Chapter 6: Paleoclimate

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Table of Contents

Executive Summary

What do pre-Quaternary climates reveal about the nature of atmospheric carbon dioxide and climate change?

- Many pre-Quaternary climates feature higher levels of atmospheric CO_2, and all those of the last 100 million years are associated with significantly warmer temperatures, both for climate states stable over millions of years (e.g., the mid-Pliocene, 3.5 ma) and for warm events lasting a few hundred thousand years (i.e., the Paleocene-Eocene Thermal Maximum, 55 ma).

What is the significance of glacial-interglacial variability in atmospheric composition and climate?

- Post-industrial levels of atmospheric CO_2 and CH_4 have risen far above the levels found in the longest (up to 800,000 years) ice-core records, highlighting the fact that the recent unprecedented rise in these trace gases does not stem from natural mechanisms. Over these multi-millennia time scales, Antarctic temperature and CO_2 co-vary with each other.

- There is no evidence that the current warming will be mitigated by a natural cooling trend towards glacial conditions. Understanding of orbital forcing indicates that the earth would not naturally enter another ice age for at least 30,000 years. Rising atmospheric CO_2 may delay or prevent the earth from entering the next anticipated ice age.

- The Last Glacial Maximum (ca. 21,000 years ago) featured reduced atmospheric concentrations of greenhouse gases, increased atmospheric aerosols, and altered land ice and vegetation. In a coordinated international multi-model experiment (PMIP-2), models simulate a change in global mean surface air temperature between the Last Glacial Maximum and the current interglacial of 3.1 to 5.1°C when considering greenhouse forcing and continental ice changes (radiative forcing change of -4 to 7 W m^{-2}). The other factors (vegetation and aerosol changes) have not yet been considered in most models, but initial results suggest they could cause additional cooling of ~ 2ºC.

- Global cooling at the Last Glacial Maximum is comparable in magnitude with the projected global mean warming over the 21st century, although the warming after the Last Glacial Maximum happened over thousands of years rather than the 100 years expected for future warming.

- Climate models are able to simulate the observed surface ocean temperature changes of the Last Glacial Maximum (ca. 21,000 years ago) of 1 to 4°C cooling in the tropics and 3 to 9°C cooling in the North Atlantic. Significant uncertainty still exists on the details of Atlantic thermohaline responses to Last Glacial Maximum forcings.

- During the past 120,000 years, many large and abrupt climate shifts have occurred, including more than 20 abrupt warmings known as Dansgaard-Oeschger events, and several cold Heinrich events. Temperatures changed by up to 16°C within a decade around the northern Atlantic (e.g., in Greenland). These events persisted for centuries and had global repercussions, such as major shifts in tropical rainfall patterns. They were probably not associated with large changes in global mean temperature, but rather with a redistribution of heat between northern and southern hemisphere.

- Strong evidence, both from sediment data and from modeling, links abrupt climate events to changes in the Atlantic Ocean circulation, although details of the mechanism are still under discussion. Our current understanding suggests that the ocean circulation can become unstable and change rapidly when critical thresholds are crossed. While it is unclear where these thresholds are and how much they differ between glacial and modern climate, it cannot be ruled out that future warming and meltwater inflow could again trigger major ocean circulation changes.

- At most 5% of the 20th century *global* sea level rise can be attributed to the disappearance of glacial ice sheets. *Regionally*, the contribution of post-glacial crustal adjustment can be dominant.

- The recent collapse of the Antarctic Larsen B ice shelf is likely unprecedented in the last 10,000 years, and partly the result of recent prolonged warming in the Antarctic Peninsula region.

- Large-scale retreat of the Greenland Ice Sheet during the previous interglacial (129 to 116 ka), confirmed by data and models, contributed 3 to 4 meters of sea level rise. Warming in the Arctic during the previous interglacial is comparable to warming expected at the end of this century, and may have also triggered partial, yet extremely rapid (>1 m/century), deglaciation of the West Antarctic Ice Sheet.

What does the climate of the current interglacial tell us about 20th century climate change?

- During the last 10,000 years, different regions of the Earth underwent periods warmer than the 20th century because of changes in the Earth's orbit and the resulting seasonal and latitudinal distribution of incoming solar radiation.

- Commonly cited warm periods, including the Medieval Warm Period, Holocene Climate Optimum, Holocene Thermal Optimum, Altithermal, Hypsithermal and others, appear to have been distinct only regionally and asynchronously. There are no known periods of synchronous global warmth comparable to the late 20th century during the Holocene, and this is consistent with our understanding of orbital forcing.

- For the mid-Holocene (6000 years ago), coupled climate models are able to simulate most robust large-scale features of observed climate change, including mid-latitude warming, with little change in global mean temperature (<0.4°C), and enhanced monsoons, consistent with our understanding of orbital forcing. Coupled climate models perform generally better than atmospheric models alone, and reveal the amplifying roles of ocean and land surface feedbacks in natural climate change

- The present global retreat of glaciers is unparalleled during the Holocene, and disagrees with the expected natural forcing mechanisms favouring larger regional variability due to changes in temperature and precipitation.

- Small variations of atmospheric greenhouse gas concentrations observed during the pre-industrial Holocene are predominantly attributed to natural processes that cannot account for the industrial era greenhouse gas increase.

- There is no evidence for centennial to millennial modes of natural climate variability generating *global* warming and cooling in the past, or that could explain global warming of the last 150 years.

- The ability of climate and vegetation models to simulate past northward shifts of reconstructed expansion of the treeline under warming conditions in the past supports the simulated significant northward (and upward) expansion of boreal trees in the NH under global warming.

- The strength and frequency of ENSO extremes have varied in response to past changes in orbital forcing, indicating that the impacts of ENSO outside the tropical Pacific are not stable as background climate and forcings change.

- During the present interglacial, slow changes in orbital forcing appear to have triggered abrupt changes in the frequency of hurricanes and floods, the frequency, extent and duration of droughts, and the spatial and temporal character of tropical precipitation. The degree of forcing required to trigger such events remains uncertain, as causal mechanisms are not well understood.

What does the climate of the last 2000 years tell us about 20th century climate change?

- Since the TAR, there has been an expansion in the length and geographical coverage of high-resolution proxy data, as well as in the number of hemispheric temperature reconstructions using the available data.

- Some of the post-TAR studies indicate greater multi-centennial Northern Hemisphere variability than was shown in the TAR, due to the particular proxies used, and the specific statistical methods of processing and/or scaling them to represent past temperatures. The additional variability implies cooler temperatures, predominantly during the 12th to 14th, the 17th, and the 19th centuries. Only one reconstruction suggests slightly warmer conditions than were shown in the TAR, in the 11th century.

- The TAR pointed to the "*exceptional warmth of the late 20th century, relative to the past 1000 years*". Subsequent evidence reinforces this conclusion. Indeed, it is *very likely* that average Northern Hemisphere temperatures during the second half of the 20th century were warmer than any other 50-year period in the last 500 years. It is also *likely* that this was the warmest period in the past 1000 years and unusually warm compared with the last 2000 years. The uneven coverage and characteristics of the proxy data mean that these conclusions are most robust over summer, extra-tropical, land areas.

- Taken together, the very sparse evidence for Southern Hemisphere temperatures prior to the period of instrumental records indicates that it is *likely* that the warmth of the last 50 years is unusual in a 350 to 1000 year context.

- Paleoclimatic reconstructions and simulations of the last 1000 years point to the increasing importance of greenhouse gases as the cause of unprecedented recent warming.

- There is proxy evidence that solar magnetic activity has changed in the past, but how this affected the total solar irradiance and climate remains uncertain.

- The ice core CO_2 record over the past millennium provides an additional constraint on natural climate variability. The available hemispheric temperature reconstructions are broadly consistent with the ice core CO_2 record over the last millennium and the strength of the carbon cycle-climate feedback as found in the models used in the projection chapter. The CO_2 record is not compatible with the existence of a large solar forcing effect on climate over the last millennium.

- Early instrumental data, mostly from Europe, show that 1980–2004 is *very likely* the warmest 25-year period during the last 280 years.

- Most coral records currently available from the tropical Indo-Pacific, and stretching continuously back from the late 20th century, suggest unusual warmth in the late 20th century.

- Reconstructions of the behaviour of ENSO over the past millennium indicate greater variability in the frequency, amplitude, and climate teleconnections than is represented in the period of instrumental record.

- There is evidence that the strength of the Asian monsoon, and hence precipitation amount, can change abruptly, perhaps as a non-linear response to more gradual changes in forcing.

- The paleoclimate records of northern and eastern Africa and of North America indicate that droughts lasting decades to centuries are a recurrent feature of climate in these regions, and that recent droughts in North America and Northern Africa are not unprecedented. Current understanding cannot rule out the occurrence of decadal and longer drought in the future.

What does the paleoclimatic record reveal about biogeochemical and biophysical processes?

- Paleoclimatic data are consistent with the view that an increase in the atmospheric concentration of CO_2 and other greenhouse gases causes global warming.

- Paleoclimatic data suggest that biogeochemical and biophysical feedbacks have amplified changes in incoming solar energy caused by changes in the earth's orbit around the sun. It is thus likely that biogeochemical and biophysical feedbacks will amplify future direct anthropogenic greenhouse gas forcing, thereby causing larger climatic changes than in the absence of these feedbacks.

- Paleoclimatic data suggest limited roles of aeolian iron deposition into the Southern Ocean and the strength of North Atlantic Deep Water in regulating past atmospheric CO_2.

- Paleoenvironmental data suggest that vegetation composition and structure are sensitive to climate change, and can, in some cases, respond to climate change within decades, or even years.

- The incorporation of vegetation-atmosphere interactions within climate models improves the realism of climate simulations. Current models are capable of simulating the vegetation structure and terrestrial carbon storage for the Last Glacial Maximum and the Holocene.

6.1 Introduction

This chapter assesses paleoclimatic data and knowledge to inform how the climate system changes across interannual to millennial time-scales, and how well these variations can be simulated with climate models. We also highlight potential implications this knowledge has for the future. Paleoclimatic insights have contributed in central ways to previous IPCC assessments, but have never formed the basis for an entire chapter; this is the first IPCC chapter devoted to paleoclimate. Even so, a complete assessment of all relevant paleoclimatic data and knowledge is precluded by page limits. For these reasons, numerous scoping and lead author meetings have led to the current configuration of the chapter: one that taps the fullness of geologic time to assess what the most relevant and well-understood paleoclimatic lessons reveal about recent and possible future climate change. A central motivation of the chapter is the realization that we must understand the full range of natural variability in order to understand how humans might, or might not, alter the climate system. Satellite and instrumental records are simply too short to do this job alone.

Paleoclimate science has made significant advances since the 1970's, when a primary focus was on the origin of the ice ages, the possibility of an imminent future ice age, and the first explorations of the so-called Little Ice Age and Medieval Warm Period. Even in the first IPCC assessment (1990), climatic variations prior to the instrumental record were not that well known or understood. Fifteen years later, our understanding is much improved, quantitative and integrated with respect to observations and modeling. It is the intent of this chapter to bring the broader climate change community and policy-makers up to date.

After an overview of paleoclimatic methods, including their strengths and weaknesses, we examine the paleoclimatic record in chronological order, from oldest to youngest. This approach was selected for a number of reasons, but primarily because the climate system varies and changes over all time scales, and it is instructive – e.g., for policy debates – to understand the contributions lower frequency patterns of climate change might make in influencing higher-frequency patterns of variability and change. Also, an examination of how the climate system has responded to large changes in climate forcing in the past is useful in assessing how the same climate system might respond to the large anticipated forcing changes in the future. We also devote the most chapter space to recent paleoclimatic history because uncertainties become smaller toward the present. Moreover, climate variation and change of the last 2000 years is of great interest well beyond the discussions of the paleoclimatic community. Lastly, additional focused paleoclimatic perspectives are also included in other chapters of this volume: for example, Chapter 4, 9 and 10.

Cross-cutting our chronologically-based presentation are assessments of climate forcing and response, and of the ability of state-of-the-art climate models to simulate the responses. Perspectives from paleoclimatic observations, theory and modelling are integrated wherever possible to reduce uncertainty in our assessment. We attempt to balance both broad-scale (e.g., hemispheric) foci with more regional foci. In several sections, we also assess the latest developments in the rapidly advancing area of abrupt climate change: i.e., *forced* or *unforced* climatic change that involve crossing a threshold to a new climate regime (e.g., new mean state or character of variability), where the transition time to the new regime is short relative to duration of the regime (Rahmstorf, 2001; Overpeck and Trenberth, 2004).

6.2 Paleoclimatic Methods

6.2.1 Methods – Observations of Forcing and Response

The field of paleoclimatology has seen significant methodological advances since the TAR, and the purpose of this section is to emphasize these advances while giving an overview of the methods underlying the data used in this chapter. Many critical methodological details are presented in subsequent sections where needed. Thus, this methods section is designed to be more general, and to give readers of this chapter more insight and confidence in the findings of the chapter. Readers are referred to several useful books and special issues of journals for more additional methodological detail (Bradley, 1999; Cronin, 1999; Ruddiman and Thomson, 2001; Alverson et al., 2003; Mackay et al., 2003; Kucera et al., 2005)

How do we know how climate forcing changed in the past?
Perhaps one of the most important aspects of modern paleoclimatology is that it is possible to derive time series of hypothesized climate forcing stretching back centuries, millennia and farther. Box 6.1 cites several

papers that reveal how atmospheric CO_2 concentrations can be inferred back millions of years. As is common across all aspects of the field, paleoclimatologists seldom rely on one method or proxy, but rather several. In this way, results can be cross-verified and uncertainties better understood. In the case of pre-Quaternary CO_2, multiple geochemical and biological methods provide reasonable constraints on past CO_2 variations, but, as pointed out in Box 6.1, the quality of the estimates is somewhat limited. In contrast, measurements of atmospheric trace gases and aerosols for the period ca. 800,000 years to present (see Section 6.3 and 6.4 for more methodological citations) are much more exact because they can be measured directly from air trapped in polar ice. As is common in paleoclimatic studies of the Late Quaternary, the quality of forcing and response series are verified against recent (e.g., post 1950) measurements made by direct instrumental sampling.

[START OF BOX 6.1]

Box 6.1: The Pre-Quaternary - Forcing and Response

The relevance of the "Deep Past" climates (e.g., prior to 3 million years ago (myr ago)) is that they were often, by and large, warmer than today, and associated with higher CO_2 levels – in that sense they are the only paleoclimates that may be thought of as partial analogs for future climate change (partial in the sense that global biology and geography were increasingly different farther back in time). In general, they verify that warmer climates are to be expected with increased greenhouse gas concentrations. However, when we go back in time beyond the reach of ice cores, i.e., beyond a million years in the past, data on climate and especially greenhouse gas concentrations in the atmosphere become much more uncertain. There are ingenious efforts to obtain quantitative reconstructions of climate variations over the past 500 million years, but in view of the still considerable uncertainties in this field, we will only discuss a few particularly relevant qualitative conclusions here.

How accurately do we know paleo CO_2 levels? There are four primary proxies used for pre-Quaternary CO_2 levels (Royer et al., 2001; Royer, 2003). Two proxies apply the fact that biological entities in seawater (Cerling, 1991; Freeman and Hayes, 1992; Yapp and Poths, 1992; Pagani et al., 2005) have carbon isotope ratios that are distinct from the atmosphere. The third proxy uses the ratio of boron isotopes (Pearson et al., 2001), while the fourth uses the empirical relationship between stomatal pores on tree leaves and CO_2 levels (McElwain and Chaloner, 1995; Royer, 2003). As shown in Box 6.1, Figure 1, while there is a wide range of reconstructed CO_2 values, magnitudes are generally higher than the interglacial, pre-industrial values seen in ice core data. Changes in CO_2 on these long time scales are thought to be driven by changes in tectonic processes (e.g., volcanic activity and weathering). Indicators for the presence of continental ice on Earth show that the planet was mostly ice-free during geologic history. Two periods of major glaciation occurred around 300 million years ago and in the past 3 million years, coinciding with phases of relatively low CO_2 concentrations compared with surrounding epochs (Box 6.1, Figure 1). The major expansion of Antarctic glaciations starting around 35-40 ma may have been associated with a continuing decline in CO_2 levels during the Tertiary.

[INSERT BOX 6.1, FIGURE 1 HERE]

What does the record of the Mid-Pliocene tell us?
The sampling frequency for CO_2 reconstruction is highest, and the envelope or range of estimates smallest for the mid-Pliocene (approximately 3.5 myr ago). This is the most recent time in Earth's history when CO_2 concentrations were higher (estimated as 360–400 ppm) and temperatures warmer (estimated by GCMs to be 2ºC–3ºC above pre-industrial), providing an example of a world similar to what is anticipated for the future. The Pliocene is also recent enough for the configuration of continents and ocean basins to be close to the present. Hence this time period is currently studied intensively both by collecting more data and with the help of model simulations. What was the climate like for this time period? The middle Pliocene presents us with the mature state of a warmer world, essentially the resulting climate impact of a prior and continuing global warming. Sea level was probably some 15–25 m higher than modern (Dowsett and Cronin, 1990); ice sheets where correspondingly smaller.

Temperature reconstructions for this time period from both terrestrial and marine paleoclimate proxies (Thompson, 1991; Thompson and Fleming, 1996; Dowsett et al., 1999) suggest substantial high latitude warming without any significant tropical SST change, resulting in a substantial decrease in the latitudinal temperature gradient. For example, atmospheric simulations driven by reconstructed SSTs from the Pliocene Research Interpretations and Synoptic Mapping (PRISM) Group produced winter surface air temperature warming of 10–20°C at high northern latitudes with 5–10°C increase in the northern North Atlantic (~60°N), whereas there was essentially no tropical temperature change (or even slight cooling) (Chandler et al., 1994; Sloan et al., 1996; Haywood et al., 2000). In contrast, a coupled atmosphere-ocean experiment with 400 ppm CO_2 produced warming relative to pre-industrial times of 3–5°C in the northern North Atlantic, and 1–3°C in the tropics (Haywood et al., 2005).

The lack of tropical warming results from the tropical SST reconstructions based on marine microfaunal evidence; as in the case of the Last Glacial Maximum, we are uncertain whether tropical sensitivity is really as small as such reconstructions show. Haywood et al. (2005) found that alkenone estimates of tropical and subtropical temperatures do indicate warming in these regions, in better agreement with GCM reconstructions from increased CO_2 forcing. As in the previous example, the climate models cannot produce a response to increased CO_2 with large high latitude warming and yet minimal tropical temperature change unless strong increases in ocean heat transport also occur (Rind and Chandler, 1991).

The substantial high latitude response is shown by many different paleo-indicators, and it may indicate that high latitudes are more sensitive to increased CO_2 than indicated by model simulations for the 21st century. Alternatively, it may be the result of increased ocean heat transports due to either an enhanced thermohaline circulation (Raymo et al., 1989; Rind and Chandler, 1991), or increased flow of surface ocean currents due to greater wind stresses (Ravelo, 1997; Haywood et al., 2000), or associated with the reduced extent of land and sea ice (Jansen et al., 2000; Knies et al., 2002; Haywood et al., 2005). Currently available proxy data are equivocal concerning a possible thermohaline increase for the equilibrium climate of the Pliocene. An increase would, however, contrast with the transient coupled model simulations for the 21st century which tend to favor reduced deep water formation and ocean transports as climate warms. Understanding the climate distribution for this time period may help improve our predictions of the likely response to increased CO_2 in the future.

What does the record of the Paleocene-Eocene Thermal Maximum tell us?

Approximately 55 million years ago, a spectacular global warming occurred. The climate anomaly, along with an accompanying carbon isotope excursion, occurred at the boundary between the Paleocene-Eocene epochs, and is therefore often referred to as the Paleocene-Eocene Thermal Maximum (PETM. The thermal maximum clearly stands out in high-resolution records of that time (Box 6.1, Figure 2). A "rapid" warming (in this case on the order of ten thousand years, lasting 100 kyr) by several degrees C is indicated by changes in ^{18}O isotope and Mg/Ca records (Kennett and Stott, 1991; Zachos et al., 2003; Tripati and Elderfield, 2004). At the same time, ^{13}C isotopes in marine and continental records show that a large mass of carbon with low ^{13}C concentration must have been released into the atmosphere and ocean. The mass of carbon was sufficiently large to lower the pH of the ocean and drive widespread dissolution of seafloor carbonates (Zachos et al., 2005). Possible sources for this carbon could have been methane from decomposition of clathrates on the sea floor, CO_2 from volcanic activity, or oxidation of organic rich sediments (Dickens et al., 1997; Kurtz et al., 2003; Svensen and al., 2004). The PETM, which altered ecosystems world-wide (Kennett and Stott, 1991) (Koch et al., 1992) (Bowen et al., 2004), is being intensively studied as it has some similarity with the ongoing rapid release of carbon into the atmosphere by humans. It shows a case of past carbon release, and the related strong climatic warming, although there is still too much uncertainty in the data to derive a quantitative estimate of climate sensitivity from this event.

[INSERT BOX 6.1, FIGURE 2 HERE]

Time series of astronomically driven insolation change are well known and can be calculated from celestial mechanics as described in sections 6.3 and 6.4. Reconstructions of past volcanic forcing based on historical sources and acidity in ice cores often do well in estimating the timing and rough magnitude of past events, but can be more limited in estimating exact optical depths, as well as the geographic and seasonal distribution of forcing associated with individual eruptions. However, the quality of volcanic forcing series likely improves with each new polar ice core drilled; see Section 6.5 for more methodological background. Estimates of past solar forcing are not as well constrained as other forcing series, but nevertheless there is evidence that solar variability does play a role in driving climate variability – see section 6.5.

How good is the time control associated with paleoclimatic records of forcing and response?
Much has been researched and written on the time control associated with paleoclimatic records, and readers are referred to the background books cited in the first paragraph of this section for more detail. In general, time control gets weaker farther back in time. Tree-ring records are generally the best, and are accurate to the year, or season of a year (even back thousands of years). There are a host of other proxies that also have annual layers or bands – e.g., corals, varved sediments, some cave deposits, some ice cores – but the age models are not always exact to a specific year. Again, paleoclimatologists always strive to generate age information from multiple sources to reduce age uncertainty, and most paleoclimatic interpretations must take into account uncertainties in time control.

There continue to be significant advances in radiometric dating, and the books cited above are a good place to start for more background. Each radiometric system has ranges over which the system is useful, and paleoclimatic studies almost always publish analytical uncertainties. Because there can be additional uncertainties, methods have been developed for checking assumptions and cross-verifying with independent methods. For example, secular variations in the radiocarbon clock over the last 15,000 years are very well known, and fairly well understood over the last 35,000 years. These variations, and the quality of the radiocarbon clock, have both been well demonstrated via comparisons with age models derived from precise tree-ring and varved sediment records, as well as with age determinations derived from independent radiometric systems such as uranium-series; note, however, that for each specific proxy record, the quality of the radiocarbon chronology also depends on the density of dates, the material available for dating and knowledge about the radiocarbon age of the carbon that was incorporated into the dated material.

It is difficult to state the age-model uncertainties associated, in general, with certain time control systems, but there are some general rules of thumb that can be used in the absence of more specific published information associated with a given paleoclimatic record of forcing or response. Non-tree-ring annually layered or banded proxies usually have chronologies accurate to a few percent and perhaps more further back in time. Unless otherwise stated the errors associated with radiometric age models are generally somewhat larger, particularly when ages near the upper time limit of the method (e.g., in the interval 30ka to 50ka for radiocarbon). With proper care, current methodologies allow more accurate age models.

How good are the methods used to reconstruct past climate dynamics?
Paleoclimatology is now a mature field, and this is reflected in the great assortment of well-tested methods used to reconstruct past climate system variability and change. Most of the methods behind the paleoclimatic reconstructions assessed in this chapter are described in some detail in the aforementioned books, as well as in the citations of each chapter section. In some sections, important methodological controversies are discussed where such discussions help assess paleoclimatic uncertainties.

Paleoclimatic reconstruction methods range from direct measurements of past change, as in the case of ground temperature variations, ocean sediment pore-water change, and glacier extent changes, to proxy measurements involving the change in chemical, physical and biological parameters that reflect – often in a highly quantitative and well-understood manner – past change in the environment where the proxy grew or existed. It is now well accepted and verified that many biological organisms (e.g., trees, corals, plankton, animals) alter their growth and/or population dynamics in response to changing climate, and that these climate-induced changes are well-recorded in past growth in living and dead (fossil) specimens or assemblages of organisms. Trees, ocean plankton and pollen are some of the best-known and best-developed proxy sources of past climate going back centuries and millennia. Networks of tree-ring width and tree-ring density chronologies are used to infer past temperature changes based on comprehensive calibration with temporally overlapping instrumental data. Past distributions of pollen and plankton from sediment cores can be used to derive quantitative estimates of past temperature, salinity or precipitation via statistical transfer functions which are well calibrated against their modern distribution and associated climate parameters. The chemistry of several biological and physical entities reflect well understood thermodynamic processes that can be transformed into estimates of climate parameters such as temperature. Key examples are: O-isotope ratios in coral and foraminiferal carbonate to infer past temperature and salinity, Mg/Ca and Sr/Ca ratios in carbonate for temperature estimates, alkenone saturation indices from marine organic molecules to infer past sea surface temperature (SST), O and H-isotopes in ice cores to infer temperature and atmospheric transport. Lastly, many physical systems (e.g., sediments and aeolian deposits) change in predictable ways that can be used to infer past climate change. While these methods are heavily used, there is ongoing work on further development and refinement, and there are remaining research issues concerning the degree to which the methods have spatial and seasonal biases. Therefore, in many recent paleoclimatic studies, a combination of methods is applied since multi-proxy series provide more rigorous estimates than single proxy and this approach may identify possible seasonal biases in the estimates. No paleoclimatic method is foolproof, and knowledge of the underlying methods is required when using paleoclimatic data.

Not surprisingly, the field of paleoclimatology depends heavily on replication and cross-verification between paleoclimate records from independent sources in order to build confidence in inferences about past climate variability and change. In this chapter, the most weight is placed on those inferences that have been made with particularly robust or replicated methodologies; the assessed quality of methods used is reflected in the confidence placed on the paleoclimatic inferences.

[END OF BOX 6.1]

6.2.2 Methods – Paleoclimate Modeling

Climate models are used to simulate episodes of past climate (e.g., climate during the Last Glacial Maximum or the last interglacial, or the time evolution during abrupt climate events) to help understand the mechanisms of past climate changes. Models are the only way to test physical hypotheses quantitatively, such as the Milankovich theory that glaciations were initiated because orbital cycles reduced the summer insolation on the northern continents (Khodri et al., 2001). Models allow us to link cause and effect in past climate change. Models also help to fill the gap between the local and global scale in paleoclimate, as paleoclimatic information is often sparse, patchy and seasonal. For example, the Vostok ice core data show a strong correlation between local temperature in Antarctica and the globally mixed gases CO_2 and methane, but the causal connections between these variables can only be understood with the help of models. A quantitative understanding of mechanisms is the best way to learn from past climate for the future, since there are no direct analogues of the future in the past.

At the same time, paleoclimate reconstructions offer the possibility of testing climate models, particularly if the climate forcing can be appropriately specified, and the response is sufficiently well-constrained. For earlier climates (i.e., before the current interglacial or Holocene), forcing and response are much larger, but data are more sparse and uncertain, while for recent millennia data are better, but forcing and signal are much smaller. Testing models with paleoclimatic data is important, as not all aspects of climate models can be tested against instrumental climate data. Good performance for present climate is not a conclusive test for a realistic sensitivity to CO_2, for example – to test this, simulation of a climate with very different CO_2 level must be used. Also, many empirical parameterizations describing sub-grid scale processes (e.g., cloud parameters, turbulent mixing) have been developed using present-day observations; hence climate states not used in model development provide an independent test-bed that increases the confidence in models.

In principle the same climate models that are used to simulate present-day climate, or scenarios for the future, are also used to simulate episodes of past climate. The difference is in the forcing (e.g., solar radiation or greenhouse gas concentrations), and for the deep past (tens of millions of years ago), also in the configuration of oceans and continents. The full spectrum of models (see Chapter 8) is used, ranging from simple conceptual models, through Earth system models of intermediate complexity (EMIC's) (Claussen et al., 2002) and coupled general circulation models. Since long simulations (thousands of years) are typically required for most paleoclimatic applications, and computer power is still a limiting factor, relatively "fast" coupled models are often preferred. Additional components, such as those that are not standard in models for present climate, are also increasingly added for paleoclimate applications, e.g., continental ice sheet models for simulation of glaciations (Hyde et al., 2000), or components describing the build-up of ocean sediments, or the behaviour of isotopes such as ^{18}O for a comparison with proxy data (Schmidt, 1999). Vegetation and ecosystem models are increasingly included, both to capture vegetation feedbacks on climate, and to allow for validation with ecological (e.g., pollen) data. The representation of biogeochemical tracers and processes is a particularly important new advance for paleoclimatic model simulations, as a rich body of information on past climate has emerged from proxy data from a variety of archives (see 6.2.2) that are intrinsically linked to the cycling of carbon and other nutrients.

6.3 Glacial-Interglacial Variability and Dynamics

6.3.1 Climate Forcings and Responses Over Glacial-Interglacial Cycles

Paleoclimatic records from deep sea sediments, continental sediments, and ice cores document a sequence of glacial-interglacial cycles covering the last 800,000 years in ice cores (Figure 6.1) and several million years in deep oceanic sediments. The last 500,000 years, which are the best documented, are characterized by 100 ka glacial-interglacial cycles of very large amplitude, as well as large climate changes at other orbital frequencies (Hays et al., 1976), and at millennial time scales (McManus et al., 2002; North Greenland Ice Core Project, 2004). Long glacial periods are interrupted by shorter interglacial warm periods lasting for 10 to 30 ka. We are now living in the Holocene period, the latest of these interglacials.

[INSERT FIGURE 6.1 HERE]

What were the main climate forcings and responses during the glacial-interglacial cycles?
Recent studies since TAR (see for instance Watanabe et al., 2003; Augustin et al., 2004) reconfirm the validity of the astronomical hypothesis of Milankovitch (Hays et al., 1976) and the major role, for at least the last 500,000 years, of variations in the Earth's orbit in driving large-scale climatic change; the primary variations are driven by variations in the tilt of the Earth's axis (41 kyr cycle), precession (19 and 23 kyr cycles) and eccentricity of the orbit (100 kyr cycle). The strong response to the 100 kyr cycle, which is associated with only weak insolation forcing, implies that the climate system reacts in a highly nonlinear manner with large positive feedbacks. Climate models indicate that the changes from glacial to interglacial conditions during the last deglaciation, which occurred between 20 and 10 ka ago, can be consistently explained by the orbital forcing working in concert with observed changes in greenhouse trace gases, northern ice sheet albedo and, to a lesser extent, dust and vegetation albedo.

The sequence of climatic forcings and responses during deglaciations (transitions from full glacial conditions to warm interglacials) are relatively well documented. High resolution ice core records of temperature proxies and CO_2 during degaciation indicates that Antarctic temperature starts to rise several hundred years before CO_2 (Monnin et al., 2001; Caillon et al., 2003). During the last deglaciation, and likely also the three previous ones, the onset of warming at both high southern and northern latitudes preceded by several thousand years the first signals of significant sea level increase resulting from the melting of the northern ice sheets linked with the rapid and major Bølling/Allerød warming at high northern latitudes (Petit et al., 1999a; Shackleton, 2000; Pépin et al., 2001). Current data are not accurate enough to identify whether warming started earlier in the Southern or Northern Hemisphere, but a major deglacial feature is the difference between North and South in terms of the magnitude and timing of strong reversals in the warming trend, which are out of phase between the hemispheres. These are much more pronounced in the Northern Hemisphere (Blunier and Brook, 2001).

[START OF BOX 6.2]

Box 6.2: What Caused the Low Atmospheric CO_2 Concentrations During Glacial Times?

Ice core records show that atmospheric CO_2 varied within about 180 to 300 ppm over the glacial-interglacial cycles of the last 650 thousand years (see main text and Figure 6.1) (Petit et al., 1999a; Siegenthaler et al., 2005). The CO_2 variations broadly followed Antarctic temperature, typically with a time lag of several centuries to a millennium. The evolution of Greenland temperature was often asynchronous with respect to Antarctic temperature (and CO_2) over the past 120,000 years, the period spanned by the Greenland ice core record. The CO_2 evolution is different for the different transitions from glacial to interglacials conditions and the rate of change in atmospheric CO_2 varies considerably over time. For example, the CO_2 increase from ~190 ppm at the Last Glacial Maximum to ~265 ppm in the early Holocene can be distinguished in different phases (Monnin et al., 2001; Figure Box 6.2). CO_2 increased relatively rapidly during periods of warming in Antarctica, slightly decreased during Antarctic cooling, and increased within a few decades only by about 10 ppm at the onset of warm phases in the Northern Hemisphere.

The quantitative and mechanistic explanation of these CO_2 variations remains one of the big unsolved questions in climate research. The reason is the complexity of the problem. Processes in the atmosphere, ocean, marine sediments, on land, and the dynamic of sea ice and ice sheets must be considered. A number of hypotheses for the low glacial CO_2 concentrations have emerged over the past 20 years and a rich body of literature is available (for overviews and further references see Broecker and Henderson, 1998; Archer et al., 2000; Sigman and Boyle, 2000; Kohfeld et al., 2005). Many processes have been identified that could potentially regulate atmospheric CO_2 on glacial-interglacial time scales. However, the existing proxy data are relatively scarce, uncertain, and partly conflicting.

Most explanations propose changes in oceanic processes as the cause for low glacial CO_2. The ocean is by far the largest of the relatively fast (<1000 yr) exchanging carbon reservoirs, and terrestrial changes cannot explain the low glacial values because terrestrial storage was also low at the Last Glacial Maximum (see section 6.6). On glacial-interglacial time scales, atmospheric CO_2 is mainly governed by the interplay between ocean circulation, marine biological activity, ocean-sediment interactions, seawater carbonate chemistry, and air-sea exchange. Atmospheric CO_2 is approximately in equilibrium with the CO_2 partial pressure of the global surface ocean. Upon dissolution in seawater, CO_2 maintains an acid/base equilibrium with bicarbonate (HCO_3^-) and carbonate (CO_3^{--}) ions that depends on the acid-titrating capacity of seawater, i.e., alkalinity. The production of biological material in the surface and its remineralization in the deep, deplete the surface ocean of dissolved inorganic carbon and alkalinity relative to the deep. The formation of calcium carbonates ($CaCO_3$) causes a higher CO_2 partial pressure, whereas the formation of organic material lowers the CO_2 partial pressure. Globally, the latter process dominates and atmospheric CO_2 would be higher in an ocean without biological activity. CO_2 is more soluble in colder than in warmer waters; therefore changes in surface and deep ocean temperature have the potential to alter atmospheric CO_2. Most hypotheses focus on the Southern Ocean, where today the cold deep-water masses of the world ocean are formed to a large extent, and large amounts of biological nutrients (phosphate and nitrate) upwelled to the surface remain unused.

One family of hypotheses of low glacial CO_2 values invokes an increase or redistribution in the ocean alkalinity as a primary cause. Potential mechanisms are (i) the increase of $CaCO_3$ weathering on land, (ii) a decrease of coral reef growth in the shallow ocean, or (iii) a change in the export ratio of $CaCO_3$ and organic material. These mechanisms require large changes in the deposition pattern of $CaCO_3$ to explain the full amplitude of the glacial-interglacial CO_2 difference through a mechanism called carbonate compensation. This is in conflict with the available sediment data. Furthermore, carbonate compensation may only explain slow CO_2 variation, as its typical time scale is multi-millennial.

Another family of hypotheses invokes changes in the sinking of marine plankton. Possible mechanisms include (iv) fertilization of phytoplankton growth in the Southern Ocean by increased deposition of iron-containing dust from the atmosphere, and a subsequent redistribution of limiting nutrients, (v) an increase in the whole ocean nutrient content, e.g., through input of material exposed on shelves or nitrogen fixation, and (vi) an increase in the ratio between carbon and other nutrients assimilated in organic material, resulting in a higher carbon export per unit of limiting nutrient exported. As with the first family of hypotheses, this family of mechanisms also suffers from the inability to account for the full amplitude of the reconstructed CO_2 variations when constrained by the available information. For example, periods of enhanced biological production and increased dustiness (iron supply) are coincident with only 20 to 50 ppm changes (see Box 6.2, Figure 1 for an illustration).

[INSERT BOX 6.2, FIGURE 1 HERE]

Physical processes also likely contributed to the observed CO_2 variations. Possible mechanisms include (vii) changes in ocean temperature (and salinity), (viii) suppression of air-sea gas exchange by sea ice, and (ix) increased stratification in the Southern Ocean. The combined changes in temperature and salinity increased the solubility of CO_2, causing a depletion in atmospheric CO_2 of perhaps 30 ppm. Simulations with general circulation ocean models do not fully support the gas exchange-sea ice hypothesis. One explanation (ix) conceived in the 1980s invokes more stratification, less upwelling of carbon and nutrient-rich waters to the surface of the Southern Ocean, and increased carbon storage at depth during glacial times. The stratification may have caused a depletion of nutrients and carbon at the surface, but proxy evidence for surface nutrient utilization is controversial. Qualitatively, the slow ventilation is consistent with very saline and very cold deep waters reconstructed for the last glacial maximum (Adkinson et al., 2002) and low glacial stable carbon isotope ratios ($^{13}C/^{12}C$) in the deep South Atlantic. However, it is apparently in conflict with radiocarbon data suggesting that the ventilation rate of the deep Pacific was similar to today (Broecker et al., 2004).

In conclusion, the explanation of glacial-interglacial CO_2 variations remains a difficult attribution problem. It appears likely that a range of mechanisms have acted in concert. The challenge is not only to explain the amplitude of glacial-interglacial CO_2 variations, but also the complex temporal evolution of atmospheric CO_2 in a way that is consistent with the underlying changes in climate.

The mechanisms described also play a role for the anthropogenic CO_2 perturbation. On millennial time scales, it is very likely that the sediment-compensation mechanism operating over glacial-interglacial cycles will also act to neutralize excess anthropogenic carbon and remove anthropogenic carbon from the atmosphere (Archer et al., 1999). On decadal-to-century time scales, changes in the marine and terrestrial biological cycles, as well as in ocean physics, are likely to impact on future atmospheric CO_2 as further discussed in Chapter 7.

[END OF BOX 6.2]

This evidence, combined with the efforts made for quantifying the different perturbations to the energy balance of the planet during a glacial-interglacial transition (Lorius et al., 1990; Hewitt and Mitchell, 1997; Hansen, 2003; Lea, 2004), leads to the conclusion that both the increase in greenhouse gases (especially in CO_2) and the shrinkage of the northern hemisphere ice sheets played major roles in amplifying the weak initial forcing due to the orbital eccentricity both at high latitudes and in the tropics. Greenhouse gas forcing was thus an important prerequisite for providing the strong 100,000 year cycle in tropical SST records (Lea, 2004). The high latitude temperature changes occur as a first response to the orbitally induced forcing. CO_2 forcing lags temperature, but leads sea level changes. Greenhouse gases were thus an important feedback critical for causing the high amplitude of deglacial warming and wasting of the continental ice sheets during deglaciations.

What does the last ice age tell us?

Understanding and modeling past glacial cold periods provide a counterpart to projections of future warm climates. The Last Glacial Maximum (LGM, approximately 21,000 years ago) has been widely studied because the radiative forcings and boundary conditions are relatively well known and because numerous proxy records document large climate changes. Model intercomparisons from the first phase of the Paleoclimate Modeling Intercomparison Project (PMIP-1), using atmospheric models (either with prescribed SST or with simple slab ocean models), were featured in the IPCC TAR. The second phase (PMIP-2) LGM simulations use coupled atmosphere-ocean models and provide a benchmark for simulations presented in Chapters 8 and 10. LGM results from 5 models are available for this assessment. Recent progress in paleodata and modelling have enabled evaluation of glacial-interglacial changes in radiative forcing and temperature, providing new empirical measures of climate sensitivity and confirming the role of CO_2 as an amplifier of glacial-interglacial climate change.

The LGM featured reduced concentrations of well-mixed greenhouse gases, relative to preindustrial values, amounting to a global radiative forcing decrease of 2.8 W m^{-2}, approximately equal to, but opposite from, the radiative forcing of these gases for year 2000, relative to 1750 (Chapter 2). The annual mean global solar insolation change due to altered orbital parameters is small (0.014 W m^{-2}). Increased land ice covered significant parts of North America and Europe at the LGM. The radiative forcing of the LGM ice is –3 W m^{-2}, but with a large uncertainty range associated with uncertainties in the coverage and height of LGM continental ice (Mangerud et al., 2002; Peltier, 2004; Toracinta et al., 2004) and the parameterization of ice albedo in climate models (Taylor et al., 2000). Altered vegetation, in particular the observed expansion of tundra over the northern continents and the reduction of tropical rain forest, and increased atmospheric aerosols (dust primarily) each contribute about –1 W m^{-2} (Claquin et al., 2003; Crucifix and Hewitt, 2005). Overall, the radiative forcing for the LGM is approximately –6 to –11 W m^{-2}, although as for present forcings, the LGM radiative forcings cannot be strictly added up because of distinct spatial and seasonal features (Figure 6.2).

[INSERT FIGURE 6.2 HERE]

Paleoclimate data indicate that the climate over land was generally colder and drier at the LGM. Annual mean surface temperature averaged over the Northern Hemisphere continents is estimated to be 7.2 ± 1.5°C colder than preindustrial (Bintanja and Van de Wal, 2005). Strong reduction in temperatures in northern latitudes produced a southward displacement and major reduction in area of forest (Bigelow and al., 2003), an expansion of permafrost limits over NW Europe (Renssen and Vandenberghe, 2003), a fragmentation of temperate forests (Prentice et al., 2000; Williams et al., 2000), and a predominance of steppe-tundra in Western Europe (Peyron et al., 2005). Reconstructions also show reduced extent of tropical forests and increased extent of grasslands and shrub lands (Elenga and al., 2000). Polar ice core temperature reconstructions indicate cooling in eastern Antarctica of 9 ± 2°C (Stenni et al., 2001) and in Greenland of 21 ± 2°C (Dahl-Jensen et al., 1998). Cooling on the order of 5°C over land in the tropics is suggested by lowered mountain glaciers, vegetation changes and noble gas results (Ballantyne et al., 2005).

The magnitude of ocean cooling at the LGM has been hotly debated, particularly in the tropics. Reconstructions from the 1970s and 1980s gave ~3°C cooling in the tropical Atlantic and little or no cooling in the tropical Pacific (CLIMAP project members, 1981) More recent reconstructions indicate more pronounced cooling. Regional coral records suggest cooling of 4–5°C (Guilderson et al., 1994). More recent syntheses suggest tropical SST cooling is generally 0–3.5°C (Rosell-Mele et al., 2004; Ballantyne et al., 2005; Barker et al., 2005; Barrows and Juggins, 2005; Chen et al., 2005). Newly developed databases indicate a more seasonally ice-free northern North Atlantic with a more meridional surface circulation during the LGM, compared to the CLIMAP reconstruction (Sarnthein et al., 2003; deVernal et al., 2005; Meland et al., 2005), and a large seasonal migration of sea ice around Antarctica (Gersonde et al., 2005). The strength and depth extent of the LGM Atlantic thermohaline circulation have been examined through the application of a variety of new marine proxy indicators (Rutberg et al., 2000; Duplessy et al., 2002; Marchitto et al., 2002; McManus et al., 2004). These tracers indicate that the boundary between North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW) was much shallower during LGM. Changes in the overturning strength are more difficult to determine and vary among the proxy indicators. Reconstructions of deep water conditions in the LGM indicate much colder and saltier conditions than present in the Atlantic and that Southern Ocean deep waters were saltier than those in the North Atlantic (Adkins et al., 2002).

How realistic are results from climate model simulations of the Last Glacial Maximum?
Model simulations using the CLIMAP-prescribed glacial SST and cryosphere boundary conditions (CLIMAP, 1981) result in a global cooling of 3.5–4.5°C (IPCC, 2001). Models run with slab oceans and specified ocean heat transports have produced global mean temperature changes on the order of 2–6°C (IPCC, 2001; Kim et al., 2003; Otto-Bliesner et al., 2005). Since the TAR, coupled ocean-atmosphere models have been used to simulate the LGM (Figure 6.2; Table 6.1), although with varying prescriptions of the changed forcings (Kitoh et al., 2001; Hewitt et al., 2003; Kim et al., 2003; S.I. Shin et al., 2003; Peltier and Solheim, 2004). These simulations generally produce a global cooling that falls within the range of the PMIP-1 results that were discussed in the TAR (IPCC, 2001), except for one simulation that exhibits very cold conditions, 10°C cooling (Kim et al., 2003). Recent PMIP-2 LGM simulations in which standard specified forcings for greenhouse gases and continental ice sheets are employed (radiative forcing decrease of 4 to 7 W m^{-2}) give a range of global cooling of 3.1–5.1°C (Masson-Delmotte et al., 2005). Other important factors (vegetation and aerosol changes) have not yet been considered in most models, but initial results suggest they could cause additional global cooling of ~2ºC (Schneider, submitted).

[INSERT TABLE 6.1 HERE]

PMIP-2 models show better agreement with pollen-based temperature reconstructions over western Siberia than PMIP-1 models because of the reduced European ice sheet extent (Kageyama et al., 2005). Polar amplification of global cooling in Antarctica is simulated by at least some of the PMIP-2 models, but the large LGM cooling over Greenland is still underestimated (Masson-Delmotte et al., 2005). Changes in vegetation appear to improve the realism of LGM simulations and point to important climate-vegetation feedbacks (see Section 6.6). As noted in the TAR, cooling of tropical land regions computed by the PMIP-1 models was ~1.8–3.4°C for specified CLIMAP SSTs and ~1.8–5.3°C for slab ocean simulations, both ranges somewhat colder than the cooling obtained from observations.

The PMIP-2 models give a cooling of 1–2.3°C for the tropical oceans, on the cold end of proxy estimates (Figure 6.2; Table 6.1). Changes in greenhouse gas concentrations may account for about half of the tropical cooling (S.I. Shin et al., 2003; Shin et al., 2003a; Otto-Bliesner et al., 2005, 2005a). The PMIP-2 models predict a strengthening of the SST meridional gradient in the LGM North Atlantic and cooling and expanded sea ice within the range of proxy indicators (Kageyama et al., 2005). Models indicate that the shoaling of the North Atlantic meridional overturning circulation can be explained by increasing oceanic stability in the LGM ocean associated with the response of the Southern Ocean sea-ice cover to lower glacial atmospheric CO_2 levels (S.-I. Shin et al., 2003; Liu et al., 2005). The PMIP-2 models simulate a range of responses of the Atlantic deep ocean and overturning circulation to the LGM forcings.

In conclusion, PMIP-2 simulations of the glacial-interglacial changes in greenhouse gas and ice sheet forcing (4 to 7 W m^{-2}) and mean global temperature (3.1 to 5.1°C) lead to a climate sensitivity of 0.4 to 1.2°C per W m^{-2}, which is similar to the range of climate sensitivity estimated for a doubling of CO_2. The PMIP-2 results also confirm that current coupled models are able to simulate the response to large-scale climate forcing change.

How do the radiative forcings of previous interglacials compare with the pre-industrial Holocene?
Modulation by the 400 kyr orbital eccentricity period leads to similar and small insolation changes during marine isotope stage 11 (an interglacial about 400 ka ago) and the present Holocene interglacial. This situation will continue for the next 50,000 years (Berger and Loutre, 2003). In contrast, insolation changes had very large amplitudes around the last interglacial (129 to 116 ka). Orbital variations induced large changes latitudinally and with season (on the order of 50 W m^{-2}). Observed greenhouse gas concentrations are similar during the pre-industrial Holocene and the four previous interglacials (Petit et al., 1999a; Figure 6.1), suggesting a fairly constant GHG radiative forcing during these interglacials.

How much did the Arctic warm during the previous interglacial?
Average insolation in NH summer was higher than today's by about 12% during the Last Interglacial (LIG) period (129 to 116 ka). However, the global annual radiation change for LIG from the present day insolation was small. Based on various lines of evidence, the climate of the LIG in both the SH and NH is inferred to be warmer than today's (Kukla and al., 2002), including warmer-than-present coastal waters in the Pacific, Atlantic, and Indian Oceans and in the Mediterranean Sea (Muhs et al., 2002), greatly reduced sea ice in the coastal waters around Alaska (Brigham-Grette and Hopkins, 1995), and extension of boreal forest into areas now occupied by tundra in interior Alaska and Siberia (Brigham-Grette and Hopkins, 1995; Lozhkin and Anderson, 1995; Muhs et al., 2001). Such an advance of forest implies July temperature increases of 4–8°C in Siberia and 0–2°C in Alaska (CAPE Last Interglacial Project Members, 2005). Globally, there was less glacial ice on Earth during the LIG period than now, as evidence on tectonically stable coasts indicates that sea-level was 3–7 m above present (Stirling et al., 1998; Fruijtier et al., 2000; Muhs et al., 2002).

Coupled atmosphere-ocean models forced by known orbital changes can be used to evaluate polar amplification in the Arctic. When forced with LIG orbital forcing, these models show a seasonal Arctic warming of ~1–2°C, with large warming in the circum-North Atlantic region associated with sea ice retreat (Montoya et al., 2000; Otto-Bliesner et al., 2005a). Ice core data indicate a large response over Greenland and Antarctica with warming of 5°C during the LIG (North Greenland Ice Core Project, 2004). The NorthGRIP core indicates that the Greenland ice sheet did not fully disappear during the LIG. Greenland ice sheet models forced with increased temperatures as deduced from Antarctic isotope records of LIG warming give a much-reduced ice cap and likely range of sea level rise of 2.7–4.5 m (Tarasov and Peltier, 2003). When simulated Arctic temperatures and precipitation are used to force a coupled ice sheet-heat flow model, negative mass balance results, although the Greenland ice sheet does not completely melt (Otto-Bliesner et al., 2005b). Current coupled models may underestimate the rate of melting and warming because of missing feedbacks and incomplete ice sheet physics. Summer warming in the Arctic by the end of this century may be comparable to the summer half-year warmth that melted much of the Greenland ice sheet in the past (Overpeck et al., 2005).

How long did the previous interglacials last?
The four interglacials of the last 450 ky preceding the Holocene (marine isotope stages 5, 7, 9 and 11) were all different in multiple aspects, including duration (Figure 6.1). The shortest (Stage 7) lasted a few thousands years, and the longest (Stage 11) lasted almost 30 kyr. Evidence for an unusually long Stage 11 has been recently reinforced by new ice core and marine sediment data. The EPICA Dome C Antarctic ice core record suggests that Antarctic temperature remained approximately as warm as the Holocene for 28 kyr

(Augustin et al., 2004). A new stack of 57 globally-distributed benthic $\delta^{18}O$ records presents age estimates at Stage 11 nearly identical to those provided by the EPICA results (Lisiecki and Raymo, 2005).

It has been suggested that Stage 11 was an extraordinary long interglacial period because of weak astronomically induced changes in the distribution of solar energy reaching the surface of the Earth at this time (Berger and Loutre, 2003). But the weak orbital forcing allows other candidate forcings, such as CO_2, to play an important climatic role. The recently revisited Vostok record offers a CO_2 record covering the complete Stage 11 warm period, and this record shows CO_2 concentrations similar to pre-industrial Holocene values over all of Stage 11 (Raynaud et al., 2005). Thus, both the orbital forcing and the CO_2 feedback were providing favourable conditions for an unusually long interglacial. Moreover, the length of Stage 11 has been simulated by conceptual models of the Quaternary climate, based on threshold mechanisms (Paillard, 1998). For Stage 11, these conceptual models show that the deglaciation is triggered by the insolation maximum at ~427 kyr, but that the next insolation minimum is not sufficiently low to start another glaciation. The interglacial thus lasts an additional precessional cycle, yielding a total duration of 28 kyr.

The long duration of Stage 11 results from the interplay between orbital forcing and the effects of elevated CO_2 (Loutre, 2003). In terms of climate forcings and responses, Stage 11 appears quite similar to the elapsed part of the Holocene. Because of that, attempts have been made to estimate the length of the Holocene interglacial if it were free of anthropogenic perturbation. Different results are obtained depending on how the proxy record is aligned with the orbital forcing conditions (Ruddiman, 2005) or the definition of the start of the interglacial conditions (Augustin et al., 2004).

Can we predict transitions out of interglacials and into ice ages?

Decreased radiation in NH summer at high latitudes, due to a combination of perihelion occurring near the winter solstice, a strongly elliptical orbit, and a small axial tilt, is the critical climatic factor for transition into an ice age, as cold summers allow the survival of snow cover from the previous winter season. At the end of the last interglacial (~116 kyr BP), when the summer incoming solar radiation in NH at high latitudes reached minimum values, the climate started to cool down, marking the beginning of the latest glacial stage (Chapman and Shackleton, 1999; Shackleton et al., 2002). When forced with orbital insolation changes only, past model studies have failed to find the proper magnitude of response to allow for perennial snow cover. Recent modeling results including additional factors have been more promising. These include vegetation feedbacks (Crucifix and Loutre, 2002; Meissner et al., 2003), increased sea ice (Jackson and Broccoli, 2003), a coupled dynamical ice sheet model (Pollard and Thompson, 1997), increased northward atmospheric moisture transport from warm low-to-mid latitude oceans (Khodri et al., 2003; Khodri et al., 2005), or increased NADW production allowing for increased snowfall (Wang and Mysak, 2002; Otterå et al., 2004). Recent land and marine records show that the initiation of a northern high latitude cooling took place in a warm North Atlantic/Nordic Seas and mid-latitude land environment (Cortijo et al., 1999; Goni et al., 1999; McManus et al., 2002; Risebrobakken et al., 2005).

When will the current interglacial end?

There is no evidence that current global warming will be mitigated by a natural cooling trend. Only a strong reduction in summer insolation at high northern latitudes and associated feedbacks can end the current interglacial. The Earth's orbit around the Sun can be calculated with high precision for the future, as well as the past (Gallee et al., 1991). With low orbital eccentricity, the effects of precession are minimized, and extreme cold-northern-summer orbital configurations like that of the last glacial initiation at 116 kyr BP do not occur. Under a natural CO_2 regime (i.e., with the global temperature-CO_2 correlation continuing as in the Vostok ice-core), the next glacial period could not be expected to start within the next 30 kyr (Loutre and Berger, 2000; Berger and Loutre, 2002; Augustin et al., 2004). Sustained high atmospheric greenhouse concentrations, comparable to a mid-range CO_2 stabilization scenario, may lead to a complete melting of the Greenland ice cap (Church et al., 2001) and further delay the onset of the next glacial period (Loutre and Berger, 2000; Archer and Ganopolski, 2005).

6.3.2 Abrupt Climatic Changes in the Glacial-Interglacial Record

What is the evidence for past abrupt climate changes?

Abrupt climate changes have been variously defined either simply as large changes within less than 30 years (Clark et al., 2002), or in a physical sense, as a threshold transition or a response that is fast compared to forcing (Rahmstorf, 2001; Alley et al., 2003) or duration of the subsequent climatic regime (Overpeck and Trenberth, 2004); note that not all abrupt change need be forced. Numerous climatic records from sediments, ice cores, corals and other sources show that large, widespread, abrupt climate changes have occurred repeatedly throughout the past glacial interval (see recent review by (Rahmstorf, 2002)). These records, particularly those from high latitudes, show that ice-age abrupt events were larger and more widespread than those of the Holocene. The most dramatic of these abrupt climate changes are the warm Dansgaard-Oeschger (DO) events, characterised by a warming in Greenland by 8 to 16ºC within a decade or so (Severinghaus and Brook, 1999; Huber et al., 2004). Another type of abrupt change is the cold Heinrich event; characteristic of these is a large discharge of icebergs into the northern Atlantic leaving diagnostic drop-stones in the ocean sediments (Hemming, 2004). While the cooling during these events appears to have occurred on a century time-scale, the warming that ended them took place within decades (Figure 6.3; Cortijo et al., 1997; Voelker, 2002).

[INSERT FIGURE 6.3 HERE]

The repercussions of these abrupt climate changes were global, although out-of-phase responses in the two hemispheres (Blunier et al., 1998) suggest that they were not primarily changes in global mean temperature. The highest amplitude of the changes, in terms of temperature, appears centred around the North Atlantic. Strong changes are found in tropical wetlands (Chappellaz et al., 1993; Brook et al., 2000) and in the Asian monsoon (Wang et al., 2001). The Northern Hemisphere cold phases were linked with a southward shift of the inter-tropical convergence zone (ITCZ) and thus the location of the tropical rainfall belts (Peterson et al., 2000; Lea et al., 2003). Cold, dry, and windy conditions generally occurred together in the Northern Hemisphere cold events. Major changes in atmospheric methane content (on the order of 100–150 ppbv) (Chappellaz et al., 1993) and dust aerosols also characterize these abrupt changes, with lower methane content and higher dust aerosols characterizing cold phases (Mayewski et al., 1994). The accompanying changes in atmospheric CO_2 content were relatively small (up to 20 ppm; Figure 6.3). The record in N_2O is less complete and shows an increase of ~50 ppb a decrease of ~30 ppb during warm and cold periods respectively, in the deglaciation record from Antarctic ice cores (Flückiger et al., 2004).

In the North Atlantic, Heinrich events are accompanied by a lowering of sea surface salinity (Bond et al., 1993). Most DO events have a similar, but lower-amplitude, influx of icebergs than the Heinrich events, as well as significantly reduced surface water salinities. At the end of the last glacial, as the climate warmed and ice sheets melted, climate went through a number of abrupt cool phases, notably the Younger Dryas and the 8.2ka cold event. These events were also associated with a southward shift of the boreal treeline and other rapid vegetation responses (Peteet, 1995; Shuman et al., 2002; Williams et al., 2002).

What do we know about the mechanism of these abrupt changes?

There is solid evidence now from sediment data for a link between these glacial-age abrupt changes in surface climate and ocean circulation changes (Clark et al., 2002). Proxy data show that the South Atlantic cooled when the north warmed, and vice versa (Voelker, 2002), a see-saw of northern and southern hemisphere temperatures which indicates an ocean heat transport change. During DO events, salinity in the Irminger Sea increased strongly (Kreveld et al., 2000), indicative of saline Atlantic waters advancing northward. Abrupt changes in deep water properties of the Atlantic have been documented from both proxy data reconstructing the ventilation of the deep water masses (e.g. ^{13}C, $^{231}Pa/^{230}Th$) and kinematic proxies that reconstruct changes in the overturning rate and flow speed of the deep waters (Vidal et al., 1998; Dokken and Jansen, 1999; McManus et al., 2003). Despite this evidence many features of the abrupt changes are still not well constrained due to a lack of precise temporal control of the sequencing and phasing of events between the surface, the deep ocean and ice sheets.

Heinrich events are thought to have been caused by ice-sheet instability (Hemming, 2004). Iceberg discharge would have provided a large freshwater forcing to the Atlantic, which can be estimated from changes in the oxygen isotope ^{18}O. These yield a volume of freshwater addition typically corresponding to a few (up to 15) meters of global sea-level rise occurring over several centuries (200-2,000 years), i.e. a flux of the order of 0.1 Sv – but uncertainty, especially in the duration of the events, means that an order of magnitude smaller or larger (in case of rapid influx in a only a few years) cannot be ruled out (Hemming, 2004). (For comparison: the current Greenland ice sheet holds the equivalent of 7 meters of sea level; melting this amount in a millennium corresponds to an average flux of ~ 0.1 Sv). Models suggest that such a freshwater influx could be enough to shut down deepwater formation in the Atlantic.

A similar mechanism is likely for the cold events at the end of the last ice age (i.e. Younger Dryas, 8.2 k event). Rather than sliding ice, it is the inflow of meltwater from melting ice due to the climatic warming at this time which could have interfered with the meridional overturning circulation and heat transport in the Atlantic – global meltwater flux during deglaciation peaked at ~0.4 Sv (Fairbanks, 1989), while the 8.2 k event was probably linked to an inflow of 1.6 x 10^{14} m^3 (i.e., 40 cm of sea level) within a few years (Teller et al., 2002). This is an important difference relative to the DO events, for which no large forcing of the ocean is known; model simulations suggest that small forcing may be sufficient if the ocean circulation was close to a threshold (Ganopolski and Rahmstorf, 2001). The exact cause and nature of these ocean circulation changes, however, is not universally agreed.

Can climate models simulate these abrupt changes?

Modeling the ice sheet instabilities that are the likely cause of Heinrich events is a difficult problem where the physics is not sufficiently understood, although recent results show some promise (Calov et al., 2002). Many model studies have been performed in which an influx of freshwater from an ice sheet instability (Heinrich event) or a meltwater release (8.2 k event) has been assumed and prescribed, and its effects on ocean circulation and climate have been simulated. These experiments suggest that freshwater input of the order of magnitude deduced from paleoclimatic data could indeed have caused NADW formation to shut down, and that this is a physically viable explanation for many of the climatic repercussions found in the data: e.g., the high-latitude northern cooling, the shift in the ITCZ and the hemispheric see-saw (Vellinga and Wood, 2002; Dahl et al., 2005; Zhang and Delworth, 2005). The asynchrony between Greenland and Antarctic temperature has been explained by a reduction in the North Atlantic Deep Water formation rate and oceanic heat transport into the North Atlantic region, producing cooling in the North Atlantic and warming in the southern hemisphere (Knutti et al., 2004). In freshwater simulations where the North Atlantic meridional overturning circulation is forced to collapse, the consequences also include an increase in nutrient-rich water in the deep Atlantic Ocean, higher $^{231}Pa/^{230}Th$ ratios in North Atlantic sediments, (Marchal et al., 1999; Marchal et al., 2000), a retreat of the northern treeline (Scholze et al., 2003; Higgins, 2004; Köhler et al., in press), a small (10 ppm) temporary increase in atmospheric CO_2 in response to a reorganization of the marine carbon cycle (Marchal et al., 1999), and CO_2 changes of a few ppm due to carbon stock changes in the land biosphere (Köhler et al., in press). A 10 ppb reduction in atmospheric N_2O is found in an ocean-atmosphere model (Goldstein et al., 2003), suggesting that a large part of the observed N_2O variations is of terrestrial origin. In summary, model simulations broadly reproduce the observed variations during abrupt events of this type.

Some authors have argued that climate models tend to underestimate the effect of past abrupt climate changes (Alley et al., 2003) and hence may underestimate the risk of future ones. However, drawing such a general conclusion is probably too simple, and a case-by-case evaluation is required to understand which effects may be misinterpreted in the paleoclimatic record, which mechanisms may be underestimated in current models, and what this means for assessments of the future. For example, climate models which (due to limited resolution) have an ocean heat transport that is too weak may underestimate the effects of past changes in ocean circulation, while overestimating the ocean's sensitivity to future warming scenarios. In addition, for models to reproduce these events in all their particulars requires experiments to be run with the glacial climate conditions under which the real world events occurred.

DO events appear to be associated with latitudinal shifts in oceanic convection between the Nordic Seas and the open mid-latitude Atlantic (Alley and Clark, 1999). Models invoking such a mechanism suggest that the temperature evolution in Greenland, the see-saw response in the South Atlantic, the observed Irminger Sea salinity changes and other observed features of the events may be explained by such a mechanism (Ganopolski and Rahmstorf, 2001), although it remains unclear what the trigger of these ocean circulation changes was. Alley et al. (2001) show evidence for a stochastic resonance process at work in the timing of these events, which means that a regular cycle together with random "noise" could have triggered them. This can be reproduced in models as long as a threshold mechanism is involved in causing the events.

6.3.3 Sea Level Variations Over the Last Glacial-Interglacial Cycle

What is the influence of past ice volume change on modern sea level changes?

Paleo-records of sea level history provide a crucial basis on which to understand the background variations upon which the modern sea level rise is superimposed. Even if no changes were currently operating in the climate system, highly significant changes of relative sea level (RSL) would still be occurring.

The primary, although not the sole, cause of this natural variability in sea level has to do with the planet's memory of the last deglaciation event of the Late Pleistocene ice age. This memory exists as a consequence of the extremely high effective viscosity of the planetary interior that governs the manner in which, through glacial isostatic adjustment (GIA), gravitational equilibrium is restored following the transfer of large volumes of water from the continent to the ocean, as continental ice sheets melt during deglaciation. A detailed theory currently exists with which this influence on RSL history may be accurately modelled, allowing the influence of the glaciation-deglaciation cycle to be isolated from the climatic effect in modern tide gauge and satellite records of relative sea level change.

A recent review and discussion of the several significant enhancements of the theory produced since the TAR has appeared in Peltier (2004a). The theory provides a prediction based upon the current ICE-5G (VM2) model of the present day rate of secular change of sea level measured with respect to the centre of mass of the planet (Peltier, 2004a), a signal that is often referred to as "absolute" sea level (see Chapter 6). This signal is currently being measured by the Gravity Recovery and Climate Experiment (GRACE) satellite system. The extent to which the climate change component of this signal will be contaminated by the GIA effect may thus be estimated and the GIA contamination is computed to be approximately –0.36 mm per year. This means that the Topex/Poseidon measurement (see Ch. 5) is biased downwards due to the GIA influence by this amount and therefore that the impact of modern climate change is larger by this same amount.

The theoretical model of ice-ocean-solid earth interactions may also assist in the identification of possible contributors to the modern rate of RSL rise. For example, it has been suggested that a significant contribution to the modern global rate could be due to a continuing (paleo)influence of the slow melting of the Antarctic ice sheet extending through the entire Holocene period (Fleming et al., 1998). Observed records of Holocene RSL history have been invoked, together with the theory, to show that if this had occurred at any significant rate then the mid-Holocene highstands of the sea that exist on all islands in the equatorial Pacific Ocean would have been eliminated. However, the elevations of these features are accurately predicted by the theory based upon the conventional assumption that global deglaciation had ceased by approximately 4 ka (Peltier and Solheim, 2002). This analysis, also produced since the TAR, places the onus for the explanation of the modern rate of RSL rise squarely upon climate change processes.

What was the magnitude of glacial-interglacial sea level change?

Model based paleo-sea level analysis also helps to refine estimates of the eustatic (globally averaged) rise of sea level that occurred during the most recent glacial-interglacial transition from LGM to Holocene. The ICE-5G (VM2) model fit to the RSL curve from the island of Barbados (Fairbanks, 1989) has been shown to provide a good approximation to the eustatic curve itself, and the ICE-5G(VM2) model fits to this data set (Figure 6.4; (Peltier, 2005).

[INSERT FIGURE 6.4 HERE]

The good fit of the model to the data (Figure 6.4) constrains the net eustatic rise since the LGM to a value near 120m, very close to that conventionally inferred (Shackleton, 2000) on the basis of deep sea oxygen isotopic information (Waelbroeck et al., 2002). These data, together with those from the Sunda Shelf of Indonesia (Hanebuth et al., 2000), appear to rule out the occurrence of any rapid rise in melting rate earlier than the meltwater pulse that occurred at 14.2 ka. This is an important though still actively debated issue (Yokoyama et al., 2000), as the paleo-constraint on the net sea level rise that occurred during the last deglaciation strongly constrains the model of the surface of the ice age earth and its usability for future projections.

Also shown on Figure 6.4 is the extension of the record of eustatic sea level change to the LIG at ~125,000 years before present. The portion of this record prior to Last Glacial Maximum is strongly constrained by oxygen isotopic measurements from deep sea sedimentary cores. During the LIG itself, direct sea level measurements based upon coastal sedimentary deposits and tropical coral sequences have clearly established that eustatic sea level was higher than present during this last interglacial by approximately 5–7m (Rostami et al., 2000; Muhs et al., 2002). Detailed analyses of the evolution of the Greenland ice sheet during this interval of time (Cuffey and Marshall, 2000; Tarasov and Peltier, 2003) suggest that this additional rise of sea level above the present eustatic level was associated with a significant diminution of both the Greenland and West Antarctic ice sheets.

What is the significance of the last major sea level rise above modern day?
As noted above, global sea level was 4–7m above modern during the LIG. One recent study (Overpeck et al., in press) assessed this sea level high-stand, and concluded that the associated LIG circum-Greenland warming was comparable with that of a future doubled-CO_2 climate; thus, the Earth could soon be warm enough to melt 4–7m of sea level equivalent. For reasons described above (Section 6.3.1; see also Overpeck et al., in press), it is likely that the Greenland ice sheet could only have contributed a maximum 3 to 4m of sea level during the LIG. Because there was no positive *global* temperature anomaly during the LIG, the contribution of ocean thermal expansion to the total LIG global sea level rise was likely small. This means that the melting of the Greenland ice sheet, and perhaps associated oceanographic change, may have triggered the melting of a portion of the West Antarctic Ice Sheet (Overpeck et al., in press), in agreement with evidence found under this ice sheet (Scherer et al., 1998). This raises the possibility that future Greenland melting could also contribute to a rapid disintegration of the West Antarctic Ice Sheet, and that associated rates of sea level rise could exceed one (Overpeck et al., in press) or more (McCulloch and Esat, 2000) meters per century. The ongoing collapse of Antarctic ice shelves appears to be both unprecedented in the current interglacial, and a response – in part – to anthropogenic warming (Domack et al., 2005).

6.4 The Current Interglacial

A variety of proxy records provide detailed temporal and spatial information concerning climate change during the current interglacial, the Holocene, a *ca* 11,600-year long period of increasingly intense human settling and anthropogenic modifications of the local (e.g. land use) to global (e.g. atmospheric composition) environment. The past impacts of climate change on human societies are outside the scope of this section. The well-dated reconstructions of the past 2000 years are covered in Section 6.5. In the context of both climate forcing and response, the Holocene is far better documented in terms of spatial coverage, dating and temporal resolution than previous interglacials. The evidence is clear that significant changes in climate forcing during the Holocene induced significant changes in climate response, including long-term and abrupt changes in temperature, precipitation, monsoon strength and ENSO. For selected periods such as the mid-Holocene, *ca* 6 ka, intensive efforts have been dedicated to the synthesis of paleoclimatic observations and modeling intercomparisons. Such large data coverage provides a sound basis to evaluate the capacity of climate models to capture the response of the climate system to the orbital forcing.

6.4.1 Climate Forcing and Response During the Current Interglacial

What were the main climate forcings during the Holocene?
During the current interglacial, changes in the Earth's orbit modulated the latitudinal and seasonal distribution of incoming solar radiation, while the global annual mean solar radiation remained stable (Berger, 1978). Superimposed on small, but significant, zonal annual mean insolation changes (1 to 5 W m^{-2}) - with opposite signs in the low and high latitudes - the dominant seasonal redistribution of incoming solar radiation involved change on the order of 15 to 35 W m^{-2} (Figure 6.5). Ongoing efforts to quantify Holocene changes in stratospheric aerosol content recorded in the chemical composition of ice cores from both poles (Zielinski, 2000; Castellano et al., 2005) confirm that volcanic forcing amplitude and occurrence varied significantly during the Holocene (see also section 6.5). Fluctuations of cosmogenic isotopes (ice core ^{10}Be and tree ring residual ^{14}C) have been used as proxies for Holocene changes in solar activity (e.g., Bond et al., 2001), but substantial work remains to be done to disentangle solar from non-solar influences on these proxies over the full Holocene (Muscheler et al., submitted). Residual continental ice sheets formed during the last ice age were retreating during the first half of the current interglacial period, inducing a significant sea-level increase until just after 6 ka BP. The associated ice sheet albedo is thought to have locally modulated the regional climate response to the orbital forcing (e.g., Davis et al., 2003).

[INSERT FIGURE 6.5 HERE]

The evolution of atmospheric trace gases during the Holocene is well known from ice core analyses (Figure 6.6). A first decrease in atmospheric CO_2 of about 7 ppm from 11 to 8 ka is followed by a 20 ppm CO_2 increase until the onset of the industrial revolution (Monnin, 2004). Atmospheric methane decreased from a Northern Hemisphere value of ~730 ppb around 10 ka to about 580 ppb around 6 ka, and increased again slowly to the same level at preindustrial times (Chappellaz et al., 1997; Flückiger, 2002) Atmospheric N_2O largely followed the evolution of atmospheric CO_2 and shows an early Holocene decrease of about 10 ppb and an increase of the same magnitude between 8 and 2 ka (Flückiger, 2002). Implied radiative forcing changes from Holocene greenhouse gas variations are 0.4 W m^{-2} (CO2) and 0.1 W m^{-2} (N_2O and CH_4), relative to preindustrial.

[INSERT FIGURE 6.6 HERE]

How and why did Holocene atmospheric greenhouse gases vary before the industrial period?

Recent transient carbon cycle-climate model simulations with a predictive global vegetation model have attributed the early Holocene CO_2 decrease to forest regrowth in areas of the waning Laurentide ice shield, partly counteracted by ocean sediment carbonate compensation (Joos et al., 2004b). Carbonate compensation of terrestrial carbon uptake during the glacial-interglacial transition and the early Holocene, as well as coral reef build-up during the Holocene, have likely contributed to the subsequent CO_2 rise (Broecker and Clark, 2003; Ridgwell et al., 2003; Joos et al., 2004b), whereas recent carbon isotope data (Eyer, 2004) and model results (Brovkin et al., 2002; Kaplan et al., 2002; Joos et al., 2004b) suggest that the terrestrial carbon inventory has been rather stable over the past 7000 years. Such natural mechanisms cannot account for the much more significant industrial trace gas increases; atmospheric CO_2 would have remained well below 290 ppm in the absence of anthropogenic emissions (Gerber et al., 2003b).

It has been hypothesized, based on Vostok ice core CO_2 data (Petit et al., 1999a), that atmospheric CO_2 would have dropped naturally by 20 ppm during the Holocene (in contrast with the observed 20 ppm increase), just as it did during the previous three glacial-interglacial cycles, if human activities had not caused a release of terrestrial carbon and methane during the Holocene (Ruddiman, 2003; Ruddiman et al., 2005); this hypothesis also suggests that incipient late Holocene high-latitude glaciation was prevented by these pre-Industrial greenhouse gas emissions. However, this hypothesis is in conflict with several, independent lines of evidence, including the analogy of the three previous interglacials with the Holocene (see Section 6.3.1). It requires much larger changes in the Holocene atmospheric stable carbon isotope ratio ($^{13}C/^{12}C$) than found in ice cores (Eyer, 2004) as well as a carbon release by anthropogenic land use that is larger than estimated by comparing carbon storage for natural vegetation and present day land cover (Joos et al., 2004b). The combination of Holocene orbital forcing (Figure 6.5) and the trace-gas forcing postulated by (Ruddiman, 2003) also fails to induce a glaciation in two climate models (Crucifix et al., 2005; Claussen et al., 2005).

Was any part of the current interglacial period warmer than today?

The temperature evolution over the Holocene has been established for many different regions (Figure 6.7). In the North Atlantic and adjacent Arctic, there was a tendency for temperature maxima to be earlier and shorter with increasing latitude, pointing to the direct influence of the summer insolation maximum on sea ice extent (Koç, 1994; Kim et al., 2004). Climate reconstructions in the mid-northern latitudes exhibit a long-term decline in SST from the warmer early- to mid-Holocene to the cooler late-Holocene pre-industrial period (Johnsen et al., 2001; Marchal et al., 2002; Andersen et al., 2004; Kim et al., 2004), most likely in response to annual mean and summer orbital forcings at these latitudes. Near ice sheet remnants in northern Europe or western North America, peak warmth is locally delayed, probably as a result of the interplay between ice elevation, albedo, atmospheric and oceanic heat transport and orbital forcing (MacDonald, 2000; Kaufman et al., 2004). The warmest period in northern Europe and western north America occurs from 7 to 5 ka (Davis et al., 2003; Kaufman et al., 2004). During this mid-Holocene period, global pollen-based reconstructions (Prentice, 1998; Prentice, 2000) show a widespread northward expansion of northern temperature forest (Bigelow et al., 2003; Kaplan et al., 2003), as well as substantial glacier retreat (Box 6.3). Other early warm periods are identified in China (He et al., 2004) and Antarctica (Masson et al., 2000). At high southern latitudes, the early warm period cannot be accounted for by local summer insolation changes, suggesting that a large-scale reorganisation of latitudinal heat transport may have been responsible. In contrast, tropical Atlantic and east Pacific SSTs exhibit a progressive warming from the beginning of the current interglacial onwards, (Rimbu et al., 2004; Stott et al., 2004), possibly a reflection of annual mean insolation change (Figure 6.5). All together, paleoclimatic records of the Holocene provide no evidence for globally synchronous warm periods.

[INSERT FIGURE 6.7 HERE]

When forced by mid-Holocene orbital parameters, state-of-the-art coupled climate models capture observed regional temperature and precipitation changes, whereas simulated global mean temperatures remain essentially unchanged, just as expected from the seasonality of the orbital forcing (Y. Wang et al., 2005a; Masson-Delmotte et al., submitted). PMIP2 simulations suggest that the global temperature changed by +0.06°C to -0.40°C (relative to present) between mid-Holocene and pre-industrial periods (Masson-Delmotte et al., submitted). It is obvious that there were places, seasons and periods in the Holocene where local temperature was likely as warm as or warmer than at the end of the 20th century. However, these warm periods were not of global scale, nor consistent through seasons, in contrast to the observed post-industrial warming. Due to different regional temperature responses from the tropics to high latitudes, as well as between hemispheres, commonly used concepts such as “mid-Holocene thermal optimum,” “Altithermal,” etc. should abandoned, or only be applied in a well-articulated regional context.

[START OF BOX 6.3]

Box 6.3: Holocene Glacier Variability

The near global retreat of mountain glaciers is among the most visible evidence for 20th century climate change (see Chapter 4), and the question arises as to the significance of this current retreat within a longer time perspective. A balance between temperature, solid precipitation and prevailing wind directions usually determines the presence of glaciers, and non-linear influences on glacier mass balance have been demonstrated between summer temperature and winter precipitation (Ahlmann, 1924). This is important in a paleoperspective because either of these factors can be reconstructed if the other is known (Dahl and Nesje, 1996). The climatic forcing that causes an advance, or a retreat, may thus be different for glaciers located in different climate regimes. This distinction is crucial if past reconstructions of glacier activity are to be understood properly.

In most mountain regions, including those in the North and South America, the Arctic, Central Asia, tropical Africa, and New Zealand, records documenting past glacier variations exist as discontinuous series. This reduces the possibility of investigating both the magnitude and rate of change associated with a specific glacier advance or retreat. Continuous records hence provide the most detailed and coherent information on Holocene glacier variations; these types of records are mainly available for the European Alps (e.g., Leemann and Niessen, 1994), Scandinavia (e.g., Nesje et al., 2005), and North America (e.g., Leonard and Reasoner, 1999) (Box 6.3, Figure 1).

[INSERT BOX 6.3, FIGURE 1 HERE]

Northern Scandinavia experienced continuous mountain glaciation up to 10.4–10.3 ka, with a later readvance between 9.8–9.4 and 9.3–8.9 ka (Bakke et al., 2005a). These early Holocene expansions may have been associated with freshwater perturbations of the Nordic Seas, including the Preboreal Oscillation and the 8.2 ka event (see Section 6.4.2). Between 8.8 and 3.8 ka the glaciers of Northern Scandinavia were less extensive than today, and many were most likely melted away. The glaciers reformed some time after 4 ka, marking the onset of “Neoglaciation” in the Northern Hemisphere, a development that culminated in a late ‘Little Ice Age’ (LIA) advance at the beginning of the 20th century. Southern Scandinavia has numerous continuous glacial records covering maritime to continental climate regimes (Nesje et al., 2005). The pattern of Northern Scandinavia is partly mirrored in the South, where glaciers reformed after 6 ka, and exhibited a general increasing mass balance trend towards a LIA advance around 1750 A.D. The higher mountains of Southern Scandinavia may explain the somewhat earlier mid-Holocene glacial inception. The data from the European Alps reveal high-frequency glacier variations with similar amplitude of advance throughout the second half of the Holocene, as well as two periods of glacier recession between ca. 7.5 to 5.2 ka, and 4.5 to 3.8 ka. In the Southern part of the Alps, some glaciers were absent or very small from 9.4 to 3.4 ka when glaciers began to grow again until a LIA maximum between 1790 and 1870 A.D. (Leemann and Niessen, 1994).

Glaciers on arctic Svalbard did not re-appear before ca. 5–4 ka, reaching their maximum extension during the "Little Ice Age" as late as in the 1920s (Svendsen and Mangerud, 1997). Between 2 and 1 ka some of the glaciers on Svalbard were substantially smaller than today (Humlum et al., 2005), or absent (Snyder et al., 2000). Moraine chronologies from Franz Josef Land (Lubinski et al., 1999) show a similarity with the Svalbard glacier history, as do data from the Baffin Islands (Miller et al., 2005). Elsewhere in North America, several separate glacier advances are recorded in the early and mid Holocene (e.g., Barclay et al., 2001), with increased glacier activity around 6, 4, and 3 ka, as well as during the LIA. Through the last millennium, the extra-tropical regions of the North and South America show similar glacier behaviour (Luckman and Villalba, 2001). In the tropical Andes, glaciers were relatively small during the Early Holocene, until they began to expand between 5.2 and 4.8 ka. A few retreating glaciers are now exposing mid-Holocene plant fossils allowing a more precise dating control of this glacier expansion (Mark et al., 2002; Thompson, 2004). There is only scarce information on the African glacier history, but ice cores retrieved from the Kilimanjaro ice cap reveal that the current retreat is unprecedented in the Holocene (Thompson et al., 2002).

What do glaciers tell us about climate change during the Holocene?
Most glacier archives from the Northern Hemisphere and the tropics show small or demised glaciers between 9.0 and 6.0 ka, whereas during the second half of the Holocene glaciers reformed and expanded. The main forcing of this tendency is related to changes in summer and winter insolation due to the configuration of orbital parameters (see main text). Transient glacier changes superimposed on this trend suggest significant reorganizations of the climate system. On shorter timescales, climate phenomenon such as the North-Atlantic Oscillation (NAO) and El Niño/Southern Oscillation (ENSO) have profound impact on glaciers mass balance, explaining some of the discrepancies found between regions (Kovanen, 2003). This is exemplified in the anti-phasing between glacier records from the Alps and Scandinavia (Six et al., 2001), but is also illustrated by the recent growth of glaciers in western Norway during the last decades – in contrast to the overall global glacier retreat. The latter example is probably the result of increased winter accumulation following an elevated state of the NAO during the 1990s.

[END OF BOX 6.3]

What is the significance of changing monsoon strength over the Holocene?
Terrestrial records for the Holocene indicate large changes in precipitation associated with the monsoons over Asia and northern Africa. The mid-Holocene systematic shift of Sahelian vegetation belts (Jolly, 1998) requires increases in precipitation of 150–300 mm/yr from 18°N–30°N to support the replacement of desert by steppe vegetation (Braconnot, 2004). Atmospheric general circulation models were first used to explore the response of monsoons to changes of orbital parameters. Because of intensified summer land-sea contrasts, atmosphere-only models were able to capture a northward shift of the desert-steppe transition, but underestimated the northward extent of the African monsoon belt and the magnitude of the precipitation change required for a "green Sahara" (Joussaume et al., 1999; Coe and Harrison, 2002).

In coupled atmosphere-ocean simulations, the ocean response to the orbital forcing modifies the seasonal cycle of land-ocean contrasts over the Atlantic and Africa. The ocean component of the climate system possesses the thermal inertia required to integrate the annual mean obliquity-driven insolation forcing (Clement et al., 1999; Braconnot, 2004; Z. Liu, Harrison, SP, Kutzback, JE & Otto-Bleisner, B., 2004; Zhao and Liu, submitted). The late summer and autumn warming of the surface ocean in response to the insolation forcing enhances the land-sea contrast and affects both the Indian and African monsoons. In Africa, the timing and length of the monsoon season is more consistent among the coupled simulations (Zhao and Liu, submitted), and changes in annual mean precipitation simulated by coupled models over West Africa are in better agreement with data reconstructions (Braconnot, 2004). The late surface warming in autumn also favours increased precipitation in the northeastern Indian Ocean as the monsoon retreats from the continent to the ocean. This ocean feedback causes a further northward expansion of the monsoon precipitation belt and lengthens the simulated monsoon season (Braconnot, 2004). Coupled atmosphere-ocean-vegetation models further include moisture and albedo feedbacks of vegetation and soils; they are now able to simulate the mid-Holocene growth of sparse grasses from the Sahel to the Mediterranean coast in agreement with paleobotanical evidence (Levis, 2004). Climate models therefore enable the disentangling of the climate response to the orbital forcing in generating more intense monsoons during the Holocene, and also reveal the amplifying roles of ocean and land surface feedbacks in the response of monsoons to altered climate forcing.

What do we learn from equilibrium simulations in terms of climate response to the orbital forcing?
Since the IPCC TAR, coupled atmosphere-ocean models, including the recent PMIP-2 simulations, have investigated the response of the climate system to orbital forcing for 6 ka BP during the mid-Holocene (Table 6.1). These models indicate a small global annual mean surface temperature change (Masson-Delmotte et al., submitted). Ocean changes simulated for this period are also generally small and difficult to quantify from data due to uncertainties in the way proxy methods respond to the seasonality and stratification of the surface waters (Waelbroeck et al., 2005). However, models can be evaluated for robust regional continental changes, including high-latitude warmth and hydrologic changes across Africa, Asia, and North America.

Terrestrial records of the mid-Holocene indicate an expansion of forest at the expense of tundra at mid- to high-latitudes of the Northern Hemisphere (Prentice, 2000). Atmosphere-alone models cannot simulate asymmetries in the mid-Holocene tundra-taiga boundary (Kohfeld and Harrison, 2000). Fully coupled atmosphere-ocean models do produce the northward shift in the position of the northern limit of boreal forest, in response to simulated summer warming, and the northward expansion of temperate forest belts in North America, in response to simulated winter warming (Wohlfahrt et al., 2004). At high latitudes the vegetation and ocean feedbacks enhance the warming in spring and autumn, respectively. Observed mid-continental drying in North America is also simulated, but overestimated in Eurasia. Simulations with atmosphere-only models indicate that a change in the mean tropical Pacific SSTs in the mid-Holocene to more La-Niña-like conditions can explain North American drought conditions at mid-Holocene (Shin et al., 2005).

Based on proxies of SST in the North Atlantic, it has been suggested that trends from early to late Holocene are consistent with a shift from a more meridional regime over northern Europe or positive NAO-like mean state in the early to mid Holocene (Rimbu et al., 2004). Results from PMIP-2 models find that six of nine models support a positive NAO-like shift in the mean state for the mid-Holocene as compared to pre-industrial. More positive NAO variability is not supported by the simulations (Gladstone et al., submitted).

What is the long-term contribution of polar ice sheet meltwater to the observed modern sea level rise?
Holocene sea-level observations and models (see Section 6.3.3) can be used to assess whether or not a significant part of the observed 2 mm/yr sea-level rise during the 20th century could be explained by long-term effect of the last deglaciation on polar ice sheets. Global models of isotostatic adjustment reveal that long term changes in Antarctic and Greenland ice sheets cannot contribute more than 0.5 mm/yr to the modern sea level increase of 2 mm/yr (Peltier and Solheim, 2002).

Are there long-term modes of climate variability identified during the Holocene that could be involved in the observed current warming?
An increasing number of Holocene proxy records are of sufficiently high resolution to describe the climate variability on centennial to millennial time scales, and to identify possible natural quasi-periodic modes of climate variability these time scales (Haug et al., 2001; A.K. Gupta et al., 2003). Although earlier studies suggested that Holocene millennial variability could display similar frequency characteristics as the glacial variability in the North Atlantic (Bond et al., 1997), this assumption has been questioned (Risebrobakken et al., 2003; Schulz et al., 2004; Moros et al., submitted). The suggested synchroneity of tropical and north Atlantic centennial to millennial variability (de Menocal et al., 2000; Mayewski et al., 2004; Wang et al., 2005b) is also not common to the full globe, as revealed by millennial scale variability in the southern hemisphere (Masson et al., 2000; Holmgren et al., 2003). Some authors argue that the Little Ice Age (LIA; ca. 1500–1850 AD) is the most recent cold phase of such natural fluctuations (Nesje, 2000). In several regions, such as Alaska, Spitzbergen, and parts of American Cordillera, the LIA advances were the most extensive of the Holocene, whereas in other regions, especially in the Southern Hemisphere, previous advances were larger (Grove, 2004; Box 6.3). In most records, multidecadal to millenial variability has relatively low amplitude, smaller than that observed for the last century, and therefore cannot account for the full amplitude of 20th century warming.

Based on the correlation between changes in atmospheric concentrations of cosmogenic isotopes (^{10}Be or ^{14}C) and climate proxy records, some authors argue that solar activity may be the driver for centennial to millennial variability (Karlen, 1996; Bond et al., 2001; Fleitmann et al., 2003; Wang et al., 2005b). The importance of (forced or unforced) modes of variability within the climate system, for instance related to the deep ocean circulation, has been pointed out (Bianchi and McCave, 1999; Duplessy et al., 2001; Marchal et al., 2002; Oppo et al., 2003). In many records there is no apparent consistent pacing at specific centennial to millennial frequencies through the Holocene period, but rather shifts between different frequencies (Moros et al., submitted). The current lack of consistency between various data sets makes it difficult, based on current knowledge, to attribute the century and longer time scale climate variations to solar variability, episodes of intense volcanism, or simple modes of variability internal to the climate system.

6.4.2 Abrupt Climate Change During the Current Interglacial

What do abrupt changes in oceanic and atmospheric circulation at mid- and high-latitudes tell us?

At the beginning of the Holocene, approximately 11,600 years ago, significant residual continental ice cover still existed in the Northern Hemisphere. Significant volumes of fresh water were also impounded in pro-glacial lakes adjacent to the remnants of this ice. In particular, the residual ice cover over the North American continent, together with adjacent pro-glacial Lake Agassiz to the southwest is believed to have been responsible for the occurrence of one of the most dramatic events of the early Holocene, namely the "8.2 ka event" that was first recognized as a prominent feature in the Summit, Greenland ice cores and other records (Alley et al., 1997). This event is widely believed to have occurred during a cooling period (Rohling and Palike, 2005) and as a consequence of an "outburst flood" during which Lake Agassiz drained into Hudson Bay (Renssen et al., 2001; Nesje et al., 2004), which by that time was entirely free of ice cover. The 8.2 ka event is recorded as a brief adjustment of the Atlantic meridional overturning circulation (Bianchi and McCave, 1999; Risebrobakken et al., 2003; McManus et al., 2004), as well as a cooling of the North Atlantic region identified in Greenland and Europe (Klitgaard-Kristensen et al., 1998; von Grafenstein et al., 1998; Barber et al., 1999; Nesje, 2000; McDermott et al., 2001). The magnitudes and rates of response are estimated to be 6°C, with most of this cooling over 5 years, in Greenland (Alley et al., 1997). There was an associated decrease in precipitation or runoff, taking 30–40 years to reach a minimum, in Northern South America (Hughen et al., 1996).

A large decrease in methane (several tens of ppb) (Spahni et al., 2003) reveals the widespread consequences of the abrupt 8.2 ka event associated with large scale atmospheric circulation change recorded from polar northern hemisphere to the tropics (Stager and Mayewski, 1997; Haug et al., 2001; Fleitmann et al., 2003). Simulations conducted with intermediate complexity climate models (see Alley and Agustsdottir, 2005 for a review) point to a southward shift of the NADW, associated with equilibrium states of the models (Bauer et al., 2004), and to responses to the freshwater forcing depending on the specific model's high frequency variability (Renssen et al., 2002). Within PMIP2, standardized freshwater forcing experiments are being conducted to assess the vulnerability of the ocean and atmospheric circulation to a large freshwater release under different mean states.

The end of the first half of the Holocene – between ca. 5,000 and 4,000 ka – is punctuated by rapid events at various latitudes, such as an abrupt change in northern hemisphere sea-ice cover (Jennings et al., 2001), transport of moisture to central Greenland (Masson-Delmotte et al., in press), change in European climate (Seppa and Birks, 2001; Lauritzen, 2003), widespread century-scale North American drought (Booth et al., 2005), and changes in South American climate (Marchant and Hooghiemstra, 2004). These observations clearly show that under gradual climate forcings (here, orbital) the climate system can change abruptly. The processes behind these observed abrupt shifts are not well understood, highlighting the possibility that the future could include unanticipated abrupt climate change.

What is the significance of past abrupt monsoon change?

In the tropics, precipitation-sensitive records and models indicate that summer monsoons in Africa, India and southeast Asia were enhanced in the early- to mid-Holocene due to orbital forcing, a resulting increase in land-sea temperature gradients, and displacement of the intertropical convergence zone. All high resolution precipitation-sensitive records reveal that the transitions (not synchronous) from wetter conditions in the early Holocene to drier modern conditions occurred in one or more abrupt steps (Guo et al., 2000; Fleitmann et al., 2003; Morrill et al., 2003; Y. Wang et al., 2005b).

In the early Holocene, large increases in African monsoon precipitation and/or wetter conditions over the Mediterranean are associated with dramatic changes in Mediterranean Sea ventilation, as evidenced by sapropel layers (Ariztegui, 2000) . Transient simulations of the Holocene have also been performed with models of intermediate complexity, although usually after the final disappearance of ice sheets. These models have pointed to the operation of mechanisms that can generate rapid events in response to orbital forcing, such as changes in African monsoon intensity due to nonlinear interactions between vegetation and monsoon dynamics (Claussen et al., 1999; Renssen et al., 2003).

How and why has ENSO changed over the present interglacial?

Paleoclimate records clearly indicate fundamental changes in ENSO during the Holocene; model simulations suggest that these changes resulted from altered radiative forcing and background climate states. Data from diverse sources (corals, archaeological middens, and lake and ocean sediments) indicate that the early-mid Holocene experienced weak ENSO variability, with a transition to a stronger modern regime occurring in the past few thousand years (Shulmeister and Lees, 1995; Gagan et al., 1998; Rodbell et al., 1999; Tudhope et al., 2001; Moy et al., 2002). Most data sources are discontinuous, providing only snapshots of mean conditions or interannual variability, and making it difficult to precisely characterize the rate and timing of the transition to the modern regime. Fossil coral records from New Guinea indicate clearly that interannual variability was weaker between about 5.4–7.7 ka than it was at 1.7–2.7 ka (Tudhope et al., 2001; McGregor and Gagan, 2004a, 2004b). A continuous lake record from Ecuador, which tracks the rainstorms characteristic of strong El Niño events, notes a stepped increase in the frequency of such events at around 7000 and at 5000 years ago (Moy et al., 2002). Pollen records from Australasia reveal the absence of drought-adapted taxa before 5ky BP; drought is a regular feature of El Niño events there today (Shulmeister and Lees, 1995).

Paleoclimate model simulations using models of varying complexity support a mechanism by which orbital forcing leads to a weakening of ENSO variability. A simple model of the coupled Pacific ocean-atmosphere, forced with orbital insolation variations, suggests that seasonal changes in insolation can produce systematic changes in ENSO behavior. Key elements of the Holocene ENSO response are the Bjerknes feedback mechanism (Bjerknes, 1969) and ocean dynamical thermostat (Clement et al., 1996; Clement et al., 2000; Cane, 2005). These studies indicate a progressive, somewhat irregular, increase in both event frequency and amplitude throughout the Holocene. Coupled general circulation models also reproduce the intensification of ENSO over the Holocene, although with some disagreement as to the magnitude of change. Both model results and data syntheses suggest that before the mid-Holocene, the tropical Pacific exhibited a more La Niña-like background state (Clement et al., 2000; Liu et al., 2000; Kitoh and Murakami, 2002; Otto-Bliesner et al., 2003; Z. Liu, Harrison, SP, Kutzback, JE & Otto-Bleisner, B., 2004). In paleoclimate simulations with general circulation models, ENSO teleconnections robust in the modern system show signs of weakening under changed forcing and background state (Otto-Bliesner, 1999; Otto-Bliesner et al., 2003).

6.5 The Last 2000 Years

6.5.1 Northern Hemisphere Temperature Variability

What do reconstructions based on palaeoclimate proxies tell us?

Figure 6.8 illustrates various instrumental and proxy-climate evidence of the variations in average large-scale surface temperatures during the last two millennia. Figure 6.8a shows two instrumental compilations representing the mean annual surface temperature of the Northern Hemisphere since 1850, one based on land data only, and one using land and surface ocean data combined (see Chapter 3). The uncertainties associated with one of these series are also shown (30-year smoothed combined land and marine). These arise primarily from the incomplete spatial coverage of instrumentation through time (Jones et al., 1997), and while these uncertainties are larger in the 19th compared to the 20th century, the significance of the recent warming, especially in the last two to three decades of the record, is clearly apparent in this 150-year context. The land-only record shows similar variability, although the rate of warming is greater than in the combined record after about 1980. The land-only series can be extended back beyond the 19th century, and is shown plotted from 1781 onwards. The early section is based on a much sparser network of available station data with at least 23 European, but only one North American station spanning the first two decades and the first Asian station beginning only in the 1820s. Four European records (Central England, De Bilt, Berlin and

Uppsala) provide an even longer, though necessarily regionally-restricted, indication of the context for the warming observed in the last ~20–30 years, which is even greater in this area than is observed over the Northern Hemisphere land as a whole.

[INSERT FIGURE 6.8 HERE]

The instrumental temperature data that exist before 1850, although increasingly biased towards Europe in earlier periods, show that the warming observed after 1980 is unprecedented compared to the levels measured in the previous 280 years. A recent analysis of instrumental, documentary and proxy climate records focusing on European temperatures has also shown that the extreme summer of 2003 was very likely warmer than any that has occurred in at least 500 years.

If the behavior of recent temperature change is to be understood, and the mechanisms and causes correctly attributed, parallel efforts are needed to reconstruct the longer, pre-instrumental, history of climate variability, as well as the detailed changes in the various factors that might influence climate (Bradley et al., 2003; Jones and Mann, 2004).

The TAR discussed various attempts to use multiple proxy data to reconstruct changes in the average temperature of the Northern Hemisphere for the period after A.D. 1000, but focused on three reconstructions, all with yearly resolution. The first (Mann et al., 1999), represents mean annual temperatures, and is based on a range of proxy types, including data extracted from tree rings, ice cores and historical documentary sources; this reconstruction also incorporates a number of instrumental (temperature and precipitation) records from the 18th century onwards. For 900 years, this series exhibits multi-decadal fluctuations with amplitudes up to 0.3°C superimposed on a negative trend of 0.15°C, followed by an abrupt warming (~0.4°C) matching that observed in the instrumental data during the first half of the 20th century. Of the other two reconstructions, one (P.D. Jones et al., 1998) was based on a very much smaller number of proxies, whereas the other (Briffa et al., 2001; Osborn et al., 2005) was based solely on tree-ring density series from an expansive area of the extra-tropics, but reached back only to AD 1400. These two reconstructions emphasise warm season rather than annual temperatures, with a geographical focus on extra-tropical land areas. They indicate a greater range of variability on centennial timescales prior to the 20th century, and also suggest slightly cooler conditions during the 17th century than those portrayed in the Mann et al. (1998; 1999) series.

Following the emphasis placed on it in the TAR, the "hockey stick" reconstruction of Mann et al. (1999) has been the subject of several critical studies. Soon and Baliunas (2003) and Soon et al. (2003) attempted to challenge the conclusion that the 20th century was the warmest on a hemispheric average scale. They listed a qualitative catalogue of regionally diverse proxy climate evidence, either for relatively warm (or cold), or alternatively dry (or wet) conditions occurring at any time within pre-conceived periods assumed to bracket the Medieval Warm Period (Little Ice Age). Their approach precluded any quantitative summary of the evidence at precise times, and their assumption that relative dryness can be equated directly with warmth regardless of season or location, severely limits the value of their review as a basis for comparison of the relative magnitude of mean Hemispheric 20th-century warmth (Mann and Jones, 2003).

[START OF BOX 6.4]

Box 6.4: The Medieval Warm Period

At least as early as the beginning of the 20th century, different authors were already disputing the evidence for climate changes during the last two millennia, particularly in relation to North America, Scandinavia and Eastern Europe (Brooks, 1922) . With regard to Iceland and Greenland, Pettersson (1914) cites evidence by Rabot of considerable areas of Iceland being cultivated in the 10th century. At the same time, Norse settlers colonized areas of Greenland, while a general absence of sea ice allowed regular voyages at latitudes far to the north of what became possible in the colder 14th century. Brooks (1922) describes how, after some amelioration in the 15th and 16th centuries, conditions worsened considerably in the 17th century; in Iceland previously cultivated land was covered by ice. Hence, at least for the area of the northern North Atlantic, a picture was already emerging of generally warmer conditions around the centuries leading up to the end of the 1st millennium, but framed largely by comparison with strong evidence of much worse conditions in later centuries, particularly the 17th century.

Lamb (1965) seems to have been the first to coin the phrase "Medieval Warm Epoch" or "Little Optimum" to describe the totality of multiple strands of evidence principally drawn from western Europe, for a period of widespread and generally warmer temperatures which he put at between AD 1000 to 1200 (Lamb, 1982). It is important to note that Lamb also considered the warmest conditions to have occurred at different times in different areas: between 950 to 1200 in European Russia and Greenland, but somewhat later, between 1150 to 1300 (though with notable warmth also in the later 900s) in most of Europe (Lamb, 1977).

Much of the evidence used by Lamb was drawn from a very diverse mixture of sources such as historical anecdotes, evidence of vegetation changes, or records of the cultivation of cereals and vines. He also drew inferences from very preliminary analyses of some Greenland ice core data and European tree-ring records. Much of this evidence is difficult to interpret in terms of accurate quantitative temperature influences. Much is not precisely dated, or results from physical or biological systems that involve complex lags between forcing and response, as is the case for vegetation and glacier changes. Lamb's analyses also predate any formal statistical calibration of much of the evidence he considered. Largely on the basis of summer temperature inferences, he concluded that "High Medieval" temperatures were probably 1.0°C to 2.0°C above early 20th century levels at various European locations (Lamb, 1977; Bradley et al., 2003b).

A later study, based on examination of more quantitative evidence, in which efforts were made to control for accurate dating and specific temperature response, concluded that it was not possible to say anything other than "… in some areas of the Globe, for some part of the year, relatively warm conditions may have prevailed" (Hughes and Diaz, 1994).

In medieval times, as now, climate was unlikely to have changed in the same direction or by the same magnitude everywhere (Box 6.4, Figure 1). At some times, some regions may have experienced even warmer conditions than those that prevailed throughout the 20th century (e.g., see Bradley et al., 2003b). Regionally restricted evidence by itself, especially when the dating is imprecise, is of little practical relevance to the question of whether climate in medieval times was as warm or warmer than today. To define medieval warmth in a way that has more relevance for exploring the causes of recent large-scale warming, widespread and continuous paleoclimate evidence must be assimilated in a homogeneous way and scaled against recent measured temperatures to allow a meaningful quantitative comparison against 20th century warmth (Figure 6.8).

[INSERT BOX 6.4, FIGURE 1 HERE]

Local climate variations can be dominated by internal climate variability, often the result of the redistribution of heat by regional climate processes. Only very large-scale climate averages can be expected to reflect global forcings over the last 2000 years (Mann and Jones, 2003a; Goosse et al., 2005a). Studies that have attempted to do this have invariably come to the same conclusion: that medieval warmth was complex in terms of its precise timing and regional expression (Crowley and Lowery, 2000a; Folland et al., 2001; Esper et al., 2002; Bradley et al., 2003b; Jones and Mann, 2004).

The uncertainty associated with present paleoclimate estimates of Northern Hemispheric mean temperatures are significant, especially for the period prior to 1600 when data are scarce (Mann et al., 1999; Briffa and Osborn, 2002; Cook et al., 2004a; Osborn et al., 2005). However, Figure 6.8 shows that the warmest period prior to the 20th century, occurred between 950 and 1100 (between 0.1°C and 0.2°C below the 1961–1990 mean), noticeably below the levels reached in the last decade.

It is certain that further work is necessary to produce many more paleoclimate series with much wider geographic coverage. There are far from sufficient data to make any meaningful estimates of *global* medieval warmth. There are very few long records with high temporal resolution data from the oceans, the tropics or the Southern Hemisphere.

The best evidence currently available indicates that Northern Hemisphere mean temperatures during the High Medieval time (950–1000) were indeed warm in a 2000-year context and even warmer in relation to the evidence of widespread average cool conditions in the 17th century. However, this evidence does not support a conclusion that hemispheric mean temperatures were as warm as those in the late 20th century, during any period in medieval times (Jones et al., 2001; Bradley et al., 2003a; Bradley et al., 2003b).

[END OF BOX 6.4]

McIntyre and McKitrick (2003), produced a Northern Hemisphere reconstruction that differs radically from that of Mann et al. (1999), in indicating a period of significant warmth in the 15th century, even though they attempted to employ the same method and candidate proxy climate predictors. However, they omitted several important proxy series used in the original reconstruction, and arrived at a regression model which did not 'verify', in the sense that it produced temperature estimates that did not agree with independent temperature observations sufficiently well to demonstrate any likely validity in their final reconstruction. The Mann et al. (1999) series was subsequently successfully reproduced by Wahl and Ammann (2004).

Since the TAR, a number of additional proxy data syntheses based on annual or near-annually resolved data, variously representing mean Northern Hemisphere temperature changes over the last one or two thousand years, have been published (Esper et al., 2002; Crowley et al., 2003; Mann and Jones, 2003; Cook et al., 2004a; Briffa et al., 2005; Moberg et al., 2005; Rutherford et al., 2005; D'Arrigo et al., submitted; Hegerl et al., submitted). These are shown in Figure 6.8b, along with the three series from the TAR. As with the original TAR series, these new records are not entirely independent reconstructions inasmuch as there are some predictors (most often tree-ring data) that are common between them, but, in general, they represent some expansion in the length and geographical coverage of the previously available data.

Mann and Jones (2003) selected only eight normalised series (all screened for temperature sensitivity) to represent annual mean Northern Hemisphere temperature change over the last 2000 years, though the majority of these eight represent integrations of multiple proxy site records or reconstructions. An average of these series was scaled to match the mean and standard deviation of Northern Hemisphere annual mean land and marine temperatures over the period 1856-1980.

Esper et al. (2002) took tree-ring data from 14 sites, and applied a variant of an existing statistical technique (Briffa et al., 1992) designed to produce ring-width chronologies in which evidence of long-timescale climate forcing is better represented compared with earlier tree-ring processing methods. The resulting series were averaged, smoothed and then simply scaled against the multi-decadal variability in the Mann et al. (1999) reconstruction over the period 1900-1977. This produced a reconstruction with markedly cooler temperatures during the 12th to the end of the 14th century than are apparent in any other series. The relative amplitude of this series is reduced somewhat if recalibrated directly against smoothed instrumental temperatures (Cook et al., 2004a), or by using annually-resolved temperature data (Briffa and Osborn, 2002), but even then, this reconstruction remains at the coldest end of the range defined by all current available reconstructions.

Two other recent studies (D'Arrigo et al., 2005a; Hegerl et al., submitted) include substantial amounts of similar tree-ring data as Esper et al. (2002) among their predictors, and scale them using statistical methods that enhance the expression of long-timescale variability in the reconstructions. Their estimates lie closer to the lower end of the range of temperatures from the various reconstructions during almost all of the last 1000 years. Moberg et al. (2005) combine the interannual-to-decadal variations in some selected tree-ring records with the multidecadal-to-century timescale variability in a number of independent, lower-resolution, but less-precisely-dated, records. Their reconstruction displays the warmest temperatures of any reconstruction during the 10th and early 11th centuries, though still below the level of warmth observed since 1980.

Briffa et al. (2005) produced an extended history of warm season temperatures based on many tree-ring width records from sites across northern Fennoscandia and northern Siberia, using improved statistical techniques for preserving multi-centennial timescale variability. Although ostensibly representative of northern Eurasian conditions, these data have also been regressed against Northern Hemisphere land data to provide estimates of summer temperature over 2000 years.

Recently, many of the individual proxy series used in the various reconstruction studies cited above have been combined in a new reconstruction (only back to AD 1400) based on a climate field reconstruction technique (Rutherford et al., 2005). This study also involved a methodological exploration of the sensitivity of the results to the precise specification of the predictor set, as well as the predictand target region and seasonal window. They concluded that the reconstructions were reasonably robust to changes in the choice of proxy data and statistical approach.

Various statistical methods are used to convert the various sets of original palaeoclimate proxies into the different estimates of mean Northern Hemisphere temperatures shown in Figure 6.8 (see discussions in Jones and Mann, (2004); Rutherford et al., (2005)). These range from simple averaging of regional data, and scaling of the resulting series so that its mean and standard deviation matches that of the observed record over some period of overlap (P.D. Jones et al., 1998; Crowley and Lowery, 2000), to complex climate field reconstruction, where large scale modes of spatial climate variability are linked to patterns of variability in the proxy network, via a multivariate transfer function that provides estimates of the spatio-temporal changes in past temperatures explicitly, and from which large-scale average temperature changes are derived by averaging the climate estimate across the required region (Mann et al., 1998; Rutherford et al., 2003). Other reconstructions can be considered to represent what are essentially intermediate applications of these two approaches, in that they involve an explicit regionalisation of much of the data prior to the use of a statistical transfer function, and so involve fewer, but potentially more robust, regional predictors (Briffa et al., 2001; Mann and Jones, 2003; Osborn et al., 2005). Some of these studies explicitly or implicitly reconstruct tropical temperatures based on data largely from the extratropics, and assume stability in the patterns of climate association between these regions. This assumption has been questioned on the basis of both observational and model simulated data suggesting that tropical–extratropical climate variability can be decoupled (Rind et al., 2005) and that extratropical teleconnections associated with ENSO may also vary through time (Section 6.5.6).

Changes in proxy records, either physical (such as the isotopic composition of various elements in ice) or certainly biological (such as the width of a tree ring or the chemical composition of a growth band in coral), do not respond precisely or solely to changes in any specific climate parameter (such as mean temperature or total rainfall), or to the changes in that parameter as measured over a fixed calendar-based time window (such as June-August or January-December). For this reason, the proxies must be ‘calibrated’ empirically, by comparing their measured variability over a number of years with available instrumental records to identify some optimal climate association, and to quantify the statistical uncertainty associated with scaling the proxy or proxies to represent this specific climate parameter. All reconstructions, therefore, involve a degree of compromise with regard to the specific choice of ‘target’ or dependent variable. Differences between the temperature reconstructions shown in Figure 6.8b are to some extent related to this, as well as to the choice of different predictor series (including differences in the way these have been processed). The use of different statistical scaling approaches (including whether the data are smoothed prior to rescaling, and differences in the period over which this rescaling is carried out) also influences the apparent spread between the various reconstructions. (Discussions of these issues can also be found in Harris and Chapman, (2001); Beltami, (2002b); Briffa and Osborn, (2002); Trenberth and Otto-Bliesner, (2003); Zorita et al., (2003); Jones and Mann, (2004); Esper et al., (2002; 2005); Pollack and Smerdon, (2004); and Rutherford et al., (2005)).

The considerable uncertainty associated with individual reconstructions (on the order of ±0.5°C two-standard-error limits on the multi-decadal timescale), is clearly expressed in several publications, calculated on the basis of analyses of regression residuals (Mann et al., 1998, 1999; Briffa et al., 2001; Jones et al., 2001; Mann and Jones, 2003; Osborn and Briffa, 2004; Rutherford et al., 2005). In virtually all cases, these are likely to be minimum uncertainties because they do not take into account other sources of uncertainty in the predictor series themselves, and at least in some cases, any possible limitations in the capacity of the regression techniques to capture variance on timescales longer than are represented in the calibration period (Briffa and Osborn, 1999; Esper et al., 2002; Bradley et al., 2003; Von Storch et al., 2004; Osborn et al., 2005; Zorita et al., 2005).

Figure 6.8b illustrates how the currently available reconstructions indicate generally greater variability in centennial time scale trends over the last 1000 years than was apparent in the TAR. However, given that the confidence levels surrounding all of the reconstructions are wide, virtually all reconstructions are effectively encompassed within the uncertainty previously indicated in the TAR. The major differences between the various proxy reconstructions relate to the magnitude of past cool excursions, principally during the 12th to

14th and 17th to 19th centuries: only one of the reconstructions, that of Moberg et al. (2005), during the early decades of both the 11th and 12th centuries, indicates hemispheric-scale conditions that were as warm as those in the 1940s and 50s, and none shows temperatures reaching the levels observed in the last two decades of the 20th century.

However, it is important to recognise that even in the Northern Hemisphere as a whole there are relatively few long and well-dated climate proxies, particularly for the period prior to the 17th century. Those that do exist are concentrated in extra-tropical, terrestrial locations, and many have greatest sensitivity to summer rather than winter (or annual) conditions. Changes in seasonality probably limit the conclusions that can be drawn regarding annual temperatures derived from predominantly summer-sensitive proxies (Jones et al., 2003a). There are very few strongly temperature sensitive proxies from tropical latitudes. Oxygen isotope series from high-elevation ice cores provide the longest records, but most represent changes in the source region of precipitation, as well as local temperature. However, very rapid and apparently unprecedented melting of tropical ice caps has been observed in recent decades (see Box 6.3) (Thompson et al., 2000; Thompson, 2001), possibly linked to sharply rising SST observed in the tropics after 1976 (Diaz and Graham, 1996), as well as enhanced warming at high elevations (Gaffen et al., 2000). Coral oxygen isotopes and Sr/Ca ratios primarily reflect SSTs, though they are also influenced by salinity changes associated with precipitation variability. Unfortunately, these records are invariably short, on the order of centuries at best, and can be associated with age uncertainties of 1 or 2%. Virtually all coral records currently available from the tropical Indo-Pacific indicate unusual warmth in the 20th century (Cole, 2003; Wilson and al., 2005), and in the tropical Indian ocean many records show a trend towards isotopically warmer conditions (Charles et al., 1997; Kuhnert et al., 1999; Cole et al., 2000). In most multi-centennial length coral series, the late 20th century is warmer than any time in the last 100-300 years.

The weight of new multi-proxy evidence, therefore, suggests a larger 20th century warming in comparison with temperature levels of the previous 400 years, than was shown in the TAR. On the evidence of the few new reconstructions that reach back across the first millennium of the Christian era, it is likely that 20th century warmth was unusual in a 2000-year context.

What do large-scale temperature histories from ground surface temperature measurements tell us?

Ground surface temperature (GST) histories reconstructed from direct measurements of subsurface temperatures in boreholes have been presented by several geothermal research groups (Pollack and Huang, 2000; Harris and Chapman, 2001; Pollack and Smerdon, 2004); see Pollack and Huang (2000) for a review of the methodology. These have all been derived using the contents of a publicly-available database of borehole temperatures (Huang and Pollack, 1998) that presently includes 695 sites in the Northern Hemisphere. These geothermal reconstructions provide independent estimates of surface temperature changes over the past five centuries with which to compare the other proxy reconstructions, but because the Earth acts as a low-pass filter on downward-propagating temperature signals, decadal and shorter fluctuations are generally unresolved in borehole reconstructions. The coupling between above and below ground temperatures is known to be complex and affected by factors such as ground vegetation and snow cover. Within the year, summer evapotranspiration can cool the ground relative to the surface air temperature; while in winter ground surface freezing can prevent the subsurface transfer of cold air temperature signals. On longer timescales, changing vegetation and snow cover are likely to result in complex and spatially varying biases in the way surface air temperature (SAT) is reflected in local GST reconstructions (Gosnold et al., 1997; Pollack and Huang, 2000; Hinkel et al., 2001; Kane et al., 2001; Sokratov and Barry, 2002; Lin et al., 2003; Mann and Schmidt, 2003; Stieglitz et al., 2003; Bartlett et al., 2004; Chapman et al., 2004; Smerdon et al., 2004, 2005). The few studies to date, using global simulations by three-dimensional coupled models, provide contradictory evidence of the likely accuracy of deep soil temperatures as an indicator of SAT on longer timescales. Gonzalez-Rouco et al. (2003) using the ECHO-G model, concluded that deep soil temperatures were indistinguishable from continental annual SAT, but in another simulation using GISS model E, Mann and Schmidt (2003) found a significant discrepancy between cold-season GST and SAT trends on multi-decadal timescales, linked to changing snow cover trends. Neither of the model simulations used in these studies included time-varying vegetation cover.

Figure 6.8 includes one reconstruction of average Northern Hemisphere GST, that by Pollack and Smerdon (2004). They state that this (like all geothermal reconstructions) shows a somewhat muted estimate of the real 20th-century trend, because about half of the borehole sites at the time of measurement (which varies by up to decades) had not yet been exposed to the significant warming of the last two decades of the 20th century. The one standard deviation uncertainties for their series (not shown here) are 0.1 (in 1500), 0.5 (1800) and 0.3 (1900) ºC. These are minimum errors (associated with various regional aggregations of local records), and do not take account of site specific noise in individual local site reconstructions (Pollack and Huang, 2000). This reconstruction indicates temperatures during the 16th and 17th centuries that coincide with the lower range of multiproxy reconstructions, values in the centre of the range during the 19th and early 20th century and an overall warming near to 1.0ºC over the last 500 years.

Not all hemispheric analyses of these borehole data give the same magnitude of warming (Rutherford and Mann, 2004), but all (Huang et al., 2000; Harris and Chapman, 2001; Beltrami, 2002a) are more consistent with the multi-proxy reconstructions that show the greatest warming during this period, and in less agreement with those that show least.

6.5.2 Southern Hemisphere Temperature Variability

There are markedly fewer well-dated proxy records for the SH compared to the NH, and consequently little evidence of how large-scale average surface temperatures have changed in the past centuries. Mann and Jones (2003) used only three normalised series to represent annual mean Southern Hemisphere temperature change over the last 2000 years, including a tree-ring-based/warm-season temperature reconstruction for western Tasmania back to 1600BC (Cook et al., 2000), and two ice-core oxygen isotope temperature series: for Law Dome, Antarctica, back to 100BC (van Ommen et al., 2005) and Quelccaya, Peru back to AD 470 (Thompson, 1992). The average of these series was scaled to match the mean and the standard deviation of Southern Hemisphere annual mean land-and-marine temperatures, over the period 1856–1980. The resulting reconstruction does show a distinct warming trend in the 20^{th} Century. However, the recent warmth (as shown in both instrumental and the proxy data) was exceeded during previous warm periods around AD 650–750 and 950–1000. The paucity of Southern Hemisphere proxy data also means that uncertainties associated with hemispheric temperature estimates are also much greater than for the Northern Hemisphere, and it is probably more appropriate at this time to consider the evidence in terms of limited regional indicators of temperature change (Figure 6.9).

[INSERT FIGURE 6.9 HERE]

Concentrating on the long-term oscillations in the warm-season temperature reconstruction for Tasmania (Cook et al., 2000), it is apparent that the warming trend over Tasmania is still a significant event when viewed in the context of multi-decadal variability covering the past 2000 years. Over that period, it remains the warmest event, but only to a marginal degree, and a much longer warm period is indicated in the period 900–1500 (Figure 6.9). The early 1900s period likewise remains a significant cold event, although it was apparently exceeded several times over the past 2000 years, with the coldest event occurring around AD 50. The 1500–1900 period is characterised by reduced multi-decadal variability.

A tree-ring reconstruction of Austral summer temperatures from the South Island of New Zealand, covering the past 1,100 years, is the longest yet produced for that region (Cook et al., 2002). Summer temperatures have warmed since 1900, with most of the warming occurring somewhat after 1950. This change is superimposed on a more general trend of increasing temperatures since the 16th Century. The rapid warming since 1950 is certainly unusual, but it is not statistically unprecedented when compared to the previous 1000 years of reconstructed temperatures (Figure 6.9), especially the warm interval indicated in the first half of the 13th Century. Tree-ring based temperature reconstructions across the Southern Andes (37–55°S) of South America indicate that the annual temperatures during the 20th century have been anomalously warm in the context of the past four centuries (Figure 6.9). The mean annual temperatures for northern and southern Patagonia during the interval 1900–1990 are 0.53°C and 0.86°C above the 1640–1899 means, respectively. In Southern Patagonia, the year 1998 was the warmest of the past four centuries (Villalba et al., 2003). The rate of temperature increase from 1850 to 1920 was the highest over the past 360 years.

Figure 6.9 also shows the evidence of ground surface temperature changes over the last 500 years, provided by regionally aggregated borehole temperature inversions, from southern Africa (92 records) and Australia (57 records). Within the limitations of their resolvable temporal resolution, these both indicate unusually warm conditions prevailing in the 20th century (Pollack and Smerdon, 2004).

Taken together, the very sparse evidence for Southern Hemisphere temperatures prior to the period of instrumental records indicates that it is likely that the warmth of the last 50 years is unusual in a 350 to 1000 year context.

6.5.3 Climate Forcing Over the Last Millennium

Figure 6.10 illustrates some of the major historical forcings used in a range of increasingly complex climate models to simulate Northern Hemisphere temperatures over the last 500-1000 years. These models include an energy balance formulation (Crowley et al., 2003), two- and three- dimensional, reduced complexity models (Bertrand et al., 2002; Bauer et al., 2003; Gerber et al., 2003a), and three fully coupled ocean-atmosphere general circulation models (Ammann et al., 2003; Von Storch et al., 2004; Tett et al., submitted).

Comparison and evaluation of the output from paleoclimate simulations is complicated by their use of different historical forcings, as well as by the way indirect evidence of the history of various forcings is translated into geographical and seasonally specific radiative inputs within the models. Some factors, such as orbital variations of the Earth in relation to the Sun can be calculated accurately (e.g. Berger, 1977; Bradley et al., 2003), and also directly implemented in terms of regional and seasonal insolation. For the last 2000 years, although they are incorporated in most models, these have not changed significantly and their impact on climate can be neglected as compared to the other forcings (Bertrand et al., 2002).

[INSERT FIGURE 6.10 HERE]

Over recent millennia, the analysis of the gas bubbles in high-deposition rate ice cores provides good evidence of greenhouse gas changes at near decadal resolution (Figure 6.6). Other factors, such as land-use changes (Ramankutty and Foley, 1999), and the concentrations and distribution of tropospheric aerosols and ozone, are not as well understood (Mickley et al., 2001). However, because of their magnitude, uncertainties in the history of solar irradiance and volcanic effects are more significant for the preindustrial period.

The direct measurement of solar irradiance by satellite began less than 30 years ago, and over this period only very small changes are apparent (0.08% between the peak and trough of recent sunspot cycles which equates to only a ~0.2 W m^{-2} change in radiative forcing). Earlier extensions of irradiance change used in most model simulations are estimated by assuming a direct correlation with evidence of changing sunspot numbers and cosmogenic isotope production as recorded in ice cores (^{10}Be) and tree rings (^{14}C) (J.L. Lean et al., 1995; Crowley, 2000).

There is generally reasonable-to-good temporal agreement between the different proxy records of solar activity such as cosmogenic isotopes, sunspot numbers or aurora observations and the annually-resolved records clearly depict the well-known 11-year solar cycle (Muscheler et al., 2005). For example, paleoclimatic ^{10}Be and ^{14}C values are higher during times of low or absent sunspot numbers. During these periods, their production is high as the shielding of the earth's atmosphere from cosmic rays provided by the solar magnetic field is weak (Beer et al., 1998). However, the relationship between sunsport numbers and solar magnetic field is not fully understood (Wang and Sheeley, 2003). In many climate simulations, the cosmogenic isotope production records have been scaled linearly to estimate solar energy output following Bard et al., 2000. More recent studies utilize physics-based models to estimate solar activity from the production rate of cosmogenic isotopes taking into account non-linearities between isotope production and the sun's open magnetic flux and variations in the the geomagnetic field (Solanki et al., 2004; Muscheler et al., 2005). Following this approach, Solanki et al., 2004 suggest that the current level of solar activity has been without precedence over the last 8000 years. An even more recent analysis in a similar vein that links the isotope proxy records to instrumental data identifies, for the last millennium, three periods of equally high or higher solar activity than for the satellite era (Muscheler et al., 2005a).

It remains unclear to what extent cosmogenic isotopes track variations in total solar irradiance and the magnitude of the long-term trend in solar irradiance remains uncertain. A reassessment of the stellar data (Hall and Lockwood, 2004) has been unable to confirm the analysis by Baliunas and Jastrow (1990) that implied significant long-term solar irradiance changes, and also underpinned some of the earlier reconstructions. Several new studies (Lean et al., 2002; Y.M. Wang et al., 2005) suggest little similarity between variations in solar activity (i.e., in the open magnetic field flux) and total solar irradiance (closed magnetic field).

These studies, as well as several others (Foster, 2004; Foukal et al., 2004), suggest that long-term irradiance changes were notably less than in the reconstructions of Hoyt et al. (1993), Lean et al. (1995; 2000), Lockwood and Stamper (1999), Fligge and Solanki (2000) and Bard et al. (2000) that were employed in a number of IPCC TAR climate change simulations and in many of the simulations shown in Figure 6.10. In the previous reconstructions, the seventeenth century Maunder Minimum total irradiance was 0.15% to 0.65% (irradiance change: ~2 to 8.7 W m^{-2}; radiative forcing: ~0.36 to 1.55 W m^{-2}) below the present-day mean. Most of the recent studies (with the exception of Solanki and Krivova (2003)) calculate a reduction of only around 0.1% (irradiance change on the order of 1 W m^{-2}) (see also the discussion of climate forcings in Chapter 2). Following these results, the radiative forcing used in Chapter 9 for the Maunder Minimum time period (~1700) is at the low end of this reconstruction (–0.2 W m^{-2}).

There is also uncertainty in the estimates of volcanic forcing during recent millennia. This derives from the necessity to infer atmospheric optical depth changes (including geographic details as well as temporal accuracy and persistence), where there is only indirect evidence in the form of levels of acidity measured in ice cores. All of the volcanic histories used in current model-based paleoclimate simulations are based on analyses of polar ice cores containing (in some cases) minor dating uncertainty and obvious geographical bias.

The considerable difficulties in calculating hemispheric and regional volcanic forcing changes (Robock and Free, 1995; Robertson et al., 2001; Crowley et al., 2003) result from differences in the choice of which ice cores are considered, assumptions as to the extent of stratosphere penetration by eruption products, and the radiative properties of different volcanic aerosols and their residence time in the stratosphere. Even after producing some record of volcanic activity, there are major differences in the way models implement this, as a direct reduction in global radiative forcing with no altitudinal or spatial discrimination, while other models prescribe geographical changes in radiative forcing (Crowley et al., 2003; Von Storch et al., 2004; Goosse et al., 2005a). Models with more sophisticated radiative schemes are able to incorporate prescribed aerosol optical depth changes, and also interactively calculate the perturbed (long and short wave) radiation budgets (Tett et al., submitted). The effective level of (prescribed or diagnosed) volcanic forcing therefore varies considerably between the simulations (Figure 6.10a).

6.5.4 Simulations of Northern Hemisphere Mean Temperature

Figure 6.10d shows the Northern Hemisphere mean (land and marine) surface temperatures simulated by a range of climate models using the forcings shown in Figures 6.10a-c. Despite differences in the detail and implementation of the different forcing histories, there is generally good qualitative agreement between the simulations as regards the major features: warmth during much of the 12th through 14th centuries, with lower temperatures being sustained during the 17th, mid 15th and early 19th centuries, and the subsequent sharp rise to unprecedented levels of warmth at the end of the 20th century. The spread of this multi-model ensemble is constrained to be small during the 1500–1899 reference period (selected following (Osborn et al., submitted)), but the inter-model spread also remains small back to 1000, with the exception of the ECHO-G simulation (Von Storch et al., 2004). The implications of the greater inter-model spread in the rate of warming after 1840 will be clear only after determining the extent to which it can be attributed to differences in prescribed forcings and individual model sensitivities (Goosse et al., 2005b). The ECHO-G simulation (dotted red line in Figure 6.10d) is atypical compared to the ensemble as a whole. (Osborn et al., submitted) show that this is likely the result of the large initial disequilibrium and the lack of anthropogenic troposphere aerosol in that simulation (see Figure 6.10c).

All of these simulations appear to be consistent with the available evidence from reconstructions of past Northern Hemisphere temperatures (Figure 6.8b). This evidence is shown on Figure 6.10d in the form of an inner envelope encompassing the reconstructed temperatures, surrounded by grey shading indicating the average uncertainty ranges. The ECHO-G simulation (dotted red line) is the only one that repeatedly leaves the range of reconstruction uncertainties, but after adjusting for the causes of early and recent excess warmth (Osborn et al., submitted), this is also consistent (solid red line) with the reconstructed data. Nevertheless, it is important to note that many of the variations during the pre-industrial time period shown in these figures have been driven by the assumed solar forcing, whose magnitude is currently in doubt. Therefore, although the data and simulations appear consistent at this hemispheric scale, they are not a powerful test of the models because of the large uncertainty in both the reconstructed Northern Hemisphere changes and the total radiative forcing. However, even with the estimated solar forcing changes, simulations that did not include anthropogenic rises in greenhouse gas concentrations were unable to reproduce the post-1850 warming evident in the instrumental and reconstructed NH temperatures (Crowley, 2000; Bertrand et al., 2002; Hegerl et al., submitted; Tett et al., submitted).

An overall conclusion can be drawn from the available instrumental and proxy evidence for the history of Hemispheric average temperature change over the last 500-2000 years, as well as the modeling studies exploring the possible roles of various causal factors: that is, greenhouse gases must be included among the forcings in order to simulate Hemispheric mean temperatures that are compatible with the empirical evidence of unusual warmth observed in the second half of the 20th century.

6.5.5 Consistency Between the Temperature, Greenhouse Gas, and Forcing Records and Compatibility of Coupled Carbon Cycle – Climate Models with the Proxy Records

As noted above, the evidence from the different proxy records for hemispheric temperature change, atmospheric trace greenhouse gases, inferred solar forcing, and reconstructed volcanic forcing, each contain some inconsistencies. The available temperature reconstructions suggest that decadally-averaged northern hemisphere temperature varied within 1°C or less over the two millennia preceding the 20th century (Figure 6.8), but the magnitude of the reconstructed variations differs by a factor of two. The greenhouse gas records of CO_2, CH_4, and N_2O show only small changes over the same time period (Figure 6.6), and are the best-known aspect of the record. The reconstructions of natural forcings (solar, volcanic) are uncertain for this timeframe. If they produced substantial negative energy balances (reduced solar, increased volcanic activity), then low-to-medium estimates of climate sensitivity are compatible with the reconstructed temperature variations (Figure 6.8); however, if solar and volcanic forcing were similar to today, then moderate-to-high climate sensitivity would be consistent with the temperature reconstructions, especially those showing larger cooling (see also Chapter 9).

The sensitivity of atmospheric CO_2 to climatic changes as simulated by coupled carbon cycle-climate models is broadly consistent with the ice core CO_2 record and the reconstructed variability in Northern Hemisphere temperature (Joos and Prentice, 2004). The CO_2-climate sensitivity can be numerically defined as the change in atmospheric CO_2 relative to a nominal change in NH temperature in units of ppm/°K. Its strength depends on several factors, including the change in solubility of CO_2 in seawater, and the responses of productivity and heterotrophic respiration on land to temperature and precipitation. The sensitivity was estimated for modest (NH temperature change <~1°C) temperature variations from simulations with the Bern Carbon Cycle-Climate model driven with solar and volcanic forcing over the last millennium (Gerber et al., 2003b), and from simulations with the range of models participating in the coupled carbon cycle-climate model intercomparison project (C4MIP) over the industrial period (Friedlingstein et al., 2005). The range of the CO_2-climate sensitivity is 4 to 16 ppm/°K for the ten models participating in the C4MIP intercomparison (evaluated as the difference in atmospheric CO_2 for the 1990 decade between a simulation with, and without, climate change, divided by the increase in NH temperature from the 1860 decade to the 1990 decade). This is comparable to a range of 10 to 17 ppm/°K as obtained for CO_2 variations in the range of 6 to 10 ppm (Siegenthaler et al., 2004; Etheridge et al., 1996), and assuming that (decadally-averaged) NH temperature varied within 0.6°C.

6.5.6 Regional Variability in Quantities Other than Temperature

What do changes in the El Niño-Southern Oscillation (ENSO) system tell us?
Considerable interest in the El Niño-Southern Oscillation (ENSO) system has encouraged numerous attempts at its paleoclimatic reconstruction. These include a boreal winter (December-February) reconstruction of the Southern Oscillation Index (SOI) based on ENSO-sensitive tree ring indicators (D.W. Stahle et al., 1998), two multiproxy reconstructions of annual and October-March Niño-3 index (average SST anomalies over 5°N–5°S, 150°W–90°W (Mann et al., 2000a; Mann et al., 2000b), and a tropical coral-based Niño 3.4 SST reconstruction (Evans et al., 2002). Fossil coral records from Palmyra Island in the tropical Pacific also provide 30–150-year windows of ENSO variability within the last 1100 years (Cobb et al., 2003). Finally, a

new 600-yr reconstruction of December–February Niño-3 SST has recently been developed (D'Arrigo et al., 2005b),which is considerably longer than previous series. These reconstructions share significant common variance (typically more than 30% during their respective cross-validation periods), suggesting a relatively consistent history of El Niño in past centuries (Jones and Mann, 2004). In addition, reconstructions of extratropical temperatures and atmospheric circulation features (e.g., the North Pacific Index) correlate significantly with tropical estimates, supporting evidence for tropical/high-latitude Pacific links during the past 3–4 centuries (Evans et al., 2002; Linsley et al., 2004; D'Arrigo et al., 2005a).

Several coral and tree-ring studies indicate that interannual ENSO weakened during the cooler and drier late 19th century (i.e., in the central Pacific), while decadal variability intensified, suggesting that the frequency-domain characteristics of ENSO are sensitive to background conditions (Urban et al., 2000). Moreover, in most coral records from the Indian and western Pacific Oceans, late 20th-century warmth is unprecedented over the past 100–300 years (see Cole, 2003). The superposition of this trend on interannual variability has led to increasingly warm/wet ENSO events in the central Pacific during recent decades.

ENSO may have responded to radiative forcing induced by solar and volcanic variations over the past millennium (Adams et al., 2003; Mann et al., 2005b). Model simulations support a statistically significant response of ENSO to radiative changes such that during higher radiative inputs, La Niña-like conditions result from an intensified zonal SST gradient that drives stronger trade winds, and vice versa (Mann et al., 2005b). Comparing data and model results over the past millennium suggests that warmer background conditions are associated with higher variability (Cane, 2005). Numerical experiments suggest that the dynamics of ENSO may have played an important role in the climatic response to past changes in radiative forcing (Mann et al., 2005a). However, ENSO reconstructions from Palmyra corals show considerable variability that appears to be internal to the ENSO system, and not tightly coupled to global mean climate (Cobb et al., 2003). This large internal variability complicates efforts to identify simple relationships among mean climate state, ENSO variability, and radiative forcing.

Proxy records suggest that ENSO's global climate imprint evolves over time, complicating predictions. Comparisons of ENSO and drought indices clearly show changes in the linkage between ENSO and U.S. moisture balance over the past 150 years. Significant ENSO-drought correlations occur consistently in the southwest U.S., but the strength of moisture penetration into the continent varies substantially over time (Cole and Cook, 1998; Cook et al., 2000). Comparing reconstructed Niño 3 SST with global temperature patterns suggests that some features are robust through time, such as the warming in the eastern tropical Pacific and western coasts of North and South America, whereas teleconnections into North America, the Atlantic and Eurasia are variable (Mann et al., 2000b). The spatial correlation pattern for the period 1801-1850 provides striking evidence of nonstationarity in ENSO teleconnections, showing a distinct absence of the typical pattern of tropical Pacific warming (Mann et al., 2000a).

What does the record of past Atlantic variability tell us?

Climate variations over the North Atlantic are related to changes in the North Atlantic Oscillation (NAO) (Hurrell, 1995) and the Atlantic Multidecadal Oscillation (Delworth and Mann, 2000; Sutton and Hodson, 2005). From 1980 to 1995, the NAO tended to remain in one extreme phase and accounted for a substantial part of the wintertime warming over Europe and northern Eurasia. Although the North Atlantic region has a unique combination of long instrumental observations, many documentary records and multiple sources of proxy records, it still remains difficult to document past variations in the dominant modes of climate variability in the region, including NAO, due to problems of establishing proxies for atmospheric pressure, as well as the lack of stationarity in the NAO frequency and in storm tracks. Several reconstructions of NAO have been proposed (Cook et al., 2002; Cullen et al., 2002; Luterbacher et al., 2002). Although the reconstructions differ in many aspects, there is a general tendency for more negative NAO during the 17th and 18th centuries than in the 20th century, thus indicating that the colder mean climate was characterized by a less zonal atmospheric pattern than in the 20th century. The coldest reconstructed European winter in 1708/1709, and the strong warming trend between 1684 and 1738 (+0.32°C per decade), have been related to a negative NAO index and the NAO response to increasing radiative forcing, respectively (Luterbacher et al., 2004). It is well known that NAO exerts a dominant influence on wintertime temperature and precipitation over Europe, but the strength of the relationship can change over time and region (Jones et al., 2003a). The strong trend towards a more positive NAO in the early part of the 18th century in the (Luterbacher et al., 2002) NAO-reconstruction appears connected with positive winter precipitation anomalies over NW Europe and marked expansions of maritime glaciers in a manner similar to the effect of positive winter precipitation anomalies over the recent decades for the same glaciers (Nesje and Dahl, 2003).

What does the record of Asian monsoon variability tell us?
Paleoclimate records of the Asian monsoon indicate substantial decade- to century-scale variability superimposed on the orbitally-driven monsoon change that took place over the last ca. 10,000 years (see Section 6.4 for the orbital change). Although there is no evidence that the entire Asian monsoon region changes in a simple synchronous manner over decade to century time scales, there is evidence that the strength of the monsoon, and hence precipitation amount can change abruptly, perhaps as a non-linear abrupt response to more gradual (e.g., orbital) changes in forcing (Morrill et al., 2002); (Zhang et al., 2004).

Another intriguing finding is that the South Asian (Indian) monsoon has, in the drier areas of its influence, recently reversed its millennia-long orbitally-driven low-frequency trend toward less rainfall. This recent reversal in monsoon rainfall also appears to coincide with a synchronous increase in inferred monsoon winds over the western Arabian Sea (Anderson et al., 2002), a change that could be related to increased summer heating over and around the Tibetan Plateau (Brauning and Mantwill, 2004; Morrill et al., 2005), or perhaps to a persistent interglacial millennial-scale mode of monsoon variability (A. K. Gupta et al., 2003).

What does the record of northern and eastern Africa hydrologic variability tell us?
Lake sediment and historical documentary evidence indicates that northern Africa and the Sahel region have for a long time experienced substantial droughts lasting from decades to centuries (Kadomura, 1992; Verschuren, 2001); (Russell et al., 2003; Stager et al., 2003; Nguetsop et al., 2004); (Brooks et al., 2005; Stager et al., 2005). Although there have been attempts to link these dry periods to solar variations, the evidence is not conclusive (Stager et al., 2005), particularly given that the relationship between hypothesized solar proxies and variation in total solar irradiance remains unclear (see Section 6.5.3). However, the paleoclimate record indicates that persistent droughts should continue to be a feature of climate in northern and eastern Africa, although the impacts of drought in the future will likely be exacerbated by higher temperatures (see Chapter 10).

What does the record of North American hydrologic variability and change tell us?
The paleohydrologic record of North America is the most complete and diverse of any in the world, in part due to the proximity to many well equipped laboratories, but also due the concern over the frequent changes in drought, flood and hurricane variability that appear to have occurred during the current interglacial. Multiple proxies, including tree-rings, sediments, historical documents, and lake sediment records make it clear that the past 2000 years included periods with more frequent, longer and/or geographically more extensive drought in North America (D. W. Stahle et al., 1998; Woodhouse and Overpeck, 1998; Stahle et al., 2000; Forman et al., 2001; Haug et al., 2003; Cook et al., 2004b; Hodell et al., 2005; MacDonald and Case, 2005). Past droughts, including decadal-length "megadrought" (Woodhouse and Overpeck, 1998; Stahle et al., 2000), are most likely due to extended periods of anomalous SST (Hoerling and Kumar, 2003; Schubert et al., 2004; MacDonald and Case, 2005; Seager et al., submitted), but still remain difficult to simulate with coupled ocean-atmosphere models. Thus, the paleoclimatic record suggests that multi-year, decadal, and even century-scale drier periods could be a feature of future North American climate, particularly in the area to the west of the Mississippi River.

There is some evidence that North American drought was more regionally extensive, severe and frequent during past intervals that were characterized by warmer than average Northern Hemisphere summer temperatures (e.g., the Medieval Warm Period and the Mid-Holocene (Forman et al., 2001; Cook et al., 2004b)).

There is evidence that changes in North American hydrologic regime can occur abruptly relative to the rate of change in climate forcing and duration of subsequent climate regime. Abrupt shifts in drought frequency and duration have been found in paleohydrologic records from western North America (Cumming et al., 2002; Laird et al., 2003; Cook et al., 2004b). Similarly, the Upper Mississippi River basin and elsewhere has seen abrupt shifts in the frequency and size of the largest flood events (Knox, 2000). Lastly, recent investigations of past large-hurricane activity in the southeast United States suggests that changes in the frequency of large hurricanes can shift abruptly in response to more gradual forcing (K.B. Liu, 2004).

Altogether, the paleoclimatic record indicates that future hydrologic shifts in drought, floods and tropical storms could occur abruptly (i.e., within years), and in ways that are currently difficult to predict.

6.6 Biogeochemical and Biophysical Interactions in the Paleoclimate Record

It is more than a hundred years since Arrhenius (e.g., Arrhenius, 1896) calculated that a doubling of atmospheric CO_2 would cause Earth's surface temperature to rise by about 4°C. Arrhenius was searching for mechanisms to explain the known glacial-interglacial climate variations and invoked the greenhouse gas theory developed by Fourier, Tyndall and others in the 19th century. Today, we know from the Antarctic ice core record that the evolution of Antarctic temperature is highly correlated with that of atmospheric CO_2 and other greenhouse gases glacial-interglacial cycles (see section 6.3). As discussed earlier, the evolution of atmospheric CO_2 appears to lag that of Antarctic temperature, and ice sheet extent lags atmospheric CO_2. This is consistent with the view that natural CO_2 variations and ice albedo constitute feedbacks in the glacial-interglacial cycle, rather than a primary driver (Shackleton, 2000). Changes in the Earth's orbit around the Sun appear to be the pacemaker for glacial-interglacial cycles (Milankovich, 1941; Hays et al., 1976; Berger, 1978), but these, globally rather subtle, orbital changes must be amplified by climate feedbacks in order to explain the large differences in global average surface temperature and ice volume, as well as the relative abruptness of the transitions between glacial and interglacial periods (Berger et al., 1998; Clark et al., 1999).

Biogeochemical cycles played an important role for the amplification of orbital changes. The radiative forcing by changes in the atmospheric concentration of the greenhouse gases CO_2, CH_4, and N_2O, in atmospheric dust loading, and vegetation cover constitute more than half of the known radiative forcing during the Last Glacial Maximum (Figure 6.2). Model simulations suggest that the direct radiative forcing by atmospheric CO_2 concentrations have contributed to about half of the reconstructed glacial-interglacial surface temperature difference in the tropical ocean (e.g. Shin et al., 2003b). Other major factors involved in maintaining cold conditions during glacial periods include the high albedo of the continental ice sheets, the high albedo (especially when snow-covered) of extensive non-forested regions at high latitudes (Gallée et al., 1992; Levis et al., 1999; Wyputta and McAvaney, 2001; Yoshimori et al., 2001), and the reflection of shortwave radiation by the enhanced atmospheric content of mineral dust (Claquin et al., 2003) – itself a consequence of reduced vegetation cover (Mahowald et al., 1999; Werner et al., 2002). It is plausible that such feedbacks amplify the direct anthropogenic forcing, just as they have amplified orbital changes in the past.

Changes in the marine biogeochemical cycles are mainly responsible for the glacial-interglacial CO_2 variations of up to 100 ppm (Box 6.3). This suggests that future changes in ocean circulation and biogeochemical cycles may amplify the anthropogenic perturbation (see Chapter 7 for further discussion). There is paleoclimatic evidence revealing how changes in the North Atlantic Deep Water production affected CO_2, and CO_2 changes during the glacial Antarctic warm events, linked to changes in North Atlantic Deep Water (Knutti et al., 2004) were small (less than 20 ppm, Figure 6.3). Consistently, a relatively small positive feedback between atmospheric CO_2 and changes in the rate of North Atlantic Deep Water formation are found in paleo and global warming simulations (Joos et al., 1999; Marchal et al., 2002). Thus, paleodata and available model simulations agree that possible future changes in the North Atlantic Deep Water formation rate would have only modest effects on atmospheric CO_2. This finding does not, however, preclude the possibility that circulation changes in other ocean regions, in particular in the Southern Ocean, could have a larger impact on atmospheric CO_2 (Greenblatt and Sarmiento, 2004).

There is evidence of the impact on atmospheric CO_2 of aeolian iron deposition into the Southern Ocean. Iron is a limiting micronutrient for phytoplankton, especially diatom, growth in the modern Southern Ocean (Martin et al., 1994; Boyd et al., 2000). An enhanced (reduced) iron input into the ocean could enhance (reduce) nutrient utilization in the surface ocean and lead to enhanced (reduced) uptake of excess atmospheric CO_2. The large variations in dust deposition recorded in Antarctica ice during glacial climate variations (Röthlisberger et al., 2004) are accompanied by only small variations in atmospheric CO_2. Consistently, model simulations suggest a limited role for iron in regulating atmospheric CO_2 concentration (e.g., Bopp et al., 2003). This provides some paleoclimate evidence that potential changes in marine iron supply may have a limited impact on atmospheric CO_2 over the coming centuries and that fertilization of the ocean with iron to mitigate anthropogenic climate change may not be very effective.

Paleoclimatic evidence shows that vegetation reacts sensitively to climate change and suggests that changes in vegetation cover affect climate. For example during the mid-Holocene, a more extensive distribution of temperate deciduous forests and grasslands is found in the continental interior, together with a northward shift of the boreal treeline in response to the warming in northern and middle to high latitudes (COHMAP, 1988; Bigelow et al., 2003). There is extensive evidence that parts of the Sahara desert were covered by shrubs and grasses (Jolly and al., 1998), a development that can be explained by an increase in monsoon penetration due to greater than present sea-land temperature contrast in the Northern Hemisphere (Kutzbach and Street Perott, 1985), coupled with the positive feedbacks caused by vegetation-atmosphere interactions (e.g. de Noblet Ducoudré et al., 2000) Vegetation structure was also markedly different during the Last Glacial Maximum, when tropical tree cover was reduced, and the Arctic treeline displaced southward (Adams and Faure, 1998; Prentice et al., 2000).

Rapid, decadal-scale changes in vegetation structure have been found in annually-laminated sequences during rapid climate change episodes, such as the beginning and the end of the Younger Dryas Northern Hemipshere cold event and the Holocene 8.2 ka event (Birks and Ammann, 2000; Tinner and Lotter, 2001). Marine pollen records with a typical sampling resolution of 200 years provide unequivocal evidence of the immediate response of vegetation in Southern Europe to the climate fluctuations during glacial times (Sànchez Goñi et al., 2002; Tzedakis, 2005). The same holds true for the vegetation response in Northern South America during the last deglaciation (Hughen et al., 2004).

Considering changes in vegetation appears to improve the realism of simulations of the Last Glacial Maximum (LGM), and also points to important climate-vegetation feedbacks. The biome distribution simulated with dynamic global vegetation models reproduce the broad features observed in paleodata (Crucifix and Hewitt, 2005). For example, extension of the tundra in Asia during the LGM contributes to the local surface cooling, while the tropics warm where tropical forest is replaced by savannah (Wyputta and McAvaney, 2001). Feedbacks between climate and vegetation occur locally, with a decrease in the tree fraction in central Africa reducing precipitation, and remotely with cooling in Siberia (tundra replacing trees) and Tibet (bare soil replacing grasslands) altering (diminishing) the Asian summer monsoon. Inclusion of the physiological effect of CO_2 concentration on vegetation needs to be included to properly represent changes in global forest (Harrison and Prentice, 2003), as well as to widen the climatic range where grasses and shrubs dominate.

There is evidence that terrestrial carbon storage was reduced during the Last Glacial Maximum compared to today. Mass balance calculations based on ^{13}C measurements on shells of benthic foraminifera yield a reduced terrestrial biosphere carbon inventory of about 300 to 700 GtC (Shackleton, 1977; Bird et al., 1994) at the Last Glacial Maximum compared to pre-industrial time. Estimates of terrestrial carbon storage based on ecosystem reconstructions suggest a larger difference (e.g., Crowley, 1995), however, these are very approximate due to large gaps in the data considered, and assumptions about the average carbon density of different forests. Simulations with carbon cycle models yield lower global carbon stocks of 600 to 1000 GtC at the LGM than at pre-industrial time (Francois et al., 1998; Beerling, 1999; Francois et al., 1999; Otto et al., 2002; Kaplan et al., 2003; Kaplan JO, 2003; Joos et al., 2004a). The majority of this simulated difference is due to reduced simulated growth resulting from lower atmospheric CO_2. In summary, results of terrestrial models, also used to project future CO_2 concentrations, are broadly compatible with the range of reconstructed differences in glacial-interglacial carbon storage on land.

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Question 6.1: What Caused the Ice Ages and Other Important Climate Changes Before the Industrial Era?

Climate on Earth has changed on all time scales, long before human activity could have played a role. Great progress has been made in understanding the causes and mechanisms of these climate changes. There is not one major cause or "driver" of past climate changes, but several. For each case – be it the Ice Ages, the warmth at the time of the dinosaurs or the ups-and-downs of the past millennium – the specific causes must be established individually. In many cases this can now be done with good confidence, and many past climate changes can be reproduced with quantitative models.

Our global climate is determined by the radiation balance of the planet (Question 1.1). There are three fundamental ways to change the radiation balance and hence cause a climate change: (1) changing the incoming solar radiation (e.g., by changes in the Earth's orbit or in the sun itself), (2) changing the fraction of solar radiation that is reflected (this fraction is called the albedo – it can be changed e.g., by changes in cloud cover, aerosols or land cover), and (3) altering the long-wave back-radiation (e.g., by changes in the greenhouse gas concentration). In addition, local climate also depends on how heat is distributed by winds and ocean currents. All of these factors have played a role in past climate changes.

Starting with the Ice Ages that have come and gone in regular cycles for the past nearly 3 million years, it is now well established that these are caused by regular variations in the Earth's orbit around the sun, the so-called Milankovich cycles (explained in paleoclimatology textbooks, e.g., Ruddiman (2000)). These cycles change the amount of solar radiation received at each latitude in each season (but hardly the global, annual mean), and they can be calculated with astronomical precision. There is still some discussion how exactly this starts and ends ice ages, but the most likely scenario is that the amount of summer sunshine on northern continents is crucial: if it drops below a critical value, snow from the past winter does not melt away in summer and an ice sheet starts to grow as more and more snow accumulates. Climate model simulations confirm that an Ice Age can indeed be started in this way (e.g., Khodri et al., 2001, Loutre, 2003), while simple conceptual models have been used to successfully "hind-cast" the onset of past glaciations based on the orbital changes (Paillard, 1998). The next large minimum in northern summer insolation, similar to ones that started past Ice Ages, is due in ~50,000 years.

Although it is not their primary cause, atmospheric CO_2 also plays an important role in the Ice Ages. Antarctic ice core data (Petit and et al., 1999) show that CO_2 concentration is low in the cold glacial times (~190 ppm), and high in the warm interglacials (~280 ppm); atmospheric CO_2 follows the climate changes with a lag of some hundreds of years (Caillon et al., 2003). Because the climate changes at the beginning and end of ice ages take several thousand years, most of these changes are affected by a positive CO_2 feedback; i.e., a small initial cooling due to the Milankovich cycles is subsequently amplified as the CO_2 concentration falls. Model simulations of Ice Age climate (see discussion in section 6.4.2.1) yield realistic results only if the role of CO_2 is accounted for.

Within the Ice Ages, over 20 abrupt and dramatic climate shifts have occurred that are particularly prominent in records around the northern Atlantic (see section 6.3). These differ from the glacial-interglacial cycles in that they probably do not involve large changes in global mean temperature: changes are not synchronous in Greenland and Antarctica (Blunier and Brook, 2001; Blunier et al., 1998), and they have the opposite sign in South and North Atlantic (Voelker, 2003). This means we do not need to look for a major change in global radiation balance as their cause; a redistribution of heat within the climate system will suffice. There is indeed strong evidence that changes in ocean circulation and heat transport can explain many features of these abrupt events (see reviews by Clark et al. (2002), Rahmstorf (2002)); sediment data and model simulations show that some of these changes could have been triggered by instabilities in the ice sheets surrounding the Atlantic at the time.

Much warmer times have also occurred in climate history – during most of the past 500 million years our planet was probably completely free of ice sheets (geologists can tell from the marks ice leaves on rock), unlike today, where Greenland and Antarctica are ice covered. Data on greenhouse gases going back beyond a million years, that is beyond the reach of Antarctic ice cores, are still rather uncertain, but analysis from sediment cores suggest that the warm ice-free periods coincide with high atmospheric CO_2 levels (Royer et al., 2004). On million-year time scales, CO_2 levels change due to tectonic activity, which affects the rates of

CO_2-exchange of ocean and atmosphere with the solid Earth. See Box 6.1 for more about these ancient climates.

Another likely cause of past climatic changes has been variations in the energy output of the sun. We know from measurements over recent decades that the solar output varies slightly (by close to 0.1%) in an 11-year cycle, and that these variations are correlated with the number of sunspots, as well as with cosmic rays reaching the Earth's surface. Hence, sunspot observations (going back to the 17th Century), as well as cosmogenic isotope data provide evidence for longer-term changes in solar activity. Such data show that the coldest periods of the past millennium coincide with minima in solar activity – for example, the Maunder minimum around the year 1700 (see section 6.5). Data correlation, as well as model simulations, indicate that solar variability and volcanic activity are likely to be leading reasons for climate variations of the past millennium, before the start of the industrial era.

These examples illustrate that different climate changes in the past had different causes. However, these natural causes very likely cannot explain the warming of the past few decades. Milankovich cycles or tectonic changes act too slowly; solar activity shows no clear trend since 1940 (although it has increased until then (Solanki and Krivova, 2003; Usoskin et al., 2003), changes in oceanic or atmospheric circulation could not explain a global warming trend, and neither can volcanic activity.

Question 6.2: Is the Current Climate Change Unusual Compared to Earlier Changes in Earth's history?

Climate has changed on all time scales throughout Earth's history. Some aspects of the current climate change are not unusual, but others are. CO_2 concentration in the atmosphere has reached a million-year record high at an exceptionally fast rate. Current global temperatures are as warm as they have ever been during the past eight centuries, probably even for millennia. And faster rates of global-mean warming than those of the past 30 years (about 0.19ºC per decade) are at least not documented in the records from the past. If warming continues unabated, the resulting climate change within this century would be extremely unusual even in geological terms.

When comparing the current climate change to earlier, natural ones, we need to make three distinctions. First, we need to be clear which variable we are comparing: is it greenhouse gas concentration or temperature (or some other climate parameter), and is it their absolute value or their rate of change? Second, we must not confuse local with global changes. Local climate changes are often much larger than global ones, since local factors (e.g., changes in oceanic or atmospheric circulation) can shift the delivery of heat or moisture from one place to another and local feedbacks operate (e.g., sea ice feedback). Large changes in global mean, in contrast, require some global forcing (such as a change in greenhouse gas concentration or solar activity). Third, we must distinguish between time scales. Climate changes over millions of years can be much larger and have different causes (e.g., continental drift) compared to climate changes on a century time-scale.

The main reason for the current concern about climate change is the rise in atmospheric CO_2 concentration, which is very unusual for the Quaternary (about the last 2 million years). CO_2 concentration is now known accurately almost half a million years back in time from Antarctic ice cores (Petit et al., 1999), and the new EPICA core will provide a record 700,000 years back in time, when analyses are finished. During this time, CO_2 concentration has varied between a low of 190 ppm during cold glacial times and a high of 290 ppm during warm interglacials. Over the past two centuries, it has increased to 380 ppm (see Chapter 2). For comparison, the ~80 ppm rise in CO_2 concentration at the end of the past Ice Ages generally took over 5,000 years. Higher values than at present have only occurred many millions of years ago (see Question 6.1).

Temperature is a more difficult variable to reconstruct than CO_2 (a globally well-mixed gas), as it does not have the same value all over the globe, so that a single record (e.g., an ice core) is only of limited value. Local temperature fluctuations, even those over just a few decades, can be several degrees, which is larger than the global warming signal of the past century of ~0.6ºC. Hence, in most places the global "signal" does not clearly exceed the "noise" of natural variability, and some regions of the Earth are cooling despite the global warming trend (see Chapter 3). Although they must not be over-interpreted, local records can still be interesting. For example, oxygen isotope data of the Dye 3 ice core from southern Greenland, which is the closest to the Medieval Viking settlement, shows that in the mid-20th Century the highest oxygen-18 values were reached, suggesting the warmest temperatures for several millennia in this region.

More meaningful for global changes is an analysis of large-scale (global or hemispheric) averages, where much of the local variations average out and variability is smaller. Sufficient coverage of instrumental records only goes back ~150 years. On this time scale, the current warming is clearly unusual – the globally warmest years on record are 1998, 2002, 2003, and 2001 (see Chapter 3). Further back in time we have compilations of proxy data from tree rings, ice cores etc., going back 1–2 millennia, with decreasing spatial coverage for earlier periods (see Section 6.5). While there are still differences between those reconstructions and significant uncertainties remain, all published reconstructions find that temperatures were warm during the Middle Ages, and then cooled to low values in the 17th, 18th, and 19th centuries, warming rapidly after that. The medieval level of warmth was reached again in the mid-20th Century, and has thus been exceeded since then. Independent support for this conclusion comes from models driven by reconstructed forcings, including solar variability. Since proxies indicate similar solar activity in the mid-20th Century as in medieval times, this conclusion is robust with respect to a scaling of the amplitude of solar variability, or any possible amplifying mechanisms.

Proxy data for the period before 2000 years ago have not been systematically compiled into large-scale averages, but they do not provide evidence for warmer-than-present global annual-mean temperatures going back through the Holocene (the last 11,600 years – see Section 6.4), or even at the peak of the previous interglacial period (~125,000 years ago – see Section 6.3). Models can reproduce past warm climates when orbital forcing (Milankovich cycles, see Question 6.1) is accounted for (e.g., Ganopolski et al., 1998). There are strong indications that a still warmer climate, with much reduced global ice cover, prevailed until around 3 million years ago. Hence, current warmth appears unusual in the context of the past millennia, but not unusual on longer time scales for which changes in tectonic activity become relevant (see Box 6.1).

A different matter is the current rate of warming of 0.19ºC per decade. Are more rapid *global* climate changes recorded in proxy data? The largest temperature changes of the past million years are the glacial cycles, during which the global mean temperature changed by 4ºC–7ºC between ice ages and warm interglacial periods (local changes were much larger, for example near the continental ice sheets). However, the data indicate that the global warming at the end of an ice age was a gradual process taking ~5,000 years, yielding a mean rate of around 0.01ºC per decade (see section 6.3). The much-discussed abrupt climate shifts during glacial times (also see section 6.3) are not counter-examples, since they were probably due to changes in ocean heat-transport which would hardly affect the global mean temperature.

Further back in time, beyond ice core data, the time resolution of sediment cores and other archives does not resolve changes as fast as the present warming. Hence, although large climate changes have occurred in the past, we have no evidence of these proceeding at a faster rate than present warming. Neither do we know of a mechanism other than a rapid greenhouse gas release that could lead to equally rapid global warming. If the more pessimistic projections of ~5ºC warming in this century are realised, then the Earth will have experienced the same amount of global-mean warming as it did at the end of the last Ice Age; however, this rate of future change would then very likely be much faster than any comparable global temperature increase of the last 50 million years.

Tables

Table 6.1. Consensus of PMIP-2 models in simulating proxy indications of climate changes at the LGM and mid-Holocene.

Result	Region	Period	Consensus
Atmosphere and Land Changes			
Change in surface temperature gradients over continents	Eurasia	LGM	Pollen data indicates a weakening of the E-W temperature gradient with greater cooling in SW Europe than Siberia. Models simulate this change, but underestimate cooling in western Europe.
Increase in northward extent of African summer monsoon	Sahara-Sahel	Mid-Holocene	Pollen and lake-level data indicate expansion of area influenced by African monsoon. Models simulate increased summer precipitation in northern Africa, but generally do not inccrease annual precipitation enough to allow transition from desert to steppe.
Ocean Changes			
Ocean cooling	Tropics	LGM	Models and data agree on cooling of 0–3°C over most of the tropical oceans. Some proxies find greater cooling in the eastern Atlantic and Indian Oceans and the southern Caribbean, which models cannot generally reproduce.
Ocean cooling	North Atlantic	LGM	Most models simulate the strengthening of the SST meridional gradient suggested by the data, but most do not get the cooling with the right amplitude or at the right location.
Change in structure of Atlantic meridional overturning circulation	North Atlantic	LGM	Data indicates a shoaling of the AMOC. Model responses vary in their predictions of change in strength and depth penetration of the AMOC.
Cryosphere Changes			
Cooling at polar latitudes	Greenland and Antarctica	LGM	Models and data indicate polar amplification of global cooling. All models underestimate reconstructed Greenland cooling, whereas some models do capture Antarctic cooling.
Expansion of sea ice in North Hemisphere	North Atlantic	LGM	Quasi-perennial sea ice existed along eastern Canada and Greenland margins. Central and eastern North Atlantic were seasonally ice-free. Models also simulate expanded sea ice at LGM, but sea ice limits are highly variable from one model to another in LGM and preindustrial simulations.