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Past Climate Variability and Change in the Arctic and at High Latitudes

Chapter 4. Paleoclimate Concepts

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ABSTRACT

Interpretation of paleoclimatic records involves understanding of the Earth's climate system, the causes of climate changes (forcings), and the processes that amplify or damp these changes (feedbacks). Paleoclimatologists reconstruct the history of climate from proxies, which are those characteristics of sedimentary deposits that preserve paleoclimatic information. A great range of physical, chemical, isotopic, and biological characteristics of lake and ocean sediments, ice cores, cave formations, the land surface itself, and more are used to reconstruct past climate. Ages of climate events are obtained by counting annual layers, measuring effects of the decay of radioactive atoms, assessing other changes that accumulate through time at rates that can be assessed accurately, and using time-markers to correlate sediments to others that have had their ages measured more accurately. Not all questions about the history of the Earth's climate can be answered through paleoclimatology because in some cases the necessary sediments are not preserved or the climatic variable of interest is not recorded in the sediments. Nonetheless, a huge range of questions can be answered from the available information.

An overview of the history of Arctic climate over the past 65 million years (Ma) shows a long-term irregular cooling over tens of millions of years. As ice became established in the Arctic, it grew and shrank over tens of thousands of years in regular cycles. During the most recent of these cycles, shorter-lived fluctuations occurred, especially around the north Atlantic with large and rapid onsets and terminations of glaciations. The last ~11,000 years have remained generally warm and relatively stable, but with small climate changes of varying spacing and size. Assessment of the causes of

climate changes, and the records of those causes, shows that reduction in atmospheric carbon-dioxide concentration and changes in continental positions were important in the cooling trend over tens of millions of years. The cycling in ice extent was paced by features of Earth's orbit, and amplified by the effects of the ice, changes in carbon dioxide and other greenhouse gases, and additional feedbacks. Abrupt climate changes were linked to changes in the circulation of the ocean and the extent of sea ice. Changes in the sun, volcanic eruptions, and other factors have been influential over the recent warm interval.

4.1 Introduction

Most people notice the weather. Day to day, week to week and even year to year, changes in such parameters as minimum and maximum daily temperatures, precipitation amounts, wind speeds, and flood levels are all details about the weather that nearly everyone shares in daily conversations. When all else fails, most people can talk about the weather.

Evaluating longer-term trends in the weather (tens to hundreds of years or even longer) is the realm of climate science. *Climate* is the average weather, usually defined as the average of the past 30 years. *Climate change* is the long-term change of the average weather, and it is *climate change* that is the focus of this assessment. While most people accept that the weather is always changing on the time scales of their recent memory, geologists reconstruct climate on even longer timescales and use these reconstructions to help understand *why* climate changes. This improved understanding of Earth's climate system informs our ability to predict future climate change. Reconstructions of past climate also allow us to define the range of natural climate variability throughout the Earth's history. This information provides insight into an assessment of whether observed contemporary climate change is potentially part of a natural cycle or whether human activity may play a role in the observed change. The relevance of climate science lies in the recognition that even small shifts in the climate state can and have had significant economic and societal impacts (Lamb, 1997; Ladurie, 1971).

Indications of past climate, called climate proxies, are preserved in geological records and tell us that Earth's climate has rarely been static. For example, over the past

70 million years (henceforth indicated as 'Ma') of Earth history, large changes have occurred in average global temperature and in temperature differences between tropical and polar regions, as well as ice-age cycles during which more than 100 m of sea level was stored on land in the form of giant continental ice sheets and then released back to the ocean by melting of that ice. Climate change includes long-term trends lasting tens of millions of years, and abrupt shifts occurring in as little as a decade or less, both of which have resulted in large-scale reorganizations of oceanic and atmospheric circulation patterns. As we discuss in the following sections, these climate changes are understood to be caused by combinations of the plate-tectonic forces that cause continental drift and mountain-building, variations in Earth's orbit about the Sun, and changes in atmospheric greenhouse gases, solar irradiance, and volcanism, all of which can be amplified by powerful positive feedback mechanisms, especially in the Arctic. Documenting past climates (paleoclimatology) and developing scientific explanations of the observed changes inform efforts to understand the climate, reveal features of importance that must be included in predictive models, and allow testing of the models developed.

4.2 Forcings, Feedback, and Variability

An observed change in climate may depend on one or more processes. It is often useful to divide these processes into three categories: internal variability; forcings; and feedbacks. (For additional information, see IPCC, 2007; Peixoto and Oort, 1992; or Hansen et al., 1984, among other excellent sources.)

Internal variability is familiar to weather watchers; if you don't like what is happening now, wait for tomorrow and something different often arrives. Different weather arrives even though the sun, the Earth's orbit, the composition of the atmosphere, and many other important controls are the same as yesterday, because complex systems exhibit fluctuations within themselves. This variability tends to average out over longer time periods, so climate is less variable than weather; however, even for the thirty-year averages typically used in defining the climate, internal variability exists. For example, without any external cause, some thirty-year period may have one more El Niño event in the Pacific Ocean, and thus slightly warmer average conditions, than in the previous thirty-year period.

Forced changes are those with a cause from outside of the climate system. If the output of energy from the sun increases, the Earth will warm in response. If fewer volcanoes than average erupt during some century, then less sunlight than normal will be blocked by particles from those volcanoes, and the Earth's surface will be warmer in response. If the carbon-dioxide level of the atmosphere is raised by fossil-fuel burning, then more of the planet's outgoing radiation will be blocked by that carbon dioxide, warming the Earth's surface. Depending on often random processes, different forcings may combine to cause large climate swings, or offset to cause climate changes to be small.

When a climate change occurs, whether in response to some forcing or to internal variability, other parts of the climate system respond to the initial climate change, and these responses may affect the climate further; if so, then these responses are called feedbacks. How much the temperature changes in response to a forcing of a given magnitude (or net magnitude of a set of forcings in combination) depends on the sum of all of the feedbacks. Feedbacks can be characterized as positive, serving to amplify the initial change, or negative, acting to partially offset the initial change.

As an example, some of the sunshine reaching the Earth is reflected back to space by snow without warming the planet. If warming (whether caused by an El Niño, increased output from the sun, increased carbon dioxide concentration in the atmosphere, or anything else) melts snow and ice that otherwise would have reflected sunshine, then more of the sun's energy will be absorbed, causing additional warming. This additional warming is a feedback (usually called the ice-albedo feedback). This ice-albedo feedback is termed a positive feedback, because it amplifies the initial change.

4.2.1 The Earth's Heat Budget – a balancing act

Over century to millennial timescales, the energy received by the Earth from the sun and the energy returned to space balance almost exactly; imbalance between incoming and outgoing energy is typically less than 1% over periods as short as years to decades. (**Figure 4.1**) This state of near-balance is maintained by the very strong negative feedback linked to thermal radiation. All bodies "glow", sending out radiation, and warmer bodies glow more brightly and send out more radiation. (Watching the glow of a burner on an electric stove become visible as it warms shows this effect very clearly.) Some of the sun's energy reaching the Earth is reflected without causing warming, and the rest is absorbed to warm the planet. The warmer the planet, the more energy it radiates back to space. A too-cold planet will receive more energy than is radiated, causing the planet to warm, thus increasing radiation from the Earth until the incoming and outgoing energy balance. Similarly, a too-warm planet will radiate more energy than is received from the sun, producing cooling to achieve balance. Greenhouse gases in the atmosphere block some of the outgoing radiation, transferring some of the energy from the blocked radiation to other air molecules to warm them, or radiating the energy up or down. The net effect is to cause the lower part of the atmosphere (the troposphere) and the surface of the planet to be warmer than they would have been in the absence of those greenhouse gases. The global average temperature can be altered by changes in the energy from the sun reaching the top of our atmosphere, in the reflectivity of the planet (the planet's albedo), or in strength of the greenhouse effect.

Equatorial regions receive more energy from space than they return, polar regions return more energy to space than they receive, and the atmosphere and ocean transfer sufficient energy from the equatorial to the polar regions to maintain balance. (for additional information see Peixoto and Oort, 1992, and Nakamura and Oort, 1988).

Important forcings described later in this section include changes in the sun, cyclical features of Earth's orbit (Milankovitch forcing), changes in greenhouse gas concentrations in the Earth's atmosphere, the shifting shape, size, and positions of the continents (plate tectonics), biological processes, volcanic eruptions, and other features of the climate system. Other possible forcings, such as changes in cosmic rays or in blocking of sunlight by space dust, cannot be ruled out entirely but do not appear to be as important.

4.2.2 Solar Irradiance Forcing

4.2.2a Effects of the Aging of the Sun

Energy emitted by the Sun is the primary driver of Earth's climate system. The Sun's energy, or irradiance, is not constant, and changes in solar irradiance force changes in Earth's climate. Our understanding of the physics of the sun indicates that over the 4.6-billion-year history of the Earth, the sun's energy output should have increased smoothly from about 70% of modern (see, for example, Walter and Barry, 1991). (Direct paleoclimatic evidence of this increase in solar output is not available.) Over the last 100 Ma, changes in solar irradiance are calculated to have been less than 1%, or less than 0.000001% per century. Therefore, the effects of the sun's aging have no bearing on climate change over the times of millennia or less that are normally of greatest interest to humans. For reference, the 0.000001% per century change in output from aging of the sun can be compared to other things, including: i) maximum changes of slightly under 0.1% occurring over 5-6 years as part of the sunspot cycle (Foukal et al., 2006); ii) the estimated increase from the year 1750 to 2005 in solar output averaged across sunspot cycles of 0.035% (Forster et al., 2007); and iii) the warming effect of human carbon-dioxide additions to the atmosphere from 1750 to 2005, which is estimated to have had the same warming effect globally as an increase in solar output of 0.5% (Forster et al., 2007) and thus is more significant than solar irradiance changes over this time.

4.2.2b Effects of Short-term Solar Variability

Earth-based observations, and in recent years more-accurate space-based observations, document an 11-year solar cycle that results from changes occurring within the sun (Foukal et al., 2006; Beer et al., 2006), that generate a small climate response (Hegerl et al., 2007). As discussed by Hegerl et al. (2007), the lack of any trend in solar output over the time of satellite observations excludes the sun as an important contributor to the strong warming over that same interval, but the solar variability may have contributed weakly to temperature changes early in the 20th century.

Over longer time frames, indirect proxies of solar activity (historical sun-spot records, tree-rings and ice-cores) also exhibit 11-year solar cycles, as well as longer-term changes that exceed the amplitudes of their 11-year cycles. Common longer cycles include 22, 80 and 210 years. Suggestions of linkages between periods of low solar activity and climate variations can be found in the historic climate record. For example, the solar minima known as the "Dalton", "Maunder Minima" (1790 to 1820 AD, and 1645 to 1715 AD, respectively) correspond to cold summers of the Little Ice Age. However, the magnitude of radiative forcing that can be attributed to variations in solar irradiance remains debated (see, for example Amman et al., 2007; Fleitmann, et al., 2003; Muscheler et al., 2007, Frolich and Lean, 2004; Bard et al., 2000, Baliunas and Jastrow, 1990) with the estimated minimum level of at least 0.2 Watts per m^2, and some estimates as high as 0.6 Watts per m^2, still well below the estimated radiative forcing of increased greenhouse gases of the past century (~1.7 Watts per m^2) (IPCC, 2007).

4.2.3 Orbital Forcing and Milankovitch Cycles

Irregularities in Earth's orbital parameters, often referred to as 'Milankovitch variations' or 'Milankovitch cycles', after the Serbian mathematician who suggested that these irregularities might pace ice-age cycles, result in systematic changes in the seasonal and geographic distribution of incoming solar radiation (insolation) across the planet ((Milankovitch, 1920, 1941). They have almost no effect on sunshine reaching the planet over times of years or decades, have only a small effect on total sunshine reaching the planet over tens of thousands of years and longer, but have large effects on north-south and summer-winter distribution of sunshine. These ‘Milankovitch variations’ (**Figure 4.2**) are due to changes in: i) the eccentricity (out-of-roundness) of the Earth’s orbit around the sun, which varies from nearly circular to more elliptical and back over ~100 ka (‘E’ in Figure 4.2); ii) the obliquity (how far the North Pole is tilted away from sticking “straight up” out of the plane containing the Earth’s orbit about the sun), which tilts more and then less over ~41 ka (‘T’ in Figure 4.2) , and iii) the precession (the wobble, which moves the Earth from close approach to the sun in northern summer/southern winter to being far in northern summer/southern winter and back over about 19 ka to 23 ka, ‘P’ in Figure 4.2) (e.g., Loutre et al., 2004). These orbital features are linked to the influence of the gravity of Jupiter and the moon, among others, acting on the Earth itself and on the bulge at the equator caused by the Earth’s rotation, and are relatively stable orbital features that can be calculated over millions of years with high accuracy. Paleoclimatic records show the influence of these changes very clearly (e.g., Imbrie et al., 1993).

The eccentricity (orbital ‘out of roundness’, or ‘E’) variations affect the total sunshine received by the planet in a year, but by less than 0.5% between extremes (hence giving very small changes of less than 0.001% per century). The other orbital variations have essentially no effect on the total solar energy received by the entire planet. However, large variations do occur in energy received at a particular latitude and season (with offsetting changes at other latitudes and in other seasons); changes of more than 20% over 10,000 years have occurred (which is still only 0.2% per century, again with offsetting changes in other latitudes and seasons so that the total energy received is virtually constant).

In the Arctic, the most important orbital controls are the tilt of Earth's axis ('T'), where high tilt angles result in significantly more high-latitude insolation than for low tilt angles, and the precession or wobble of the Earth's rotational axis ('P'). When Earth is closest to the Sun at the summer solstice, insolation is significantly greater than when Earth is at its greatest distance from the Sun at the summer solstice. For example, 11 ka ago, Earth was closest to the Sun at the Northern Hemisphere summer solstice, but the summer solstice has been steadily moving toward the greatest distance from the Sun since then, such that at present, Northern Hemisphere summer occurs when Earth is almost the greatest distance from the Sun, resulting in 9% less insolation in Arctic summers today than 11 ka ago (**Figure 4.3**). Based on this orbital consideration alone, Arctic summers should have been cooling over this interval in response to the precession of the equinoxes.

4.2.4 Greenhouse Gases in the Atmosphere

Roughly 70% of the incoming solar radiation is absorbed by the planet, warming the land water and air. The Earth, in turn, radiates energy to balance what it receives, but at a longer wavelength than the incoming solar radiation. Greenhouse gases are those gases present in the atmosphere that allow incoming shortwave radiation to pass largely unaffected, but that absorb some of the Earth's outgoing long-wave radiation band (**Figure 4.1**). Greenhouse gases play a key role in keeping the planetary temperature within the range conducive to life. In the absence of any greenhouse gases in Earth's atmosphere the planetary temperature would be about -18°C (~0°F); with them, it averages ~33°C warmer (~57°F) (Hansen et al, 1984; Le Treut et al., 2007). The primary pre-industrial greenhouse gases include, in order of importance, water vapor, carbon dioxide, methane, nitrous oxide and tropospheric ozone, the concentrations of which are directly impacted by anthropogenic activities, with the exception of water vapor as discussed below. Purely anthropogenic recent additions to greenhouse gases include a suite of halocarbons and fluorinated sulfur compounds (Ehhalt et al., 2001).

Typically, carbon dioxide is not as significant as water vapor as a greenhouse gas near the Earth's surface. However, while it is relatively easy to change the carbon-dioxide concentration of the atmosphere, the atmospheric concentration of water vapor is difficult to change notably except by changing the temperature. Natural fluxes of water vapor into and out of the atmosphere are very large, equivalent to a layer of water across the entire surface of the Earth of ~2 cm/week (see, e.g., Peixoto and Oort, 1992); human perturbations to this are relatively very small (Forster et al, 2007). However, the large ocean surface and moisture from plants provide an important water source that can yield more water vapor to warmer air; relative humidity tends to remain nearly constant as climate changes, so warming for any reason introduces more water vapor to the air and increases the greenhouse effect in a positive feedback (Pierrehumbert et al., 2007; Hansen et al., 1984). Hence, focus is especially on carbon dioxide, and to a lesser degree on methane and other greenhouse gases (Forster et al., 2007).

Carbon dioxide concentrations in the atmosphere are tied into an extensive natural system of terrestrial, atmospheric and oceanic sources and sinks called the global carbon cycle, an excellent discussion of which is given by Prentice et al. 2001 in the IPCC 3rd

Assessement Report. The possible impact of increasing CO_2 levels in the atmosphere was first recognized by Arrhenius (1896). By the 1930's, mathematical models linking greenhouse gases and climate change (Callendar, 1938) predicted that a doubling of atmospheric CO_2 concentration would increase the mean global temperature by 2°C, with considerably more warming at the poles. (Le Treut et al., 2007 provides a detailed historical perspective on the recognition of Earth's greenhouse effect.) By the 1970s, CH_4, N_2O and CFCs were widely recognized as important additional anthropogenic greenhouse gases (Ramanathan, 1975).

The direct relationship between climate change and greenhouse gases such as CO_2 and methane is clearly described by the recent IPCC report (IPCC, 2007). Both the pattern of observed warming in the direct observational record, especially the record of the past 30 years, as well as climate model simulations, suggest that the Arctic will be more impacted by increases in greenhouse gas concentrations than any other region on Earth (**Figure 4.4).**

4.2.5 Plate Tectonics

The drifting of continents (plate tectonics) moves land masses from equator to pole or back, opens and closes oceanic "gateways" between land masses to redirect ocean currents, raises mountain ranges to redirect winds, and causes other changes that may affect climate. These changes can have very large local to regional effects (moving a continent from the pole to the equator obviously will greatly change the climate of that continent). Moving continents around may have some effect on global-average temperature, in part through changes in the planet's albedo (Donnadieu et al., 2006).

Processes linked to continental rearrangement can strongly affect global climate by altering the composition of the atmosphere and thus the strength of the greenhouse effect, especially through control of the carbon-dioxide concentration of the atmosphere (e.g., Berner, 1991; Royer et al., 2007). Over millions of years, the atmospheric concentration of carbon dioxide is controlled primarily by the balance between carbon-dioxide removal through chemical reactions with rocks near the Earth's surface, and carbon-dioxide release from volcanoes or other pathways involving melting or heating of rocks that sequester carbon dioxide. Because higher temperatures cause carbon dioxide to react more rapidly with Earth-surface rocks, atmospheric warming tends to speed removal of carbon dioxide from the air and thus to limit further warming, in a negative feedback (Walker et al., 1981).

4.2.6 Biological Processes

Biological processes can be important in the uptake and release of carbon dioxide, such that evolutionary changes have contributed to atmospheric changes. For example, some of the carbon dioxide taken from the air by plants is released by their roots into the soil, through respiration while living and by decaying after death, thus speeding the reaction of atmospheric carbon dioxide with rocks compared to the no-plant situation (Berner, 1991; Beerling and Berner, 2005). This process could not have occurred on the early Earth before the evolution of plants with roots.

Plants are composed in part of carbon dioxide removed from the atmosphere, and decomposition or burning of plants releases most of this carbon dioxide back to the atmosphere (minus the small fraction that reacts with rocks in the soil). When plants are buried without decomposition to form fossil fuels, the atmospheric carbon-dioxide level is reduced; later, natural processes may bring the fossil fuels back to the surface to decompose, releasing the stored carbon dioxide. (Humans are greatly accelerating these natural processes; the economically viable fossil fuels accumulated over hundreds of millions of years but are being burned over hundreds of years.) Rapid burial favors preservation of organic matter, whereas leaving dead things on the surface allows them to decompose, so changes in rates of sediment deposition linked to continental rearrangement are among the processes that may affect the formation and breakdown of fossil fuels and thus the strength of the atmospheric greenhouse effect.

Continents move about as rapidly as fingernails grow, so that a major reshuffling of the continents requires ~100 million years, and the opening or closing of an oceanic gateway may require millions of years (e.g., Livermore et al., 2007). Major evolutionary changes have required millions of years or longer (e.g., d'Hondt, 2005). Thus, those changes in the greenhouse effect affecting Earth's climate and linked to continental drift or biological evolution, which have been highly influential over times of tens of millions of years, have had essentially no effect over times of centuries or millennia. (Note that over hundreds of thousands of years or longer, an increase in volcanic activity may notably increase carbon dioxide in the atmosphere, causing warming; however, volcanic release of carbon dioxide is small enough that over millennia or less the changes in volcanism have not notably affected the carbon-dioxide concentration of the atmosphere, and the main short-term effect of an increase in volcanic eruptions is to cool the planet by blocking the sun, as discussed next.)

4.2.7 Volcanic eruptions

Volcanic eruptions are an important natural cause of climate change on seasonal to multi-decadal time scales. Large explosive volcanic eruptions inject both particles and gases into the atmosphere. Although particles are removed by gravity relatively rapidly (days to weeks), sulfur gases generated by the eruption injected into the stratosphere convert rapidly to sulfate aerosols (tiny droplets of sulfuric acid) with a residence time of about 3 years, and are transported around the world and poleward by circulation within the stratosphere. Tropical eruptions typically influence both hemispheres, whereas effects of eruptions at middle to high latitudes are usually restricted primarily to the hemisphere of eruption. Consequently, the Arctic is impacted primarily by tropical and Northern Hemisphere eruptions.

The radiative and chemical effects of the global volcanic aerosol cloud produce significant responses in the climate system on short timescales (see Figure 6.5) (Briffa et al, 1998; deSilva and Zielinski, 1998, Oppenheimer, 2003). By scattering and reflecting some solar radiation back to space, the aerosols cool the planetary surface, but by absorbing both solar and terrestrial radiation, the aerosol layer also heats the stratosphere. For a tropical eruption, this heating is larger in the tropics than in the high latitudes, producing an enhanced pole-to-equator stratospheric temperature gradient, especially in winter. In the Northern Hemisphere winter, this steeper gradient produces a stronger jet stream and a characteristic stationary tropospheric wave pattern that brings warm tropical air to Northern Hemisphere continents, resulting in winter warming. Because little solar energy reaches the Arctic during winter months, the transfer of warm air from tropical sources to high latitudes has more impact on winter temperatures than does the radiative cooling effect from the aerosols. However, during the summer months, radiative cooling dominates, resulting in anomalously cold summers across most of the Arctic. The 1991 Mt. Pinatubo eruption in the Philippines resulted in volcanic aerosols covering the entire planet, producing global-average cooling, but winter warming over the Northern Hemisphere continents in the subsequent two winters (Stenchikov et al., 2004, 2006).

The impacts of three large historical Northern Hemisphere eruptions have been studied in detail. The 939 AD Eldgjá (Iceland), 1783-1784 AD Laki (Iceland), and 1912 AD Novarupta (Katmai, Alaska) eruptions all caused cooling of the Arctic during summer, but no winter warming (Oman et al., 2005, 2006; Thordarson et tal., 2001).

When widespread stratospheric volcanic aerosols fall, some of the sulfate reaches the Antarctic and Greenland ice sheets (**Figure 4.5**), allowing estimation of the sun-blocking effect of the eruption from ice-core records. Large volcanic eruptions, especially when occurring within a few years or decades of other large eruptions, are thought to have played a significant role in cooling during the Little Ice Age (~1280-1850 AD; Anderson et al, 2008). A comprehensive review of the effects of volcanic eruptions on climate and of records of past volcanism is provided by Robock (2000, 2007).

The effects of volcanic eruptions are clearly evident in ice-core records (e.g., Zielinski et al., 1994), with major eruptions giving cooling of ~1^{o}C for ~1-2 years in Greenland ice-core records (e.g., Stuiver et al., 1995, **Figure 4.6**). The growth and shrinkage of the great ice-age ice sheets, and the associated bending and loading/unloading of the Earth, may have affected the frequency of volcanic eruptions somewhat (e.g., Maclennan et al. 2002), but in general the recent timing of explosive volcanic eruptions appears to be random. There is no mechanism for a volcano in, say, Alaska to synchronize its eruptions with a volcano in Indonesia; hence, volcanic eruptions over recent millennia appear to have introduced unavoidable climatic “noise” as opposed to controlling the climate in an organized way.

4.2.8 Others

Paleoclimatic records discount some speculative mechanisms of climate change. For example, about 40,000 years ago natural fluctuations in the strength of the Earth’s magnetic field reduced it essentially to zero for about one millennium. The cosmic-ray flux into the Earth system increased greatly, as recorded by a large peak in beryllium-10 in sedimentary records. However, the climate did not track the beryllium-10, indicating that the cosmic-ray increase had little or no effect on the climate (Muscheler et al., 2005). Large changes in concentration of extraterrestrial dust between the Earth and sun might lead to changes in solar energy reaching the Earth and thus to changes in climate; however, the available evidence to date shows no significant changes in such extraterrestrial dust, based on the rate of infall recorded in sedimentary archives (Winckler and Fischer, 2006).

The climate is a complex, integrated system, with important linked feedbacks, internal variability, and numerous forcings. On timescales of centuries or less, however, many of the drivers of past climate change—including drifting continents, biological evolution, aging of the sun, and features of Earth’s orbit—have no discernible influence on the climate. Small variations in climate appear to be linked to the small variations in the sun’s output, and explosive volcanic eruptions produce occasional cooling. Greenhouse-gas changes affect the planet’s temperature directly.

4.3 Reading the History of Climate through Proxies

A modern historian trying to understand our human story cannot go back in time and replay an important event. Instead, the historian must rely on indirect evidence: eyewitness accounts (which may or may not be highly accurate), artifacts, and more. It is as if the historical figures, who cannot tell their tale directly, have given their proxies to other people and other things to deliver the story to the modern historian.

Historians of climate—paleoclimatologists—are just like other historians, reading the indirect evidence that the past sends by proxy. All trained historians are aware of the strengths and weaknesses of proxy evidence, of the value of weaving multiple strands of evidence together to form the complete fabric of the story, of the necessity of knowing when things happened as well as what happened, and of the ultimate value of using history to inform understanding and guide choices.

Some of the proxy evidence used by paleoclimatologists would be familiar to more-traditional historians. Written accounts of many different activities often include notes on the weather, on presence or absence of ice on local water bodies, and on times of planting or harvest and the crops that grew or failed. If care is taken to average across extreme events, to account for the tendency of people to report the rare rather than the commonplace, and to include the effects of changes in husbandry and other issues, written records can contribute to knowledge of climate back through written history. However, almost all of the Earth's climatic history happened at times and in places for which human accounts are lacking. The paleoclimatologist is forced to rely on evidence that is less familiar to most people than are written records. Remarkably, these natural proxies may reveal as much as, or even more than, the written records.

4.3.1 Climate's Proxies

Much of the history of a civilization can be reconstructed from the detritus its people left behind. Similarly, paleoclimate records are typically developed through analysis of sediment, broadly defined. "Sediment" may include the ice formed as years of snowfall pile up into an ice sheet, the mud accumulating at the bottom of the sea or a lake, the annual layers of a tree, the thin sheets of mineral laid one on top of another to form a stalagmite in a cave, the piles of rock bulldozed by a glacier, the piles of desert sand dropped into dunes by the wind, the odd things collected and stored by packrats, and more (e.g., Crowley and North, 1991; Bradley, 1999; Cronin, 1999). For a sediment to be useful, it must: i) preserve a record of the conditions when it formed (i.e., subsequent events cannot have erased the original story and replaced it with something else); ii) be interpretable in terms of climate (the characteristics of the deposit must uniquely relate to the climate at the time of formation); and iii) be "datable" (i.e., there must be some way to determine the time when the sediment was deposited).

As an example, long records of Earth's climate are commonly reconstructed from climate proxies preserved in deep-ocean sediments. One of the best-known proxy records of climate change is that recorded by benthic (bottom-dwelling) foraminifera, microscopic organisms that live on the sea floor and secrete their calcium-carbonate shells in equilibrium with the sea water. The isotopes of oxygen in the carbonate are a function of both the water temperature (which often does not change very rapidly over time or very steeply over space in the deep ocean) and changes in global ice volume. Global ice volume determines the relative abundances of the isotopes oxygen-16 to oxygen-18 in seawater over time. Snow has relatively less of the heavy oxygen-18 than is in its seawater source. Consequently, as ice sheets grow on land, the ocean becomes enriched in the heavy oxygen-18, and this enrichment is recorded by the oxygen isotopic composition of foraminifera shells. The proportion of the heavy and light isotopes of oxygen is usually expressed as $\delta^{18}O$, with positive $\delta^{18}O$ values reflecting extra amounts of the heavy isotope of oxygen, and negative values reflecting samples with less of the heavy isotope than average seawater. Although the $\delta^{18}O$ of foraminifera shells does not reveal where the glacial ice buildup was located, the record does provide a globally-integrated value of the amount of glacial ice on land, especially if appropriate corrections are made for temperature changes using other indicators. Positive $\delta^{18}O$ reflects colder times (more ice), whereas more negative $\delta^{18}O$ reflects interglacial (warmer- less ice) periods in Earth's history. In the absence of changes in global ice volume, changes in benthic foraminifera $\delta^{18}O$ reflect changes in ocean temperatures, with more positive $\delta^{18}O$ indicative of colder water, and more negative $\delta^{18}O$ indicating warmer water.

Written documents have sometimes been erased and rewritten, in a deliberate attempt to distort history, or simply because the paper was more valuable than the original words. Paleoclimatologists are continually watching for any signs that a climate record has been "erased" and "rewritten" by events since deposition of the sediment. Occasionally, this proves to be important. For example, water may remove isotopes carrying paleoclimatic information from shells, replacing them with other isotopes telling a different story (e.g., Pearson et al., 2001). However, except for the very oldest deposits from early in Earth's history, it is usually rather easy to tell whether a record has been altered, and this problem should not affect any of the conclusions presented in this report.

Finding the linkage between climate and some characteristic of the sediment is then required. The climate is recorded in myriad ways by physical, biological, chemical, and isotopic characteristics of sediments.

Physical indicators of past climate are often easy to read and understand. For example, a sand dune can form only if dry sand is available to be blown around by the wind, without being held down by plant roots. Except near beaches (where fluctuations in water level reveal bare sand), a dry climate is needed to keep grass off the sand so it can blow around. Today in northwestern Nebraska, the huge dune field of the Sand Hills is grass-covered (**Figure 4.7**). Thus, drier conditions in the past allowed formation of the dunes, with wetter conditions now allowing grass to grow on top (e.g., Muhs et al., 1997). Similarly, the sediments left by glaciers are readily identified, and occurrence of such sediments in areas that are ice-free today attests to changing climate. A very different physical indicator of past climate is given by temperatures measured in boreholes. Just as a Thanksgiving turkey placed in an oven takes a while to warm in the middle, the two-mile-thick ice sheet of Greenland has not finished warming from the ice age, and the cold temperatures below reveal how cold the ice age was (Cuffey and Clow, 1997).

Many paleoclimate records are based directly on living things. Tundra plants are quite different from those living in temperate forests. If the deep layers of a sediment core contain pollen, seeds, twigs, etc. of tundra plants, with temperate-forest plants in shallow layers, a warming from a formerly cold time is indicated. Trees grow more rapidly, adding thicker rings, when climatic conditions are more favorable. In very dry regions, this allows trees to be used in reconstruction of rainfall; in cold regions, growth may be more closely linked to temperature.

Chemical analysis of sediments may reveal additional information about past climates. As one example, some single-celled organisms in the ocean change the chemistry of their cell walls in response to changing temperature, using more-flexible molecules to offset the increase in brittleness caused by colder temperatures. These are sturdy chemicals that persist in sediments after death, so the history of the ratio of stiffer to less-stiff molecules in a sediment core provides a history of the temperature at which the organisms grew (in this case, the organisms are prymnesiophyte algae, the chemicals are alkenones, and the frequency of carbon double bonds controls the stiffness; Muller et al., 1998; other such indicators also exist).

Isotopic ratios are among the most-commonly used of proxy indicators of past climates. Consider just one example, providing one of the ways to determine the past concentration of carbon dioxide. All carbon atoms have 6 protons in their nuclei, most have 6 neutrons (making carbon-12), but some have 7 neutrons (carbon-13) and a few have 8 neutrons (radioactive carbon-14). The only real difference between carbon-12 and carbon-13 is that carbon-13 is a bit heavier. The lighter carbon-12 is “easier” for plants to use, so growing plants preferentially incorporate carbon from carbon dioxide containing only carbon-12 rather than carbon-13. However, if carbon dioxide is scarce in the environment, the plants cannot be picky and must use what is available. Hence, the carbon-12:carbon-13 ratio in plants provides an indicator of the availability of carbon dioxide in the environment. The sturdy cell-wall chemicals described in the previous paragraph can be recovered and analyzed for their carbon isotopes, providing an estimate of the carbon-dioxide concentration at the time the algae grew (e.g., Pagani et al., 1999).

Much of the science of paleoclimatology is devoted to calibration and interpretation of the relation between sediment characteristics and climate (see National Research Council, 2006). The relationship of some indicators to climate is relatively straightforward, but others may be complex. The width of a tree ring, for example, is especially sensitive to water availability in dry regions, but may also be influenced by changes in shade from neighboring trees, the attack of beetles or other pests that weaken a tree, the temperature, and more. Extensive efforts go into calibration of paleoclimatic indicators against the climatic variables. Because paleoclimatic data cannot be collected everywhere, additional work is devoted to determining which areas of the globe have climates that can be reconstructed from the available paleoclimatic data. Wherever possible, multiple indicators are used to reconstruct past climates, and to assess agreement or disagreement (National Research Council, 2006), with major results typically resting on multiple lines of evidence.

4.3.2 The Age of the Sediments

History requires “when” as well as “what”. Many techniques reveal the “when” of sediments, sometimes to the nearest year, other times with less precision.

Climate records developed from most trees, and from some ice cores and sediment cores, can be dated to the nearest year by counting annual layers. The yearly nature of tree rings from seasonal climates is well-known. A lot of checking goes into demonstrating that layers observed in ice cores and special sediment cores are annual, but in some cases they clearly are (Alley et al., 1997), allowing quite accurate counts. The longest-lived trees may reach 5000 years, but use of overlapping living and dead wood has allowed extension of records to more than 10,000 years (Friedrich et al. 2004), and the longest annually-layered ice cores recovered to date extend beyond 100,000 years (Meese et al., 1997). However, relatively few records can be absolutely dated in this way.

Other techniques that have been used for dating include measuring the damage that accumulates from cosmic rays striking things near the Earth's surface, observing the size of lichen colonies growing on rocks deposited by glaciers, and identifying the fallout of particular volcanic eruptions that can be dated by historical accounts or annual-layer counting.

Most paleoclimatic dating uses the decay of radioactive elements. Radiocarbon is commonly used for samples containing carbon from the most recent 40,000 years or so (older than that, so little radiocarbon survives that measurement is difficult). Many other isotopes are used for different time intervals and materials. Intercomparison with annual-layer counts, with historical records, and between different techniques shows that quite high accuracy can be obtained, so that it is often possible to have errors in age estimates of less than 1%. (That is, if an age is quoted as being 100,000 years, the event actually occurred sometime between 99,000 years and 101,000 years ago.)

4.4 Cenozoic Global History of Climate

As emphasized in the Summary for Policymakers of IPCC (2007) and in the body of the report, a paleoclimatic perspective is important for understanding the climate system including its forcings and feedbacks. Arctic records, and especially Arctic ice-core records, have provided key insights. The material that follows provides a brief overview of selected features in the history of Earth's climate, and of the forcings and feedbacks of those climate events. This is not a comprehensive treatment of the extensive literature on these topics, but is provided here as a primer to help place the main results of this report in context. Kump et al. (2003) is a more-complete yet accessible introduction.

This report focuses on the Cenozoic Era, the 65 Ma beginning with the demise of the dinosaurs and continuing through today (see section 4.5 for a discussion of the chronology used in this report). Over most of this 65 Ma interval, deep-sea records of foraminifera $\delta^{18}O$ (a powerful paleoclimatic indicator, described above in section 4.4.1) integrated from sedimentary archives of several ocean basins show that the Earth has been warmer than at present with lower ice volume (**Figure 4.8**). Yet, following the peak warming of the early Eocene, about 50-55 Ma ago, global temperatures generally declined (Miller et al., 2005). While this record is not specific about Arctic climate change, the record indicates that the global gradient (or difference) in temperature between the polar regions and the tropics was smaller when global climate was warmer, and that this gradient increased as the high latitudes progressively cooled (Barron and Washington, 1982), with effects on atmospheric and oceanic circulation. The overall cooling trend of the past 55 Ma was punctuated by intervals during which the cooling was reversed and the oceans warmed, only to cool rapidly again at a later time. For example, rapid decreases in foraminifera $\delta^{18}O$ at about 34 Ma ago, and again at about 23 Ma are thought to reflect the rapid build up of ice over Antarctica in only a few hundred thousand years (Zachos et al. 2001). The Paleocene-Eocene thermal maximum (~55Ma ago) represents a major interval of global warming with CO_2 levels estimated to have risen abruptly (Shellito et al. 2003), perhaps due to the rapid release of methane from the oceans (Bralower et al., 1995).

The style and tempo of global climate change over the past 5.3 million years is depicted well by the foraminifera $\delta^{18}O$ record of Lisiecki and Raymo (2005; **Figure 4.9,** see section 4.4.1 for a discussion of this proxy). This composite record provides a well-dated stratigraphic tool against which other records from around world can be compared. The foraminifera $\delta^{18}O$ record reflects both changes in global ice volume and oceanic bottom-water temperature change, but with the same sense—increase in global ice, or decrease in ocean temperatures, push the indicator in the same direction. The foraminifera $\delta^{18}O$ record indicates low-magnitude climate changes from 5.3 until about 2.7 Ma ago, when the amplitude of the foraminifera $\delta^{18}O$ signal increased markedly. This shift in foraminifera $\delta^{18}O$ amplitude coincides with widespread indications of onset of northern continental glaciation (see Chapter 5). The oxygen isotopic fluctuations since 2.7 Ma ago are commonly used as a global index of the frequency and magnitude of glacial-interglacial cycles. In addition to the fluctuations, the data show that over the past 3 Ma, average ocean temperatures have been dropping. Global circulation models constrained by extensive paleoclimatic data targeting the late Pliocene interval from 3.3 to 3.0 Ma ago suggest that temperatures globally were warmer by as much as 2-3 $^{\circ}C$ at that time (see Jiang et al. 2005 and IPCC, 2007).

The large fluctuations in foraminifera $\delta^{18}O$ beginning about 2.7 Ma exhibited clear periodicities matching those of the Milankovitch forcing (with those periodicities also present in smaller, older fluctuations). A 41 ka periodicity was especially apparent, as well as the 19-23 ka periodicity. More recently, over the last 0.9 Ma or so, the variations in $\delta^{18}O$ became even bigger, and while the 41 and 19-23 ka periodicities continued, a 100 ka periodicity became dominant. The reasons for this shift remain unclear, and the focus of much research (Ruddiman, 2006; Lisiecki and Raymo, 2007, Huybers, 2007).

Moving toward the present, the number of available records increases greatly, as does typical time resolution of the records (see section 4.4). The large ice-age cycling of the last 0.9 Ma produced growth and retreat of extensive ice sheets across broad regions of North America and Eurasia, as well as smaller extensions of ice in Greenland, Antarctica, and many mountainous areas. Ice in North America covered New York and Chicago, for example. The water that composed those ice sheets had been removed from the oceans, causing non-ice-covered coastlines typically to lie well beyond modern boundaries. Melting of ice sheets exposed land that had been ice-covered while covering land beneath the rising seas, with relatively small net effect (e.g., Kump and Alley, 1994). Strong temperature changes occurred with the ice-age cycling, of many degrees to tens of degrees in some places (see Chapter 5).

Large, abrupt jumps in climate (see section 6.4.3) occurred during at least the most recent of the glacial intervals and probably during earlier ones. In records from near the North Atlantic such as Greenland ice cores, roughly half of the total difference between glacial and interglacial conditions was achieved in many climate-change indicators over times of decades to years. Changes away from the North Atlantic were notably smaller, and in the far south the changes appear to have exhibited a see-saw pattern (southern warming with northern cooling). The "shape" of the climate records is interesting, with northern records typically showing abrupt warming, gradual cooling, abrupt cooling, near-stability or slight gradual warming, and then repeating.

The most recent interglacial interval has lasted slightly over 10,000 years. Generally warm conditions have prevailed compared to the average over the last 0.9 Ma. However, important changes have been observed. These include broad warming and then cooling over millennia, abrupt events probably linked to the older abrupt changes, and additional events with various spacings and sizes that have a range of causes, which will be described more in chapters 5 and 6.

4.5 Chronology

In any discussion of past climate periods, it is necessary to use a timescale with terminology that is understandable to all readers. Beyond the historic period, this means use of time periods that are within the realm of geology. In this report, we use two sets of terminology for prehistoric time periods, one for the longer history of the Earth and one for much more recent Earth history, approximately the past 2.6 Ma (the Quaternary Period). For the longer period of Earth history, we use the terminology and timescale adopted by the International Commission on Stratigraphy (Ogg, 2004). This timescale is well-established and has been widely accepted throughout the geologic community. The Quaternary Period is the youngest geologic period in this timescale and constitutes the past ~2.6 million years (http://www.stratigraphy.org/gssp.htm, **Figure 4.10,** Jansen et al., 2007). The Quaternary Period is of particular interest in this paper, because it is this time that is characterized by dramatic changes in climate, from glacial to interglacial periods.

There are problems associated with the use of timescales within the Quaternary Period. These problems are common to all geologic dating, but assume additional importance in the Quaternary because the focus during this geologically short, recent period is on relatively short-lived events. Very few geologic records for the Quaternary Period are continuous, well-dated, and applicable to all other records of climate change. Furthermore, many geologic deposits preserve records of events that are *time-transgressive* or *diachronous*. This means that a particular geologic event of paleoclimatic significance is recorded earlier at one geographic location or region than another.

A good example of time-transgression is the most recent deglaciation of midcontinental North America, namely the retreat of the Laurentide ice sheet. Although this marks a major shift in a climate state, from a glacial period to an interglacial period, by its very nature it is an event that occurred at different times in different places. In midcontinental North America, the Laurentide ice sheet had begun to retreat from its southernmost position in central Illinois after ~22.6 ka ago, but was still present over what is now northern Illinois until after ~15.1 ka ago, and was still in Wisconsin and Michigan until after ~12.9 ka ago (Johnson et al., 1997, [radiocarbon ages converted using the algorithm of Fairbanks et al., 2005]). Thus, the geologic record of when the present "interglacial" period began is older in central Illinois than it is in northern Michigan, which in turn is older than it is in southern Canada. Time-transgression as a concept also applies to phenomena other than geologic processes. Migration of plant communities or biomes as a result of climate change is not an instantaneous process over a wide geographic region. Thus, many records of climate change that reflect changes in plant communities will take place at different times across a region as taxa within that community migrate.

Another difficulty with some geologic records is that local terminology differs from continent to continent and creates confusion when a topic is being addressed in a global context. For example, "Sangamon" is the name of the last interglacial period in the mid-continent of North America (Johnson et al., 1997) and the term "Eemian" is used for the last interglacial period in Europe. However, North American workers apply the term Sangamon primarily to rock-stratigraphic records (tills deposited by glaciers, and old soils called paleosols). The Sangamon interglacial is considered to have lasted several tens of thousands of years, because no glacial ice was present in the mid-continent between the last major glacial event ("Illinoian") and the most recent one ("Wisconsinan"). In contrast, the term Eemian, for European workers, is often applied to pollen records, and is reserved for a period of time, perhaps less than 10,000 years, when climate conditions were as warm or warmer than present.

Nevertheless, it is crucial that at least some terminology is used as a common basis for discussion of geologic records of climate change during the Quaternary. In this paper, we have chosen to use the *stages* of the oxygen isotope record from foraminifera in deep-sea cores as our terminology for discussing different intervals of time within the Quaternary Period. The identification of glacial-interglacial changes in the deep-sea-core record, and the naming of stages for them, began with a landmark paper by Emiliani (1955). Foraminifera in the ocean record shifts in climate from glacial-to-interglacial states in the oxygen isotope composition of their carbonate skeletons (see section 4.4.1, above). These shifts are due both to changes in ocean temperature and changes in the isotopic composition of seawater. The latter result from the shifts in oxygen isotopic composition of seawater, in turn a function of ice volume on land. Because the temperature and ice-volume influences on foraminiferal oxygen-isotope compositions are in the same direction, the record of glacial-interglacial changes in deep-sea cores is particularly robust.

The oxygen isotope record of glacial-interglacial cycles has been studied and well documented in hundreds of deep-sea cores. The same glacial-interglacial cycles are easily identified in cores from all the world's oceans (Bassinot, 2007). It is, therefore, truly a continuous and global record of climate change within the Quaternary Period. Furthermore, a wide variety of geologic records of climate change have shown the same glacial-interglacial cycles that allow comparison and correlation to the deep-sea record, including glacial records (e.g., Andrews and Dyke, 2007; Booth et al., 2004), ice cores (e.g., NGRIP, 2004; Jouzel et al., 2007), cave carbonates (e.g., Winograd et al., 1992, 1997), and eolian sediments (e.g., Sun et al., 1999). Furthermore, deep-sea cores themselves sometimes contain, in addition to foraminifera, other records of climate change, including pollen records of past vegetation (e.g., Heusser et al., 2000) or eolian (wind-deposited) sediments that record glacial and interglacial climates on land (e.g., Hovan et al., 1991).

The timescales that have been developed for the oxygen isotope record are important to understand. The mostly widely used timescales are those that have been developed by use of "stacked" deep-sea core records (i.e., multiple core records, from more than one ocean) that are in turn, "tuned" or "dated" by a combination of identification of dated paleomagnetic events and an assumed forcing of climate change by changes in the parameters related to Earth-sun orbital geometry, precession and obliquity. (Initially, dated paleomagnetic events were used with an assumed constant sedimentation rate to provide a first estimate of the timing of the main variations in the climate. These very closely matched the known periodicities in the Earth-sun orbital geometry, to a degree that provided very high confidence that those known periodicities were affecting the climate. Then, this result was used to fine-tune the dating by adjusting the sedimentation rates to allow closer match between the data and the orbital periodicities.) The practice is often referred to as "astronomical" or "orbital" tuning. The strategy behind "stacking" multiple records is to eliminate possible local effects on a core and present a smoothed, global record. There have been several, highly similar, timescales developed using this approach, the most commonly cited being the SPECMAP studies of Imbrie et al. (1984) and Martinson et al. (1987) (**Figure 4.11**), and the more recent work of Lisiecki and Raymo (2005).

It is important to recognize, however, that there are disadvantages to using the astronomically tuned oxygen isotope records. Very few deep-sea cores are dated directly, except in the upper parts, which are within the range of radiocarbon dating, or at widely spaced depths where paleomagnetic events are recorded. In addition, after the initial tests, the astronomical tuning approach *assumes* that the orbital parameters, particularly precession and obliquity, are the primary forcing mechanisms behind climate change on glacial-interglacial timescales in the Quaternary Period. There have been challenges to this assumption, based on directly dated cave calcite records (Winograd et al., 1992,

1997) and emergent coral reef terraces (Szabo et al., 1994; Gallup et al., 2002; Muhs et al., 2002), although in general the assumption appears to be more-or-less accurate.

Recognizing that there are assumptions inherent in the use of the SPECMAP timescale, we use this timescale and the marine oxygen isotope stage terminology in this paper for four reasons. These include: (1) the wide acceptance and use in the scientific community, (2) the continuous nature of the record, (3) the global aspect of the record, and (4) the ability to subdivide the periods of time under consideration. Regarding the latter, for example, the marine record can accommodate the problem in the use of "Sangamon," as used in North America vs. "Eemian, " in Europe for the last interglacial period. The Sangamon interglacial, as used by North Americans, includes all of marine isotope Stage 5 (MIS 5), as well as perhaps parts of MIS 4. However, the Eemian, as used by most European workers, would include only MIS 5e or 5.5, an interval *within* the greater MIS 5.

FIGURE CAPTIONS

Figure 4.1 Earth's energy budget is a balance between incoming and outgoing radiation. Numbers are in watts per square meter of the Earth's surface, and some estimates may be uncertain by as much as 20%. Incoming shortwave radiation from the sun enters the Earth's atmosphere and may be reflected by clouds, or absorbed or reflected as longwave radiation by the Earth. The greenhouse effect involves the absorption and reradiation of energy by atmospheric greenhouse gases and particles, resulting in a downward flux of infrared radiation (longwave) from the atmosphere to the surface (back radiation) causing higher surface temperatures. In this figure, Earth is in energy balance with the total rate of energy lost from Earth (107 W/m2 of reflected sunlight plus 235 W/m2 of infrared [long-wave] radiation) equal to the 342 W/m2 of incident sunlight (Kiehl and Trenberth, 1997).

Figure 4.2 Schematic of the Earth's orbital variations (Milankovitch cycles) that control the amount of sunlight received (insolation) at a given place on the Earth's surface. 'T' denotes changes in the tilt (or obliquity) of the Earth's axis which has a ~41 ka periodicity, 'E' denotes changes in the eccentricity of the orbit (due to variations in the minor axis of the ellipse) with a ~100 ka periodicity, and 'P' denotes precession, that is, changes in the direction of the axis tilt at a given point of the orbit, which has a ~19 to 23 ka periodicity. (Rahmstorf and Schellnhuber ,2006; Jansen et al., 2007)

Figure 4.3. Milankovitch-driven monthly insolation anomalies (deviations from present) from 21 to 0 ka and spanning the last interglacial and end of the previous glaciation 120 to 140 ka from 60 deg N. X-axis is time in thousands of years before present and Y axis is calendar months. Contours and numbers reflect anomalies from present in Watts m^{-2} (Data from Berger and Loutre, 1992). Midsummer insolation values 11 ka ago exceeded 40 Watts per m^{2,} whereas current values are less than 10 Watts per m^{2,}.

Figure 4.4 Mean surface temperature anomalies for the Earth relative to the period 1951-1980. Panel A is the global average. Panel B shows the temperature anomalies for the period from 2000 to 2005. High northern latitudes show the largest anomalies for this time period. (Hansen et al., 2006)

Figure 4.5 Volcanic evidence derived from 44 ice cores were recently used to simulate the spatial distribution of volcanic sulfate aerosols using the NASA Goddard Institute for Space Studies (GISS) ModelE climate model. Shown here are simulated loadings for the Laki (1783), Katmai (1912), Tambora (1815), and Pinatubo (1991) eruption deposition (kg/km2) in the Arctic region. The colors are defined so blue indicates smaller than average deposition for 66 N-82 N, 50 W- 35 W and yellow, orange, and red indicate larger than average. (from Gao et al., 2007)

Figure 4.6 Isotopic record of temperature response in Greenland snow to large volcanic eruptions reconstructed from the GISP2 ice core. (modified from Stuiver et al., 1995).

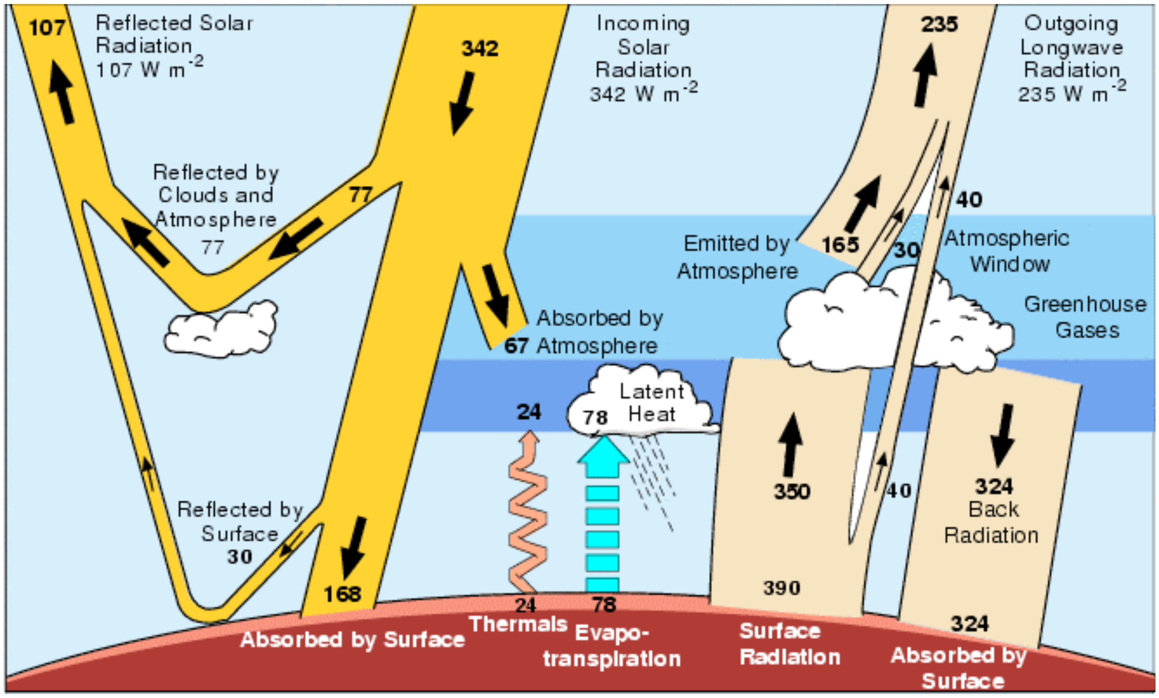
Figure 4.7 About a quarter of Nebraska is covered by the Sand Hills. These are Pleistocene sand dunes derived from glacial outwash eroded from the Rocky Mountains, and now stabilized by vegetation. The hills are characterized by crowded barchan (crescent-shaped) dunes, general absence of drainage, and numerous tiny lakes filling the closed depressions between dunes. Covering an area of 51,400 square kilometers, the Sand Hills are the largest sand dune formation in America. This ASTER simulated natural color image was acquired September 10, 2001, covers an area of about 57.9 x 61.6 km, and is centered near 42.1 degrees north latitude, 102.2 degrees west longitude. (Photo credit: NASA/GSFC/METI/ERSDAC/JAROS, and U.S./Japan ASTER Science Team)

Figure. 4.8. Global compilation of over 40 deep sea benthic $\delta^{18}O$ isotopic records taken from Zachos et al. 2001, updated with high resolution Eocene through Miocene records from Billups et al 2002; Bohaty and Zachos, 2003 and Lear et al., 2004. Dashed blue bars represent times when glaciers came and went or were smaller than now; the solid blue bars are ice sheets of modern size or larger. (Figure and text from IPCC Chapter 6 on Paleoclimate, Jansen et al., 2007)

Figure. 4.9. Composite stack of 57 different benthic oxygen isotope records from (a proxy for temperature) from a globally distributed network of marine sediment cores. (data from Lisiecki and Raymo, 2005 and associated website)**.** This foraminifera $\delta^{18}O$ record indicates low-magnitude climate changes from 5.3 until about 2.7 Ma ago, when the amplitude of the foraminifera $\delta^{18}O$ signal increases markedly.

Figure 4.10. Chronology used for geological time periods in this report. (modified from Ogg and 2004)

Figure 4.11. Marine Isotope Stage (MIS) nomenclature and chronology used in this report (after Imbrie et al. (1984) and Martinson et al. (1987)).



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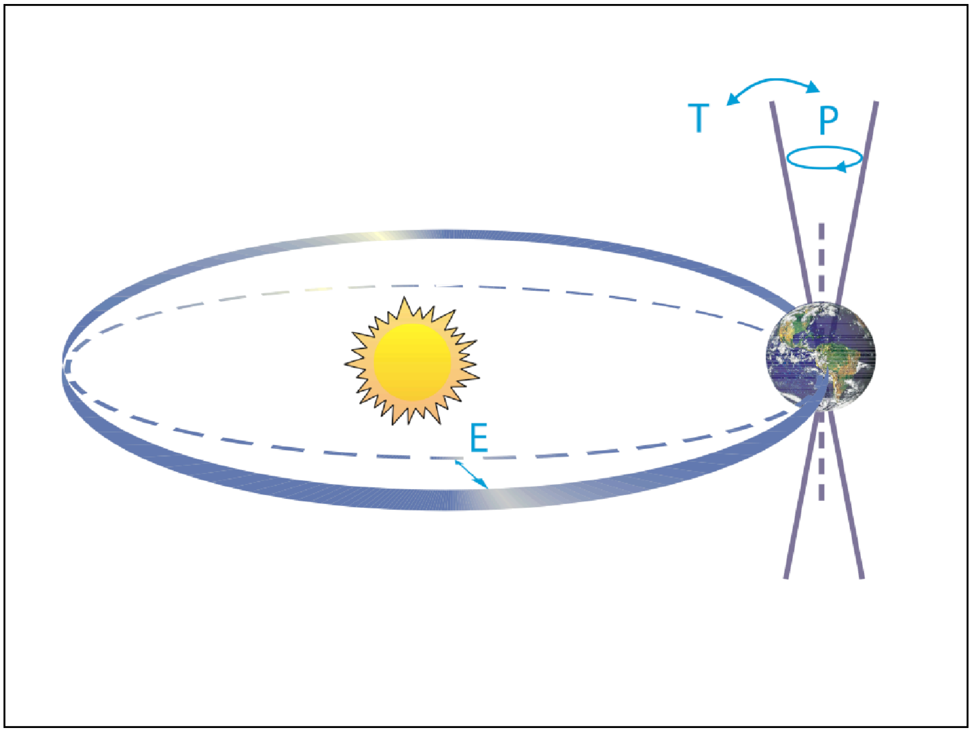

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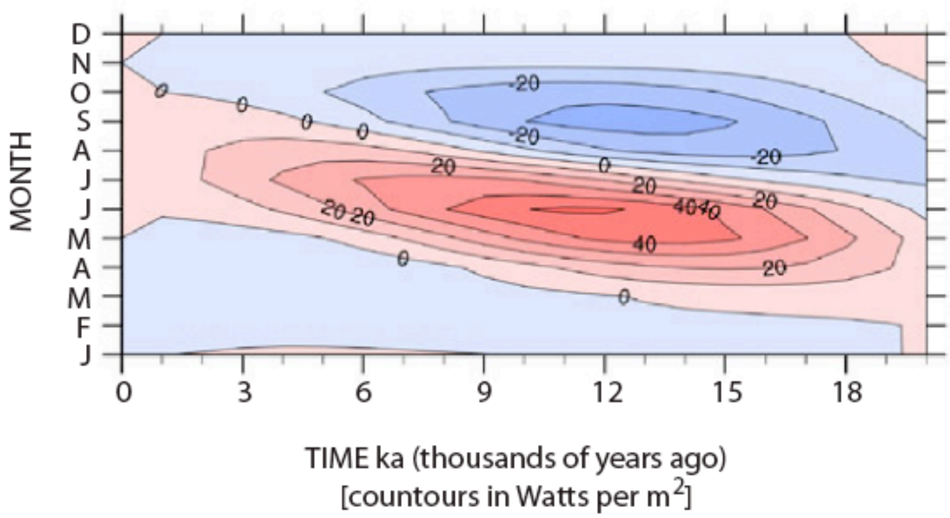

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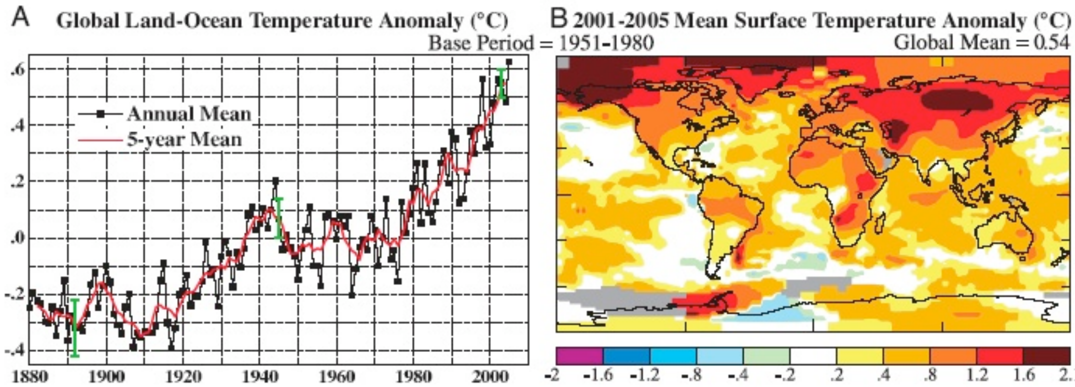

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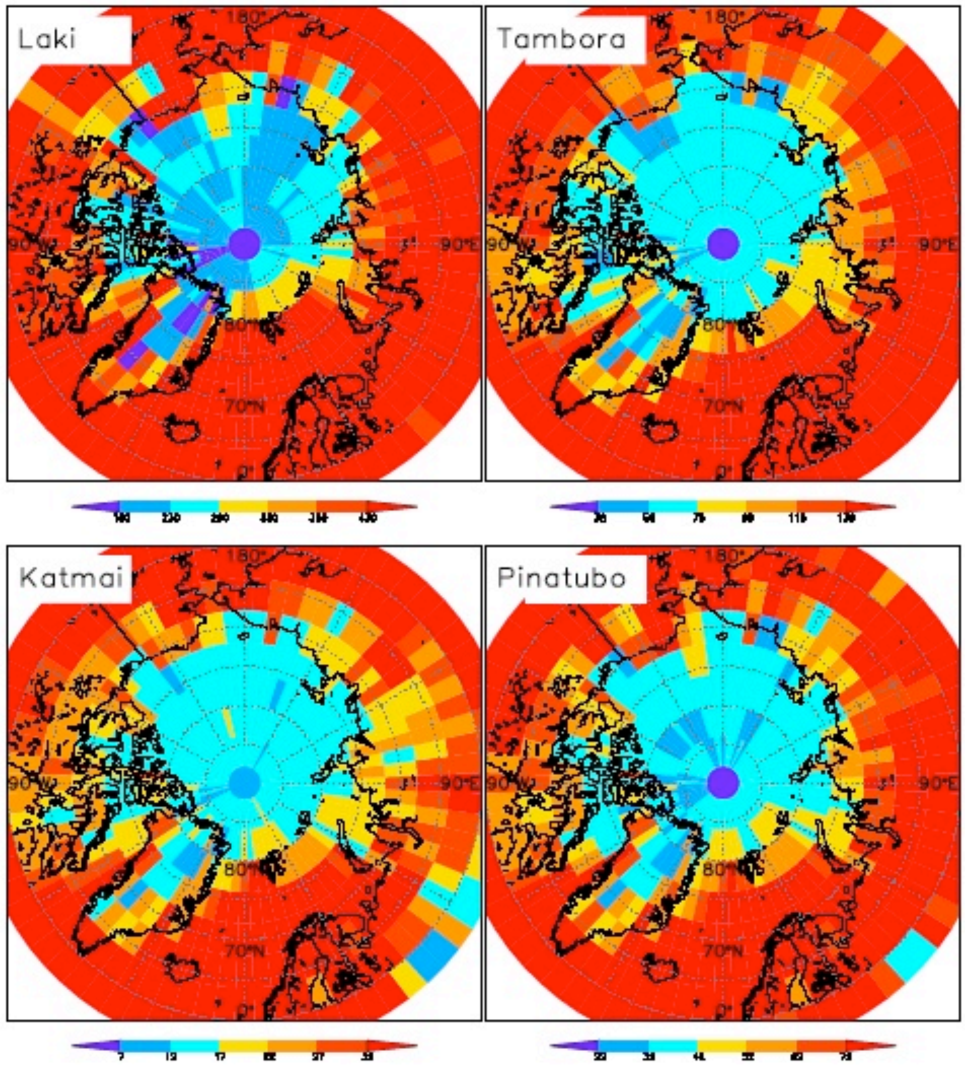

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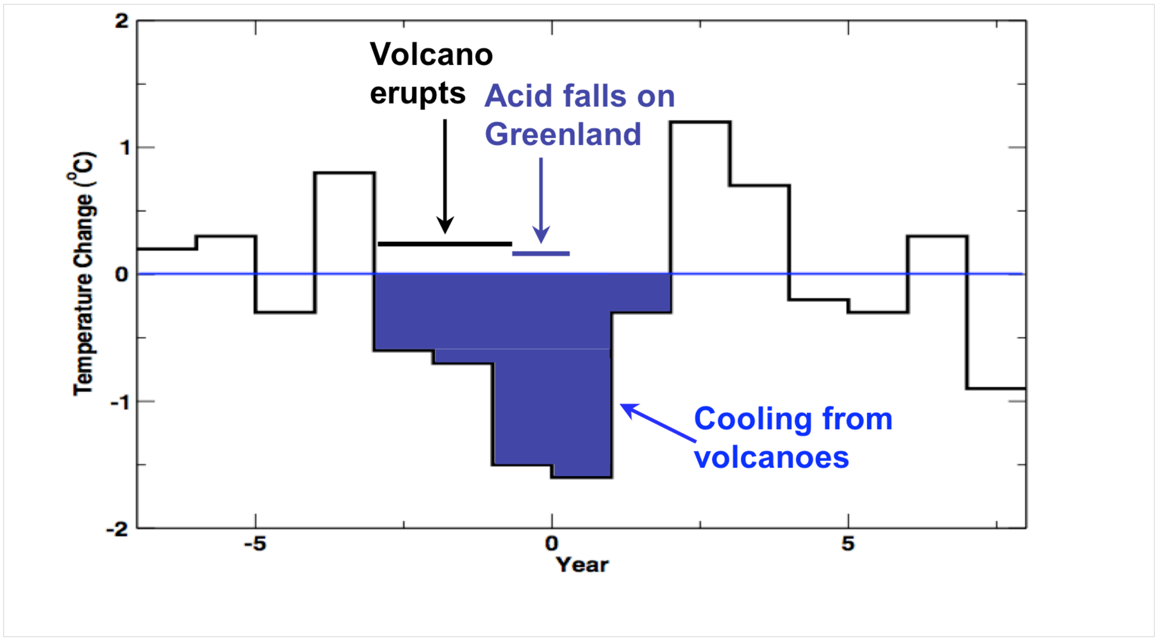

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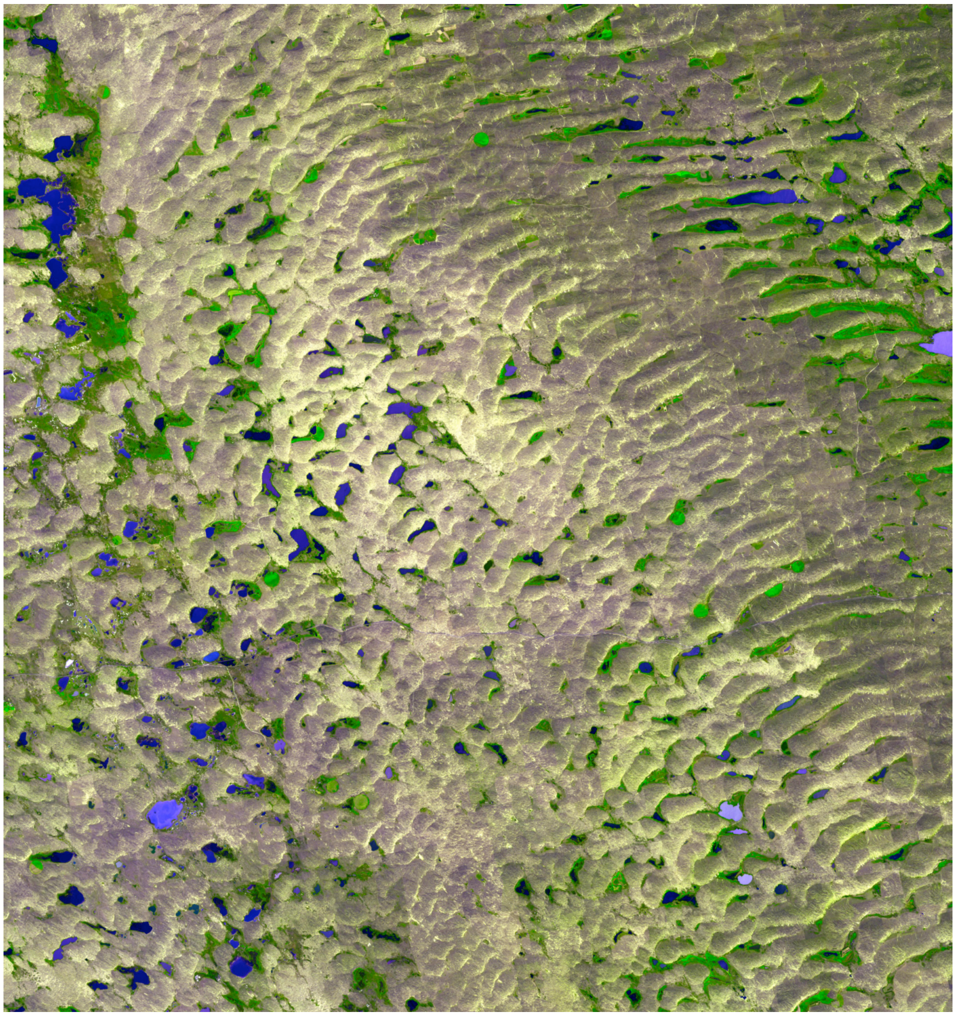

Figure 4.7 About a quarter of Nebraska is covered by the Sand Hills. These are Pleistocene sand dunes derived from glacial outwash eroded from the Rocky Mountains, and now stabilized by vegetation. The hills are characterized by crowded barchan (crescent-shaped) dunes, general absence of drainage, and numerous tiny lakes filling the closed depressions between dunes. Covering an area of 51,400 square kilometers, the Sand Hills are the largest sand dune formation in America. This ASTER simulated natural color image was acquired September 10, 2001, covers an area of about 57.9 x 61.6 km, and is centered near 42.1 degrees north latitude, 102.2 degrees west longitude. (Photo credit: NASA/GSFC/METI/ERSDAC/JAROS, and U.S./Japan ASTER Science Team)

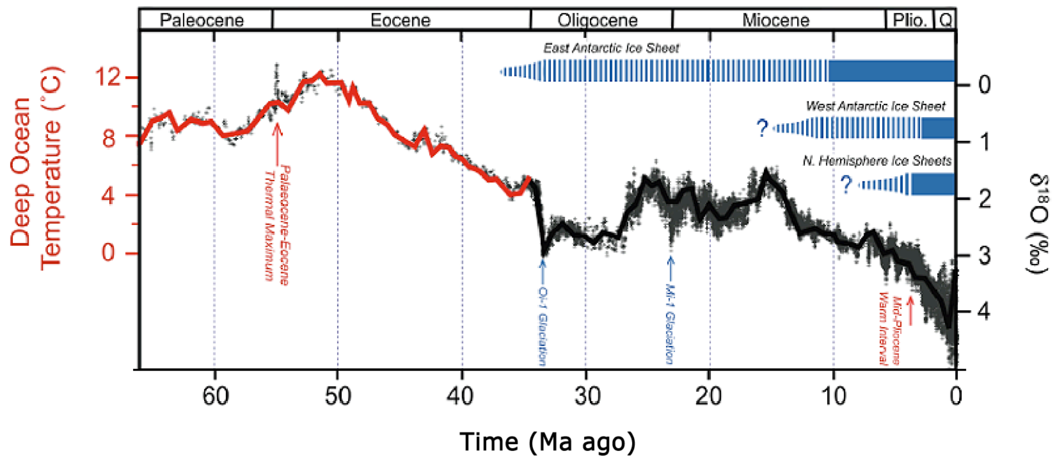

Figure. 4.8. Global compilation of over 40 deep sea benthic $\delta^{18}O$ isotopic records taken from Zachos et al. 2001, updated with high resolution Eocene through Miocene records from Billups et al 2002; Bohaty and Zachos, 2003 and Lear et al., 2004. Dashed blue bars represent times when glaciers came and went or were smaller than now; the solid blue bars are ice sheets of modern size or larger. (Figure and text modified from IPCC Chapter 6 on Paleoclimate, Jansen et al., 2007)

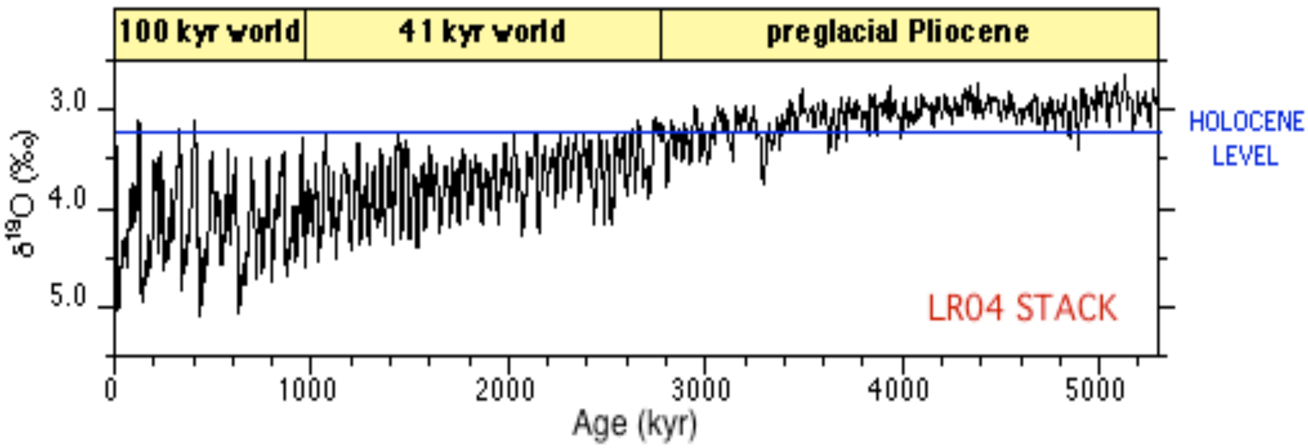

Figure. 4.9. Composite stack of 57 different benthic oxygen isotope records from (a proxy for temperature) from a globally distributed network of marine sediment cores. (data from Lisiecki and Raymo, 2005 and associated website)**.** This foraminifera $\delta^{18}O$ record indicates low-magnitude climate changes from 5.3 until about 2.7 Ma ago, when the amplitude of the foraminifera $\delta^{18}O$ signal increases markedly.

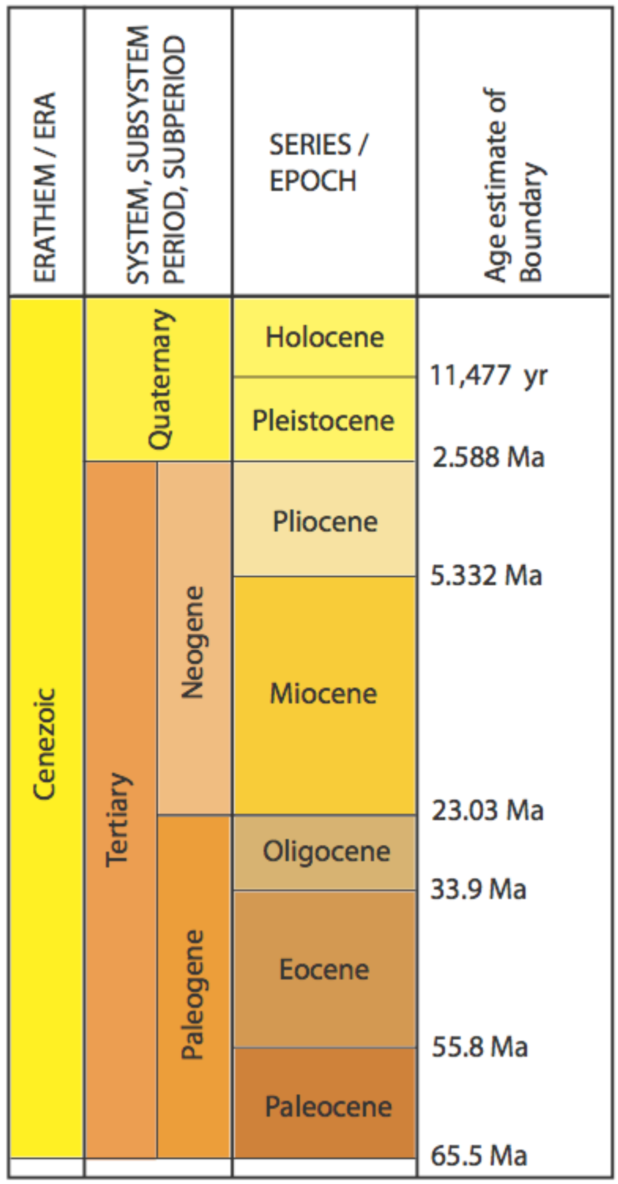

Figure 4.10. Chronology used for geological time periods in this report. (modified from Ogg and 2004)

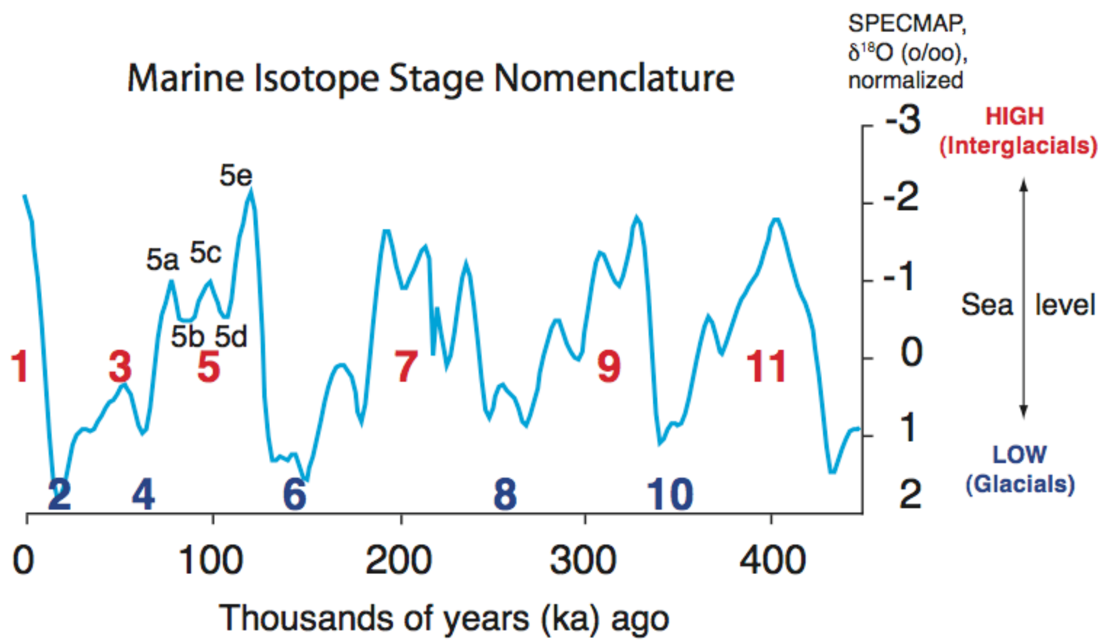

Figure 4.11. Marine Isotope Stage (MIS) nomenclature and chronology used in this report (after Imbrie et al. (1984) and Martinson et al. (1987)).

Chapter 4 References

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