U.S. Climate Change Science Program

Synthesis and Assessment Product 3.3

Weather and Climate Extremes in a Changing Climate

Regions of Focus: North America, Hawaii, Caribbean, and U.S. Pacific Islands

Lead Agency:
National Oceanic and Atmospheric Administration

Contributing Agencies:
Department of Energy
National Aeronautics and Space Administration
U.S. Geological Survey

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Abstract

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Changes in extreme weather and climate events are among the most serious challenges to society in coping with global warming. While some extremes, such as snowstorms and ice storms in the U.S., have not generally been observed to change, many others have, and are already causing impacts. For example, most of North America is experiencing more unusually hot days and nights and fewer unusually cold days and nights. There has been a decrease in frost days, particularly in the western part of North America. Extreme precipitation episodes (heavy downpours) have become more frequent and intense. All these trends are expected to continue or accelerate.

There are recent regional tendencies toward more severe droughts in the southwestern U.S., and parts of Canada, Alaska, and Mexico. In the future, droughts are likely to become more frequent and severe in some regions, leading to an increased need to respond to reduced water supplies, increased wildfires, and various ecological impacts.

Atlantic tropical storm and hurricane destructive potential has increased substantially since about 1970. Although there have been fluctuations from decade to decade, the

balance of evidence suggests that human activity has caused a discernible increase in tropical storm/hurricane and major hurricane frequency in the North Atlantic over the past century. U.S. land-falling hurricane frequency has not increased since the late 1800s. Hurricane intensity in the eastern Pacific has decreased since 1980, but rainfall from near-coastal hurricanes has increased since 1949. Hurricane rainfall rates, wind speeds, and the potential for storm surge damage are projected to increase.

For cold-season storms, in the North Pacific, storm tracks are shifting northward and the strongest storms are becoming even stronger, with increases in extreme wave heights along the Pacific Northwest coast. In the future, there are likely to be more frequent strong cold-season storms in both the Atlantic and Pacific basins, with stronger winds and more extreme wave heights. Observations and models used to determine changes in tornadoes and severe thunderstorms are inadequate to make definitive statements about past and projected changes.

Current and future impacts resulting from these changes depend not only on the changes in extremes, but also on responses by human and natural systems.

Preface

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According to the National Research Council, "an essential component of any research program is the periodic synthesis of cumulative knowledge and the evaluation of the implications of that knowledge for scientific research and policy formulation." The U.S. Climate Change Science Program (CCSP) is helping to meet that fundamental need through a series of 21 "synthesis and assessment products" (SAPs). A key component of the CCSP Strategic Plan (released July 2003), the S&A products integrate research results focused on important science issues and questions frequently raised by decision makers.

The SAPs support informed discussion and decisions by policymakers, resource managers, stakeholders, the media, and the general public. They are also used to help define and set the future direction and priorities of the program. The products help meet the requirements of the Global Change Research Act of 1990. The law directs agencies to "produce information readily usable by policymakers attempting to formulate effective strategies for preventing, mitigating, and adapting to the effects of global change" and to undertake periodic scientific assessments. This SAP (3.3) provides an in-depth assessment of the state of our knowledge about changes in weather and climate extremes in North America (and U.S. territories), where we live, work, and grow much of our food.

The impact of weather and climate extremes can be severe and wide-ranging although, in some cases, the impact can also be beneficial. Weather and climate extremes affect all sectors of the economy and the environment, including human health and well-being. During the period 1980-2006, the U.S. experienced 70 weather-related disasters in which overall damages exceeded $1 billion at the time of the event. Clearly, the direct impact of extreme weather and climate events on the U.S. economy is substantial.

There is scientific evidence that a warming world will be accompanied by changes in the intensity, duration, frequency, and spatial extent of weather and climate extremes. The Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report has evaluated extreme weather and climate events on a global basis in the context of observed and projected changes in climate. However, prior to SAP 3.3 there has not been a specific assessment of observed and projected changes in weather and climate extremes across North America (including the U.S. territories in the Caribbean Sea and the Pacific Ocean), where observing systems are among the best in the world and the extremes of weather and climate are some of the most notable occurring across the globe.

The term “weather extremes”, as used in SAP 3.3, signifies individual weather events that are unusual in their occurrence (minimally, the event must lie in the upper or lower ten percentile of the distribution) or have destructive potential, such as hurricanes and tornadoes. The term “climate extremes” is used to represent the same type of event, but viewed over seasons (e.g., droughts), or longer periods. In this assessment we are particularly interested in whether climate extremes are changing in terms of a variety of characteristics, including intensity, duration, frequency, or spatial extent, and how they are likely to evolve in the future although, due to data limitations and the scarcity of published analyses, there is little that can be said about extreme events in Hawaii, the Caribbean or the Pacific Islands outside of discussion of tropical cyclone intensity and frequency**.** It is often very difficult to attribute a particular climate or weather extreme, such as a single drought episode or a single severe hurricane, to a specific cause. It is more feasible to attribute the changing "risk" of extreme events to specific causes. For this reason, this assessment focuses on the possible changes of past and future statistics of weather and climate extremes.

In doing any assessment, it is helpful to precisely convey the degree of certainty of various findings and projections. For this reason, a lexicon expressing the likelihood of each key finding is presented below and used throughout this report. There are numerous choices for categories of likelihood and appropriate wording to define these categories. CCSP SAP 5.2, currently under review, but scheduled for public release before SAP 3.3 is finalized, provides useful guidance. Additionally, the community of scientists and policy-makers familiar with the IPCC assessments provide yet another consistent way to express likelihood statements. The US National Assessment published in 2000 provides a similar approach to likelihood statements. Figure Preface.1 provides the common terms used in this report to express likelihood of occurrence based on experience with previous likelihood statements. Because of the nature of the topic being considered, our likelihood statements do not have discrete boundaries, unlike in IPCC, but characterize likelihood in more general terms (Figure Preface.1). In some cases, there is sufficiently strong evidence to draw a conclusion, but not enough to allow a determination of 'likely'. In these few cases, the term 'the balance of evidence' is used to express our assessment of the state of the science. Statements made without likelihood qualifiers are intended to indicate a high degree of certainty.

Dedication

This Climate Change and Synthesis Product is dedicated to the memory of our colleague, friend and co-author Dr. Miguel Cortez whose untimely passing during the writing of the report was a loss to us all.

Figure Preface.1 Language in this Synthesis and Assessment Product used to express the team's expert judgment of likelihood.

Executive Summary

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ES.1 What Are Extremes and Why Do They Matter

Weather and climate extremes (Figure ES.1) have always posed serious challenges to society. Changes in extremes are already observed to be having impacts on socioeconomic and natural systems, and future changes associated with continued warming will present additional challenges. Increased frequency of heat waves and drought, for example, could seriously affect human health, agricultural production, water resources, and water quality (chapter 1, section 1.1).

Extremes are a natural part of even a stable climate system and have associated costs (Figure ES.2) and benefits. For example, extremes are essential in some systems to keep insect pests under control. While hurricanes cause significant death, injury, damage, and disruption, they also provide needed rainfall to certain areas, and some tropical plant communities are dependent on hurricane winds toppling tall trees, allowing more sunlight to rejuvenate low-growing trees. But on balance, because systems have adapted to their historical range of extremes, the majority of the impacts of events outside this range are negative impacts (chapter 1, section 1.4 and 1.5).

The impacts of changes in extremes depend on both changes in climate and ecosystem and societal vulnerability. Vulnerability is shaped by factors such as population dynamics and economic status as well as developing and utilizing adaptation measures such as appropriate building codes, disaster preparedness, and water use efficiency. Some actions taken to lessen the risk from extreme events can lead to increases in vulnerability to even larger extremes. For example, moderate flood control measures on a river can stimulate development in a now "safe" floodplain, only to see those new structures damaged when a very large flood occurs (chapter 1, section 1.6).

Human activities are known to affect climate averages. This is relevant to extremes because small changes in the averages of many variables result in larger changes in their extremes. Thus, within a changing climate system, what are now considered to be extreme events will occur more frequently (Figures ES.3, ES.4). More frequent extreme events occurring over a shorter period reduce the time available for recovery and adaptation. In addition, extreme events often occur in clusters. The cumulative effect of compound or back-to-back extremes has far larger impacts than the same events spread out over a longer period of time. For example, heat waves, droughts, air stagnation, and resulting wildfires often occur concurrently and have more severe impacts than any of these alone (chapter 1, section 1.2).

ES.2 Temperature–related Extremes

ES.2.1 Observed Changes

Since 1998, over half of the U.S. annual average temperatures have been extremely high, including the hottest two years on record (1998 and 2006). Accompanying a general rise in the average temperature, most of North America is experiencing more unusually hot days and nights. The number of heat waves (extended periods of extremely hot weather) also has been increasing since 1950. However, the heat waves of the 1930s remain the most severe in the U.S. historical record (chapter 2, section 2.2.1).

There have been fewer unusually cold days during the last few decades. The last 10 years have seen fewer severe cold waves than any other 10-year period in the historical record, which dates back to 1895. There has been a decrease in frost days and a lengthening of the frost-free season over the past century (chapter 2, section 2.2.1).

In summary, there is a shift towards a warmer climate with an increase in extreme high temperatures and a reduction in extreme low temperatures. These changes have been especially apparent in the western half of North America (chapter 2, section 2.2.1).

ES.2.2 Attribution of Changes

Human-induced warming has likely caused much of the average temperature increase in North America over the past fifty years and, consequently, changes in temperature extremes. The very hot year of 2006 has now been scientifically attributed primarily to human influences (chapter 3, section 3.2.1 and 3.2.2).

ES.2.3 Projected Changes

Future changes in extreme temperatures will generally follow changes in average temperature. 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. The number of days with frost is very likely to decrease (chapter 3, section 3.3.1 and 3.3.2).

Climate models indicate that currently rare extreme events will become more commonplace. For example, for a mid-range emissions scenario, a day so hot that it is currently experienced 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 (chapter 3, section 3.3.1).

Episodes of what are now considered to be unusually high sea-surface temperature are very likely to become more frequent and widespread. Sustained periods (e.g., months) of unusually high temperatures could lead, for example, to more coral bleaching and death of the corals (chapter 3, section 3.3.1).

Sea ice extent is expected to continue to decrease and may even disappear entirely in the Arctic Ocean in summer in the coming decades. This increases extreme coastal erosion in Arctic Alaska and Canada due to the increased exposure of the coastline to strong wave action (chapter 3, section 3.3.1 and 3.3.10).

ES.3 Precipitation Extremes

ES.3.1 Observed Changes

Extreme precipitation episodes (heavy downpours) have become more frequent and more intense in recent decades over most of North America and now account for a larger percentage of total precipitation. For example, intense precipitation (the heaviest 1%) in the continental U.S. increased by 20% over the past century while total precipitation increased by 7% (chapter 2, section 2.2.2.2).

The monsoon season is beginning about 10 days later than usual in Mexico. In general, for the summer monsoon in southwestern North America, there are fewer rain events, but the events are more intense (chapter 2, section 2.2.2.3).

ES.3.2 Attribution of Changes

Heavy precipitation events averaged over North America have increased over the past 50 years, consistent with the increased water holding capacity of the atmosphere in a warmer climate and the observed increase in water vapor over the oceans (chapter 3, section 3.2.3).

ES.3.3 Projected Changes

On average, precipitation is likely to be less frequent but more intense, and precipitation extremes are very likely to increase. For example, for a mid-range emission scenario, daily precipitation so heavy that it now occurs only once every 20 years is projected by climate models to occur every eight years or so by the end of this century over much of Eastern North America (chapter 3, section 3.3.5).

ES.4 Drought

ES.4.1 Observed Changes

Drought can be defined in many ways, from acute short-term to chronic long-term hydrological drought, agricultural drought, meteorological drought, and so on. The assessment in this report focuses primarily on drought as measured by the Palmer Drought Severity Index (PDSI), which best represents multi-seasonal aspects of drought. Some other indices are included where and when available especially because of known deficiencies in the PDSI (Box 2.1).

Averaged over the continental U.S. and southern Canada the most severe droughts occurred in the 1930s and there is no indication of an overall trend in the observational record, which dates back to 1895. In Mexico and the U.S. Southwest, the 1950s were the driest period, though droughts in the past 10 years now rival the 1950s drought. There are also recent regional tendencies toward more severe droughts in parts of Canada and Alaska (chapter 2, section 2.2.2.1).

ES.4.2 Attribution of Changes

No formal attribution studies for greenhouse warming and changes in drought severity in North America have been attempted. However, it is likely that the increasing temperatures (and associated increasing evaporation potential over land) are already contributing to droughts that are longer and more intense. The location and severity of droughts are also affected by the spatial pattern of sea surface temperatures, which appear to have been a factor in the severe droughts of the 1930s and 1950s (chapter 3, section 3.2.3).

ES.4.3 Projected Changes

A contributing factor to droughts becoming more frequent and severe is higher air temperatures increasing the potential for evaporation. It is likely that droughts will become more severe in the southwestern U.S. and parts of Mexico. In other places where precipitation increases cannot keep pace with increased evaporation, droughts are also likely to become more severe (chapter 3, section 3.3.7).

It is likely that droughts will continue to be exacerbated by earlier and possibly lower spring snowmelt run-off in the mountainous West, which results in less water available in late summer (chapter 3, section 3.3.4 and 3.3.7).

In southwestern North America, the precipitation in the winter rainy season is projected to continue to decrease, increasing the risk of drought (chapter 3, section 3.3.7).

ES.5 Storms

ES.5.1 Hurricanes and Tropical Storms

ES.5.1.1 Observed Changes

Atlantic tropical storm and hurricane destructive potential as measured by the Power Dissipation Index (which combines storm intensity, duration, and frequency) has increased. This increase is substantial since about 1970, and is likely substantial since the 1950s and 60s, in association with warming Atlantic sea surface temperatures (Figure ES.5) (chapter 2, section 2.2.3.1).

There have been fluctuations from decade to decade, and data uncertainty is larger in the early part of the record compared to the satellite era beginning in 1965. Taking these into account, 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 sea surface temperatures also increased. The evidence is less compelling for trends beginning in the late 1800s because uncertainties due to data issues preclude any definitive conclusions prior to 1900. Despite the likely increase in basin-wide hurricane counts over the past 100 years, there is no observational evidence for an increase in North American mainland land-falling hurricanes since the late 1800s (Chapter 2, section 2.2.3.1).

The hurricane Power Dissipation Index in the eastern Pacific, affecting the Mexican west coast and shipping lanes, has decreased since 1980. However, coastal station observations show that rainfall from hurricanes has increased since 1949, in part due to slower moving storms (chapter 2, section 2.2.3.1)

ES.5.1.2 Attribution of Changes

It is likely that human activities have caused a discernible increase in sea surface temperatures in the hurricane formation region of the tropical Atlantic Ocean over the past 100 years. The balance of evidence suggests that human activity has caused a discernable increase in tropical storm/hurricane and major hurricane frequency in the North Atlantic (chapter 3, section 3.2.4.3).

ES.5.1.3 Projected Changes

For North Atlantic and North Pacific hurricanes (both basin-wide and land-falling) it is likely that for each one degree Celsius increase in tropical sea surface temperatures core rainfall rates will increase by 6 to 18% and the surface wind speeds of the strongest hurricanes will increase by about 2 to 10% (chapter 3, section 3.3.9.2 and 3.3.9.4).

ES.5.2 Other Storms

ES.5.2.1 Observed Changes

It is very likely that there has been a northward shift in the tracks of strong low-pressure systems (storms) in both the N. Atlantic and N. Pacific over the past fifty years. In the North Pacific, the strongest storms are becoming even stronger. Evidence in the Atlantic is insufficient to draw a conclusion about changes in storm strength (Chapter 2, section 2.2.3.2).

Increases in extreme wave heights have been observed along the Pacific Northwest coast of North America based on three decades of buoy data, and are likely a reflection of changes in cold season storm tracks (chapter 2, section 2.2.3.3).

While snow cover extent has decreased over North America, overall trends in snowstorms and episodes of freezing rain have not been observed over the past century (chapter 2, section 2.2.3.4).

The data used to examine changes in the frequency and severity of tornadoes and severe thunderstorms are inadequate to make definitive statements about actual changes (chapter 2, section 2.2.3.5).

ES.5.2.2 Attribution of Changes

Human influences on changes in sea-level pressure patterns have been detected over the Northern Hemisphere and this affects the location and intensity of storms (chapter 3, section 3.2.5).

ES.5.2.3 Projected Changes

There are likely to be more frequent deep low-pressure systems (strong storms) outside the tropics, with stronger winds and more extreme wave heights (Figure ES.6) (chapter 3, section 3.3.10).

ES.6 Recommendations: What Measures Can be Taken to Improve the Understanding of Weather and Climate Extremes?

Drawing on the material presented in this report, recommendations are described in detail in Chapter 4. Briefly summarized here, they emphasize the highest priority areas for rapid and substantial progress in improving understanding of weather and climate extremes.

4.1. The continued establishment and maintenance of high quality climate observing systems to monitor climate variability and change should be of the highest priority

4.2 Efforts to digitize, homogenize, and analyze long-term observations in the instrumental record should be expanded.

4.3 Current weather observing systems should adhere to standards of observation that are consistent with the needs of both the climate and the weather forecasting communities.

4.4 Efforts to extend reanalysis products using surface observations should be pursued.

4.5 Research is needed to create annually-resolved, regional-scale reconstructions of the climate for the past 2,000 years.

4.6 Research efforts to improve our understanding of the mechanisms that govern hurricane intensity should be increased.

4.7 Substantial increases in computational and human resources should be made available to fully investigate the ability of climate models to recreate the recent past as well as make projections under a variety of future forcing scenarios.

4.8 Modeling groups should make available high temporal resolution data (daily, hourly) from climate model simulations both of the past and for the future to allow the investigation of potential changes in weather and climate extremes.

4.9 Research needs to move beyond purely statistical analysis and focus more on linked physical processes that produce extremes and their changes with climate.

4.10 Communication between the science community and the user community should be enhanced in both directions.

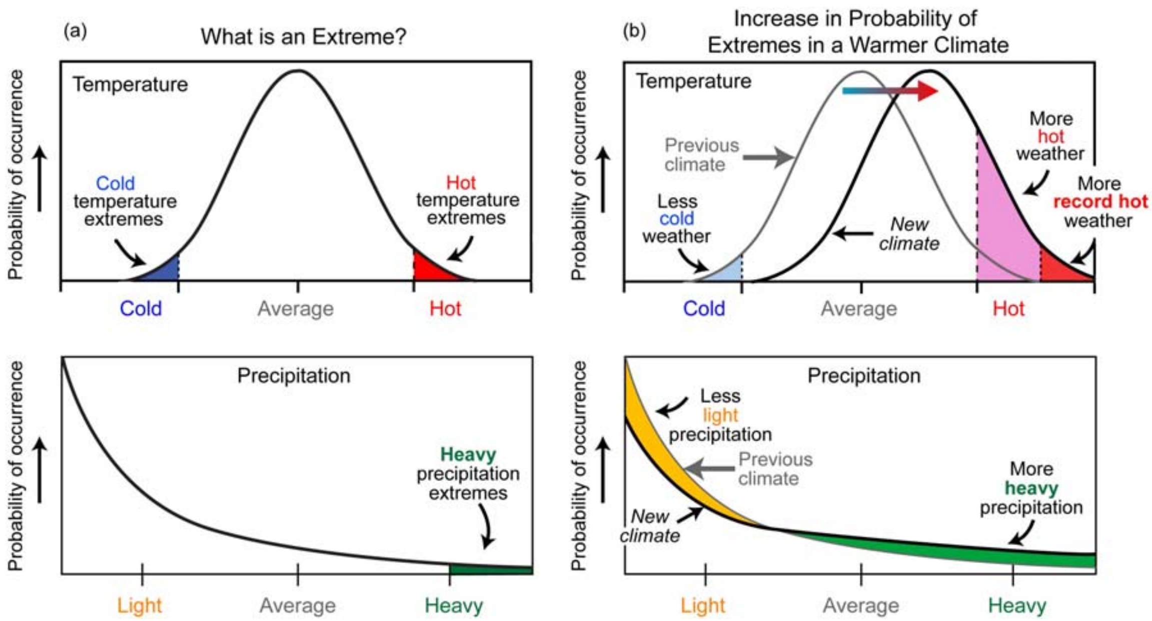

Figure ES.1 Most measurements of temperature (top) will tend to fall within a range close to average, so their probability of occurrence is high. A very few measurements will be considered extreme and these occur very infrequently. Similarly, for rainfall (bottom), there tend to be more days with relatively light precipitation and only very infrequently are there extremely heavy precipitation events, meaning their probability of occurrence is low. The exact threshold for what is classified as an extreme varies from one analysis to another, but would normally be as rare as, or rarer than, the top or bottom 10% of all occurrences. For the purposes of this report, all tornadoes and hurricanes are considered extreme.

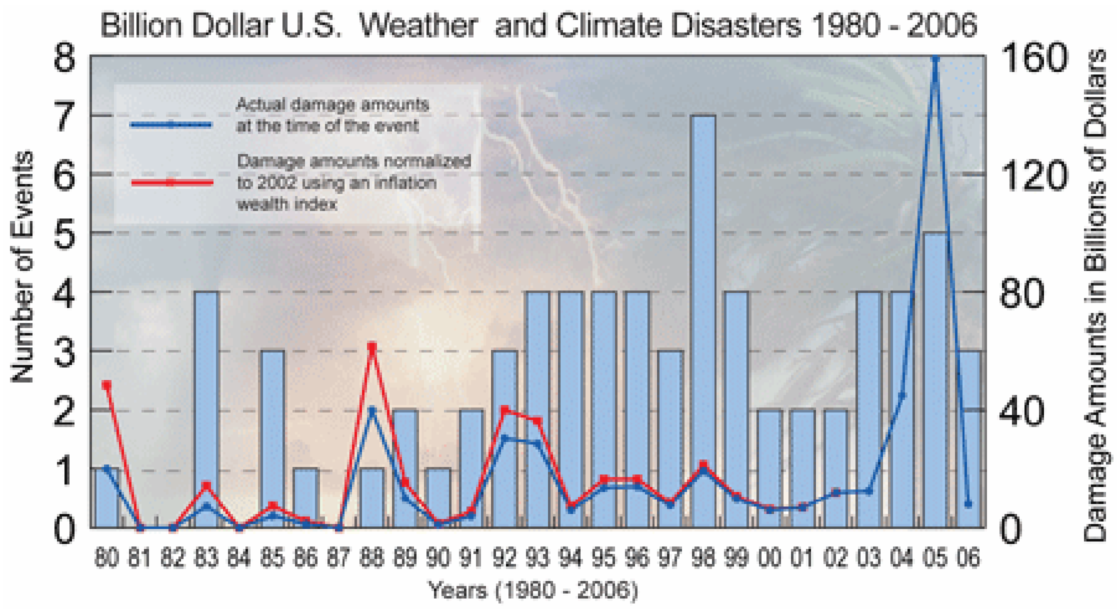

Figure ES.2 The blue bars show the number of events per year that exceed a cost of 1 billion dollars (these are scaled to the left side of the graph). The blue line (actual costs at the time of the event) and the red line (costs adjusted for wealth/inflation) are scaled to the right side of the graph, and depict the annual damage amounts in billions of dollars.

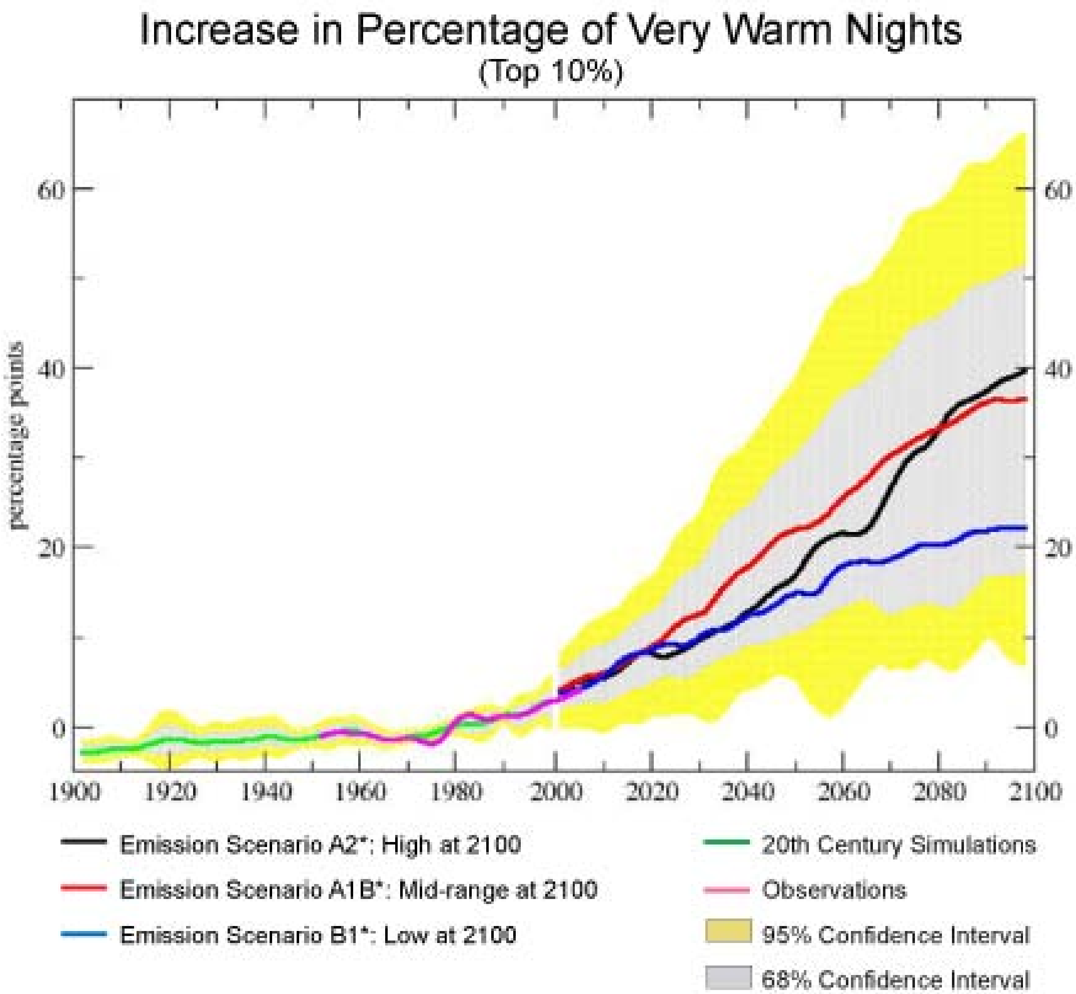

Figure ES.3 Increase in the percentage of days in a year in which the daily low temperature is unusually warm (falling in the top 10% of annual daily lows, using 1961 to 1990 as a baseline). Under the lower emissions scenario[*], the percentage of very warm nights increases about 20% by 2100 whereas under the higher emissions scenarios, it increases by about 40%.

[*]3 future emission scenarios from the IPCC Special Report on Emissions Scenarios:
B1 blue line: emissions increase very slowly for a few more decades, then level off and decline
A2 black line: emissions continue to increase rapidly and steadily throughout this century
A1B red line: emissions increase rapidly until 2050 and then decline.
There are more details on these emission scenarios in the glossary.

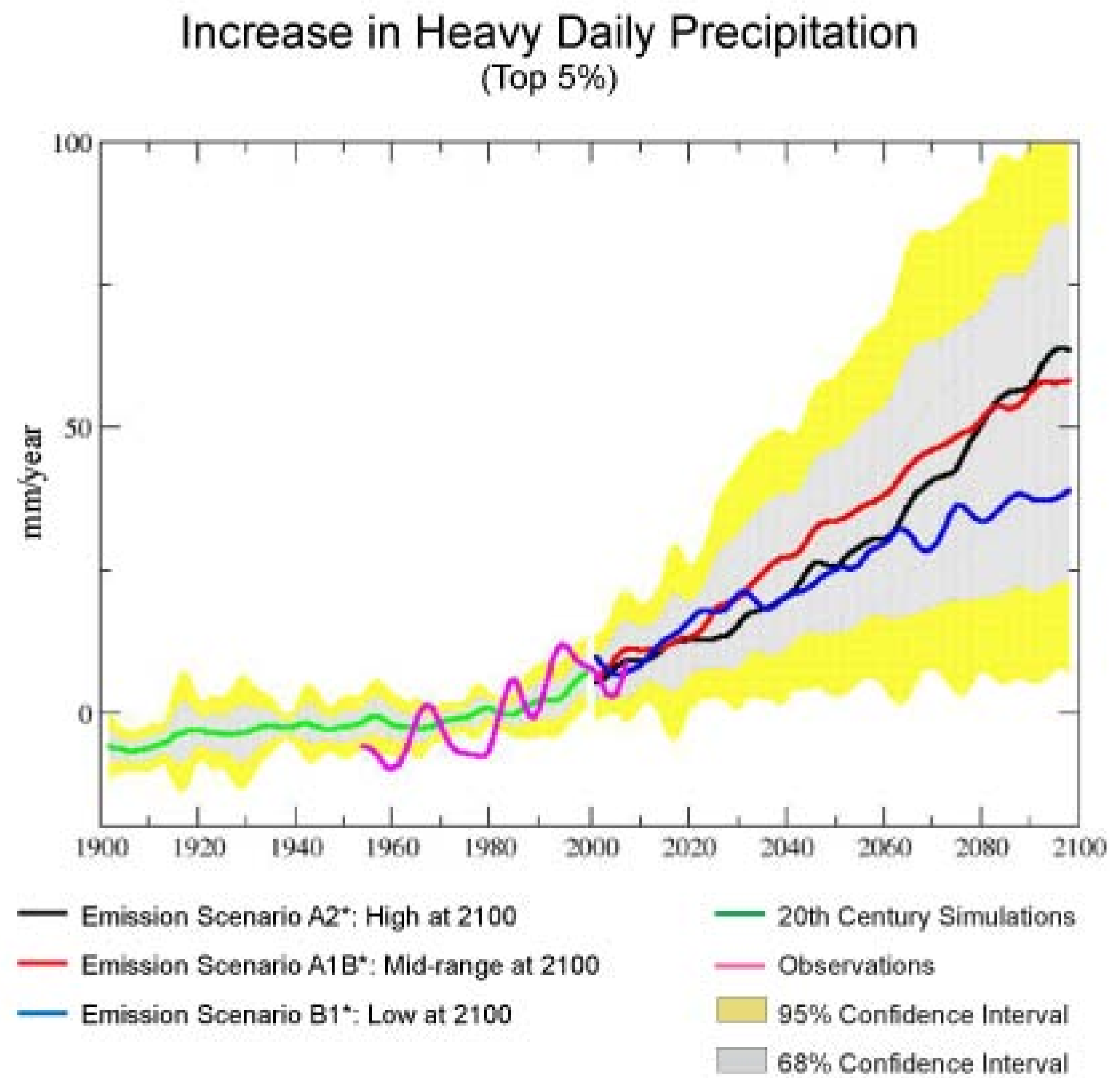

Figure ES.4 Increase in the amount of precipitation that falls in heavy events (the top 5% of all precipitation events in a year) compared to the 1961-1990 average Various emission scenarios are used for future projections*

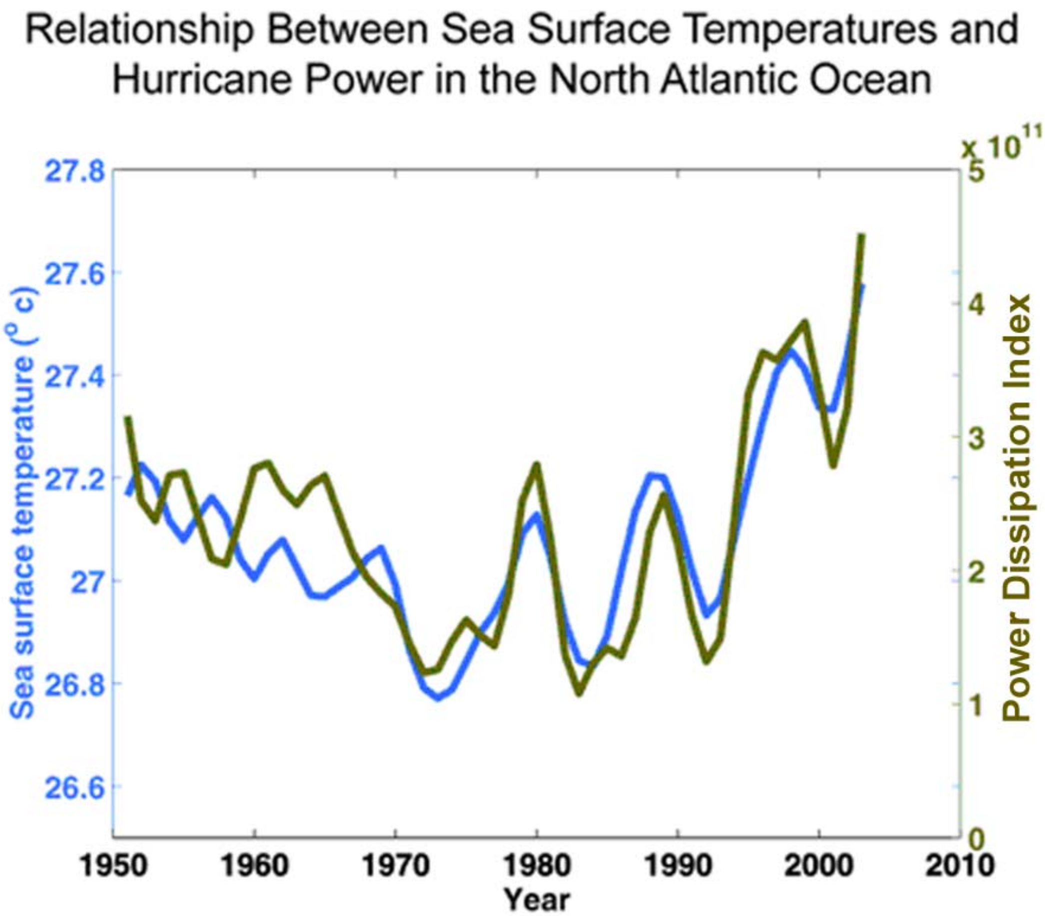

Figure ES.5 Sea surface temperatures (blue) are correlated with the Power Dissipation Index for North Atlantic hurricanes (Emanuel, 2007)

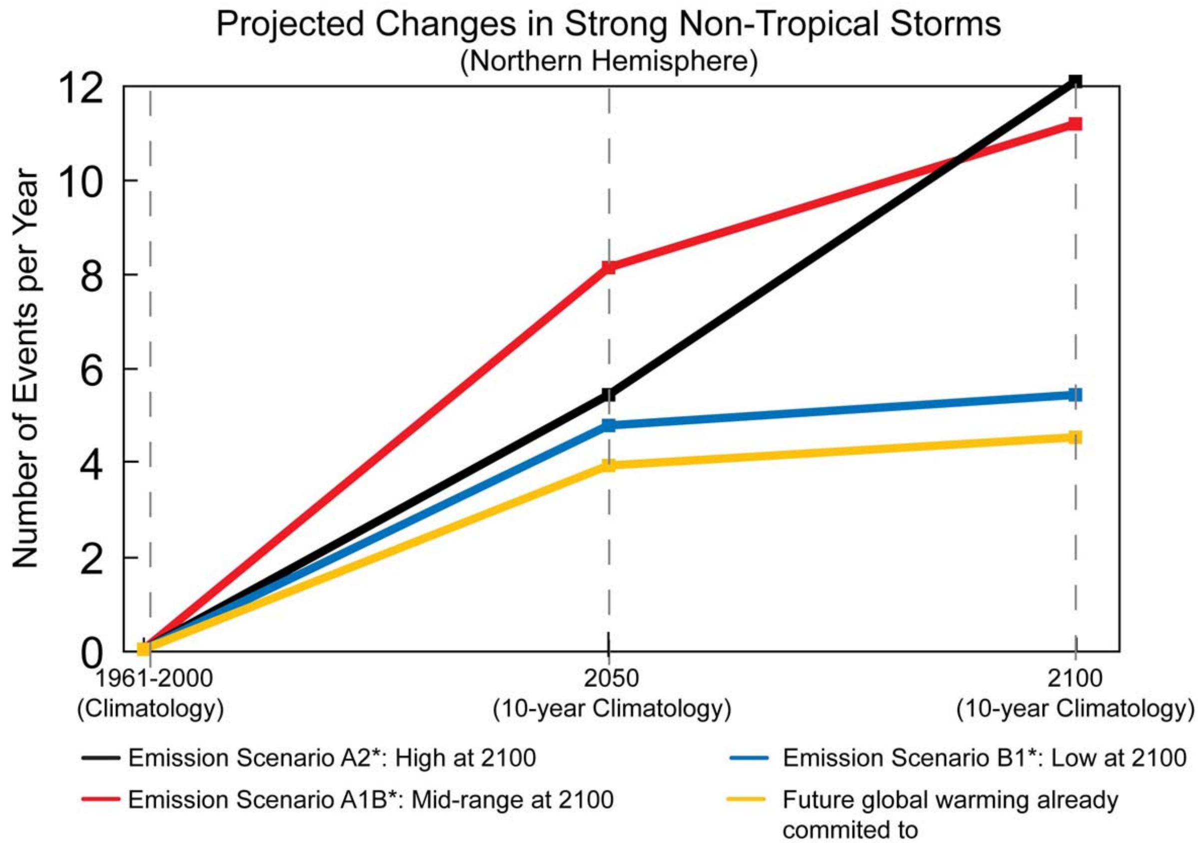

Figure ES.6 The projected change in intense low pressure systems (strong storms) during the cold season for the Northern Hemisphere for various emission scenarios* (adapted from Lambert and Fyfe; 2006).

Chapter 1 Why Weather and Climate Extremes Matter

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KEY FINDINGS

- Climate extremes expose existing human and natural system vulnerabilities.
- Changes in extreme events are one of the most significant ways socio-economic and natural systems are likely to experience climate change.
  – Systems have adapted to their historical range of extreme events.
  – The impacts of extremes in the future, some of which are expected to be outside the historical range of experience, will depend on both climate change and future vulnerability. The latter is shaped by factors such as population dynamics and poverty as well as by development and utilization of climate change adaptation measures such as appropriate building codes, disaster preparedness, and water use efficiency.
- Changes in extreme events are already observed to be having impacts on socio-economic and natural systems.
  – Two or more extreme events that occur over a short period reduce the time

available for recovery.

– The cumulative effect of back-to-back extremes is greater than if the same events are spread over a longer period.

- Extremes can have positive or negative effects. However, on balance, because systems have adapted to their historical range of extremes, the majority of the impacts of events outside this range are expected to be negative.
- Actions that lessen the risk from small or moderate events in the short-term can lead to increases in vulnerability to larger extremes in the long-term.

1.1 Extremes Matter Because They Impact People, Plants, and Animals

Observed and projected warming of North America has direct implications for the occurrence of extreme weather and climate events. It is very unlikely that the climate could change without extremes changing as well. Extreme events drive natural systems much more than average climate (Parmesan *et al.,* 2000). Extreme events cause property damage, injury, loss of life and threaten the existence of some species. Society recognizes the need to plan for the protection of communities from extreme events of various kinds. Structural measures (such as engineering works), governance measures (such as zoning and building codes), financial instruments (such as insurance and contingency funds) and emergency measures practices have all been used to lessen the impacts of historical extremes. To the extent that changes in extremes can be reliably forecast, society can engage in practices that would mitigate future impacts.

Global and regional climate patterns have changed throughout the history of our planet. Prior to the Industrial Revolution, these changes occurred due to natural causes, including variations in the Earth's orbital parameters, volcanic eruptions, and fluctuations in solar output. Since the late nineteenth century, atmospheric concentrations of carbon dioxide and other trace greenhouse gases (GHG) have been increasing due to human activity, such as fossil-fuel combustion and land-use change. On average, the world has warmed by 0.74°C over the last century with most of that coming in the last three decades, as documented by instrumental observations of air temperature over land and ocean surface temperature (IPCC, 2007a; Arguez, 2007; Lanzante *et al*., 2006). These observations are corroborated by, among many examples, the shrinking of mountain glaciers (Barry, 2006), later lake and river freeze dates and earlier thaw dates (Magnuson *et al.*, 2000), earlier blooming of flowering plants (Cayan *et al.*, 2001), earlier spring bird migrations (Sokolov, 2006), thawing permafrost and associated shift in ecosystem functioning, shrinking sea ice (Arctic Climate Impact Assessment, 2004), earlier spring events and shifts of plant and animal ranges both poleward and up mountainsides both within the U.S. (Parmesan and Galbraith, 2004) and globally (Walther *et al.*, 2002; Parmesan and Yohe, 2003; Root *et al.*, 2003; Parmesan 2006). Most of the recent warming observed around the world has very likely been due to observed changes in GHG concentrations (IPCC, 2007a). The continuing increase in GHG concentration is projected to result in additional warming of the global climate by 1.1 to 6.4°C by the end of this Century (IPCC, 2007a).

Extremes are already having significant impacts on North America. As examination of Figure 1.1 reveals, it is a rare year when the United States doesn't have any billion dollar weather and climate-related disasters. Furthermore, the costs of weather and climate-related disasters in the U.S. have been increasing faster than non-weather related disaster costs (Hazards and Vulnerability Research Institute, 2007). For the world as a whole, "weather-related [insured] losses in recent years have been trending upward much faster than population, inflation, or insurance penetration, and faster than non-weather-related events" (Mills, 2005a). Numerous studies indicate that both the climate and the socio-economic vulnerability to weather and climate extremes are changing, although their relative contributions to observed increases in disaster costs are subject to debate. For example the extent to which increases in coastal building damage is due to population growth[1] in vulnerable coastal locations versus increase in storm intensity is not easily quantified. Though the causes of the current damage increases are difficult to quantitatively assess, it is clear that any change in extremes will have a significant impact.

Hurricanes and tropical storms are the leading cause of billion dollar weather and climate events followed by floods, droughts and heat waves.. It should be noted that partitioning losses into the different categories is often not clear cut. For example, tropical storms also contribute to damages that were categorized as flooding and coastal. The annual mean loss of life from weather extremes in the U.S. exceeds 1,500 per year (Kunkel *et al.*, 1999) without including such factors as fog-related traffic fatalities. Approximately half

[1] Since 1980, the U.S. coastal population growth has generally reflected the same rate of growth as the entire nation (Crossett *et al.*, 2004).

of these deaths are related to hypothermia due to extreme cold, with extreme heat responsible for another one-fourth of the fatalities. There appears to be no trend in the number of these deaths (Goklany and Straja, 2000). However, it should be noted that these statistics were compiled before the 1,400 hurricane-related fatalities in 2004-2005 (Chowdhury and Leatherman, 2007).

Natural systems display complex vulnerabilities to climate change that sometimes are not evident until after the event. According to van Vliet and Leemans (2006), “the unexpected rapid appearance of ecological responses throughout the world” can be explained largely by the observed changes in extremes over the last few decades. Insects in particular have the ability to respond quickly to climate amelioration by increasing in abundances and/or increasing numbers of generations per year, which has resulted in widespread mortality of previously healthy trees (Logan *et al.*, 2003). The observed warming-related biological changes may have direct adverse effects on biodiversity, which in turn may impact ecosystem stability, resilience, and ability to provide societal goods and services (Parmesan and Galbraith, 2004; Arctic Climate Impact Assessment, 2004). The greater the change in global mean temperature, the greater will be the change in extremes and their consequent impacts on species and systems.

This introductory chapter addresses various questions that are relevant to the points raised above. Section 1.2 focuses on defining characteristics of extremes. Section 1.3 discusses the sensitivities of socio-economic and natural systems to changes in extremes. Factors that influence the vulnerability of systems to changes in extremes are described in section

1.4. As systems are already adapted to particular morphologies of extremes, section 1.5 explains why changes in extremes usually pose challenges. Section 1.6 describes how actions taken in response to those challenges can either increase or decrease future impacts of extremes. Lastly, in section 1.7, the difficulties in assessing extremes are discussed. The chapter also includes several text boxes, which highlight a number of topics related to particular extremes and their impacts, as well as analysis tools for assessing impacts.

1.2 Extremes Are Changing

When most people think of extreme weather or climate events, they focus on short-term intense episodes. However, this perspective ignores longer-term, more cumulative events, such as droughts. Thus, rather than defining extreme events solely in terms of temporal considerations, it is useful to look at them from a statistical point of view. If one plots all values of a particular variable, such as temperature, the values most likely will fall within a typical bell-curve with many values near average and fewer occurrences of values far away from the average. Extreme temperatures are in the tails of such distributions, as shown in the top panel of Figure 1.2.

According to the Glossary of the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (IPCC, 2007a), "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[2] of

[2] On average, one in every ten temperature values is cold enough to be at or below the 10^{th} percentile just as one in every ten temperature values is hot enough to be at or above the 90^{th} percentile.

the observed probability density function[3]. By definition, the characteristics of what is called extreme weather may vary from place to place in an absolute sense. 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)." Extreme climate events such as drought can often be viewed similarly to the tails on the temperature distribution.

Daily precipitation, however, has a distribution which is very different than the temperature distribution. Over most of North America, the majority of days have no precipitation at all. Of the days where some rain or snow does fall, many have very light precipitation while only a few have heavy precipitation, as illustrated by the bottom panel of Figure 1.2. Extreme value theory is a branch of statistics that fits a probability distribution to historical observations. The tail of the distribution can be used to estimate the probability of very rare events. This is the way the 100-year flood level can be estimated using 50 years of data. One problem with relying on historical data is that some extremes are far outside the observational record. For example, the heat wave that struck Europe in 2003 was so far outside natural variability that public health services were unprepared for the excess mortality (see Figure 1.3). Climate change is likely to increase the severity and frequency of extreme events for both statistical and physical reasons.

Wind is one parameter where statistics derived from all observations are not generally used to define what is an extreme. This is because most extreme wind events are

[3] A probability density function is the distribution of the probabilities of all different possible weather or climate events which is depicted by the heavy black lines in Figure 1.2.

generated by special meteorological conditions that are well known. For purposes of this report, all tornadoes and hurricanes are considered extreme. Extreme wind events associated with other phenomena, such as blizzards or nor'easters, tend to be defined by thresholds based on impacts rather than statistics or are just one aspect of the measure of intensity of these storms.

Most considerations of extreme weather and climate events are limited to discrete occurrences. However, in some cases, events that occur repeatedly can have impacts greater than the simple sum of each individual event. For example, the ice storm that occurred in eastern Ontario and southern Quebec in 1998 was the most destructive and disruptive storm in Canada in recent memory. The storm featured record amounts of freezing rain and sleet in a series of storms over a record number of hours. Further, the storm brutalized an area extending nearly 1000 km which included one of the largest urban areas of Canada, leaving more than 4 million people freezing in the dark for hours, if not days. The conditions were so severe that no clean-up action could be taken between storms and the ice built up, stranding even more people at airports, bringing down high-tension transmission towers, and straining food supplies. Such cumulative events need special consideration.

Also, compound extremes are conditions that depend on two or more parameters. For example, heat waves have greater impacts on human health when they are accompanied by high humidity. Additionally, problems with one extreme, such as a windstorm, may only be present if it is preceded by a different extreme, such as drought, which would, in this example, result in far more wind-blown dust than the storm would generate without the drought.

As the global climate continues to adjust to changes in radiative forcing brought on by increasing concentrations of GHG in the atmosphere, many different aspects of extremes have the potential to change as well (Easterling *et al.*, 2000a,b). The most commonly considered parameter is frequency. Is the extreme occurring more frequently? Will currently rare events become commonplace in 50 years? Changes in intensity are as important as changes in frequency. Are, for example, hurricanes becoming more intense? This is important because, as explained in the box on hurricanes, damage increases exponentially with the speed of the wind so a more intense hurricane causes much more destruction than a weak hurricane.

Frequency and intensity are only two parts of the puzzle. There are also temporal considerations, such as time of occurrence and duration. For example, the timing of peak snow melt in the western mountains has become earlier (Johnson *et al.*, 1999; Cayan *et al.* 2001). Earlier snowmelt in the Sierra Nevada Mountains means a longer dry season with far-reaching impacts on the ecologies of plant and animal communities, fire threat and human water resources. Indeed, in the American West, wildfires are strongly associated with increased spring and summer temperatures and correspondingly earlier spring snowmelt in the mountains (Westerling *et al.*, 2006). In Canada, human-induced warming of summer temperatures has a detectable influence on the increased area burned by forest fires in recent decades (Gillett *et al.*, 2004). Changing the timing and/or number of wildfires might pose threats to certain species by overlapping with their active seasons (causing increased deaths) rather than occurring during a species' dormant phase (when they are less vulnerable). Further, early snowmelt reduces summer water resources, particularly in California where summer rains are rare. The duration of events (such as heat waves, flood-inducing rains, and droughts), is also potentially subject to change. Spatial characteristics also need to be considered. Is the size of the impact area changing? In addition to the size of the individual events, the location is subject to change. For example, is the region susceptible to freezing rain moving farther north?

Therefore, the focus of this assessment is not only the meteorology of extreme events, but how climate change might alter the characteristics of extremes. Figure 1.4 illustrates how the tails of the distribution of temperature and precipitation are anticipated to change in a warming world. For temperature both the mean and the tails of the distributions are expected to warm. While the change in the number of average days may be small, the percentage change in the number of very warm and very cold days can be quite large. For precipitation, model and observational evidence indicates an increase in the number of heavy rain events which are balanced by a proportionate decrease in the number of light precipitation events.

1.3 Systems Are Sensitive to Changes in Extremes

Climate sensitivity is defined as the degree to which a system is affected by climate-related stimuli. The effect may be direct, such as changing crop yield due to variations in temperature or precipitation, or indirect, such as the decision to build a house in a location based on insurance rates, which can change due to flood risk caused by sea level rise (IPCC, 2007b). Indicators of climate sensitivity can include changes in, timing of life events, or distributions of individual species, or alteration of whole ecosystem functioning (Parmesan and Yohe, 2003; Parmesan and Galbraith, 2004).

Climate sensitivity directly impacts the vulnerability of a system or place. As a result, managed systems, both rural and urban, are constantly adjusting to changing perceptions of risks and opportunities. For example, hurricane destruction can lead to the adoption of new building codes (or enforcement of existing codes) and the implementation of new construction technology, which alter the future climate sensitivity of the community. Further, artificial selection and genetic engineering of crop plants can adjust agricultural varieties to changing temperature and drought conditions. Warrick (1980) suggested that the impacts of extreme events would gradually decline because of improved planning and early warning systems. Ausubel (1991) went further, suggesting that irrigation, air conditioning, artificial snow making, and other technological improvements, were enabling society to become more climate-proof. While North American society is not as sensitive to extremes as it was 400 years ago — for example, a megadrought in Mexico mid to late 1500s contributed to conditions that caused tremendous population declines as illustrated by Figure 1.5 — socio-economic systems are still far from being climate-proof.

Society is clearly altering relationships between climate and society, and thereby sensitivities to climate. However, this is not a unidirectional change. Societies make decisions that alter regional-scale landscapes (urban expansion, pollution, land-use change, water withdrawals) which can increase or decrease both societal and ecosystem sensitivities (*e.g.*, Mileti, 1999; Glantz, 2003). Contrary to an anticipated gradual decline in impacts, recent droughts have resulted in increased economic losses and conflicts (Riebsame *et al*., 1991; Wilhite, 2005). The increased concern about El Niño's impacts reflect a heightened awareness of its effects on extreme events worldwide, and growing concerns about the gap between scientific information and adaptive responses by communities and governments (Glantz, 1996). In the U.S. Disaster Mitigation Act of 2000, Congress specifically wrote that a "greater emphasis needs to be placed on . . . implementing adequate measures to reduces losses from natural disasters."

Many biological processes undergo sudden shifts at particular thresholds of temperature or precipitation (Precht *et al.*, 1973; Weiser, 1973; Hoffman and Parsons, 1997). The adult male/female sex ratios of certain reptile species such as turtles and snakes are determined by the extreme maximum temperature experienced by the growing embryo (Bull, 1980; Bull and Vogt, 1979; Janzen, 1994). A single drought year has been shown to affect population dynamics of many insects, causing drastic crashes in some species (Singer and Ehrlich, 1979; Ehrlich *et al.*, 1980; Hawkins and Holyoak, 1998) and population booms in others (Mattson and Haack, 1987). The nine-banded armadillo (Dasypus novemcinctus) cannot tolerate more than nine consecutive days below freezing (Taulman and Robbins, 1996). The high sea surface temperature (SST) event associated with El Niño in 1997-98 ultimately resulted in the death of 16% of the world's corals (Hoegh-Guldberg 1999, 2005; Wilkinson 2000); see the box on coral bleaching for more information. Further, ecosystem structure and function are impacted by major disturbance events, such as tornadoes, floods, and hurricanes (Pickett and White, 1985; Walker, 1999). Warming winters, with a sparse snow cover at lower elevations, have led to false springs and subsequent population declines and extirpation in certain butterfly species (Parmesan, 1996, 2005).

By far, most of the documented impacts on natural systems have been ecological in nature. Observed ecological responses to local, regional and continental warming include changes in species' distributions, changes in species' phenologies (the timing of the different phases of life events) and alterations of ecosystem functioning (Walther *et al.*, 2002; Parmesan and Yohe, 2003; Root *et al.*, 2003; Parmesan and Galbraith, 2004; Parmesan, 2006; IPCC 2007b). Changes in species' distributions include a northward and upward shift in the mean location of populations of the Edith's checkerspot butterfly in western North America of a magnitude approximately equal to the degree expected from the observed shift in thermal isotherms from 0.7 C warming – about 100 km northward and 100 m upward (Parmesan, 1996; Karl *et al.*, 1996). Phenological (*e.g.*, timing) changes includes lilac blooming 1.5 days earlier per decade and honeysuckle blooming 3.5 days earlier per decade since the 1960s in the western U.S. (Cayan *et al.*, 2001). In another example, tree swallows across the U.S. and southern Canada bred about 9 days earlier from 1959 to 1991, mirroring a gradual increase in mean May temperatures (Dunn and Winkler, 1999). One of the clearest examples of the impacts of warming on whole ecosystem functioning comes from the Arctic tundra, where warming trends have been considerably stronger than in the contiguous U.S. Melting and drying of the permafrost layer has caused an increase in decomposition rates of dead organic matter during winter, which ultimately in some areas has already resulted in a shift from the tundra being a carbon sink to being a carbon source (Oechel *et a.*, 1993; Oechel *et al.*, 2000).

Very few behavioral changes have been observed, but there is some evidence that individuals of the sooty shearwater have shifted their migration pathway from the coastal California current to a more central Pacific pathway, apparently in response to a warming-induced shift in regions of high productivity during their summer flight (Spear and Ainley, 1999; Oedekoven *et al.*, 2001). Evolutionary studies of climate change impacts are also few (largely due to dearth of data), but it is clear that genetic responses have already occurred (Parmesan, 2006). Genetic changes in local populations have taken place resulting in much higher frequencies of individuals who are warm-adapted (*e.g.*, for fruit flies; Rodriguez-Trelles and Rodriguez, 1998; Levitan, 2003; Balanya *et al.*, 2006), or can disperse better (*e.g.*, for the bush cricket; Thomas *et al.*, 2001). For species-level evolution to occur, either appropriate novel mutations or novel genetic architecture (*i.e.*, new gene complexes) would have to emerge to allow a response to selection for increased tolerance to more extreme climate than the species is currently adapted to (Parmesan *et al.*, 2000; Parmesan *et al.*, 2005). However, so far there is no evidence for change in the absolute climate tolerances of a species, and hence no indication that evolution at the species level is occurring, nor that it might occur in the near future (Parmesan, 2006).

Ecological impacts of climate change on natural systems are beginning to have carry-over impacts on human health (Parmesan and Martens, 2007). The best example comes from the bacteria which causes human cholera, *Vibrio cholerae,* which lives in brackish rivers and sea water and uses a diversity of marine life as reservoirs, including many shellfish, some fish, and even water hyacinth. Two-hundred years of observational records strong repeated patterns in which extreme warm water temperatures cause algae blooms which then promote rapid increases in zooplankton abundances, and hence also in their associated *V. cholerae* bacteria (Colwell, 1996). Analyses of long-term data sets from Peru and Bangladesch (from 18 years up to 70 years) show that cholera has recently become associated with El Niño events, suggesting a threshold for high transmission as only recently been commonly surpassed as El Niño events have become stronger and more frequent in the past three decades (Pascual *et al.,* 2000: Rodó *et al.,* 2002). Even when known epidemiological dynamics are taken into account (such as cycling of immunity in human populations), a strong El Niño signal in cholera dynamics is maintained (Koelle *et al.,* 2005). In summary, there is compelling evidence for links between climate variability, climate change (via increases in strength of El Niño), native plankton dynamics, bacterial dynamics in the wild, and cholera disease epidemics.

1.4 Future Impacts of Changing Extremes Also Depend on System Vulnerability

Climate change presents a significant risk management challenge, and dealing with weather and climate extremes is one of its more demanding aspects. In human terms, extreme events are important precisely because they expose the vulnerabilities of communities and the infrastructure on which they rely. Extreme weather and climate events are not simply hydrometeorological occurrences. They impact socio-economic systems and are often exacerbated by other stresses, such as social inequalities, disease, and conflict. Extreme events can threaten our very well-being. Understanding vulnerabilities from weather and climate extremes is a key first step in managing the risks of climate change.

According to IPCC (2007b), “vulnerability to climate change is the degree to which systems are susceptible to, and unable to cope with, adverse impacts.” Vulnerability is a function of the character, magnitude, and rate of climate variation to which a system is exposed, its sensitivity, and its adaptive capacity. A system can be sensitive to change but not vulnerable, such as agriculture in North America; or relatively insensitive but highly vulnerable. An example of the latter is incidence of diarrhea (caused by a variety of water-borne organisms) in less developed countries. Diarrhea, which is not correlated with temperatures in the U.S. because of highly-developed sanitation facilities, shows a strong correlation with high temperatures in Lima, Peru (Checkley *et al.*, 2000; WHO, 2003, 2004). Thus, vulnerability is highly dependent on robust societal infrastructures, which have been shown to break down under flood events even in the U.S. (Curreriero *et al.*, 2001). Systems that normally survive are those well adapted to the more frequent forms of low-damage events. On the other hand, the less frequent high-damage events can overwhelm the ability of any system to quickly recover.

The adaptive capacity of socio-economic systems is determined largely by their characteristics such as poverty and resource availability, which often can be managed. Communities with little adaptive capacities are those with limited economic resources, low levels of technology, weak information systems, poor infrastructure, unstable or weak institutions, and uneven access to resources. Enhancement of social capacity, effectively addressing some of the exacerbating stresses, represents a practical means of coping with changes and uncertainties in climate. However, despite advances in knowledge and technologies, costs appear to be a major factor in limiting the adoption of adaptation measures (White *et al.*, 2001).

Communities can often achieve significant reductions in losses from natural disasters by adopting land-use plans that avoid the hazards, *e.g.*, by not allowing building in a floodplain. Building codes are also effective for reducing disaster losses but they need to be enforced. For example, more than 25% of the damage from Hurricane Andrew could have been prevented if the existing building codes had been enforced (Board on Natural Disasters, 1999). The first major industry sector to pay attention to the threats posed by climate change was insurance, which recognized the steady increase in claims paralleling an increase in the number and severity of extreme weather and climate events – a trend that is expected to continue. The insurance industry in fact has an array of instruments/levers that can stimulate policy-holders to take actions to adapt to future extremes. These possibilities are increasingly being recognized by governments. When such measures take effect, the same magnitude event can have less impact, as illustrated by the top panel of Figure 1.6.

Extreme events themselves can alter vulnerability and expose underlying stresses. There are obvious response times for recovery from the effects of any extreme weather or climate event – ranging from several decades in cases of significant loss of life, to years for the salinization of agricultural land following a tropical storm, to several months for stores to restock after a hurricane. A series of extreme events that occurs in a shorter period than the time for recovery can exacerbate the impacts as illustrated in the bottom panel of Figure 1.6. For example, in 2005 there was a series of hurricanes that made landfall in Florida; these occurred close enough in time and space that it often proved impossible to recover from one hurricane before the next arrived. Hardware stores and lumberyards were not able to restock quickly enough. A multitude or sequence of extreme events can also strain the abilities of insurance and re-insurance companies to compensate victims. Extremes can also initiate adaptive responses. For example, droughts in the 1930s triggered waves of human migration that altered the demographics of the United States. After the 1998 eastern Canadian ice storm the design criteria for freezing rain on high-voltage power and transmission lines were changed to accommodate radial ice accretion of 25 mm in the Great Lakes region to 50 mm for Newfoundland and Labrador (Canadian Standards Association, 2001).

Factors such as societal exposure, vulnerability, and sensitivity to weather and climate can play a significant role in determining whether a weather or climate event is considered extreme. In fact, an extreme weather or climate event, defined solely using statistical properties, may not be perceived to be an extreme if it affects an exposure unit[4] that is designed to withstand that extreme. Conversely, a weather or climate event that is not extreme in a statistical sense might still be considered an extreme event because of the resultant impacts. Case in point, faced with an extended dry spell, consider the different effects and responses in a city with a well-developed water supply infrastructure

[4] An exposure unit can be a person, home, city, or animal or plant community.

and a village in an underdeveloped region with no access to reservoirs. These differences also highlight the role of adaptive capacity in a society's response to an extreme event. Wealthy societies will be able to devote the resources needed to construct a water supply system that can withstand an extended drought.

Given the relationship between extreme events and their resultant socio-economic impacts, it would seem that the impacts alone would provide a good way to assess changes in extremes. Unfortunately, attempts to quantify trends in the impacts caused by extreme events are hindered by the difficulty in obtaining loss-damage records. As a result, there have been many calls for improvements in how socio-economic data are collected (Changnon, 2003; Cutter and Emrich, 2005; National Research Council, 1999). However, there is no government-level coordinated mechanism for collecting data on all losses or damage caused by extreme events. A potentially valuable effort, led by the Hazards Research Lab at the University of South Carolina, is the assembly of the Spatial Hazard Events and Losses Database for the United States (Cutter *et al.*, 2007). If successful, this effort could provide standardized guidelines for loss estimation, data compilation, and metadata standards. Without these types of guidelines, a homogeneous national loss inventory will remain a vision and it will not be possible to precisely and accurately detect and assess trends in losses and quantify the value of mitigation.

To date most efforts at quantifying trends in losses caused by impacts are based on insured loss data or on total loss (insured plus non-insured losses) estimates developed by insurers. Unfortunately, the details behind most of the insured loss data are proprietary and only aggregated loss data are available. The relationship between insured losses and total losses will likely vary as a function of extreme event and societal factors such as building codes, the extent of insurance penetration, and more complex societal factors. The National Hurricane Center generally assumes that for the United States, total losses are twice insured loss estimates. However, this relationship will not hold for other countries or other weather phenomena.

Regardless of the uncertainties in estimating insured and total losses, it is clear that the absolute dollar value of losses from extreme events has increased over the past few decades, even after accounting for the effects of inflation (see Figure 1.1). However, much of the increasing trend in losses, particularly from tropical cyclones, appears to be related to an increase in population and wealth (Pielke *et al.*, 2003; Pielke, 2005; Pielke and Landsea, 1998). The counter argument is that there is a climate change signal in recent damage trends. Similarly, those damage trends have increased significantly despite ongoing adaptation efforts that have been taking place (Mills, 2005b; Stott *et al.*, 2004; Kunkel *et al.*, 1999). A number of other complicating factors also play a role in computing actual losses. For example, all other things being equal, the losses from Hurricane Katrina would have been dramatically lower if the dikes had not failed. In addition, the potential for an increase in storm intensity (*e.g.*, tropical cyclone wind speeds and precipitation) (Knutson and Tuleya, 2003) and the intensity of the hydrological cycle[5] (Trenberth *et al.*, 2003) raises the possibility that changes in climate extremes will contribute to an increase in loss.

Another confounding factor in assessing extremes through their impacts is that an extreme event that lasts for a few days or even less can have impacts that persist for decades. For example, it will take years for Honduras and Guatemala to recover from the damage caused by Hurricane Mitch in 1998 and it seems likely that New Orleans will need years to recover from Hurricane Katrina. Furthermore, extreme events not only produce “losers” but “winners” too. Examples of two extreme-event winners are the construction industry in response to rebuilding efforts and the tourism industry at locations that receive an unexpected influx of tourists who changed plans because their first-choice destination experienced an extreme event that crippled the local tourism facilities. Even in a natural ecosystem there are winners and losers. For example, the mountain pine beetle infestation in British Columbia has been warmly greeted as a dinner bell by woodpeckers.

1.5 Systems are Adapted to Particular Morphologies of Extremes so Changes in Extremes Pose Challenges

Over time, socio-economic and natural systems adapt to their climate, including extremes. Snowstorms that bring traffic to a standstill in Atlanta are shrugged off in Minneapolis. Hurricane-force winds that topple tall non-indigenous Florida trees like the Australian pine (*Casuarina equisetifolia*) may only break a few small branches from the

[5] The hydrologic cycle is the continuous movement of water on, above and below the surface of the Earth where it evaporates from the surface, condenses in clouds, falls to Earth as rain or snow, flows downhill in streams and rivers and then evaporates again.

native live oak (*Quercus virginiana*) or gumbo-limbo (*Bursera simaruba*) trees that evolved in areas frequented by strong winds. Some species even depend on major extremes happening. For example, the jack pine (*Pinus banksiana*) produces very durable resin-filled cones that remain dormant until wildfire flames melt the resin. Then the cones pop open and spread their seeds (Herring, 1999).

Therefore, it is less a question of whether extremes are good or bad, but rather, what will be the impact of their changing characteristics? For certain species and biological systems, various processes may undergo sudden shifts at specific thresholds of temperature or precipitation (Precht *et al.*, 1973; Weiser, 1973; Hoffman and Parsons, 1997), as discussed in section 1.3. Generally, managed systems are more buffered against extreme events than natural systems, but certainly are not immune to them. The heat waves of 1995 in Chicago and 2003 in Europe caused considerable loss of life in large part because building architecture and city design were adapted for more temperate climates and not adapted for dealing with such extreme and enduring heat (Patz *et al.*, 2005). On balance, because systems have adapted to their historical range of extremes, the majority of the impacts of events outside this range are negative (IPCC, 2007b).

When considering how the statistics of extreme events have changed, and may change in the future, it is important to recognize how such changes may affect efforts to adapt to them. Adaptation is important because it can reduce the extent of damage caused by extremes (*e.g.*, Mileti, 1999; Wilhite, 2005). Currently, long-term planning uses, where possible, the longest historical time series, including consideration of extreme events. The combined probabilities of various parameters that can occur at any given location can be considered the cumulative hazard of a place. Past observations lead to expectations of their recurrence, and these form the basis of building codes, infrastructure design and operation, land-use zoning and planning, insurance rates, and emergency response plans.

However, what would happen if statistical attributes of extreme events were to change as the climate changes? Individuals, groups, and societies would seek to adjust to changing exposure. Yet the climate may be changing in ways that pose difficulties to the historical decision-making approaches (Burton *et al*., 1993). The solution is not just a matter of utilizing global climate model projections. It is also involves translating the projected changes in extremes into changes in risk.

Smit *et al.* (2000) outline an "anatomy" of adaptation to climate change and variability, consisting of four elements: a) adapt to what, b) who or what adapts, c) how does adaptation occur, and d) how good is the adaptation. Changing extreme statistics will influence the adaptation. As noted earlier, a change in the frequency of extreme events may be relatively large, even though the change in mean is small. Increased frequencies of extreme events could lead to reduced time available for recovery, altering the feasibility and effectiveness of adaptation measures. Changes to the timing and duration of extremes, as well as the occurrence of new extreme thresholds (*e.g.*, greater precipitation intensity, stronger wind speeds), would be a challenge to both managed and unmanaged systems.

Trends in losses or productivity of climate-sensitive goods exhibit the influences of both climate variability/change and ongoing behavioral adjustments. For example, U.S. crop yields have generally increased with the introduction of new technologies. As illustrated by Figure 1.7, climatic variability still causes short-term fluctuations in crop production, but a poor year in the 1990s tends to have better yields than a poor year (and sometimes even a good year) in the 1960s. Across the world, property losses show a substantial increase in the last 50 years, but this trend is being influenced by both increasing property development and offsetting adaptive behavior. For example, economic growth has spurred additional construction in vulnerable areas but the new construction is often better able to withstand extremes than older construction. Future changes in extreme event will be accompanied by both autonomous and planned adaptation, which will further complicate calculating losses due to extremes.

1.6 Actions Can Increase or Decrease the Impact of Extremes

It is important to note that most people do not use climate and weather data, and forecasts directly. People who make decisions based on meteorological information typically base their decisions on the output of an intermediate model that translates the data into a form that is more relevant for their decision process (Figure 1.8). For example, a farmer will not use weather forecasts or climate data directly when making a decision on when to fertilize a crop or on how much pesticide to apply. Instead, the forecast is filtered through a model or mental construct that uses such information as one part of the decision process and includes other inputs such as crop type, previous pesticide application history, government regulations, market conditions, producer recommendations, and the prevalence and type of pest.

One useful decision tool is a plant hardiness zone map (Cathey, 1990). Plant hardiness zones are primarily dependent on extreme cold temperatures. Already due to changing locations of plant hardiness zones, people are planting fruit trees such as cherries farther north than they did 30 years ago as the probability of winterkill has diminished. This type of adaptation is common among farmers who continually strive to plant crop species and varieties well suited to their current local climate.

To a large extent, individual losses for hazard victims have been reduced as the larger society absorbs a portion of their losses through disaster relief and insurance. Clearly relevant for settings such as New Orleans is the so-called levee effect, first discussed by Burton (1962), in which construction of levees (dams, revetments, beach nourishment) induces additional development leading to much larger losses when the levee is eventually overtopped. A more general statement of this proposition is found in the safe development paradox in which increased safety (*e.g.*, flood control) induces increased development (such as in areas considered safe due to the protection provided by levees or dams) leading to increased losses when a major event hits. The notion that cumulative reduction of smaller scale risks might increase vulnerability to large events has been referred to as the *levee effect* even when the concern has nothing to do with levees (Bowden *et al*., 1981).

After particularly severe or visible catastrophes, policy windows have been identified as windows of opportunity for creating long-term risk reduction plans which can include adaptation for climate change. A policy window opens when the opportunity arises to change policy direction and is thus an important part of agenda setting (Kingdon, 1995). Policy windows can be created by triggering or focusing events, such as disasters, as well as by changes in government and shifts in public opinion. Immediately following a disaster, the social climate may be conducive to much needed legal, economic, and social change, which can begin to reduce structural vulnerabilities. Indeed, an extreme event that is far out of normal experience can wake society up to the realization that extremes are changing and that society must adapt to these changes.

The assumptions behind the utility of policy windows are that (1) new awareness of risks after a disaster leads to broad consensus, (2) agencies are reminded of disaster risks, and (3) enhanced community will and resources become available. However, during the post-recovery phase, reconstruction requires weighing, prioritizing, and sequencing of policy programming, and there are usually too many mainstreaming agendas for most decision makers and operational actors to digest with attendant requests for resources for various actions. Thus, there is pressure to quickly return to the “normal” conditions prior to the event, rather than incorporate longer-term development strategies (Berube and Katz, 2005; Christoplos, 2006). In addition, while adaptive institutions clearly matter, they are often not there in the aftermath (or even before the occurrence) of a disaster.

In contrast to the actual reconstruction plans, the *de facto* decisions and rebuilding undertaken ten months after Katrina clearly demonstrate the rush to rebuild the familiar, as found after other major disasters in other parts of the world (Kates *et al*., 2006). This perspective helps explain the evolution of vulnerability of settings such as New Orleans, where smaller events have been mitigated, but with attendant increases in long-term vulnerability. As in diverse contexts such as El Niño-Southern Oscillation (ENSO) related impacts in Latin America, induced development below dams or levees in the United States, and flooding in the United Kingdom, the result is that focusing only on short-term risk reduction can actually produce greater vulnerability to future events (Pulwarty *et al*., 2003). Thus, the evolution of responses in the short-term after each extreme event can appear logical, but might actually increase long-term risk to larger or more frequent events. Adaptation to climate change must be placed within the context of adaptation to climate across time scales (from extremes and interannual variability through long-term change) if it is to be embedded into effective response strategies.

According to the Stern Review on the economics of climate change (Stern, 2006), "many developing countries are already struggling to cope with their current climate. Both the economic costs of natural disasters and their frequency have increased dramatically in the recent past. Global losses from weather-related disasters amounted to a total of around $83 billion during the 1970s, increasing to a total of around $440 billion in the 1990s with the number of 'great natural catastrophe' events increasing from 29 to 74 between those decades. The financial costs of extreme weather events represent a greater proportion of GDP loss in developing countries, even if the absolute costs are more in developed countries given the higher monetary value of infrastructure. And over 96% of all disaster-related deaths worldwide in recent years have occurred in developing countries. Climatic shocks can - and do - cause setbacks to economic and social development in developing countries. The IMF, for example, estimates costs of over 5% of GDP per large disaster on average in low-income countries between 1997 and 2001." Given the high costs, wise adaptation has ample opportunity to save money in the long run.

1.7 Assessing Impacts of Changes in Extremes Is Difficult

As has been mentioned, assessing consequences relevant to extreme weather and climate events is not simply a function of the hydrometeorological phenomena but depends critically on the vulnerability of the system being impacted. Thus, the context in which these extreme events take place is crucial. This means that while the changes in extreme events are consistent with a warming climate (IPCC, 2007a), any analysis of past events or projection of future events has to carefully weigh non-climatic factors. In particular, consideration must be given to changes in demographic distributions and wealth. It is likely that part of the increase in economic losses shown in Figure 1.1 has been due to increases in population in regions that are vulnerable such as coastal communities affected by hurricanes, sea-level rise, and storm surges. In addition, property values have risen. These factors increase the sensitivity of our infrastructure to extreme events. Together with the expected increase in the frequency and severity of extreme events (IPCC 2007a), our vulnerability to extreme events is very likely to increase. Unfortunately, because many extreme events occur at small temporal and spatial scales, where model skill is currently limited and local conditions are highly variable, projections of future impacts cannot always be made with a high level of confidence.

While anthropogenic climate change is very likely to affect the distribution of extreme events, it can be misleading to attribute any particular event solely to human causes. Nevertheless, scientifically valid statements regarding the increased risk can sometimes be made. A case in point is the 2003 heat wave in Europe, where it it is very likely that human influence at least doubled the risk of such a heat wave occurring (Stott *et al*., 2004). Furthermore, over time, there is expected to be some autonomous adaptation to experienced climate variability and other stresses. Farmers, for example, have traditionally altered their agricultural practices, such as planting different crop varieties, based on experience and water engineers have built dams and reservoirs to better manage resources during recurring floods or droughts. Such adaptation needs to be considered when assessing the importance of future extreme events.

Assessing historical extreme weather and climate events is more complicated than just the statistical analysis of available data. Intense rain storms are often of short duration and not always captured in standard meteorological records; however, they can often do considerable damage to urban communities, especially if the infrastructure has not been enhanced as the communities have grown. Similarly, intense wind events (hurricanes are a particular example), may occur in sparsely populated areas or over the oceans, and it is only since the 1960s, with the advent of satellite observations, that a comprehensive picture can be put together. Therefore, it is important to continually update the data sets and improve the analyses. For example, probabilistic estimates of rainfall intensities for a range of durations, from 5 minutes to 24 hours for return periods, or recurrence intervals of 20, 50, and 100 years, have long been employed by engineers when designing many types of infrastructure. In the United States, these probabilistic estimates of intense precipitation are in the process of being updated. Newer analysis based on up-to-date rainfall records often differ by more than 45% from analyses done in the 1970s (Bonnin *et al*., 2003).

1.8 Summary and Conclusions

For good and for ill, weather and climate extremes have always been present. Both socio-economic and natural systems are adapted to historical extremes. Changes from this historical range matter because people, plants, and animals tend to be more impacted by changes in extremes compared to changes in average climate. Extremes are changing, and in some cases impacts on socio-economic and natural systems have been observed. The vulnerability of these systems is a function not only of the rate and magnitude of climate change but also depends on the sensitivity of the system, the extent to which it is exposed, and its adaptive capacity. Vulnerability can be exacerbated by other stresses such as social inequalities, disease, and conflict, and can be compounded by changes in other extremes events (e.g., drought and heat occurring together) and by rapidly-recurring events.

Despite the widespread evidence that humans have been impacted by extreme events in the past, predicting future risk to changing climate extremes is difficult. Extreme phenomena are often more difficult to predict than changes in mean climate. In addition, systems are adapting and changing their vulnerability to risk in different ways. The ability to adapt differs among systems and changes through time. Decisions to adapt to or mitigate the effect of changing extremes will be based not only on our understanding of climate processes but also on our understanding of the vulnerability of socio-economic and natural systems.

BOX 1.1: Warm Temperature Extremes and Coral Bleaching

Corals are marine animals that obtain much of their nutrients from symbiotic unicellular algae that live protected within the coral's calcium carbonate skeleton. Elevated sea surface temperatures (SST), one degree C above long-term summer averages, lead to the loss of algal symbionts resulting in bleaching of tropical corals (Hoegh-Guldberg, 1999). While global SST has risen an average of 0.13°C per decade since 1950 (IPCC, 2007a), a more acute problem for coral reefs is the increase in episodic warming events such as El Niño. High SSTs associated with the strong El Niño event in 1997-98 caused bleaching in every ocean (up to 95% of corals bleached in the Indian Ocean), ultimately resulting in 16% of corals dying globally (Hoegh-Guldberg, 1999, 2005; Wilkinson, 2000).

Recent evidence for genetic variation in temperature thresholds among the obligate algal symbionts suggests that some evolutionary response to higher water temperatures may be possible (Baker, 2001; Rowan, 2004). Changes in genotype frequencies toward increased frequency of high temperature-tolerant symbionts appear to have occurred within some coral populations between the mass bleaching events of 1997/1998 and 2000/2001 (Baker *et al*., 2004). However, other studies indicate that many entire reefs are already at their thermal tolerance limits (Hoegh-Guldberg, 1999). Coupled with poor dispersal of symbionts between reefs, this has led several researchers to conclude that local evolutionary responses are unlikely to mitigate the negative impacts of future temperature rises (Donner *et al*., 2005; Hoegh-Guldberg *et al*., 2002). Interestingly, though, hurricane-induced ocean cooling can temporarily alleviate thermal stress on coral reefs (Manzello *et al.*, 2007).

Examining coral bleaching in the Caribbean, Donner *et al.* (2007) concluded that “the observed warming trend in the region of the 2005 bleaching event is unlikely to be due to natural climate variability alone.” Indeed, “simulation of background climate variability suggests that anthropogenic warming may have increased the probability of occurrence of significant thermal stress events for corals in this region by an order of magnitude. Under scenarios of future greenhouse gas emissions, mass coral bleaching in the eastern Caribbean may become a biannual event in 20–30 years.” As coral reefs make significant contributions to attracting tourists to the Caribbean, coral bleaching has adverse socio-economic impacts.

BOX 1.2: Cold Temperature Extremes and Forest Beetles

Forest beetles in western North America have been responding to climate change in ways that are destroying large areas of forests (see Figure 1.9). The area affected is 50 times larger than the area affected by forest fire with an economic impact nearly five times as great (Logan *et al*., 2003). Two separate responses are contributing to the problem. The first is a response to warm summers, which enable the mountain pine beetle (*Dendroctonus ponderosae*), in the contiguous United States, to have two generations in a year, when previously it had only one (Logan *et al.*, 2003). In south-central Alaska, the spruce beetle (*Dendroctonus rufipennis*) is maturing in one year, where previously it took two years (Berg *et al*., 2006).

The second response is to winter temperatures, specifically extremely cold winter temperatures, which strongly regulate over-winter survival of the spruce beetle in the Yukon (Berg *et al*., 2006) and the mountain pine beetle in British Columbia. The supercooling threshold, which is the temperature at which the insect freezes and dies, for spruce beetle larvae, is −41°C[6] and for adults −37°C (Werner *et al*., 2006). Recent warming, limiting the frequency of sub−40°C occurrences, has reduced over-winter mortality of mountain pine beetle larvae in British Columbia. It has led to an explosion of the beetle population, with tree losses covering an area of 8.7 million hectares[7] in 2005, a doubling since 2003, and a 50-fold increase since 1999 (British Columbia Ministry of Forests and Range, 2006a). It is estimated that at the current rate of spread, 80% of British Columbia's mature lodgepole pine trees, the province's most abundant commercial tree species, will be dead by 2013 (Natural Resources Canada, 2007). Similarly in Alaska, approximately 847,000 hectares of south-central Alaska spruce forests were infested by spruce beetles from 1920 to 1989 while from 1990 to 2000, an extensive outbreak of spruce beetles caused mortality of spruce across 1.19 million

[6] The freezing point of water is 0°C or 32°F. The boiling point of water is 100 degrees higher in Celsius (100°C) and 180 degrees higher in Fahrenheit (212°F). Therefore, to convert from Celsius to Fahrenheit, multiply the Celsius temperature by 1.8 and then add 32.
[7] One hectare is 10,000 square meters or the area in a square with sides of 100 meters and equals 2.5 acres.

hectares, approximately 40% more forest area than had infested the state the previous 70 years (Werner *et al*., 2006). The economic loss goes well beyond the millions of board feet of dead trees as tourism revenue is highly dependent on having healthy, attractive forests. Hundreds of millions of dollars are being spent to mitigate the impacts of beetle infestation in British Columbia alone (British Columbia Ministry of Forests and Range, 2006b).

The beetle-forest relationships are much more complex than just climate and beetle survival and life cycle. In the contiguous United States, increased beetle populations have increased incidences of a fungus they transmit (pine blister rust, *Cronartium ribicola*) (Logan *et al*., 2003). Further, in British Columbia and Alaska, long-term fire suppression activities have allowed the area of older forests to double. Older trees are more susceptible to beetle infestation. The increased forest litter from infected trees has, in turn, exacerbated the forest fire risks. Forest managers are struggling to keep up with changing conditions brought about by changing climate extremes.

BOX 1.3: Heavy Precipitation and Human Health

Anthropogenic climate change is already affecting human health (WHO 2002, 2003, 2004). For the year 2000, the World Health Organization estimated that 6% of malaria infections, 7% of dengue fever cases and 2.4% of diarrhea could be attributed to climate change (Campbell-Lendrum *et al.*, 2003). Increases in these water borne diseases has been attributed to increases in intensity and frequency of flood events, which in turn has been linked to greenhouse-gas driven climate change (Easterling *et al.,* 2000a,b; IPCC

2007a). Floods directly promote transmission of water-borne diseases by causing mingling of untreated or partially treated sewage with freshwater sources, as well as indirectly from the breakdown of normal infrastructure causing post-flood loss of sanitation and fresh water supplies (Atherholt *et al.,* 1998*;* Rose *et al.*, 2000; Curriero *et al.,* 2001; Patz *et al.*, 2003). Precipitation extremes also cause increases in malnutrition due to drought and flood-related crop failure. For all impacts combined, WHO estimated that for a single year, total deaths due to climate change of 150,000 people (WHO 2002).

There is general agreement that the health sectors are strongly buffered against responses to climate change, and that a suite of more traditional factors is often responsible for both chronic and epidemic health problems. These include quality and accessibility of health care, sanitation infrastructure and practices, land use change (particularly practices which alter timing and extent of standing water), pollution, population age structure, presence and effectiveness of vector control programs, and general socio-economic status (Patz *et al.,* 2001; IPCC 2001b; Gubler *et al.,* 2001; Campbell-Lendrum *et al.,* 2003; Wilkinson *et al.,* 2003; WHO 2004, IPCC 2007b).

It is generally assumed that diarrhea incidence in developed countries, which have much better sanitation infrastructure, has little or no association with climate (WHO 2003, 2004). Studies for the U.S., however, indicate that the assumption that developed countries have low vulnerability may be premature, as independent studies have repeatedly concluded that water and food-borne pathogens (that cause diarrhea) will likely increase with projected increases in regional flooding events, primarily by contamination of main waterways (Rose *et al.*, 2000; Ebi *et al.*, 2006).

A U.S. study documented that 51% of waterborne disease outbreaks were preceded by precipitation events above the 90th percentile, with 68% of outbreaks preceded by precipitation above the 80th percentile (Curriero *et al.*, 2001). These outbreaks comprised mainly intestinal disorders due to contaminated well water or water treatment facilities that allowed microbial pathogens, such as *E. coli*, to enter drinking water. In 1993, 54 people in Milwaukee, Wisconsin died in the largest reported flood-related disease outbreak (Curriero *et al.*, 2001). The costs associated with this one outbreak were $31.7 million in medical costs and $64.6 million in productivity losses (Corso et al., 2003).

Another heavy precipitation-human health link comes from the southwestern desert of the United States. This area experienced extreme rainfalls during the intense 1992/1993 El Niño. Excess precipitation promoted lush vegetative growth, which led to population booms of deer mice *(Peromyscus maniculatus*). This wild rodent carries the hantavirus which is transmissible to humans and causes a hemorrhagic fever that is frequently lethal. The virus is normally present at moderate levels in wild mouse populations. In most years, humans in nearby settlements experienced little exposure. However, in 1993, local overcrowding arising from the wet-year/population boom, caused greater spillover rodent activity. Subsequent increased human contact and higher transmission rates led to a major regional epidemic of the virus (Engelthaler *et al.*, 1999; Glass *et al.*, 2000). Similar dynamics have been shown for plague in the western United States (Parmenter *et al*., 1999).

BOX 1.4: Drought

Drought should not be viewed as merely a physical phenomenon. Its impacts on society result from the interplay between a physical event (less precipitation than expected) and the demand people place on water supply. Human beings often exacerbate the impact of drought. Recent droughts in both developing and developed countries and the resulting economic and environmental impacts and personal hardships have underscored the vulnerability of all societies to this natural hazard (National Drought Mitigation Center, 2006).

Over the past century, the area affected by severe and extreme drought in the United States each year averages around 14% with the affected area as high as 65% in 1934. In recent years, the drought-affected area ranged between 35 and 40% as shown in Figure 1.10. FEMA (1995) estimates that average annual drought-related losses at $6-8 billion (based on relief payments alone). Losses were as high as $40 billion in 1988 (Riebsame *et al*., 1991). Available economic estimates of the impacts of drought are difficult to reproduce. This problem has to do with the unique nature of drought relative to other extremes, such as hurricanes. The onset of drought is slow. Further, the secondary impacts may be larger than the immediately visible impacts and often occur past the lifetime of the event (Wilhite and Pulwarty, 2005).

In recent years, the western United States has experienced considerable drought impacts, with 30% of the region under severe drought since 1995. Widespread declines in springtime snow water equivalent in the U.S. West have occurred over the period 1925–2000, especially since mid-century. While non-climatic factors, such as the growth of forest canopy, might be partly responsible, the primary cause is likely changing climate because the patterns of climatic trends are spatially consistent and the trends are dependent on elevation (Mote *et al*., 2005). Increased temperature appears to have led to increasing drought (Andreadis and Lettenmaier, 2006). In the Colorado River Basin, the 2000-2004 period had an average flow of 9.9 million acre feet[8] (maf) per year, lower than the driest period during the Dust Bowl years of (1931-35 with 11.4 maf), and the 1950s with (10.2 maf) (Pulwarty *et al*., 2005). For the winter of 2004-5, average precipitation in the Basin was around 100% of normal. However, the combination of low antecedent soil moisture (absorption into soil and depleted high mountain aquifers) and the warmest January-July period on record (driving evaporation) resulted in a reduced flow of 75% of average.

At the same time, states in the U.S. Southwest experienced some of the most rapid economic and population growth in the country, with attendant demands on water resources and associated conflicts. It is estimated that as a result of the 1999-2004 drought and increased water resources extraction, Lake Mead and Lake Powell[9] will take 13 to 15 years of average flow conditions to refill. In the Colorado River Basin, high-elevation snow pack contributes approximately 70% of the annual runoff. Because the

[8] One acre foot is equal to 325,853 U.S. gallons or 1233.5 cubic meters.
[9] Lake Mead and Lake Powell are reservoirs on the Colorado River. Lake Mead is the largest man-made lake in the United States.

Colorado River Compact[10] prioritizes the delivery of water to the Lower Basin states of Arizona, California, and Nevada, the largest impacts may be felt in the Upper Basin states of Wyoming, Utah, Colorado, and New Mexico. With increased global warming, the compact requirements may only be met 59% to 75% of the time (Christensen *et al*., 2004).

While there are multi-billion dollar estimates for annual agricultural losses (averaging about $4 billion a year over the last ten years), it is unclear whether these losses are directly related to crop production alone or other factors. Wildfire suppression costs to the United States Department of Agriculture (USDA) alone have surpassed $1 billion each of the last four years but it is unclear how much of this is attributable to dry conditions. Little or no official loss estimates exist for the energy, recreation/tourism, timber, livestock, or environmental sectors, although the drought impacts within these sectors in recent years is known to be large. Better methods to quantify the cumulative direct and indirect impacts associated with drought need to be developed. The recurrence of a drought today of equal or similar magnitude to major droughts experienced in the past will likely result in far greater economic, social, and environmental losses and conflicts between water users.

BOX 1.5: Hurricanes

There are substantial vulnerabilities from hurricanes along the Atlantic seaboard of the United States. Four major concentrations of economic vulnerability (capital stock greater

[10] The Colorado River Compact is a 1922 agreement among seven U.S. states in the basin of the Colorado River which governs the allocation of the river's water.

than $100 billion) are along the Miami coast, New Orleans, Houston, and Tampa. Three of these four areas have been hit by major storms in the last fifteen years (Nordhaus, 2006). A simple extrapolation of the current trend of doubling losses every ten years suggests that a storm like the 1926 Great Miami Hurricane could result in perhaps $500 billion in damages as early as the 2020s (Pielke *et al*., 2007; Collins and Lowe, 2001).

Property damages are well correlated to hurricane intensity. The formula for the kinetic energy of a moving object, be it a baseball or the wind, is one half the mass times the square of the speed. The mass of the wind in a hurricane does not change significantly. However because the kinetic energy increases with the square of the wind speed, faster winds have much more energy, dramatically increasing damages, as shown in Figure 1.11. Only 21% of the hurricanes making landfall in the United States are in Saffir-Simpson categories 3, 4, or 5, yet they cause 83% of the damage (Pielke and Landsea, 1998). Nordhaus (2006) argues that hurricane damage does not increase with the square of the wind speed as kinetic energy does, but rather, damage appears to rise with the eighth power of maximum wind speed. The 2005 total hurricane economic damage of $174 billion was primarily due to the cost of Katrina ($135 billion). As Nordhaus (2006) notes, 2005 was an economic outlier not because of extraordinarily strong storms but because the cost as a function of hurricane strength was high.

A fundamental problem within many economic impact studies lies in the unlikely assumption that there are no other influences on the macro-economy during the period analyzed for each disaster (Pulwarty *et al*., 2007). However, more is at work than aggregate indicators of population and wealth. It has long been known that different social groups, even within the same community, can experience the same climate event quite differently. In addition, economic analysis of capital stocks and densities does not capture the fact that many cities, such as New Orleans, represent unique corners of American culture and history (Kates *et al*., 2006). Importantly, the implementation of past adaptations (such as levees) actually conditions the degree of present and future impacts (Pulwarty *et al*., 2003). At least since 1979, the reduction of mortality over time has been noted, including drought in the United States and Africa, tropical cyclones in Bangladesh, and floods and hurricanes in the United States. On the other hand, a reduction in property damage is less clear because aggregate property damages have risen along with increases in the population, material wealth, and development in hazardous areas.

BOX 1.6: Impacts Tools

There are a variety of impact tools that help users translate climate information into an assessment of what the impacts will be and provide guidance on how to plan accordingly. These tools would be part of the filter/medium circle in Figure 1.8. However, as illustrated, using the example of a catastrophe risk model, the model has clear linkages to all the other boxes in Figure 1.8.

A catastrophe risk model can be divided into four main components, as shown in Figure 1.12. The hazard component provides information on the characteristics of a hazard. For probabilistic calculations, this component would include a catalog with a large number of simulated events with realistic characteristics and frequencies. Event information for each hazard would include the frequency, size, location, and other characteristics. The overall statistics should agree with an analysis of historical events.

The inventory component provides an inventory of structures that are exposed to a hazard and information on their construction. The vulnerability component simulates how structures respond to a hazard. This component requires detailed information on the statistical response of a structure to the forces produced by a hazard. This component would also account for secondary damage such as interior water damage after a structure's windows are breached. The fourth component in the risk model estimates losses produced by a hazard event and accounts for repair or replacement costs. In cases of insurance coverage, the loss component also accounts for business interruption costs and demand surge. If the model is used for emergency management purposes, the loss component also accounts for factors such as emergency supplies and shelters.

It should be noted, though, that how the loss component is treated impacts the vulnerability and inventory components, as indicated by the curved upward pointing arrows. Is a house destroyed in a flood rebuilt in the same location or on higher ground? Is a wind damaged building repaired using materials that meet higher standards? These actions have profound effects on future catastrophe risk models for the area.

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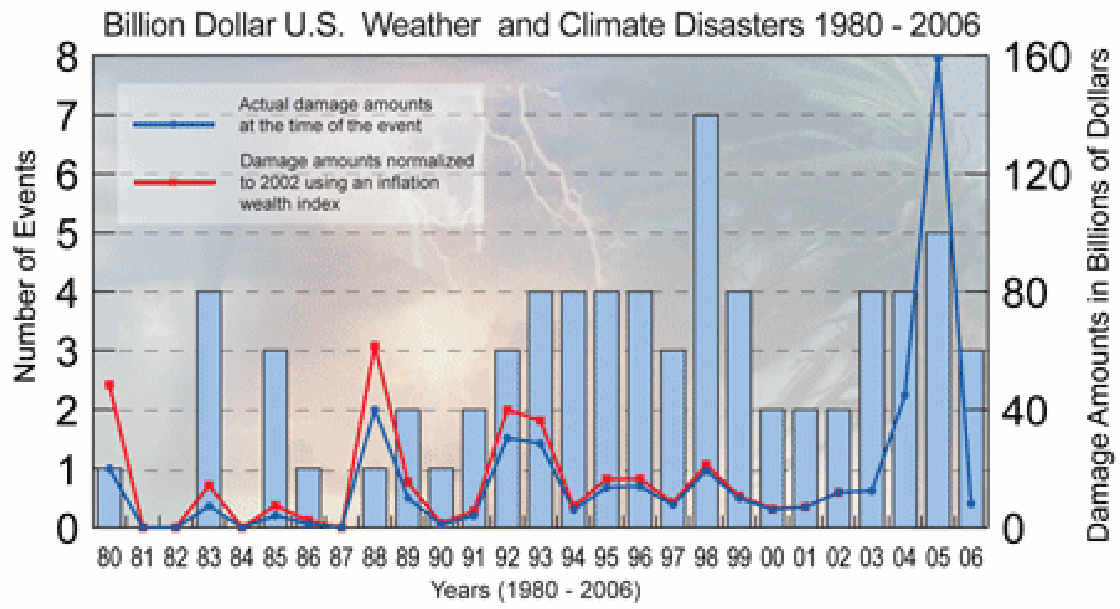

Figure 1.1 The blue bars show the number of events per year that exceed a cost of 1 billion dollars (these are scaled to the left side of the graph). The blue line (actual costs at the time of the event) and the red line (costs adjusted for wealth/inflation) are scaled to the right side of the graph, and depict the annual damage amounts in billions of dollars. Over the last 27 years, the U.S. averaged between two and three weather and climate-related disasters a year that exceeded one billion dollars in cost. Data from NOAA's National Climatic Data Center.

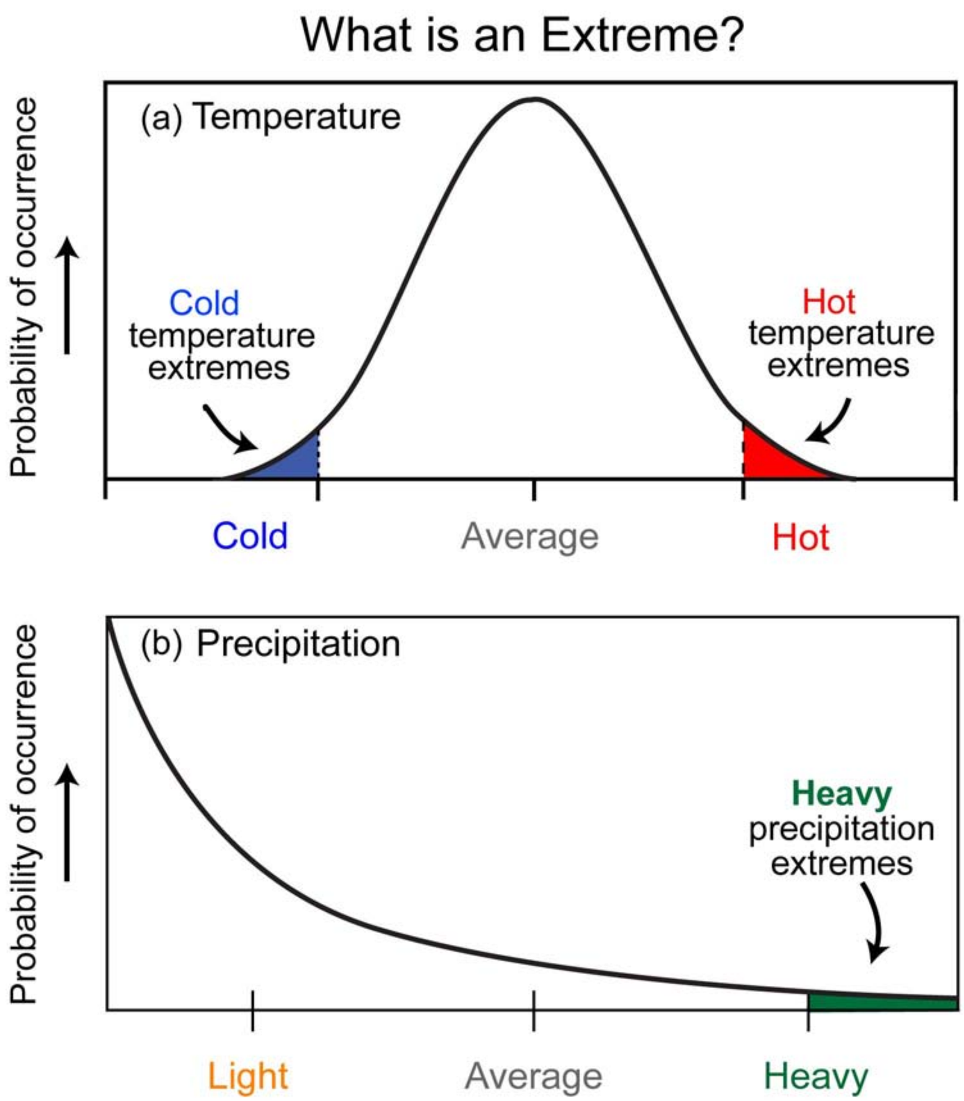

Figure 1.2 Probability distributions of daily temperature and precipitation. The higher the black line, the more often weather with those characteristics occurs.

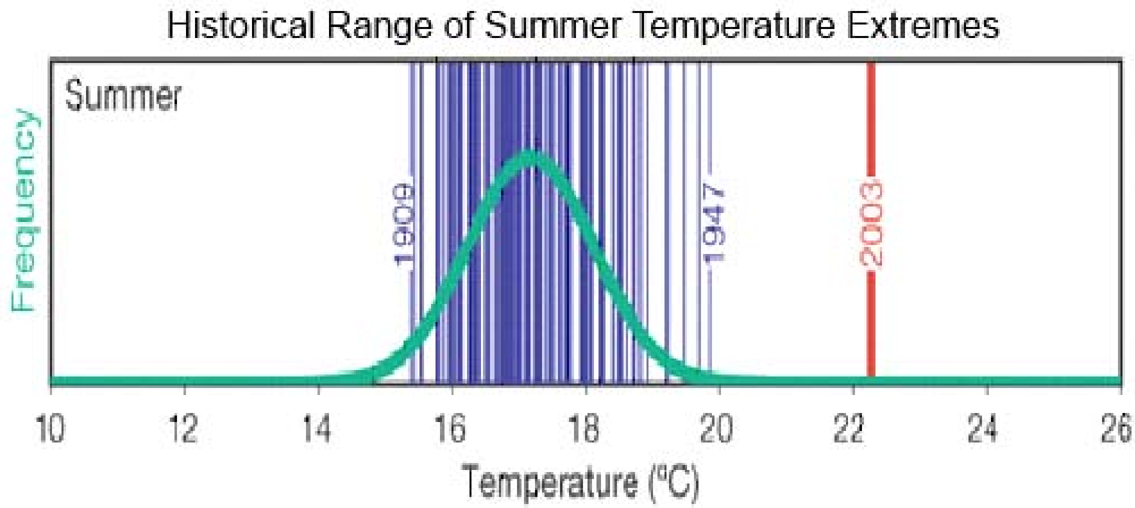

Figure 1.3 Like the European summer temperature of 2003, some extremes that are more likely to be experienced in the future will be far outside the range of historical observations. Each vertical line represents the summer temperature for a single year with the extreme values from the years 1909, 1947 and 2003 identified. From Schär *et al*., 2004.

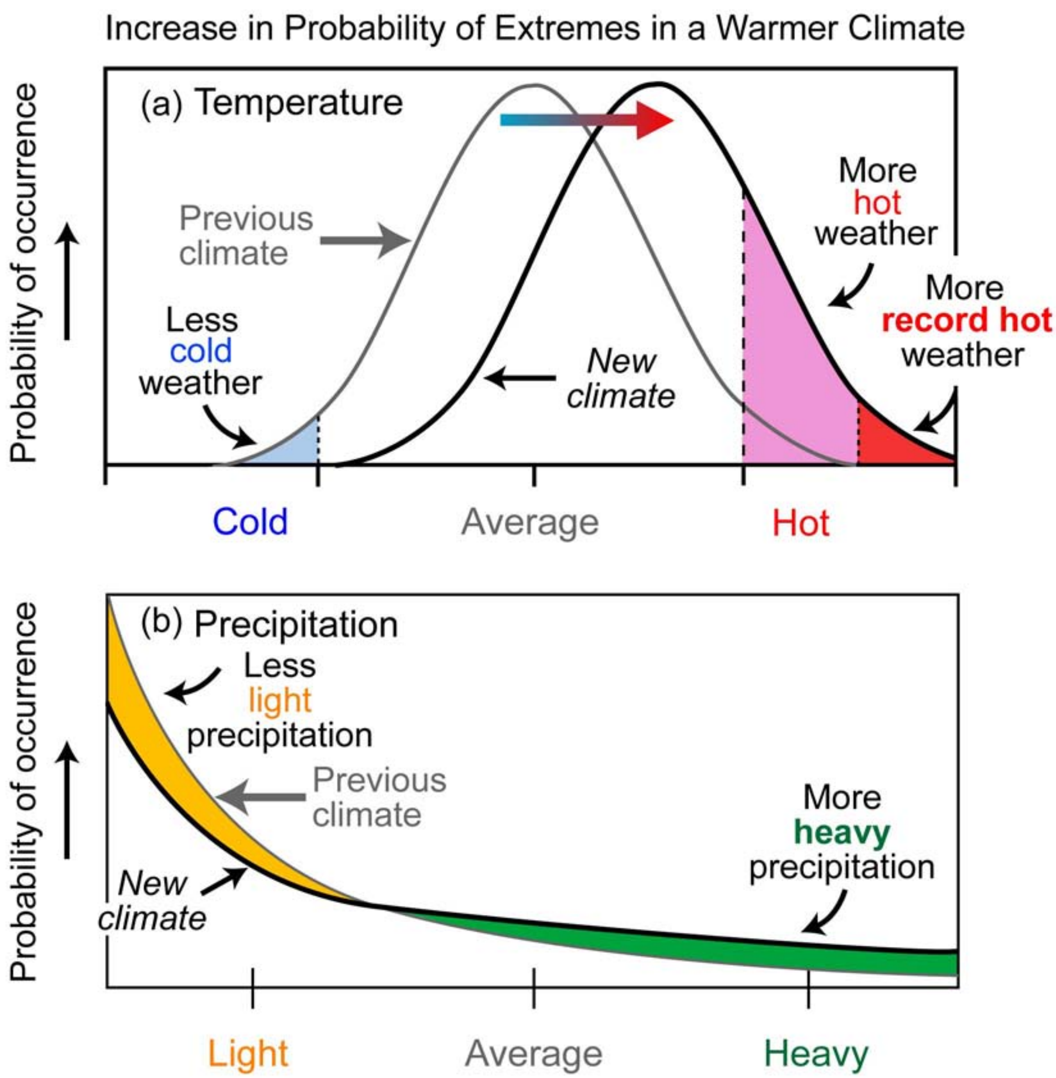

Figure 1.4 Simplified depiction of the changes in temperature and precipitation in a warming world.

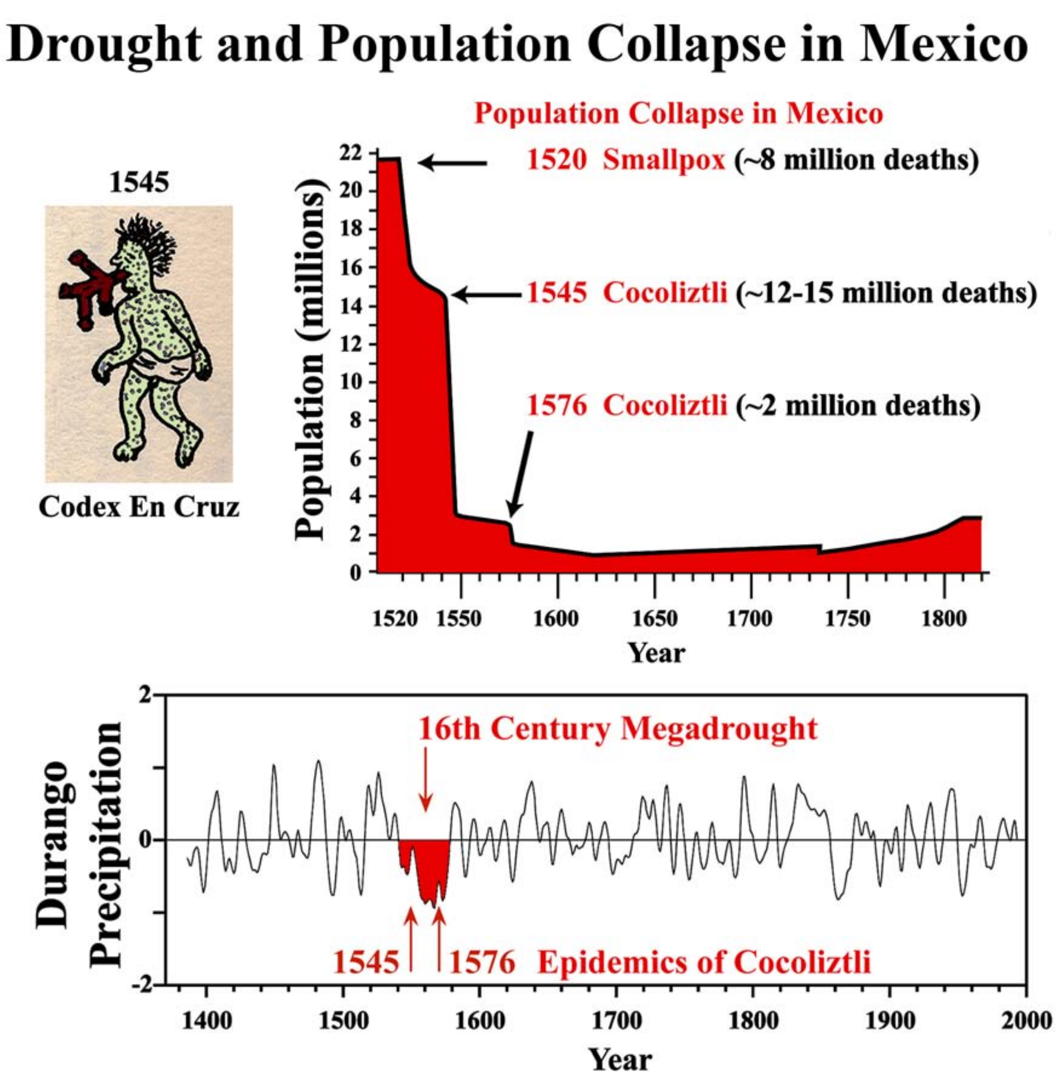

Figure 1.5 Megadrought and megadeath in 16th Century Mexico. Four hundred years ago the Mexican socio-economic and natural systems were so sensitive to extremes that a mega-drought in Mexico led to a massive population declines (Acuna-Soto *et al.*, 2002). The 1545 Codex En Cruz depicts the effects of the cocoliztli epidemic which has symptoms similar to rodent-borne hantavirus hemorrhagic fever.

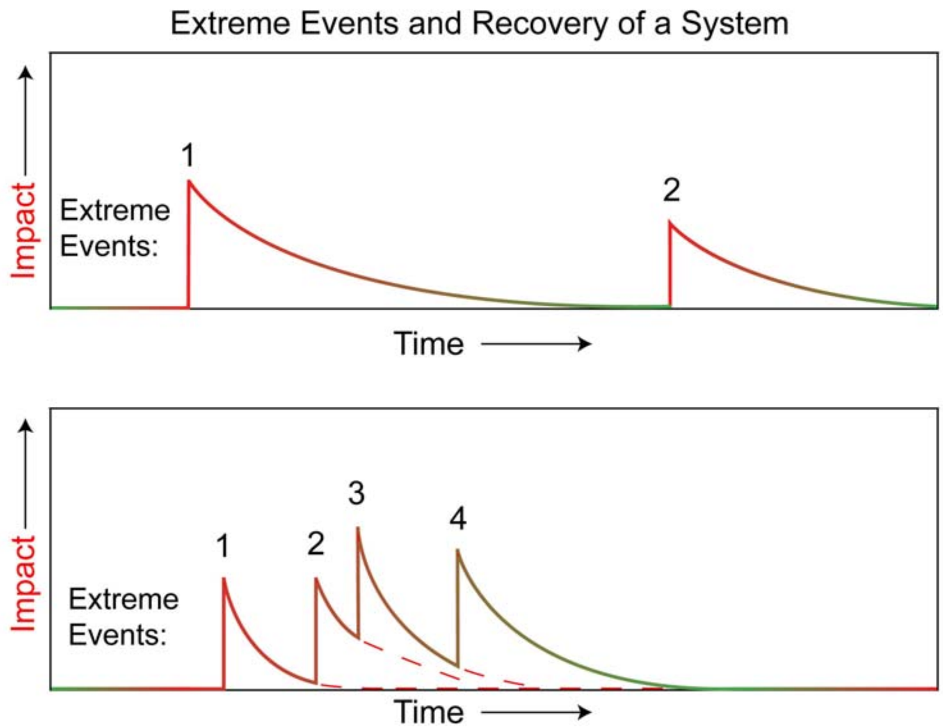

Figure 1.6 Extreme events such as hurricanes can have significant sudden impacts that take some time to recover from. Top: Two similar magnitude events take place but after the first one, new adaptation measures are undertaken, such as changes in building codes, so the second event doesn't have as great an impact. Bottom: An extreme that occurs before an area has completely recovered from the previous extreme can have a total impact in excess of what would have occurred in isolation.

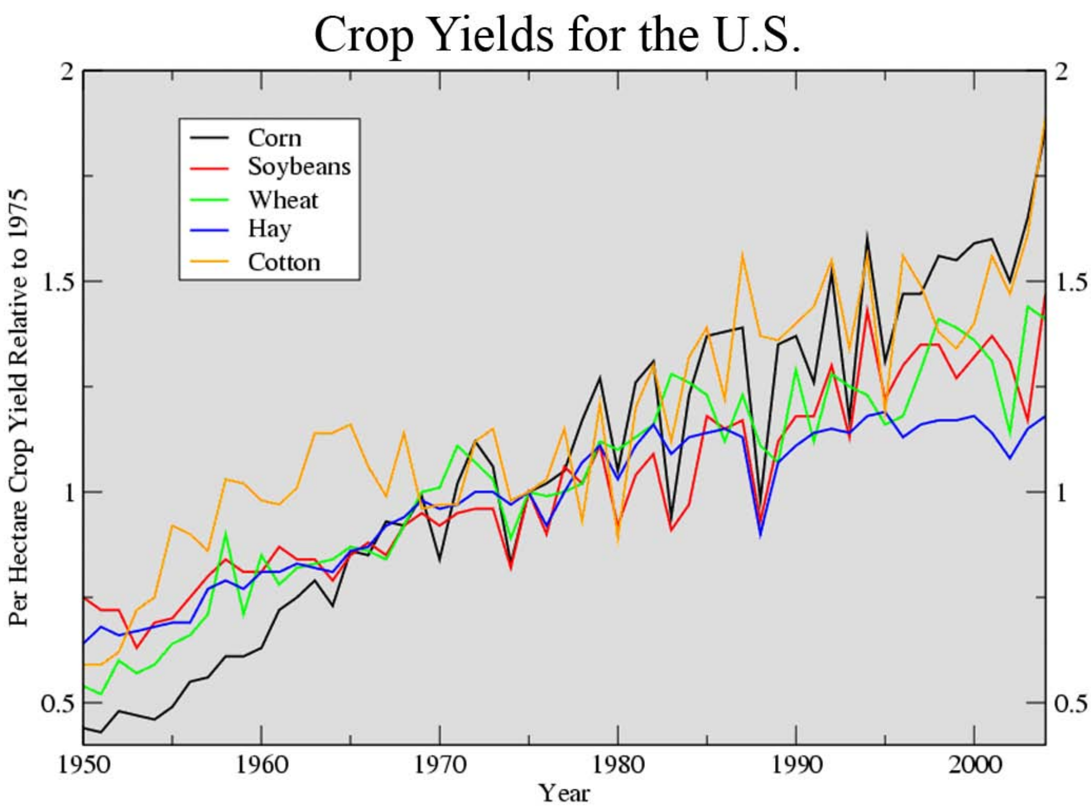

Figure 1.7 Climate variability may reduce crop yield, but because of technological improvements, a poor yield in the 1990s can still be higher than a good yield in the 1950s indicating a changing relationship between climate and agricultural yield. Data are in units of cubic meters or metric tons per unit area with the yield in 1975 defined as 1. Data from USDA National Agricultural Statistics Service via update to Heinz Center (2002).

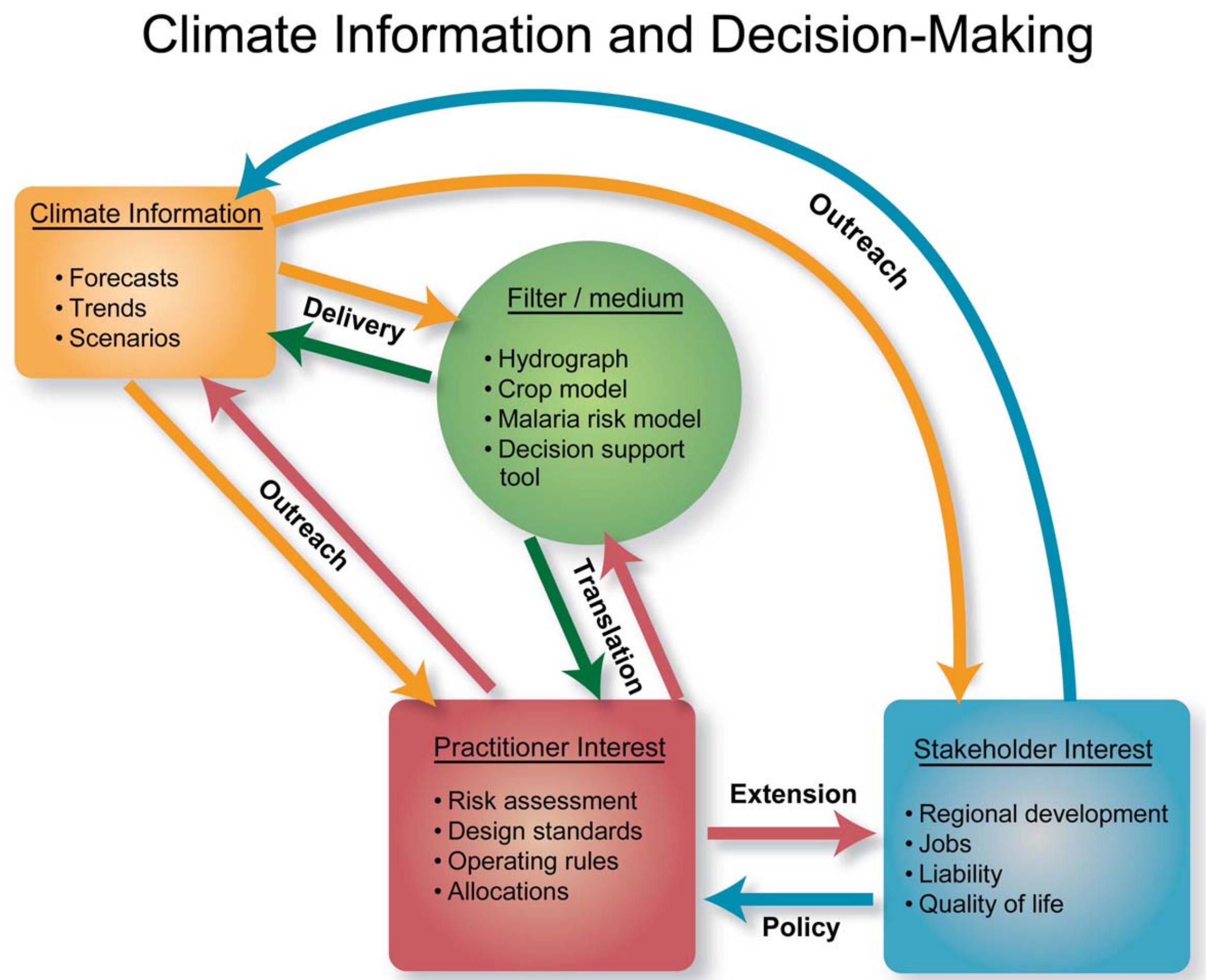

Figure 1.8 Illustration of how climate information is processed, filtered, and combined with other information in the decision process relevant to stakeholder interests.

Beetle Damage to Pine Trees in Canada

Figure 1.9 Photograph of a pine forest showing pine trees dying (red) from beetle infestation in the Quesnel-Prince George British Columbia area. Fewer instances of extreme cold winter temperatures that control beetle populations as well as hotter summers that increase populations are leading to a greater likelihood of beetle infestations. (Figure inclusion in Final Document subject to copyright permission).

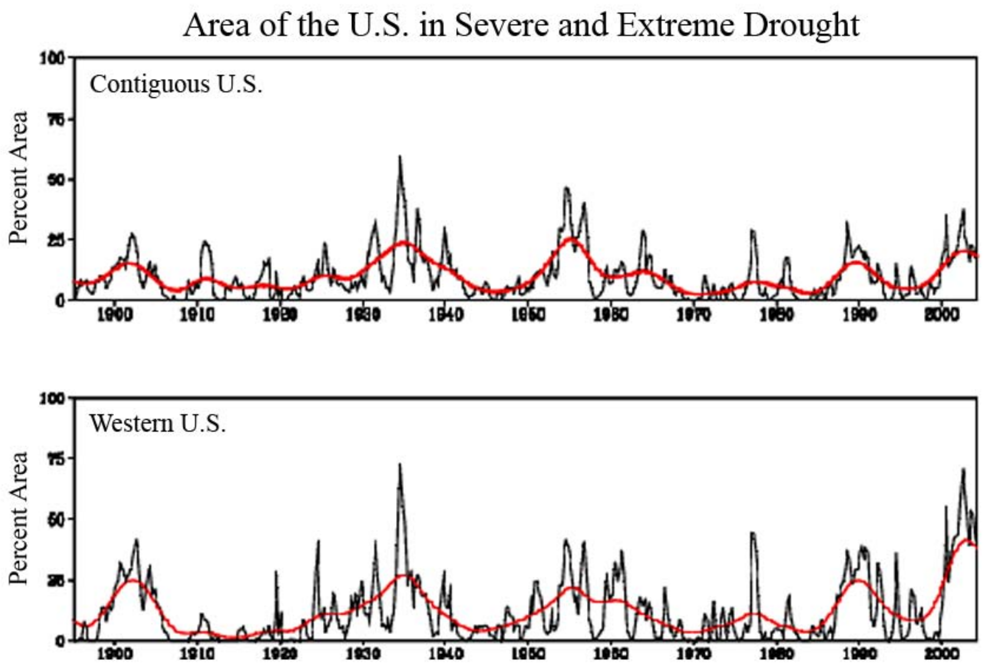

Figure 1.10 Percent of area in the contiguous U.S. and western U.S. affected by severe and extreme drought as indicated by Palmer Drought Severity Index (PDSI) values of less than or equal to −3. Data from NOAA's National Climatic Data Center.

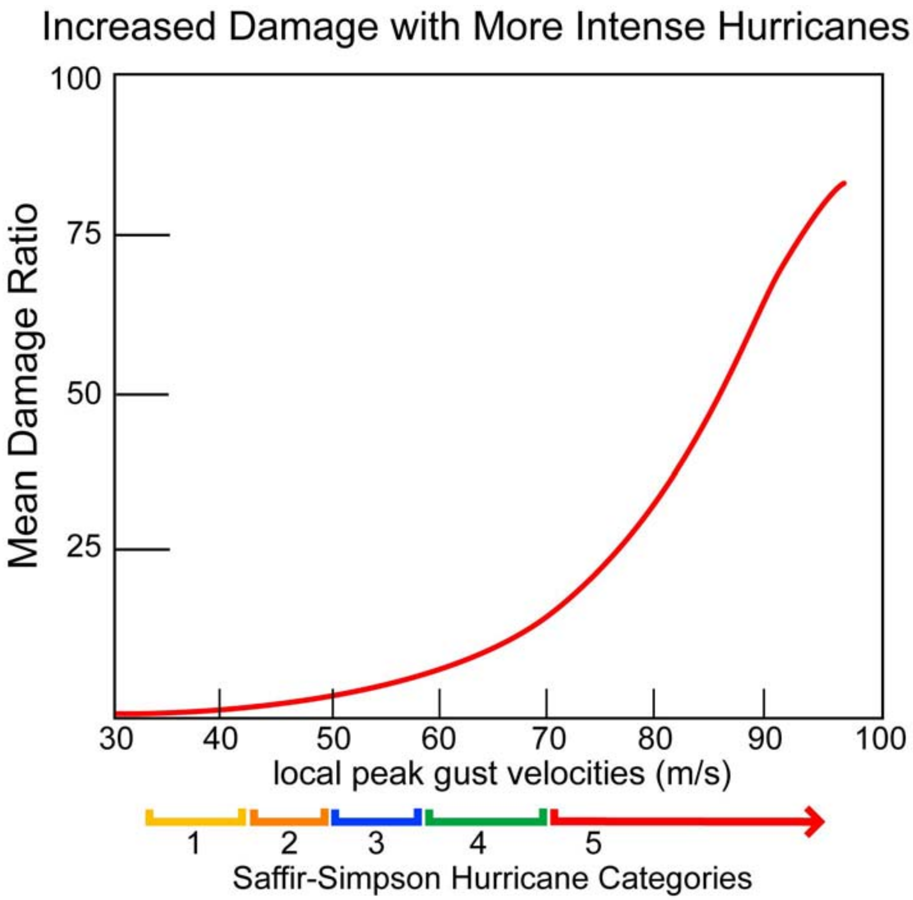

Figure 1.11 More intense hurricanes cause much greater losses. Mean damage ratio is the average expected loss as a percent of the total insured value. Adapted from Meyer *et al*. (1997).

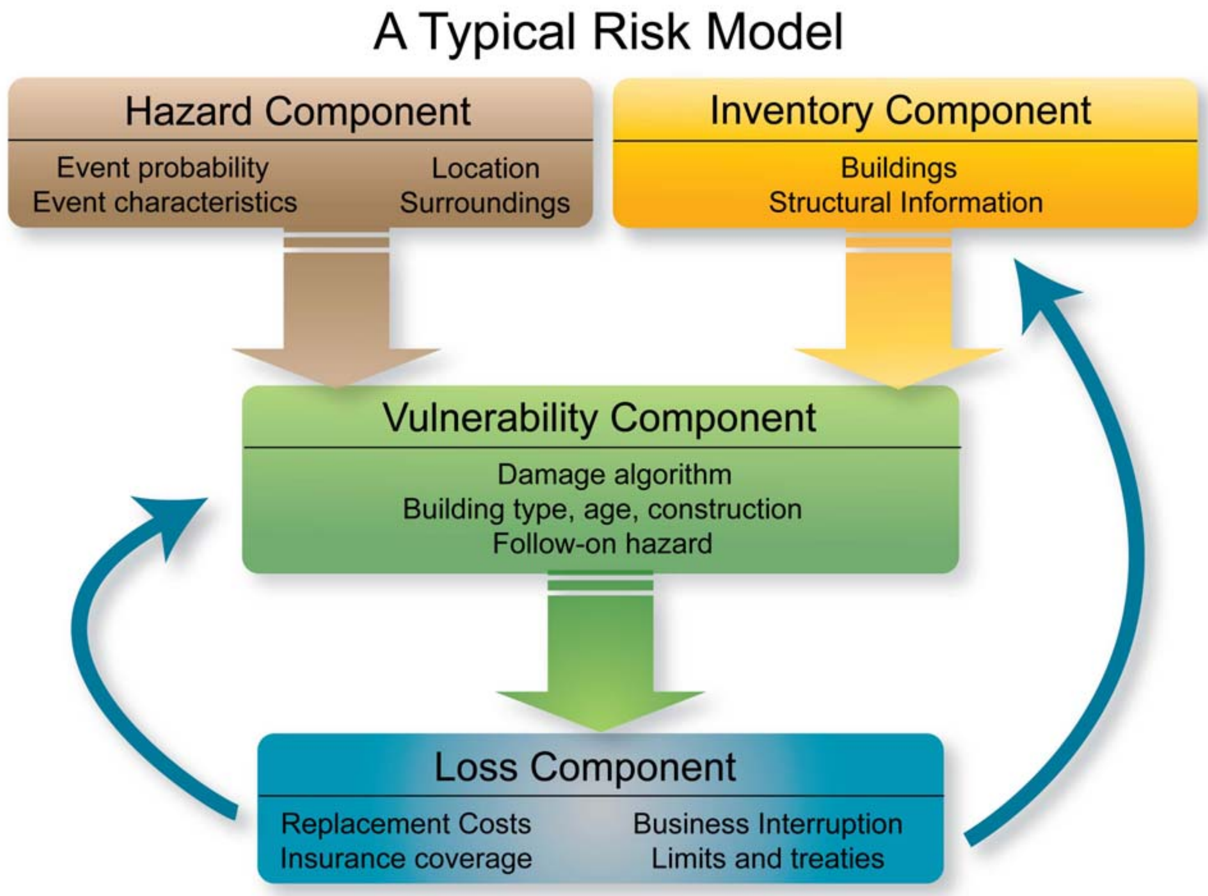

Figure 1.12 Schematic diagram of a typical risk model used by the insurance industry. The diagram highlights the three major components (hazard, damage, and loss) of a risk model. What happens to the loss component feedbacks to the vulnerability and inventory components.

Chapter 2 Observed Changes of Weather and Climate Extremes

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KEY FINDINGS

Observed Changes

Upward trends in the frequency of unusually warm nights, extreme precipitation episodes, the frequency of North Atlantic tropical cyclones (hurricanes), the length of the frost-free season, and extreme wave heights along the west coast are notable changes in the North American climate record.

- Most of North America is experiencing more unusually hot days. The number of warm spells has been increasing since 1950. However, the heat waves of the 1930s remain the most severe in the U.S. historical record back to 1895.

- There are fewer unusually cold days during the last few decades. The last 10 years have seen a lower number of severe cold waves than for any other 10-yr period in the historical record which dates back to 1895. There has been a decrease in the number of frost days and a lengthening of the frost-free season, particularly in the western part of North America.
- Extreme precipitation episodes (heavy downpours) have become more frequent and more intense in recent decades than at any other time in the historical record dating back to the late 19th Century and account for a larger percentage of total precipitation. The most significant changes have occurred in most of the U.S., northern Mexico, southeastern, northern and western Canada, and southern Alaska.
- There are recent regional tendencies toward more severe droughts in the southwestern U.S., parts of Canada and Alaska, and Mexico.
- For the continental U.S. and southern Canada, the most severe droughts occurred in the 1930s and there is no indication of an overall trend since 1895; in Mexico, the 1950s and 1994-present were the driest period.
- Atlantic tropical cyclone (hurricane) activity, as measured by both frequency and the Power Dissipation Index (which combines storm intensity, duration and frequency) has increased.

  – The increases are substantial since about 1970, and are likely substantial since the 1950s and 60s, in association with warming Atlantic sea surface temperatures. There is less confidence in data prior to 1900.

  – It is likely that there has been an increase in tropical cyclone *frequency* in the North Atlantic over the past 100 years, which has closely followed warming tropical Atlantic sea surface temperatures. There is increasing uncertainty in the data as one proceeds further back in time.

– The frequency of major hurricanes has increased coincident with overall tropical cyclone numbers.

- There is no observational evidence for an increase in North American mainland land-falling hurricanes since the late 1800s.
- The hurricane Power Dissipation Index in the eastern Pacific, affecting the Mexican west coast and shipping lanes, has decreased since 1980, but rainfall from near-coastal hurricanes has increased since 1949.
- The balance of evidence suggests that there has been a northward shift in the tracks of strong low pressure systems (storms) in both the N. Atlantic and N. Pacific basins. There is a trend toward stronger intense low pressure systems in the North Pacific.
- Increases in extreme wave height characteristics have been observed along the Atlantic and Pacific coasts of North America during recent decades based on 3 decades of buoy data.

  – Increases along the West coast have been greatest in the Pacific Northwest, and are likely a reflection of changes in storm tracks.

  – Increases along the U.S. east coast are evident during the hurricane season.
- Although snow cover extent has decreased over North America, there is no indication of continental-scale trends in snowstorms and episodes of freezing rain during the 20th Century.
- There is no trend in the frequency of tornadoes and other severe convective storms when the data are adjusted for changes in observing practices.

2.1 Background

Weather and climate extremes exhibit substantial spatial variability. It is not unusual for severe drought and flooding to occur simultaneously in different parts of North America (e.g. catastrophic flooding in the Mississippi River basin and severe drought in the southeast U.S. during summer 1993). These reflect temporary shifts in large-scale circulation patterns that are an integral part of the climate system (Chapter 2, Box 2.3). The central goal of this chapter is to identify long-term shifts/trends in extremes and to characterize the continental-scale patterns of such shifts. Such characterization requires data that is homogeneous, of adequate length, and with continental-scale coverage. Many datasets meet these requirements for limited periods only. For temperature and precipitation, rather high quality data are available for the conterminous U.S. back to the late 19th Century. However, shorter data records are available for parts of Canada, Alaska, Hawaii, Mexico, the Caribbean, and U.S. territories. In practice, this limits true continental-scale analyses of temperature and precipitation extremes to the middle part of the 20th Century onward. Other phenomena have similar limitations and continental-scale characterizations are generally limited to the last 50-60 years or less, or must confront data homogeneity issues which add uncertainty to the analysis. We consider all studies that are available, but in many cases these studies have to be interpreted carefully because of these limitations. A variety of statistical techniques are used in the studies cited here. General information about statistical methods along with several illustrative examples are given in the Appendix.

2.2 Observed Changes and Variations in Weather and Climate Extremes

2.2.1 Temperature Extremes

Extreme temperatures do not always correlate with average temperature, but they often change in tandem; thus, average temperature changes provide a context for discussion of extremes. In 2005, virtually all of North America was above to much above average[11] (Shein et al. 2006) and 2006 was one of the warmest years on record in the conterminous United States (Arguez et al., 2007). The areas experiencing the largest temperature anomalies included the higher latitudes of Canada and Alaska. Annual average temperature time series for Canada, Mexico and the United States all show substantial warming since the middle of the 20^{th} century (Shein et al. 2006). Since 1998 over half of the U.S. annual average temperatures have been extremely high, including the hottest two years on record (1998 and 2006).

Since 1950, the annual percent of days exceeding the 90^{th}, 95^{th}, and 97.5^{th} percentile thresholds[12] for both maximum (daytime highs) and minimum (nighttime lows) temperature has increased when averaged over all the land area (Figure 2.1; Peterson et al. 2007). Although the changes are greatest in the 90^{th} percentile (increasing from about 10% of the days to about 13% for maximum and almost 15% for minimum) and decrease as the threshold temperatures increase indicating more rare events (the 97.5^{th} percentage increases from about 3% of the days to 4% for maximum and 5% for minimum), the

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[11] NOAA's National Climatic Data Center uses the following terminology for classifying its monthly/seasonal/annual U.S. temperature and precipitation rankings: "near-normal" is defined as within the *mid-tercile*, "above/below normal" is within the *top-tercile*, and "much-above/much-below normal" is within the *top-decile* of all such periods on record.

[12] An advantage of the use of percentile, rather than absolute, thresholds is that they account for regional climate differences

relative changes are similar. There are important regional differences in the changes. For example, the largest increases in the 90th percentile threshold temperature occur in the western part of the continent from northern Mexico through the western U.S. and Canada and across Alaska, while some areas, such as eastern Canada, show declines of as many as 10 days per year from 1950 to 2004 (Fig. 2.2).

Other regional studies have shown similar patterns of change. For the U.S., the number of days exceeding the 90th, 95th and 99th percentile thresholds (defined monthly) have increased in recent years[13], but are also dominated earlier in the 20th century by the extreme heat and drought of the 1930s[14] (DeGaetano and Allen 2002). Changes in cold extremes (days falling below the 10th, 5th, and 1st percentile threshold temperatures) show decreases, particularly since 1960[15]. For the 1900-1998 period in Canada, there are fewer cold extremes in winter, spring and summer in most of southern Canada and more high temperature extremes in winter and spring, but little change in warm extremes in summer[16] (Bonsal et al. 2001). However, for the more recent (1950-1998) period there are significant increases in warm extremes over western Canada, but decreases in eastern Canada. Similar results averaged across all of Canada are found for the longer 1900-2003 period, with 28 fewer cold nights, 10 fewer cold days, 21 more extreme warm nights and 8 more warm days per year now than in 1900[17] (Vincent and Mekis 2006). For the U.S. and Canada, the largest increases in daily maximum and minimum temperature are

[13] The number of stations with statistically significant positive trends for 1960-1996 passed tests for field significance based on resampling.
[14] The number of stations with statistically significant negative trends for 1930-1996 was greater than the number with positive trends.
[15] The number of stations with statistically significant downward trends for 1960-1996 passed tests for field significance based on resampling, but not for 1930-1996.
[16] Statistical significance of trends was assessed using Kendall's tau test
[17] These trends were statistically significant at more than 20% of the stations based on Kendall's tau test

occurring in the colder days of each month (Robeson 2004). For the Caribbean region, there is an 8% increase in the number of very warm nights and 6% increase in the number of very warm days for the 1958-1999 period. There also has been a corresponding decrease of 7% in the number of cold days and 4% in the number of cold nights (Peterson et al. 2002). The number of warm nights has increased by 10 or more per year for Hawaii and 15 or more per year for Puerto Rico from 1950 to 2004 (Fig. 2.2).

Analysis of multi-day very extreme heat and cold episodes[18] in the U.S. were updated[19] from Kunkel et al. (1999) for the period 1895-2005. The most notable feature of the pattern of the annual number of the extreme heat waves (Fig. 2.3a) through time is the high frequency in the 1930s compared to the rest of the years in the 1895-2005 period. This was followed by a decrease to a minimum in the 1960s and 1970s and then an increasing trend since then. There is no trend over the entire period, but a highly statistically significant upward trend since 1960. The heat waves during the 1930s were characterized by extremely high daytime temperatures while nighttime temperatures were not as unusual (Fig. 2.3b,c). An extended multi-year period of intense drought undoubtedly played a large role in the extreme heat of this period, particularly the daytime temperatures, by depleting soil moisture and reducing the moderating effects of evaporation. By contrast, the recent period of increasing heat wave index is distinguished by the dominant contribution of a rise in extremely high nighttime temperatures (Fig. 2.3c). Cold waves show a decline in the first half of the 20th century, then a large spike of

[18] The threshold is approximately the 99.9 percentile.

[19] The data were first transformed to create near-normal distributions using a log transformation for the heat wave index and a cube root transformation for the cold wave index. The transformed data were then subjected to least squares regression. Details are given in the Appendix, Example 2.

events during the mid-1980s, then a decline[20]. The last 10 years have seen a lower number of severe cold waves in the U.S. than in any other 10-yr period since 1895, consistent with observed impacts such as insect populations (Chapter 1, Box 1.2). Decreases in the frequency of extremely low nighttime temperatures have made a somewhat greater contribution than extremely low daytime temperatures to this recent low period of cold waves. Over the entire period there is a downward trend but it is not statistically significant at the p=0.05 level.

The annual number of warm spells[21] averaged over North America has increased since 1950 (Peterson et al. 2007). In the U.S. the annual number of warm spells[22] has increased by about 1 ½ per year, and the duration has increased by about 1 day since 1950 (Easterling et al. 2007a). Regionally the largest increases, up to about 2 ½ per year, were found in the western U.S., with many parts of the south and southeast showing little change. Seasonal results show the largest increases in the spring and winter, with little change in the number of events for the fall or summer. These results for warm spells are roughly consistent with those for the much more extreme heat waves illustrated in Fig. 2.3a for the common period of analysis (1950-present); the warm spell analyses do not extend back to the 1930s when very extreme heat was frequent. The frequency and extent of hot summers[23] was highest in the 1930s, 1950s, and 1995-2003; the geographic pattern

[20] Details of this analysis are given in the Appendix, Example 1.
[21] Defined as at least 3 consecutive days above the 90th percentile threshold done separately for maximum and minimum temperature.
[22] Defined as at least 3 consecutive days with both the daily maximum and succeeding daily minimum temperature above the 80th percentile.
[23] Based on percentage of North American grid points with summer temperatures above the 90th or below the 10th percentiles of the 1950-1999 summer climatology.

of hot summers during 1995-2003 was similar to that of the 1930s (Gershunov and Douville 2007).

The occurrence of temperatures below the biologically- and societally-important freezing threshold (0°C, 32°F) is an important aspect of the cold season climatology. Studies have typically characterized this either in terms of the number of frost days (days with the minimum temperature below freezing) or the length of the frost-free season[24]. The number of frost days decreased by 4 days per year in the U.S. during the 1948-1999 period, with the largest decreases, as many as 13 days per year, occurring in the Western U.S.[25] (Easterling 2002). In Canada, there have been significant decreases in frost day occurrence over the entire country from 1950 to 2003, with the largest decreases in extreme western Canada where there have been decreases of up to 40 or more frost days per year, and slightly smaller decreases in eastern Canada (Vincent and Mekis 2006). The start of the frost-free season in the Northeastern U.S. occurred 11 days earlier in the 1990s than in the 1950s (Cooter and LeDuc 1995). For the entire U.S., the average length of the frost-free season over the 1895-2000 period for the U.S. increased by almost 2 weeks[26] (Figure 2.4; Kunkel et al. 2004). The change is characterized by 4 distinct regimes, with decreasing frost-free season length from 1895 to1910, an increase in length of about 1 week from 1910 to 1930, little change during1930-1980, and large increases since1980. The frost-free season length has increased more in the western U.S. than in the eastern U.S. (Easterling 2002; Kunkel et al. 2004), which is consistent with the

[24] The difference between the date of the last spring frost and the first fall frost
[25] Trends in the western half of the U.S. were statistically significant based on simple linear regression
[26] Statistically significant based on least-squares linear regression

finding that the spring pulse of snow melt water in the Western U.S. now comes as much as 7-10 days earlier than in the late 1950s (Cayan et al. 2001).

Ice cover on lakes and the oceans is a direct reflection of the number and intensity of cold, below freezing days. Ice cover on the Laurentian Great Lakes of North American usually forms along the shore and in shallow areas in December and January, and in deeper mid-lake areas in February due to their large depth and heat storage capacity. Ice loss usually starts in early to-mid-March and lasts through mid-to-late April (Assel 2003).

Annual maximum ice cover on the Great Lakes has been monitored since 1963. The maximum extent of ice cover over the past 4 decades varied from less then 10% to over 90%. The winters of 1977-1982 were characterized by a higher ice cover regime relative to the prior 14 winters (1963-1976) and the following 24 winters (1983-2006) (Assel et al. 2003, Assel 2005a, Assel personal communication for winter 2006). A majority of the mildest (lowest) seasonal average ice cover winters (Assel 2005b) over the past 4 decades occurred during the most recent 10-year period (1997-2006). Analysis of ice breakup dates on other smaller lakes in North America with at least 100 years of data (Magnuson et al. 2000) show a uniform trend toward earlier breakup dates (up to 13 days earlier per 100 years)[27].

Reductions in Arctic sea ice, especially near-shore sea ice, allow strong storm and wave activity to produce extensive coastal erosion resulting in extreme impacts. Observations from satellites starting in 1978 show that there has been a substantial decline in Arctic sea

[27] Statistically significant trends were found for 16 of 24 lakes

ice, with a statistically significant decreasing trend in annual Arctic sea ice extent of -33 ± 8.8 x 10^3 km^2 per year (equivalent to approximately -2.7% ± 0.7% per decade). Seasonally the largest changes in Arctic sea ice have been observed in the ice that survives the summer, where the trend in the minimum Arctic sea ice extent, between 1979 and 2005, was -60 ± 24 x 10^3 km^2 per year (–7.4 ± 2.9% per decade) (Lemke et al. 2007).

Rising sea surface temperatures have led to an increase in the frequency of extreme high SST events causing coral bleaching (see Box 1.1, Chapter 1). Mass bleaching events were not observed prior to 1980. However, since the 1970s, there have been 6 major global cycles of mass bleaching, with increasing frequency and intensity (Hoegh-Guldberg 2005). Almost 30% of the world’s coral reefs have disappeared in that time.

Less scrutiny has been focused on Mexico temperature extremes, in part, because much of the country can be classified as a ‘tropical climate’ where temperature changes are presumed fairly small, or semi-arid to arid climate where moisture availability exerts a far greater influence on human activities than does temperature.

Most of the sites in Mexico’s oldest temperature observing network are located in major metropolitan areas and there is considerable evidence to indicate that trend behaviors at least partly reflect urbanization and urban heat island influences (Englehart and Douglas, 2003). To avoid such issues in analysis, a monthly rural temperature dataset has recently been developed[28]. Examined in broad terms as a national aggregate, a couple of basic behaviors emerge. First, long period temperature trends over Mexico are generally compatible with continental-scale trends which indicate a cooling trend over North America from about the mid-1940s to the mid-1970s, with a warming trend thereafter.

The rural gridded data set indicates that much of Mexico experienced decreases in both T_{max} and T_{min} during 1941-1970 (-0.27°C/decade for T_{max} and -0.19°C/decade for T_{min}) while the later period of 1971-2001 is dominated by positive trends that are most strongly evident in T_{max} (0.35°C/decade for T_{max} and 0.10°C/decade for T_{min}). Based on these results it appears very likely that much of Mexico has experienced an increase in average temperature driven in large measure by increases in T_{max}. The diurnal temperature range (T_{max} minus T_{min}) for the warm season (June-September) averaged over all of Mexico has increased by 0.26°C/decade since 1970 with particularly rapid rises since 1990 (Fig. 2.5) reflecting a comparatively rapid rise in T_{max} with respect to T_{min} (Englehart and Douglas 2005)[29]. This behavior departs from the general picture for many regions of the world, where warming is attributable mainly to a faster rise in T_{min} than in T_{max} (e.g. Easterling et al., 1997).

Given Mexico's largely tropical/sub-tropical climate and the influence of nearby oceans, a reasonable expectation would be that changes in the behavior of temperature extremes

[28] It consists of monthly historical surface air temperature observations (1940-2001) compiled from stations (n=103) located in places with population <10,000 (2000 Census). To accommodate variable station record lengths and missing monthly observations, the dataset is formatted as a grid-type (2.5° x 2.5° lat.-long.) based on the climate anomaly method (Jones and Moberg, 2003)
[29] Statistically significant trends were found in the northwest, central, and south, but not the northeast regions

could be small and difficult to detect as compared to at many mid-and high latitude locations. However, the cold surge[30] phenomena – the equatorward penetration of modified cold air, known as *nortes* in Mexico – is an integral part of the country's cool season climatology. The frequency of both cold surge days and cold surge events tends to vary depending in part on Pacific Decadal Oscillation (PDO) phase: under negative PDO phase cold surge activity tends to be more prevalent. However, the intensity of cold surge events as indicated by the maximum daily drop in T_{min} tends to be greater under positive PDO phase. Analysis of linear trends indicates that from the early 1950s onward, it is very likely that southern Mexico has experienced a trend toward decreasing frequency of both cold surge days by 2.4 cold days/decade and cold surge events by 0.88 events/decade (Englehart and Douglas 2007).

2.2.2 Precipitation Extremes

2.2.2.1 Drought

Droughts are one of the most costly natural disasters (Chapter 1, Box 1.4), with estimated annual U.S. losses of $6 – 8 billion (Federal Emergency Management Agency, 1995). An extended period of deficient precipitation is the root cause of a drought episode, but the intensity can be exacerbated by high evaporation rates arising from excessive temperatures, high winds, lack of cloudiness, and/or low humidity. Drought can be defined in many ways, from acute short-term to chronic long-term hydrological drought, agricultural drought, meteorological drought, and so on. The assessment in this report focuses mainly on meteorological droughts based on the Palmer (1965) Drought Severity Index (PDSI), though other indices are also documented in the report (Chapter 2, Box 2.1).

[30] Cold surges are defined for the period 1925-2002 based on daily station observations of T_{min} from two locations – stations in south Texas and near coastal stations from the southern Mexican state of Veracruz. Cold surge days have T_{min} below its climatological values by 1 standard deviation. Cold surge events are runs of 1 or more consecutive cold surge days.

Individual droughts can occur on a range of spatial scales, but they often affect rather large areas and can persist for many months and even years. Thus, the aggregate impacts can be very large. For the U.S., the percentage area affected by severe to extreme drought (Fig. 2.6) highlights some major episodes of extended drought. The most widespread and severe drought conditions occurred in the 1930s and 1950s (Andreadis et al. 2005). The early 2000s were also characterized by severe droughts in some areas, notably in the western U.S. When averaged across the entire U.S. (Fig. 2.6), there is no clear tendency for a trend based on the PDSI. Similarly, long-term trends (1925-2003) of hydrologic droughts based on model derived soil moisture and runoff show that droughts have, for the most part, become shorter, less frequent, and cover a smaller portion of the U. S. over the last century (Andreadis and Lettenmaier, 2006). The main exception is the Southwest and parts of the interior of the West, where increased temperature has led to positive drought trends (Andreadis and Lettenmaier, 2006). The trends averaged over all of North America since 1950 (Fig. 2.6) are similar to U.S. trends for the same period, indicating no overall trend.

Since the contiguous United States has experienced an increase in both temperature and precipitation during the 20th century, one question is whether these increases are impacting the occurrence of drought. Easterling et al (2007b) examined this possibility by looking at drought, as defined by the PDSI, for the United States using detrended temperature and precipitation. Results indicate that without the upward trend in precipitation the increase in temperatures would have lead to an increase in the area of the U.S. in severe-extreme drought of up to 30% in some months.

Summer conditions, which relate to fire danger, have trended toward lesser drought in the upper Mississippi, Midwest, and Northwest, but the fire danger has increased in the Southwest, in California in the spring season (not shown), and, surprisingly, over the Northeast, despite the fact that annual precipitation here has increased. A century-long warming in this region is quite significant in summer, which reverses the tendencies of the precipitation contribution to soil wetness (Groisman et al. 2004). Westerling *et al.* (2006) document that large wildfire activity in the Western U.S. increased suddenly and markedly in the mid-1980s, with higher large-wildfire frequency, longer wildfire durations, and longer wildfire seasons. The greatest increases occurred in mid-elevation, Northern Rockies forests, where land-use histories have relatively little effect on fire risks and are strongly associated with increased spring and summer temperatures and an earlier spring snowmelt.

For the entire North American continent, there is a north-south pattern in drought trends (Dai et al. 2004). Since 1950, there is a trend toward wetter conditions over much of the conterminous U.S., but a trend toward drier conditions over southern and western Canada, Alaska, and Mexico. The summer PDSI averaged for Canada indicates dry conditions during the 1940s and 1950s, generally wet conditions from the 1960s to 1995, but much drier after 1995 (Shabbar and Skinner, 2004). In Alaska and Canada, the upward trend in temperature, resulting in increased evaporation rates, has made a substantial contribution to the upward trend in drought (Dai et al. 2004). In agreement with this drought index analysis, the area of forest fires in Canada has been quite high since 1980 compared to the previous 30 years and Alaska experienced a record high year for forest fires in 2004 followed by the third highest in 2005 (Soja et al. 2007). During the mid-1990s and early 2000s, central (Stahle *et al.* 2007) and western Mexico (Kim *et al.* 2002; Nicholas and Battisti, 2006; Hallack and Watkins, 2007) experienced continuous cool-season droughts having major impacts in agriculture, forestry, and ranching, especially during the warm summer season. In 1998, "El Niño" caused one of the most severe droughts in Mexico since the 1950s (Ropelewski, 1999), creating the most difficult wildfire season in Mexico's history. Mexico had 14,445 wildfires affecting 849,632 hectares - the largest area ever burned in Mexico in a single season (SEMARNAP, 2000).

Reconstructions of drought prior to the instrumental record based on tree-ring chronologies indicate that the 1930s may have been the worst drought since 1700 (Cook et al. 1999). There were three major multiyear droughts in the U.S. during the latter half of the 1800s: 1856-1865, 1870-1877 and 1890-1896 (Herweijer et al. 2006). Similar droughts have been reconstructed for northern Mexico (Therrell et al. 2002). There is evidence of earlier, even more intense drought episodes (Woodhouse and Overpeck 1998). A period in the mid to late 1500s has been termed a "mega-drought" and was longer-lasting and more widespread than the 1930s Dust Bowl (Stahle et al. 2000). Several additional mega-droughts occurred during 1000-1470 (Herweijer et al. 2007). These droughts were about as severe as the 1930s Dust Bowl episode but much longer, lasting 20-40 years. In the western U.S., the period of 900-1300 was characterized by widespread drought conditions (Fig. 2.7; Cook et al. 2004). In Mexico, reconstructions of seasonal precipitation (Stahle et al. 2000, Acuña-Soto et al. 2002, Cleaveland et al. 2004) indicate that there have been droughts more severe than the 1950s drought, e.g., the mega-drought in the mid- to late- 16th century, which appears as a continental-scale drought.

During the summer months, excessive heat and drought often occur simultaneously because the meteorological conditions typically causing drought are also conducive to high temperatures. The impacts of the Dust Bowl droughts and the 1988 drought were compounded by episodes of extremely high temperatures. The month of July 1936 in the central U.S. is a notable example. To illustrate, Lincoln, NE received only 0.05” of precipitation that month (after receiving less than 1 inch the previous month) while experiencing temperatures reaching or exceeding 110°F on 10 days, including 117°F on July 24. Although no studies of trends in such “compound” extreme events have been performed, they represent a significant societal risk.

BOX 2.1: Drought Indicators and Resources

- **Palmer Drought Severity Index (PDSI; Palmer, 1965) –** meteorological drought. The PDSI is a commonly-used drought index that measures intensity, duration, and spatial extent of drought. It is derived from measurements of precipitation, air temperature, and local estimated soil moisture content. Categories range from less than -4 (extreme drought) to more than +4 (extreme wet conditions), and have been standardized to facilitate comparisons from region to region. Alley (1984) identified some positive characteristics of the PDSI that contribute to its popularity: (1) it is an internationally recognized index; (2) it provides decision makers with a measurement of the abnormality of recent weather for a region; (3) it provides an opportunity to place current conditions in historical perspective; and (4) it provides spatial and temporal representations of historical droughts. However, the PDSI has some limitations (1) it may lag emerging droughts by several months; (2) it is less well suited for mountainous land or areas of frequent climatic extremes; (3) it does not take into account streamflow, lake and reservoir levels, and other long-term hydrologic impacts (Karl and Knight, 1985), such as snowfall and snow cover; (4) the use of temperature alone to estimate potential evapotranspiration (PET) can introduce biases in trend estimates because humidity, wind and radiation also affect PET and changes in these elements are not accounted for. In fact, Hobbins et al. (2007) show that the PDSI trends in Australia and New Zealand are exaggerated compared to trends using more realistic methods to estimate evapotranspiration. The use of temperature alone is a practical consideration since measurements of these other elements are often not available.

- **Crop Moisture Index (CMI; Palmer, 1968) –** short-term meteorological drought. Whereas the PDSI monitors long-term meteorological wet and dry spells, the CMI was designed to evaluate short-term moisture conditions across major crop-producing regions. It is based on the mean temperature and total precipitation for each week, as well as the CMI value from the previous week. Categories range from less than -3 (severely dry) to more than +3 (excessively wet). The CMI responds rapidly to changing conditions, and it is weighted by location and time so that maps, which commonly display the weekly CMI across the United States, can be used to compare moisture conditions at different locations. Weekly maps of the CMI are available as part of the USDA/JAWF Weekly Weather and Crop Bulletin.

- **Standardized Precipitation Index (SPI; McKee *et al.*, 1993) –** precipitation-based drought. The SPI was developed to categorize rainfall as a standardized departure with respect to a rainfall probability distribution function; categories range from less than -3 (extremely dry) to more than +3 (extremely wet). The SPI is calculated on the basis of selected periods of time (typically from 1 to 48 months of total precipitation) and it indicates how the precipitation for a specific period compares with the long-term record at a given location (Edwards and McKee, 1997). The index correlates well with other drought indices. Sims et al. (2002) suggested that the SPI was more representative of short-term precipitation and a better indicator of soil wetness than the PDSI. The 9-month SPI corresponds closely to the PDSI (Heim 2002; Guttman 1998).
- **Keetch-Byram Index (KBDI; Keetch and Byram, 1968) –** meteorological drought and wildfire potential index. This was developed to characterize the level of potential fire danger. It uses daily temperature and precipitation information and estimates soil moisture deficiency. High values of KBDI are indicative of favorable conditions for wildfires. However, the index needs to be regionalized, as values are not comparable among regions Groisman et al. 2004, 2007a).
- **No-rain episodes –** meteorological drought. Groisman and Knight (2007) proposed to directly monitor frequency and intensity of prolonged no-rain episodes (greater

than 20, 30, 60, etc. days) during the warm season, when evaporation and transpiration are highest and the absence of rain may affect natural ecosystems and agriculture. They found that during the past four decades the duration of prolonged dry episodes has significantly increased over the Eastern and Southwestern United States and adjacent areas of Northern Mexico and Southeastern Canada.

- **Soil Moisture and Runoff Index (SMRI; Andreadis and Lettenmaier, 2006) –** hydrologic and agricultural droughts. The SMRI is based on model-derived soil moisture and runoff as drought indicators; it uses percentiles and the values are normalized from 0 (dry) to 1 (wet conditions). The limitation of this index is that it is based on land-surface model-derived soil moisture. However, long-term records of soil moisture – a key variable related to drought – are essentially non-existent (Andreadis and Lettenmaier, 2006). Thus, the advantage of the SMRI is that it is physically based and with the current sophisticated land-surface models it is easy to produce multimodel average climatologies and century-long reconstructions of land surface conditions, which could be compared under drought conditions.

Resources: A list of these and other drought indicators, data availability, and current drought conditions based on observational data can be found at the National Climatic Data Center (NCDC, http://www.ncdc.noaa.gov). The North American Drought Monitor at NCDC monitors current drought conditions in Canada, the United States, and Mexico. Tree-ring reconstruction of PDSI across North America over the last 2000 years can be also found at NCDC

2.2.2.2 Short Duration Heavy Precipitation

2.2.2.2.1 Data Considerations and Terms

Intense precipitation often exhibits higher spatial variability than many other extreme phenomena. This poses challenges for the analysis of observed data since the heaviest area of precipitation in many events may fall between stations. This adds uncertainty to estimates of regional trends based on the climate network. The uncertainty issue is explicitly addressed in some recent studies.

Precipitation extremes are typically defined based on the frequency of occurrence [by percentile (e.g., upper 5%, 1%, 0.1%, etc) or by return period (e.g. an average occurrence of once every 5 years, once every 20 years, etc.)] of rain events and/or their absolute values (e.g., above 50 mm, 100 mm, 150 mm, or more). Values of percentile or return period thresholds vary considerably across North America. For example, in the U.S., regional average values of the 99.9 percentile threshold for daily precipitation are lowest in the Northwest and Southwest (average of 55 mm) and highest in the South (average of 130mm)[31].

As noted above, spatial patterns of precipitation have smaller spatial correlation scales (for example, compared to temperature and atmospheric pressure) which means that a denser network is required in order to achieve a given uncertainty level. While monthly precipitation time series for flat terrain have typical radii of correlation[32] (ρ) of ~300 km

[31] The large magnitude of these differences is a major motivation for the use of regionally-varying thresholds based on percentiles.
[32] Spatial correlation decay with distance, r, for many meteorological variables, X, can be approximated by

or even more, daily precipitation may have ρ less than 100 km with typical values for convective rainfall in isolated thunderstorms of ~15 to 30 km (Gandin and Kagan 1976). Values of ρ can be very small for extreme rainfall events and sparse networks may not be adequate to detect a desired minimum magnitude of change that can result in societally-important impacts and can indicate important changes in the climate system.

2.2.2.2.2 United States

One of the clearest trends in the U.S. observational record is that of an increasing frequency and intensity of heavy precipitation events (Karl and Knight 1998; Groisman et al. 1999, 2001, 2004; Kunkel et al. 1999; Easterling et al. 2000; IPCC 2001; Semenov and Bengtsson 2002, Kunkel 2003). For example, the area of the United States affected by a much above normal contribution to total annual precipitation of daily precipitation events exceeding 50.8 mm (2 inches) increased by a statistically significant amount from about 9% in the 1910s to about 11% in the 1980s and 1990s (Karl and Knight 1998). Total precipitation also increased during this time, due in large part to increases in the intensity of heavy precipitation events (Karl and Knight 1998). In fact, there has been little change or decreases in the frequency of light and average precipitation days (Easterling et al. 2000; Groisman et al. 2004, 2005) during the last 30 years while heavy precipitation frequencies have increased (Sun and Groisman 2004). For example, the amount of precipitation falling in the heaviest 1% of rain events has increased by 20% during the 20th Century while total precipitation has increased by only 7% (Groisman et al. 2004). Although the exact character of those changes has been questioned (e.g.

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an exponential function of distance: Corr (X(A), X(B)) ~ $e^{-r/\rho}$ where r is a distance between point A and B and ρ is a radius of correlation, which is a distance where the correlation between the points is reduced to 1/e compared to an initial "zero" distance.

Michaels et al. 2004), it is highly likely that in recent decades extreme precipitation events have increased more than light to medium events.

A statistically significant 50% increase during the 1900s in the frequency of days with precipitation exceeding 101.6 mm (4 inches) was found in the upper Midwest U.S. (Groisman et al. 2001). Upward trends in the amount of precipitation occurring in the upper 0.3% of daily precipitation events are statistically significant for the period of 1908-2002 within three major regions (the South, Midwest, and Upper Mississippi; see Fig. 2.8) of the central United States (Groisman et al. 2004). The upward trends are primarily a warm season phenomenon when the most intense rainfall events typically occur. A time series of the frequency of events in the upper 0.3% averaged for these 3 regions (Fig 2.8) shows a 20% increase over the period of 1893-2002 with all of this increase occurring over the last third of the 1900s (Groisman et al. 2005).

Examination of intense precipitation events defined by return period, covering the period of 1895-2000, indicates that the frequencies of extreme precipitation events before 1920 were generally above the long-term averages for durations of 1 to 30 days and return periods 1 to 20 years and only slightly lower than values during the 1980s and 1990s (Kunkel et al. 2003). The highest values occur after about 1980, but the elevated levels prior to about 1920 are an interesting feature suggesting that there is considerable variability in the occurrence of extreme precipitation on multi-decadal time scales

There is a seeming discrepancy between the results for the 99.7th percentile (which do not show high values early in the record in the analysis of Groisman et al. 2004) and for 1 to 20-yr return periods (which do in the analysis of Kunkel et al. 2003). The number of stations with available data is only about half (about 400) in the late 1800s of what is available in most of the 1900s (800-900). Furthermore, the spatial distribution of stations throughout the record is not uniform; the density in the western U.S. is relatively lower than in the central and eastern U.S. It is possible that the resulting uncertainties in heavy precipitation estimates are too large to make unambiguous statements about the recent high frequencies.

Recently, this question was addressed (Kunkel et al. 2007a) by analyzing the modern dense network to determine how the density of stations affects the uncertainty and then to estimate the level of uncertainty in the estimates of frequencies in the actual (sparse) network used in the long-term studies. The results were unambiguous. For all combinations of three durations (1-day, 5-days and 10-days) and 3 return periods (1-yr, 5-yr, and 20-yr), the frequencies for 1983-2004 were significantly higher than those for 1895-1916 at a high level of confidence. In addition, the observed linear trends were all found to be upward, again with a high level of confidence. Based on these results, it is highly likely that the recent elevated frequencies in the U.S. are the highest since 1895.

2.2.2.2.3 Alaska and Canada

The sparse network of long-term stations in Canada increases the uncertainty in estimates of extremes. Changes in the frequency of heavy events exhibit considerable multi-decadal variability since 1900, but no long-term trend for the entire century (Zhang et al. 2001). However, according to Zhang et al. (2001), there are not sufficient instrumental data to discuss the nationwide trends in precipitation extremes over Canada prior to 1950. Nevertheless, there are changes that are noteworthy. For example, the frequency of 99.7% events exhibits a statistically significant upward trend of 19%/50yr in British Columbia since 1910 (Fig. 2.8; Groisman et al. 2005). For Canada, increases in precipitation intensity during the second half of the 1900s are concentrated in heavy and intermediate events, with the largest changes occurring in Arctic areas (Stone et al. 2000). The tendency for increases in the frequency of intense precipitation while the frequency of days with average and light precipitation does not change or decreases has also been observed in Canada over the last 30 years (Stone et al. 2000), mirroring U.S. changes. Recently, Vincent and Mekis (2006) repeated analyses of precipitation extremes for the second half of 1900s (1950-2003 period). They reported a statistically significant increase of 1.8 days per 54 years in heavy precipitation days (defined as the days with precipitation above 10 mm) and statistically insignificant increases in the maximum 5-day precipitation (by ~5%) and in the number of “very wet days” defined as days with precipitation above the upper 5th percentiles of local daily precipitation (by 0.4 days).

There is an upward trend of 37%/50yr in southern Alaska since 1950 although this trend is not statistically significant (Fig. 2.8; Groisman et al. 2005).

2.2.2.2.4 Mexico

On an annual basis, the number of heavy precipitation (P > 10 mm) days has increased in northern Mexico and the Sierra Madre Occidental and decreased in the south-central part of the country (Alexander et al. 2006). The percent contribution to total precipitation from heavy precipitation events exceeding the 95th percentile threshold has increased in the monsoon region (Alexander et al., 2006) and along the southern Pacific coast (Aguilar et al. 2005), while some decreases are documented for south-central Mexico (Aguilar et al. 2005).

On a seasonal basis, the maximum precipitation reported in 5 consecutive days during winter and spring has increased in Northern Mexico and decreased in south-central Mexico (Alexander et al. 2006). Northern Baja California, the only region in Mexico characterized by a Mediterranean climate, has experienced an increasing trend in winter precipitation exceeding the 90th percentile, especially after 1977 (Cavazos and Rivas, 2004). Heavy winter precipitation in this region is significantly correlated with El Niño events (Pavia and Badan, 1998; Cavazos and Rivas, 2004); similar results have been documented for California (e.g., Gershunov and Cayan, 2003). During the summer there has been a general increase of 2.5 mm in the maximum 5-consecutive-day precipitation in most of the country and an upward trend in the intensity of events exceeding the 99th and 99.7th percentiles in the high plains of Northern Mexico during the summer season (Groisman et al. 2005).

During the monsoon season (June-September) in northwestern Mexico, the intensity and seasonal contribution of rainfall events exceeding the 95^{th} percentiles significantly increased ($p<0.05$) in the core monsoon region and at mountain sites (Fig. 2.8; Cavazos et al. 2007). The mean intensity of 95^{th} percentile events in the monsoon region increased significantly by 0.6 mm dec^{-1} during 1950-2003. It went from 17.9 mm d^{-1} in the 1950-1976 period to 19.6 mm d^{-1} in 1977-2003 while at mountain sites the increase was from 40.8 mm d^{-1} to 43.9 mm d^{-1}, respectively. These increases are mainly due to an increase in tropical cyclone-derived rainfall after 1980. The frequency of heavy events does not show a significant trend (Englehart and Douglas 2001; Neelin et al. 2006; Cavazos *et al*., 2007). Similarly, Groisman et al. (2005) report that the frequency of very heavy summer precipitation events (above the 99^{th} percentile) in the high plains of Northern Mexico (east of the core monsoon) has not increased, whereas their intensity has increased significantly.

The increase in the mean intensity of heavy summer precipitation events in the core monsoon region during the 1977-2003 period are significantly correlated with the Oceanic El Niño Index (ONI[33]) conditions during the cool season. El Niño SST anomalies antecedent to the monsoon season are associated with less frequent, but more intense, heavy precipitation events[34] (exceeding the 95^{th} percentile threshold), and vice versa.

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[33] ONI INDEX:
http://www.cpc.ncep.noaa.gov/products/analysis_monitoring/ensostuff/ensoyears.shtml
Warm and cold episodes based on a threshold of +/- 0.5°C for the Oceanic Niño Index (ONI) [3 month running mean of ERSST.v2 SST anomalies in the Niño 3.4 region (5°N-5°S, 120°-170°W)], based on the 1971-2000 base period.

[34] The correlation coefficient between ONI and heavy precipitation frequency (intensity) is -0.37 (+0.46)

There has been an insignificant decrease in the number of consecutive dry days in northern Mexico, while an increase is reported for south-central Mexico (Alexander et al., 2006), and the southern Pacific coast (Aguilar et al. 2005).

2.2.2.2.5 Summary

All studies indicate that changes in heavy precipitation frequencies are *always* higher than changes in precipitation totals and, in some regions, an *increase* in heavy and/or very heavy precipitation occurred while no change or even a decrease in precipitation totals was observed (e.g., in the summer season in central Mexico). There are regional variations in where these changes are statistically significant (Fig. 2.8). The most significant changes occur in the central U.S., central Mexico, southeastern, northern and western Canada, and southern Alaska. These changes have resulted in a wide range of impacts, including human health impacts (Chapter 1, Box 1.3).

2.2.2.3 Monthly to Seasonal Heavy Precipitation

On the main stems of large river basins, significant flooding will not occur from short duration extreme precipitation episodes alone. Rather, excessive precipitation must be sustained for weeks to months. The 1993 Mississippi River flood, which resulted in an estimated $17 billion in damages, was caused by several months of anomalously high precipitation (Kunkel et al. 1994).

A time series of the frequency of 90-day precipitation totals exceeding the 20-year return period (a simple extension of the approach of Kunkel et al. 2003) indicates a statistically significant upward trend (Fig. 2.9). The frequency of such events during the last 25 years is 20% higher than during any earlier 25-year period. Even though the causes of multi-month excessive precipitation are not necessarily the same as for short duration extremes, both show moderately high frequencies in the early 20th Century, low values in the 1920s and 1930s, and the highest values in the past 2-3 decades. The trend[35] over the entire period is highly statistically significant.

2.2.2.4 North American Monsoon

Much of Mexico is dominated by a monsoon type climate with a pronounced peak in rainfall during the summer (June through September) when up to 60% to 80% of the annual rainfall is received (Douglas et al., 1993; Higgins et al., 1999 and Cavazos et al., 2002). Monsoon rainfall in southwest Mexico is often supplemented by tropical cyclones moving along the coast. Farther removed from the tracks of Pacific tropical cyclones, interior and northwest sections of Mexico receive less than 10% of the summer rainfall from passing tropical cyclones (Fig. 2.10; Englehart and Douglas 2001). The main influences on total monsoon rainfall in these regions rests in the behavior of the monsoon as defined by its start and end date, rainfall intensity and duration of wet and dry spells (Englehart and Douglas 2006). Extremes in any one of these parameters can have a strong effect on the total monsoon rainfall.

The monsoon in northwest Mexico has been studied in detail because of its singular importance to that region and because summer rainfall from this core monsoon region

[35] The data were first subjected to a square root transformation to produce a data set with an approximate normal distribution; then least squares regression was applied. Details can be found in the Appendix, Example 4.

spills over into the U.S. Desert Southwest (Douglas et al., 1993; Higgins et al. 1999, Cavazos et al. 2002)**.** Based on long term data from 8 stations in southern Sonora, the summer rains have become increasingly late in arriving (Englehart and Douglas 2006) and this has had strong hydrologic and ecologic repercussions for this northwest core region of the monsoon. Based on linear trend**,** the mean start date for the monsoon has been delayed almost 10 days (9.89 days with a significant trend of 1.57 days per decade) over the past 63 years (Figure 2.11a). Because extended periods of intense heat and desiccation typically precede the arrival of the monsoon, the trend toward later starts to the monsoon will place additional stress on the water resources and ecology of the region if continued into the future.

Accompanying the tendency for later monsoon starts, there also has been a notable change in the **"**consistency**"** of the monsoon as indicated by the average duration of wet spells in southern Sonora (Figure 2.11b). Based on a linear trend**,** the average wet spell[36] has decreased by almost one day (0.88 days with a significant trend of -0.14 days per decade) from nearly four days in the early 1940s to slightly more than three days in recent years. The decrease in wet spell length indicates a more erratic monsoon is now being observed. Extended periods of consecutive days with rainfall are now becoming less common during the monsoon. These changes can have profound influences on surface soil moisture levels which affect both plant growth and runoff in the region.

[36] For southern Sonora, Mexico, wet spells are defined as the mean number of consecutive days with mean regional precipitation ≥1mm**.**

A final measure of long term change in monsoon activity is associated with the change in rainfall intensity over the past 63 years (Figure 2.11c). Based on linear trend**,** rainfall intensity[37] in the 1940s was roughly 5.6mm per rain day, but in recent years has risen to nearly 7.5mm per rain day[38]. Thus, while the summer monsoon has become increasingly late in arriving and wet spells have become shorter, the average rainfall during rain events has actually increased very significantly by 17% or 1.89mm over the 63 year period (0.3 mm per decade) as well as the intensity of heavy precipitation events (Fig. 2.9). Taken together, these statistics indicate that the rainfall in the core region of the monsoon (i.e., northwest Mexico) has become more erratic with a tendency towards high intensity rainfall events countering the tendency towards a shorter monsoon with shorter wet spells.

Variability in Mexican monsoon rainfall shows modulation by large-scale climate modes. Englehart and Douglas (2002) demonstrate that a well developed inverse relationship exists between ENSO and total seasonal rainfall (June-September) over much of Mexico, but the relationship is only operable in the positive phase of the PDO. Evaluating monsoon rainfall behavior on intraseasonal time scales, Englehart and Douglas (2006) demonstrate that rainfall intensity (mm/rain day) in the core region of the monsoon is related to PDO phase with the positive (negative) phase favoring relatively high (low) intensity rainfall events. Analysis indicates that other rainfall characteristics of the monsoon respond to ENSO with warm events favoring later starts to the monsoon and

[37] Daily rainfall intensity during the monsoon is defined as the regional average rainfall for all days with rainfall $\geq$ 1mm.
[38] The linear trend in this time series is significant at the p=0.01 level

shorter length wet spells (days) with cold events favoring opposite behavior (Englehart and Douglas 2006).

2.2.2.5 Tropical Storm Rainfall in Western Mexico

Across southern Baja California and along the southwest coast of Mexico, 30% to 50% of warm season rainfall (May-November) is attributed to tropical cyclones (Fig. 2.10) and in years heavily affected by tropical cyclones (upper 95th percentile) 50% to 100% of the summer rainfall comes from tropical cyclones. In this region of Mexico, there is a long term, upward trend in tropical cyclone-derived rainfall at both Manzanillo (41.8mm/decade; Fig. 2.12a) and Cabo San Lucas (20.5mm/decade)[39]. This upward trend in tropical cyclone rainfall has led to an increase in the importance of tropical cyclone rainfall in the total warm season rainfall for southwest Mexico (Fig. 2.12b) and this has resulted in a higher ratio of tropical cyclone rainfall to total warm season rainfall. Since these two stations are separated by more than 700km, these significant trends in tropical cyclone rainfall imply large scale shifts in the summer climate of Mexico.

This recent shift in emphasis on tropical cyclone warm season rainfall in western Mexico has strong repercussions as rainfall becomes less reliable from the monsoon and becomes more dependent on heavy rainfall events associated with passing tropical cyclones. Based on the large scale and heavy rainfall characteristics associated with tropical cyclones, dams in the mountainous regions of western Mexico are often recharged by strong

[39] The linear trends in tropical cyclone rainfall at these two stations are significant at the p=0.01 and p=0.05 level, respectively.

tropical cyclone events which therefore have positive benefits for Mexico despite any attendant damage due to high winds or flooding.

This trend in tropical cyclone-derived rainfall is consistent with a long term analysis of near-shore tropical storm tracks along the west coast of Mexico (storms passing within 5° of the coast) which indicates an upward trend in the number of near shore storms over the past 50 years (Fig. 2.12c). While the number of tropical cyclones occurring in the entire east Pacific Basin is uncertain prior to the advent of satellite tracking in about 1967, it should be noted that the long term data sets for near shore storm activity (within 5° of the coast) are considered to be much more reliable due to coastal observatories and heavy ship traffic to and from the Panama Canal to Pacific ports in Mexico and the United States. The number of near shore storm days (storms less than 550km from the station) has increased by 1.3 days/decade in Manzanillo and about 0.7days/decade in Cabo San Lucas (1949-2006)[40]. The long term correlation between tropical cyclone days at each station and total tropical cyclone rainfall is r = 0.61 for Manzanillo and r = 0.37 for Cabo San Lucas, illustrating the strong tie between passing tropical cyclones and the rain that they provide to coastal areas of Mexico.

Interestingly, the correlations between tropical cyclone days and total tropical cyclone rainfall actually drop slightly when based only on the satellite era, 1967-2006 (r = 0.54 for Manzanillo and r = 0.31 for Cabo San Lucas). The fact that the longer time series has the higher set of correlations shows no reason to suggest problems with near shore

[40] The linear trends in near shore storm days are significant at the p=0.05 level and p=0.10 level, respectively.

tropical cyclone tracking in the pre-satellite era. The lower correlations in the more recent period between tropical cyclone days and total tropical cyclone rainfall may be tied to tropical cyclone derived rainfall rising at a faster pace compared to the rise in tropical cyclone days. In other words, tropical cyclones are producing more rain per event than in the earlier 1949-1975 period when SSTs were colder.

2.2.2.6 Tropical Storm Rainfall in the Southeastern United States

Tropical cyclone-derived rainfall along the southeastern coast of the United States on a century time scale has changed insignificantly in summer (when no century-long trends in precipitation was observed) as well as in autumn (when the total precipitation increased by more than 20% since the 1900s; Groisman et al. 2004).

2.2.3 Storm Extremes

2.2.3.1 Tropical Cyclones

2.2.3.1.1 Introduction

Each year, about 90 tropical cyclones develop over the world's oceans, and some of these make landfall in populous regions, exacting heavy tolls in life and property. Their occurrence is often statistically modeled as a Poisson process. The global number has been quite stable since 1970, when global satellite coverage began in earnest, having a standard deviation of 10 and no evidence of any substantial trend (e.g. Webster et al 1995). However,, there is some evidence for trends in storm intensity and/or duration (e.g. Holland and Webster 2007 and quoted references for the North Atlantic; Chan 2000 for the Western North Pacific), and there is substantial variability in tropical cyclone frequency within each of the ocean basins they affect. Regional variability occurs on all resolved time scales, and there is also some evidence of trends in certain measures of tropical cyclone energy, affecting many of these regions and perhaps the globe as well.

There are at least two reasons to be concerned with such variability. The first and most obvious is that tropical cyclones rank with flash floods as the most lethal and expensive natural catastrophes, greatly exceeding other phenomena such as earthquakes. In developed countries, such as the U.S., they are enormously costly: Hurricane Katrina is estimated to have caused in excess of $80 billion 2005 dollars in damage, and killed more than 1500 people. Death and injury from tropical cyclones is yet higher in developing nations; for example, Hurricane Mitch of 1998 took more than 11,000 lives in Central America. Any variation or trend in tropical cyclone activity is thus of concern to coastal residents in affected areas, compounding trends related to societal factors such as changing coastal population.

A second, less obvious and more debatable issue is the possible feedback of tropical cyclone activity on the climate system itself. The inner cores of tropical cyclones have the highest specific entropy content of any air at sea level, and for this reason such air penetrates higher into the stratosphere than is the case with other storm systems. Thus tropical cyclones may play a role in injecting water and trace gases and microscopic airborne particles into the upper troposphere and lower stratosphere, though this idea remains largely unexamined. There is also considerable evidence that tropical cyclones vigorously mix the upper ocean, affecting its circulation and biogeochemistry, perhaps to the point of having a significant effect on the climate system. Since the current generation of coupled climate models greatly under-resolves tropical cyclones, such feedbacks are badly underrepresented, if they are represented at all.

For these reasons, it is important to quantify, understand, and predict variations in tropical cyclone activity. The following sections review current knowledge of these variations on various time scales.

2.2.3.1.2 Data Issues

Quantifying tropical cyclone variability is limited, sometimes seriously, by a large suite of problems with the historical record of tropical cyclone activity. In the North Atlantic and eastern North Pacific regions, responsibility for the tropical cyclone database rests with NOAA's National Hurricane Center (NHC), while in other regions, archives of hurricane activity are maintained by several organizations, including the U.S. Navy's Joint Typhoon Warning Center (JTWC), the Japan Meteorological Agency (JMA), the Hong Kong Observatory (HKO) and the Australian Bureau of Meteorology (BMRC). The data, known as ``best track'' data (Jarvinen et al. 1984; Chu et al. 2002), comprise a global historical record of tropical cyclone position and intensity, along with more recent structural information. Initially completed in real time, the best tracks are finalized by teams of forecasters update the best track data at the end of the hurricane season in each ocean basin using data collected during and after each hurricane's lifetime.

It should first be recognized that the primary motivation for collecting data on tropical cyclones was initially to support real-time forecasts and this remains the case in many regions today. From the 1970s onwards increasing emphasis has been placed on improving the archive for climate purposes, and on extending the record back to include historical systems (e.g. Laurensz 1977; Neumann 1993; Landsea et al 2004). Unfortunately, improvements in measurement and estimation techniques have often been implemented with little or no effort to calibrate against existing techniques and with poor documentation where such calibrations were done. Thus the available tropical cyclone data contain an inhomogeneous mix of changes in quality of observing systems, reporting policies, and the methods utilized to analyze the data

It remains a scientific tragedy that insufficient effort is expended in re-examining and quality controlling the tropical cyclone record on a year to year basis, particularly outside the Atlantic and eastern North Pacific regions. Efforts are ongoing to reanalyze the historic best track data, but such a posteriori reanalyses are less than optimal because not all of the original data that the best track was based on are readily available.

Documentation of the occurrence of tropical cyclones is thought to be reliable back to about 1945 in the Atlantic and 1970 in the Eastern Pacific (e.g. Holland and Webster 2007 and references therein), and back to about 1975 for the Western and Southern Pacific basins, thanks to earth-orbiting satellites (e.g. Holland 1981). Until the launch of MeteoSat-7 in 1998, the Indian Oceans were seen only obliquely, but storm counts may still be expected to be accurate after 1977. Before those periods, storms could and undoubtedly remain undetected, especially if they did not pass near ships at sea or land masses. For the North Atlantic it is likely that up to 3 storms per year were missing before 1900 dropping to zero by the early 1960s (Holland and Webster 2007; Chang and Guo 2007). Estimates of the duration of storms are considered to be less reliable prior to the 1970's due particularly to a lack of good information on their time of genesis. Since the 1970s storms were more accurately tracked throughout their lifetimes by geostationary satellites.

Estimates of storm intensity are far less reliable, and this remains true for large portions of the globe even today. Airborne hurricane reconnaissance flight became increasingly routine in the North Atlantic and western North Pacific regions after 1945, but was discontinued in the western North Pacific region in 1987. Some missions are today being conducted under the auspices of the government of Taiwan. However airborne reconnaissance only samples a small fraction of storms, and then only over a fraction of their lifetimes; moreover, good, quantitative estimates of wind speeds from aircraft did not become available until the late 1950s. Beginning in the mid 1970s, tropical cyclone intensity has been estimated from satellite imagery. Until relatively recently, techniques for doing so were largely subjective, and the known lack of homogeneity in both the data and techniques applied in the post-analyses has resulted in significant skepticism regarding the consistency of the intensity estimates in the data set. This lack of temporal consistency renders the data suspect for identifying trends, particularly in metrics related to intensity.

Recent studies have addressed these known data issues. Kossin et al. (2007a) constructed a more homogeneous record of hurricane activity and found remarkably good agreement in both variability and trends between their new record and the best track data in the N. Atlantic and Eastern Pacific basins during the period 1983–2005. They concluded that the best track maintained by the NHC does not appear to suffer from data quality issues during this period. On the other hand, they were not able to corroborate the presence of upward intensity trends in any of the remaining tropical cyclone-prone ocean basins. This could be due to inaccuracies in the satellite best tracks, or could be due to the training of the Kossin et al technique on North Atlantic data. This is supported by Wu et al. (2006), who considered Western Pacific best track data constructed by other agencies (HKMO and JMA) who construct best track data for the western North Pacific. Harper and Callaghan (2006) report on reanalyzed data from the Southeastern Indian Ocean and showed some biases, but a remaining upward intensity trend. These studies underscores the need for improved care in analyzing tropical cyclones and in obtaining better understanding of the climatic controls of tropical cyclone activity beyond SST-based arguments alone.

The standard tropical cyclone databases do not usually contain information pertaining to the geometric size of tropical cyclones Exceptions include the Australian region and the enhanced database for the North Atlantic over the last few decades. A measure of size of a tropical cyclone is a crucial complement to estimates of intensity as it relates directly to storm surge and damage area associated with landfalling storms. Such size measures can be inferred from aircraft measurements and surface pressure distributions, and can now be estimated from satellite imagery (e.g. Mueller et al. 2006; Kossin et al. 2007b).

2.2.3.1.3 Low-frequency Variability and Trends of Tropical cyclone Activity Indices

"Low frequency" variability is here defined as variations on time scales greater than those associated with ENSO (i.e. more than 3-4 years). Several papers in recent years have quantified interdecadal variability of tropical cyclones in the Atlantic (Goldenberg et al., 2001; Bell and Chelliah, 2006) and the western North Pacific (Chan and Shi, 1996), attributing most of the variability to natural interdecadal variability of regional climates in the Atlantic and Pacific, respectively. In the last few years, however, several papers have attributed both low frequency variability and trends in tropical cyclone activity to changing radiative forcing owing to anthropogenic sulfate aerosols and greenhouse gases. Emanuel (2005a) developed a "Power Dissipation Index" (PDI) of tropical cyclones, defined as the sum of the cubed estimated maximum sustained surface wind speeds at 6-hour intervals accumulated over each Atlantic tropical cyclone from the late 1940s to 2003. Landsea (2005) commented on the quality of data comprising the index. An updated version of this analysis (Emanuel 2007), shown in Fig. 2.13, confirms that there has been a substantial increase in tropical cyclone activity since about 1970, and indicates that the low-frequency Atlantic PDI variations are strongly correlated with low-frequency variations in tropical Atlantic SSTs. Based on this analysis, it is likely that hurricane activity, as measured by the Power Dissipation Index (PDI), has increased substantially since the 1950s and 60s in association with warmer Atlantic SSTs. The magnitude of this increase depends on the adjustment to the wind speed data from the 1950s and 60s (Landsea 2005; Emanuel 2007). It is very likely that PDI has generally tracked SST variations on decadal time scales in the tropical Atlantic since 1950 and likely that it also generally tracked the secular increase of SST. Confidence in these statistics prior to the late 1940s is low, due mainly to the decreasing confidence in hurricane duration and intensity observations. The PDI in the eastern Pacific has decreased since 1980 (Kossin et al. 2007).

The Power Dissipation Index for U.S. landfalling tropical cyclones has not increased since the late 1800s (Landsea 2005). Pielke (2005) noted that there are no evident trends in observed damage in the North Atlantic region, after accounting for population increases and coastal development. However, Emanuel (2005b) notes that a PDI series such as Landsea's (2005) based on only U.S. landfalling data, contains only about 1 percent of the data that Emanuel's (2005a) basin-wide PDI contains, which is based on all storms over their entire lifetimes. Thus a trend in basin-wide PDI may not be detectable in U.S. landfalling PDI since the former index has a factor of 10 advantage in signal to noise ratio.

Figure 2.14 (from Holland and Webster 2007), indicates that there has been no distinct trend in the mean intensity of all Atlantic storms, hurricanes, and major hurricanes. A distinct increase in the most intense storms occurred around the time of onset of aircraft reconnaissance, but this is considered to be largely due to better observing methods. Holland and Webster also found that the overall proportion of hurricanes has remained remarkably constant during the 20^{th} century at around 50%, and there has been a marked oscillation in major hurricane proportions, which has no observable trend.
Webster et al. (2005) reported that the number of category 4 and 5 hurricanes has almost

doubled globally over the past three decades. The recent reanalysis of satellite data beginning in the early 1980s by Kossin et al. (2007a) support these results in the Atlantic although the results in the remaining basins were not corroborated.

The recent Emanuel and Webster et al. studies have generated much debate in the hurricane research community, particularly with regard to homogeneity of the tropical cyclone data over time and the required adjustments (e.g. Landsea 2005; Knaff and Sampson 2006; Chan 2006; Hoyos et al. 2006; Landsea et al. 2006; Sriver and Huber 2006; Klotzbach 2006; Elsner et al. 2006; Maue and Hart 2007; Manning and Hart 2007; Holland and Webster 2007, Landsea 2007, Mann et al 2007, Holland 2007). Several of these studies argue that data problems preclude determination of significant trends in various tropical cyclone measures, while others provide further evidence in support of reported trends. In some cases, differences between existing historical data sets maintained by different nations can yield strongly contrasting results (e.g., Kamahori et al. 2006).

Several studies have examined past regional variability in tropical cyclone tracks (Wu et al. 2005; Xie et al. 2005; Vimont and Kossin 2007; Kossin and Vimont 2007). Thus far, no clear long-term trends in this metric have been reported, but there is evidence that Atlantic tropical cyclone formation regions have undergone systematic long-term shifts to more eastward developments (Holland 2007). These shifts affect track and duration, which subsequently affects intensity. The modulation of the Atlantic tropical cyclone genesis region occurs through systematic changes of the regional SST and circulation patterns. Thus SST affects intensity not just through thermodynamic pathways that are local to the storms, but also through changes in basinwide circulation patterns (Kossin and Vimont 2007).

In summary, we conclude that Atlantic tropical storm and hurricane destructive potential as measured by the Power Dissipation Index (which combines storm intensity, duration, and frequency) has increased. This increase is substantial since about 1970, and is likely substantial since the 1950s and 60s, in association with warming Atlantic sea surface temperatures.

2.2.3.1.4 Low-frequency Variability and Trends of Tropical Cyclone Counts

Mann and Emanuel (2006) reported that Atlantic tropical cyclone counts closely track low-frequency variations in tropical Atlantic SSTs, including a long-term increase since the late 1800s and early 1900s (see also Fig. 2.15 from Holland and Webster 2007). There is currently debate on the relative roles of internal climate variability (e.g., Goldenberg et al. 2001) versus radiative forcing, including greenhouse gases, and sulfate aerosols (Mann and Emanuel 2006; Santer et al 2006) in producing the multi-decadal cooling of the tropical North Atlantic. This SST variation is correlated with reduced hurricane activity during the 1970s and 80s relative to the 1950s and 60s or to the period since 1995 (see also Zhang et al. 2007).

On a century time scale, time series of tropical storm frequency in the Atlantic (Fig. 2.15) show substantial interannual variability and a marked increase (of over 100%) since about 1900. This increase occurred in two sharp jumps of around 50%, one in the 1930s and another that commenced in 1995 and has not yet stabilized. Holland and Webster (2007) have suggested that these sharp jumps are transition periods between relatively stable climatic periods of tropical cyclone frequency (Fig. 2.15). Figure 2.15 uses unadjusted storm—an issue which will be addressed further below.

For tropical cyclone frequency, the finding that the largest recorded increases over the past century have been in the eastern North Atlantic (e.g., see recent analysis in Vecchi and Knutson 2007; Holland 2007), which historically has been the least well observed, has led to questions of whether this may be due to data issues (Landsea et al. 2004; Landsea 2007). The major observing system change points over the past century have been:

- The implementation of routine aircraft reconnaissance in 1944-45;
- The use of satellite observations and related analysis procedures from the late 1960s onwards; and,
- A change in analysis practice by the National Hurricane Center from 1970 to include more mid-latitude systems.

In addition, there have steady improvements in techniques and instrumentation, which may also introduce some spurious trends.

Landsea (2007) has used the fraction of storms striking land in the satellite and pre-satellite era to estimate the number of missing storms per year in the pre-satellite era (1900 to 1965) to be about 3.2 storms per year. This assumes that the fraction of all storms that strike land in the real world has been relatively constant over time, which has been shown to be incorrect by Holland (2007). Holland also shows that the smaller fraction of storms that made landfall during the past fifty years (1956-2005) compared to the previous fifty years (1906-1955) is directly related to changes in the main formation location regions, with a decrease in western Caribbean and Gulf of Mexico developments and an increase in the eastern Atlantic.

Alternative approaches to estimating the earlier data deficiencies have been used by Chang and Guo (2007), Vecchi and Knutson (2007) and Mann et al (2007). The first two studies use historical ship tracks from the pre-satellite era, combined with storm track information from the satellite era, to infer an estimated adjustment for missing storms in the pre-satellite era (assumed as all years prior to 1965). Mann et al used statistical climate relationships to estimate potential errors. Vecchi and Knutson found 2.5 storms per year were missing prior to 1900, decreasing to zero by 1960. Chang and Guo found 1.2 storms missing around 1910 also decreasing to zero by 1960. Mann et al, estimated a more modest undercount bias of 1 per year back to 1970. The adjusted time series by Vecchi and Knutson (Fig. 2.16) suggest a statistically significant (p=0.003 or less) positive linear trend in adjusted storm counts of 55%/century since 1900. However, beginning the trend from 1878, the trend through 2006 is smaller (+15%/century) and not statistically significant at the p=0.05 level (p-value of about 0.3)[41]. It is notable that the degree of increase over the past century depends on the analysis methodology. When using a linear trend, as above, the increase from 1900 to 2005 is around 55% in the adjusted storm counts. However, using the essentially non-linear approach by Holland

[41] Details of the statistical analysis are given in the Appendix, Example 5.

and Webster (2007) of separate climatic regimes, the increase in adjusted storm counts from the 1900-1920 regime to the 1995-2006 regime is 85%. The trend from 1900 begins near a local minimum in the time series and ends with the recent high activity, perhaps exaggerating the significance of the trend due to multidecadal variability. On the other hand high levels of activity during the late 1800s, which lead to the insignificant trend result, are indirectly inferred in large part from lack of ship track data, and the uncertainty in the late 1800s storm counts is greater than that during the 1900s.

Hurricane frequency closely follows the tropical cyclone variability, with a stable 50% of all cyclones developing to hurricane strength over much of the past century (Holland and Webster 2007). However, there has been a concomitant increase in both overall storm frequency and the proportion of major hurricanes since 1995. Taken together, these result in a very sharp increase in major hurricane numbers, which can be associated with changes of SST (Holland and Webster 2007, Webster et al 2005). The PDI trend reported by Emanuel (2007) is largely due to this increase in major hurricane numbers.

Atlantic basin total hurricane counts, major hurricane counts, and U.S. landfalling hurricane counts as recorded in the HURDAT data base for the period 1851-2006 are shown in Fig. 2.17. These have not been adjusted for missing storms, as there was likely less of a tendency to miss both hurricanes and major hurricanes in earlier years compared to tropical storms, largely because of their intensity and damage potential. There is a slight negative trend in U.S. landfalling hurricane frequency. The basin-wide major hurricane counts increase over the long-term. For total hurricanes, trends to 2005 beginning in 1881 through 1921 are positive and statistically significant (p=0.05) whereas trends beginning in 1851 through 1871 are positive but not statistically significant, owing to the prolonged active period in the late 1800s. For major hurricanes, trends beginning in 1851 through 1911 were positive and statistically significant, whereas the trend beginning from 1921 was positive but not statistically significant[42].

Regional storm track reconstructions for the basin (Vecchi and Knutson 2007; Holland and Webster 2007b(?)) indicate a decrease in tropical storm occurrence in the western part of the basin, consistent with the minimal change or slight decrease in U.S. landfalling tropical storm or hurricane counts. These analyses further suggest that—after adjustment for missing storms-- a century-scale increase in basin-wide Atlantic tropical storm occurrence has occurred, with increases mainly in the central and eastern parts of the basin (also consistent with Chang and Guo 2007). From a climate variability perspective, Kossin and Vimont (2007) have shown that a positive phases of the Atlantic Meridional Mode is correlated to an systematic eastward extension of the genesis region in the Atlantic. Elsner (1996) and Holland and Webster (2007) have shown that the increasing frequency over the past 30 years is associated with a changeover to equatorial developments and particularly to developments in the eastern equatorial region.

In summary, we conclude that there have been fluctuations from decade to decade in tropical cyclone numbers, and data uncertainty is larger in the earlier parts of the record, particularly prior to aircraft reconnaissance beginning in the mid-1940s. While there are undoubtedly data deficiencies and missing storms in the early record, they appear

[42] Further details of the statistical analysis are given in the Appendix, Example 6.

insufficient to remove the observed positive trends in basin-wide tropical storm counts. It is likely that that the annual numbers of tropical storms/hurricanes and major hurricanes in the North Atlantic basin have increased significantly over the past 100 years in close relationship with warming equatorial Atlantic sea surface temperatures. The positive linear trend in all storm categories extends back to the 1800s, but is generally not significant prior to 1890. The increasingly decreased confidence in the data before 1900 precludes any definitive conclusions from this era. The increases in basin-wide storm counts has occurred primarily from an eastward shift in the formation and occurrence regions and there has been a distinct decrease in western Caribbean and Gulf of Mexico developments. As a result, North American mainland land-falling hurricanes have remained quasi-static over the past century.

2.2.3.1.5 Paleoclimate Proxy Studies of Past Tropical Cyclone Activity

Paleotempestology is an emerging field of science that attempts to reconstruct past tropical cyclone activity using geological proxy evidence or historical documents. This work attempts to expand knowledge about hurricane occurrence back in time beyond the limits of conventional instrumental records, which cover roughly the last 150 years. A broader goal of paleotempestology is to help researchers explore physically based linkages between prehistoric tropical cyclone activity and other aspects of past climate.

Among the geologically based proxies, overwash sand layers deposited in coastal lakes and marshes have proven to be quite useful (Liu and Fearn, 1993, 2000; Liu 2004; Donnelly and Webb 2004). Similar methods have been used to produce proxy records of hurricane strikes from back-barrier marshes in Rhode Island and New Jersey extending back about 700 years (Donnelly et al. 2001a, 2001b; Donnelly et al. 2004; Donnelly and Webb 2004), and more recently in the Caribbean (Donnelly 2005). Stable isotope signals in tree rings (Miller et al. 2006), cave deposits (Frappier et al. 2007) and coral reef materials are also being actively explored for their utility in providing paleoclimate information on tropical cyclone activity. Historical documents apart from traditional weather service records (newspapers, plantation diaries, Spanish and British historical archives, etc.) can also be used to reconstruct some aspects of past tropical cyclone activity (Ludlam, 1963; Millas, 1968; Fernandez-Partagas and Diaz, 1996; Chenoweth, 2003; Mock 2004; Garcia Herrera et al. 2004; 2005; Liu et al. 2001; Louie and Liu 2003; Louie and Liu 2004).

Donnelly and Woodruff's (2007) proxy reconstruction the past 5,000 years of intense hurricane activity in the western North Atlantic suggests that hurricane variability has been strongly modulated by El Nino during this time and that the past 250 years has been relatively active in the context of the past 5,000 years. Nyberg et al. (2007) suggest that major hurricane activity in the Atlantic was anomalously low in the 1970s and 1980s relative to the past 270 years. As with Donnelly and Woodruff, their proxy measures were located in the western part of the basin (near Puerto Rico), and in their study, hurricane activity was inferred indirectly through statistical associations with proxies for vertical wind shear and SSTs.

2.2.3.2 Strong Extratropical Cyclones Overview

Extra-tropical cyclone (ETC)[43] is a generic term for any non-tropical, large-scale low pressure storm system that develops along a boundary between warm and cold air masses. These types of cyclonic[44] disturbances are the dominant weather phenomenon occurring in the mid- and high-latitudes during the cold season because they are typically large and often have associated severe weather. The mid-latitude North Pacific and North Atlantic basins, between ~30^{o}N-60^{o}N, are regions where large-numbers of ETC's develop and propagate across the ocean basins each year. Over land or near populous coastlines, strong or extreme ETC events generate some of the most devastating impacts associated with extreme weather and climate, and have the potential to affect large areas and dense population centers. A notable example was the blizzard of 12-14 March 1993 along the East Coast of the U.S. that is often referred to as the "super-storm" or "storm of the century"[45] (e.g., Kocin et al.1995;). Over the ocean, strong ETCs generate high waves that can cause extensive coastal erosion when combined with storm surge as they reach the shore, resulting in significant economic impact. Rising sea level extends the zone of impact from storm surge and waves farther inland, and will likely result in increasingly greater coastal erosion and damage from storms of equal intensity.

[43] The fundamental difference between the characteristics of extra-tropical and tropical cyclones is that ETC's have a cold core and their energy is derived from baroclinic instability, while tropical cyclones have a warm core and derive their energy from barotropic instability (Holton 1979).
[44] A term applied to systems rotating in the counter-clockwise direction in the Northern Hemisphere.
[45] The phrase "Storm of the Century" is also frequently used to refer to the 1991 Halloween ETC along the Northeast US coast, immortalized in the movie *The Perfect Storm*

Studies of changes in strong ETC's and associated frontal systems have focused on locations where ETCs form and the resulting storm tracks, frequencies, and intensities[46]. The primary constraint on these studies has been the limited period of record available that has the best observation coverage for analysis and verification of results, with most research focused on the latter half of the 20th century. Model reanalysis data is used in the majority of studies, either NCEP-NCAR (Kalnay et al. 1996) or ERA-40 (Upalla et al. 2005) datasets, although prior to 1965 data quality have been shown to be less reliable

It is important to stress that any observed changes in ETC storm tracks, frequencies or intensities are highly dependent on broad-scale atmospheric modes of variability, and the noise associated with this variability is large in relation to any observed linear trend. Therefore, detection and attribution of long-term (decadal- to century-scale) changes in ETC activity is extremely difficult.

2.2.3.2.1 Variability of Extra-Tropical Cyclone Activity

Inter-annual and inter-decadal variability of ETC's is primarily driven by the location and other characteristics associated with the Polar jet stream. The mean location of the Polar jet is often referred to as the "storm track". The large-scale circulation is governed by the equator-to-pole temperature gradient, which is strongly modulated by SST's over the oceans. The magnitude of the equator-to-pole temperature gradient is of utmost

[46] These studies use *in situ* observations (both surface and upper-air), re-analysis fields, and Atmospheric-Ocean Global Climate Model (GCM) hind-casts

importance in determining the intensity of storms: the smaller (larger) the gradient in temperature, the weaker (stronger) the potential energy available for extra-tropical cyclone formation. The observed intensity of ETC's at the surface is related to the amplitude of the large-scale circulation pattern, with high-amplitude, negatively tilted troughs favoring stronger development of ETC's at the surface (Sanders and Gyakum 1980).

From a seasonal perspective, the strongest ETC's are temporally out of phase in the Pacific and Atlantic basins, since the baroclinic wave energy climatologically reaches a peak in late fall in the North Pacific and in January in the North Atlantic (Nakamura 1992; Eichler and Higgins 2006). While it remains unclear what the physical basis is for the offset in peak storm activity between the two basins, Nakamura (1992) showed statistically that when the Pacific jet exceeds 45 m s^{-1} there is a suppression of baroclinic wave energy, even though the low-level regional baroclinicity and strength of the Pacific jet are at a maximum (this effect is not evident in the Atlantic basin, since the peak strength of the jet across the basin rarely exceeds 45 m s^{-1}). Despite the observed seasonal difference in the peak of ETC activity, Chang and Fu (2002) found a strong positive correlation between the Pacific and Atlantic storm tracks using monthly mean reanalysis data covering 51 winters (1949 to 1999). They found the correlations between the two basins remained positive and robust over individual months during winter (DJF) or over the entire season (Chang and Fu 2002).

It has been widely documented that the track position, intensity and frequency of ETC's is strongly modulated on inter-annual time-scales by different modes of variability, such as the El Niño/Southern Oscillation (ENSO) phenomenon (Gershunov and Barnett 1998; An et al. 2007). In a recent study, Eichler and Higgins (2006) used both NCEP-NCAR and ERA-40 reanalysis data to diagnose the behavior of ETC activity during different ENSO phases. Their results showed that during El Niño events there is an equator-ward shift in storm tracks in the North Pacific basin, as well as an enhancement of the storm track along the U.S. East Coast. However, they found significant variability related to the magnitude of the El Niño event. During strong El Niños, ETC frequencies peak over the North Pacific and along the eastern U.S., from the southeast coast to the Maritime Provinces of Canada (Eichler and Higgins 2006), with a weaker track across the Midwest from the lee of the Rocky Mountains to the Great Lakes. During weak to moderate El Niños, the storm tracks are similar to the strong El Niños, except there is a slight increase in the number of ETC's over the northern Plains and the frequency of ETC activity decreases over the mid-Atlantic region. Similar to other previous studies (e. g. Hirsch et al. 2001; Noel and Changnon 1998), an inverse relationship typically exists during La Niñas; as the strength of La Niña increases, the frequency maxima of East Coast storms shifts poleward, the North Pacific storm track extends eastward toward the Pacific Northwest, and the frequency of cyclones increases across the Great Lakes region (Eichler and Higgins (2006).

In addition to ENSO, studies have shown that the Arctic Oscillation (AO) can strongly influence the position of storm tracks and the intensity of ETC's. Previous studies have shown that during positive AO conditions Northern Hemisphere cyclone activity shifts poleward (Serreze et al. 1997; Clark et al. 1999). Inversely, during negative AO conditions the polar vortex is weaker and cyclone activity shifts southward. Since the North Atlantic Oscillation (NAO) represents the primary component of the AO, it has a similar affect on storm tracks position, especially over the eastern North Atlantic basin (McCabe et al. 2001). For futher information on the different atmospheric modes of variability (Chapter 2, Box 2.3).

2.2.3.2.2 Changes in Storm Tracks and Extra-Tropical Cyclone Characteristics

Many studies have documented changes in storm track activity. Specifically, a significant pole-ward shift of the storm track in both the Pacific and Atlantic ocean basins has been verified by a number of recent studies that have shown a decrease in ETC frequency in mid-latitudes, and a corresponding increase in ETC activity in high-latitudes (Wang et al. 2006a; Simmons and Keay 2002; Paciorek et al. 2002; Graham and Diaz 2001; Geng and Sugi 2001; McCabe et al. 2001; Key and Chan 1999; Serreze et al. 1997). Several of these studies have examined changes in storm tracks over the entire Northern Hemisphere (i.e. McCabe et al. 2001; Paciorek et al. 2002; Key and Chan 1999), while several others have focused on the storm track changes over the Pacific (i.e., Graham and Diaz 2001) and Atlantic basins (i.e., Geng and Sugi 2001), or both (i.e., Wang and Swail 2001). Most of these studies focused on changes in frequency and intensity observed during winter (DJF) or the entire cold season (Oct-Mar). However, for spring, summer and autumn, Key and Chan (1999) found opposite trends in 1000-hPa and 500-hPa cyclone frequencies for both the mid- and high latitudes of the Northern Hemisphere.

The standardized annual departures[47] of ETC frequency for the entire Northern Hemisphere over the period 1959-1997 (Fig. 2.18a,b; McCabe et al. 2001) shows that cyclone frequency has decreased for the mid-latitudes (30^{o}-60^{o}N) and increased for the high latitudes (60^{o}-90^{o}N). For the 55-year period of 1948-2002, a metric called the Cyclone Activity Index (CAI)[48] was developed by Zhang et al. (2004) to document the variability of Northern Hemisphere cyclone activity. The CAI has increased in the Arctic Ocean (70^{o}-90^{o}N) during the latter half of the 20th century, while it has decreased in mid-latitudes (30^{o}-60^{o}N) from 1960 to 1993, which is evidence of a pole-ward shift in the average storm track position. Interestingly, the number and intensity of cyclones entering the Arctic from the mid-latitudes has increased, particularly during summer (Zhang et al. 2004). The increasing activity in the Arctic was more recently verified by Wang et al. (2006a), who analyzed ETC counts by applying two separate cyclone detection thresholds to ERA-40 reanalysis of mean sea level pressure data. Their results showed an increase in high latitude storm counts, and a decrease in ETC counts in the mid-latitudes during the latter half of the 20th century.

Northern Hemisphere ETC intensity has increased over the period 1959-1997 across both mid- and high-latitudes cyclone intensity (McCabe et al. 2001; Fig. 2.18c,d), with the upward trend more significant for the high latitudes (0.01 level) than for the mid-latitudes

[47] Standardized departures (*z* scores) were computed for each 5^{o} latitudinal band by subtracting the respective 1959-1997 mean from each value and dividing by the respective 1959-1997 standard deviation (McCabe et al. 2001).
[48] The CAI integrates information on cyclone intensity, frequency, and duration into a comprehensive index of cyclone activity. The CAI is defined as the sum over all cyclone centers, at a 6-hourly resolution, of the differences between the cyclone central SLP and the climatological monthly mean SLP at corresponding grid points in a particular region during the month (Zhang et al. 2004).

(0.10 level). From an ocean basin perspective, the observed increase in intense ETC's appears to be more robust across the Pacific than the Atlantic. Using reanalysis data covering the period 1949-1999, Paciorek et al. (2002) found that extreme wind speeds have increased significantly in both basins (Fig. 2.19a,d). Their results also showed that the observed upward trend in the frequency of intense cyclones has been more pronounced in the Pacific basin (Fig. 2.19c), although the inter-annual variability is much less in the Atlantic (Fig. 2.19f). Surprisingly, they found that the overall counts of ETC's showed either no long-term change, or a decrease in the total number of cyclones (Fig. 2.19b,e). However, this may be a result of the large latitudinal domain used in their study (20^{o}-70^{o}N), which included parts of the tropics, sub-tropics, mid- and high latitudes.

On a regional scale, ETC activity has increased in frequency, duration and intensity in the lower Canadian Arctic during 1953-2002 with the most statistically significant trends during winter[49] (p=0.05 level; Wang et al. 2006b). In contrast to the Arctic region, cyclone activity was less frequent and weaker along the southeast and southwest coasts of Canada. Winter cyclone deepening rates (i.e. rates of intensification) have increased in the zone around 60^{o}N, but decreased further south in the Great Lakes area and southern Prairies-British Columbia region of Canada. This is also indicative of a pole-ward shift in ETC activity, and corresponding weakening of ETC's in the mid-latitudes and an increase in observed intensities in the high latitudes. For the period of 1949-1999, the intensity of Atlantic ETC's increased from the 1960's to the 1990's during the winter

[49] Results based on hourly average sea level pressure data observed at 83 stations

season[50] (Harnik and Chang 2003). Their results showed no significant trend in the Pacific region but this is a limited finding because of a lack of upper-air (i.e. radiosonde) data over the central North Pacific[51] in the region of the storm track peak (Harnik and Chang 2003).

There have been very few studies that have analyzed the climatological frequencies and intensities of ETC's across the central U.S., specifically in the Great Lakes region (e.g., Lewis 1987; Harmon et al. 1980; Garriott 1903). Over the period 1900 to 1990 the number of strong cyclones (≤992 mb) increased significantly across the Great Lakes (Angel and Isard 1998). This increasing trend was evident (at the p=0.05 level) both annually and during the cold season,. In fact, over the 91-yr period analyzed, they found that the number of strong cyclones per year more than doubled during both November and December.

In addition to studies using reanalysis data, which have limited record lengths, other longer-term studies of the variability of *storminess* typically use wave or water level measurements as proxies for storm frequency and intensity. Along the U.S. West Coast, one of the longest continuous climate-related instrumental time series in existence is the hourly tide gauge record at San Francisco that dates back to 1858. A derived metric called non-tide residuals (NTR)[52], which are related to broad-scale atmospheric

[50] Results based on gridded rawinsonde observations covering the Northern Hemisphere
[51] Besides the few radiosonde sites located on islands (i.e., Midway or the Azores), most upper-air observations over the vast expanses of the North Pacific and Atlantic are from automated pilot reports (pireps) that measure temperature, wind speed, and barometric pressure onboard commercial aircraft traveling at or near jet stream level (between 200-300 hPa).
[52] Non-tide residuals are obtained by first removing the known tidal component from the water level variations using a spectral method; then, variations longer than 30 days and shorter than 2.5 days are

circulation patterns across the eastern North Pacific that affect storm track location, provides a measure of *storminess* variability along the California coast (Bromirski et al. 2003). Average monthly variations in NTR, which are associated with the numbers and intensities of all ETCs over the eastern North Pacific, did not change substantially over the period 1858-2000 or over the period covered by most ETC reanalysis studies, 1951-2000. However, the highest 2% of extreme winter NTR (Fig. 2.20), which are related to the intensity of the most extreme ETCs, had a significant upward trend since ~1950, with a pronounced quasi-periodic decadal-scale variability that is relatively consistent over the last 140 yr. Changes in storm intensity from the mid-1970s to early 1980s are also suggested by a substantial pressure decreases at an elevation above sea level of about 3000 m over the eastern North Pacific and North America (Graham 1994), indicating that the pattern of variability of extreme storm conditions observed at San Francisco (as shown in Fig. 2.20) likely extends over much of the North Pacific basin and the U.S. The oscillatory pattern of variability is thought to be influenced by teleconnections from the tropics, predominately during ENSO events (Trenberth and Hurrell 1994), resulting in a deepened Aleutian low shifted to the east that causes both ETC intensification and a shift in storm track. It is interesting to note that peaks in the 5-year moving average in Fig. 2.20 generally correspond to peaks in extreme rainfall in Fig. 2.10 suggesting that the influence of El Niño and broad-scale atmospheric circulation patterns across the Pacific that affect sea level variability along the West Coast are associated with storm systems that affect rainfall variability across the U.S..

The amplitude and distribution of ocean wave energy measured by ocean buoys is

removed with a bandpass filter.

determined by ETC intensity and track location. Changes in long period (>12 sec), intermediate period, and short period (<6 sec) components in the wave-energy spectra permit inferences regarding the changes over time of the paths of the storms, as well as their intensities and resulting wave energies (Bromirski et al. 2005). Analysis of the combination of observations from several buoys in the eastern North Pacific supports a progressive northward shift of the dominant Pacific storm tracks to the central latitudes (section 2.2.3.3).

2.2.3.2.3 Nor'easters

Those ETCs that develop and propagate along the East Coast of the U.S. and southeast Canada are often termed colloquially as *Nor'easters*[53]. In terms of their climatology and any long-term changes associated with this subclass of ETCs, there are only a handful of studies in the scientific literature that have analyzed their climatological frequency and intensity (Jones and Davis 1995), likely due to a lack of any formal objective definition of this important atmospheric phenomenon (Hirsch et al. 2001).

Because waves generated by ETCs are a function of storm size and the duration and area over which high winds persist, changes in significant wave heights can also be used as a proxy for changes in Nor'easters. Using hindcast wave heights and assigning a minimum criterion of open ocean waves greater than 1.6 m in height (a commonly used threshold for storms that caused some degree of beach erosion along the mid-Atlantic coast) to qualify as a nor'easter, the frequency of nor'easters along the Atlantic coast peaked in the

[53] According to the *Glossary of Meteorology* (Huschke 1959), a *nor'easter* is any cyclone forming within 167 km of the East Coast between 30°-40°N and tracking to the north-northeast

1950's, declined to a minimum in the 1970's, and then increased again to the mid-1980's (Dolan et al. 1988; Davis et al. 1993).

An alternate approach utilized by Hirsch et al. (2001) uses pressure, direction of movement and wind speed to identify such systems and generically names them as East Coast Winter Storms (ECWS)[54]. Hirsch et al. (2001) defined an ECWS as "strong" if the maximum wind speed is greater than 23.2 m s^{-1} (45 kt). During the period of 1951-1997, their analysis showed that there were an average of 12 ECWS events occurring each winter (October-April), with a maximum in January, and an average of 3 strong events (Fig. 2.21a). They found a general tendency toward weaker systems over the past few decades, based on a marginally significant (at the 90% confidence level) increase in average storm minimum pressure (not shown). However, their analysis found no statistically significant trends in ECWS frequency for all nor'easters identified in their analysis, for those storms that occurred over the northern portion of the domain (>35^{o}N), or those that traversed full coast (Fig. 2.21b,c) during the 46-year period of record used in this study.

Because strong storms over the open ocean generate high amplitude waves, buoy measurements of wave height and wave period can be used to infer characteristics of ETC variability. The wave power index (WPI) of strong storm-forced wave events

[54] According to Hirsch et al. (2001), in order to be classified as an ECWS, an area of low pressure is required to (1) have a closed circulation; (2) be located along the east coast of the United States, within the quadrilateral bounded at 45^{o}N by 65^{o} and 70^{o}W and at 30^{o}N by 75^{o} and 85^{o}W; (3) show general movement from the south-southwest to the north-northeast; and (4) contain winds greater than 10.3 m s^{-1} (20 kt) during at least one 6-h period.

(significant wave heights > 3 m) measured at deep-water open-ocean NOAA buoys 44004, 41001, 41002 along the U.S. Atlantic coast (see Figure 2.25 for locations) during winter months (October-March, excluding tropical cyclone wave events) shows a decreasing trend that is significant at the p=0.05 level amounting to a decrease in ETC-forced wave power of about 1%/yr (Bromirski 2007). Coupled with no statistically significant change in either mean wave height or the number of measurements exceeding 3 m (implying no change in storm duration and/or the number of strong storms), the downward trend in the WPI suggests that winter ETC intensity has decreased since 1980, in general agreement with Hirsch et al. (2001).

BOX 2.2: Extreme Coastal Storm Impacts: "The Perfect Storm" as a True Nor'easter: From a coastal impacts perspective, damage is greatest when large storms are propagating *towards* the coast, which generally results in both a larger storm surge and more long period wave energy (resulting in greater run-up causing more beach/coastal erosion/damage). Storm intensity (winds) is usually greatest in the right-front quadrant of the storm (based on the cyclone's forward movement), so the typical track of east coast winter storms propagating parallel to the coast leaves the most intense part of the storm out to sea. In contrast to storms propagating parallel to the coast, Nor'easters (such as "the Perfect Storm") that propagate from east-to-west in a retrograde track at some point in their lifetime (Fig. 2.22) can generate much greater surge and greater long period wave energy, and also potentially have the most intense associated winds making landfall along the coast.

2.2.3.3 Coastal Waves: Trends of Increasing Heights and Their Extremes

The high wind speeds of hurricanes and extratropical cyclones over bodies of water cause extremes in the heights and energies of the waves they generate. Seasonal and long-term changes in storm intensities and their tracks produce corresponding variations in wave heights and periods along coasts, defining their wave climates. Waves generated by extratropical storms dominate the oceans at higher latitudes, including the Northeast Pacific along the shores of Canada and the west coast of the United States, and along the Atlantic coast of North America where they originate from destructive Nor'easters. Tropical cyclones dominate the wave climates at lower latitudes during the warm season (June-September), including the southeast Atlantic coast of the United States, Gulf of Mexico, and the Caribbean, while hurricanes in the East Pacific generate waves along the western shores of Mexico and Central America . The magnitude of associated damage from storm waves depends to a large extent on whether the storms make landfall, when storm surge, high winds, and heavy rainfall combined with high waves cause severe impacts. However, high waves from strong tropical cyclones that reach hurricane strength and then track northward along the East Coast as they weaken, can combine with extratropical systems, such as the 1991 Halloween Storm (Bromirski 2001; Chapter 2, Box 2.2), and cause severe coastal erosion and have significant economic impacts (Davis et al. 1993; Dolan et al. 1988; Mather et al. 1967).

2.2.3.3.1 The Waves of Extratropical Storms and Hurricanes

The heights and periods of waves generated by a storm depend on the speed of its winds, the area over which the winds blow (the storm's fetch), and on the duration of the storm, factors that determine the amount of energy transferred to the waves. Wave climate variability has been estimated from: (1) direct measurements by buoys; (2) visual observations from ships; (3) wave hindcast analyses where wave heights and periods are assessed using forecast models that are run retrospectively using observed meteorological data; and (4) in recent years from satellite altimetry. The reliability of the wave records ranges widely for these different sources, and changes in data-collection methodologies and processing techniques can affect the data consistency. However, long records from these sources make it possible to identify long-term trends, and to investigate underlying climate controls.

In the Northern Hemisphere the hurricane winds are strongest on the right-hand side of the storm relative to it track, where its cyclonic winds coincide with the direction of the storm's propagation, in turn producing the highest waves on that side of the storm. They achieve their greatest heights in proximity to the wall of the storm's eye where the winds reach their maximum, and systematically decrease outward as the wind speeds are reduced. Extreme heights are closely associated with the Saffir-Simpson hurricane classification system, where the central atmospheric pressures are lower and the associated wind speeds are higher for the higher hurricane categories. A correlation between the meausred wave heights and the central atmospheric pressure (Hsu et al. 2000) allows the magnitude of the significant wave height[55], H_S,to be related to the hurricane categories[56]. Estimates of the maximum H_S generated close to the wall of the hurricane's eye on the storm's leading right quadrant where the wind speeds are greatest, range from 6 to 7 m for Category 1 storms to about 20 m and greater for Category 5 storms. The decrease in observed H_S outward from the hurricane's eye in response to the

[55] The "significant wave height" is a commonly used statistical measure for the waves generated by a storm, defined as the average of the highest one-third of the measured wave heights

[56] Hsu et al. (2000) have developed the empirical formula H_{smax}=0.2(P_n-P_c) where P_c and P_n ~ 1013 mbar are respectively the atmospheric pressures at the center and edge of the tropical cyclone, and H_{smax} is the maximum value of the significant wave height

outward decrease in wind speeds, demonstrates that H_S is reduced by 50% at approximately a distance of 5 times the radius of the eye, typically occurring about 250 km outward from the storm's center (Hsu, et al. 2000).

The impression has been, however, that such models under-predict the highest waves of Category 4 and 5 storms, and this has led to recent investigations that included the direct measurement of waves generated by hurricanes. For example, measurements obtained by six wave gauges deployed by the Naval Research Laboratory (NRL) at depths of 60 to 90 m in the Gulf of Mexico, when the Category 4 Hurricane Ivan passed directly over the array on 15 September 2004, recorded significant wave heights ranging from 16.1 to 17.9 m; the largest individual wave height reached 27.7 m (Wang et al. 2005). The simple model of Hsu et al. (2000) yields a maximum significant wave height of 15.6 m for Ivan's 935-mbar central pressure, seemingly in agreement with the 16-m measured waves. However, the NRL gauges were about 30 km outward from the zone of strongest winds and were positioned toward the forward face of Ivan rather than in its right-hand quadrant, so Wang et al. (2005) concluded it is likely that the maximum significant wave height was greater than 21 m, with the largest individual wave heights having been greater than 40 m, indicating that the Hsu et al. (2000) empirical formula somewhat under predicts the waves generated by high-category hurricanes. On the other hand, hurricane waves from more complex models that use spatially distributed surface wind measurements (Tolman et al. 2002) compare well with satellite and buoy observations both in deep water and in shallow water as hurricanes make landfall (Moon et al. 2003).

Any trend over the years of increasing intensities of hurricanes or of extratropical storms should on average be reflected in similar upward trends in associated wave heights. Analyses of wave-buoy data along both the Atlantic and Pacific coasts of the United States document that wave-height increases have occurred at some locations since the late 1970s.

2.2.3.3.2 Atlantic Coast Waves

Two analyses have recently been undertaken of the hourly measurements of the significant wave heights collected by the buoys of NOAA's National Data Buoy Center (NDBC) along the U.S. Atlantic shore. These analyses, while differing in some important methodological aspects that affect some of the results, both show changes in waves generated by hurricanes while the ranges of wave heights created by extratropical storms appear to have undergone little change.

Komar and Allan (2007a) analyzed the data from three buoys located in deep water to the east of Cape May, New Jersey, Cape Hatteras, North Carolina, and offshore from Charleston, South Carolina. These buoys were selected due to their long record lengths and because the sites represent a range of latitudes where the wave climate is expected to be affected by both tropical hurricanes and extratropical storms (Nor'easters). Separate analyses were undertaken for the winter season dominated by extratropical storms and the summer season of hurricanes[57]. There was not a statistically significant change over the decades in the heights of waves generated by extratropical storms, but statistically significant increases have occurred for the hurricane-generated waves. The increases in annual-averaged significant wave heights measured by the

[57] The hurricane waves were analyzed for the months of July through September, expected to be dominated by tropical cyclones, while the waves of extratropical storms were based on the records from November through March; transitional months such as October were not included, when both types of storms could be expected to be important in wave generation. Also, strict missing data criteria eliminated some years from the analysis.

three Atlantic buoys for the summer hurricane seasons are graphed in Figure 2.23. These annual averages have included only occurrences when the significant wave heights were greater than 3 m, it having been found that those higher waves can be directly attributed to specific hurricanes, whereas the lower waves represent the calmer periods between storms. It is seen in Figure 2.23 that there has been a dependence on the latitude, with the highest rate of increase having occurred in the south; 0.059 m/yr (1.8 m in 30 years) for the Charleston buoy, 0.024 m/yr for the Hatteras buoy, and 0.017 m/yr for Cape May[58].

Figure 2.24 provides a comparison of histograms for the numbers of significant wave heights measured during the hurricane season by the Cape Hatteras buoy, one histogram for data from early in its record (1977-1990) and the second from 1996-2005, this comparison further documenting the decadal increase seen in Figure 2.23, especially of the more-extreme waves[59]. The histogram for the early decade in the wave record shows that the maximum significant wave height measured was 7.8 m, providing an approximate estimate for the height expected to have a 10-year recurrence interval. From this, we could expect that the 100-year extreme (1% probability) would have been on the order of 9.5 m significant wave height. In contrast, during 1996-2005 there has been a considerably larger number of occurrences having significant wave heights greater than 4 m, with the most extreme heights measured ranging up to 10.3 m. The 100-year extreme is now on the order of 12 m, about 3 m higher than in the 1980s. Similar

[58] The regressions in Figure 2.38 for the Charleston and Cape Hatteras buoy data are statistically significant at the p=0.05 level according to the Wilcoxon Test, whereas the value of the trend for the Cape May does not pass that test.

[59] Traditionally a wave histogram is graphed as the percentages of occurrences, but here the actual numbers of occurrences for the range of wave heights have been plotted, using a log scale that emphasizes the most-extreme heights.

results have been found in analyses of the wave-height histograms for the Cape May and Charleston buoys (Komar and Allan, 2007a).

This analysis of the three U.S. East Coast buoys (Figures 2.23 and 2.24) demonstrate that there has been a 30-year increase in wave heights measured during the hurricane season. This increase could depend on several factors, including changes from year to year in the numbers and intensities of storms, their tracks that determine whether they traveled northward through the Atlantic where their generated waves could be recorded by these buoys, and how distant the hurricanes were from the buoys, whether they passed far offshore within the central Atlantic, or approached the east coast and possibly made landfall. Analyses by Komar and Allan (2007b) indicate that all of these factors have been important to the observed wave-height increases, but the increased hurricane intensities found by Emanuel (2005) based on the measured wind speeds provide the best explanation for the progressive increase in wave heights seen in Figure 2.23, since the numbers and tracks of the storms show considerable variability from year to year.

In the second study (Bromirski and Kossin, 2007)[60], extreme tropical cyclone-associated H_S events (deep water H_S exceeding 3 m) measured at buoys in both the Atlantic and Gulf regions (Figure 2.25a) show a general tendency for more significant tropical cyclone-associated wave events since 1995 (Figure 2.25b), consistent with increasing overall counts of named storms during recent years [Webster et al. 2005; Klotzbach 2006]. As would be expected, the intense 2005 hurricane season had the highest incidence of significant H_S events over the data record in

[60] In this study, the entire hurricane season (June-November) was analyzed. Hurricane track data were used to restrict the analysis to time periods when hurricanes were likely the cause of extreme waves, the goal being to minimize the effects of ETCs during the transition months of October and November. Less stringent missing data were applied, resulting in the inclusion of more years than in Komar and Allan (2007).

the Gulf when Hurricanes Katrina, Rita, and Wilma occurred. Since 1978, there were substantially more significant H_S events along the Atlantic coast than in the Gulf, with almost three times as many events during September (Figure 2.25c). The monthly distribution along both coasts peaks in September, with an equally likely chance of a significant tropical cyclone wave event occurring during October as during August over the 1978-2006 data record. About 3 times as many extreme events occurred in September in the Atlantic compared with the Gulf from 1978-2006. However, inclusion of all tropical cyclone generated wave events (listed in http://www.nhc.noaa.gov/pastall.shtml) for the entire June though November hurricane season indicates that there is no significant trend in mean tropical cyclone associated H_S at either the western North Atlantic or Gulf buoys (Bromirski and Kossin, 2007; Figure 2.25b).

A tropical cyclone wave power index, WPI[61], shows an increase in the Atlantic during the mid-1990s (Bromirski and Kossin, 2007; Figure 2.26), associated with an increase in the number of significant tropical cyclone forced wave events, that is proportionally consistent with the increase observed for the tropical cyclone power dissipation index (PDI, Emanuel 2005]\). The Gulf WPI indicates that only the 2005 hurricane season was exceptional in the Gulf, but is highly correlated with the Atlantic multidecadal oscillation (AMO, Goldenberg et al. 2001) over the 1980-2006 period. In contrast, the Atlantic WPI is not well correlated with the AMO, suggesting that tropical sea surface temperature variability has a greater influence on the characteristics of tropical cyclones that reach the Gulf.

[61] The WPI for the Atlantic and Gulf regions is obtained as the average of the total wave power for all tropical cyclone associated wave events during the June – November hurricane season at the three southernmost Atlantic buoys and the three Gulf buoys in Figure 2.waves.1a (Bromirski, 2007).

To summarize, these 2 studies both detect changes in tropical cyclone-related waves, but in different aspects. Komar and Allan (2007a) show statistically significant increases in extreme wave heights during July-September, while Bromirski and Kossin (2007) do not find the trends over the entire hurricane season to be statistically significant. However, Bromirski and Kossin (2007) do find a statistically significant increase in tropical cyclone-caused wave power, a trend that is attributed to an increase in numbers of events rather than intensity.

In contrast to the changes in the hurricane waves, analyses of the winter wave heights generated by extratropical storms and recorded since the mid-1970s by the three buoys along the central U.S. Atlantic shore have shown little change (Komar and Allan, 2007a). The records from the Cape Hatteras and Charleston NDBC buoys yield regressions indicating that they have actually experienced a slight decrease over the decades (-0.005 m/yr), while the Cape May buoy shows a lower rate of reduction (-0.001 m/yr). These trends are not statistically significant, but may a reflection in the changes in storm tracks over the decades, with the storms having shifted to the north.

Analyses of the winter waves generated by extratropical storms demonstrate that the highest measured occurrences are on the order of 10.5-m significant wave heights, with the extreme-value assessments placing the 100-year event at on the order of 11.5 m, effectively the same as seen in the histogram of Figure 2.24 for the summer hurricane waves recorded by the Hatteras buoy during the 1996-2005 decade, so the wave climates of the two seasons are now quite similar. However, thirty-years ago when these buoys first became operational, the significant wave heights generated in the summer by hurricanes were much lower than those of the extratropical storms during the winter; while the heights of hurricane-generated waves have progressively increased since the 1970s, the wave heights due to extratropical storms have not.

Although minimal change in the heights of waves generated by extratropical storms have been measured by buoys along the U.S. shore in the Western Atlantic, progressive increases have occurred in the Northeast Atlantic extending back to at least the 1960s, documented by the Seven Stones ship-borne wave recorder located in deep water off the southwest coast of England (Carter and Draper, 1988; Bacon and Carter, 1991). Of interest, the rate of increase (0.022 m/yr) in the annual averages are closely similar to those measured by buoys along the northwest coast of the United States in the Pacific Ocean, discussed below.

The documentation by buoys of trends in wave heights in the North Atlantic are limited by their relatively short records, hindering a determination of the longevity of the identified trends and the possible presence of any decadal cycles in climate-determined variability. To supplement the buoy data, visual observations from ships in transit provide longer time series of estimated ocean wave-heights; although the quality of the data may be questionable, its availability extends back through the entire 20th century, and in general appears to yield reasonably consistent trends when compared with the buoy data and with wave hindcasts. Gulev and Grigorieva (2004) have undertaken detailed analyses of the visual assessments of wave heights from ships, covering the years 1895-2002 except for a gap in the data during World War II. The observations for the northeast Atlantic showed a distinct increase in wave heights after about 1955, corresponding to the wave-sensor measurements since the 1960s collected southwest of England. Earlier in the 20th century, however, there were distinct cycles in the visual wave heights observed from ships, with years during which the average wave heights were some 0.2 m higher than at present. These cycles correlate with the North Atlantic Oscillation (NAO), with the higher wave heights having been associated with high NAO indices.

Hindcasts by Wang and Swail (2001) of the wave climates based on the meteorological records of extratropical storms have been analyzed with respect to changes in the 90th and 99th percentiles of the significant wave heights, thereby representing the trends for the more extreme wave conditions. The results indicate a lack of change along the east coast of North America, in agreement with the buoy data for waves generated by extratropical storms.

2.2.3.3.3 Pacific Coast Waves

Analyses of the wave data from NDBC buoys have also been undertaken along the U.S. Pacific coast, similar to those discussed above for the Atlantic but with the focus having been on the waves generated by extratropical storms in the Northeast Pacific. The principal investigations of the trends of changing wave heights and their potential climate controls are those of Allan and Komar (2000, 2006), who analyzed the records from 6 buoys along the coast from Washington to south-central California (Point Conception). The analyses were limited to the “winter” waves, October though the following March, the season with the most intense storms and highest waves. Trends of increasing wave heights spanning the past 30 years were found, with the greatest rate of increase having occurred off the coast of Washington where the regression yielded an average rate of 0.032 m/yr for the winter, with a regular pattern of lesser rates of increase for the latitudes to the south, such that off the coast of south-central California there has not been a statistically significant trend[62].

Analyses of the more extreme wave heights measured off the Washington coast were undertaken due to their importance to coastal-erosion occurrences (Allan and Komar, 2006). Figure 2.27 contains graphs of the annual averages of the winter wave heights, and the averages of the five largest significant wave heights measured each winter, the latter showing a higher rate of increase (0.095 m/yr, a 2.85-m increase in the significant wave heights in 30 years). The full series of analyses are listed in Table 2.1, demonstrating that there is an orderly progression with the more extreme the assessment the greater the rate of increase, up to a rate of 0.108 m/yr for the single highest measured significant wave height each year. While the data in Figure 2.27 for the averages of the largest five storm-wave occurrences each year are statistically significant at the $p=0.05$ level, the trends for the more extreme waves do not meet this criterion (Table 2.1). However, for applications to engineering design of coastal structures and in coastal management assessments of hazards, these extremes for the measured wave heights are of greatest relevance, and therefore are sometimes used in applications as is the trend for the assessment of the 100-year projected extreme, which has increased at a still greater rate over the decades, from about 11 m in 1975 to 16 m at present. This use in applications is further supported by the fact that much of the scatter in the diagrams, as seen in Figure 2.27, can be accounted for by considering the range of climate events from El Niños to La Niñas (Allan and Komar, 2000, 2006).

[62] Where trends of increasing wave heights do exist, they have again been verified by application of the Wilcoxon test, a statistical analysis that basically compares the first half the record with the second half to establish that there has been a meaningful change.

The intensities of North Pacific extratropical storms and their associated tracks are strongly affected by the depth and position of the Aleutian Low, which tends to intensify and shift southward and eastward during strong El Nino events (Mo and Livezey, 1986). This southward shift results in increased occurrences of extreme waves throughout the eastern North Pacific, particularly along the south-central California coast (Seymour et al. 1984; Allan and Komar 2000, 2006; Bromirski et al. 2005). Correlations between the measured wave heights and the Multivariate ENSO Index show that increased wave heights occur at all latitudes along the U.S. Pacific coast during major El Niños, but with the greatest increases along the shore of southern California (Allan and Komar, 2006). Along the coast of California where the trends of decadal increases are small to non-existent, it is this cycle between El Niños and La Niñas that exerts the primary climate control on the storm-wave heights and their extremes (and also on the monthly-mean winter water levels, which are elevated by 20 to 50 cm during a major El Niño above the long-term mean sea levels).

The documentation of increasing wave heights in the North Pacific is given limited by the relatively short records from buoys, extending back only to the 1970s. Similar to discussed for the Atlantic, visual observations from ships in transit provide longer time series of ocean wave height estimates, but of questionable quality. Gulev and Grigorieva (2004)examined this source of wave data for the North Pacific, finding that there has been a general increase in the significant wave heights throughout the 20th century, with a rapid increase from 1900 to about 1925, and a leveling off from 1925 to about 1950-60 but with an apparent maximum during the 1940s (there being a gap in the data during World War II). There was a renewed increase beginning in about 1960, corresponding to that documented by the wave buoy measurements

(Fig. 2.27). The wave hindcasts[63] by Wang and Swail (2001), representing the more extreme significant wave-height occurrences (the 90th and 99th percentiles), largely also confirm the general increase in wave heights throughout the central to eastern North Pacific.

There is the potential for the use of proxy evidence to examine the changes in wave heights back beyond that provided by the wave data, the proxy having the clearest potential being measurements by seismometers installed to monitor earthquake activity. During the “quiet” intervals between earthquakes it has been noted that there is a consistent level of "noise" in the recorded ground motions, termed "microseisms". It has been shown that much of this energy is derived from surf on the coast, with the microseisms increasing at times of storms. Analyses have been undertaken by Bromirski et al. (1999) correlating buoy measurements of ocean waves along the coast of central California and the microseisms measured by the seismometer at the University of California, Berkeley. The results of that study yielded a calibration between the ocean wave heights and the microseism energy, demonstrating the potential use of the archived seismic data that dates back to 1930, to investigate changes in the U.S. West Coast wave climate.

2.2.3.4 Winter Storms

2.2.3.4.1 Snowstorms

The amount of snow that causes serious impacts varies depending on a given location’s usual snow conditions. A snowstorm is defined here as an event in which more than 15 cm of snow falls in 24 hours or less at some location in the U.S. This is an amount

[63] Hindcasts are model estimates of waves using forecast models that are run retrospectively using observed meteorological data

sufficient to cause societally-important impacts in most locations. During the 1901-2001 period, 2,257 snowstorms occurred (Changnon et al. 2006). Temporal assessment of the snowstorm incidences during 1901-2000 revealed major regional differences. Comparison of the storm occurrences in 1901-1950 against those in 1951-2000 revealed that much of the eastern U.S. had more storms in the early half of the 20th Century, whereas in the West and New England, the last half of the century had more storms. Nationally, 53% of the weather stations had their peaks in 1901-1950 and 47% peaked in 1951-2000.

The South and lower Midwest had distinct statistically significant downward trends in snowstorm frequency from 1901 to 2000. In direct contrast, the Northeast and upper Midwest had statistically significant upward linear trends. These contrasting regional trends suggest a northward shift in snowstorm occurrence. Nationally, the regionally varying up and down trends resulted in a national storm trend that was slightly upward for 1901-2000, but not statistically significant. The long-term increases in the upper Midwest and Northeast occurred where snowstorms are most frequent, and thus had an influence on the upward trend in national snowstorm activity. Research has shown that cyclonic activity was low during 1931-1950, a period of few snowstorms in the U.S.

Nationally, 39 of 231 stations with long-term records had their lowest frequencies of storms during 1931-1940, whereas 29 others had their peak of incidences then. The second ranked decade with numerous stations having low snowstorm frequencies was 1981-1990. Very few low storm occurrences were found during 1911-1920 and in the

1961-1980 period, times when storms were quite frequent. The 1911-1920 decade had the greatest number of high station values with 38 stations. The fewest peak values occurred in the next decade, 1921-1930. Comparison of the decades of high and low frequencies of snowstorms reveals, as expected, an inverse relationship. That is, when many high storm values occurred, there are few low storm frequencies.

Generally, the decades with high snowstorm frequencies were characterized by cold winters. The three highest decades for snowstorms (1911-1920, 1961-1970, and 1971-1980) were ranked 1st, 4th, and 3rd coldest, respectively while the two lowest decades (1921-1930 and 1931-1940) were ranked as 3rd and 4th warmest. One exception to this general relationship is the warmest decade (1991-2000), which experienced a moderately high number of snowstorms.

Very snowy seasons (those with seasonal snowfall totals exceeding the 90th percentile threshold) were infrequent in the 1920s and 1930s and have also been rare since the mid-1980s (Kunkel et al. 2007b). There is a high correlation with average winter temperature. Warm winters tend to have few stations with high snowfall totals and most of the snowy seasons have also been cold.

Some of the snowiest regions in North America are the southern and eastern shores of the Great Lakes where cold northwesterly winds flowing over the warmer lakes pick up moisture and deposit on the shoreline areas. There is evidence of upward trends in snowfall since 1951 in these regions even while locations away from the snowy shoreline areas have not experienced increases (Burnett et al. 2003). An analysis of historical heavy lake-effect snowstorms identified several weather conditions to be closely related to heavy lake-effect snowstorm occurrence including moderately high surface wind speed, wind direction promoting a long fetch over the lakes, surface air temperature in the range of -10 to 0°C, lake surface to air temperature difference of at least 7°C, and an unstable lower troposphere (Kunkel et al. 2002). It is also necessary that the lakes be mostly ice-free.

Snow cover extent for North America based on satellite data (Robinson et al. 1993) abruptly decreased in the mid-1980s and generally has remained low since then (http://climate.rutgers.edu/snowcover/chart_anom.php?ui_set=0&ui_region=nam&ui_month=6).

2.2.3.4.2 Ice Storms

Freezing rain is a phenomenon where even light amounts can have substantial impacts. All days with freezing rain (ZR) were determined during the 1948-2000 period based on data from 988 stations across the U.S. (Changnon and Karl 2003). The national frequency of freezing rain days (FZRA) exhibited a downward trend, being higher during 1948-1964 than in any subsequent period.

The temporal distributions of FZRA for three climate regions (Northeast, Southeast, and South) reveal substantial variability. They all were high in 1977-1980, low in 1985-1988, and lowest in 1973-1976. The 52-year linear trends for all three regions were downward over time. The time distributions for the Central, West North Central, and East North Central regions are alike, all showing that high values occurred early, 1949-1956. All climate regions had their lowest FZRA during 1965-1976. The East north central, Central, Northwest, and Northeast regions, which embrace the northern half of the conterminous U.S., all had statistically significant downward linear trends. This is in contrast to trends in snowstorm incidences.

Both snowstorms and ice storms are often accompanied or followed by extreme cold because a strong ETC (which is the meteorological cause of the snow and ice) is one of the meteorological components of the flow of extreme cold air from the Arctic. This compounds the impacts of such events in a variety of ways, including increasing the risks to human health and adversely affecting the working environment for snow removal and repair activities. While there have been no systematic studies of trends in such compound events, observed variations in these events appear to be correlated. For example, the late 1970s were characterized both by a high frequency of extreme cold (Kunkel et al. 1999) and a high frequency of high snowfall years (Kunkel et al. 2007b).

2.2.3.5 Convective Storms

Thunderstorms in the United States are defined to be severe by the National Weather Service (NWS) if they produce hail of at least 1.9 cm (3/4 inch) in diameter, wind gusts of at least 25.5 m s^{-1} (50 kt) or a tornado. Currently, reports come from a variety of sources to the local NWS forecast offices that produce a final listing of events for their area. Over the years, procedures and efforts to produce that listing have changed. Official data collection in near real-time began in 1953 for tornadoes and 1955 for hail and wind. Prior to 1973, tornado reports were verified by state climatologists (Changnon 1982). In addition, efforts to improve verification of severe thunderstorm and tornado warnings, the introduction of Doppler radars, changes in population, and increases in public awareness, have led to increases in reports over the years. Changes in reporting practices have also led to inconsistencies in many aspects of the records (e.g., Brooks 2004). Changnon and Changnon (2000) identified regional changes in hail frequency from reports made at official surface observing sites. With the change to automated surface observing sites in the 1990s, the number of hail reports at those locations dropped dramatically because of the loss of human observers at the sites. As a result, comparisons to the Changnon and Changnon work cannot be continued, although Changnon et al. (2001) have attempted to use insurance loss records as a proxy for hail occurrence.

The raw reports of annual tornado occurrences show an approximately doubling from 1954-2003 (Brooks and Dotzek 2007), a reflection of the changes in observing and reporting. When detrended to remove this artificial trend, the data show large interannual variability, but a persistent minimum in the late 1980s (Fig. 2.28). There were changes in assigning intensity estimates in the mid-1970s that resulted in tornadoes prior to 1975 being rated more strongly than those in the later part of the record (Verbout et al. 2006). More recently, there have been no tornadoes rated F5, the highest rating, since 3 May 1999, the longest gap on record. Coupled with a large decrease in the number of F4 tornadoes (McCarthy et al. 2006), it has been suggested that the strongest tornadoes are now being rated lower than practice prior to 2000.

A dataset of F2 and stronger tornadoes extending back before the official record (Grazulis 1993) provides an opportunity to examine longer trends. This examination[64] of the record from 1921-1995 indicates that the variability between periods was large, without significant long-term trends (Concannon et al. 2000).

The fraction of strong tornadoes (F2 and greater) that have been rated as violent (F4 and greater) has been relatively consistent in the US from the 1950s through the 1990s[65] (Brooks and Doswell 2001)[66]. There were no significant changes in the high-intensity end of these distributions from 1950s through the 1990s, although the distribution from 2000 and later may differ.

Nontornadic reports have increased even more rapidly than tornadic reports (Doswell et al. 2005, 2006). Over the period 1955-2004, this increase was approximately exponential, resulting in an almost 20-fold increase over the period. The increase is mostly in marginally severe thunderstorm reports (Brooks 2007. An overall increase is seen, but the distribution by intensity is similar in the 1970s and post-2000 eras for the strongest 10% of reports of hail and wind. Thus, there is no evidence for a change in the severity of events, and the large changes in the overall number of reports make it impossible to detect if meteorological changes have occurred.

[64] This analysis used the technique described in Brooks et al. (2003a) to estimate the spatial distribution over different periods
[65] Note that consistent overrating will not change this ratio.
[66] Feuerstein et al. (2005) showed that the distribution in the US and other countries could be fit to Weibull distributions with the parameters in the distribution converging as time goes along, which they associated with more complete reporting of events.

Environmental conditions that are most likely associated with severe and tornadic thunderstorms have been derived from reanalysis data (Brooks et al. 2003b) and counts of the frequency of favorable environments for significant severe thunderstorms[67] have been determined for the area east of the Rocky Mountains in the US for the period 1958-1999 (Brooks and Dotzek 2007). The count of favorable environments decreased from the late 1950s into the early 1970s and increased after that through the 1990s, so that the frequency was approximately the same at both ends of the analyzed period. Given the high values seen at the beginning of the reanalysis era, it is likely that the record is long enough to sample natural variability, so that it is possible that even though the 1973-1999 increase is statistically significant, it does not represent a departure from natural variability. The time series of the count of reports of very large hail (7 cm diameter and larger) shows an inflection at about the same time as the inflection in the counts of favorable environments. A comparison of the rate of increase of the two series suggested that the change in environments could account for approximately 7% of the change in reports from the mid-1970s through 1999, with the rest coming from non-meteorological sources. Changes in tornado reports do not correspond to the changes in overall favorable severe thunderstorm environment, in part because the discrimination of tornadic environments in the reanalysis data is not as good as the discrimination of severe thunderstorm environments (Brooks et al. 2003a).

[67] Hail of at least 5 cm diameter, wind gusts of at least 33 m s^{-1}, and/or a tornado of F2 or greater intensity

BOX 2.3: Changes in Modes of Variability

The atmosphere-ocean system has a wide variety of circulation patterns, or modes, of climate variability that pulse on time scales ranging from days, to many decades, or longer. For example, the well-known winter weather pattern of a storm followed by clear skies and then another storm a week later is, part of an atmospheric wave (wind) pattern that circles the Earth. As these waves move over the ocean, heat from the ocean is given up to the air, which impacts both the intensity and the movement of the atmospheric waves (weather) as well as ocean circulations. Weather and climate extremes are often linked to one or more of these modes of climate variability, and following is a brief description of the most important circulation regimes. However, it is important to keep in mind that these modes of variability are not independent of each other.

El Niño-Southern Oscillation (ENSO)

The ENSO phenomenon is the result of coupled ocean-atmosphere dynamics and is the largest source of interannual variability in global weather and climate. It is characterized by changes in eastern equatorial Pacific sea surface temperature (SST) and surface air pressure in the tropical Pacific region. Warm (cold) eastern Pacific SST anomalies are associated with El Niño (La Niña) events. El Niños occur at irregular intervals of approximately 2 to 7 years, and generally persists for 12 to 18 months. The Southern Oscillation component of ENSO is defined by air pressure differences between the eastern and western tropical Pacific (typically between Darwin and Tahiti) and is characterized by changes in tropical atmospheric flow patterns which are caused by and can enhance tropical Pacific SST variations. These tropical atmospheric circulation changes can alter both the intensity and tracks of North American storms. For example, El Niño is often associated with heavy winter rains in southern California.

The nature of ENSO has varied considerably through time. Strong ENSO events occurred with regularity from the late 19th Century through 1925 and again after 1950. Between 1976 and 1977 rapid warming occurred in the Tropical Pacific with concurrent cooling in the Central Pacific that has been termed the climate shift of 1976/1977 (Trenberth 1990, Miller et al. 1994). The shift has been associated with increased El Niño activity, changes in storm tracks, increased storm intensity and is at the start of the period of rapid warming in global temperatures, and the 1997-1998 El Niño was the strongest on record.

The North Atlantic Oscillation (NAO)

The NAO is the most important mode of winter climate variability in the North Atlantic region and is measured by an index that is based on air pressure differences between Iceland/Greenland and the Azores in the north Atlantic. As Figure 2.29 illustrates, high values of the NAO index are associated with intensified westerly winds around the arctic. Changes in the strength and location of the westerlies produce characteristic shifts in temperature, rainfall, and winds. Low NAO values correspond with cold extremes in central North America and high NAO index values increase the chances of warm winter extremes. Proxy and instrumental data show evidence for intervals of decadal and longer positive and negative NAO index in the last few centuries (Cook et al., 2002; Jones et al., 2003). A reversal occurred from minimum winter index values in the late 1960s to strongly positive NAO index values in the mid-1990s but since have declined to near the long-term mean.

Atlantic Multidecadal Oscillation (AMO)

The Atlantic Ocean meridional overturning circulation carries warm salty surface waters into far-northern latitudes around Greenland where it cools, sinks, and returns cold deep waters southward across the equator. An oscillating pattern of SSTs in the northern Atlantic that is related to this overturning circulation, called the Atlantic Multidecadal Oscillation (AMO), has been identified by a number of researchers (Delworth and Mann, 2000; Folland *et al*., 1986; Mann and Park, 1994). The AMO is commonly identified by subtracting a linear trend from a time series of the North Atlantic SST. This trend subtraction is intended to remove, or at least reduce, the influence of greenhouse-gas induced global warming from the AMO so that the bulk of the variability in the remainder is due to natural causes. The warm phase, the decades when the temperature is above the trend line, is associated with increased Atlantic hurricane activity, and the cool phase is associated with reduced Atlantic hurricane activity. Instrumental data has been used to identify warm phases roughly between 1860-1880, 1930-1960, and one beginning in the mid-1990s which continues to present. Cool phases were present during 1905-1925 and 1970-1990 (Schlesinger and Ramankutty, 1994).

Some scientists, however, question the validity of subtracting a linear trend from a time series created by non-linear forcings and wonder if the AMO as commonly calculated is primarily an artifact of this creation process rather than a real change in the ocean circulation. Some suggest that subtracting the global SST time series from the North Atlantic SST time series removes a global climate change signal better than subtracting a linear trend and produces a very different historical AMO record (Trenberth and Shea, 2006). Proxy and modeling studies have identified an AMO-like signal and found that multidecadal eras in hurricane activity in the North Atlantic are correlated with the AMO (Bell and Chelliah, 2006). No matter how it is calculated, the AMO has such a long period that the observational SST data only records about 1.5 cycles which makes it difficult to determine whether the AMO is truly a natural oscillation or caused in whole or at least in part by greenhouse-gas induced climate change.

Pacific Decadal Oscillation (PDO)

The Pacific Decadal Oscillation (PDO) is a multidecadal pattern of monthly SST anomalies in the North Pacific Ocean poleward of 20°N. Two full PDO cycles occurred through the twentieth century with each phase persisting for 20 to 30 years. The typical spatial pattern of the “warm” phase of the PDO has negative SST anomalies in the central and eastern North Pacific and positive SST anomalies along the coast of North America. Sea level pressure (SLP) anomalies during the warm phase tend to have a basin-scale low centered over the Aleutian Islands and high sea-level pressure over western North America. The cool phase of the PDO has SST and SLP patterns that are essentially the opposite of the warm phase. Because the PDO influences various weather systems it can affect the chances of, for example, winter temperatures cold enough to cause mountain pine beetle mortality in British Columbia (Stahl *et al*., 2006). When an El Niño event occurs during a warm phase of the PDO, the characteristic El Niño-related temperature and precipitation anomalies in North America tend to be accentuated. The PDO had extended periods of negative values indicative of weakened circulation from 1900 to 1924 and 1947 to 1976, and positive values indicative of strengthened circulation from 1925 to 1946 and 1977 to 2005. The 1976-1977 climate shift in the Pacific described above was associated with a phase change in the PDO from negative to positive (Trenberth et al., 2002; Deser et al., 2004).

Pacific North American Pattern (PNA)

The PNA can be defined as a secondary pattern in the variability of monthly atmospheric pressure anomalies for the latitude range 20-90°N. When the PNA is positive, the mid-tropospheric winds over North America and the North Pacific have a strong meridional (north-south) wave pattern while the negative PNA has more zonal (west to east) flow. Strong wave patterns tend to bring extreme weather; whether the extremes are warm, cold, wet or dry at a particular location depends on the shape of the wave. A positive PNA is associated with El Niños and negative PNA is associated with La Niña.

The Madden-Julian Oscillation (MJO)

The atmospheric response to convection on the equator, which heats the atmosphere, is the creation of circulation cells, which then move eastward. These cells have a period of about 50 days and either enhance tropical convection or help suppress it. Referred to as the Madden-Julian Oscillation (MJO), after the two scientists who discovered it (Madden and Julian, 1971 and 1972), it is the dominant source of tropical atmospheric variability on intraseasonal time scales. The MJO is related to North American extremes through its influence on the dynamics of tropical cyclone formation (Hartmann and Maloney, 2001; Maloney and Hartmann, 2000a; 2000b, 2001) as well as western North American winter rainfall variability. The MJO can enhance or suppress either depending on which part of the circulation cell is active in the region.

As the climate changes, some of the atmospheric circulation patterns or modes of atmospheric variability described above have changed as well. However, only one circulation pattern, the MJO, would not be expected to have long-term changes since it is a localized circulation response to convection on the equator.

2.3 Key Uncertainties Related to Measuring Specific Variations and Change

In this section we review the statistical methods that have been used to assess uncertainties in studies of changing extremes. The focus of the discussion is on precipitation events, though similar methods have also been used for temperature.

2.3.1 Methods Based on Counting Exceedances Over a High Threshold

Most existing methods follow some variant of the following procedure, given by Kunkel et al. (1999). First, daily data are collected, corrected for biases such as winter undercatchment. Only stations with nearly complete data are used (typically, “nearly complete” means no more than 5% missing values). Different event durations (for example, 1-day or 7-day) and different return periods (such as 1 year or 5 years) are considered. For each station, a threshold is determined according to the desired return value – for example, with 100 years of data and a 5-year return value, the threshold is the

20th largest event. The number of exceedances of the threshold is computed for each year, and then averaged either regionally or nationally. The averaging is a weighted average in which, first, simple averaging is used over climate divisions (typically there are about 7 climate divisions in each state), and then, an area-weighted average is computed over climate divisions, either for one of the nine U.S. climate regions or the whole contiguous U.S. This averaging method ensures that parts of the country with relatively sparse data coverage are adequately represented in the final average. Sometimes (e.g. Groisman et al. 2005, Kunkel et al. 2007a) the climate divisons are replaced by 1^o by 1^o grid cells. Two additional refinements used by Groisman et al. (2005) are (i) to replace the raw exceedance counts for each year by anomalies from a 30-year reference period, computed separately for each station, (ii) to assess the standard error of the regional average using spatial statistics techniques. This calculation is based on an exponentially decreasing spatial covariance function with a range of the order 100-500 km. and a nugget:sill ratio (the proportion of the variability that is not spatially correlated) between 0 and 85%, depending on the region, season and threshold.

Once these spatially averaged annual exceedance counts or anomalies are computed, the next step is to compute trends. In most studies, the emphasis is on linear trends computed either by least squares regression or by the Kendall slope method, in which the trend is estimated as the median of all possible slopes computed from pairs of data points. The standard errors of the trends should theoretically be corrected for autocorrelation, but in the case of extreme events the autocorrelation is usually negligible (Groisman et al., 2004).

One of the concerns about this methodology is the effect of changing spatial coverage of the data set, especially for comparisons that go back to the late years of the 19th century. Kunkel et al. (2007a) generated simulations of the 1895-2004 data record by first randomly sampling complete years of data from a modern network of 6351 stations for 1971-2000, projecting to a random subnetwork equivalent in size and spatial extent to the historical data network, then using repeat simulations to calculate means and 95% confidence intervals for five 22-year periods. The confidence intervals were then superimposed on the actual 22-year means calculated from the observational data record. The results for 1-year, 5-year and 20-year return values show clearly that the most recent period (1983-2004) has the highest return values of the five periods, but they also show the second highest return values in 1895-1916 with a sharp drop thereafter, implying a still not fully explained role due to natural variability.

Some issues that might justify further research include the following:

1. Further exploration of why extreme precipitation apparently decreases after the 1895-1916 period before the recent (post-1983) rise when they exceeded that level. For example, if one breaks the data down into finer resolution spatially, does one still see the same effect?
2. What about the effect of large-scale circulation effects such as ENSO events, AMO, PDO, etc? These could potentially be included as covariates in a time series regression analysis, thus allowing one to “correct” for circulation effects in measuring the trend.

3. The spatial analyses of Groisman et al. (2005) allow for spatial correlation in assessing the significance of trends, but they don't do the logical next step, which is to use the covariance function to construct optimal interpolations (also known as kriging) and thereby produce more detailed spatial maps. This is something that might be explored in the future.

2.3.2 The GEV Approach

An alternative approach to extreme value assessment is though the Generalized Extreme Value (GEV) distribution[68], and its variants. The GEV combines together three "types" of extreme value distributions that in earlier treatments were often regarded as separate families (e.g. Gumbel 1958). The distribution is most frequently applied to the annual maxima of a meteorological or hydrological variable, though it can also be applied to maxima over other time periods (e.g. one month or one season). With minor changes in notation, the distributions are also applicable to minima rather than maxima. The parameters may be estimated by maximum likelihood, though there are also a number of more specialized techniques such as L-moments estimation. The methods have been applied in climate researchers by a number of authors including Kharin and Zwiers (2000), Wehner (2004,2005), Kharin et al. (2007).

[68] The basic GEV distribution is given by the formula (see, e.g. Zwiers and Kharin (1998)) $F(x) = exp\{-[1-k(x-\xi)/\alpha]^{1/k}\}$ in which ξ plays the role of a centering or location constant, α determines the scale, and k is a key parameter that determines the shape of the distribution. (Other authors have used different notations, especially for the shape parameter.) The range of the distribution is $x<\xi+\alpha/k$ when $k<0$, $x>\xi+\alpha/k$ when $k<0$, $-\infty<x<\infty$ when $k=0$, in which case the formula reduces to $F(x) = exp\{-exp[-(x-\xi)/\alpha]\}$ and is known as the Gumbel distribution.

The potential advantage of GEV methods over those based on counting threshold exceedances is that by fitting a probability distribution to the extremes, one obtains more information that is less sensitive to the choice of threshold, and can also derive other quantities such as the T-year return value X_T, calculated by solving the equation $F(X_T)=1-1/T$. Trends in the T-year return value (for typical values of T, e.g. 1, 10, 25 or 100 years) would be particularly valuable as indicators of changing extremes in the climate. Direct application of GEV methods is often inefficient because they only use very sparse summaries of the data (typically one value per year), and need reasonably long time series before they are applicable at all. Alternative methods are based on exceedances over thresholds, not just counting exceedances but also fitting a distribution to the excess over the threshold. The most common choice of distribution of excess is the Generalized Pareto distribution or GPD, which is closely related to the GEV (Pickands 1975, Davison and Smith 1990). Some recent overviews of extreme value distributions, threshold methods, and a variety of extensions are by Coles (2001) and Smith (2003).

Much of the recent research (e.g. Wehner 2005, Kharin et al. 2007) has used model output data, using the GEV to estimate for example a 20-year return value at each grid cell, then plotting spatial maps of the resulting estimates. Corresponding maps based on observational data must take into account the irregular spatial distribution of weather stations, but this is also possible using spatial statistics (or “kriging”) methodology. For example, Cooley et al. (2007) have applied a hierarchical modeling approach to precipitation data from the Front Range of Colorado, fitting a GPD to threshold exceedances at each station and combining results from different stations through a spatial model to compute a map of 25-year return values. Smith et al. (2007) applied similar methodology to data from the whole contiguous U.S., producing spatial maps of return values and also calculating changes in return values over the 1970-1999 period.

Chapter 2 References

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Table 2.1 Regressions for the decadal trends of increasing wave heights measured off the Washington coast (NDBC buoy #46005). [after Allan and Komar (2006)]

Wave Heights	Rate (m/yr)	Ratio of Rate to Annual Average	Statistical Significance*
Annual Average	0.024	1.0	SS
Winter Average	0.032	1.3	SS
Five Largest	0.095	4.0	SS
Three Largest	0.103	4.3	NSS
Maximum	0.108	4.5	NSS
100-yr Projection	≈0.13	≈5.5	estimate

SS = statistically significant at the 0.05 level; NSS = not statistically significant.

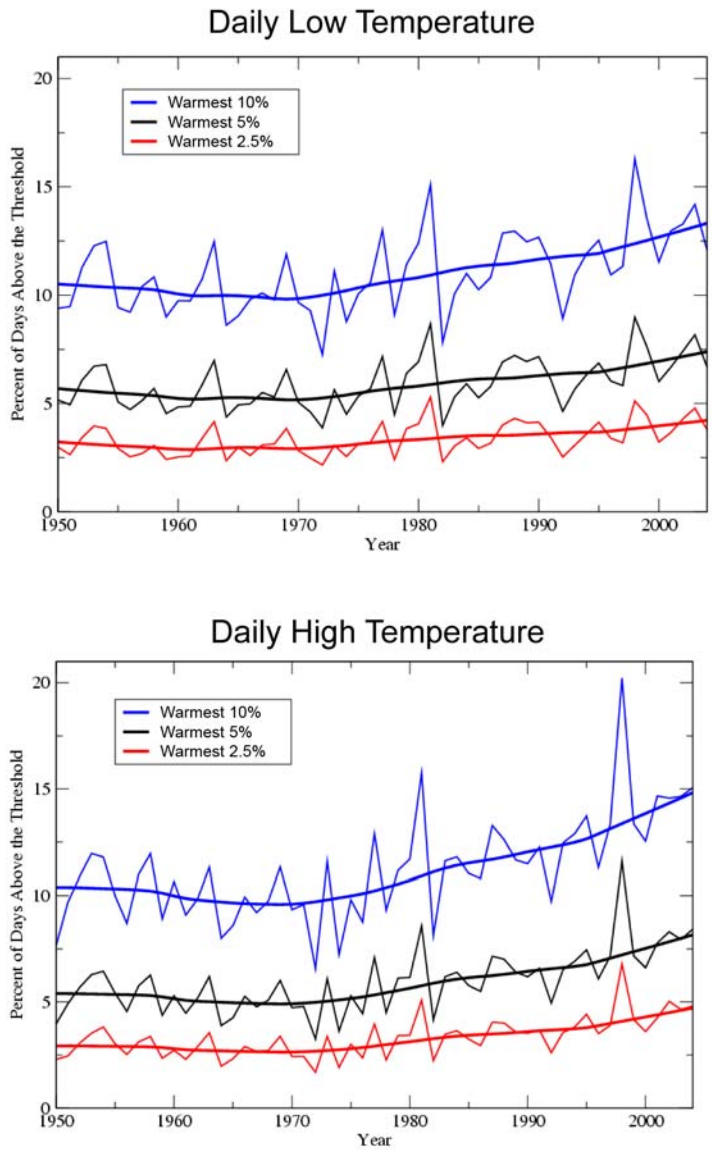

Figure 2.1 Changes in the percent of days in a year above three thresholds for North America for daily high temperature (top) and daily low temperature (bottom) from Peterson et al. (2007).

Trends in Number of Days With Unusually Warm Daily Low Temperature

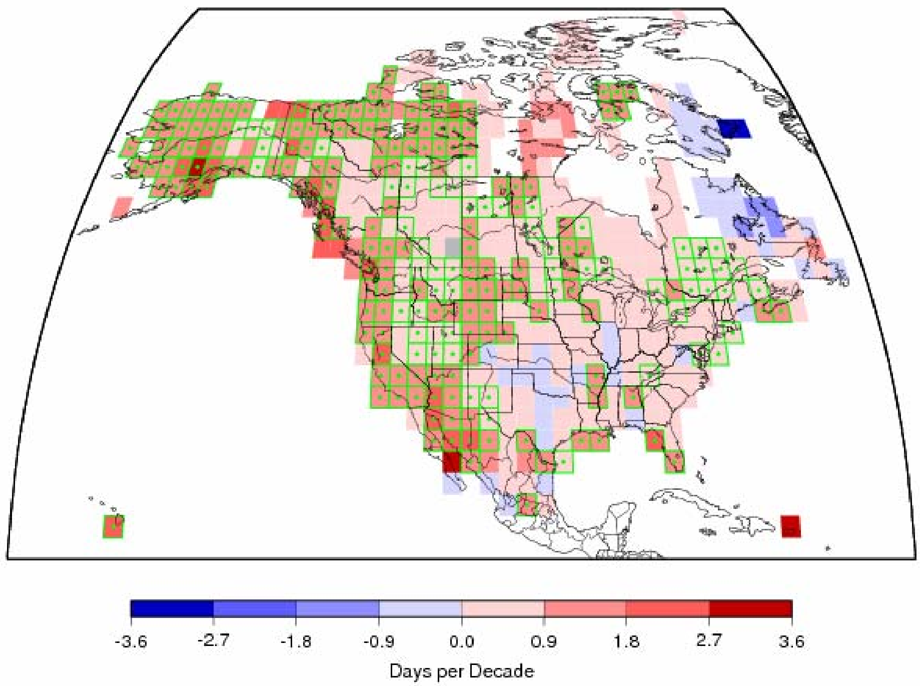

Figure 2.2 Trends in the number of days in a year when the daily low is unusually warm (ie. In the top 10% of warm nights for the 1950-2004 period). Grid boxes with green squares are statistically significant at the p=0.05 level, (from Peterson et al. 2007). A trend of 1.8 days/decade translates to a trend of 9.9 days over the entire 55-year (1950-2004) period, meaning that 10 days more a year will have unusually warm nights.

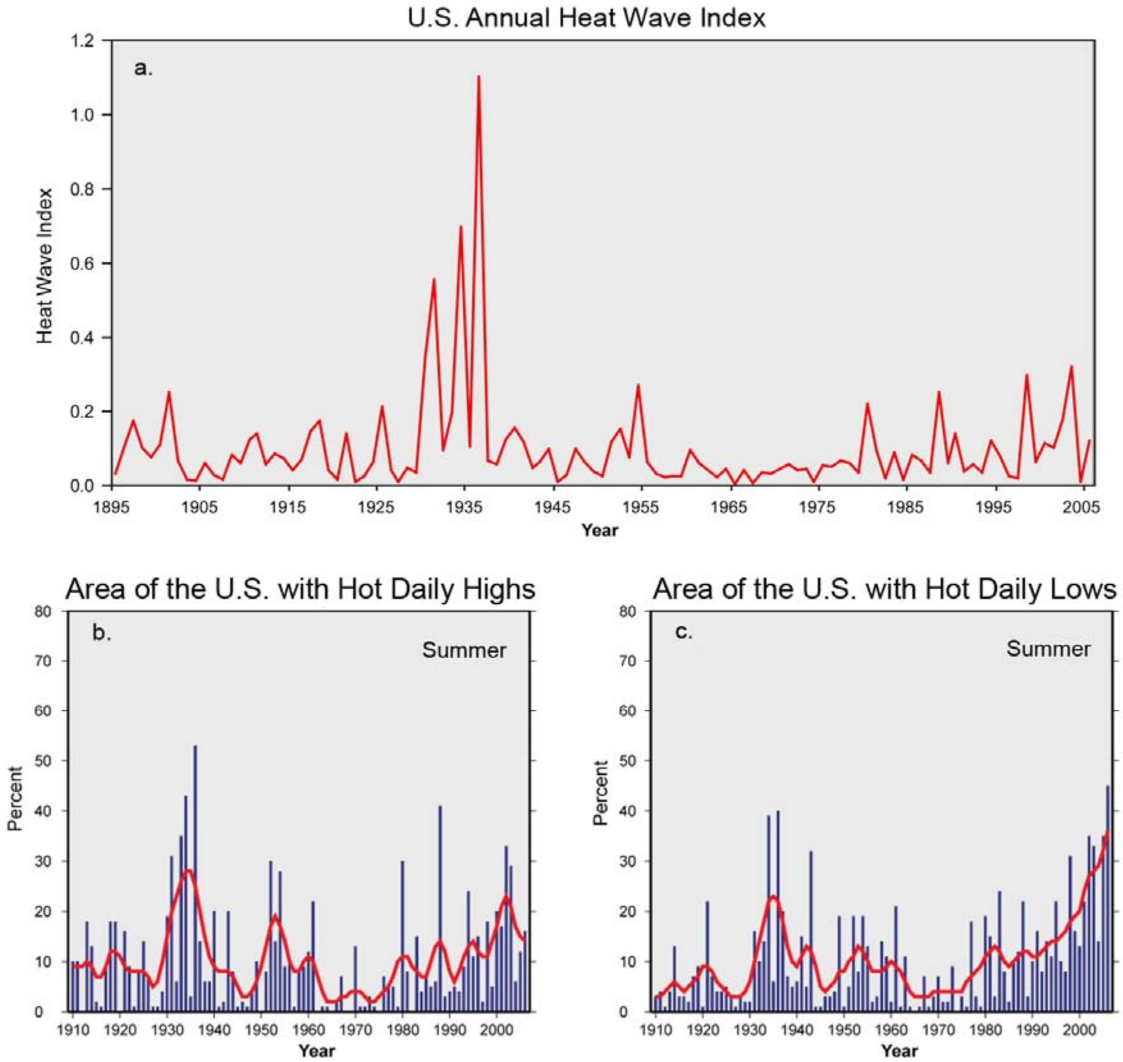

Figure 2.3 Time series of (a) annual values of a U.S. national average "heat wave" index. Heat waves are defined as warm spells of 4 days in duration with mean temperature exceeding the threshold for a 1 in 10 year event. (updated from Kunkel et al. 1999); (b)Area of the U.S. (in percent) with much above normal daily high temperatures in summer; (c) Area of the U.S. (in percent) with much above normal daily low temperatures in summer. Blue vertical bars give values for individual seasons while red lines are smoothed (9-yr running) averages.

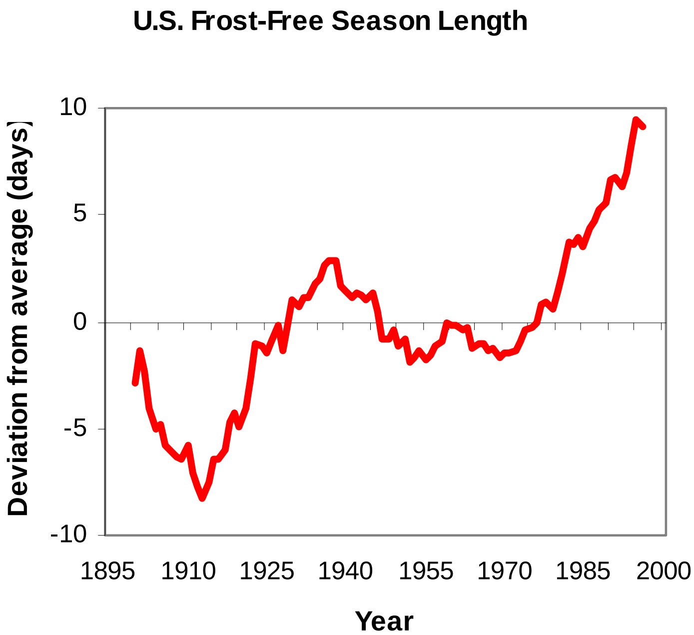

Figure 2.4 Change in the length of the frost free season averaged over the U.S. (from Kunkel et al. 2003). The frost-free season is at least 10 days longer on average than the long-term average.

Changes in the Daily Range of Temperature for Mexico

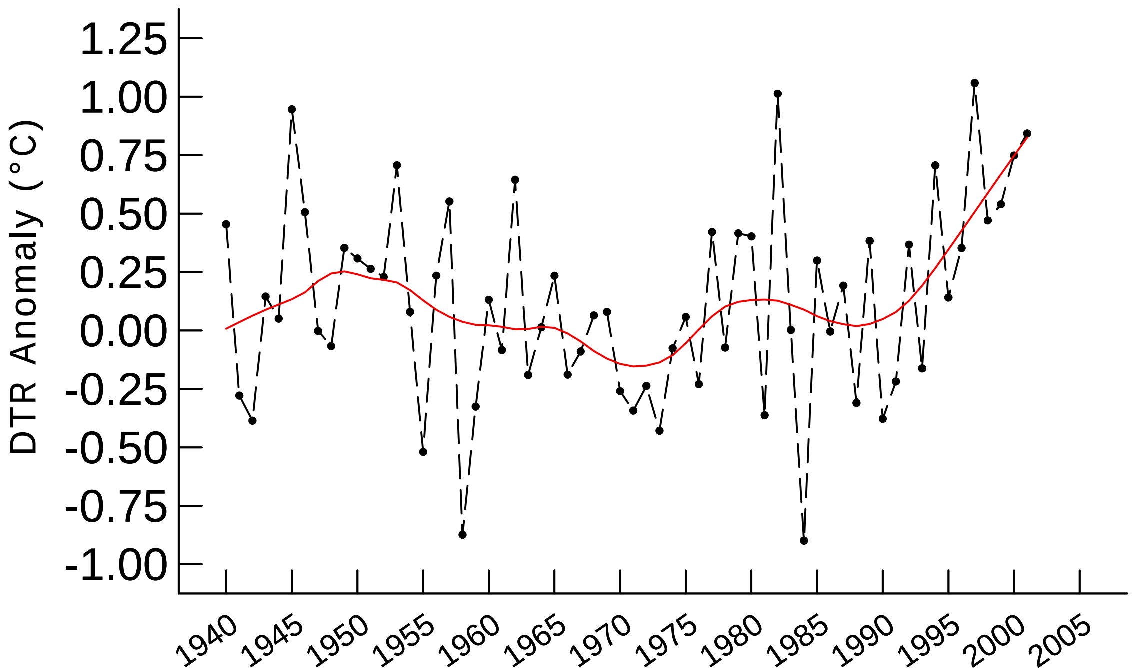

Figure 2.5 Change in the daily range of temperature (difference between the daily low and the daily high temperature) during the warm Season (June-Sept) for Mexico. This difference is known as a Diurnal Temperature Range (DTR). The recent rise in the daily temperature range reflects hotter daily summer highs. The time series represents the average DTR taken over the four temperature regions of Mexico as defined in Englehart and Douglas, 2004. Trend line (red) based on LOWESS smoothing (*n*=30).

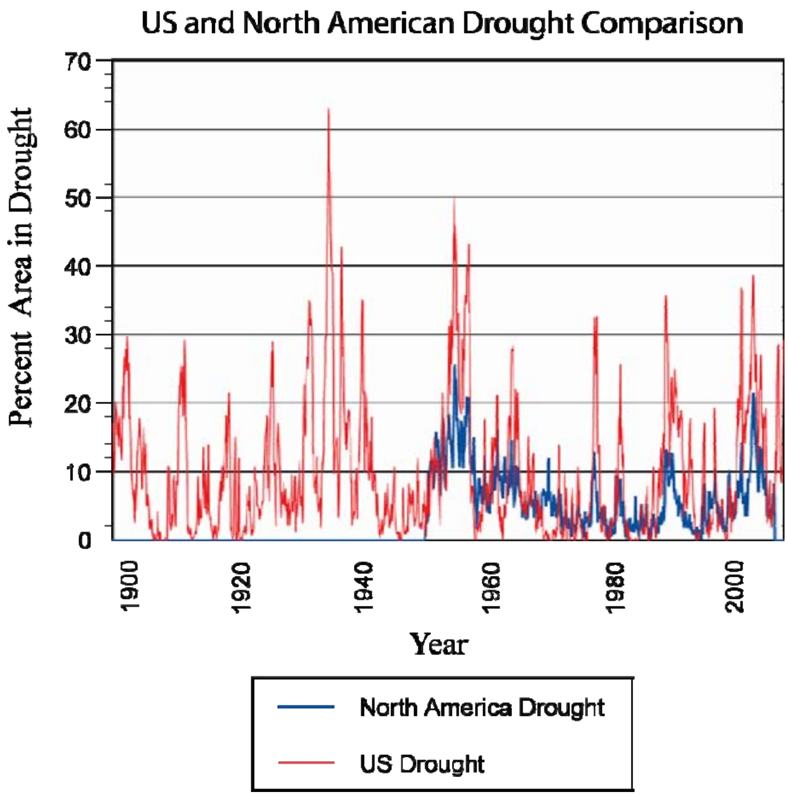

Figure 2.6 the area (in percent) of area in severe to extreme drought as measured by the Palmer Drought Severity Index for the U.S. (red) from 1900 to present and for North America (blue) from 1950 to present.

Western U.S. Drought Area for the last 1200 years

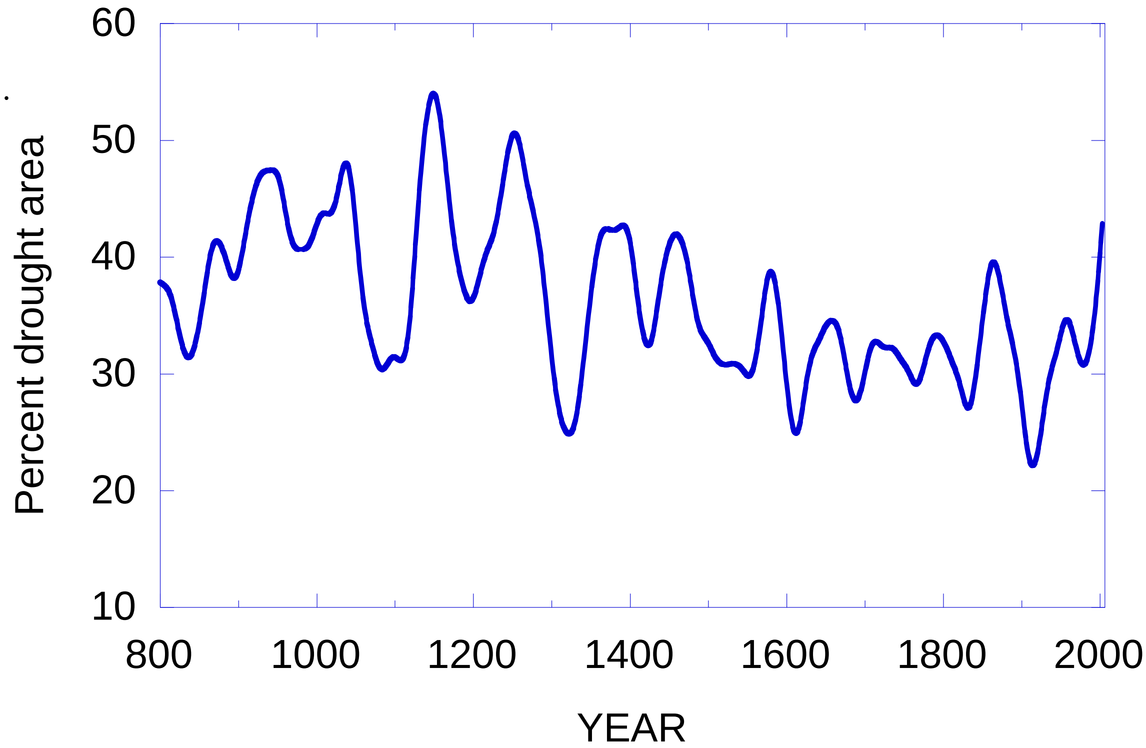

Figure 2.7 Area of drought in the western U.S. as reconstructed from tree rings (Cook et al. 2004).

Regions of N. America where Heavy and Very Heavy Precipitation has Increased

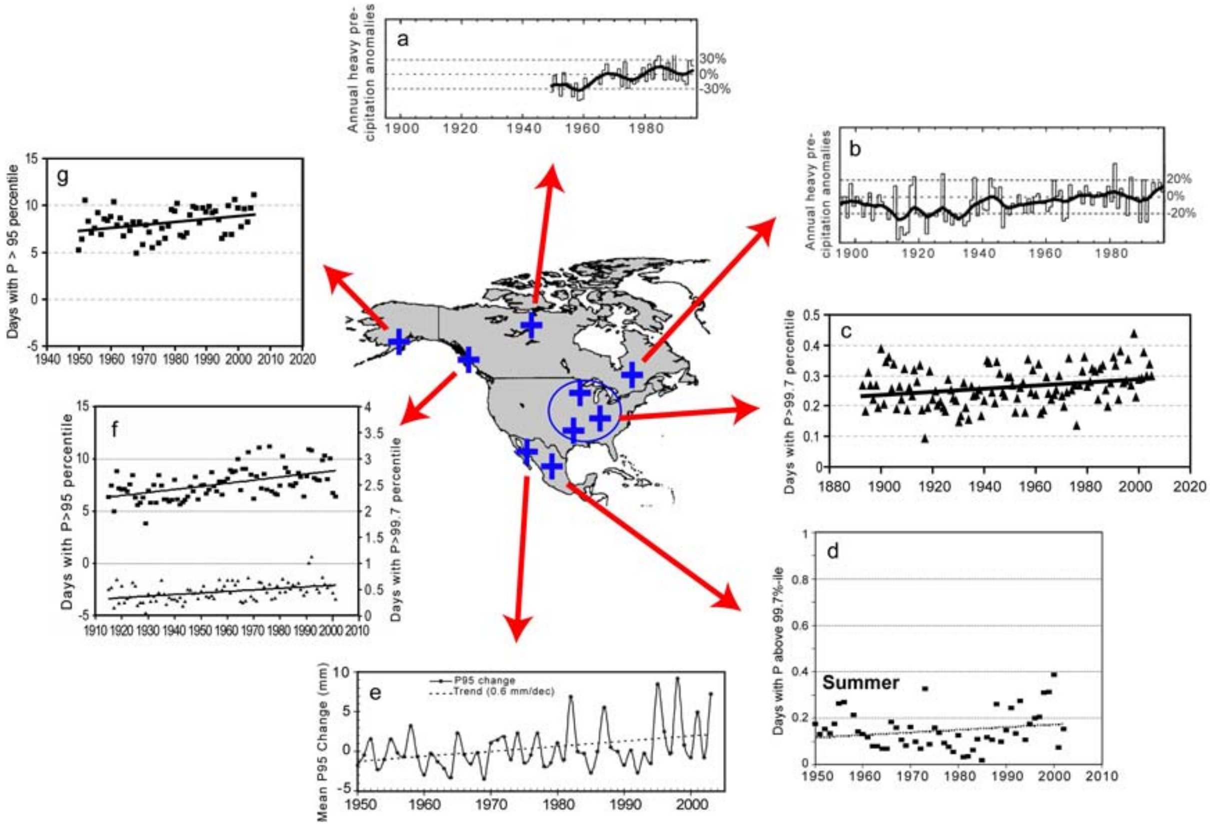

Figure 2.8 Regions where disproportionate increases in heavy and very heavy precipitation during the past decades were documented compared to the change in the annual and/or seasonal precipitation. Because these results come from different studies, the definitions of extreme precipitation vary. (a) annual anomalies (% departures) of heavy precipitation for northern Canada. (b) as (a), but for southeastern Canada. (c) the top 0.3% of daily rain events over the central United States and the trend (22%/113 yrs) (updated from Groisman et al. 2005). (d) as for (c), but for southern Mexico. (e) change of intensity of the upper 5% of daily rain events in the core monsoon region of Mexico, relative to the 1961-1990 base period. (Cavazos et al., 2007) (f) upper 5%, top points, and upper 0.3%, bottom points, of daily precipitation events and linear trends for British Columbia south of 55°N. (g) upper 5% of daily precipitation events and linear trend for Alaska south of 62°N.

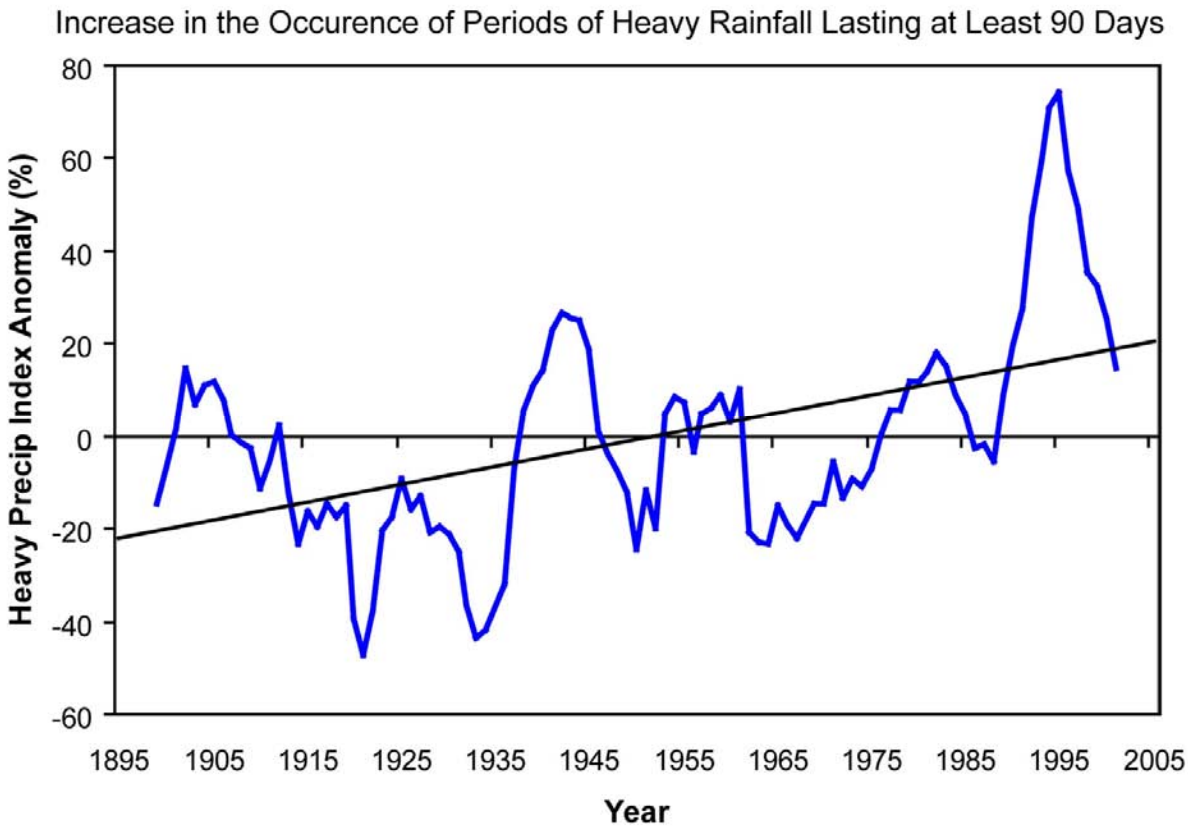

Figure 2.9 Frequency (expressed as a percentage anomaly from the period of record average) of excessive precipitation periods of 90 day duration exceeding a 1-in-20-year event threshold for the U.S. Annual values have been smoothed with a 9-yr running average filter. The black line shows the trend (a linear fit) for the annual values.

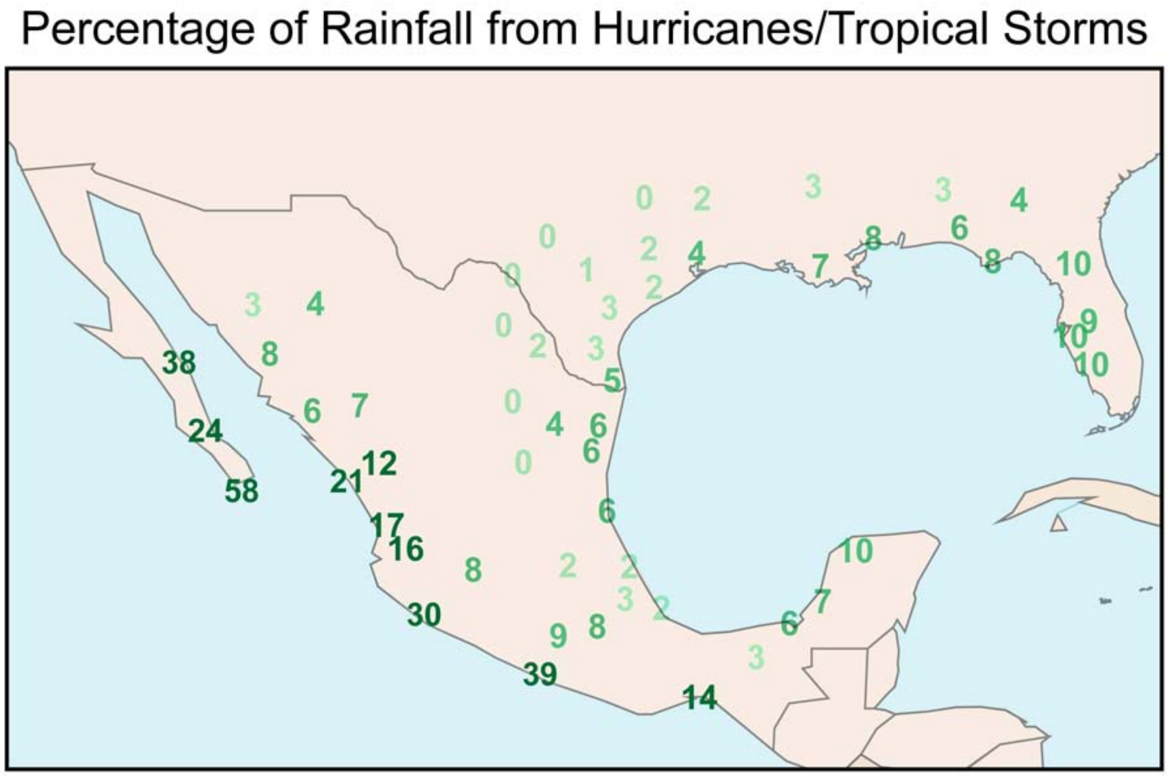

Figure 2.10 Average (median) percentage of warm season rainfall (May-November) from Hurricanes and tropical storms affecting Mexico and the Gulf Coast of the United States. Figure updated from Englehart and Douglas 2001.

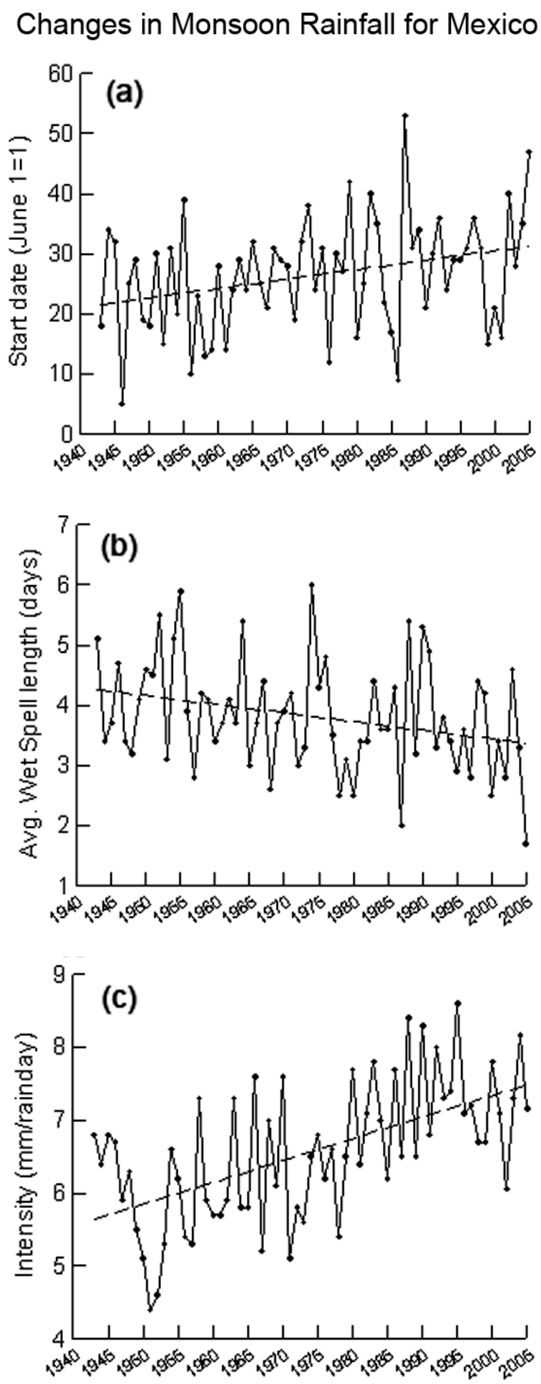

Figure 2.11 Variations and linear trend in various characteristics of the summer monsoon in southern Sonora, Mexico, including (a) the mean start date June 1 = Day 1 on the graph; (b) the mean wet spell length defined as the mean number of consecutive days with mean regional precipitation $\geq$1mm; and (c) the mean daily rainfall intensity for wet days defined as the regional average rainfall for all days with rainfall $\geq$ 1mm.

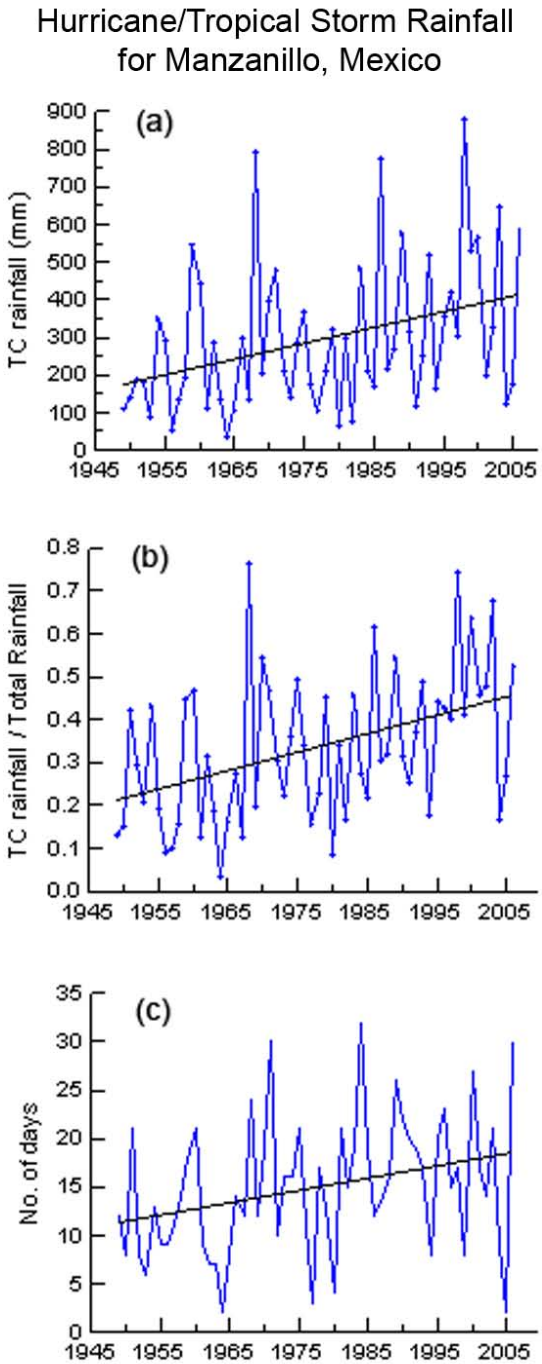

Figure 2.12 Trends in hurricane/tropical storm rainfall statistics at Manzanillo, Mexico, including (a) the total warm season rainfall from hurricanes/tropical storms; (b) the ratio of hurricane/tropical storm rainfall to total summer rainfall; and (c) the number of days each summer with a hurricane or tropical storm within 550km of the stations

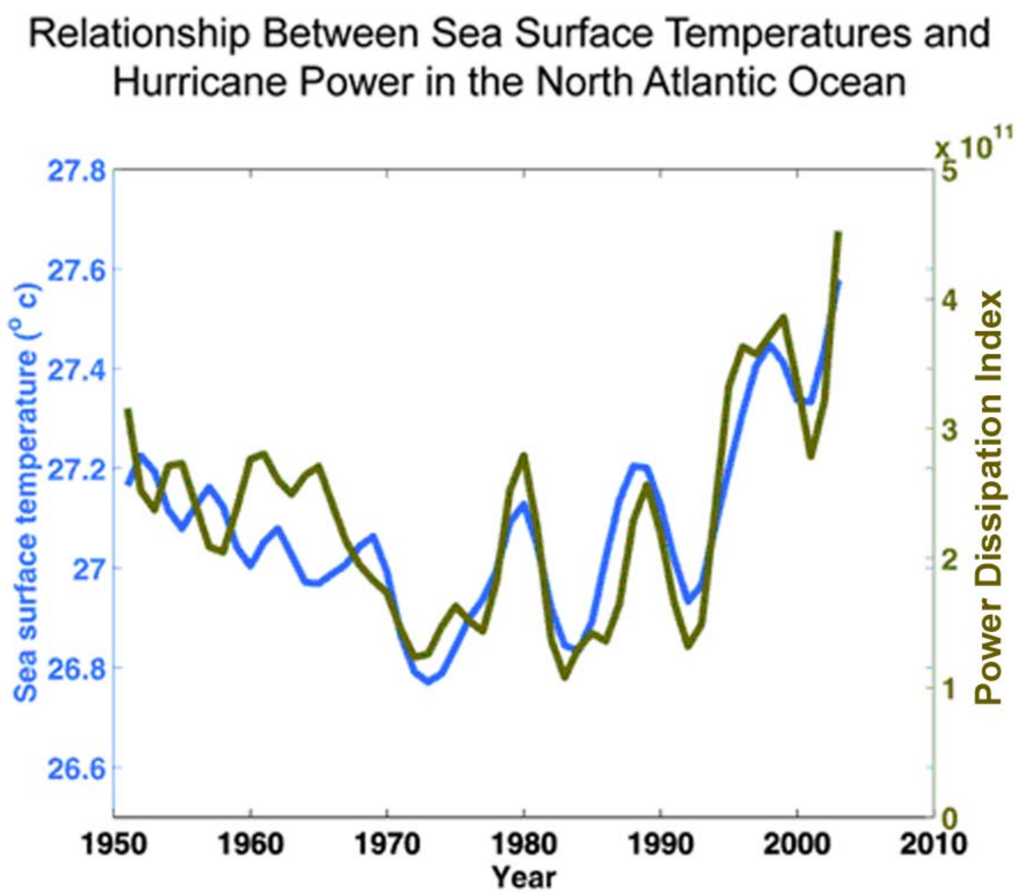

Figure 2.13 Sea surface temperatures (blue) correlated with the Power Dissipation Index for North Atlantic hurricanes (Emanuel, 2007). Sea Surface Temperature is from the Hadley Centre dataset and is for the Main Development Region for tropical cyclones in the Atlantic, defined as 6-18°N, 20-60°W. The time series have been smoothed using a 1-3-4-3-1 filter to reduce the effect of interannual variability and highlight fluctuations on time scales of 3 years and longer.

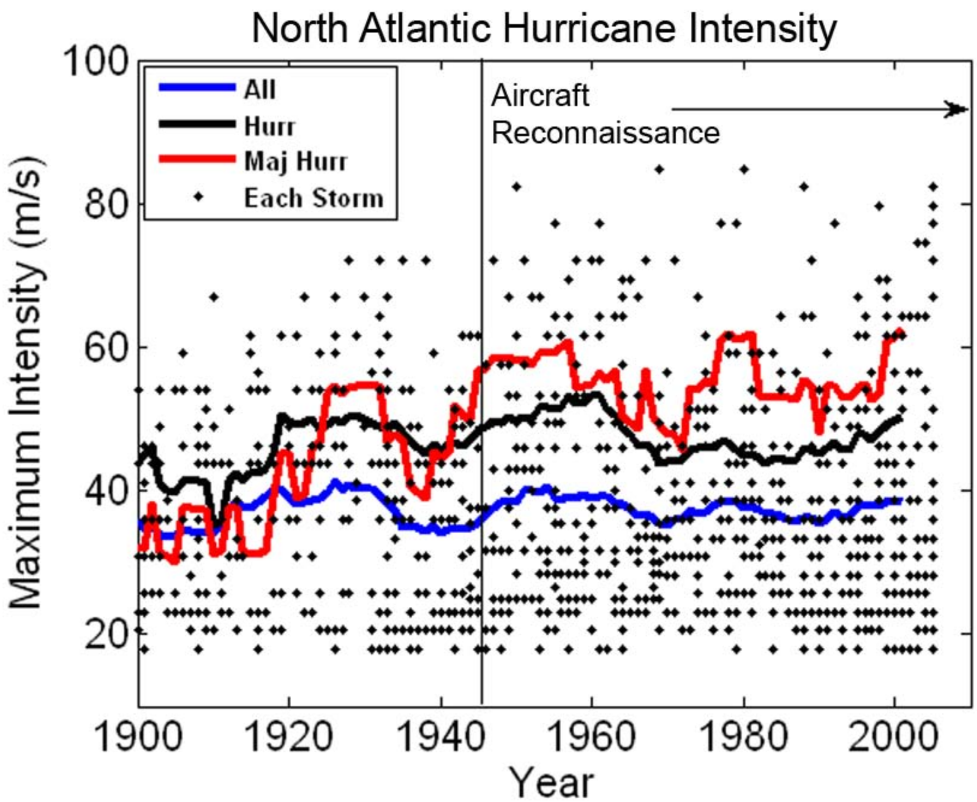

Figure 2.14 Century changes in the intensity of North Atlantic tropical cyclones, hurricanes and major hurricanes. Also shown are all individual tropical cyclone intensities. (From Holland and Webster 2007).

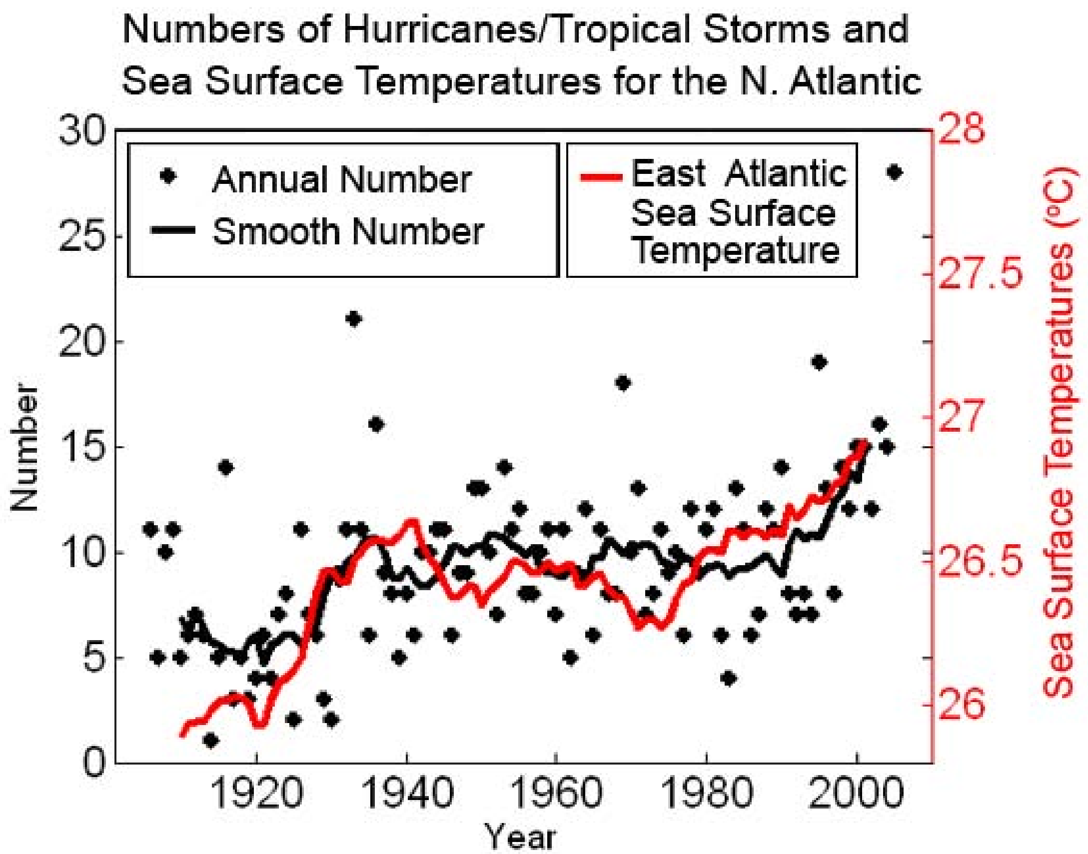

Figure 2.15 Combined annual numbers of hurricanes and tropical storms for the North Atlantic (black dots), together with a 9-year running mean filter (black line) and the 9-year smoothed sea surface temperature in the eastern North Atlantic (red line). Adapted from Holland and Webster (2007).

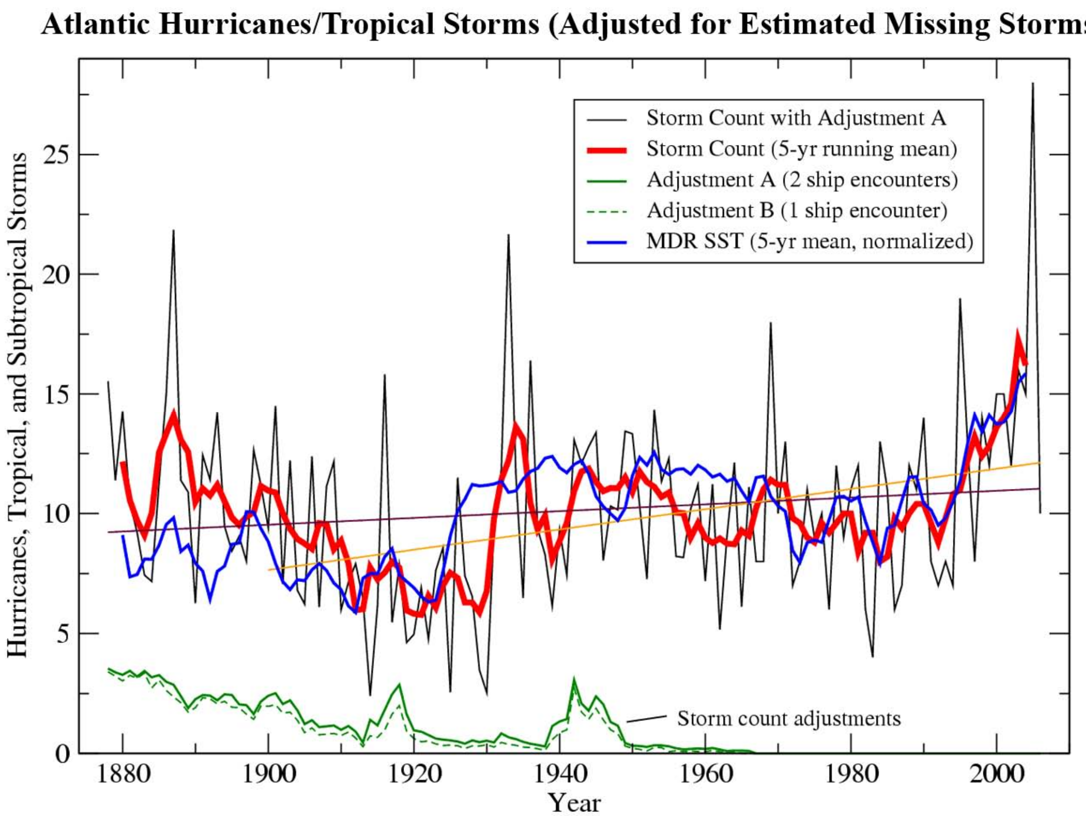

Figure 2.16 Atlantic hurricanes and tropical storms for 1878-2006, using the adjustment method A for missing storms described in the text. Black curve is the adjusted annual storm count, red is the 5-yr running mean, and solid blue curve is a normalized (same mean and variance) 5-yr running mean sea surface temperature index for the Main Development Region of the tropical Atlantic (HadISST, 80-20W, 10-20N; Aug.-Oct.). Green curves show the adjustment that has been added for missing storms to obtain the black curve, assuming two simulated ship-storm "encounters" are required for a modern-day storm to be "detected" by a historical ship traffic for a given year. Dashed green curve is an alternative adjustment sensitivity test requiring just one ship-storm simulated encounter for detection. Straight lines are least squares trend lines for the adjusted storm counts. (Adapted from Vecchi and Knutson, 2007).

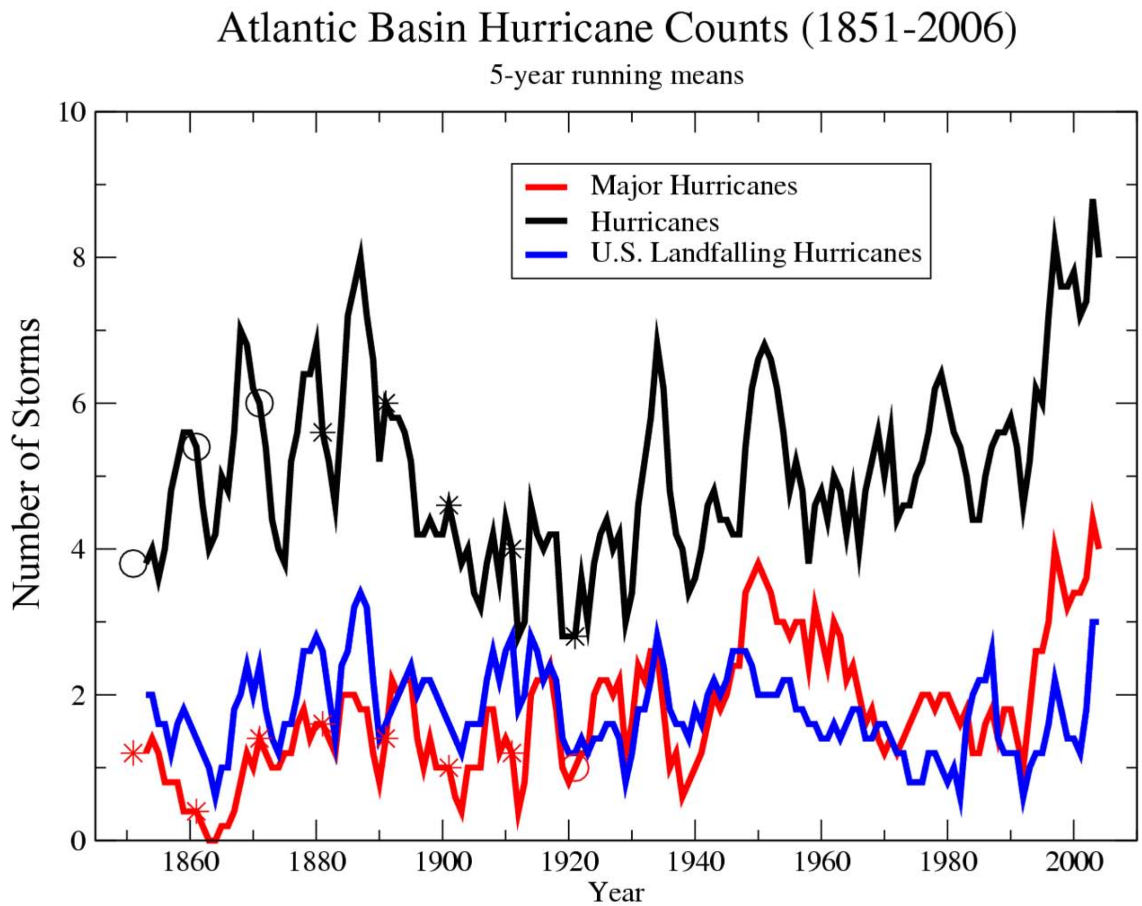

Figure 2.17 Counts of total North Atlantic basin hurricanes (black), major hurricanes (red) and U.S. landfalling hurricanes (blue) based on annual data from 1851 to 2006 and smoothed (using a 5-year running mean). Asterisks on the time series indicate years where trends beginning in that year and extending through 2005 are statistically significant (p=0.05) based on annual data; circles indicate non-significant trend results.

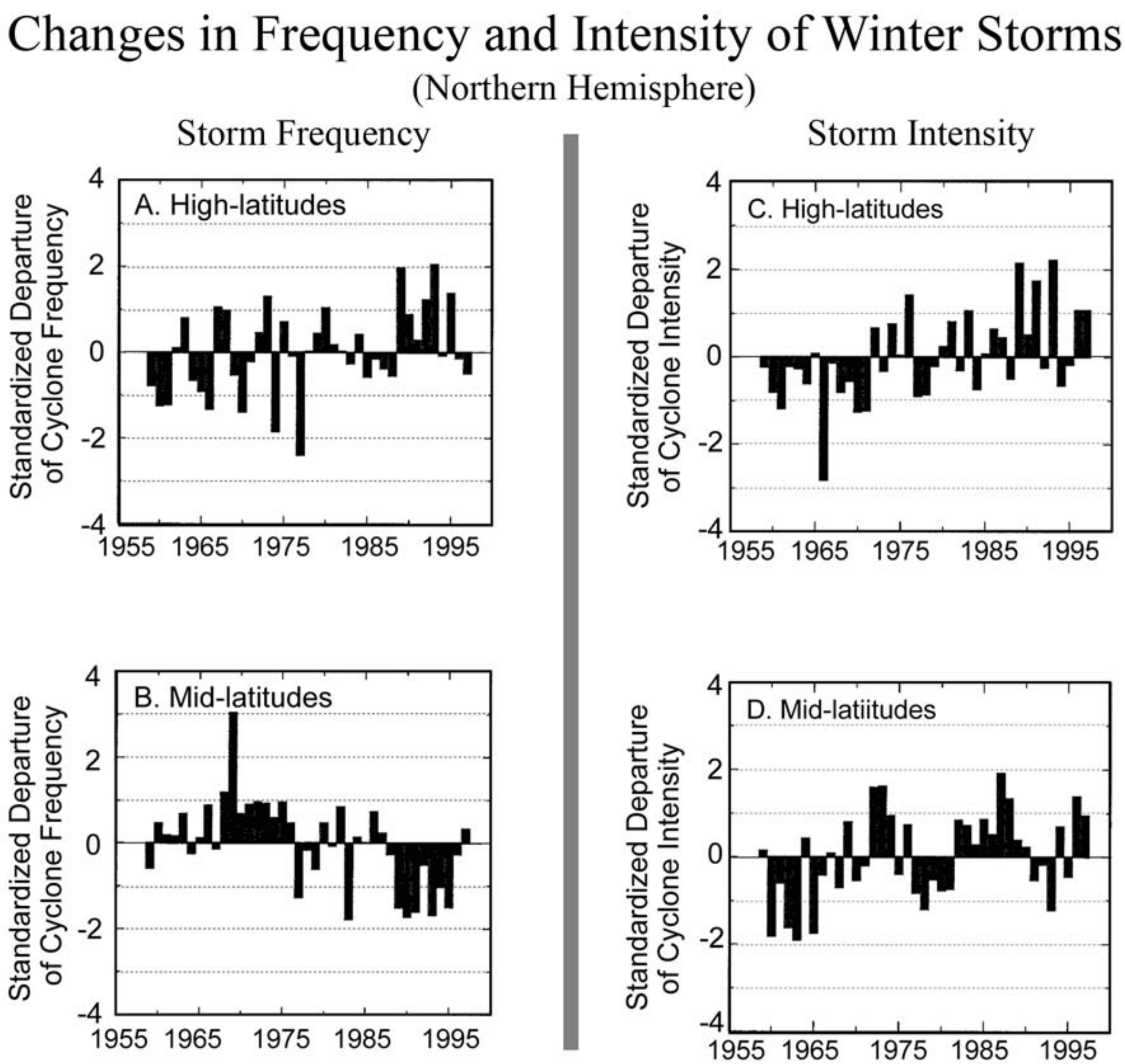

Figure 2.18 Changes from average (1959-1997) in the number of winter (Nov-Mar) storms each year in the Northern Hemisphere for (a) high latitudes (60^{o}-90^{o}N), and (b) mid-latitudes (30^{o}-60^{o}N), and the change from average of winter storm intensity in the Northern Hemisphere each year for (c) high latitudes (60^{o}-90^{o}N), and (d) mid-latitudes (30^{o}-60^{o}N). [Adapted from McCabe et al. 2001].

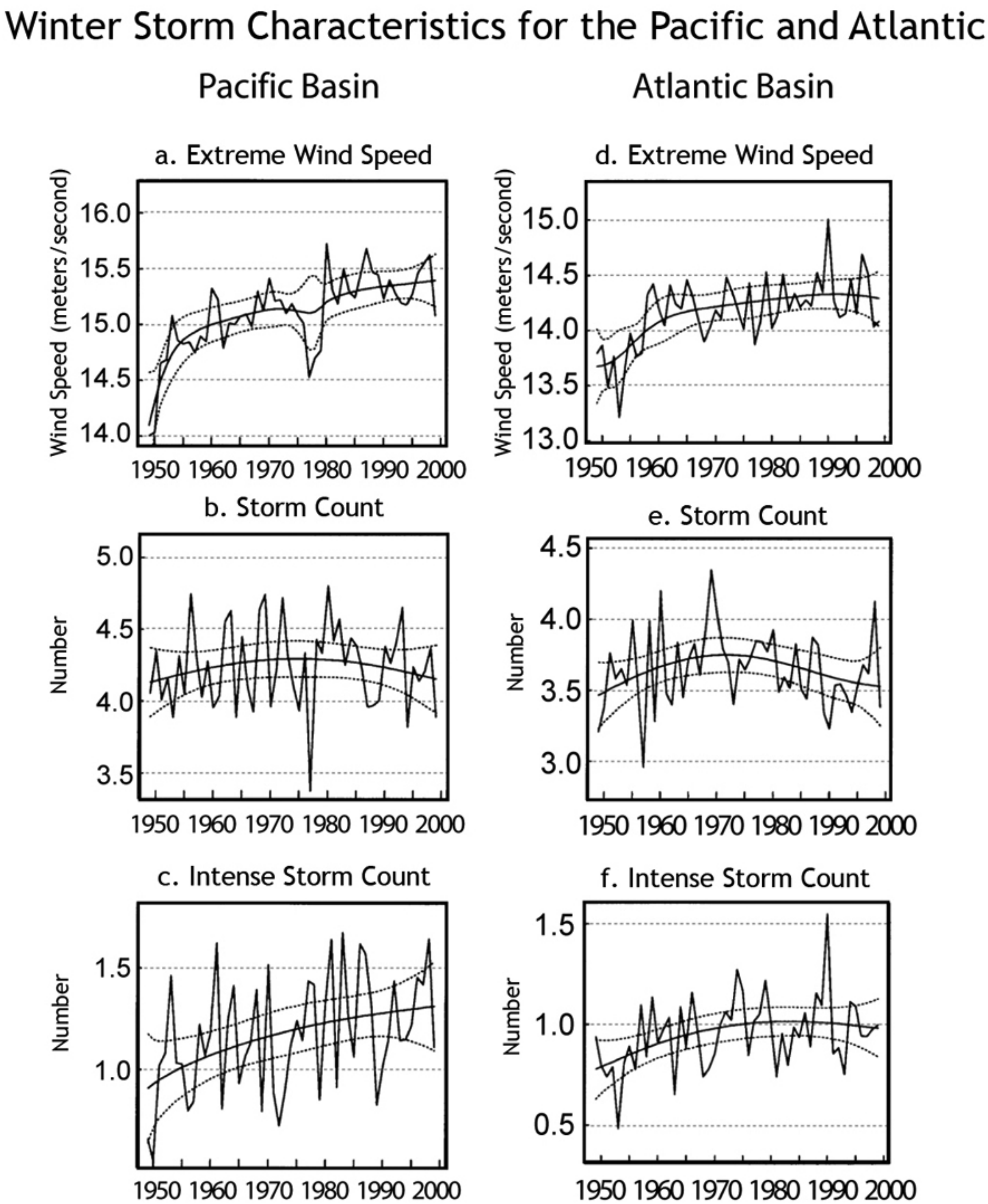

Fig. 2.19 Extreme wind speed (meters per second), number of winter storms, and number of intense (≤980 hPa) winter storms for the Pacific region (20^{o}-70^{o}N, 130^{o}E-112.5^{o}W; panels a-b-c) and the Atlantic region (20^{o}-70^{o}N, 7.5^{o}E-110^{o}W; panels d-e-f):. The thick smooth lines are the trends determined using a Bayesian spline model, and the thin dashed lines denote the 95% confidence intervals. [Adapted from Paciorek et al. 2002].

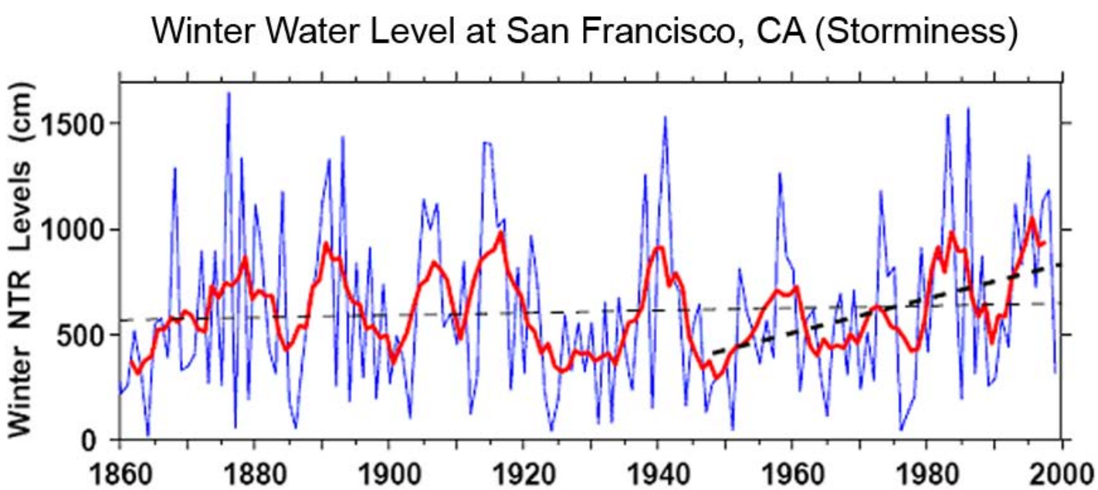

Figure 2.20 Cumulative extreme Non-Tide Residuals (NTR) (water level) exceeding the 98th percentile level of hourly NTR levels at San Francisco, during winter months (Dec-Mar), with the 5-yr running mean (red line). Least squares trend estimates for the entire winter record (light dashed line) and since 1948 (bold dashed line), the period covered by NCEP reanalysis and ERA-40 data used in most ETC studies. [Adapted from Bromirski et al. 2003].

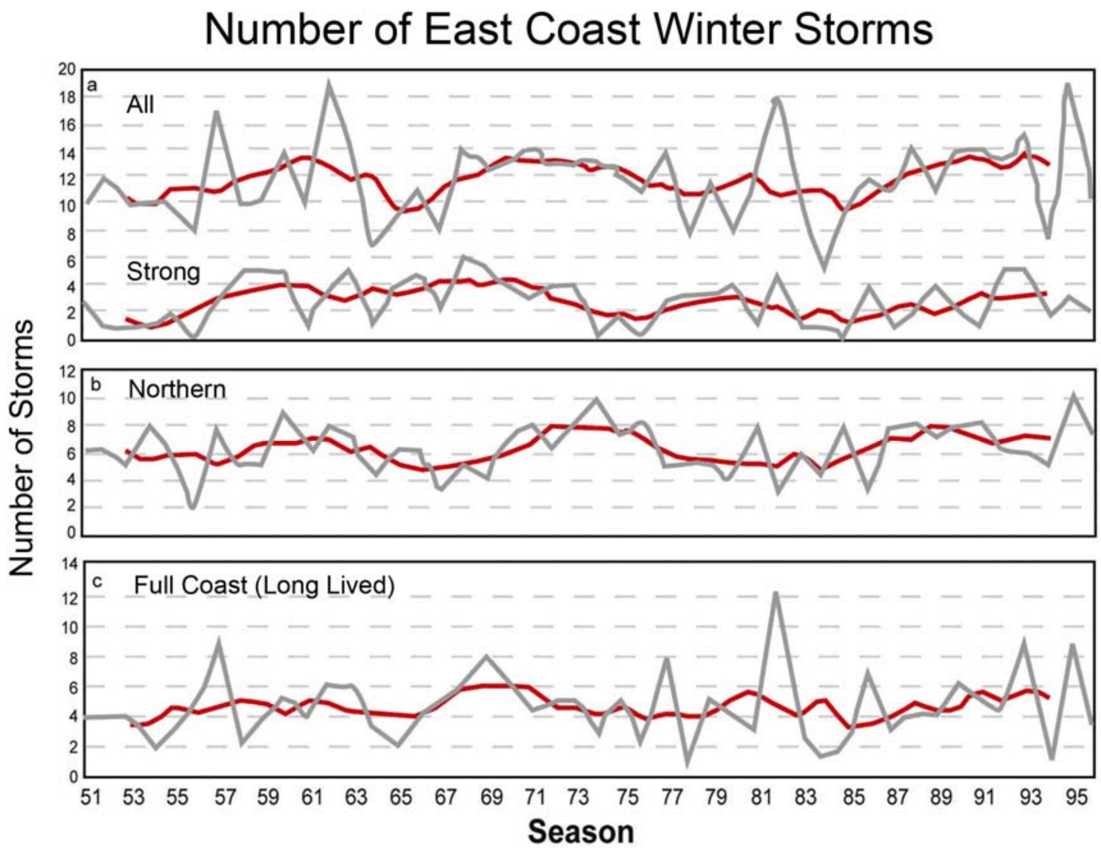

Figure 2.21 Seasonal totals (gray line) covering the period of 1951-1997 for (a) all East Coast Winter Storms (ECWS; top curve) and strong ECWS (bottom curve), (b) northern ECWS (>35^{o}N), and (c) those ECWS tracking along the full coast. Data points along the 5-yr moving average (black) correspond to the middle year. [Adapted from Hirsch et al. 2001].

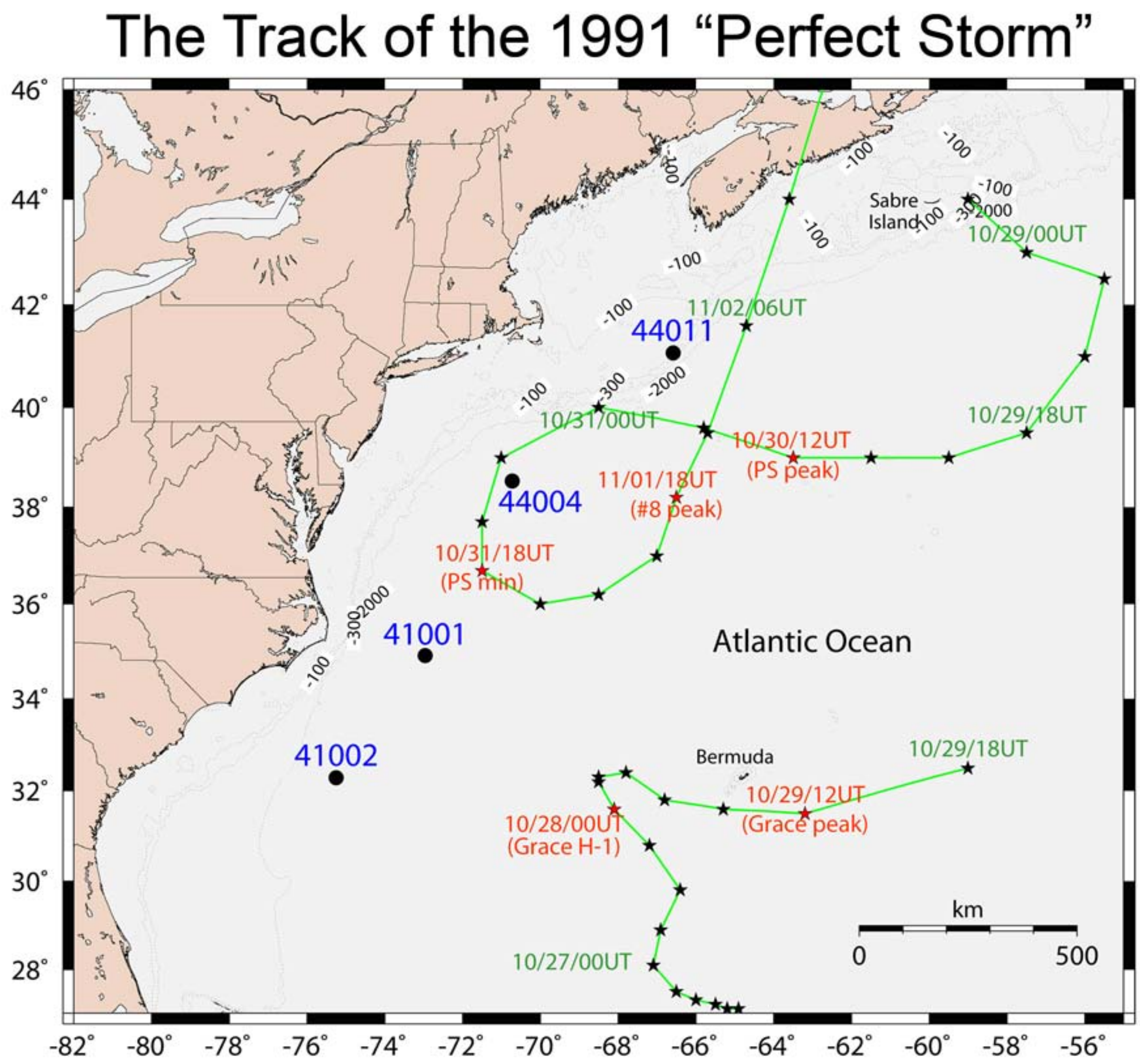

Figure 2.22 Track of the October 1991 "Perfect Storm" (PS) center showing the east-to-west retrograde propagation of a non-typical Nor'easter. The massive ETC was reenergized as it moved southward by absorbing northward propagating remnants of Hurricane Grace, becoming unnamed Hurricane #8 and giving rise to the name "Perfect Storm" for this composite storm. Storm center locations with date/hr time stamps at 6-hr intervals are indicated by stars. Also shown are locations of open ocean NOAA buoys that measured the extreme waves generated by these storms. [Adapted from Bromirski 2001].

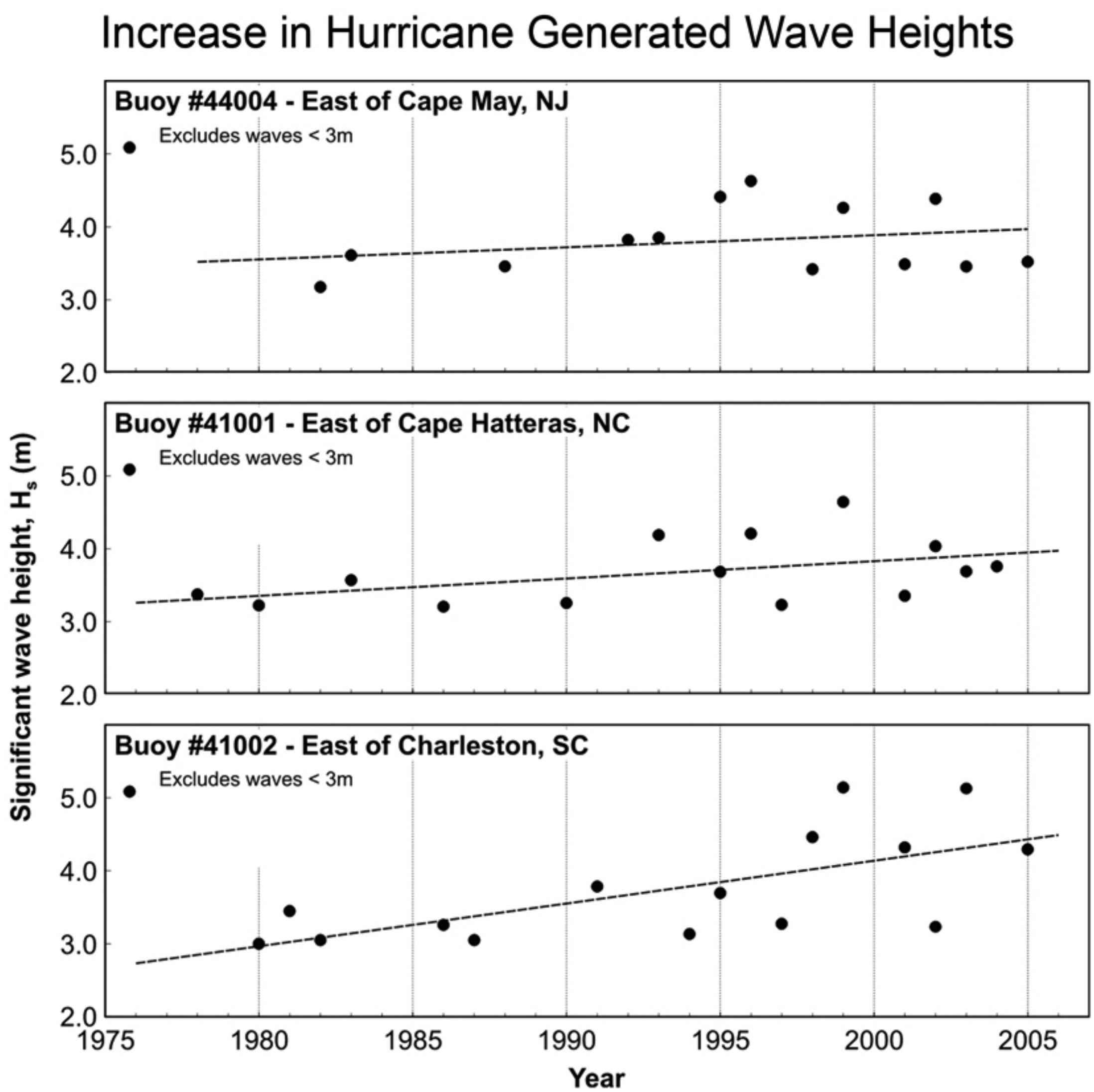

Figure 2.23 Increases in the summer, hurricane-generated wave heights of 3 meters and higher significant wave heights (from Komar and Allan 2007, and in review).

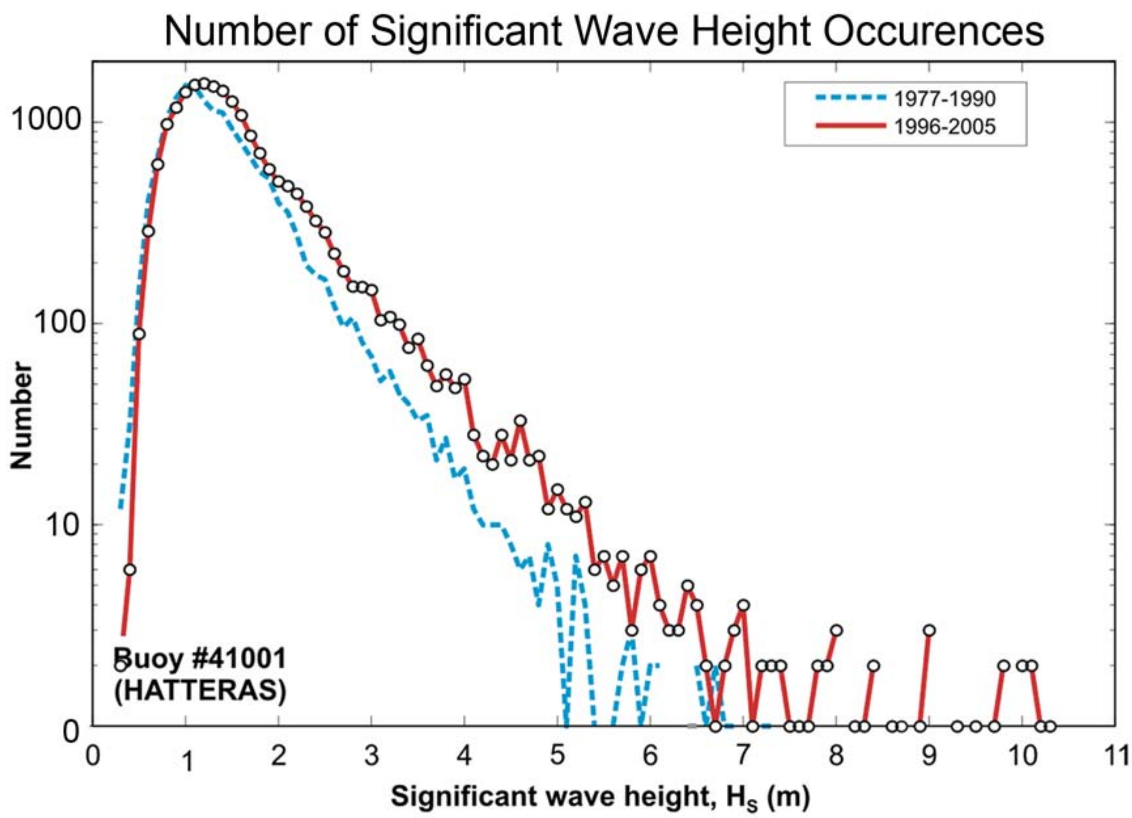

Figure 2.24 Number of significant wave heights measured by the Cape Hatteras buoy during the July-September season, early in its record 1976-1991 and during the recent decade,1996-2005 (from Komar and Allan 2007a,b).

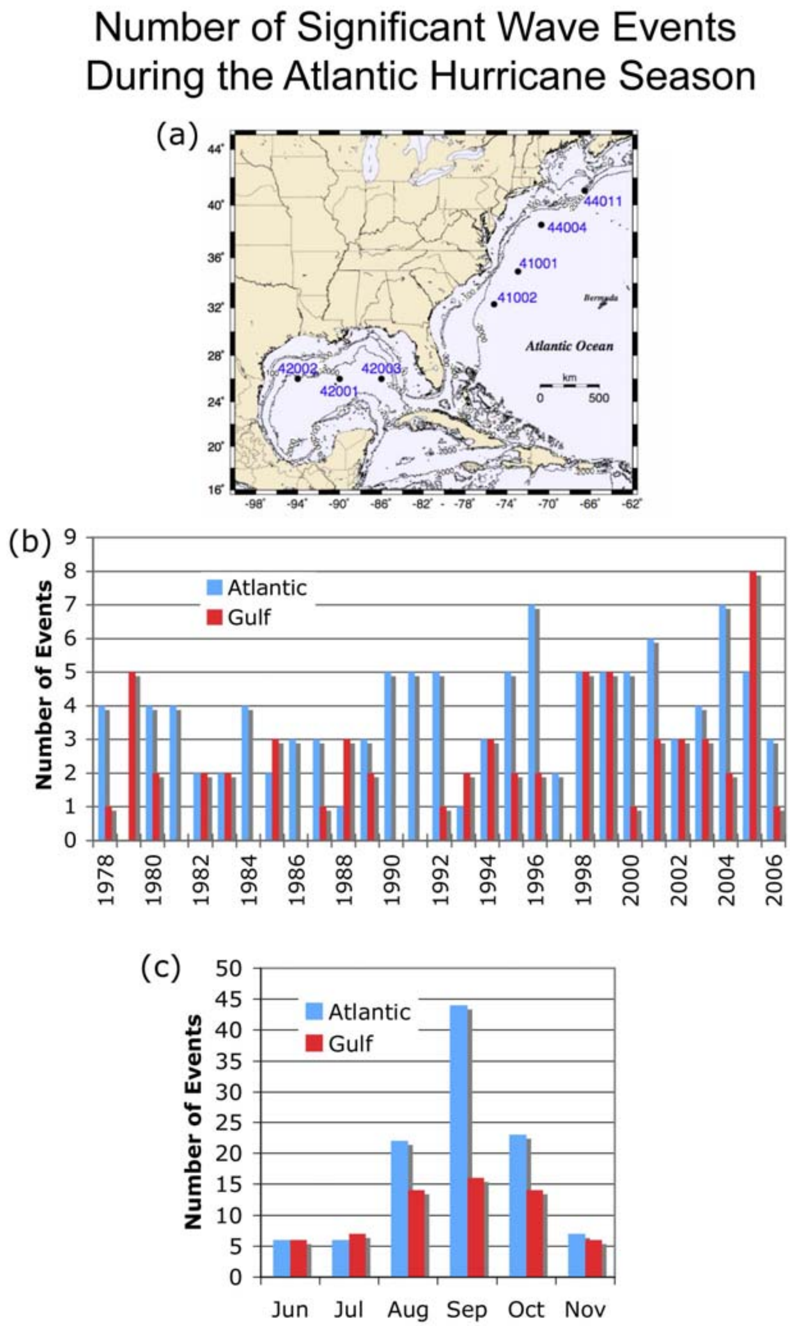

Figure 2.25 (a) Location of the NOAA Atlantic and Gulf buoys discussed. Bathymetric contours identify the continental shelf boundary. (b) Total number of significant wave events per hurricane season. (c) Total number of wave events identified during each month of the June-November hurricane season for all buoy data available from NOAA's National Ocean Data Center (NODC) from 1978-2006. Panels (b) and (c) show the

number of wave events associated with hurricanes/tropical storms with wave heights that exceeded 3 m at a minimum of one of the buoys in each group. Each event was counted only once, even if observed at multiple buoys in a group. No data were available from NODC for any of the Atlantic buoys during the 1979 hurricane season. [Adapted from Bromirski, 2007b]

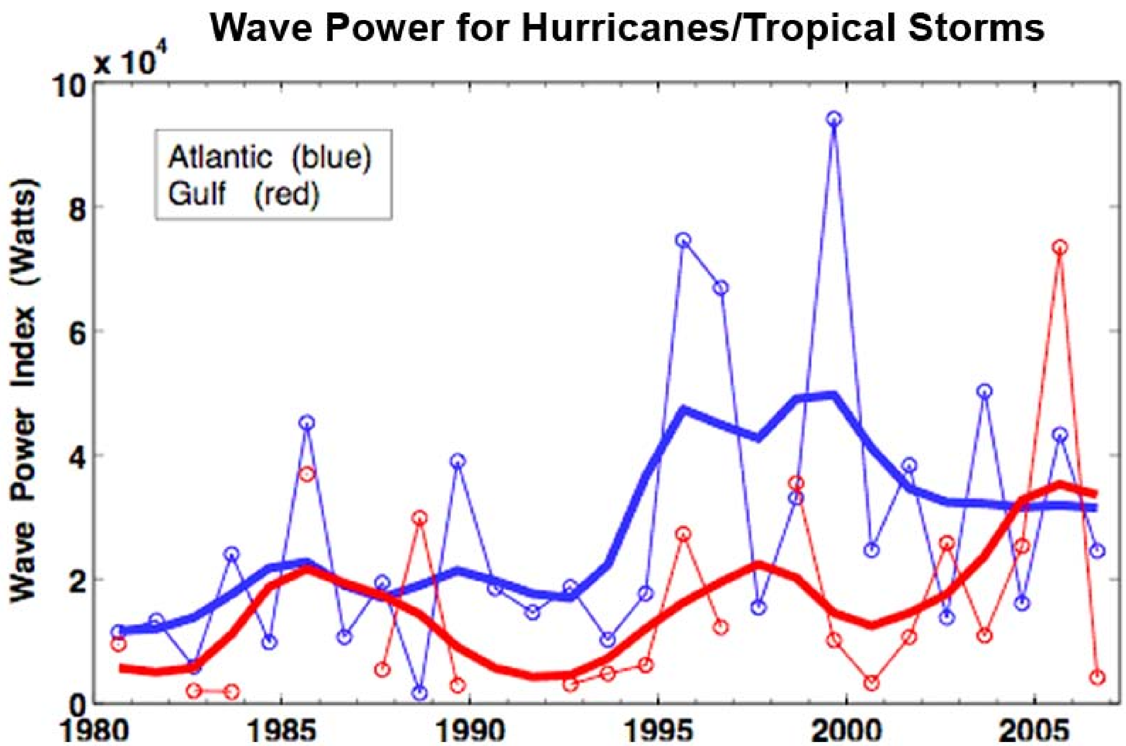

Figure 2.26 A measure of the total annual tropical cyclone wave power in the western North Atlantic and Gulf regions obtained as the mean of the available annual deep water wave power (the wave power index, WPI). Longer period variability is emphasized by lowpass filtering the annual data with three iterations of a 1-2-1 smoothing operator, giving the Atlantic and Gulf region WPI (thick lines). [Adapted from Bromirski, 2007b)

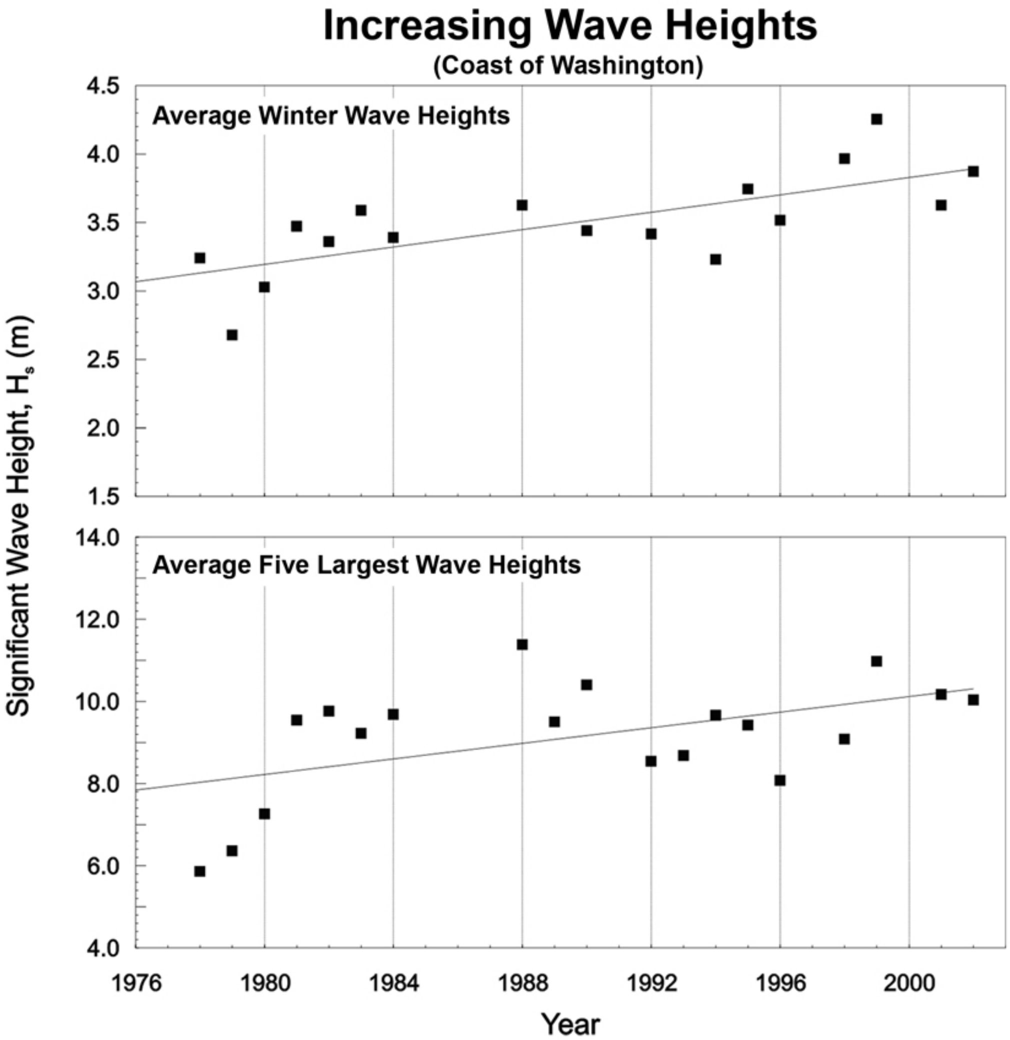

Figure 2.27 The trends of increasing wave heights measured by NOAA's National Data Buoy Center (NDBC) buoy #46005 off the coast of Washington [after Allan and Komar (2006)]

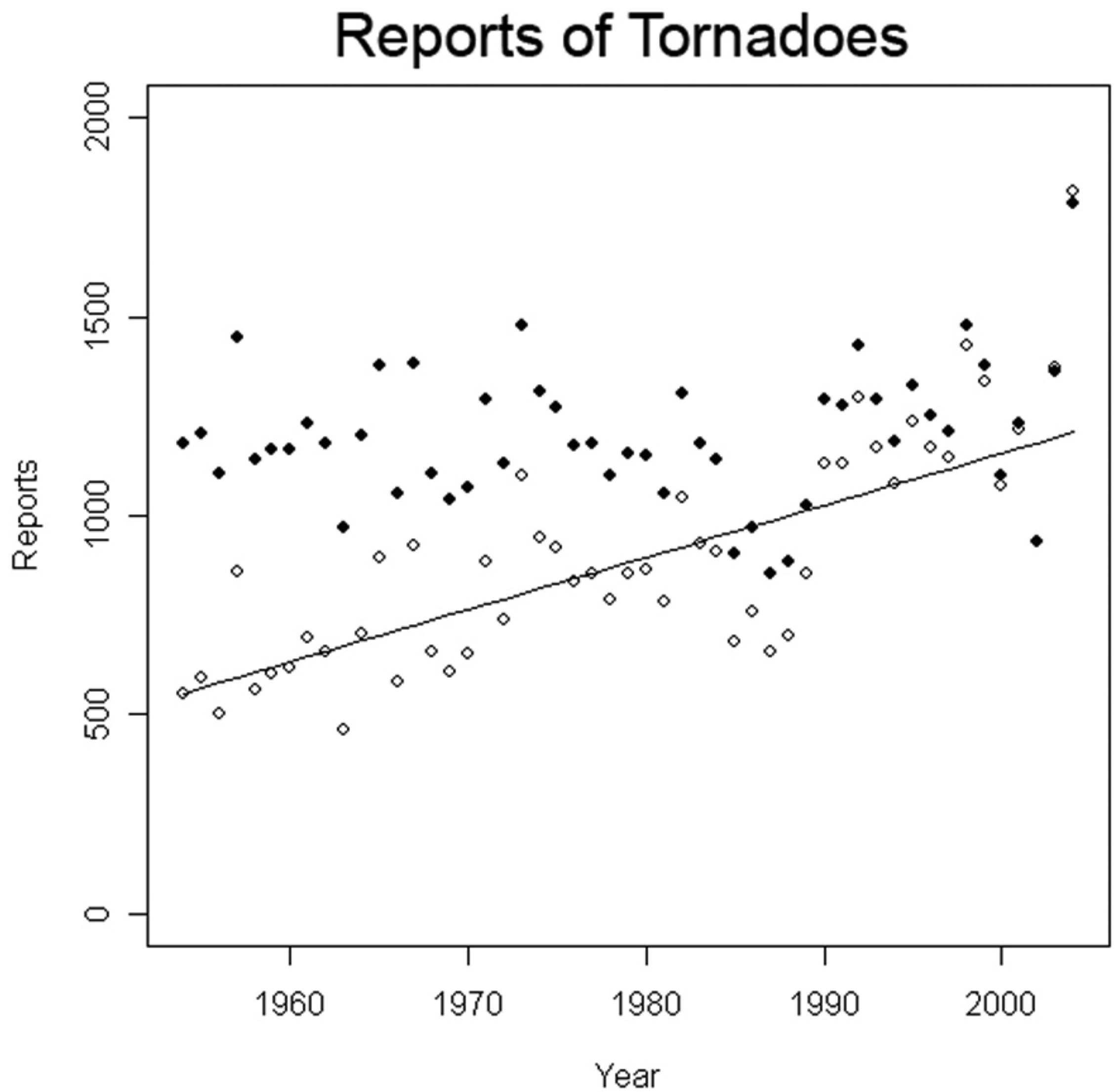

Figure 2.28 Tornado reports in official database in USA from 1954-2004. Open circles are raw reports, solid line (linear regression) is the trend for raw reports, solid circles are reports adjusted to 2002 reporting system. The adjusted data show little or no trend in reported tornadoes. The trend in raw reports reflects an increasing density of population in tornado-prone areas, and therefore more opportunity for sightings, rather than a real increase in the occurrences of tornadoes.

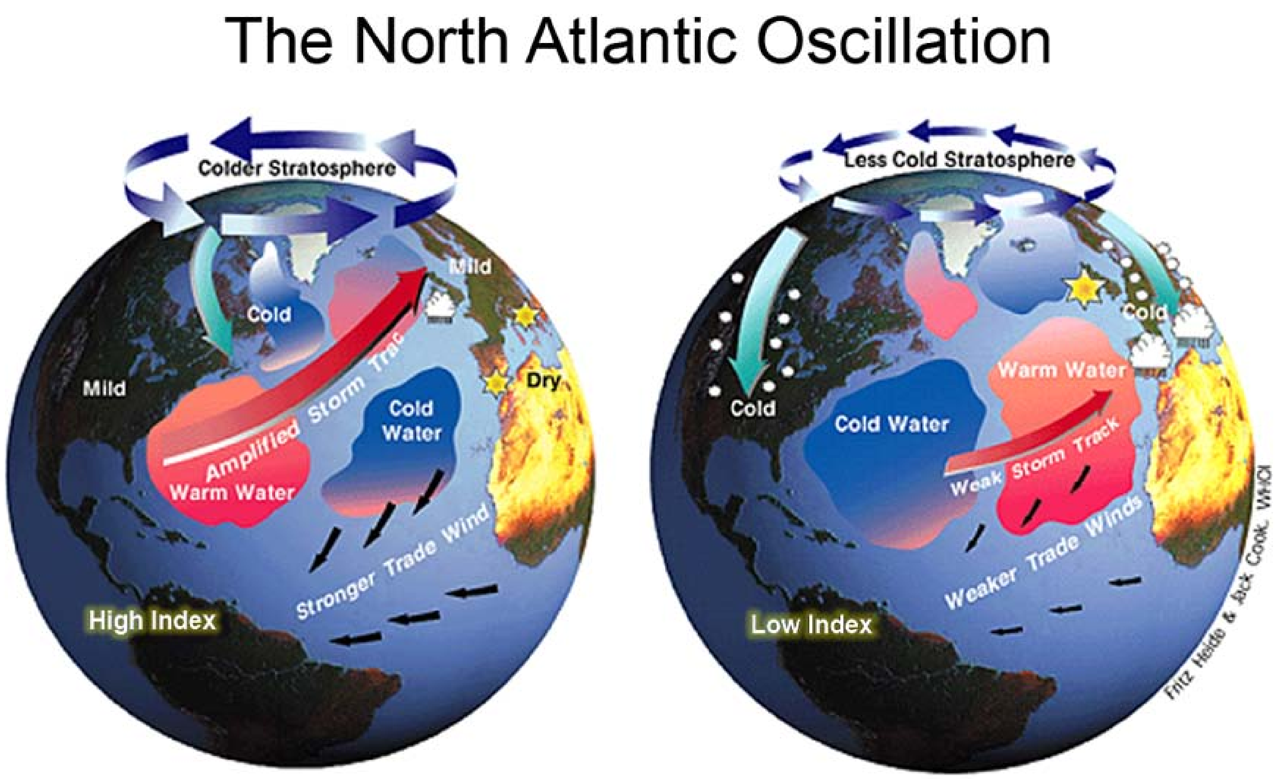

Figure 2.29 Schematic of the North Atlantic Oscillation (NAO) showing its effect on extremes. Illustrations by Fritz Heidi and Jack Cook, Woods Hole Oceanographic Institution.

Chapter 3 How Well Do We Understand the Causes of Observed Changes in Extremes, and What Are the Projected Future Changes?

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KEY FINDINGS

Observed Changes

Changes in some weather and climate extremes are attributable to human-induced changes in greenhouse gases.

- Human-induced warming has likely caused much of the average temperature increase in North America over the past 50 years. This affects changes in temperature extremes.
- Heavy precipitation events averaged over North America have increased over the past 50 years, consistent with the increased water holding capacity of the atmosphere in a warmer climate and observed increases in water vapor over the oceans.

- It is likely that human activities have caused a discernable increase in sea surface temperatures in the hurricane formation region of the tropical Atlantic Ocean over the past 100 years. The balance of evidence suggests that human activity has caused a discernable increase in tropical storm/hurricane and major hurricane frequency in the North Atlantic.

Projected Changes

- Future changes in extreme temperatures will generally follow changes in average temperature:
  – 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.
  – The number of days with frost is very likely to decrease.
- Droughts are likely to become more frequent and severe in some regions as higher air temperatures increase the potential for evaporation.
- Over most regions, precipitation is likely to be less frequent but more intense, and precipitation extremes are very likely to increase.
- According to theory and models for North Atlantic and North Pacific hurricanes and typhoons (both basin-wide and land-falling):
  – Hurricane/typhoon core rainfall rates will likely increase by about 6 to 18% per degree Celsius tropical sea surface warming.
  – It is likely that surface wind speeds of the strongest hurricanes/typhoons will increase by about 2 to 10% per degree Celsius tropical sea surface warming.

– Frequency changes are too uncertain for confident projections.

– The spatial distribution of hurricanes/typhoons will likely change.

– Due to projected sea level rise, the potential for storm surge damage will very likely increase.

- There are likely to be more frequent deep low-pressure systems (strong storms) outside the tropics, with stronger winds and more extreme wave heights.

3.1 Introduction

Understanding physical mechanisms of extremes involves processes governing the timing and location of extreme behavior, such as ENSO cycles, as well as the mechanisms of extremes themselves (e.g., processes producing heavy precipitation). In other words, processes creating an environment conducive to extreme behavior and processes of the extreme behavior itself. This includes not only the extreme events but also the factors governing their timing and location.

A deeper understanding of physical mechanisms is of course important for understanding why extremes have occurred in the past and for predicting their occurrence in the future. Understanding physical mechanisms serves a further purpose for projected climate changes. Because the verification time for climate-change projections can be many decades into the future, strict verification of projections is not always possible. Other means of attaining confidence in projections are therefore needed. Confidence in projected changes in extremes increases when the physical mechanisms producing extremes in models are consistent with observed behavior. This requires careful analysis of the observed record as well as model output. Assessment of physical mechanisms is also necessary to determine the physical realism of changes in extremes. While physical consistency of simulations with observed behavior is not sufficient evidence for accurate projection, it is necessary.

3.2 What Are the Physical Mechanisms of Observed Changes in Extremes?

3.2.1 Detection and Attribution of Anthropogenic Influences on Climate Extremes Over North America

Climate change detection, as discussed in this chapter, is distinct from the concept that is used in Chapter 2. In that chapter, detection refers to the identification of change in a climate record that is statistically distinguishable from the record's previous characteristics. A typical example is the detection of a statistically significant trend in a temperature record. Here, detection and attribution involves the assessment of observed changes in relation to those that are expected to have occurred in response to external forcing. Detection of climatic changes in extremes involves demonstrating statistically significant changes in properties of extremes over time. Attribution further links those changes with variations in climate forcings, such as changes in greenhouse gases, solar radiation or frequency of volcanoes. Attribution is a necessary step toward identifying the physical causes of changes in extremes. Attribution often uses quantitative comparison between climate-model simulations and observations, comparing expected changes due to physical understanding integrated in the models with those that have been observed. By comparing observed changes with those anticipated to result from external forcing, detection and attribution studies also provide an assessment of the performance of climate models in simulating climate change. The relationships between observed and simulated climate change that are diagnosed in these studies also provide an important means of constraining projections of future change made with those models.

3.2.1.1 Detection and Attribution of Anthropogenic Changes in Mean Climate That Affect Climate Extremes

This section discusses the present understanding on the causes of large-scale changes in the climatic state over North America. Simple statistical reasoning indicates that substantial changes in the frequency and intensity of extreme events can result from a relatively small shift in the average of a distribution of temperatures, precipitation or other climate variables (Katz and Brown 1992). Expected changes in temperature extremes are largely but not entirely due to changes in seasonal mean temperatures. Some differences between the two arise because moderate changes are expected in the shape of the temperature distribution affecting climate extremes, for example, due to changes in snow cover, soil moisture, and cloudiness (e.g., Hegerl et al., 2004; Kharin et al., 2007). In contrast, increases in mean precipitation are expected to increase the precipitation variance , thus increasing precipitation extremes, but decreases in mean precipitation do not necessarily imply that precipitation extremes will decrease, because of the different physical mechanisms that control mean and extreme precipitation (e.g., Allen and Ingram, 2002; Kharin et al., 2007). Therefore, changes in the precipitation background state are also interesting for interpreting changes in extremes, although more difficult to interpret (Groisman et al., 1999). Relevant information about mean temperature changes appeared in Chapter 2. More detailed discussion of historical mean changes appears in CCSP Synthesis and Assessment Products 1-1, 1-2 and 1-3.

Global-scale analyses using space-time detection techniques have robustly identified the influence of anthropogenic forcing on the 20th century near-surface temperature changes. This result is robust to applying a variety of statistical techniques and using many different climate simulations (Hegerl et al., 2007). Detection and attribution analyses also indicate that over the past century there has likely been a cooling influence from aerosols and natural forcings counteracting some of the warming influence of the increasing concentrations of greenhouse gases. Spatial information is required in addition to temporal information to reliably detect the influence of aerosols and distinguish them from the influence of increased greenhouse gases.

A number of studies also consider sub-global scales. Studies examining North America find a detectable human influence on 20th century temperature changes, either by considering the 100-year period from 1900 (Stott 2003) or the 50-year period from 1950 (Zwiers and Zhang 2003, Zhang et al. 2006). Based on such studies, a substantial part of the warming over North America has been attributed to human influence (Hegerl et al., 2007).

Further analysis has compared simulations using changes in both anthropogenic (greenhouse gas and aerosol) and natural (solar flux and volcano eruption) forcings with others that neglect anthropogenic changes. There is a clear separation in North American temperature changes of ensembles of simulations including just natural forcings from ensembles of simulations containing both anthropogenic and natural forcings (Karoly et al. 2003, IDAG 2005, Karoly and Wu 2005, Wang et al. 2006, Knutson et al. 2006, Hegerl et al. 2007), especially for the last quarter of the 20th century, indicating that the warming in recent decades is inconsistent with natural forcing alone.

Attribution of observed changes on regional (subcontinental) scales has generally not yet been accomplished. One reason is that as spatial scales considered become smaller, the uncertainty becomes larger (Stott and Tett 1998, Zhang et al., 2006) because internal climate variability is typically larger than the expected responses to forcing on these scales. Also, small-scale forcings and model uncertainty make attribution on these scales more difficult. Therefore, interpreting changes on sub-continental scales is difficult (see discussion in Hegerl et al., 2007). For example, in Alaska warming has been large but high levels of internal variability lead to an overlap of naturally forced and all-forcing simulations even at the end of the 20th century (Wang et al. 2007). In central North America, there is a relatively small warming over the 20th century compared to other regions around the world (Hegerl et al. 2007) and the observed changes lie (just) within the envelop of changes simulated by models using natural forcing alone. In this context, analysis of a multi-model ensemble by Kunkel et al. (2006) for a central U.S. region suggests that the region’s warming from 1901to 1940 and cooling from 1940 to 1979 may have been a consequence on unforced internal variability.

Burkholder and Karoly (2007) detected an anthropogenic signal in multidecadal trends of a U.S. climate extremes index. The observed increase is largely due to an increase in the number of months with monthly mean daily maximum and daily minimum temperatures that are much above normal and an increase in the area of the US that experienced a greater than normal proportion of their precipitation from extreme one-day events. Twentieth century simulations from coupled climate models show a similar, significant increase in the same US climate extremes index for the late twentieth century. There is some evidence of an anthropogenic signal in regions a few hundred kilometers across (Karoly and Wu 2005, Knutson et al. 2006, Zhang et al. 2006, Burkholder and Karoly 2007), suggesting the potential for progress in regional attribution if careful attention is given to the choice of appropriate time scales, region sizes and fields analyzed and if all relevant forcings are considered.

Warming from greenhouse gas increases is expected to increase the moisture content of the atmosphere and lead to a small increase in global mean precipitation. More important, the increase in water holding capacity of the atmosphere is expected to affect more strongly changes in heavy precipitation, for which the Clausius-Clapeyron relation provides an approximate physical constraint (e.g., Allen and Ingram, 2002). Observed changes in moisture content and mean and extreme precipitation are generally consistent with these expectations (Chapter 2 of this document, Trenberth et al. 2007). In addition, greenhouse gas increases are also expected to cause enhanced horizontal transport of water vapor that is expected to lead to a drying of the subtropics and parts of the tropics

(Kumar et al., 2004; Neelin et al., 2006), and a further increase in precipitation in the equatorial region and at high latitudes (Emori and Brown, 2005; Held and Soden, 2006).

Several studies have demonstrated that simulated global land mean precipitation in climate model integrations including both natural and anthropogenic forcings is significantly correlated with that observed (Allen and Ingram, 2002; Gillett et al., 2004b; Lambert et al., 2004), thereby detecting external influence in observations of precipitation. This external influence on global land mean precipitation during the 20th century is dominated by volcanic forcing. Anthropogenic influence on the spatial distribution of global land precipitation, as represented by zonal-average precipitation changes, has also been detected (Zhang et al. 2007). Both changes are significantly larger in observations than simulated in climate models, raising questions about whether models underestimate the response to external forcing in precipitation changes (see also Wentz et al 2007). Changes in North American continental-mean rainfall have not yet been attributed to anthropogenic influences. A large part of North America falls within the latitude band identified by Zhang et al. (2007) where the model simulated response to forcing is not in accord with the observed response. However, both models and observations show a pattern of wetting north of 50N and drying between 0-30N, and this together with agreement on moistening south of the equator provides support for the detection of a global anthropogenic influence.

3.2.1.2 Detection and Attribution of Other Modes of Climate-system Behavior Affecting Climate Extremes

North American extreme climate is also substantially affected by changes in atmospheric circulation (e.g., Thompson and Wallace 2001). Natural low frequency variability of the climate system is dominated by a small number of large-scale circulation patterns such as the El Niño Southern Oscillation (ENSO), the Pacific Decadal Oscillation (PDO), and the Northern Annular Mode (NAM). The impact of these modes on terrestrial climate on annual to decadal time scales can be profound. In particular, there is considerable evidence that the state of these modes affects substantially the risk of extreme temperature (Thompson and Wallace 2002, Kenyon and Hegerl 2007), droughts (Hoerling and Kumar 2003), and short-term precipitation extremes (e.g., Gershunov and Cayan 2003, Eichler and Higgins 2006) over North America.

Some evidence of anthropogenic influence on these modes appears in surface-pressure analyses. Gillett et al. (2003, 2005, 2006) and Wang et al. (2007) diagnosed anthropogenic influence on Northern Hemisphere sea level pressure change, although the model-simulated change is not as large as has been observed. Model-simulated changes in extremes related to circulation changes may therefore be affected. The change in sea level pressure largely manifests itself through an intensification of the Northern and Southern Annular Modes with reduced pressure above both poles and equatorward displacement of mass. However, apart from these modes, the extent to which modes of variability are excited or altered by external forcing remains uncertain. While some modes might be expected to change as a result of anthropogenic effects such as the enhanced greenhouse effect, there is little a priori expectation about the direction or magnitude of such changes. In addition, models may not simulate well the behavior of these modes in some regions and seasons.

ENSO is the leading mode of variability in the tropical Pacific, and it has impacts on climate around the globe (Trenberth et al. 2007, see also Chapter 1 of this report). There have been multi-decadal oscillations in the ENSO index throughout the 20th century, with more intense El Niño events since the late 1970s, which may reflect in part a mean warming of the eastern equatorial Pacific (Mendelssohn et al., 2005). There is presently no clear consensus on the possible impact of anthropogenic forcing on observed ENSO variability (Merryfield 2006, Meehl et al. 2007).

Decadal variability in the North Pacific is characterised by variations in the strength of the Aleutian Low coupled to changes in North Pacific SST. The leading mode of decadal variability in the North Pacific is usually termed the Pacific Decadal Oscillation (PDO) and has a spatial structure in the atmosphere and upper North Pacific Ocean similar to the pattern that is associated with ENSO. Pacific Decadal variability can also be characterized by changes in sea level pressure in the North Pacific, termed the North Pacific Index (Deser et al., 2004). One recent study showed a consistent tendency towards the positive phase of the PDO in observations and model simulations that included anthropogenic forcing (Shiogama et al., 2005), though differences between the observed and simulated PDO patterns, and the lack of additional studies, limit confidence in these findings.

ENSO and Pacific decadal variability affect the mean North American climate and its extremes (e.g., Kenyon and Hegerl, 2007), particularly when both are in phase, at which time considerable energy is propagated from tropical and northern Pacific sources towards the North American land mass (Yu et al. 2007, Zwiers and Yu 2007).

The Northern Annular Mode (NAM) is an approximately zonally symmetric mode of variability in the Northern Hemisphere (Thompson and Wallace, 1998, Chapter 1 of this report), and the North Atlantic Oscillation (NAO) (Hurrell, 1996) may be viewed as its Atlantic counterpart. The NAM index exhibited a pronounced trend towards its positive phase between the 1960s and the 1990s, corresponding to a decrease in surface pressure over the Arctic and an increase over the subtropical North Atlantic (e.g., Hurrell, 1996; Thompson et al., 2000; Gillett et al., 2003a). Several studies have shown this trend to be inconsistent with simulated internal variability (Osborn et al., 1999; Gillett et al., 2000; Gillett et al., 2002b; Osborn, 2004; Gillett, 2005) and similar to, although larger than, simulated changes in coupled climate models in response to 20th century forcing, particularly, greenhouse gas forcing and ozone depletion (Gillett et al., 2002b, Osborn, 2004, Gillet 2005, Hegerl et al. 2007). The mechanisms underlying Northern Hemisphere circulation changes also remain open to debate (see e.g., Hoerling et al., 2005; Hurrell et al., 2005, Scaife et al. 2005).

Over the period 1968–1997, the trend in the NAM was associated with approximately 50% of the winter surface warming in Eurasia, a decrease in winter precipitation over

Southern Europe and an increase over Northern Europe, due the northward displacement of the storm track (Thompson et al., 2000). Such a change would have substantial influence on North America, too, reducing the probability of cold extremes in winter even over large areas (for example, Thompson and Wallace, 2001; Kenyon and Hegerl, 2007), although part of the northeastern U.S. tends to show a tendency for more cold extremes with the NAO trend (Wettstein and Mearns, 2002).

3.2.2 Changes in Temperature Extremes

As discussed in Chapter 2, observed changes in temperature extremes are consistent with the observed warming of the climate (Alexander et al., 2006). Globally, there has been a widespread reduction in the number of frost days in mid-latitude regions in recent decades, an increase in the number of warm extremes, particularly warm nights, and a reduction in the number of cold extremes, such as cold nights.

There is now evidence that anthropogenic forcing has likely affected extreme temperatures. Christidis et al. (2005) analyzed a new dataset of gridded daily temperatures (Caesar et al., 2006) using the indices shown by Hegerl et al. (2004) to have potential for attribution, namely the average temperature of the most extreme 1, 5, 10 and 30 days of the year. Christidis et al. (2005) detected robust anthropogenic changes in a global analysis of indices of extremely warm nights using fingerprints from the HadCM3 model, with some indications that the model over-estimates the observed warming of warm nights. Human influence on cold days and nights was also detected, but in this case the model underestimated the observed changes, significantly so in the case of the coldest day of the year. Anthropogenic influence was not detected in observed changes in extremely warm days. Tebaldi et al. (2006) find that changes simulated by an ensemble of eight global models that include anthropogenic and natural forcing changes agrees well with observed global trends in heat waves, warm nights and frost days over the last four decades.

North American observations also show a general increase in the number of warm nights, but with a decrease in the center of the continent that models generally do not reproduce (e.g., Christidis et al 2005). However, analysis for North America of models (Table 3.1) used by Tebaldi et al. (2006) shows reasonable agreement between observed and simulated changes in the frequency of warm nights, number of frost days and growing season length over the latter half of the 20th century when averaged over the continent (Fig. 3.1a,b,c). There is also good agreement between the observed and ensemble mean simulated spatial pattern of change in frost days (Fig.3.2a,b) over the latter half of the 20th century. Note that the observational estimate has a much greater degree of temporal (Fig. 3.1) and spatial (Fig. 3.2) variability than the model result. The model result is derived from an ensemble of simulations produced by many models, some of which contributed multiple realizations. Averaging over many simulations reduces much of the spatial and temporal variability that arises from internal climate variability. The variability of individual model realizations is comparable to the single set of observations, which is well bounded by the two standard deviation confidence interval about the model ensemble average. Furthermore, Meehl et al. (2007b) demonstrate that ensemble simulations using two coupled climate models driven with human and natural forcings approximate well the observed changes, but when driven with natural forcings only cannot reproduce the observed changes, indicating a human contribution to observed changes in heat waves, frost days and warm nights. Output from one of these ensembles, produced by the Parallel Climate Model, also shows significant trends in the Karl-Knight heat-wave index (Karl and Knight, 1997) in the eastern half of the U.S. for 1961-1990 that are similar to observed trends (Fig. 3.3).

There have also been some methodological advances whereby it is now possible to estimate the impact of external forcing on the risk of a particular extreme event. For example, Stott et al (2004), assuming a model-based estimate of temperature variability, estimate that past human influence may have more than doubled the risk of European mean summer temperatures as high as those recorded in 2003. Such a methodology has not yet been applied to North American extremes, though Hoerling et al. (2007) have used the method to conclude that the very hot 2006 in the United States was primarily due to human influences.

3.2.3 Changes in Precipitation Extremes

3.2.3.1 Heavy Precipitation

Allen and Ingram (2002) suggest that while global annual mean precipitation is constrained by the energy budget of the troposphere, extreme precipitation is constrained by the atmospheric moisture content, as governed by the Clausius-Clapeyron equation, though this constraint may be most robust in extratropical regions and seasons where the circulation's fundamental dynamics are not driven by latent heat release (Pall et al. 2007). For a given change in temperature the constraint predicts a larger change in extreme precipitation than in mean precipitation, which is consistent with changes in precipitation extremes simulated by the ensemble of GCMs available for the IPCC Fourth Assessment Report (Kharin et al, 2007). Emori and Brown (2005) discuss physical mechanisms governing changes in the dynamic and thermodynamic components of mean and extreme precipitation and conclude that changes related to the dynamic component (i.e., that due to circulation change) are secondary factors in explaining the larger increase in extreme precipitation than mean precipitation seen in models. On the other hand, Meehl et al. (2005) demonstrate that while tropical precipitation intensity increases are related to water vapour increases, mid-latitude intensity increases are related to circulation changes that affect the distribution of increased water vapor.

Climatological data show that the most intense precipitation occurs in warm regions (Easterling et al., 2000) and diagnostic analyses have shown that even without any change in total precipitation, higher temperatures lead to a greater proportion of total precipitation in heavy and very heavy precipitation events (Karl and Trenberth, 2003). In addition, Groisman et al. (1999) have demonstrated empirically, and Katz (1999) theoretically, that as total precipitation increases a greater proportion falls in heavy and very heavy events if the frequency of raindays remains constant. Trenberth et al. (2005) point out that a consequence of a global increase in precipitation intensity should be an offsetting global decrease in the duration or frequency of precipitation events, though some regions could have differing behavior, such as reduced total precipitation or increased frequency of precipitation.

Simulated changes in globally averaged annual mean and extreme precipitation appear to be quite consistent between models. The greater and spatially more uniform increases in heavy precipiation as compared to mean precipitation may allow extreme precipitation change to be more robustly detectable (Hegerl et al., 2004).

Evidence for changes in observations of short-duration precipitation extremes varies with the region considered (Alexander et al., 2006) and the analysis method that is employed (e.g., Trenberth et al., 2007). Significant increases in observed extreme precipitation have been reported over the United States, where the increase is qualitatively similar to changes expected under greenhouse warming (e.g., Karl and Knight, 1998; Semenov and Bengtsson, 2002; Groisman et al., 2005). However, a quantitative comparison between area-based extreme events simulated in models and station data remains difficult because of the different scales involved (Osborn and Hulme, 1997, Kharin et al. 2005) and the pattern of changes does not match observed changes. Part of this difference is expected since most current GCMs do not simulate small-scale (< 100 km) variations in precipitation intensity, as occurs with convective storms. Nevertheless, when compared with a gridded reanalysis product (ERA40), the ensemble of currently available AOGCMs reproduces observed precipitation extremes reasonably well over North America (Kharin et al., 2007). An attempt to detect anthropogenic influence on precipitation extremes using global data based on the Frich et al. (2002) indices used fingerprints from atmospheric model simulations with prescribed sea surface temperature (Kiktev et al., 2003). This study found little similarity between patterns of simulated and observed rainfall extremes. This is in contrast to the qualitative similarity found in other studies (Semenov and Bengtsson, 2002; Groisman et al., 2005; Fig. 3.4). Tebaldi et al. (2006) reported that an ensemble of eight global climate models simulating the 20th century showed a general tendency toward more frequent heavy-precipitation events over the past four decades, most coherently in the high latitudes of the Northern Hemisphere, broadly consistent with observed changes (Groisman et al., 2005). This is also seen when analyzing these models for North America (Fig. 3.1d). The pattern similarity of change in precipitation extremes over this period is more difficult to assess, particularly on continental and smaller scales.

3.2.3.2 Runoff and Drought

Changes in runoff have been observed in many parts of the world, with increases or decreases corresponding to changes in precipitation. Climate models suggest that runoff will increase in regions where precipitation increases faster than evaporation, such as at high Northern latitudes (Milly et al., 2005; Wu et al., 2005). Gedney et al. (2006a) attributed increased continental runoff in the latter decades of the 20th century in part to suppression of transpiration due to CO_2-induced stomatal closure. However, their result is subject to considerable uncertainty in the runoff data (Peel and McMahon, 2006; Gedney et al. 2006b). Qian et al. (2006) simulate observed runoff changes in response to observed temperature and precipitation alone, and Milly et al. (2005) demonstrate that 20th century runoff trends simulated by several global climate models are significantly correlated with observed runoff trends. Wu et al. (2005) find that observed increases in Arctic river discharge are simulated in a global climate model with anthropogenic and natural forcing, but not in the same model with natural forcings only. Anthropogenic changes in runoff may be emerging, but attribution studies specifically on North American runoff are not available.

Mid-latitude summer drying is another anticipated response to greenhouse gas forcing (Meehl et al., 2006) and drying trends have been observed in the both the Northern and Southern hemispheres since the 1950's (Trenberth et al., 2006). Burke et al. (2006), using the HadCM3 model with all natural and anthropogenic external forcings and a global Palmer Drought Severity Index (PDSI) dataset compiled from observations by Dai et al. (2004), detect the influence of anthropogenic forcing in the observed global trend towards increased drought in the second half of the 20th century, although the model trend was weaker than observed and the relative contributions of natural external forcings and anthropogenic forcings was not assessed. Nevertheless, this supports the conclusion that anthropogenic forcing has influenced the global occurrence of drought. However, the spatial pattern of observed PDSI change over North America is dissimilar to that in the coupled model, so no anthropogenic influence has been detected for North America alone.

Nevertheless, the long term trends in the precipitation patterns over North America are well reproduced in atmospheric models driven with observed changes in sea-surface temperatures (Schubert et al., 2003; Seager et al., 2005), indicating the importance of sea-surface temperatures in determining North American drought (see also, for example, Hoerling and Kumar, 2003). Specifically, Schubert et al. (2003) and Seager et al. (2005), using AGCMs forced with observed SSTs, show that some SST anomaly patterns, particularly in the tropical Pacific, can produce drought over North America. Using the observed SST anomalies, both studies successfully reproduce many aspects of the 1930's drought. Only the Seager et al. (2005) model simulates the 1950's drought over North America, indicating that more modelling studies of this kind are needed.

3.2.4 Tropical Cyclones

Long-term (multidecadal to century) scale observational records of tropical cyclone activity (frequency, intensity, power dissipation, etc.) were described in Chapter 2. Here discussion focuses on whether the any changes can be attributed to particular causes, including anthropogenic forcings. Tropical cyclones respond to their environment in quite different manners for initial development, intensification, determination of overall size, and motion. Therefore this section begins with a brief summary of the major physical mechanisms and understanding.

3.2.4.1 Development Criteria and Mechanisms

Gray (1968) drew on a global analysis of tropical cyclones and a large body of earlier work to arrive at a set of criteria for tropical cyclone development, which he called Seasonal Genesis Parameters:

- Sufficient available oceanic energy for the cyclone to develop, usually defined as a requirement for ocean temperatures > 26 ^{o}C down to a depth of 60 m;
- Sufficient cyclonic (counterclockwise in Northern Hemisphere, clockwise in Southern Hemisphere) rotation to enhance the capacity for convective heating to accelerate the vertical winds,

- A small change in horizontal wind with height (weak shear) so that the upper warming can become established over the lower vortex.
- A degree of atmospheric moist instability to enable convective clouds to develop;
- A moist mid-level atmosphere to inhibit the debilitating effects of cool downdrafts; and
- Some form of pre-existing disturbance, such as an easterly wave, capable of development into a tropical cyclone.

A more recent study by Camargo et al. (2007) has developed a new genesis index, which is based on monthly mean values of 850 hPa relative vorticity, 700 hPa humidity, 850-250 hPa wind shear, and Potential Intensity (Bister and Emanuel, 1998). Some skill has been demonstrated in applying it to re-analysis data and global climate models to estimate the frequency and location of storms.

In the North Atlantic, the bulk of tropical cyclone developments arise from easterly waves, though such development is a relatively rare event, with only around 10-20% of waves typically developing into a tropical cyclone (Dunn 1940, Frank and Clarke 1980, Pasch et al 1998, Thorncroft and Hodges 2001). Thus, any large-scale mechanism that can help produce more vigorous easterly waves leaving Africa or provide an environment to enhance their development is of importance. ENSO is a major influence; during El Nino years, tropical cyclone development is suppressed by a combination of associated increased vertical wind shear, general drying of the mid-levels and oceanic cooling (e.g., Gray 1984). The Madden-Julian Oscillation (MJO) influences cyclogenesis in the Gulf of

Mexico region on 1-2 month time scales (Maloney and Hartmann 2000). Approximately half of the North Atlantic tropical cyclone developments are associated with upper-level troughs migrating into the tropics (e.g. Pasch et al 1998; Davis and Bosart, 2001; 2006). The large scale zonal wind flow may also modulate development of easterly wave troughs into tropical cyclones (Holland 1995, Webster and Chang 1988). The easterly wave development process is particularly enhanced in the wet, westerly phase of the MJO.

The eastern and central North Pacific experience very little subtropical interaction and appear to be dominated by easterly wave development (e.g. Frank and Clarke 1980). The two major environmental influences are the ENSO and MJO, associated with the same effects as described for the North Atlantic. The MJO is a particularly large influence, being associated with a more than 2:1 variation in tropical cyclone frequency between the westerly-easterly phases (Liebmann et al 1994, Molinari and Vollaro 2000).

Suitable conditions in the western Pacific development region are present throughout the year. Developments in this region are associated with a variety of influences, including easterly waves, monsoon development and mid-latitude troughs (e.g. Ritchie and Holland 1999). The dominant circulation is the Asiatic monsoon, and tropical cyclones typically form towards the eastern periphery of the main monsoonal trough, or further eastwards (Holland, 1995), though development can occur almost anywhere (e.g. Lander 1994). ENSO has a major impact, but it is opposite to that in the eastern Pacific and Atlantic, with western Pacific tropical cyclone development being enhanced during the El Nino phase (Chan 1985, Lander 1994, Wang and Chan 2002).

Mesoscale influences include those that occur on scales similar to, or smaller than the tropical cyclone circulation and seem to be operative in some form or other to all ocean basins. These influences include interactions amongst the vorticity fields generated by Mesoscale Convective Complexes (MCCs), which may enhance cyclogenesis under suitable atmospheric conditions, but also may introduce a stochastic element in which the interactions may also inhibit short-term development (Houze 1977; Zipser 1977; Ritchie and Holland 1997; Simpson et al. 1997; Ritchie 2003; Bister and Emanuel 1997; Hendricks et al. 2004; Montgomery et al. 2006) and inherent barotropic instability (e.g. Schubert et al. 1991; Ferreira and Schubert 1997).

3.2.4.1.1 Factors Influencing Intensity and Duration

Once a cyclone develops it proceeds through several stages of intensification. The maximum achievable intensity of a tropical cyclone appears to be limited by the available energy in the ocean and atmosphere. This has led to various thermodynamic assessments of the Potential Intensity (PI) that can be achieved by a cyclone for a given atmospheric/oceanic thermodynamic state (Emanuel 1987, 1995, 2000, Holland 1997, Tonkin et al 1999, Rotunno and Emanuel 1987). The basis for these assessments is characteristically the sea surface temperature and the thermodynamic structure of the near-cyclone atmospheric environment, with particular emphasis on the temperature at the outflow level of air ascending in the storm core.

In most cases tropical cyclones do not reach this thermodynamic limit, due to a number of processes that have a substantial negative influence on intensification. Major negative impacts may include: vertical shear of the horizontal wind (Frank and Ritchie 1999, DeMaria 1995), oceanic cooling by cyclone-induced mixing of cool water from below the mixed layer to the surface (Price 1981, Bender and Ginis 2000, Schade and Emanuel 1999); potential impacts of sea spray on the surface exchange process (Wang et al. 2001, Andreas and Emanuel 2001); processes that force the cyclone into an asymmetric structure (Wang 2002, Corbosiero and Molinari 2003); ingestion of dry air, perhaps also with suspended dust (Neal and Holland 1976, Dunion and Velden 2004); and internal processes. Since many of these factors tend to be transitory in nature, the longer a cyclone can spend in a region with plentiful thermodynamic energy, the better its chances of approaching the PI. This is reflected in, for example, the observation that over 80% of major hurricanes in the North Atlantic occur in systems that formed at low latitudes in the eastern region, the so-called Cape Verde storms.

A weakening tropical cyclone may merge with an extratropical system, or it may redevelop into a baroclinic system (Jones et al. 2003). Since the system carries some of its tropical vorticity and moisture, it can produce extreme rains and major flooding. The transition is also often accompanied by a rapid acceleration in translation speed, which leads to an asymmetric wind field with sustained winds that may be of hurricane force on the right (left) side of the storm track in the northern (southern) hemisphere, despite the overall weakening of the cyclone circulation.

3.2.4.1.2 Movement Mechanisms

Tropical cyclones are steered by the mean flow in which they are embedded, but they also propagate relative to this mean flow due to dynamical effects (Holland 1984, Fiorino and Elsberry 1989). This combination leads to the familiar hyperbolic (recurving) track of tropical cyclones as storms initially move westward, embedded in the low-latitude easterly flow, then more poleward and eventually eastward as they encounter the mid-latitude westerlies.

An important result of this pattern of movement is that storms affecting the Caribbean, Mexico, Gulf States, Lower Eastern Seaboard and Pacific Trust Territories have mostly developed in low-latitudes (which also comprise the most intense systems). Eastern Pacific cyclones tend to move away from land, and those that recurve are normally suffering from combined negative effects of cold water and vertical shear. Upper Eastern US Seaboard and Atlantic Canada cyclones are typically recurving and undergoing various stages of extratropical transition.

3.2.4.2 Attribution Preamble

Determining the causal influences on the observed changes in tropical cyclone characteristics is currently subject to vigorous community debate. Chief amongst the more contentious topics are data deficiencies in early years, natural variability on decadal time scales, and trends associated with greenhouse warming. A summary of the published contributions to this debate at the end of 2006 is contained in a report and accompanying statement that was put together by the attendees at a World Meteorological Society Workshop on Tropical Cyclones held in November 2006 (WMO 2006, Knutson et al 2006b). Of direct relevance in the WMO statement are the following:

- Though there is evidence both for and against the existence of a detectable anthropogenic signal in the tropical cyclone climate record to date, no firm conclusion can be made on this point.
- It is likely that some increase in tropical cyclone intensity and rainfall will occur if the climate continues to warm. Model studies and theory project a 3-5% increase in wind speed per degree increase of tropical sea surface temperatures.
- No individual tropical cyclone can be directly attributed to climate change.
- Some studies of the observational record conclude that the reported global increase in tropical cyclone activity is questionable owing to data problems, since tropical cyclone monitoring has improved continuously.
- There is an observed multi-decadal variability of Atlantic hurricanes whose causes, whether natural, anthropogenic or a combination, are currently being debated. This variability makes detecting any long-term trends in tropical cyclone activity difficult.
- Current theories and computer models predict an increase in wind speed and rainfall of tropical cyclones in a warmer climate.
- Recent climate model simulations project a decrease or no change in global tropical cyclone numbers in a warmer climate, although there is low confidence in this projection.

We note that these were consensus views, and no attempt was made to assign likely probabilities to the possible outcomes. The International Panel for Climate Change (IPCC) arrived at similar findings, but also focused on the observed changes in the North Atlantic (IPCC 2007):

- There is observational evidence for an increase of intense tropical cyclone activity in the North Atlantic since about 1970, correlated with increases of tropical sea surface temperatures.
- 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 global numbers of tropical cyclones.

The IPCC also made the following probability assessments on intense tropical cyclone activity:

- Likely that increases have occurred in some regions since 1970;
- More likely than not a human contribution to the observed trend;
- Likely that there will be future trends in tropical cyclone intensity and heavy precipitation associated with ongoing increases of tropical SSTs;
- Less confidence in projections of a global decrease in the numbers of tropical cyclones.

Emanuel (2005) and Webster et al. (2005) showed a clear increase in the more intense Northwest Pacific cyclones (as shown by category 4 and 5 frequency or PDI) since the commencement of the satellite era. These increases have been closely related to concomitant changes in SSTs in this region. On the other hand there are also concerns about the quality of the data (WMO 2006) and there has been little focused research on attributing the changes in this region. For these reasons this report accepts the overall findings of WMO (2006) and IPCC (2007) as they relate to the North Pacific.

One area where there is consensus is on tropical cyclone rainfall. WMO (2007) and IPCC (2007) concur on there being a likely increase in heavy rainfall associated with tropical cyclones, though the actual level of increase is not clear.

The remainder of the attribution section on tropical cyclones concentrates on attribution in the North Atlantic, where the available data and published work enables more detailed attribution analysis compared to other basins.

3.2.4.3 Attribution of North Atlantic Changes

Chapter 2 provides an overall summary of the observed variations and trends in storm frequency, section 3.3.9.6 considers future scenarios, and Holland and Webster (2007) present a detailed analysis of the changes in North Atlantic tropical cyclones, hurricanes and major hurricanes over the past century, together with a critique of the potential attribution mechanisms. Here we examine these changes in terms of the potential causative mechanisms.

3.2.4.3.1 Storm Intensity

There has been no distinct trend in the mean intensity of all storms, hurricanes, or major hurricanes (Chapter 2). Holland and Webster (2007) also found that there has been a marked oscillation in major hurricane proportions, which has no observable trend. The attribution of this oscillation has not been adequately defined, but it is known that it is associated with a similar oscillation in the proportion of hurricanes that develop in low latitudes and thus experience environmental conditions that are more conducive to development into an intense system than those at more poleward locations. The lack of a mean intensity trend or a trend in major hurricane proportions is in agreement with modeling and theoretical studies that predict a relatively small increase of around 1 to 7% for the observed 0.5 to 0.7^{o}C trend in tropical North Atlantic SSTs (Henderson-Sellers et al 1998, Knutson et al 1998; 2001; Knutson and Tuleya 2004; 2007).

Multidecadal increases of maximum intensity due to multidecadal increases of SST may play a relatively small role in increases of overall hurricane activity, and increases in frequency (discussed in the next section), for which variations in duration due to large-scale circulation changes may be the dominant factors. The relationship between SST, circulation patterns, and hurricane activity variability is not as well understood as the thermodynamic relationships that constrain maximum intensity.

3.2.4.3.2 Storm Frequency and Integrated Activity Measures

Emanuel (2005a; 2007a) examined a Power Dissipation Index (PDI), which combines the frequency, lifetime and intensity, and is related to the cube of the maximum winds summed over the lifetime of the storm. In Chapter 2, it was concluded that there has been a substantial increase in tropical cyclone activity, as measured by the Power Dissipation Index (PDI), since about 1970, strongly correlated with low-frequency variations in tropical Atlantic SSTs. It is likely that hurricane activity (PDI) has increased substantially since the 1950s and 60s in association with warmer Atlantic SSTs. It is also likely that PDI has generally tracked SST variations on multidecadal time scales in the tropical Atlantic since 1950. Holland and Webster (2007) have shown that the PDI changes have arisen from a combination of increasing frequency of tropical cyclones of all categories: tropical storms, hurricanes and major hurricanes; and a multi-decadal oscillation in the proportion of major hurricanes. They found no evidence of a trend in the major hurricane proportions or in overall intensity, but a marked trend in frequency.

While there is a close statistical relationship between low frequency variations of tropical cyclone activity (e.g., the PDI and storm frequency) and SSTs (Ch. 2), this almost certainly arises from a combination of factors, including joint relationships to other atmospheric process that effect cyclone development, such as vertical windshear (Shapiro 1982, Kossin and Vitmer 2007, Goldenberg et al 2001, Shapiro and Goldenberg 1998). It is also notable that the recent SST increases have been associated with a concomitant shift towards increased developments in low latitudes and the eastern Atlantic, regions where the conditions are normally more conducive to cyclogenesis and intensification (Holland and Webster 2007, Ch. 2).

Low-frequency variations in Atlantic tropical cyclone activity have previously been attributed to a natural variability in Atlantic SSTs associated with the Atlantic Multi-decadal Oscillation (Bell and Chelliah 2006, Goldenberg et al. 2001). However, these studies either did not consider the trends over the 20th century in SST (Goldenberg et al. 2001) or did not cover a long enough period to confidently distinguish between oscillatory (internal climate variability) behavior and radiatively forced variations or trends. For example, the multi-decadal AMM2 mode in Bell and Chelliah (2006) first obtains substantial amplitude around 1970. Their circulation-based indices are of insufficient length to determine whether they have a cyclical or trend-like character, or some combination thereof.

While there is undoubtedly a natural variability component to the observed tropical Atlantic SSTs, it is also likely that a discernable warming trend, due to greenhouse gases, has occurred, especially over the past 30-40 years. For example, Santer et al. (2006; see also Gillett et al. 2007) have shown that the observed trends in Atlantic tropical SSTs are unlikely to be caused entirely by internal climate variability, and that the pronounced Atlantic warming since around 1970 that is reproduced in their model is predominantly due to increased greenhouse gases. These conclusions are supported by several other studies that use different methodologies (e.g., Knutson et al. 2006; Trenberth and Shea 2006; Mann and Emanuel 2006; Karoly and Wu 2005). There is also evidence for a detectable greenhouse gas-induced SST increase in the NW Pacific tropical cyclogenesis region (Santer et al. 2006, Gillett et al. 2007; see also Knutson et al. 2006 and Karoly and Wu 2005).

We conclude that there has been an observed SST increase of 0.5-0.7^{o}C over the past century in the main development region for tropical cyclones in the Atlantic. Based on comparison of observed SST trends and corresponding trends in climate models with and without external forcing, it is likely that increased greenhouse gases have caused a discernible increase in SSTs both the North Atlantic and the NW Pacific tropical storm basins over the past 100 yrs and also for the period since about 1950.

Chapter 2 also concludes that it is likely that there has been an increase in tropical storm/hurricane and major hurricane frequency in the North Atlantic over the past century or so, a time during which tropical Atlantic SSTs also increased. Ongoing efforts to reconstruct a complete record of Atlantic tropical cyclone counts back to 1900 or the late 1800s find evidence (in several analyses) for a statistically significant increasing trend since 1900. The evidence is much less compelling for significant positive trends beginning in the late 1800s, although there is increasing uncertainty in the data as one proceeds further back in time. There has not been a significant trend in U.S. landfalling activity since the late 1880s as the overall impacts of the increasing trend in North Atlantic tropical cyclones appear to have been to some extent ameliorated by concomitant shifts into eastern North Atlantic developments, which are less likely to directly affect US coastal regions.

Attribution of these past changes in tropical storm/hurricane activity (e.g., PDI) and frequency to various climate forcings is hampered by the lack of adequate model simulations of tropical cyclone climatologies. In the case of global scale temperature increase formal detection-attribution studies have detected strong evidence for the presence of the space-time pattern of warming expected due to greenhouse gas increases. These studies find that other plausible explanations, such as solar and volcanic forcing together with climate variability alone, fail to explain the observed changes sufficiently. The relatively good agreement between observed and simulated trends based on climate model experiments with estimated past forcings lends substantial confidence to attribution statements for SST. However, since adequate model-based reconstructions of historical tropical cyclone variations are not currently available, we do not have estimates of expected changes in tropical cyclone variations due to a complete representation of the changes in the physical system that would have been caused by greenhouse gas increases and other forcing changes. We therefore must rely on statistical analyses and expert judgement to make attribution assessments. Further discussion of these issues is contained in section 3.3.9.6 (Reconciliation of Future Projections and Past Variations).

The strong relationship with SST—and particularly the large increase in both tropical cyclone activity (PDI) and SST since 1970, along with the observed increases in tropical storm/hurricane and major hurricane frequency and SSTs over the past century or so—provides evidence in support of a discernible impact of anthropogenic forcing on Atlantic tropical cyclone activity. Although there is evidence both for and against this interpretation (e.g., lack of trend in U.S. landfalling hurricanes), the balance of evidence now suggests that human activity has caused a discernible increase in tropical storm, hurricane and major hurricane frequency. It is more difficult to judge whether anthropogenic forcing will cause further increases in activity as the climate continues to warm, since the precise physical reasons for the relationship have not been fully elucidated. It is noted that relevant anthropogenic forcing includes increasing greenhouse gases, as well as changes in aerosol forcing, and possibly decreasing stratospheric ozone and other factors associated with cooling upper atmospheric (~100mb) temperatures in recent decades (Emanuel 2007a).

This assessment is consistent with the IPCC (2007) conclusion that it is more likely than not that there has been a human contribution to the observed increase in intense tropical cyclone activity. It is further supported by several recent related studies, including Trenberth and Shea (2006), Mann and Emanuel (2006), Santer et al (2006), Elsner (2006), Emanuel (2007a), Gillett et al. (2007), Kossin and Vitmer (2007), Vitmer and Kossin (2007), Vecchi and Knutson (2007), and Holland and Webster (2007a).

3.2.4.3.3 Storm lifetime, Track and Extratropical Transition

There has been insufficient work done on the changes, or otherwise, in these important aspects of tropical cyclones to arrive at any firm conclusions.

3.2.5 Extratropical Storms

Chapter 2 documents changes in strong extratropical storms during the twentieth century, especially for oceanic storm track bordering North America. Changes include altered intensity and tracks of intense storms (Wang et al. 2006, Caires and Sterl 2005). Analysis of physical mechanisms is lacking. Natural cycles of large-scale circulation affect variability, through the North Atlantic Oscillation (e.g., Lozano and Swail, 2002, Caires and Sterl 2005) or the related Northern Annular Mode (Hurrell 1995, Ostermeier and Wallace 2003). Changes in sea-surface temperature (Graham and Diaz 2001) and baroclinicity (Fyfe 2003) may also play a role. Analysis of a multi-century GCM simulation by Fischer-Bruns et al. (2005) suggests that changes in solar activity and volcanic activity have negligible influence on strong-storm activity. However, it is likely that anthropogenic influence has contributed to extratropical circulation change during the latter half of the 20th century (Hegerl et al, 2007; see also Gillett et al., 2003, 2005, 2006; Wang et al 2007), which would have influenced storm activity. There is also some evidence that anthropogenic forcing has affected related variables such as geostrophic wind energy and significant wave height (Wang et al 2007) during the latter half of the 20th century, although as with sea-level pressure change, the model simulated response to forcing is not as large as observed. On the other hand, the WASA Group (1998), using long records of station data, suggest that observed changes in storminess in Northern Europe over the latter part of the 20th century are not inconsistent with natural internal low-frequency variability. However, analyses based on direct observations suffer from incomplete spatial and temporal coverage, especially in storm-track regions over adjacent oceans, and generally cover regions that may be too small to allow detection of externally forced signals (Hegerl et al., 2007). Studies of global reanalysis products generally cover less than 50 years. While 50-year records are generally considered adequate for detection and attribution research (Hegerl et al, 2007), a difficulty with reanalysis products is that they are affected by inhomogeneities resulting from changes over time in the type and quantity of data that is available for assimilation (e.g., Trenberth et al. 2005).

A number of investigations have considered the climate controls on the storm intensities or on the decadal trends of wave heights generated by those storms. Most of this attention has been on the North Atlantic, and as noted above the important role of the North Atlantic Oscillation has been recognized (e.g., Neu, 1984; WASA, 1998; Gulev and Grignorieva, 2004). Fewer investigations have examined the climate controls on the storms and waves in the North Pacific, and with less positive conclusions (Graham and Diaz, 2001; Gulev and Grignorieva, 2004). In particular, definite conclusions have not been reached concerning the climate factor producing the progressive increase seen in wave heights, apparently extending at least back to the 1960s. However, Wang et al. (2007) indicate that anthropogenically forced circulation change may have been an important factor in changes of significant wave heights.

A definite control on the wave conditions experienced along the west coast of North America is occurrences of major El Niños such as those in 1982-83 and 1997-98. Both of these events in particular brought extreme wave conditions to south-central California, attributed primarily to the more southerly tracks of the storms compared with non-El Niño years. Allan and Komar (2006) found a correlation between the winter-averaged wave heights measured along the west coast and the multivariate ENSO index (MEI), showing that while the greatest increase during El Niños takes place at the latitudes of south-central California, some increase occurs along the entire west coast, evidence that the storms are stronger as well as having followed more southerly tracks. The wave climates of the west coast therefore have been determined by the decadal increase found by Allan and Komar (2000, 2006), but further enhanced during occurrences of major El Niños.

3.2.6 Convective Storms

Trenberth et al. (2005) point out that since the amount of moisture in the atmosphere is likely to rise much faster as a consequence of rising temperatures than the total precipitation, this should lead to an increase in the intensity of storms, offset by decreases in duration or frequency of events. Environmental conditions that are most likely associated with severe and tornadic thunderstorms have been derived from reanalysis data (Brooks et al. 2003b). Brooks and Dotzek (2007) applied those relationships to count the frequency of favorable environments for significant severe thunderstorms (hail of at least 5 cm diameter, wind gusts of at least 33 m s^{-1}, and/or a tornado of F2 or greater intensity) for the area east of the Rocky Mountains in the US for the period 1958-1999. The count of favorable environments decreased by slightly more than 1% per year from 1958 until the early-to-mid 1970s, and increased by approximately 0.8% per year from then until 1999, so that the frequency was approximately the same at both ends of the analyzed period. They went on to show that the time series of the count of reports of very large hail (7 cm diameter and larger) shows an inflection at about the same time as the inflection in the counts of favorable environments. A comparison of the rate of increase of the two series suggested that the change in environments could account for approximately 7% of the change in reports from the mid-1970s through 1999, with the rest coming from non-meteorological sources.

3.3 Projected Future Changes in Extremes, Their Causes, Mechanisms, and Uncertainties

Projections of future changes of extremes are relying on an increasingly sophisticated set of models and statistical techniques. Studies assessed in this section rely on multi-member ensembles (3 to 5 members) from single models, analyses of multi-model ensembles ranging from 8 to 15 or more AOGCMs, and a perturbed physics ensemble with a single mixed layer model with over 50 members. The discussion here is intended to identify the characteristics of changes of extremes in North America and set in the broader global context.

3.3.1 Temperature

The IPCC Third Assessment Report concluded there was a very likely risk of increased high temperature extremes (and reduced risk of low temperature extremes), with more extreme heat episodes in a future climate. This latter result has been confirmed in subsequent studies (e.g., Yonetani and Gordon, 2001). An ensemble of more recent global simulations projects marked increase in the frequency of very warm daily-temperature minima (Fig. 3.1a). Kharin and Zwiers (2005) show in a single model that future increases in temperature extremes follow increases in mean temperature over most of the world including North America. They show a large reduction in the wintertime cold temperature extremes in regions where snow and sea ice decrease due to changes in the effective heat capacity and albedo of the surface. They also show that summertime warm temperature extremes increase in regions where the soil dries due to a smaller fraction of surface energy used for evaporation. Furthermore, that study showed that in most instances warm-extreme changes are similar in magnitude to the increases in daily maximum temperature, but cold extremes shift to warmer temperatures faster than daily minimum temperatures, though this result is less consistent when model parameters are varied in a perturbed physics ensemble where there are increased daily temperature maxima for nearly the whole land surface. However, the range in magnitude of increases was substantial indicating a sensitivity to model formulations (Clark et al., 2006).

Events that are rare could become more commonplace. Recent studies using both individual models (Kharin and Zwiers, 2005) and an ensemble of models (Wehner 2006, Kharin, et al 2007) show that events that currently reoccurr on average once every 20 years (i.e., have a 5% chance of occurring in a given year) will become signficantly more frequent over North America. For example, by the middle of the 21st century, in simulations of the SRES A1B scenario, the recurrence period (or expected average waiting time) for the current 20-year extreme in daily average surface-air temperature reduces to three years over most of the continental United States and five years over most of Canada (Kharin, et al 2007). By the end of the century (Fig. 3.5a), the average reoccurrence time may further reduce to every other year or less (Wehner, 2006).

Similar behavior occurs for seasonal average temperatures. For example, Weisheimer and Palmer (2005) examined changes in extreme seasonal (DJF and JJA) temperatures in 14 models for 3 scenarios. They showed that by the end of 21st century, the probability of such extreme warm seasons is projected to rise in many areas including North America. Over the North American region, an extreme seasonal temperature event that occurs 1 out of 20 years in the present climate becomes a 1 in 3 year event in the A2 scenario by the end of this century. This result is consistent with that from the perturbed physics ensemble of Clark et al. (2006) where, for nearly all land areas, extreme JJA temperatures were at least 20 times and in some areas 100 times more frequent compared to the control ensemble mean, making these changes greater than the ensemble spread.

Others have examined possible future cold-air outbreaks. Vavrus et al. (2006) analysed 7 AOGCMs run with the A1B scenario, and defined a cold air outbreak as 2 or more consecutive days when the daily temperatures were at least 2 standard deviations below the present-day winter-time mean. For a future warmer climate, they documented a decline in frequency of 50 to 100% in NH winter in most areas compared to present-day, with some of the smallest reductions occurring in western North America due to atmospheric circulation changes (blocking and ridging on West Coast) associated with the increase of GHGs.

Several recent studies have addressed explicitly possible future changes in heat waves (very high temperatures over a sustained period of days), and found that in a future climate there is an increased risk of more intense, longer-lasting and more frequent heat waves (Meehl and Tebaldi, 2004; Schär et al., 2004; Clark et al., 2006). Meehl and Tebaldi (2004) related summertime heat waves to circulation patterns in the models and observations. They found that the more intense and frequent summertime heat waves over the southeast and western U.S. were related in part to base state circulation changes due to the increase in GHGs. An additional factor for extreme heat is drier soils in a future warmer climate (Brabson et al., 2005; Clark et al., 2006). The "Heat Index", a measure of the apparent temperature felt by humans that includes moisture influences, was projected in a GFDL model study to increase substantially more than the air temperature in a warming climate in many regions (Delworth et al. 1999). The regions most prone to this effect included humid regions of the tropics and summer hemisphere extratropics, including the Southeast U.S. and Caribbean. A multi-model ensemble showed that simulated heat waves increase during the latter part of the 20th century, and are projected to increase globally and over most regions including North America (Tebaldi et al., 2006), though different model parameters can influence the range in the magnitude of this response (Clark et al., 2006).

Warm episodes in ocean temperatures can stress marine ecosystems, causing impacts such as coral bleaching (e.g., Liu et al. 2006). Key factors appear to be clear skies, low winds and neap tides occurring near annual maximum temperatures since they promote heating with little vertical mixing of warm waters with cooler, deeper layers (Strong et al. 2006). At present, widespread bleaching episodes do not appear to be related to variability such as ENSO cycles (Arzayus and Skirving 2004) or Pacific Decadal Oscillation (Strong et al. 2006). The 2005 Caribbean coral bleaching event has been linked to warm ocean temperatures that appear to have been partially due to long-term warming associated with anthropogenic forcing and not a manifestation of unforced climate variability alone (Donner et al. 2007). Warming trends in the ocean increase the potential for temperatures to exceed thresholds for mass coral bleaching, and thus may greatly increase the frequency of bleaching events in the future, depending on the ability of corals and their symbionts to adapt to increasing water temperatures (see Donner et al. 2007 and references therein).

A decrease in diurnal temperature range in most regions in a future warmer climate was reported in Cubasch et al. (2001) and is substantiated by more recent studies (e.g., Stone and Weaver, 2002), which are assessed in the 2007 IPCC report (Meehl et al. 2007a, Christensen et al. 2007). However, noteworthy departures from this tendency have been found in the western portion of the US (particularly the Southwest), where increased diurnal temperature ranges occur in several regional (e.g., Bell et al. 2004, Leung et al. 2004) and global (Christensen et al., 2007) climate-change simulations. Increased diurnal temperature range often occurs in areas that experience drying in the summer.

3.3.2 Frost

As the mean climate warms, the number of frost days are expected to decrease (Cubasch et al. 2001). Meehl et al (2004a) have shown that there would indeed be decreases in frost days in a future warmer climate in the extratropics, particularly along the northwest coast of North America, with the pattern of the decreases dictated by the changes in atmospheric circulation from the increase in GHGs. Results from a multi-model ensemble show simulated and observed decreases in frost days for the 20th century continuing into the 21st century over North America and most other regions (Meehl et al. 2007a, Fig. 3.1b). By then end of the 21st century, the number of frost days averaged over North America has decreased by about 1 month in the 3 future scenarios considered here.

In both the models and the observations, the number of frost days is decreasing over the 20th century (Fig. 3.1b). This decrease is generally related to warming climate, although the pattern of the warming and pattern of the frost-days changes (Fig. 3.2) are not well correlated. The decrease in the number of frost days per year is biggest in the Rockies and along the west coast of North America. The 21st century frost day pattern of change is similar to the 20th century pattern, just much larger in magnitude. In some places by 2100, the number of frost days decrease by more than 2 months.

These changes would have a large impact on biological activity both positive and negative (See chapter 1 for more discussion). An example of a positive change is that there would be increase in growing season length directly related to the decrease in frost days per year. A negative example is fruit trees, which need a certain number of frost periods per winter season to set their buds. In places, this threshold would no longer be exceeded. Note also that changes in wetness and CO_2 content of the air would also impact the biological changes.

3.3.3 Growing Season Length

A quantity related to frost days in many mid and high latitude areas, particularly in the Northern Hemisphere, is growing season length as defined by Frich et al. (2002), and this has been projected to increase in future climate in most areas (Tebaldi et al., 2006). This result is also shown in a multi-model ensemble where the simulated increase in growing season length in the 20th century continues into the 21st century over North America and most other regions (Meehl et al. 2007a, Fig. 3.1c). The growing season length has increased by about 1 week over the 20th century when averaged over all of North America in the models and observations. By the end of the 21st century, the growing season is on average more than 2 weeks longer than present day. (For more discussion on the reasons these changes are important, see chapter 1)

3.3.4 Snow Cover and Sea Ice

Warming generally leads to reduced snow and ice cover (Meehl et al. 2007a). Reduction in perennial sea ice may be large enough to yield a summertime, ice-free Arctic Ocean by the end of the 21st century (Arzel et al. 2006; Zhang and Walsh 2006). Summer Arctic Ocean ice also may undergo substantial, decadal-scale abrupt changes rather than smooth retreat (Holland et al. 2006). The warming may also produce substantial reduction in the duration of seasonal ice in lakes across Canada and the U.S. (Hodgkins et al. 2002, Gao and Stefan 2004, Williams et al. 2004, Morris et al. 2005) and in rivers (Hodgkins et al. 2003, Huntington et al. 2003). Reduced sea ice in particular, may produce more strong storms over the ocean (Section 3.3.10). Reduced lake ice may alter the occurrence of heavy lake-effect snowfall (Section 3.3.8). The annual cycle of snow cover and river runoff may be substantially altered in western U.S. basins (Miller et al. 2003, Leung et al. 2004), affecting water-resource management and potentially exacerbating the impacts of droughts.

3.3.5 Precipitation

Climate models continue to confirm the earlier results that in a future climate warmed by increasing GHGs, precipitation intensity (i.e., precipitation amount per event) is projected to increase over most regions (Wilby and Wigley, 2002; Kharin and Zwiers, 2005; Meehl et al., 2005a; Barnett et al., 2006), and the increase of precipitation extremes is greater than changes in mean precipitation (Kharin and Zwiers, 2005). Rare events precipitation events could become more commonplace in North America (Wehner, 2006, Kharin et al. 2007). For example, by the middle of the 21st century, in simulations of the SRES A1B scenario, the recurrence period (or expected average waiting time) for the current 20-year extreme in daily total precipitation reduces to between 12 and 15 years over much of North America (Kharin, et al 2007). By the end of the century (Fig. 3.5b), the expected average reoccurrence time may further reduce to every six to eight years (Wehner, 2006, Kharin, et al 2007). Note the area of little change in expected average reoccurrence time in the central United States in Fig. 3.5b.

As discussed in section 3.2.3 of this chapter and in Hegerl et al. (2007), the substantial increase in precipitation extremes is related to the fact that the energy budget of the atmosphere constrains increases of large-scale mean precipitation, but extreme precipitation responds to increases in moisture content and thus the nonlinearities involved with the Clausius-Clapeyron relationship. This behavior means that for a given increase in temperature, increases in extreme precipitation can be relatively larger than the mean precipitation increase (e.g., Allen and Ingram, 2002), so long as the character of the regional circulation does not change substantially (Pall et al., 2007). Additionally, timescale can play a role whereby increases in the frequency of seasonal mean rainfall extremes can be greater than the increases in the frequency of daily extremes (Barnett et al., 2006). The increase of mean and extreme precipitation in various regions has been attributed to contributions from both dynamic (circulation) and thermodynamic (moisture content of the air) processes associated with global warming (Emori and Brown, 2005) although the precipitation mean and variability changes are largely due to the thermodynamic changes over most of North America. Changes in circulation also contribute to the pattern of precipitation intensity changes over northwest and northeast North America (Meehl et al., 2005a). Kharin and Zwiers (2005) showed that changes to both the location and scale of the extreme value distribution produced increases of precipitation extremes substantially greater than increases of annual mean precipitation. An increase in the scale parameter from the gamma distribution represents an increase in precipitation intensity, and various regions such as the Northern Hemisphere land areas in winter showed particularly high values of increased scale parameter (Semenov and Bengtsson, 2002; Watterson and Dix, 2003). Time slice simulations with a higher resolution model (~1°) show similar results using changes in the gamma distribution, namely increased extremes of the hydrological cycle (Voss et al., 2002).

3.3.6 Flooding and Dry Days

Changes in the precipitation extremes have a large impact on both flooding and the number of precipitation free days. The discussion of both is combined because their changes are related, in spite of the apparent contradiction.

A number of studies have noted that increased rainfall intensity may imply increased flooding. McCabe et al. (2001) and Watterson (2005) showed there was an increase in extreme rainfall intensity in extratropical surface lows, particularly over Northern Hemisphere land. However, analyses of climate changes from increased greenhouse gases gives mixed results, with increased or decreased risk of flooding depending on the model analyzed (Arora and Boer 2001,Milly et al. 2002, Voss et al. 2002).

Global and North American averaged time series of the Frich et al. (2002) indices in the multi-model analysis of Tebaldi et al. (2006) show simulated increases in heavy precipitation during the 20th century continuing through the 21st century (Meehl et al. 2007a, Fig. 3.1d), along with a somewhat weaker and less consistent trend for increasing dry periods between rainfall events for all scenarios (Meehl et al. 2007a). Part of the reason for these results is that precipitation intensity increases almost everywhere, but particularly at mid and high latitudes, where mean precipitation increases (Meehl et al., 2005a).

There are regions of increased runs of dry days between precipitation events in the subtropics and lower midlatitudes, but a decreased number of consectutive dry days at higher midlatitudes and high latitudes where mean precipitation increases. Since there are areas of both increases and decreases of consecutive dry days between precipitation events in the multi-model average, the global mean trends are smaller and less consistent across models. Consistency of response in a perturbed physics ensemble with one model shows only limited areas of increased frequency of wet days in July, and a larger range of changes of precipitation extremes relative to the control ensemble mean in contrast to the more consistent response of temperature extremes (discussed above), indicating a less consistent response for preciptitation extremes in general compared to temperature extremes (Barnett et al., 2006).

Associated with the risk of drying is a projected increase in chance of intense precipitation and flooding. Though somewhat counter-intuitive, this is because precipitation is projected to be concentrated into more intense events, with longer periods of little precipitation in between. Therefore, intense and heavy episodic rainfall events with high runoff amounts are interspersed with longer relatively dry periods with increased evapotranspiration, particularly in the subtropics (Frei et al., 1998; Allen and Ingram, 2002; Palmer and Räisänen, 2002; Christensen and Christensen, 2003; Beniston, 2004; Christensen and Christensen, 2004; Pal et al., 2004; Meehl et al., 2005a). However, increases in the frequency of dry days do not necessarily mean a decrease in the frequency of extreme high rainfall events depending on the threshold used to define such events (Barnett et al., 2006). Another aspect of these changes has been related to the mean changes of precipitation, with wet extremes becoming more severe in many areas where mean precipitation increases, and dry extremes becoming more severe where the mean precipitation decreases (Kharin and Zwiers, 2005; Meehl et al., 2005a; Räisänen, 2005a; Barnett et al., 2006). However, analysis of a 53-member perturbed-physics ensemble indicates that the change in the frequency of extreme precipitation at an individual location can be difficult to estimate definitively due to model parameterization uncertainty (Barnett et al., 2006).

3.3.7 Drought

A long-standing result from global coupled models noted in Cubasch et al. (2001) has been a projected increase of summer drying in the midlatitudes in a future warmer climate, with an associated increased risk of drought. The more recent generation of models continues to show this behavior (Burke et al., 2006; Meehl et al., 2006b, 2007a; Rowell and Jones, 2006). For example, Wang (2005) analyzed 15 recent AOGCMs to show that in a future warmer climate, the models simulate summer dryness in most parts of northern subtropics and midlatitudes, but there is a large range in the amplitude of summer dryness across models. Hayhoe et al. (2007) found in an ensemble of AOGCMs an increased frequency of droughts lasting a month or longer in the northeastern U.S. Droughts associated with summer drying could result in regional vegetation die-offs (Breshears et al., 2005) and contribute to an increase in the percentage of land area experiencing drought at any one time. For example, extreme drought increases from 1% of present day land area (by definition) to 30% by the end of the century in the Hadley Centre AOGCM's A2 scenario (Burke et al., 2006). Drier soil conditions can also contribute to more severe heat waves as discussed above (Brabson et al., 2005).

A recent analysis of Milly et al. (2005) shows that several AOGCMs project greatly reduced annual water availability over the southwest US and northern Mexico in the future (Fig. 3.6). In the historical context, this area is subject to very severe and long lasting droughts (Cook et al. 2004). The tree-ring record indicates that the late 20th century was a time of greater than normal water availability. However, the consensus of most climate models is for a reduction of cool season precipitation across the Southwest and northwest Mexico (Christensen et al., 2007). This is consistent with a recent 10-year shift to shorter and weaker winter rainy seasons and an observed northward shift in northwest Pacific winter storm tracks (Yin, 2005). Reduced cool season precipitation promotes drier summer conditions by reducing the amount of soil water available for evapotranspiration in summer.

The model projections of reduced water availability over the southwest US and Mexico in the future needs further study. The uncertainty associated with these projections is related to the ability of models to simulate the precipitation distribution and variability in the present climate and to correctly predict the response to future changes. For example, the uncertainty associated with the ENSO response to climate change (Zelle et al. 2005, Meehl et al. 2007a) also impacts the projections of future water availability in southwest US and northern Mexico (e.g., Meehl and Tebaldi 2007). See Chapter 1 for more discussion on the importance of drought.

3.3.8 Snowfall

Extreme snowfall events could change as a result of both precipitation and temperature change. Although reductions in North American snow depth and areal coverage have been projected (Frei and Gong, 2005; Bell and Sloan, 2006; Déry and Wood, 2006), there appears to be little analysis of changes in extreme snowfall. An assessment of possible future changes in heavy lake-effect snowstorms (Kunkel et al. 2002) from the Laurentian Great Lakes found that surface air temperature increases are likely to be the dominant factor. They examined simulations from 2 different climate models and found that changes in the other factors favorable for heavy snow events were relatively small. In the snowbelts south of Lakes Ontario, Erie and Michigan, warming decreases the frequency of temperatures in the range of -10 ˚C to 0 ˚C that is favorable for heavy lake-effect snowfall. Thus, decreases in event frequency are likely in these areas. However, in the northern, colder snowbelts of the Great Lakes, such as the Upper Peninsula of Michigan, moderate increases in temperature have minor impacts on the frequency of favorable temperatures because in the present climate temperatures are often too cold for very heavy snow; warming makes these days more favorable, balancing the loss of other days that become too warm. Thus, the future frequency of heavy events may change little in the northern snowbelts.

Increased temperature suggests that heavy snow events downwind of the Great Lakes will begin later in the season, and on most lakes end earlier. Also, increased temperature with concomitant increased atmospheric moisture implies that in central and northern Canada, Alaska, and other places cold enough to snow (e.g., high mountains) the intensity of heavy snow events may increase.

3.3.9 Tropical Storms

3.3.9.1 Introduction

In response to future anthropogenic climate warming (IPCC 2001) tropical cyclones could potentially change in a number of important ways, including frequency, intensity, size, duration, tracks, area of genesis or occurrence, precipitation, and storm surge characteristics.

Overarching sources of uncertainty in future projections of hurricanes include uncertainties in future emission scenarios for climatically important radiative forcings, global-scale climate sensitivity to these forcings and the limited capacity of climate models to adequately simulate intense tropical cyclones. The vulnerability to storm surge flooding from future hurricanes will very likely be enhanced to some degree due to continuing global sea level rise associated with anthropogenic warming, modulated by local sea level changes due to other factors such as local land elevation changes and regionally varying sea level rise patterns. These related topics are covered in more detail in other CCSP Synthesis and Assessment Products 2-1, 3-2, and 4-1, or IPCC Fourth Assessment Report chapters on climate sensitivity, future emission scenarios, and sea level rise. An assessment of the state of understading of tropical cyclones and climate change as of 2006 has been prepared by the tropical cyclone community (IWTC VI, 2006; section 3.2.4 of this document). Although not published in the literature as yet, the full summary statement and condensed summary are available online at http://www.wmo.ch/web/arep/arep-home.html.

Future projections of hurricanes will depend upon not only on global mean climate considerations, but also on regional-scale projections of a number of aspects of climate that can potentially affect tropical cyclone behavior. These include:

- The local potential intensity (Emanuel 2005a; 2006a, Holland 1997), which depends on sea surface temperatures, atmospheric temperature and moisture profiles, and near-surface ocean temperature stratification;
- Influences of vertical wind shear, large-scale vorticity, and other circulation features (Gray 1968; 1984; Goldenberg et al. 2001; Bell and Chelliah 2006); and,
- The characteristics of precursor disturbances such as easterly waves and their interaction with the environment (Dunn 1940, Frank and Clarke 1980, Pasch et al 1998, Thorncroft and Hodges 2001).

Details of future projections in regions remote from the tropical storm basin in question may also be important. For example, El Nino fluctuations in the Pacific influence Atlantic basin hurricane activity (Chapter 2, Section 3.2 of this chapter). West African monsoon activity has been correlated with Atlantic hurricane activity (Gray 1990), as have African dust outbreaks (Evans et al. 2006). Zhang and Delworth (2006) show how a warming of the northern tropical Atlantic SST relative to the southern tropical Atlantic produces atmospheric circulation features, such as reduced vertical wind shear of the mean wind field, that are correlated with low-frequency variations in major hurricane activity (Goldenberg et al. 2001).

The high sensitivity of tropical storm and hurricane activity in the Atlantic basin to modest environmental variations suggests the possibility of strong sensitivity of hurricane activity to anthropogenic climate change, though the nature of such changes remains to be determined. Confidence in any future projections of anthropogenic influence on Atlantic hurricanes will depend on the reliability of future projections of the local thermodynamic state (e.g., potential intensity) as well as circulation changes driven by both local and remote influences, as described above. Projected effects of global warming on El Niño remain uncertain (Timmermann, 1999; Zelle et al., 2005; Meehl et al. 2007a). There is climate model-based evidence that the time-mean climate late in the 21st century will be characterized by higher tropical-cyclone potential intensity in most tropical-cyclone regions, and also tend toward having a decreased east-west overturning circulation in the Pacific sector in the 21st century, with likely consequences for vertical wind shear and other characteristics in the tropical Atlantic (Vecchi and Soden 2007).

Even assuming that the climate factors discussed above can be projected accurately, additional uncertainties in hurricane future projections arise from uncertainties in understanding and modeling the response of hurricanes to changing environmental conditions. This is exacerbated by projections that the large-scale conditions for some factors, such as decadal means and seasonal extremes of SSTs, will be well outside the range of historically experienced values. This raises questions of the validity of statistical models trained in the present day climate (Ryan et al. 1992; Royer et al. 1998), thus the emphasis here is placed on physical models and inferences as opposed to statistical methods and extrapolation. Thus, we consider projections based on global and regional nested modeling frameworks as well as more idealized modeling or theoretical frameworks developed specifically for hurricanes. The idealized approaches include potential intensity theories as well as empirical indices which attempt to relate tropical cyclone frequency to large-scale environmental conditions. Global and regional nested models simulate the development and life cycle of tropical storm-like phenomena that are typically much weaker and with a larger spatial scale than observed tropical cyclones. These model storms are identified and tracked using automated storm tracking algorithms, which typically differ in detail between studies but include both intensity and "warm-core" criteria which must be satisfied. Models used for existing studies vary in horizontal resolution, with the low-resolution models having a grid spacing of about 300 km, medium resolution with grid spacing of about 120 km, and high resolution with grid spacing of 20-50 km.

3.3.9.2 Tropical Cyclone Intensity

Henderson-Sellers et al. (1998), in an assessment of tropical cyclones and climate change, concluded that the warming resulting from a doubling of CO_2 would cause the potential intensity of tropical cyclones to remain the same or increase by 10 to 20%. (Their estimate was given in terms of central pressure fall; all other references to intensity in this section will refer to maximum surface winds, except where specifically noted otherwise.) They also noted limitations of the potential intensity theories, such as sea spray influences and ocean interactions. Further studies using a high resolution hurricane prediction model for case studies or idealized experiments under boundary conditions provided from high CO_2 conditions (Knutson et al. 1998; Knutson and Tuleya 1999; 2004; 2007) have provided additional model-based evidence to support these theoretical assessments. For a CO_2-induced tropical SST warming of 1.75C, they found a 14% increase in central pressure fall (Fig. 3.7) and a 6% increase in maximum surface wind or a maximum wind speed sensitivity of about 4% per degree Celsius (Knutson and Tuleya 2007). In a related study, Knutson et al. (2001) demonstrated that inclusion of an interactive ocean in their idealized hurricane model did not significantly affect the percentage increase in hurricane intensity associated with CO_2-induced large-scale SST warming. Caveats to these idealized studies are the simplified climate forcing (CO_2 only versus a mixture of forcings in the real world) and neglect of potentially important factors such as vertical wind shear and changes in tropical cyclone distribution.

Global climate model experiments have historically been performed at resolutions which precluded the simulation of realistic hurricane intensities (e.g., major hurricanes). To date, the highest resolution tropical cyclone/climate change experiment published is that of Oouchi et al. (2006). Under present climate conditions, they simulated tropical cyclones with central pressures as low as about 935 hPa and surface wind speeds as high as about 53 m/sec. Oouchi et al. report a 14% increase in the annual maximum tropical cyclone intensity globally and a 20% increase in the Atlantic, both in response to a greenhouse-warming experiment with global SSTs increasing by about 2.5˚C. A notable aspect of their results is the finding that the occurrence rate of the most intense storms increased despite a large reduction in the global frequency of tropical cyclones. Statistically significant intensity increases in their study were limited to two of six basins

(North Atlantic and South Indian Ocean). Bengtsson et al. (2007) also find a slightly reduced tropical storm frequency in the Atlantic coupled with an increase in the intensities (measured in terms of relative vorticity) of the most intense storms. The latter finding only became apparent at relative high model resolution (~30-40 km grid).

Other studies using comparatively lower resolution models have reported tropical-cyclone intensity results. However, the simulated response of intensity to changes in climate in lower resolution models may not be reliable as they have not been able to simulate the marked difference in achievable tropical-cyclone intensities for different SST levels (e.g., Yoshimura et al. 2006) as documented for observed tropical cyclones (DeMaria and Kaplan 1994; Whitney and Hobgood 1997; Baik and Paek 1998). Given this important caveat, the lower resolution model results for intensity are mixed: Tsutsui (2002) and McDonald et al. (2005) report intensity increases under warmer climate conditions, while Sugi et al. (2002), Bengtsson et al. (2006), and Hasegawa and Emori (2005; western North Pacific only) , and Chauvin et al. (2006; North Atlantic only) found either no increase or a decrease of intensity.

Vecchi and Soden (2007) present maps of projected late 21st century changes in Emanuel's potential intensity, vertical wind shear, vorticity, and mid-tropospheric relative humidity as obtained from the latest (IPCC AR4, 2007) climate models (Fig. 3.8). While their results indicate an increase in potential intensity in most tropical cyclone regions, the Atlantic basin in particular displays a mixture with about two-thirds of the area showing increases and about one-third slight decreases. In some regions, they also found a clear tendency for increased vertical wind shear and reduced mid-tropospheric relative humidity – factors that are detrimental for tropical storm development. In the Gulf of Mexico and closer to the U.S. and Mexican coasts the potential intensity generally increases. The net effect of these composite changes remains to be modelled in detail, although existing global modelling studies (Oouchi et al. 2006; Bengtsson et al. 2007) suggest increases in the intensities and frequencies of the strongest storms. In the Eastern Pacific, the potential intensity is predicted to increase across the entire basin, although the vertical wind shear increases may counteract this to some extent.

A more recent idealized calculation by Emanuel et al. (2006) finds that artificially increasing the modelled potential intensity by 10% leads to a marked increase in the occurrence rate of relatively intense hurricanes (Fig. 3.9a), and to a 65% increase in the PDI. Increasing vertical wind shear by 10% leads to a much smaller decrease in the occurrence rate of relatively intense hurricanes (Fig. 3.9b) and a 12% reduction in the PDI. This suggests that increased potential intensity in a CO_2-warmed climate implies a much larger percentage change in potential destructiveness of storms from wind damage than the percentage change in wind speed itself.

In summary, theory and high-resolution idealized models indicate increasing intensity and frequency of the strongest hurricanes/typhoons in a CO_2-warmed climate. Parts of the Atlantic basin may have small decreases in the upper limit intensity, according to one multi-model study of theoretical potential intensity. Expected changes in tropical cyclone intensity and their confidence is therefore assessed as follows: in the Atlantic and North

Pacific basins, some increase of maximum surface wind speeds of the strongest hurricanes and typhoons is likely. We estimate the likely range for the intensity change (in terms of maximum surface winds) to be +2% to +10% per degree Celsius tropical sea surface warming over most tropical storm regions. This range is based on our subjective judgement that the likely range is from about half to twice the sensitivity found in current hurricane models and theory. Furthermore, the balance of evidence suggests that maximum intensities may decrease in some regions, particularly in parts of the Atlantic basin, even though sea surfaces are expected to warm in all regions.

This assessment assumes that there is no change in geographical distribution of the storms (i.e. the storms move over the same locations, but with a generally warmer climate). On the other hand, there is evidence (Holland and Webster 2007a) that changes in distribution (e.g. tropical-cyclone development occurring more equatorward, or poleward of present day) have historically been associated with large changes in the proportion of major hurricanes. It is uncertain how such distributions will change in the future (see below), but such changes potentially could strongly modify the projections reported here.

3.3.9.3 Tropical Cyclone Frequency and Area of Genesis

In contrast to the case for tropical-cyclone intensity, the existing theoretical frameworks for relating tropical-cyclone frequency to global climate change are relatively less well-developed. Gray (1979) developed empirical relationships that model the geographical variation of tropical-cyclone genesis in the present climate relatively well, but several investigators have cautioned against the use of these relationships in a climate change context (Ryan et al. 1992, Royer et al. 1998). Royer et al. proposed a modified form of the Gray relationships based on a measure of convective rainfall as opposed to SST or oceanic heat content, but this alternative has not been widely tested. They showed that tropical-cyclone frequency results for a future climate scenario depended strongly on whether the modified or unmodified genesis parameter approach was used. More recently, Emanuel and Nolan (2004) and Nolan et al. (2006) have developed a new empirical scheme designed to be more appropriate for climate change application (see also Camargo et al. 2006), but tropical-cyclone frequency/climate change scenarios with this framework have not been published to date.

Vecchi and Soden (2007) have assessed the different components of the Emanuel and Nolan (2004) scheme using outputs from the IPCC AR4 models. Their results suggest that a decrease in tropical cyclone frequency may occur over some parts of the Atlantic basin associated with a SW-NE oriented band of less favorable conditions for tropical cyclogenesis and intensification, including enhanced vertical wind shear, reduced mid-tropospheric relative humidity, and slight decrease in potential intensity. The enhanced vertical shear feature (present in about 14 of 18 models in the Caribbean region) also extends into the main cyclogenesis region of the Eastern Pacific basin. Physically, this projection is related to the weakening of the east-west oriented Walker Circulation in the Pacific region, similar to that occurring during El Nino events. During El Nino conditions in the present-day climate, hurricane activity is reduced, as occurred for example in the latter part of the 2006 season. While this projection may appear at odds with observational evidence for an increase in Atlantic tropical storm counts during the past century (Holland and Webster 2007a; Vecchi and Knutson 2007), there is evidence that this has occurred in conjunction with a regional decreasing trend in storm occurrence and formation rates in the western part of the Caribbean and Gulf of Mexico (Vecchi and Knutson 2007; Holland 2007). Earlier, Knutson and Tuleya (2004) had examined the vertical wind shear of the zonal wind component for a key region of the tropical Atlantic basin using nine different coupled models from the CMIP2+ project. Their analysis showed a slight preference for increased vertical shear under high CO_2 conditions if all of the models are considered, and a somewhat greater preference for increased shear if only the six models with the most realistic present-day simulation of shear in the basin are considered. Note that these studies are based on different sets of models, and that a more idealized future forcing scenario was used in the earlier Knutson and Tuleya study.

Alternative approaches to the empirical analysis of large-scale fields are the global and regional climate simulations, in which the occurrence of model tropical cyclones can be tracked. Beginning with the early studies of Broccoli and Manabe (1990), Haarsma et al. (1993), and Bengtsson et al. (1996), a number of investigators have shown that global models can generate tropical storm-like disturbances in roughly the correct geographical locations with roughly the correct seasonal timing. The annual occurrence rate of these systems can be quite model dependent (Camargo et al. 2005) and is apparently sensitive to various aspects of model physics (e.g., Vitart et al. 2001).

The notion of using global models to simulate the climate change response of tropical cyclone counts is given some support by several studies showing that such models can successfully simulate certain aspects of interannual to interdecadal variability of tropical-cyclone occurrence seen in the real world (Vitart et al. 1997; Carmargo et al. 2005; Vitart and Anderson 2001). A recent regional model dynamical downscaling study (Knutson et al. 2007) with an18 km grid model, and a more idealized modelling approach (Emanuel et al. 2007) both indicate that the increase in hurricane activity in the Atlantic from 1980-2005 can be reproduced in a model using specified SSTs and large-scale historical atmospheric information from reanalyses.

Since tropical storms are relatively rare events and can exhibit large interannual to interdecadal variability, large samples sizes (i.e. many seasons) are typically required to test the significance of any changes in a model simulation against the model's "natural variability".

The most recent future projection results obtained from medium and high resolution (120 km-20 km) GCMs are summarized in Table 3.2. Among these models, the higher resolution ones indicate a consistent signal of fewer tropical cyclones globally in a warmer climate, while two lower resolution models find essentially no change. There are, however, regional variations in the sign of the changes, and these vary substantially between models (Table 3.2). For the North Atlantic in particular, more tropical storms are projected in some models, despite a large reduction globally (Sugi et al. 2002; Oouchi et al. 2006), while fewer Atlantic tropical cyclones are projected by other models (e.g., McDonald et al. 2005; Bengtsson et al. 2007). It is not clear at present how the Sugi et al. (2002) and Oouchi et al. (2006) results for the Atlantic reconciles with the tendency for increased vertical wind shear projected for parts of that basin by most recent models (Vecchi and Soden 2007). For example, Oouchi et al. (2006) did not analyze how Atlantic vertical wind shear changed in their warming experiment. However, their results suggest that a future increase in tropical cyclone frequency in the Atlantic is at least plausible, based on current models. Chauvin et al. (2006) and Emanuel et al. (2007) find, in multi-model experiments, that the sign of the changes in tropical cyclone frequency in the north Atlantic basin depends on the climate model used . All of these results cited here should be treated with some caution, as it is not always clear that these changes are greater than the model's natural variability, or that the natural variability or the tropical-cyclone genesis process are being properly simulated in the models.

From the above summarized results, it is not clear that current models provide a confident assessment of even the sign of change of tropical storm frequency in the Atlantic, East Pacific, or Northwest Pacific basins. From an observational perspective, recent studies (Chapter 2) report that there has been a long term increase in Atlantic tropical-cyclone counts since the late 1800s, although the magnitude and in some cases statistical significance of the trend depends on adjustments for missing storms early in the record.

Based on the above available information, we assess that it is unknown how late 21st century tropical cyclone frequency in the Atlantic and North Pacific basins will change, compared to the historical period (~1950-2006).

3.3.9.4 Tropical Cyclone Precipitation

The notion the tropical cyclone precipitation rates could increase in a warmer climate is based on the hypothesis that moisture convergence into tropical cyclones will be enhanced by the increased column integrated water vapor – with the increased water vapor being extremely like to accompany a warming of tropical SSTs. The increased moisture convergence would then be expected to lead to enhanced precipitation rates. This mechanism has been discussed in the context of extreme precipitation in general by Trenberth (1999), Allen and Ingram (2002), and Emori and Brown (2005). In contrast to the near-storm or storm core precipitation rate, accumulated rainfall at a locality along the storm's path is strongly dependent upon the speed of the storm, and there is little guidance at present on whether any change in this factor is likely in a future warmed climate.

An enhanced near-storm tropical rainfall rate for high CO2 conditions has been simulated, for example, by Knutson and Tuleya (2004, 2007) based on an idealized version of the GFDL hurricane model. The latter study reported an increase of 21.6% for a 1.75˚C tropical SST warming (Fig. 3.10), or about 12% per degree Celsius SST increase. Using a global model, Hasegawa and Emori (2005) found an increase in tropical-cyclone-related precipitation in a warmer climate in the western North Pacific basin, despite a decrease in tropical-cyclone intensity there in their model. Chauvin et al (2006) found a similar result in the North Atlantic in their model, and Yoshimura et al. (2006) found a similar result on a global domain. There are issues with all of these modelling studies as they are of course resolution and thus generally depend on parameterization of much of the rainfall within the grid box. Further there is a tendency towards tropical cyclone rainfall simulations that have a high bias in core rainfall rates (e.g. Marchok et al. 2007). Nevertheless, the consistent result of an increased rainfall with greenhouse warming over a number of models, together with the theoretical expectations that this will occur lends credibility to there being a real trend.

Based on the modeling studies to date, the relatively straightforward proposed physical mechanism, and the observed increases in extremely heavy rainfall in the U.S. (although not established observationally for hurricane-related rainfall (Groismann et al. 2004)) we assess the projections that hurricane related rainfall (per storm) will increase in the 21st century as likely. Note that if the frequency of tropical cyclones decreases, the total rainfall from tropical cyclones may decrease. The expected general magnitude of the change for storm core rainfall rates is about +6% to +18% per degree Celsius increase in tropical sea surface temperature.

3.3.9.5 Tropical Cyclone Size, Duration, Track, Storm Surge, and Regions of Occurrence

In this section, other possible impacts of greenhouse gas induced climate warming on tropical cyclones are briefly assessed. The assessment is highly preliminary and the discussion for these relatively brief owing to the lack of detailed studies on these possible impacts at this time.

Wu and Wang (2004) explored the issue of tropical cyclone track changes in a climate change context. Based on experiments derived from one climate model, they found some evidence for inferred track changes in the NW Pacific, although the pattern of changes was fairly complex.

Concerning storm duration, using an idealized hurricane simulation approach in which the potential intensity of a large sample of Atlantic basin storms with synthetically generated storm tracks was artificially increased by 10%, Emanuel (2006b) found that the average storm lifetime of all storms increased by only 3%, whereas the average duration at hurricane intensity for those storms that attained hurricane intensity increased by 15%. However, in the Atlantic and NE Pacific, future changes in duration are quite uncertain, owing to the uncertainties in formation locations and potential circulation changes mentioned previously.

Few studies have attempted to assess possible future changes in hurricane size. Knutson and Tuleya (1999) noted that the radius of hurricane-force winds increased a few percent in their experiments in which the intensities also increased a few percent.

An important question for regions along the periphery of tropical cyclone basins is whether regions with have never or only infrequently experienced tropical cyclones in recorded history may experience them more frequently in the future owing to climate change. Little guidance is available at present on this important question.

Storm surge depends on many factors, including storm intensity, size and track, local bathymetry and the structure of coastal features such as wetlands and river inlets. Unknowns in storm frequency, tracks, size and future changes to coastal features lead to considerable uncertainty in assessing storm surge changes. However, the high confidence of there being future sea level rise as well as the likely increase of intensity of the strongest hurricanes, leads to an assessment that the potential for storm surge damage (per hurricane) is very likely to increase.

In summary, tropical cyclone size, duration, track and regions of occurrence are important questions that need to be addressed. However, based on available published work and previous assessments, it is unknown how these will change in the future (IWTC-VI 2006). Storm surge damage (per hurricane) is likely to rise.

3.3.9.6 Reconciliation of Future Projections and Past Variations

In this section, we attempt to reconcile the future projections discussed above with the past observed variations in TC activity. The balance of evidence suggests that human activity has caused a discernible increase in tropical storm/hurricane and major hurricane frequency in the North Atlantic. U.S. landfalling hurricane frequency has not increased. However, it is more difficult to judge whether anthropogenic forcing will cause further increases in basin-wide activity as the climate continues to warm, since the precise physical reasons for the observed increases have not been fully elucidated. It is noted that relevant anthropogenic forcing includes increasing greenhouse gases, as well as changes in aerosol forcing, and possibly decreasing stratospheric ozone and other factors associated with cooling upper atmospheric (~100mb) temperatures in recent decades (Emanuel 2007a). A recent modeling study (Knutson et al. 2007) indicates that the increase in hurricane activity in the Atlantic from 1980-2005 can be reproduced using a high-resolution nested regional model downscaling approach. However the various changes in the large-scale atmospheric and SST forcings used to drive their regional model were prescribed from observations.

No published model study has directly simulated a substantial century-scale rise in Atlantic tropical cyclone counts similar to those reported for the observations (e.g., Ch. 2). In fact the 20th century behavior in TC frequency has not yet been documented for existing models. One exception is Bengtsson et al. (2007) who simulate little change in tropical storm frequencies comparing the late 1800s and late 1900s. Given the future regional climate projections arising from the models, including the multi-model consensus increase of vertical wind shear in the IPCC AR4 models (Vecchi and Soden 2007), the substantial variability among existing models of such projected characteristics as Atlantic vertical wind shear and the differing mixtures of climate forcings that may be relevant in the two periods, we anticipate that it would be difficult to confidently extrapolate the strong increasing trend in 20th century storm counts using future consensus projections available from existing models. Nonetheless, a significant trend (or anthropogenic signal, whether trend-like or not) detected in observed tropical cyclone activity and attributed to increasing greenhouse gases could imply that a future increase in tropical cyclone frequency in the Atlantic is much more likely than assessed here.

3.3.10 Extratropical Storms

Scientists have used a variety of methods for diagnosing extratropical storms in GCM projections of future climate. These include sea-level pressure (Lambert and Fyfe 2006), strong surface winds (Fischer-Bruns et al. 2005), lower atmosphere vorticity (Bengtsson et al. 2006) and significant wave heights (Wang et al. 2004; Caires et al. 2006). Consequently, there are no consistent definitions used to diagnose extreme extratropical storms. Some analyses do not, for example, determine events in extreme percentiles but rather consider storms that deepen below a threshold sea-level pressure (e.g., Lambert and Fyfe, 2006), though such thresholds may effectively select the most extreme percentiles.

Wave heights of course indicate strong storms only over oceans, but the strongest extratropical storms typically occur in ocean storm tracks, so all three methods focus on similar regions. Ocean storms in the North Atlantic and North Pacific are relevant for this study because they affect coastal areas and shipping to and from North America. GCMs projecting climate change can supply sea-level pressure and surface winds, but they typically do not compute significant wave heights. Rather, empirical relationships (Wang et al. 2004; Caires et al. 2006) using sea-level pressure anomalies and gradients provide estimates of significant wave heights.

Despite the variety of diagnoses, some consistent changes emerge in analyses of extratropical storms under anthropogenic greenhouse warming. Projections of future climate indicate strong storms will be more frequent (Fig. 3.11; Wang et al. 2004, Fischer-Bruns et al. 2005, Bengtsson et al. 2006, Caires et al. 2006, Lambert and Fyfe 2006, Pinto et al. 2007), though the overall number of storms may decrease. These changes are consistent with observed trends over the last half of the twentieth century (Paciorek et al. 2002). More frequent strong storms may reduce the frequency of all extratropical storms by increasing the stability of the atmosphere (Lambert and Fyfe 2006). Analyses of strong winds (Fischer-Bruns et al. 2005, Pinto et al. 2007), lower atmosphere vorticity (Bengtsson et al. 2006) and significant wave heights (Wang et al. 2004; Caires et al. 2006) from single models suggest increased storm strength in the northeast Atlantic, but this increase is not apparent an analysis using output from multiple GCMs (Lambert and Fyfe 2006). Differences may be due to the focus on cold season behavior in the wind and wave analyses, whereas Lambert and Fyfe's (2006) analysis includes the entire year.

The warming projected for the 21st century is largest in the high latitudes due to a poleward retreat of snow and ice resulting in enhanced warming (Meehl et al. 2007a). Projected seasonal changes in sea ice extent show summertime ice area declining much more rapidly than wintertime ice area and that sea ice thins largest where it is initially the thickest, which is consistent with observed sea ice thinning in the late 20th century (Meehl et al. 2007a). Increased storm strength the northeast Atlantic found by some may be linked to the poleward retreat of arctic ice (Fischer-Bruns et al. 2005) and a tendency toward less frequent blocking and more frequent positive phase of the Northern Annular mode (Pinto et al. 2007), though further analysis is needed to diagnose physical associations with ice line, atmospheric temperature and pressure structures and storm behavior. Whether or not storm strength increases, the retreat of sea ice together with changing sea levels will likely increase the exposure of arctic coastlines to damaging waves and erosion produced by strong storms (Lynch et al. 2004, Brunner et al. 2004, Cassano et al. 2006), continuing an observed trend of increasing coastal erosion in arctic Alaska (Mars and Houseknecht, 2007). Rising sea levels, of course, may expose all coastlines to more extreme wave heights (e.g., Cayan et al., 2007).

3.3.11 Convective Storms

Conclusions about possible changes in convective precipitating storms (CPSs) and associated severe-weather hazards under elevated greenhouse gas concentrations have remained elusive. Perhaps the most important reason for this is the mesoscale (10s of km) and smaller dynamics that control behavior of these storms, particularly the initiation of storms. Marsh et al. (2007) and Trapp et al. (2007) have evaluated changes in the frequency of environments that are favorable for severe thunderstorms in GCM simulations of greenhouse-enhanced climates. In both cases, increases in the frequency of environments favorable to severe thunderstorms are seen, but the absence of the mesoscale details in the models means that the results are preliminary. Nevertheless, the approach and the use of nested models within the GCMs show promise for yielding estimates of changes in extreme convective storms.

Chapter 3 References

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Table 3.1 Models and scenarios used for computing the Frich et al. (2002) indices for North America that appear in this document.

Scenario	**Models**
SRES A1B	ccsm3.0 cnrm gfdl2.0 gfdl2.1 inmcm3 ipsl miroc3_2_medres miroc3_2_hires mri_cgcm2_3_2a
SRES A2	cnrm gfd2.0 gfdl2.1 inmcm3 ipsl miroc3_2_medre mri_cgcm2_3_2a
SRES B1	ccsm3.0 cnrm gfdl2.0 gfdl2.1 inmcm3 ipsl miroc3_2_medres miroc3_2_hires

Table 3.2 Summary of tropical storm frequency, expressed as a percent of present day levels, as simulated by several climate GCMs under global warming conditions.

Reference	Model	Resolution	Experiment	Global	N Atlantic	NW Pacific	NE Pacific
Sugi et al. 2002	JMA timeslice	T106 L21 (~120km)	10y 1xCO2, 2xCO2	*66*	**161**	*34*	33
Tsutsui 2002	NCAR CCM2	T42 L18	10y 1xCO2 2xCO2 from 115y CO2 1% pa	102	86	111	91
McDonald et al. 2005	HadAM3 timeslice	N144 L30 (~100km)	15y IS95a 1979-1994 2082-2097	*94*	*75*	*70*	**180**
Hasegawa and Emori 2005	CCSR/NIES /FRCGC timeslice	T106 L56 (~120km)	5x20y at 1xCO2 7x20y at 2xCO2			96	
Yoshimura et al. 2006	JMA timeslice	T106 L21 (~120km)	10y 1xCO2, 2xCO2	*85*			
Bengtsson et al. 2006	ECHAM5-OM	T63 L31 1.5° L40	A1B 3 members 30y 20C and 21C	94			
Oouchi et al. 2006	MRI/JMA timeslice	TL959 L60 (~20km)	10y A1B 1982-1993	*70*	**134**	*62*	*66*

			2080-2099				
Chauvin et al. 2006	ARPEGE-Climat time slice	Stretched non-uniform grid (~50 km)	10y CNRM SRES-B2: Hadley SRES-A2:		 **118** *75*		
Bengtsson et al. 2007	ECHAM5 time slice	up to T319 (down to ~30-40 km grid)	20yr, A1B scenario	---	87	72	107

Bold = significantly **more** tropical storms in the future simulation

Italic = significantly *fewer* tropical storms in the future simulation

Plain text = not significant or significance level not tested

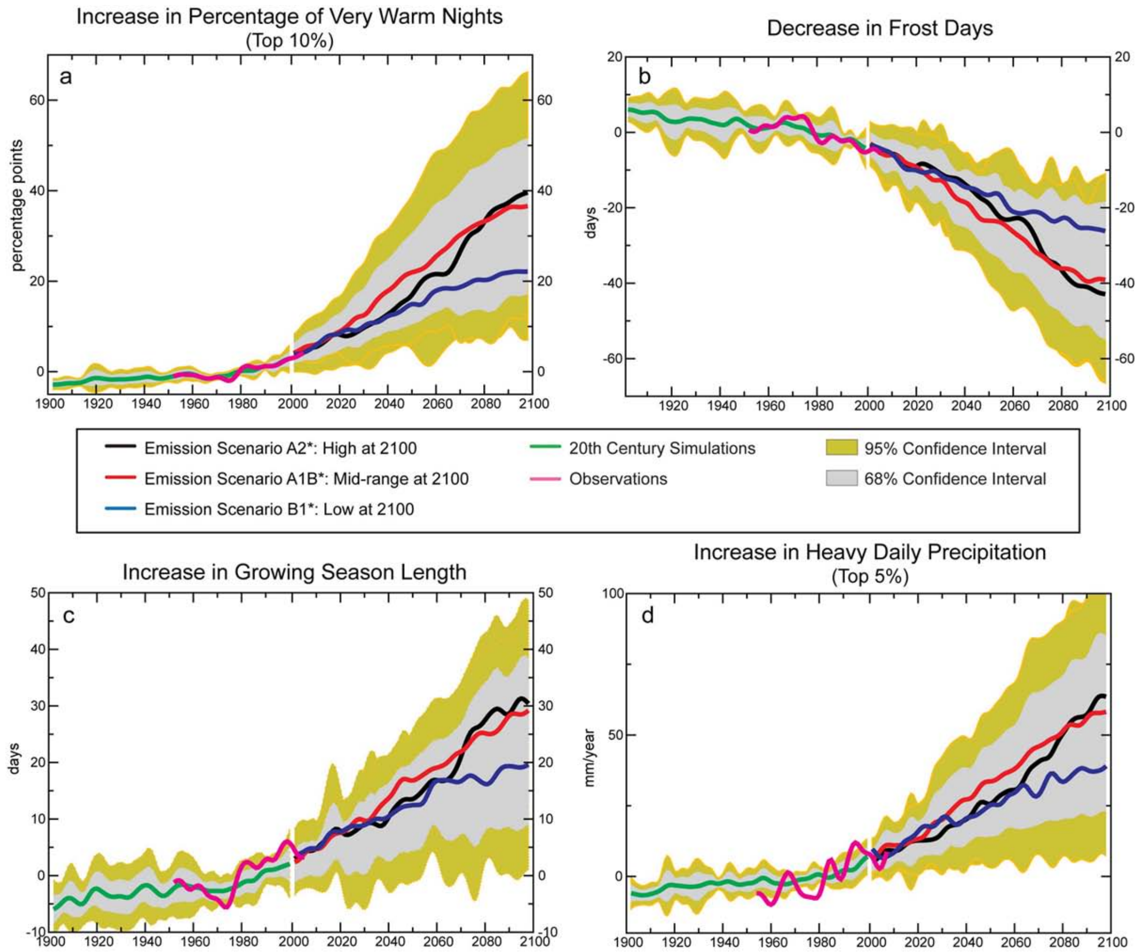


Figure 3.1 Indices (Frich et al. ,2002) averaged over North America for model simulations and observations for the 20th and 21st centuries showing changes relative to 1961-1990 in the a) percentage of days in a year for which daily low temperature is in the top 10% of warm nights for the period 1961-1990, b) number of frost days per year, c) growing season length (days) and d) sum of precipitation on days in the top 5% of heavy precipitation days for the period 1961-1990. In the 20th century, the 1-σ and 2-σ bounds are computed from the ensemble of 20th century simulations. In the future, the bounds are

from an ensemble of simulations that used the A1B, A2 or B1 scenarios[*]. The bounds are the max (or min) standard deviation plus (or minus) signal over all three scenarios. The model plots are obtained from the CMIP-3 multi-model data set at PCMDI and the observations are from Peterson et al. (2007).

[*]3 future emission scenarios from the IPCC Special Report on Emissions Scenarios:
B1 blue line: emissions increase very slowly for a few more decades, then level off and decline
A2 black line: emissions continue to increase rapidly and steadily throughout this century
A1B red line: emissions increase rapidly until 2050 and then decline.
There are more details on these emission scenarios in the glossary.

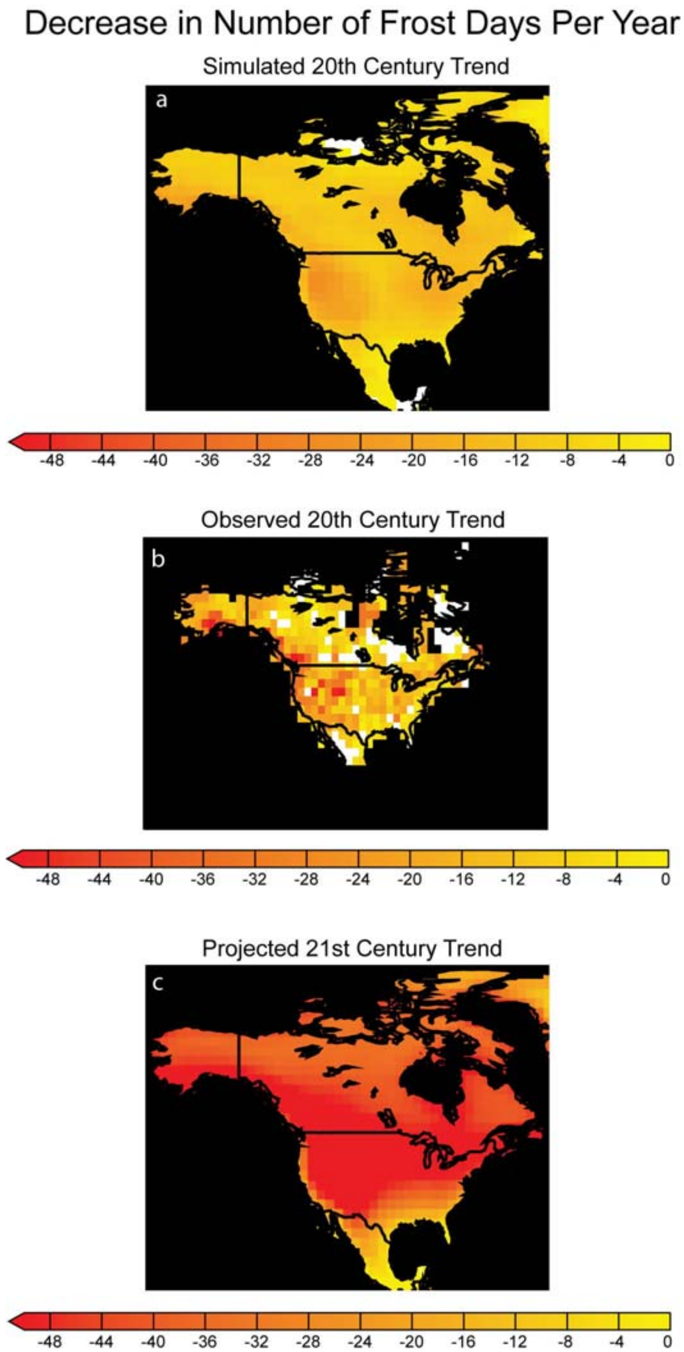


Figure 3.2 Indices (Frich et al., 2002) for frost days over North America for model simulations and observations: a) 20th century trend for model ensemble, b) Observed 20th century trend and c) 21st century trend for emission scenario A2 from model ensemble. The model plots are obtained from the CMIP-3 multi-model data set at PCMDI and the observations are from Peterson et al. (2007).

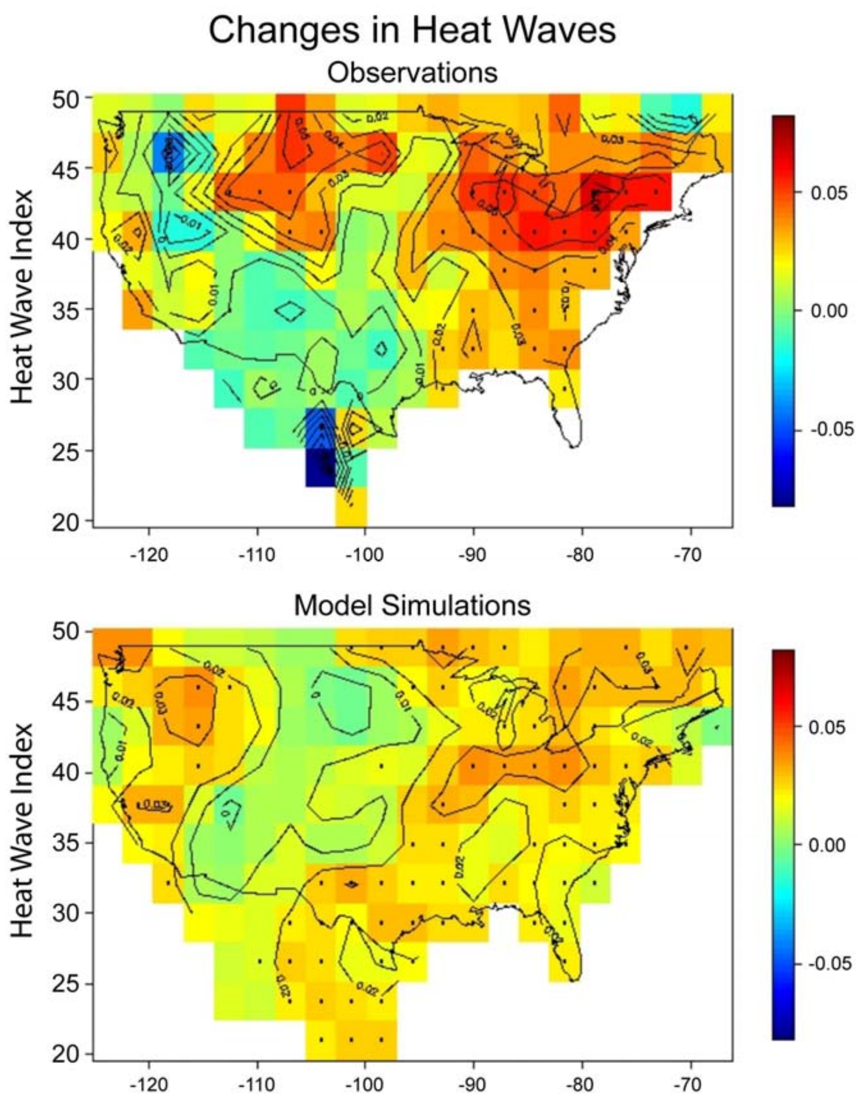


Figure 3.3 Trends in the Karl-Knight heat-wave index (Karl and Knight, 1997) for 1961-1990 in observations (top panel) and in an ensemble of climate simulations by the Parallel Climate Model (bottom panel). Dots mark trends that are significant at the 95% level.

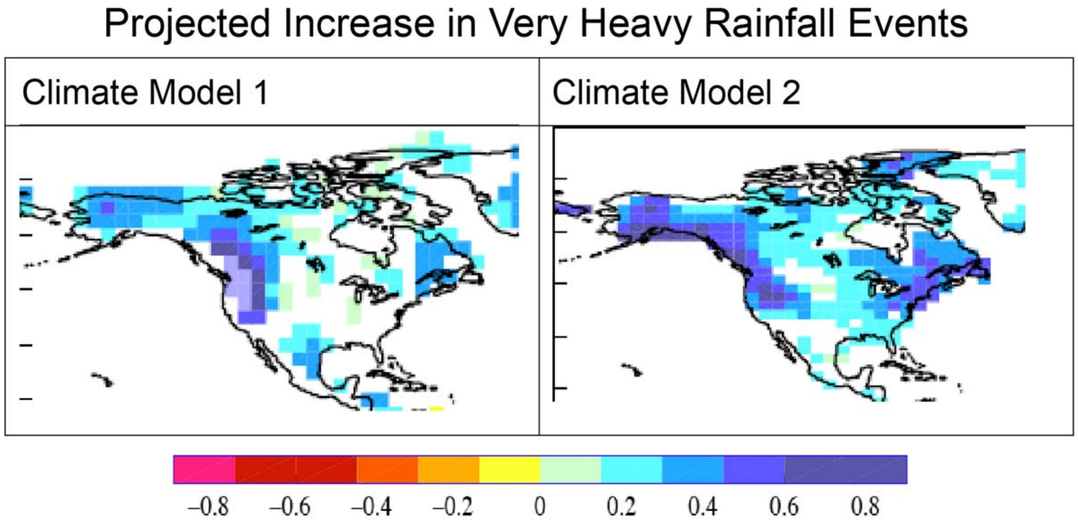


Figure 3.4 Comparison between regions with disproportionate trends in the number of exceedances of the heaviest rainfall events (99.7th percentile) in two climate models at the time of CO2 doubling. See figure 2.8 for areas of N. America which show observed increases in very heavy rainfall Model 1 is the CGCM2 and model 2 is the HadCM3. After Groisman et al. (2005).

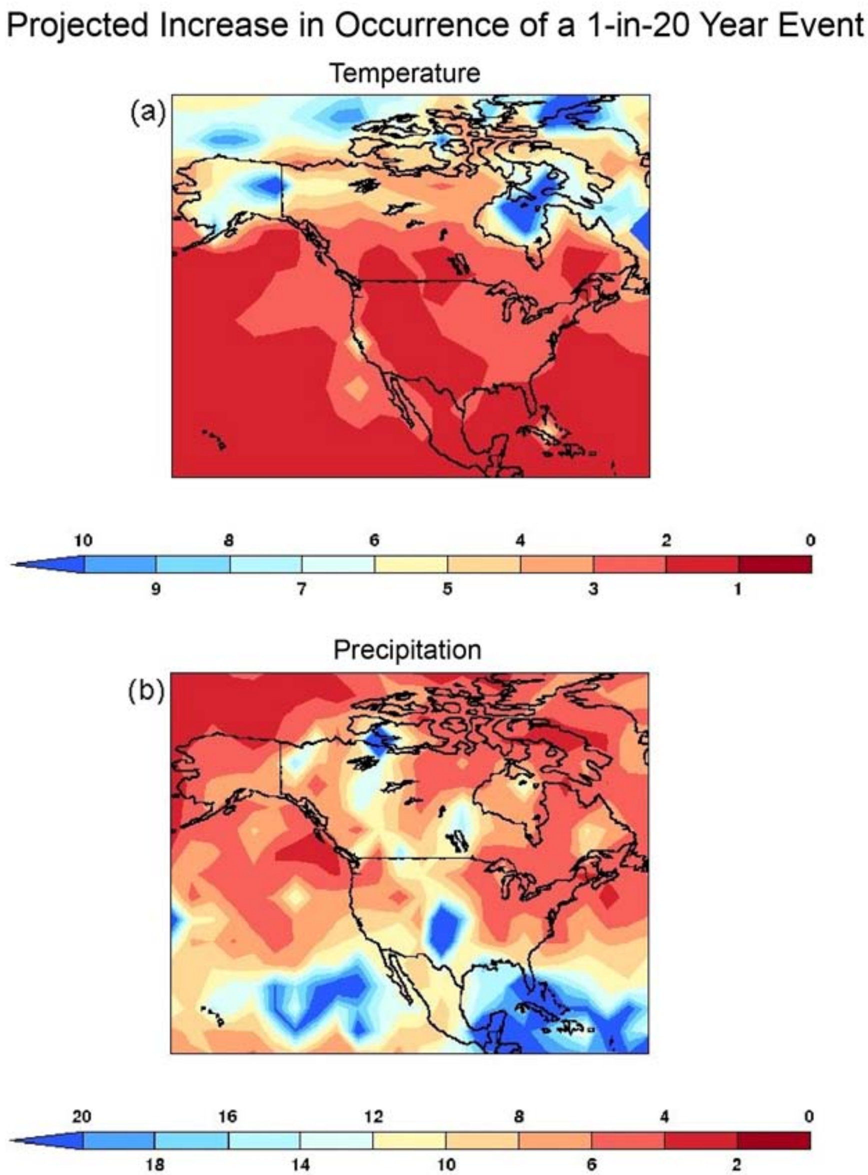


Figure 3.5 Simulations for 2090-2099 indicating how currently rare extremes (a 1-in-20-year event) are projected to become more commonplace. a) Temperature - a day so hot that it is currently experienced once every 20 years would occur every other year or more by the end of the century, (b) daily total precipitation events that occur on average every 20 years in the present climate would, for example, occur once in every 4-6 years for N.E. North America. These results are based on a multi-model ensemble of global climate models simulating the midrange A1B emission scenario*. (from Wehner 2005). [units: years].

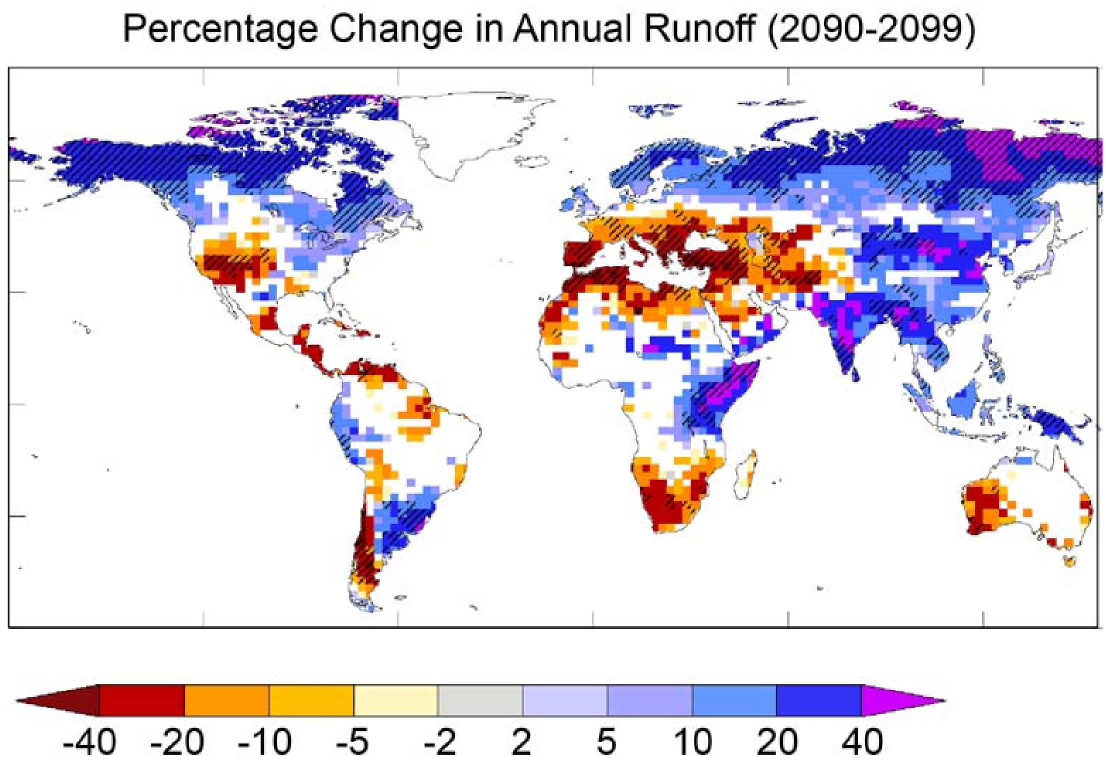


Figure 3.6 Change in annual runoff (%) for the period 2090-2099, relative to 1980-1999. Values are obtained from the median in a multi-model dataset that used the A1B emission scenario. White areas are where less than 66% of the models agree in the sign of change and stippled areas are where more than 90% models agree in the sign of change. [Derived from the analysis of Milly et al. (2005)]

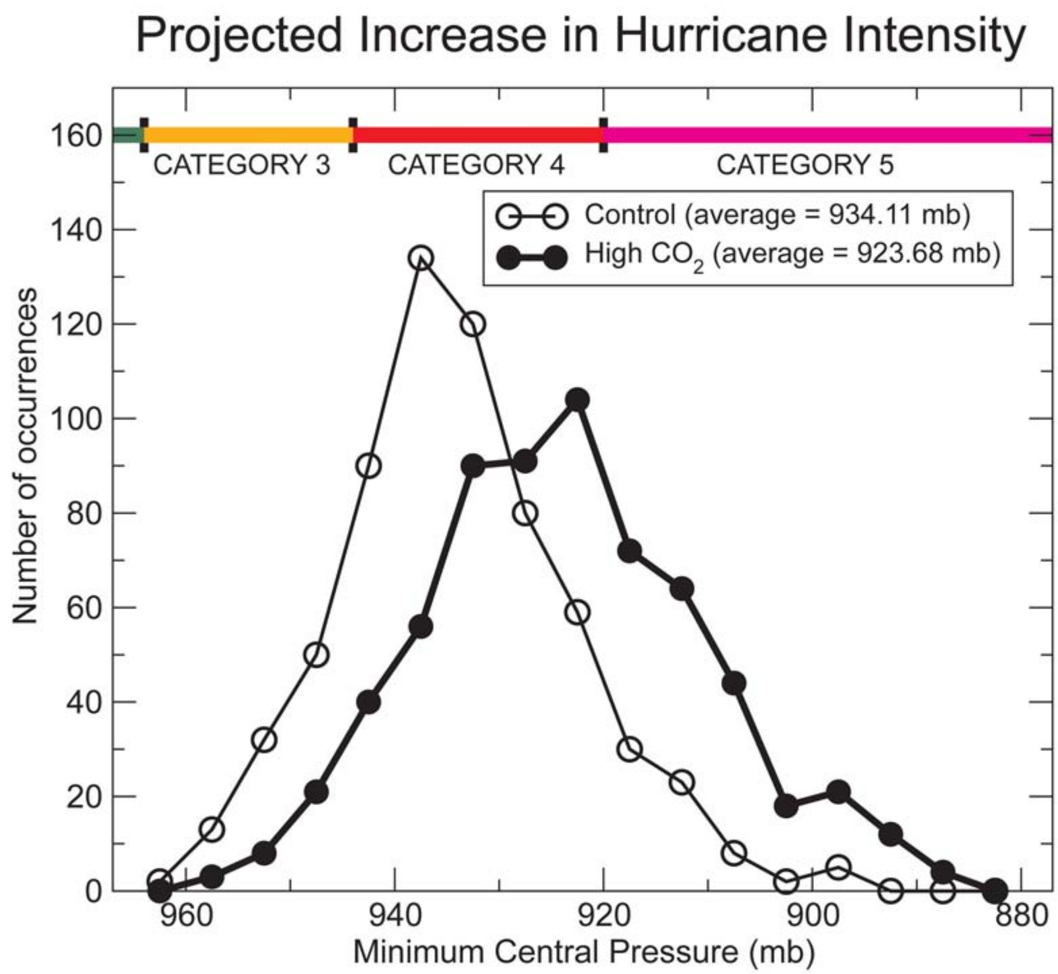


Figure 3.7 Frequency histograms of hurricane intensities in terms of central pressure (mb) aggregated across all idealized hurricane experiments in the Knutson and Tuleya (2004) study. The light curve shows the histogram from the experiments with present-day conditions, while the dark curve is for high CO_2 conditions (after an 80 yr warming trend in a +1%/yr CO_2 experiment). The results indicate that hurricanes in a CO_2-warmed climate will have significantly higher intensities (lower central pressures) than hurricanes in the present climate.

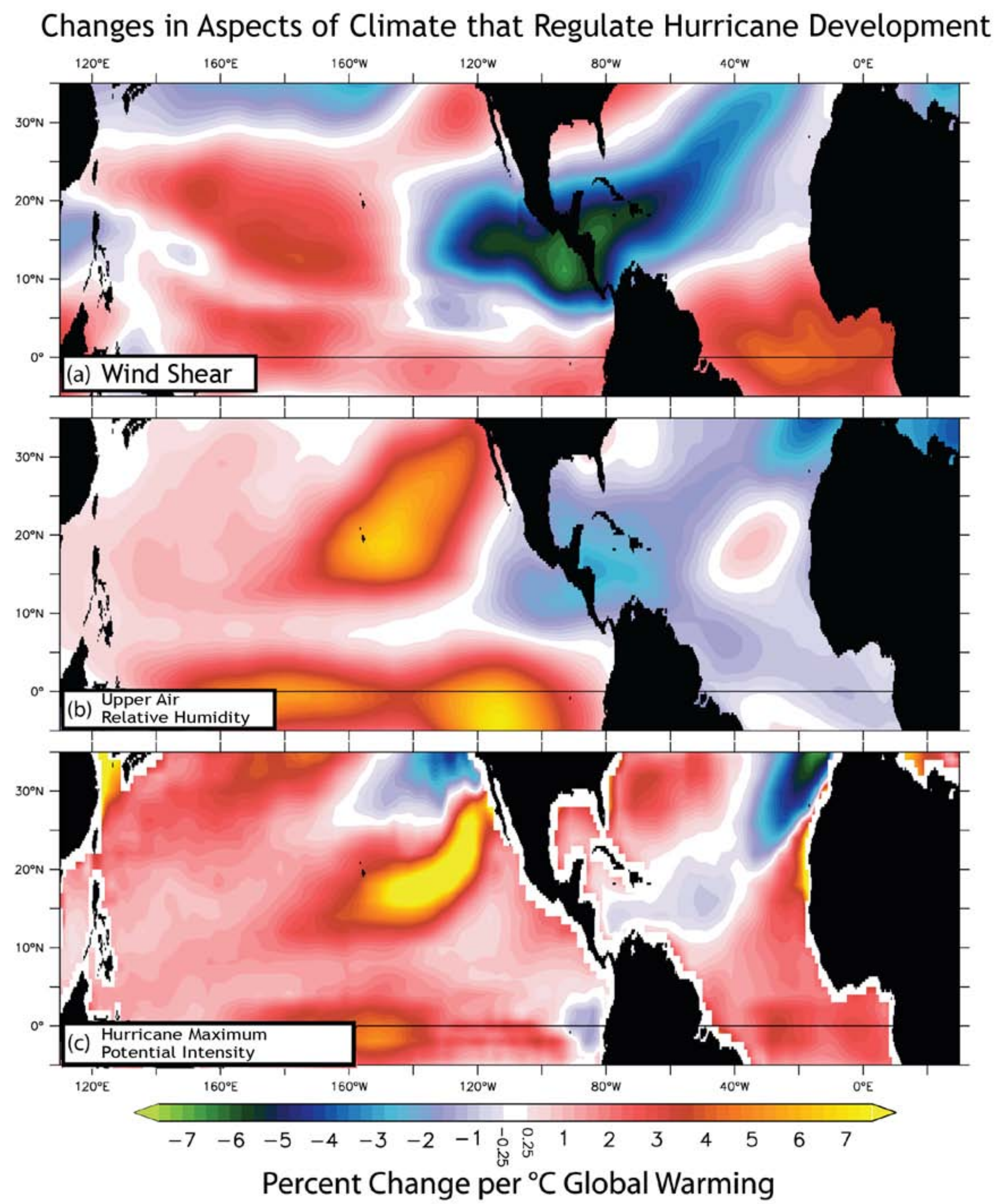


Figure 3.8 Percent changes in June-November ensemble mean a) vertical wind shear, b) mid-tropospheric relative humidity, and c) maximum potential intensity of tropical cyclones for the period 2081-2100 minus the period 2001-2021 for an ensemble of 18

GCMs, available in the IPCC AR4 archive, using the A1B scenario. The percentage changes are normalized by the global surface air temperature increase projected by the models. From Vecchi and Soden (2007).

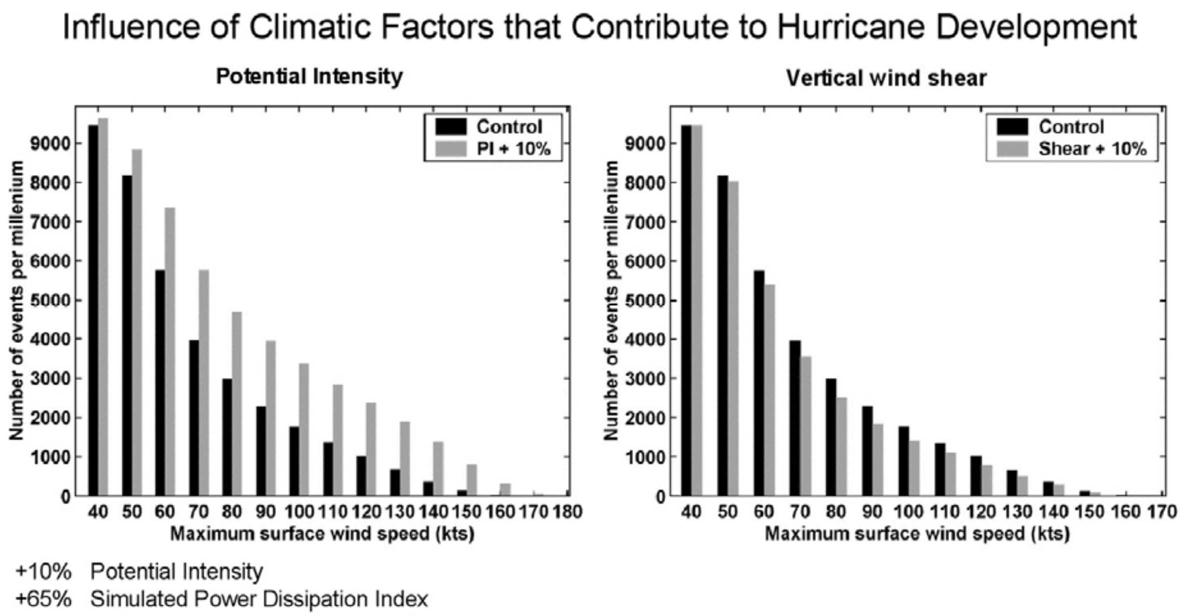


Figure 3.9 Number of events per 1000 yrs with peak wind speeds exceeding the value on the x-axis. Results obtained by running a simple coupled hurricane intensity prediction model over a set of 3000 synthetic storm tracks for the North Atlantic. The grey bars depict storms for present day climate conditions. The black bars depict storms for similar conditions except that the potential intensity (a) or vertical wind shear (b) of the environment is increased everywhere by 10%. From Emanuel (2006).

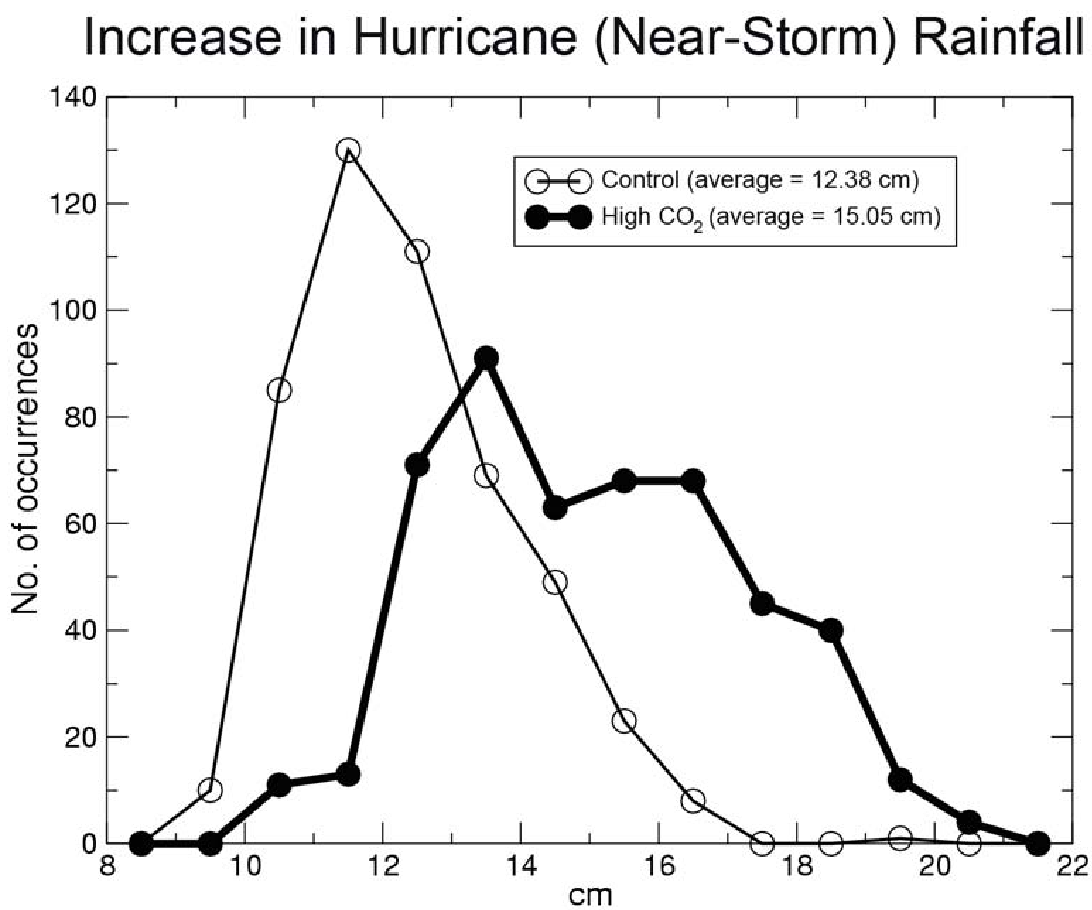
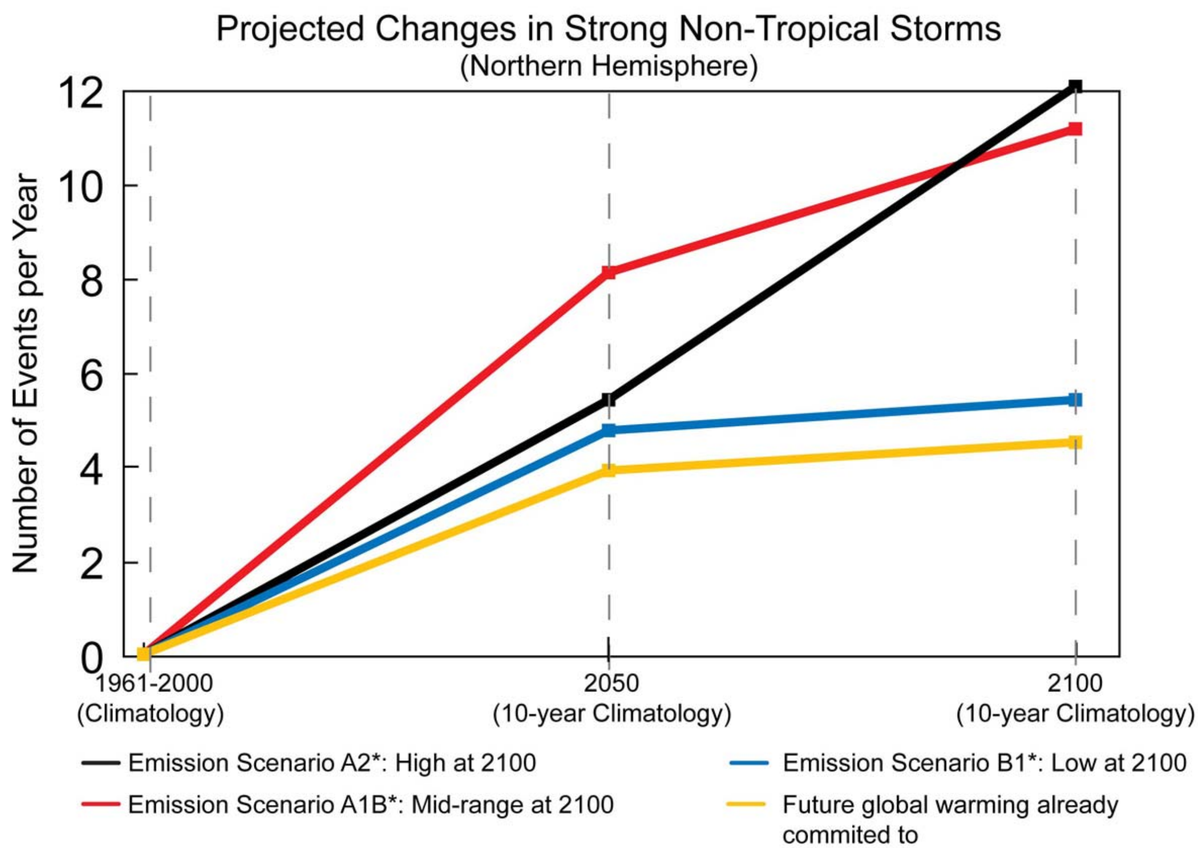


Figure 3.10 As in Fig. 3.7, but for near-hurricane precipitation, estimated as the average precipitation rate for the 102 model grid points (32,700 km^2 area) with highest accumulated rainfall over the last 6 hours of the 5-day idealized hurricane experiments in Knutson and Tuleya (2004). The results indicate that hurricanes in a CO2-warmed climate will have substantially higher core rainfall rates than those in the present climate. (From Knutson and Tuleya, 2007).



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8161 **Figure 3.11** The projected change in intense low pressure systems (strong storms) during
8162 the cold season for the Northern Hemisphere for various emission scenarios* (adapted
8163 from Lambert and Fyfe; 2006). Storms counted have central pressures less than 970 mb
8164 and occur poleward of 30˚N during 120-day season starting November 15. Adapted from
8165 Lambert and Fyfe (2006).

Chapter 4 Recommendations for Improving our Understanding

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In this chapter we provide a set of key recommendations for improving our understanding that stem from the previous three chapters. Many of these findings and recommendations are consistent with previous reports, especially the CCSP 1.1 report on reconciling temperature trends between the surface and free atmosphere.

Many types of extremes, such as excessively hot and cold days, drought, and heavy precipitation show changes over North America consistent with observed warming of the climate. Regarding future changes, model projections show large changes in warm and cold days consistent with projected warming of the climate by the end of the 21^{st} century. However, there remains uncertainty in both observed changes, due to the quality and homogeneity of the observations, and in model projection, due to constraints in model formulation, in a number of other types of climate extremes, including tropical cyclones, extratropical cyclones, tornadoes, and thunderstorms.

4.1 The continued establishment and maintenance of high quality climate observing systems to monitor climate variability and change should be of the highest priority.

Recently, more emphasis has been placed on the development of true climate observing networks that adhere to the Global Climate Observing System (GCOS) Climate Monitoring Principles. This is exemplified by the establishment in the U.S. of the Climate Reference Network, in Canada of the Reference Climate Network, and recent efforts in Mexico to establish a climate observing network. Stations in these networks are carefully sited and instrumented and are designed to be benchmark observing systems adequate to detect the true climate signal for the region being monitored.

Similar efforts to establish a high-quality, global upper-air reference network have been undertaken under the auspices of GCOS. However, this GCOS Reference Upper-air Network (GRUAN) is dependent on the use of current and proposed new observing stations, whose locations will be determined through observing system simulation experiments (OSSEs) that use both climate model simulations and observations to determine where best to locate new observing stations

However, at the present these efforts generally are restricted to a few countries and large areas of the world, even large parts of North America remain under observed. A commitment to developing climate observing networks, especially in areas that traditionally have not had long-term climate observations, is critical for monitoring and detecting future changes in climate, including extremes.

4.2 Efforts to digitize, homogenize and analyze long-term observations in the instrumental record should be expanded.

Research using homogeneity-adjusted observations will provide a better understanding of climate system variability in extremes. Observations of past climate have, by necessity, relied on observations from weather observing networks established for producing and verifying weather forecasts. In order to make use of these datasets in climate analyses, non-climatic changes in the data, such as changes due to station relocations, land use change, instrument changes, and observing practices must be accounted for through data adjustment schemes.

The intent of these data adjustments is to approximate homogeneous time series where the variations are only due to variations in climate and not due to the non-climatic changes discussed above. However, the use of these adjustment schemes introduces another layer of uncertainty into the results of analyses of climate variability and change. Thus, research into both the methods and quantifying uncertainties introduced through use of these methods is critical for understanding observed changes in climate.

Even with the recent efforts to develop true climate observing networks, an understanding of natural and anthropogenic effects on historical weather and climate extremes is best achieved through study of very long (century-scale) records because of the presence of multi-decadal modes of variability in the climate system. For many of the extremes discussed here, including temperature and precipitation extremes, storms, and drought, there are significant challenges in this regard because long-term, high quality, homogeneous records are not available. For example, recent efforts have been made in the U.S. to digitize surface climate data for the 19th Century; however, using these data poses several problems. The density of stations is considerably less than in the 20th Century. Equipment and observational procedures were quite variable and different than the standards established within the U.S. Cooperative Network (COOP). Thus, the raw data are not directly comparable to COOP data. However, initial efforts to homogenize these data have been completed and analysis shows interesting features, including high frequencies of extreme precipitation and low frequencies of heat waves for the 1850-1905 period over the conterminous U.S.

In some cases, heterogeneous records of great length are available and useful information has been extracted. However, there are many opportunities where additional research may result in longer and better records to better characterize the historical variations. For example, the ongoing uncertainty and debate about tropical cyclone trends is rooted in the heterogeneous nature of the observations and different approaches toward approximating homogeneous time series. Therefore, efforts to resolve the existing uncertainties in tropical cyclone frequency and intensity should continue by re-examining the heterogeneous records, and paleotempestological studies should be pursued to better understand variations on multi-century time scales.

4.3 Current weather observing systems should adhere to standards of observation that are consistent with the needs of both the climate and the weather forecasting communities.

Smaller-scale storms, such as thunderstorms and tornadoes are particularly difficult to observe since historical observations have been highly dependent on population density. For example, the U.S. record of tornadoes shows a questionable upward trend that appears to be due mainly to increases in population density in tornado-prone regions. With more people in these regions, tornadoes that may have gone unobserved in earlier parts of the record are now being recorded, thus hampering any analysis of true climate trends of these storms. Since many of the observations of extreme events are collected in support of operational weather forecasting, changes in policies and procedures regarding those observations need to take climate change questions into account, in order to collect high-quality, consistently collected data over time and space. Therefore, consistent standards of collection of data about tornadoes and severe thunderstorms need to be developed and applied. Included in this process is a need for the collection of information about reports that allows users to know the confidence levels that can be applied to reports.

However, in the absence of homogeneous observations of extremes, such as thunderstorms and tornadoes, one promising method to infer changes is through the use of surrogate measures. For example, since the data available to study past trends in these kinds of storms suffer from the problems outlined above an innovative way to study past changes lies in techniques that relate environmental conditions to the occurrence of thunderstorms and tornadoes. Studies along these lines could then produce better relationships, than presently exist, between favorable environments and storms. Those relationships could then be applied to past historical environmental observations and reanalysis data to make improved estimates of long-term trends.

4.4 Efforts to extend reanalysis products using surface observations should be pursued.

Studies of the temporal variations in the frequency of strong extratropical cyclones have typically examined the past 50 years and had to rely on reanalysis fields due to inconsistencies with the historical record. But a much longer period is desirable to gain a better understanding of possible multi-decadal variability in strong storms. There are surface pressure observations extending back to the 19th Century and, although the spatial density of stations decreases backwards in time, it may be possible to identify strong extratropical cyclones and make some deductions about long-term variations. Additionally, efforts to extend reanalysis products back to the early 20th Century using only surface observations have recently begun. These efforts should continue since they provide physically-consistent depictions of climate behavior and will contribute to an understanding of causes of observed changes in climate extremes.

4.5 Research is needed to create annually-resolved, regional-scale reconstructions of the climate for the past 2,000 years.

The development of a wide-array of climate reconstructions for the last two millennia, such as temperature, precipitation, and drought will provide the longer baseline needed to analyze infrequent extreme events, such as those occurring once a century or less. This and other paleoclimatic research can also answer the question of how extremes change when the global climate was warmer and colder than today.

The instrumental record of climate is generally limited to the past 150 years or so. Although there are observations of temperature and precipitation as recorded by thermometers and rain gauges for some locations prior to the early to mid-1800s, they are few and contain problems due to inconsistent observing practices thus their utility is limited. However, the paleoclimate record covering the past 2,000 years and beyond reveals extremes of greater amplitude and longer duration compared to events observed in the instrumental record of the past 100 years (e.g. Woodhouse and Overpeck 1998). The paleoclimate record also reveals that some events occur so infrequently that they may be observed only once, or even not at all during the instrumental period. An improved array of paleo time series is essential to understanding the repeat frequency of rare events, for example events occurring only once a century.

The frequency of some extremes appears tied to the background climate state, according to some paleoclimate records. For example, century-scale changes in the position of the subtropical high may have affected hurricane tracks and the frequency of hurricanes in the Gulf of Mexico (Elsner, et al., 2000). Throughout the western United States, the area exposed to drought may have been elevated for four centuries from 900-1300 AD, according the Palmer Drought Severity Index reconstructed from tree rings (Cook, et al., 2004). The period from 900-1300 AD was a period when the global mean temperature was above average (Mann et al. 1999), consistent with the possibility that changes in the background climate state can affect some extremes. The paleoclimatic record can be used to further understand the possible changes in extremes during warmer and colder climates of the past.

4.6 Research efforts to improve our understanding of the mechanisms that govern hurricane intensity should be increased.

A major limitation of our current knowledge lies in the understanding of hurricane intensity together with surface wind structure and rainfall, and particularly how these relate to a combination of external forcing from the ocean and surrounding atmosphere, and potentially chaotic internal processes. This lack of understanding and related low predictive capacity has been recognized by several expert committees set up in the wake of the disastrous 2005 Atlantic hurricane season:

- The National Science Board recommended that the relevant Federal agencies commit to a major hurricane research program to reduce the impacts of hurricanes and encompassing all aspects of the problem: physical sciences, engineering, social, behavioral, economic and ecological (NSB 2006);
- The NOAA Science Advisory Board established an expert Hurricane Intensity Research Working Group that recommended specific action on hurricane intensity and rainfall prediction (NOAA SAB 2006);

- The American Geophysical Union convened a meeting of scientific experts to produce a white paper recommending action across all science-engineering and community levels (AGU 2006)]; and,
- A group of leading hurricane experts convened several workshops to develop priorities and strategies for addressing the most critical hurricane issues (HiFi 2006).

While much of the focus for these groups was on the short-range forecasting and impacts reduction aspects of hurricanes, the research recommendations also apply to longer term projections. Understanding the manner in which hurricanes respond to their immediate atmospheric and oceanic environment is critical to prediction on all scales.

A critical issue common to all of these expert findings is the need for understanding and parameterization of the complex interactions occurring at the high wind oceanic interface and for very high model resolution in order for forecast models to be able to capture the peak intensity and fluctuations in intensity of major hurricanes. Climate models are arriving at the capacity to resolve regional structures but not relevant details of the hurricane core region. As such, some form of statistical inference will be required to fully assess future intensity projections.

4.7 Substantial increases in computational and human resources should be made available to fully investigate the ability of climate models to recreate the recent past as well as make projections under a variety of future forcing scenarios.

The continued development and improvement of numerical climate models, and observational networks for that matter, is highly related to funding levels of these activities. A key factor, which is often overlooked, is the recruitment and retention of people necessary to perform the analysis of models and observations. For the development and analysis of models, scientists are drawn to institutions with supercomputing resources which require large sources of funding to sustain them. For example, the high resolution global simulations of Oouchi, et al. (2006) to predict future hurricane activity are currently beyond the reach of US tropical cyclone research scientists. This limitation is also true for other smaller-scale storm systems, such as severe thunderstorms and tornadoes. Yet, to understand how these extreme events might change in the future it is critical that climate models are developed that can realistically resolve these types of weather systems. Given sufficient computing resources current U.S. climate models can achieve very high horizontal resolution. Current generation high performance computing (HPC) platforms are also sufficient provided that enough access to computational cycles is made available. Furthermore, many other aspects of the climate system relevant to extreme events, such as extra-tropical cyclones, would be much better simulated in such integrations than they are at typical global model resolutions of today.

Even atmospheric models at ~20 kilometer horizontal resolution are still not finely resolved enough to simulate the high wind speeds and low pressure centers of the most intense hurricanes (Category 5 on the Saffir-Simpson scale). Realistically capturing details of such intense hurricanes, such as the inner eye-wall structure, will require models up to one kilometer horizontal resolution. Such ultra-high resolution global models will require very high computational rates to be viable (Wehner, et al 2006). This is not beyond the reach of next generation HPC platforms but will need significant investments in both model development (human resources) as well as in dedicated computational infrastructure (Randall, 2005).

4.8 Modeling groups should make available high temporal resolution data (daily, hourly) from climate model simulations both of the past and for the future to allow the investigation of potential changes in weather and climate extremes.

In order to achieve high levels of statistical confidence in analyses of climate extremes using methods such as those based on generalized extreme value theory, lengthy stationary datasets are required. Although climate model output is well suited to such analysis, the datasets are often unavailable to the research community at large. Many of the models utilized for the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4) were integrated as ensembles permitting more robust statistical analysis. The simulations were made available at the Program for Climate Model Diagnostics and Intercomparison (PCMDI) at Lawrence Livermore National Laboratory. However, the higher temporal resolution data necessary to analyze extreme events is quite incomplete in the PCMDI database with only four models represented in the daily averaged output sections with ensemble sizes greater than three realizations and many models not represented at all. Lastly, a critical component of this work is the development of enhanced data management and delivery capabilities such as those in the NOAA Operational Model Archive and Distribution System (NOMADS), not only for archive and delivery of model simulations, but for reanalysis and observational data sets as well (NRC 2006).

4.9 Research needs to move beyond purely statistical analysis and focus more on linked physical processes that produce extremes and their changes with climate. Analyses should include attribution of probability distribution changes to natural or anthropogenic influences, comparison of individual events in contemporary and projected climates and the synoptic climatology of extremes and its change in projected climates. The ultimate goal should be a deeper understanding of the physical basis for changes in extremes that improves modeling and thus lends confidence in projected changes.

Literature is lacking that analyzes the physical processes producing extremes and their changes as climate changes. One area that is particularly sparse is analysis of so-called “compound extremes”, events that contain more than one type of extreme such as drought and extremely high temperatures occurring simultaneously.

A substantial body of work has emerged on attribution of changes, with a growing subset dealing with attribution of changes in extremes. Such work shows associations between climate forcing mechanisms and changes in extremes, which is an important first step toward understanding what changes in extremes are attributable to climate change. However, such work typically does not examine the coordinated physical processes linking the extreme behavior to the climate in which it occurs.

More effort should be dedicated to showing how the physical processes producing extremes are changing. Good examples are studies by Meehl and Tebaldi (2004) on severe heat waves, Meehl et al. (2004) on changes in frost days and Meehl and Hu (2006) on megadroughts. Each of these examples involves diagnosing a coherent set of climate-system processes that yield the extreme behavior. An important aspect of the work is demonstrating correspondence between observed and simulated physical processes that yield extremes and, in some of these cases, evaluation of changes in the physical processes in projected climates.

More broadly, the need is for greater analysis of the physical climatology of the climate system leading to extremes. Included in this are further studies of the relationship in projected future climates between slow oscillation modes, such as PDO and AO, and variation in extremes (e.g., Thompson and Wallace 2001). Methods of synoptic climatology (e.g., Cassano et al. 2006, Lynch et al. 2006) could also provide deeper physical insight into the processes producing extremes and their projected changes. Also the development and use of environmental proxies for smaller storm systems such as severe thunderstorms and tornadoes from regional and nested climate models is encouraged. Finally, more probability analysis of the type applied by Stott et al. (2004) to the 2003 European heat wave is needed to determine how much the likelihood of individual extreme events has been altered by human influences on climate.

4.10 Communication between the science community and the user community should be enhanced in both directions.

Because extremes can have major impacts on socio-economical and natural systems, changes in climate extremes will affect the ability of states, provinces and local communities to cope with rare weather events. The process of adaptation to climate change begins with addressing existing vulnerabilities to current and near-term climatic extremes and is directly linked to disaster risk management. Research and experience have shown that mitigating the impacts of extremes and associated complex multiple-stress risks, involve improvements in early warning systems, information for better land-use planning and resource management, building codes, and, coordination of contingency planning for pre- and post-event mitigation and response.

Many adaptations can be implemented at low cost, but comprehensive estimates of adaptation costs and benefits are currently limited partly because detailed information about costs of extreme events are not adequately archived and made available to researchers. To address this problem, guidelines should be developed to improve the methods to collect, archive and quality control detailed information on impacts of extreme events and sequences of extremes, including costs, loss estimates, loss of life, and ecological damage as well as the effectiveness of post event responses. Additionally, networks of systematic observations of key elements of physical, biological, and socio-economical systems affected by climate extremes should be developed, particularly in regions where such networks are already known to be deficient.

Because the links between impacts and changes in extremes can be complex, unexpected and highly nonlinear, especially when modified by human interventions over time, research into these linkages should be strengthened to better understand system vulnerabilities and capacity, to develop a portfolio of best practices, and to implement better response options. But best practices guidelines do not do any good unless they are adequately communicated to the relevant people. Therefore, mechanisms for collaboration and exchange of information among climate scientists, impacts researchers, decision makers (including resources managers, insurers, emergency officials and planners) and the public should be developed and supported. Such mechanisms would involve multi-way information exchange systems and pathways. Better communication between these groups would help communities and individuals make the most appropriate responses to changing extremes. As climate changes, making the complexities of climate risk management explicit can transform event to event response into a learning process for informed proactive management. In such learning-by-doing approaches, the base of knowledge is enhanced through the accumulation of practical experience for risk scenario development and disaster mitigation and preparedness.

4.11 Summary

Figure 4.1 shows the complex interrelationships between the different sections and recommendations in this chapter. Enhanced observing systems and data sets allow better analyses of the observed climate record for patterns of observed variability and change. This provides information for the climate modeling community to verify that their models produce realistic simulations of the observed record, providing increased confidence in simulations of future climate. Both of these activities help improve our physical understanding of the climate which, linked with model simulations through observing system simulation experiments (OSSEs), helps understand where we need better observations, and leads to better formulation of model physics through process studies of observations. This link between observed and modeling patterns of climate change also provides the basis for establishing the cause and effect relationships critical for attribution of climate change to human activities. Since the ultimate goal of this assessment is to provide better information to policy and decision makers, a better understanding of the relationships between climate extremes and their impacts is critical information for reducing the vulnerability of societal and natural systems to climate extremes.

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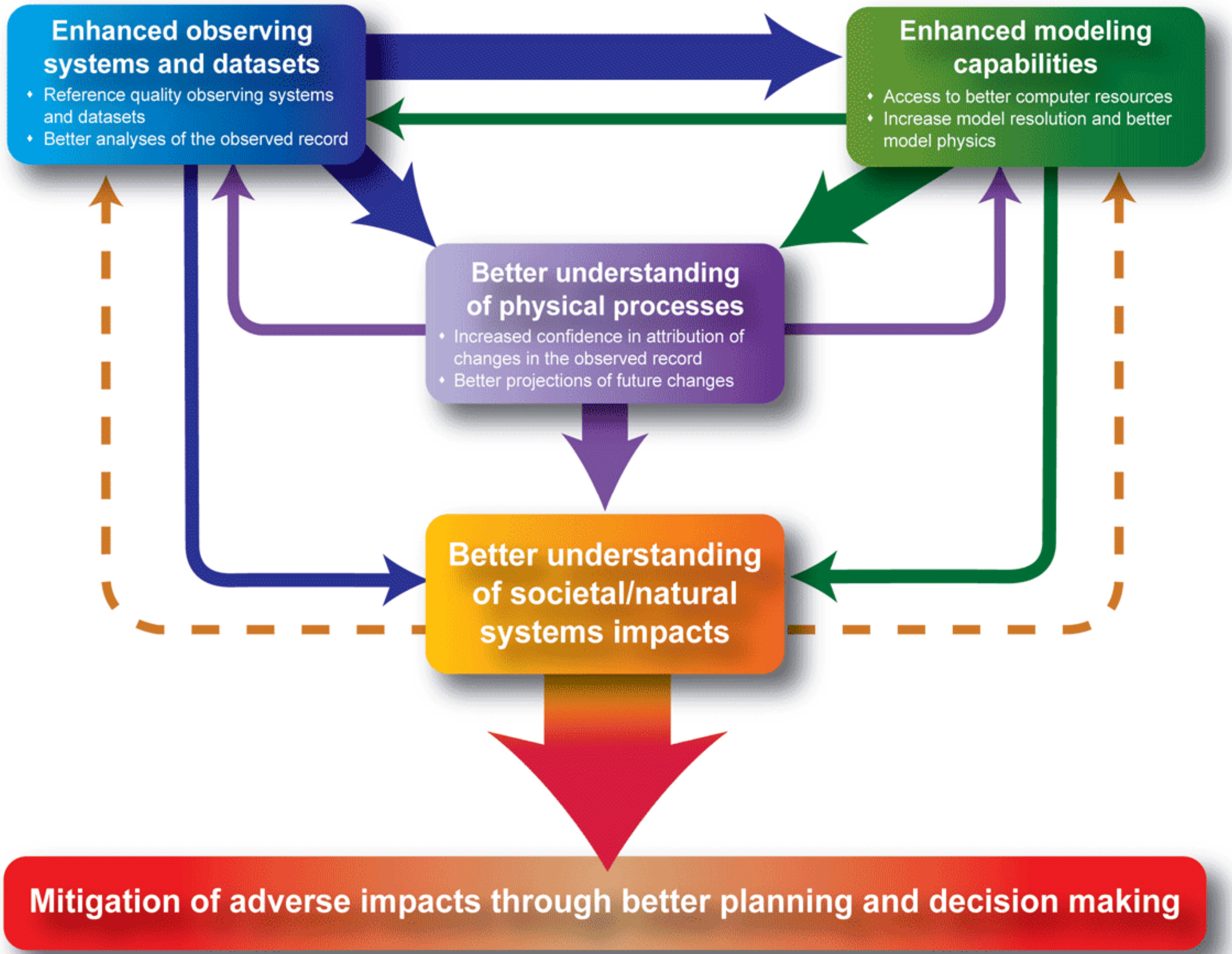

Figure 4.1 Interrelationships between recommendations. Thick arrows indicate major linkages included in this assessment. Better observing systems result in improved analyses which helps improve modeling, physical understanding, and impacts through clearer documentation of observed patterns in climate. Similarly, improved modeling helps improve physical understanding and together can point to deficiencies in observing systems as well as helping to understand future impacts. Lastly, a better understanding of the relationships between climate extremes and impacts can help improve observations by identifying deficiencies in observations (e.g. under-observed areas), and improve modeling efforts by identifying specific needs from model simulations for use in impacts studies.

Appendix A Statistical Trend Analysis

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In many places in this report, but especially Chapter 2, trends have been calculated, either based directly on some climatic variable of interest (e.g. hurricane or cyclone counts) or from some index of extreme climate events. Statistical methods are used in determining the form of a trend, estimating the trend itself along with some measure of uncertainty (e.g. a standard error), and in determining the statistical significance of a trend. A broad-based introduction to these concepts has been given by Wigley (2006). The present review extends Wigley's by introducing some of the more advanced statistical methods that involve time series analysis.

Some initial comments are appropriate about the purpose, and also the limitations, of statistical trend estimation. Real data rarely conform exactly to any statistical model, such as a normal distribution. Where there are trends, they may take many forms. For example, a trend may appear to follow a quadratic or exponential curve rather than a straight line, or it may appear to be superimposed on some cyclic behavior, or there may be sudden jumps (also called changepoints) as well or instead of a steadily increasing or decreasing trend. In these cases, assuming a simple linear trend (equation (1) below) may be misleading. However, the slope of a linear trend can still represent the most compact and convenient method of describing the overall change in some data over a given period of time.

In this appendix, we first outline some of the modern methods of trend estimation that involve estimating a linear or non-linear trend in a correlated time series. Then, the methods are illustrated on a number of examples related to climate and weather extremes.

The basic statistical model for a linear trend can be represented by the equation

(1) $y_t = b_0 + b_1 t + u_t$

where t represents the year, y_t is the data value of interest (e.g. temperature or some climate index in year t), b_0 and b_1 are the intercept and slope of the linear regression, and u_t represents a random error component. The simplest case is when u_t are uncorrelated error terms with mean 0 and a common variance, in which case we typically apply the standard ordinary least squares (OLS) formulas to estimate the intercept and slope, together with their standard errors. Usually the slope (b_1) is interpreted as a trend so this is the primary quantity of interest.

The principal complication with this analysis in the case of climate data is usually that the data are autocorrelated, in other words, the terms cannot be taken as independent. This brings us within the field of statistics known as time series analysis, see e.g. the book by Brockwell and Davis (2002). One common way to deal with this is to assume the values form an autoregressive, moving average process (ARMA for short). The standard ARMA(p,q) process is of the form

(2) $u_t - \varphi_1 u_{t-1} - \ldots - \varphi_p u_{t-p} = \varepsilon_t + \theta_1 \varepsilon_{t-1} + \ldots + \theta_q \varepsilon_{t-q}$

where $\varphi_1 \ldots \varphi_p$ are the autoregressive coefficients, $\theta_1 \ldots \theta_q$ are the moving average coefficients and the ε_t terms are independent with mean 0 and common variance. The orders p and q are sometimes determined empirically or sometimes through more formal model-determination techniques such as the Akaike Information Criterion (AIC) or the Bias-Corrected Akaike Information Criterion (AICC). The autoregressive and moving average coefficients may be determined by one of several estimation algorithms (including maximum likelihood) and the regression coefficients b_0 and b_1 by the algorithm of generalized least squares or GLS. Typically, the GLS estimates are not very different from the OLS estimates that arise when autocorrelation is ignored, but the standard errors can be very different. It is quite common that a trend that appears to be statistically significant when estimated under OLS regression is not statistically significant under GLS regression, because of the larger standard error that is usually though not invariably associated with GLS. This is the main reason why it is important to take autocorrelation into account.

An alternative model which is an extension of (1) is

(3) $y_t = b_0 + b_1 x_{t1} + \ldots + b_k x_{tk} + u_t$

where $x_{t1} \ldots x_{tk}$ are k regression variables (covariates) and $b_1 \ldots b_k$ are the associated coefficients. A simple example is polynomial regression, where $x_{tj} = t^j$ for $j=1,\ldots,k$.

However, a polynomial trend, when used to represent a non-linear trend in a climatic dataset, often has the disadvantage that it behaves unstably at the endpoints, so alternative representations such as cubic splines are usually preferred. These can also be represented in the form of (3) with suitable $x_{t1} \ldots x_{tk}$. As with (1), the u_t terms can be taken as uncorrelated with mean 0 and common variance, in which case OLS regression is again appropriate, but it is also common to consider the u_t as autocorrelated.

There are, by now, several algorithms available that fit these models in a semi-automatic fashion. The book by Davis and Brockwell (2002) includes a CD containing a time series program, ITSM, that among many other features, will fit a model of the form (1) or (3) in which the u_t terms follow an ARMA model as in (2). The orders p and q may be specified by the user or selected automatically via AICC. Alternatively, the statistical language R (R Development Core Team, 2007) contains a function “arima” which allows for fitting these models by exact maximum likelihood. The inputs to the arima function include the time series, the covariates, and the orders p and q. The program calculates maximum likelihood/GLS estimates of the ARMA and regression parameters, together with their standard errors, and various other statistics including AIC. Although R does not contain an automated model-selection procedure, it is straightforward to write a short subroutine that fits the time series model for various values of p and q (for example, all values of p and q between 0 and 10) and then identifies the model with minimum AIC. This method has been routinely used for several of the following analyses.

However, it is not always necessary to search through a large set of ARMA models. In very many cases, the AR(1) model in which p=1, q=0, captures almost all of the autocorrelation, in which case this would be the preferred approach.

In other cases, it may be found that there is cyclic behavior in the data corresponding to large-scale circulation indexes such as the Southern Oscillation Index (SOI – often taken as an indicator of El Niño) or the Atlantic Multidecadal Oscillation (AMO) or the Pacific Decadal Oscillation (PDO). In such cases, an alternative to searching for a high-order ARMA model may be to include SOI, AMO or PDO directly as one of the covariates in (2).

Two other practical features should be noted before we discuss specific examples. First, the methodology we have discussed assumes the observations are normally distributed with constant variances (homoscedastic). Sometimes it is necessary to make some transformation to improve the fit of these assumptions. Common transformations include taking logarithms or square roots. With data in the form of counts (such as hurricanes) a square root transformation is often made, because count data are frequently represented by a Poisson distribution, and for that distribution, a square root transformation is a so-called variance-stabilizing transformation, making the data approximately homoscedastic.

The other practical feature that occurs quite frequently is that the same linear trend may not be apparent through all parts of the data. In that case, it is tempting to select the start and finish points of the time series and recalculate the trend just for that portion of the series. There is a danger in doing this, because in formally testing for the presence of a trend, the calculation of significance levels typically does not allow for the selection of a start and finish point. Thus, the procedure may end up selecting a spurious trend. On the other hand, it is sometimes possible to correct for this effect, for example using a Bonferroni correction procedure. An example of this is given in our analysis of the heatwave index dataset below.

Example 1: Cold Index Data (Section 2.2.1)

The data consist of the "cold index", 1895-2005. A density plot of the data shows that the original data are highly right-skewed, but a cube-root transformation leads to a much more symmetric distribution (Figure A.1).

We therefore proceed to look for trends in the cube root data.

A simple OLS linear regression yields a trend of -.00125 per year, standard error .00068, for which the 2-sided p-value is .067. Recomputing using the minimum-AIC ARMA model yields the optimal values p=q=3, trend -.00118, standard error .00064, p-value .066. In this case, fitting an ARMA model makes very little difference to the result, compared with OLS. By the usual criterion of a .05 significance level, this is not a statistically significant result, but it is close enough that we are justified in concluding there is still some evidence of a downward linear trend. Figure A.2 illustrates the fitted linear trend on the cube root data.

Example 2: Heat Wave Index Data (Section 2.2.1 and Fig. 2.3(a))

This example is more complicated to analyze because of the presence of several outlying values in the 1930s which frustrate any attempt to fit a linear trend to the whole series. However, a density plot of the raw data show that they are very right-skewed, whereas taking natural logarithms makes the data look much more normal (Figure A.3). Therefore, for the rest of this analysis we work with the natural logarithms of the heat wave index.

In this case there is no obvious evidence of a linear trend, either upwards or downwards. However, nonlinear trend fits suggest an oscillating pattern up to about 1960, followed by a steadier upward drift in the last four decades. For example, the solid curve in Figure A.4, which is based on a cubic spline fit with 8 degrees of freedom, fitted by ordinary linear regression, is of this form.

Motivated by this, a linear trend has been fitted by time series regression to the data from 1960-2005 (dashed straight line, Figure A.4). In this case, searching for the best ARMA model by the AIC criterion led to the ARMA(1,1) model being selected. Under this model, the fitted linear trend has a slope of 0.031 per year and a standard error of .0035. This is very highly statistically significant – assuming normally distributed errors, the probability that such a result could have been reached by chance, if there were no trend, is of the order 10^{-18}.

We should comment a little about the justification for choosing the endpoints of the linear trend (in this case, 1960 and 2005) in order to give the best fit to a straight line. The potential objection to this is that it creates a bias associated with multiple testing. Suppose, as an artificial example, we were to conduct 100 hypothesis tests based on some sample of data, with significance level .05. This means that if there were in fact no trend present at all, each of the tests would have a .05 probability of incorrectly concluding that there was a trend. In 100 such tests, we would typically expect about 5 of the tests to lead to the conclusion that there was a trend.

A standard way to deal with this issue is the *Bonferroni correction*. Suppose we still conducted 100 tests, but adjusted the significance level of each test to .05/100=.0005. Then even if no trend were present, the probability that at least one of the tests led to rejecting the null hypothesis would be no more than 100 times .0005, or .05. In other words, with the Bonferroni correction, .05 is still an upper bound on the overall probability that one of the tests falsely rejects the null hypothesis.

In the case under discussion, if we allow for all possible combinations of start and finish dates, given a 111-year series, that makes for 111x110/2=6105 tests. To apply the Bonferroni correction in this case, we should therefore adjust the significance level of the individual tests to .05/6105=.0000082. However, this is still very much larger than 10^{-18} The conclusion is that the statistically significant result cannot be explained away as merely the result of selecting the endpoints of the trend.

This application of the Bonferroni correction is somewhat unusual – it is rare for a trend to be so highly significant that selection effects can be explained away completely, as has been shown here. Usually, we have to make a somewhat more subjective judgment about what are suitable starting and finishing points of the analysis.

Example 3: 1-day Heavy Precipitation Frequencies (Section 2.1.2.2)

In this example we considered the time series of 1-day heavy precipitation frequencies for a 20-year return value. In this case, the density plot for the raw data is not as badly skewed as in the earlier examples (Figure A.5, left plot), but is still improved by taking square roots (Figure A.5, right plot). Therefore, we take square roots in the subsequent analysis.

Looking for linear trends in the whole series from 1895-2005, the overall trend is positive but not statistically significant (Figure A.6). Based on simple linear regression, the estimated slope is .00023 with a standard error of .00012, which just fails to be significant at the 5% level. However, time series analysis identifies an ARMA (5, 3) model, when the estimated slope is still .00023, the standard error rises to .00014, which is again not statistically significant.

However, a similar exploratory analysis to that in Example 2 suggested that a better linear trend could be obtained starting around 1935. To be specific, we have considered the data from 1934-2005. Over this period, time series analysis identifies an ARMA(1,2) model, for which the estimated slope is .00067, standard error .00007, under which a formal test rejects the null hypothesis of no slope with a significance level of the order of 10^{-20} under normal theory assumptions. As with Example 2, an argument based on the Bonferroni correction shows that this is a clearly significant result even allowing for the subjective selection of start and finish points of the trend.

Therefore, our conclusion in this case is that there is an overall positive but not statistically significant trend over the whole series, but the trend post-1934 is much steeper and clearly significant.

Example 4: 90-day Heavy Precipitation Frequencies (Section 2.1.2.3 and Fig. 2.9)

This is a similar example based on the time series of 90-day heavy precipitation frequencies for a 20-year return value. Once again, density plots suggest a square root transformation (the plots look rather similar to Figure A.5 and are not shown here).

After taking square roots, simple linear regression leads to an estimated slope of .00044, standard error .00019, based on the whole data set. Fitting ARMA models with linear trend leads us to identify the ARMA(3,1) as the best model under AIC: in that case the estimated slope becomes .00046 and the standard error actually goes down, to .00009. Therefore, we conclude that the linear trend is highly significant in this case (Figure A.7).

Example 5: Tropical cyclones in the North Atlantic (Section 2.1.3.1)

This analysis is based on historical reconstructions of tropical cyclone counts described in the recent paper of Vecchi and Knutson (2007). We consider two slightly different reconstructions of the data, the “one-encounter” reconstruction in which only one intersection of a ship and storm is required for a storm to be counted as seen, and the "two-encounter" reconstruction that requires two intersections before a storm is counted. We focus particularly on the contrast between trends over the 1878-2005 and 1900-2005 time periods, since before the start of the present analysis, Vecchi and Knutson had identified these two periods as of particular interest.

For 1878-2005, using the one-encounter dataset, we find by ordinary least squares a linear trend of .017 (storms per year), standard error .009, which is not statistically significant. Selecting a time series model by AIC, we identify an ARMA(9,2) model as best (an unusually large order of a time series model in this kind of analysis), which leads to a linear trend estimate of .022, standard error .022, which is clearly not significant.

When the same analysis is repeated from 1900-2005, we find by linear regression a slope of .047, standard error .012, which is significant. Time series analysis now identifies the ARMA(5,3) model as optimal, with a slope of .048, standard error .015, very clearly significant. Thus, the evidence is that there is a statistically significant trend over 1900-2005, though not over 1878-2005.

A comment here is that if the density of the data is plotted as in several earlier examples, this suggests a square root transformation to remove skewness. Of course the numerical values of the slopes are quite different if a linear regression is fitted to square root cyclones counts instead of the raw values, but qualitatively, the results are quite similar to those just cited – significant for 1900-2005, not significant for 1878-2005, after fitting a time series model. We omit the details of this.

The second part of the analysis uses the "two-encounter" data set. In this case, fitting an ordinary least-squares linear trend to the data 1878-2005 yields an estimated slope .014 storms per year, standard error .009, not significant. The time series model (again ARMA(9,2)) leads to estimated slope .018, standard error .021, not significant.

When repeated for 1900-2005, ordinary least-squares regression leads to a slope of .042, standard error .012. The same analysis based on a time series model (ARMA(9,2)) leads to a slope of .045 and a standard error of .021. Although the standard error is much bigger under the time series model, this is still significant with a p-value of about .03.

Example 6: U.S. Landfalling Hurricanes (Section 2.1.3.1)

The final example is a time series of U.S. landfalling hurricanes for 1851-2006 taken from the website http://www.aoml.noaa.gov/hrd/hurdat/ushurrlist18512005-gt.txt. The data consist of annual counts and are all between 0 and 7. In such cases a square root transformation is often performed because this is a variance stabilizing transformation for the Poisson distribution. Therefore, square roots have been taken here.

A linear trend was fitted to the full series and also for the following subseries: 1861-2006, 1871-2006 and so on up to 1921-2006. As in preceding examples, the model fitted was ARMA (p, q) with linear trend, with p and q identified by AIC.

For 1871-2006, the optimal model was AR(4), for which the slope was -.00229, standard error .00089, significant at p=.01.

For 1881-2006, the optimal model was AR(4), for which the slope was -.00212, standard error .00100, significant at p=.03.

For all other cases, the estimated trend was negative but not statistically significant.

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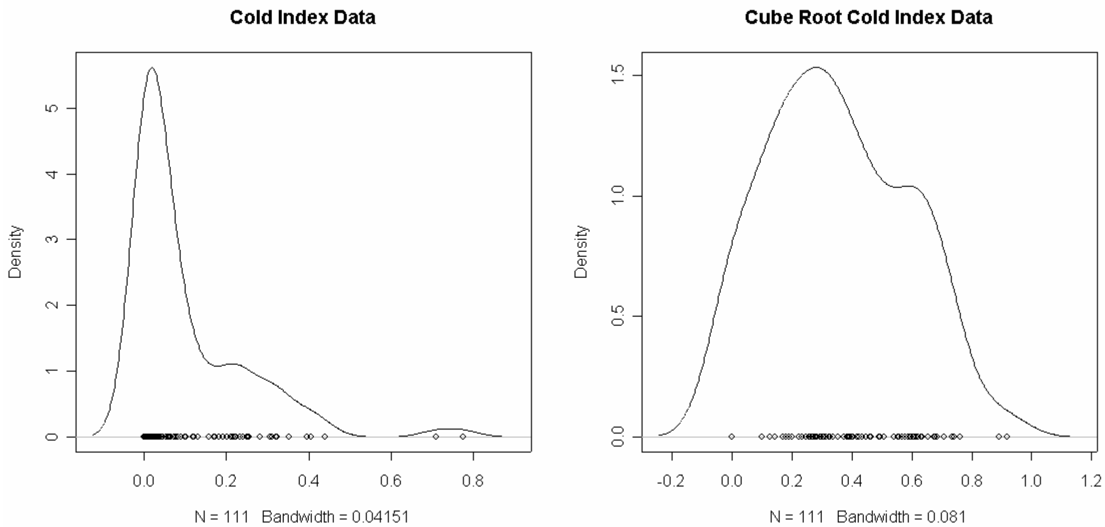


Figure A.1 Density plot for the cold index data (left), and for the cube roots of the same data (right).

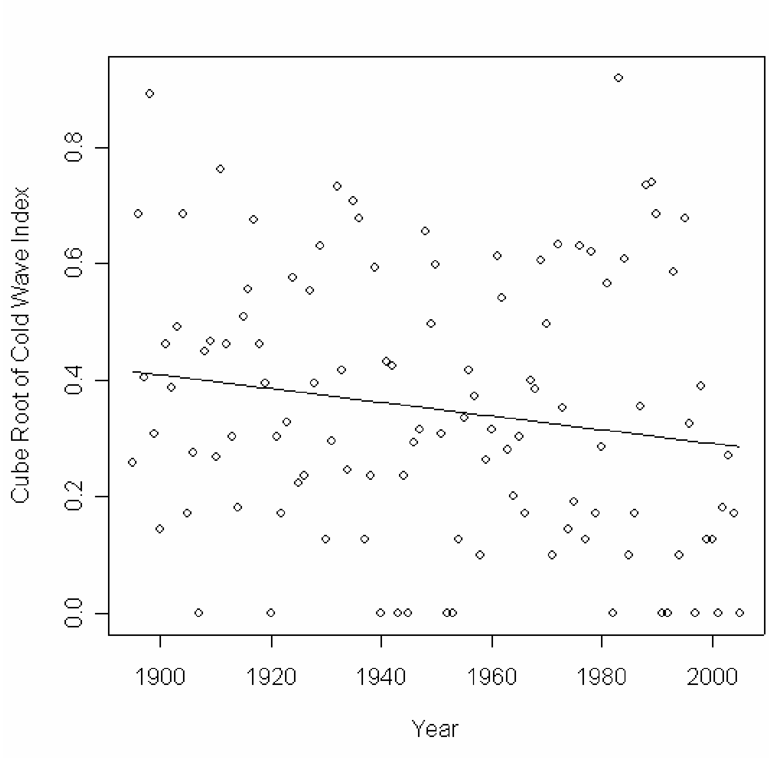


Figure A.2 Cube root of cold wave index with fitted linear trend.

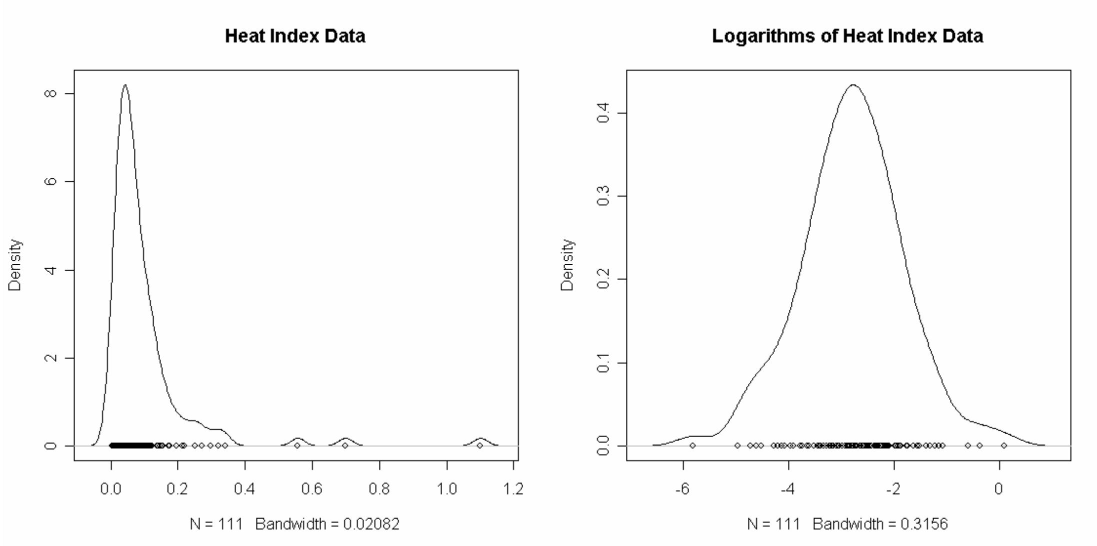


Figure A.3 Density plot for the heat index data (left), and for the natural logarithms of the same data (right).

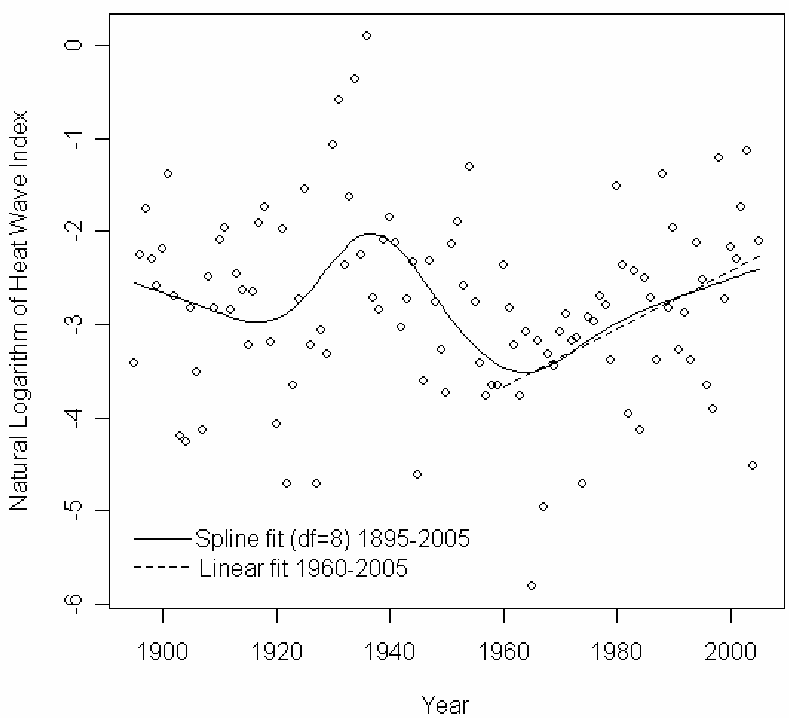
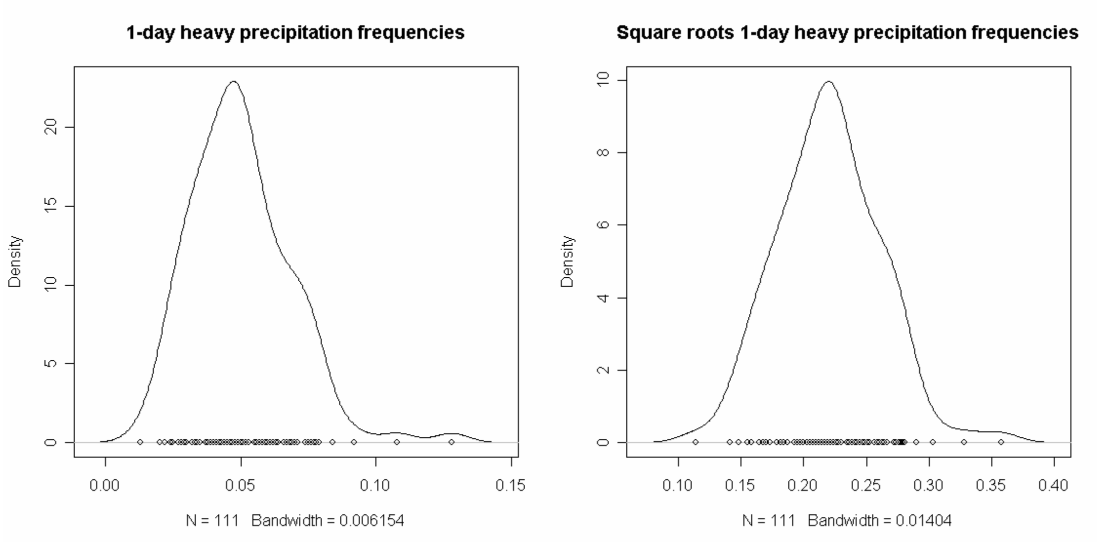


Figure A.4 Trends fitted to natural logarithms of heat index. Solid curve: non-linear spline with 8 degrees of freedom fitted to the whole series. Dashed line: linear trend fitted to data from 1960-2005.



8919
8920

8921 **Figure A.5** Density plot for 1-day heavy precipitation frequencies for a 20-year return
8922 value (left), and for square roots of the same data (right).

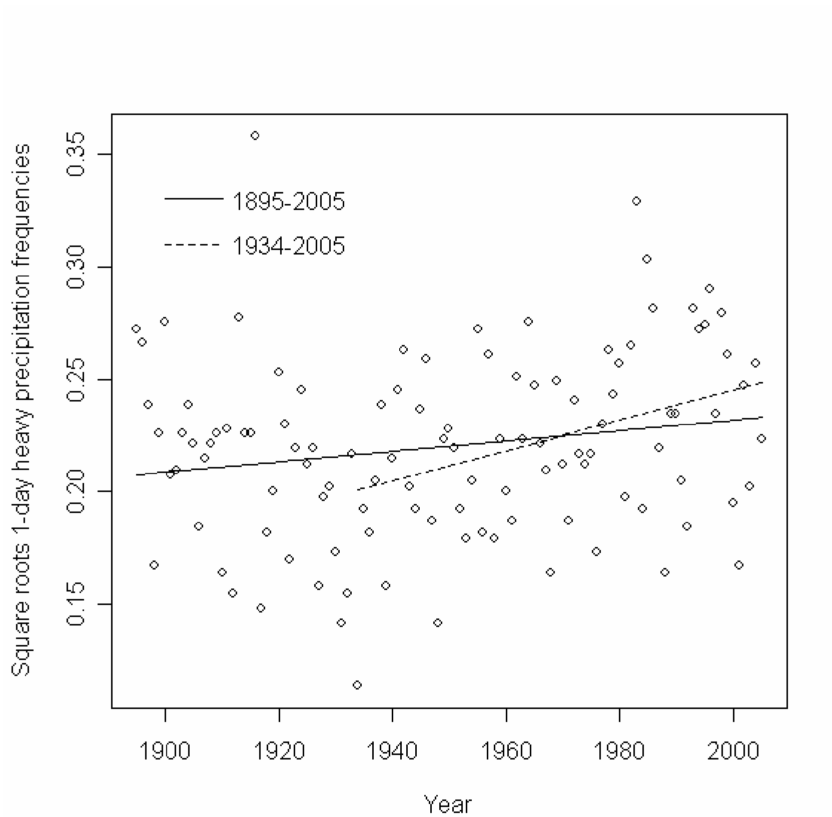


Figure A.6 Trend analysis for the square roots of 1-day heavy precipitation frequencies for a 20-year return value, showing estimated linear trends over 1895-2005 and 1934-2005.

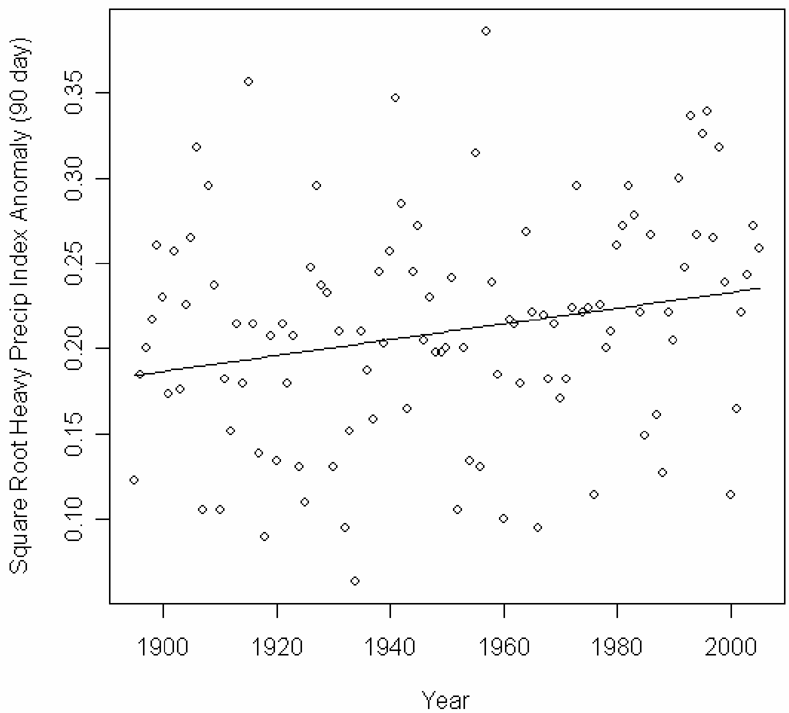


Figure A.7 Trend analysis for the square roots of 90-day heavy precipitation frequencies.