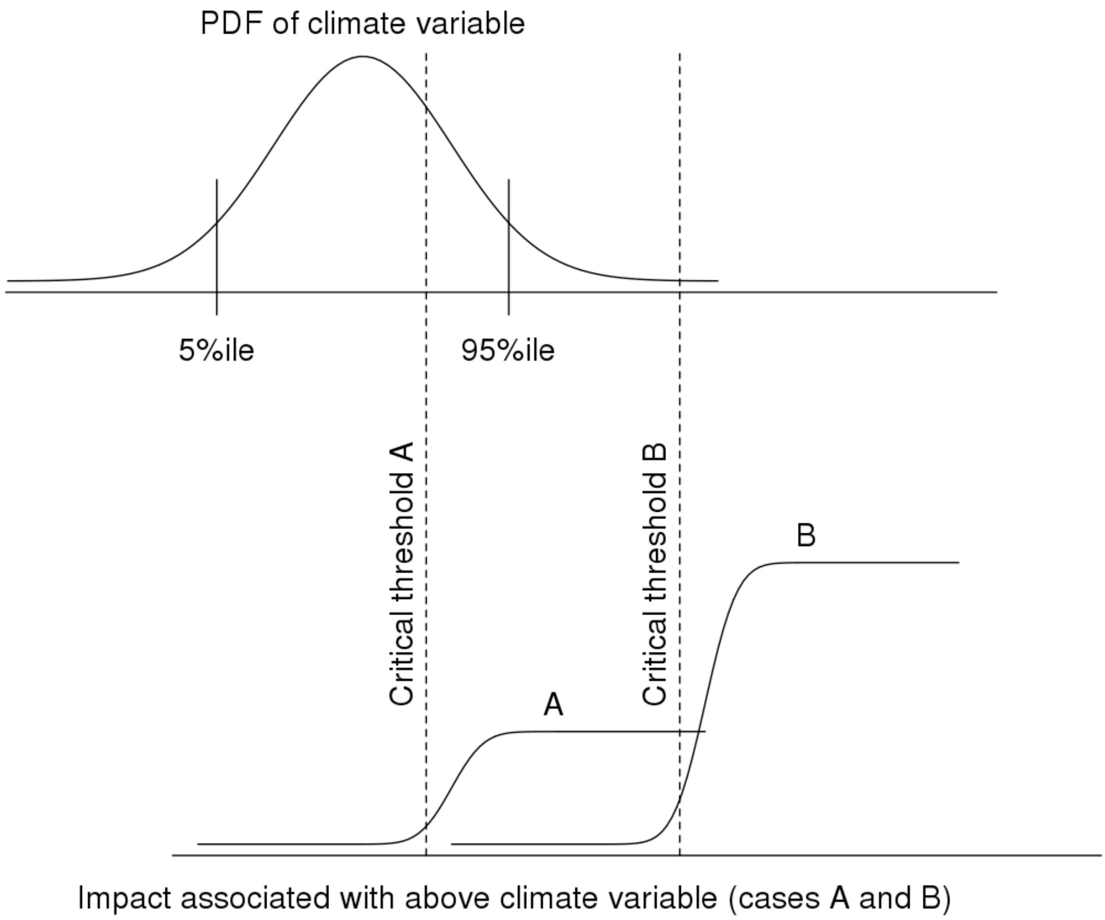
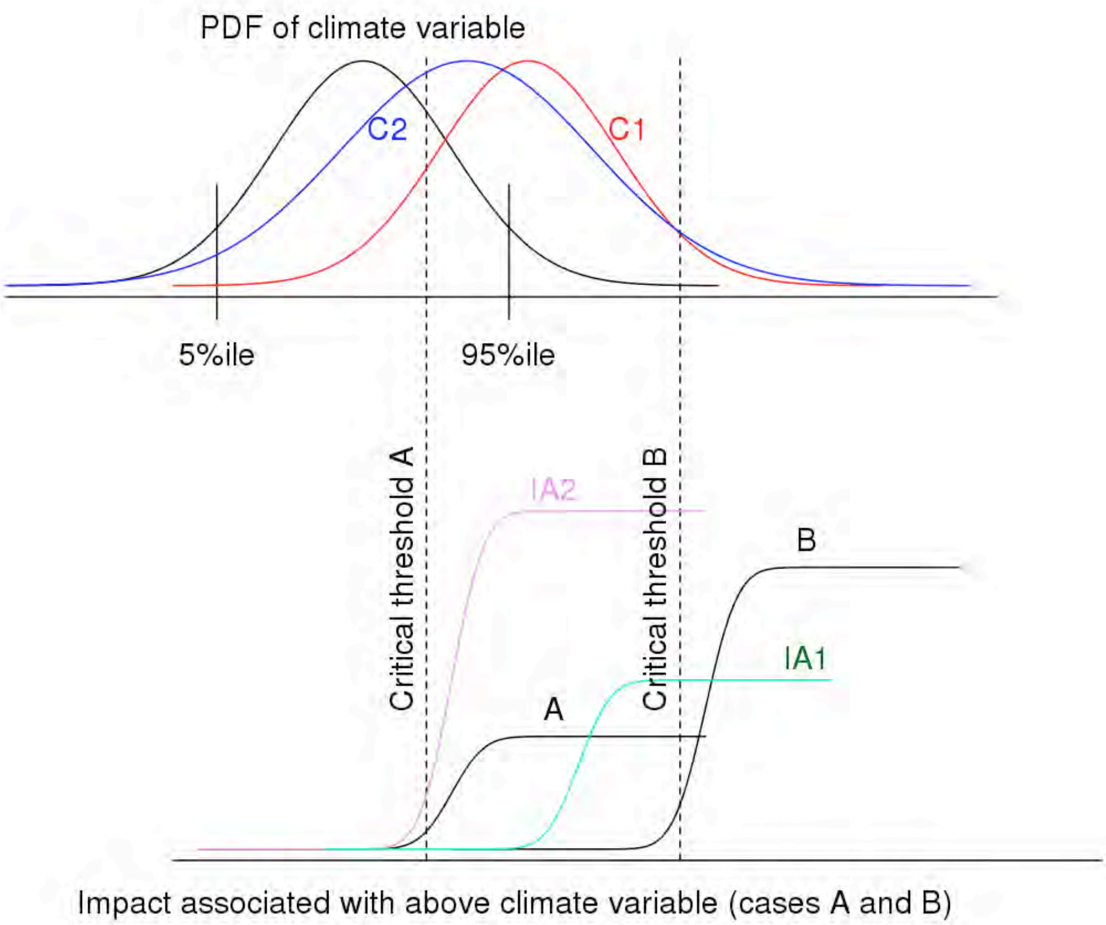


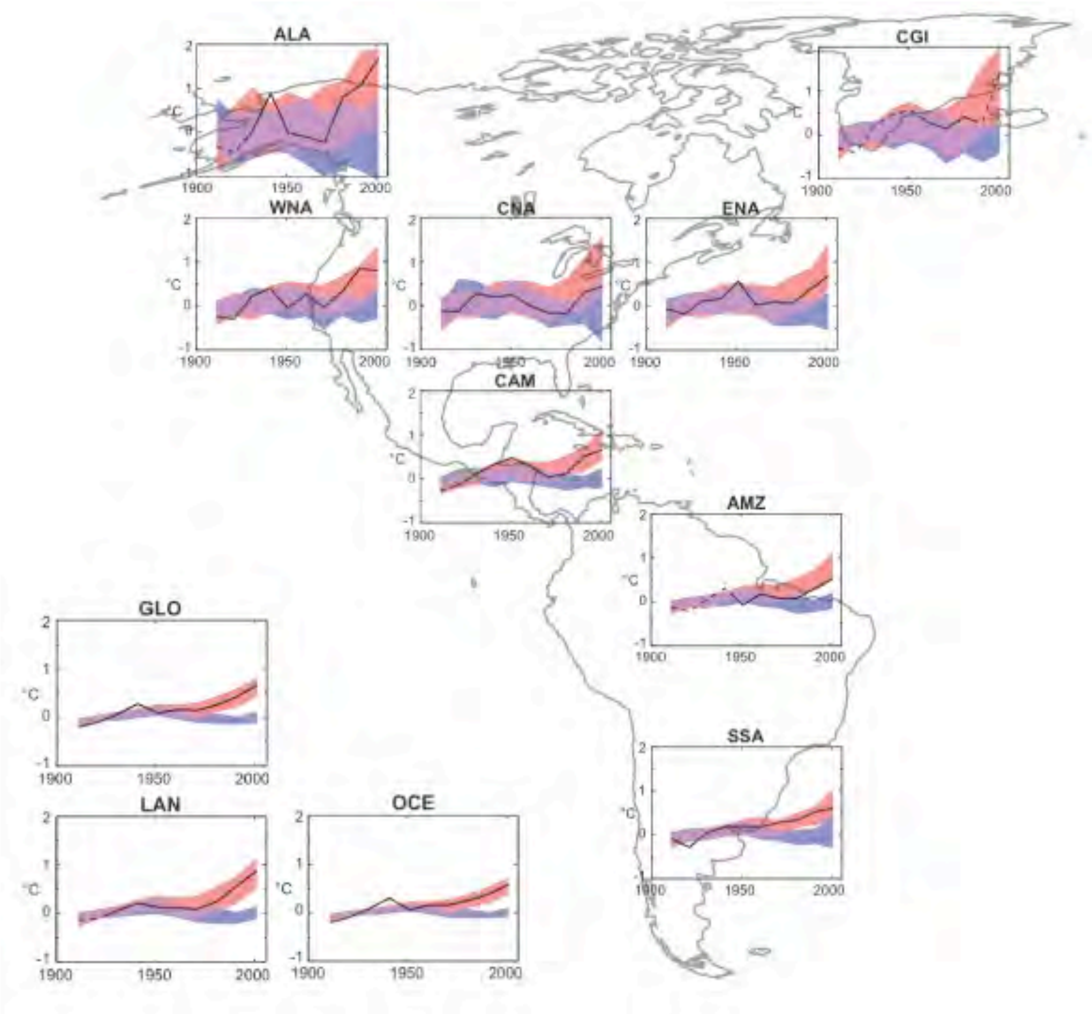
Figure 3.12: Relationships between climate, weather phenomena and physical impacts in the coastal zone.



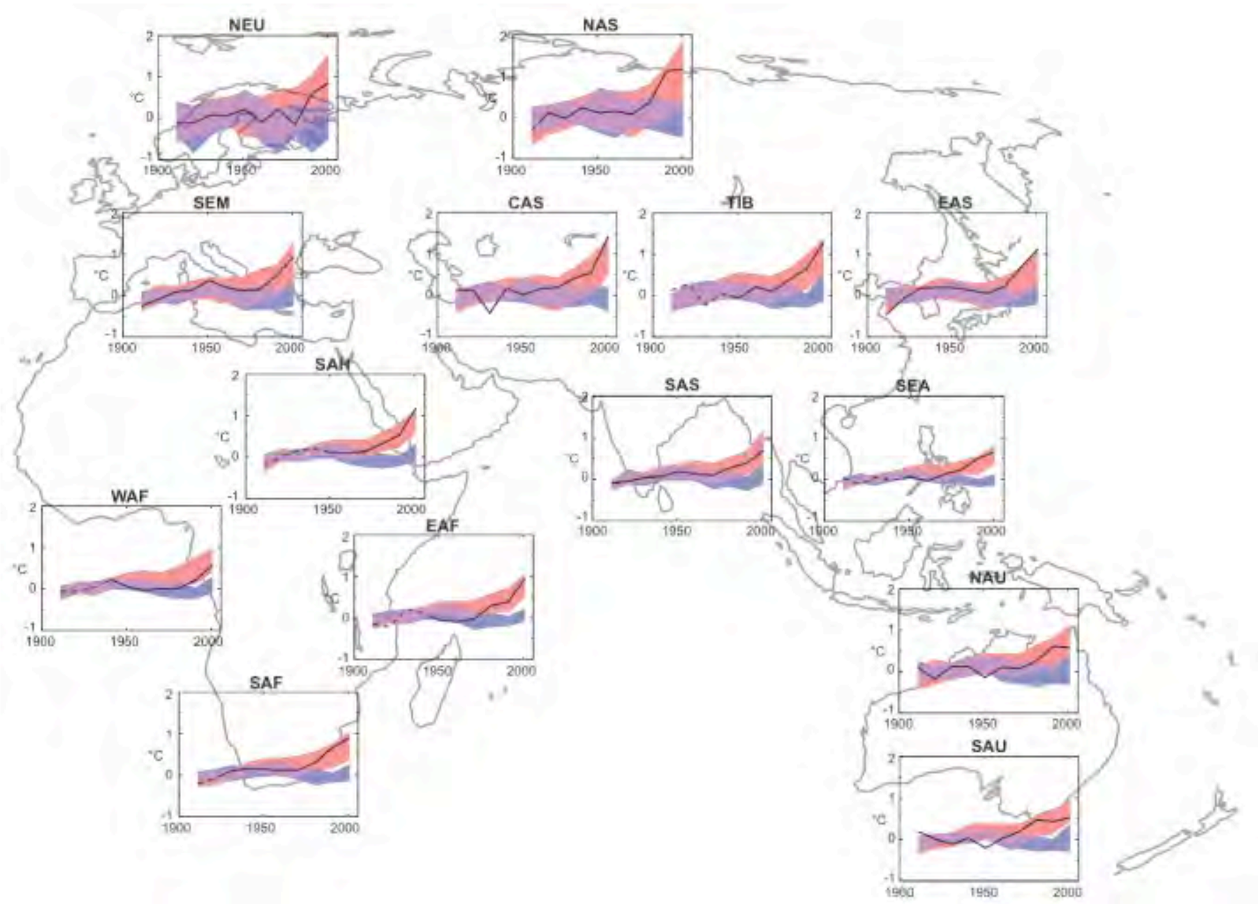
Box 3.1, Figure 1: Link between climate/weather variable probability distribution function (PDF) and associated impacts (A and B), and implication for definition of "climate extremes" (see discussion in text). Note that the PDF of a climate variable is not necessarily Gaussian.



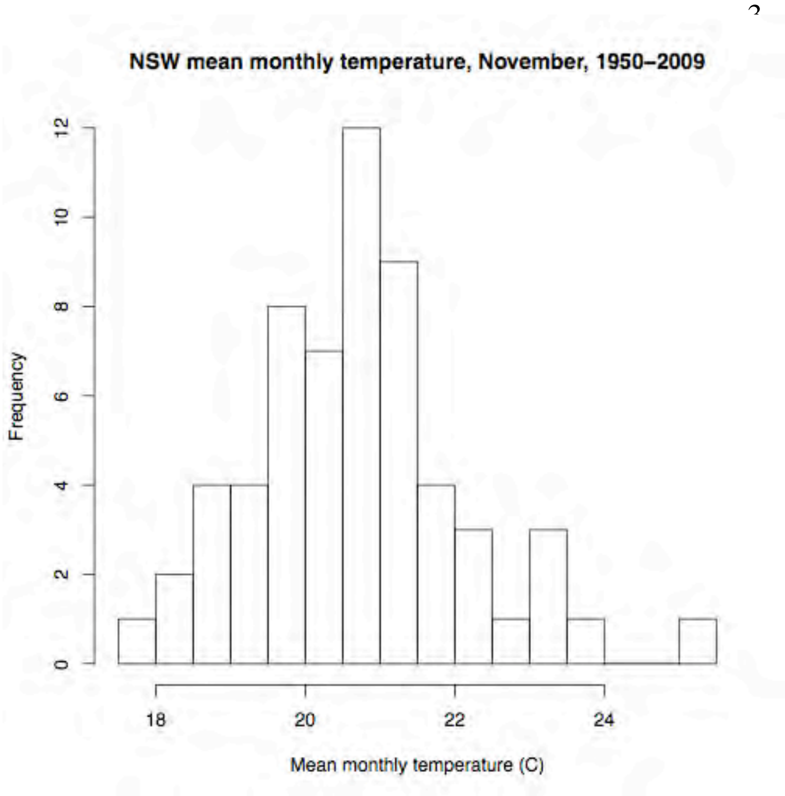
Box 3.1, Figure 2: Link between climate/weather variable PDF and associated impacts under climate change (see discussion in text). Note that the PDF of a climate variable is not necessarily Gaussian.



Box 3.3, Figure 1: Comparison for the Americas of multi-model data set of model simulations containing all forcings (red shaded regions) and containing natural forcings only (blue shaded regions) with observed decadal mean temperature changes (°C) from 1906 to 2005 from the Hadley Centre/Climatic Research Unit gridded surface temperature data set (HadCRUT3, Brohan et al., 2006). The panel labelled GLO shows comparison for global mean; LAN, global land; and OCE, global ocean data. Remaining panels display results for 22 sub-continental scale regions. Shaded bands represent the middle 90% range estimated from the multi-model ensemble. Note that the model simulations have not been scaled in any way. The same simulations are used as in Figure 9.5 of AR4 (58 simulations using all forcings from 14 models, and 19 simulations using natural forcings only from 5 models) (Hegerl et al., 2007). Each simulation was sampled so that coverage corresponds to that of the observations, and was centred relative to the 1901 to 1950 mean obtained by that simulation in the region of interest. Observations in each region were centred relative to the same period. The observations in each region are generally consistent with model simulations that include anthropogenic and natural forcings, whereas in many regions the observations are inconsistent with model simulations that include natural forcings only. Lines are dashed where spatial coverage is less than 50%. From Hegerl et al., (2007).



Box 3.3, Figure 2: Same as Box 3.3, Figure 1 for Europe, Africa, Asia and Oceania. From Hegerl et al., (2007).



FAQ 3.2, Figure 1: The distribution of monthly mean November temperatures averaged across the State of New South Wales in Australia, using data from 1950–2009. Data from Australian Bureau of Meteorology. The mean temperature for November 2009 (the bar on the far right hand end of the Figure) was more than three standard deviations from the long-term mean (calculated from 1950–2008 data).