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

Chapter 3 — Paleoclimate Concepts

Chapter Lead Authors

Richard B. Alley, Pennsylvania State University, University Park, PA

Joan Fitzpatrick, U.S. Geological Survey, Denver, CO

Contributing Authors

Julie Brigham-Grette, University of Massachusetts, Amherst , MA

Gifford H. Miller, University of Colorado, Boulder, CO

Daniel Muhs, U.S. Geological Survey, Denver, CO

Leonid Polyak, Ohio State University, Columbus, OH

ABSTRACT

Interpretation of paleoclimate records requires an understanding of Earth's climate system, the causes (forcings) of climate changes, and the processes that amplify (positive feedback) or damp (negative feedback) these changes. Paleoclimatologists reconstruct the history of climate from proxies, which are those characteristics of sedimentary deposits that preserve paleoclimate information. A great range of physical, chemical, isotopic, and biological characteristics of lake and ocean sediments, ice cores, cave formations, tree rings, 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 with others that have had their ages measured more accurately. Not all questions about the history of Earth's climate can be answered through paleoclimatology: in some cases the necessary sediments are not preserved, or the climatic variable of interest is not recorded in the sediments. Nonetheless, many questions can be answered from the available information.

An overview of the history of Arctic climate over the past 65 million years (m.y.) 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 at least the most recent of these cycles, shorter-lived large and rapid fluctuations occurred, especially around the North Atlantic Ocean. The last 11,000 years or so 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 itself, 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’s output and in Earth’s orbit, volcanic eruptions, and other factors have contributed to the natural climate changes since the end of the last ice age.

3.1 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 climate change is the focus of this assessment report. While most people accept that the weather is always changing on the time scale of recent memory, geologists reconstruct climate on longer time scales 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 Earth's history. This information helps scientists assess whether climate changes observable now may be part of a natural cycle or whether human activity may play a role. The relevance of climate science lies in the recognition that even small shifts in climate can and have had sweeping economic and societal effects (Lamb, 1997; Ladurie, 1971).

Indications of past climate, called climate proxies, are preserved in geological records; they tell us that Earth's climate has rarely been static. For example, during the past 70 million years ("m.y."), 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 drifting of continents and mountain-building in response to plate-tectonic forces that cause continental drift and mountain-building forces, 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 and developing scientific explanations of the observed changes (paleoclimatology) inform efforts to understand the climate, reveal features of importance that must be included in predictive models, and allow testing of the models developed.

An overview of key climate processes is provided here, followed by a summary of techniques for reconstructing past climatic conditions. Additional details pertaining to specific aspects of the Arctic climate system and its history are presented in the subsequent chapters.

3.2 Forcings, Feedback, and Variability

An observed change in climate may depend on more than one process. Tight linkages and interactions exist between these processes, as described below, but it is commonly useful to divide these processes into three categories: internal variability, forcings, and feedbacks. (For additional information, see Hansen et al., 1984, Peixoto and Oort, 1992; or IPCC, 2007 among other excellent sources.)

Internal variability is familiar to weather watchers: if you don't like the weather now, wait for tomorrow and something different may arrive. Even though the Sun's energy, Earth's orbit, the composition of the atmosphere, and many other important controls are the same as yesterday, different weather arrives 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 the 30-year averages typically used in defining the climate vary internally. For example, without any external cause, a given 30-year period may have one more El Niño event in the Pacific Ocean, and thus slightly warmer average temperatures, than the previous 30-year period.

Forced changes are caused by an event outside the climate system. If the Sun puts out more energy, Earth will warm in response. If fewer volcanoes than average erupt during a given century, then less sunlight than normal will be blocked by particles from those volcanoes, and Earth's surface will warm in response. If burning fossil fuel raises the carbon-dioxide concentration of the atmosphere, then more of the planet's outgoing radiation will be absorbed by that carbon dioxide, and Earth's surface will warm in response. Depending on often-random processes, different forcings may combine to cause large climate swings or offset to cause climate changes to be small.

When one aspect of climate changes, whether in response to some forcing or to internal variability, other parts of the climate system respond, 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 in response to the net magnitude of a set of forcings) 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 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 and the melting of more snow and ice. 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.

3.2.1 The Earth’s Heat Budget—A Balancing Act

On time scales of hundreds to thousands of years, 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 3.1)**.** This state of near-balance is maintained by the very strong negative feedback linked to thermal radiation. All bodies “glow” (send out radiation), and warmer bodies glow more brightly and send out more radiation than cooler ones. (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 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 (that is, a planet colder than the temperature at which it would be in equilibrium) will receive more energy than is radiated, causing the planet to warm, thus increasing radiation from the planet 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..

FIGURE 3.1 NEAR HERE

Equatorial regions receive more energy from space than they emit to space, polar regions emit 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 Nakamura and Oort, 1988, Peixoto and Oort, 1992, and Serreze et al., 2007).

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 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 important.

3.2.2 Solar Irradiance Forcing

3.2.2a 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 during Earth's 4.6-billionyear history, the Sun's energy output should have increased smoothly from about 70% of modern output (see, for example, Walter and Barry, 1991). (Direct paleoclimatic evidence of this increase in solar output is not available.) During the last 100 m.y., 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 time periods of millennia or less. For reference, the 0.000001% per century change in output from aging of the Sun can be compared with other changes, for example:

- maximum changes of slightly under 0.1% over 5 to 6 years as part of the sunspot cycle (Foukal et al., 2006);
- the estimated increase from the year 1750 to 2005 in solar output averaged across sunspot cycles, which also is slightly under 0.1% (Forster et al., 2007; see below); and
- the warming effect of carbon dioxide added to the atmosphere from 1750 to 2005. This addition is estimated to have had the same warming effect globally as an increase in solar output of ~0.7% (Forster et al., 2007), and thus it is much larger than changes in solar irradiance during this same time interval.

3.2.2b 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 within the Sun. Changes in solar output associated with this cycle cause peak solar output to exceed the minimum value by slightly less than 0.1% (Beer et al., 2006; Foukal et al., 2006; Camp and Tung, 2007). A satellite thus measures a change from maximum to minimum of about 0.9 W/m^2, out of an average of about 1365 W/m^2. This value is usually recalculated as a "radiative forcing" for the lower atmosphere. It is divided by 4 to account for spreading of the radiation around the spherical Earth and multiplied by about 0.7 to allow for the radiation that is directly reflected without warming the planet (Forster et al., 2007). The climate response to this sunspot cycling has been estimated as less than 0.1°C (Stevens and North, 1996) to almost 0.2°C (Camp and Tung, 2007). As discussed by Hegerl et al. (2007), the lack of any trend in solar output over longer times than this sunspot cycling, as measured by satellites, excludes the Sun as an important contributor to the strong warming during the interval of satellite observations, but the solar variability may have contributed weakly to temperature trends in the early part of the 20th century.

Over longer time frames, indirect proxies of solar activity (historical sunspot records, tree-rings and ice-cores) also exhibit 11-year solar cycles as well as longer-term variability. Common longer cycles are about 22, 88 and 205 years (e.g., Frohlich and Lean, 2004). The historical climate record suggests that periods of low solar activity may be linked to climate anomalies. For example, the solar minima known as the "Dalton Minimum" and the "Maunder Minimum" (1790–1820 AD, and 1645–1715 AD, respectively) correspond to the relatively cool conditions of the Little Ice Age, suggesting a role for changes in solar activity in the climate anomalies (along with other influences; see Chapter 4). However, the magnitude of radiative forcing that can be attributed to variations in solar irradiance remains debated (e.g., Baliunas and Jastrow, 1990; Bard et al., 2000; Fleitmann, et al., 2003; Frolich and Lean, 2004; Amman et al., 2007; Muscheler et al., 2007). An extensive summary of estimates of solar increase since the Maunder Minimum is given by Forster et al. (2007), which lists a preferred value of a radiative forcing of ~0.2 W/m^2, although the report also lists older estimates of just less than 0.8 W/m^2, still well below the estimated radiative forcing of the human-caused increase in atmospheric carbon dioxide (~1.7 W/m^2) (IPCC, 2007).

3.2.3 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 control ice-age cycles, result in systematic changes in the seasonal and geographic distribution of incoming solar radiation (insolation) for the planet (Milankovitch, 1920, 1941). The Milankovitch cycles have almost no effect on total sunshine reaching the planet over time spans of years or decades; they have only a small effect on total sunshine reaching the planet over tens of thousands of years and longer; but they have large effects on north-south and summer-winter distribution of sunshine. These "Milankovitch variations" (Figure 3.2) are due to three types of changes: (1) the eccentricity (out-of-roundness) of Earth's orbit around the Sun varies from nearly circular to more elliptical and back over about 100 thousand years (k.y.) (E in Figure 3.2); (2) the obliquity (how far the North Pole is tilted away from "straight up" out of the plane containing Earth's orbit about the Sun) tilts more and then less over about 41 k.y. (T in Figure 3.2); and (3) the precession (the wobble of Earth's rotational axis, moves Earth from its position closest to the Sun in the Northern-Hemisphere summer (the southern winter) to its position farthest from the Sun in the northern summer (the southern winter and back again in cycles of about 19–23 k.y. (P in Figure 3.2) (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 Earth itself and on the bulge at the equator caused by Earth's rotation. These features are relatively stable, and can be calculated for periods of millions of years with high accuracy. Paleoclimatic records show the influence of these changes very clearly (e.g., Imbrie et al., 1993).

FIGURE 3.2 NEAR HERE

The variations in eccentricity (orbital “out of roundness” or departure from circularity) 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 planet as a whole. 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 have exceeded 20% in 10,000 years (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 in Figure 3.2), where high tilt angles result in much more high-latitude insolation than do low tilt angles, and the precession or wobble of Earth’s rotational axis (P in Figure 3.2). 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 thousand years ago (ka), 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 midsummers today than at 11 ka (Figure 3.3). On the basis of this orbital consideration alone, Arctic summers should have been cooling during this interval in response to the Earth’s precession.

FIGURE 3.3 NEAR HERE

3.2.4 Greenhouse Gases in the Atmosphere

Roughly 70% of the incoming solar radiation is absorbed by the planet, warming the land, water, and air (Forster et al., 2007). Earth, in turn, radiates energy to balance what it receives, but at a longer wavelength than that of 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 Earth's outgoing longwave radiation band (Figure 3.1). Greenhouse gases play a key role in keeping the planetary temperature within the range conducive to life. In the absence of greenhouse gases in Earth's atmosphere, the planetary temperature would be about –19°C (–2°F); with them, the average temperature is about 33°C (about 57°F) higher (with constant albedo; 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. Concentrations of these gases are directly affected by anthropogenic (human) 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 a less important greenhouse gas than water vapor near Earth's surface. Changing the carbon-dioxide concentration of the atmosphere is relatively easy, but changing the atmospheric concentration of water vapor to any appreciable degree is difficult 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 Earth of about 2 cm/week (e.g., Peixoto and Oort, 1992); human perturbations to these fluxes are relatively very small (Forster et al., 2007). However, the large ocean surface and moisture from plants provide important water sources 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 (Hansen et al., 1984; Pierrehumbert et al., 2007). Hence, discussions of forcing of changes in climate focus especially on carbon dioxide, and to a lesser degree on methane and other greenhouse gases, rather than on water vapor (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 (see Prentice et al. (2001) in the IPCC 3rd Assessment Report for a comprehensive discussion). The possible effect of increasing CO_2 levels in the atmosphere was first recognized by Arrhenius (1896). By the 1930s, mathematical models linking greenhouse gases and climate change (Callendar, 1938) projected that a doubling of atmospheric CO_2 concentration would increase the mean global temperature by 2°C and would warm the poles considerably more. (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 Intergovernmental Panel on Climate Change report (IPCC, 2007). Information summarized there highlights the likelihood that changes in concentrations of greenhouse gases will especially affect the Arctic (Figure 3.4) and focuses attention on greenhouse gases as well as other influences on the Arctic, as discussed in this report especially in Chapter 4 (temperature and precipitation history).

FIGURE 3.4 NEAR HERE

3.2.5 Plate Tectonics

The drifting of continents (explained by the theory of plate tectonics) moves land masses from equator to pole or the reverse, opens and closes oceanic “gateways” between land masses thus redirecting ocean currents, raises mountain ranges that 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 the average global 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). Because the tectonic processes causing continental drift control the rate of volcanism, and can change over millions of years, changes in atmospheric carbon-dioxide concentration can be forced by the planet beneath.

3.2.6 Biological Processes

Biological processes can both absorb and release carbon dioxide, such that evolutionary changes have contributed to atmospheric changes. For example, some carbon dioxide taken from the air by plants is released by their roots into the soil, by respiration while living and by decay after death. Thus, plants speed the reaction of atmospheric carbon dioxide with rocks (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 burning (oxidation) 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 burning and altered 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 and release the stored carbon dioxide. (Humans are greatly accelerating these natural processes; fossil fuels that required hundreds of millions of years to accumulate are being burned in hundreds of years.) Rapid burial favors preservation of organic matter, whereas dead things left on the surface will decompose. Thus, 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 more or less as rapidly as fingernails grow, so that a major reshuffling of the continents requires about 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 that modified Earth's climate or were linked to continental drift or biological evolution have been highly influential over time spans of tens of millions of years, but they have had essentially no effect over shorter intervals of centuries or millennia. (Note that if one considers 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 in a few millennia or less the changes in volcanic release have not notably affected the carbon-dioxide concentration of the atmosphere. The main short-term effect of an increase in volcanic eruptions is to cool the planet by blocking the Sun, as discussed next.)

3.2.7 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. Particles are removed by gravity in days to weeks. Sulfur gases, in contrast, are converted rapidly to sulfate aerosols (tiny droplets of sulfuric acid) that have a residence time in the stratosphere of about 3 years and are transported around the world and poleward by circulation within the stratosphere. Tropical eruptions typically influence both hemispheres, whereas eruptions at middle to high latitudes usually affect only the hemisphere of eruption (Shindell et al., 2004; Fischer et al., 2007). Consequently, the Arctic is affected primarily by tropical and Northern Hemisphere eruptions.

The radiative and chemical effects of the global volcanic aerosol cloud produce strong responses in the climate system on short time scales (see Figure 5.5) (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. A tropical eruption produces more heating in the tropics than in the high latitudes and thus a steeper temperature gradient between the pole and the equator, 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 and warms winter temperatures. Because little solar energy reaches the Arctic during winter months, the transfer of warm air from tropical sources to high latitudes has more effect 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).

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 (Thordarson et al., 2001; Oman et al., 2005, 2006).

When widespread stratospheric volcanic aerosols settle out, some of the sulfate falls onto the Antarctic and Greenland Ice Sheets (Figure 3.5). Measurements of those sulfates present in ice cores can be used to estimate the Sun-blocking effect of the eruption. Large volcanic eruptions, especially those within a few decades of each other, are thought to have promoted cooling during the Little Ice Age (about1280–1850 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).

FIGURE 3.5 NEAR HERE

The effects of volcanic eruptions are clearly evident in ice-core records (e.g., Zielinski et al., 1994); major eruptions cooled Greenland about 1°C for about 1 or 2 years as recorded in Greenland ice cores (e.g., Stuiver et al., 1995) (Figure 3.6). Tree-ring records also support the connection between climate and volcanic eruptions (LaMarche and Hirschbeck, 1984; Briffa et al., 1998; D’Arrigo et al., 1999; Salzer and Hughes, 2007). The growth and shrinkage of the great ice-age ice sheets, and the associated loading and unloading of 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 in recent millennia appear to have introduced unavoidable climatic “noise” as opposed to controlling the climate in an organized way.

FIGURE 3.6 NEAR HERE

3.2.8 Other influences

Paleoclimatic records discount some speculative mechanisms of climate change. For example, about 40,000 years ago natural fluctuations reduced the strength of Earth's magnetic field 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 record does not change in parallel with changes in beryllium-10, indicating that the cosmic-ray increase had little or no effect on climate (Muscheler et al., 2005). Large changes in concentration of extraterrestrial dust between Earth and Sun might lead to changes in solar energy reaching Earth and thus to changes in climate; however, the available sedimentary records show no significant changes in the rate of infall of such extraterrestrial dust (Winckler and Fischer, 2006).

The climate is a complex, integrated system, and it operates through strong linked feedbacks, internal variability, and numerous forcings. On time scales of centuries or less, however, many of the drivers of past climate change—such as 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 have been caused by small variations in the Sun's output, occasional short-lived cooling caused by explosive volcanic eruptions, and greenhouse-gas changes have affected the planet's temperature.

3.3 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 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: they read the indirect evidence that the past sends by proxy. All 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 the 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 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, human accounts are lacking for almost all of Earth's history. 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 even more than the written records.

3.3.1 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 shaped 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 do the following: (1) 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); (2) be interpretable in terms of climate (the characteristics of the deposit must uniquely relate to the climate at the time of formation); and (3) be "datable" (i.e., there must be some way to determine the time when the sediment was deposited). Here, we first present one well-known paleoclimatic indicator as an example, then discuss general issues raised by that example, and follow with a discussion of many types of paleoclimatic indicators.

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) foraminifers, microscopic organisms that live on the sea floor and secrete 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 with time or very steeply with space in the deep ocean) and changes in global ice volume. Global ice volume determines the relative abundances of the isotopes oxygen-16 and oxygen-18 in seawater. Snow has relatively less of the heavy oxygen-18 than 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 foraminifer shells. The proportion of the heavy and light isotopes of oxygen is usually expressed as $\delta^{18}O$; positive $\delta^{18}O$ values represent extra amounts of the heavy isotope of oxygen, and negative values represent samples with less of the heavy isotope than average seawater. Positive $\delta^{18}O$ reflects glacial times (colder, more ice), whereas more negative $\delta^{18}O$ reflects interglacial (warmer, less ice) times in Earth's history. Although the $\delta^{18}O$ of foraminifer shells does not reveal where the glacial ice 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 by use of other indicators. In the absence of changes in global ice volume, changes in **benthic foraminifer** $\delta^{18}O$ reflect changes in ocean temperatures: more positive $\delta^{18}O$ values indicate colder water, and more negative $\delta^{18}O$ values indicate warmer water.

Written documents have sometimes been erased and rewritten, in a deliberate attempt to distort history or 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 vigilance proves to be important. For example, water may remove isotopes carrying paleoclimatic information from shells and replace 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 possible to tell whether a record has been altered, and this problem should not affect any of the conclusions presented in this report.

Finding the link 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 the sand can blow around. Today in northwestern Nebraska, the huge dune field of the Sand Hills is covered in grass (Figure 3.7). The dunes formed during drier conditions in the past, but wetter conditions now allow grass to grow on top (e.g., Muhs et al., 1997). Similarly, the sediments left by glaciers are readily identified, and those sediments in areas that are ice free today attest to changing climate. A very different physical indicator of past climate is the 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 at depth reveal how cold the ice age was (Cuffey and Clow, 1997).

FIGURE 3.7 NEAR HERE

Many paleoclimate records are based directly on living things. Tundra plants are quite different from those living in temperate forests. If pollen, seeds, and twigs found in deep layers of a sediment core came from tundra plants, and those found in shallow layers came from temperate-forest plants, a formerly cold time that has warmed is indicated. Trees grow more rapidly and add thicker rings when climatic conditions are more favorable. In very dry regions, this feature allows trees to be used in reconstruction of rainfall; in cold regions, growth may be more closely linked to temperature (Fritts, 1976; Cook and Kairiukstis, 1990)

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: they use more-flexible molecules to offset the increase in brittleness caused by colder temperatures. These molecules are sturdy and persist in sediments after the organism dies, 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 exist.)

Isotopic ratios are among the most commonly used 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 their carbon isotopes analyzed, 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 other relationships may be complex. The width of a tree ring, for example, is especially sensitive to water availability in dry regions, but it may also be influenced by changes in shade from neighboring trees, an attack of beetles or other pests that weaken a tree, the temperature of the growing season, 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). Conclusions about climate typically rest on many lines of evidence.

3.3.2 The Age of the Sediments

History requires "when" as well as "what." Many techniques reveal the "when" of sediments, sometimes to the nearest year. In general, more-recent events can be dated more precisely.

Climate records that have been 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 the layering clearly is annual (Alley et al., 1997), allowing quite accurate counts. The longest-lived trees may be 5000 years old; 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 Earth's surface (those rays produce beryllium-10 and other isotopes), 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 (very little of the original radiocarbon survives in older samples, causing measurements difficulties and allowing even trace contamination by younger materials to cause large errors in estimated age, so other techniques are preferred). Many other isotopes are used for various materials and time intervals, extending back to the formation of Earth. 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 event is said to be 100,000 years old, the event can be said with high confidence to have occurred sometime between 99,000 years and 101,000 years ago.)

3.4 Cenozoic Global History of Climate

As emphasized in the Summary for Policymakers of IPCC (2007) and in the body of that report, a paleoclimatic perspective is important for understanding Earth's climate system and its forcings and feedbacks. Arctic records, and especially Arctic ice-core records, have provided key insights. The discussion that follows briefly discusses selected features in the history of Earth's climate and the forcings and feedbacks of those climate events. This discussion does not treat all of the extensive literature on these topics, but it 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 to this topic.)

This report focuses on the Cenozoic Era, which began about 65 Ma with the demise of the dinosaurs and continues today (see section 4.5 for a discussion of the chronology used in this report). During most of this 65 m.y. interval, deep-sea records of foraminifer $\delta^{18}O$ (a powerful paleoclimatic indicator, described above in section 4.4.1), which integrate the sedimentary record in several ocean basins, show that Earth was warmer than at present and supported a smaller volume of ice (Figure 3.8). Yet, following the peak warming of the early Eocene, about 50–55 Ma, global temperatures generally declined (Miller et al., 2005). Although this record is not specific about Arctic climate change, the record indicates that the global gradient (or difference) in temperature between 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). Changes in the gradient cause changes in atmospheric and oceanic circulation. The overall cooling trend of the past 55 m.y. was punctuated by intervals during which the cooling was reversed and the oceans warmed, only to cool rapidly again at a later time. Examples of such accelerated cooling include rapid decreases in foraminifer $\delta^{18}O$ about 34 Ma and again about 23 Ma, which are thought to reflect the rapid buildup of ice in Antarctica in only a few hundred thousand years (Zachos et al., 2001). The Paleocene-Eocene thermal maximum (about 55 Ma) represents a major interval of global warming when CO_2 levels are estimated to have risen abruptly (Shellito et al., 2003, Higgins and Schrag, 2006), perhaps owing to the rapid release of methane from sea-floor sediments (Bralower et al., 1995).

FIGURE 3.8 NEAR HERE

The style and tempo of global climate change during the past 5.3 m.y. is depicted well by the foraminifer $\delta^{18}O$ record of Lisiecki and Raymo (2005) (Figure 3.9**;** 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 foraminifer $\delta^{18}O$ record reflects changes in both global ice volume and ocean bottom-water temperature change, and with the same sense—An increase in global ice or a decrease in ocean temperatures pushes the indicator in the same direction. The foraminifer $\delta^{18}O$ record indicates low-magnitude climate changes from 5.3 until about 2.7 Ma, when the amplitude of the foraminifer $\delta^{18}O$ signal increased markedly. This shift in foraminifer $\delta^{18}O$ amplitude coincides with widespread indications of onset of northern continental glaciation (see Chapter 4, temperature and precipitation history). The oxygen isotope fluctuations since 2.7 Ma 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 within the past 3 m.y., 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 suggest that global temperatures were warmer by as much as 2°C or 3°C at that time (see Jiang et al., 2005; IPCC, 2007).

FIGURE 3.9 NEAR HERE

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

Moving toward the present, the number of available records increases greatly, as does typical time resolution of the records and the accuracy of dating (see section 4.4). The large ice-age cycling of the last 0.9 m.y. 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 and submerged coastal land, but with a relatively small net effect (e.g., Kump and Alley, 1994). The ice-age cycling caused large temperature changes, of many degrees to tens of degrees in some places (see Chapter 4, temperature and precipitation history).

Climate changed in large abrupt jumps (see section 5.4.3) during 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 (as recorded by many climate-change indicators) in time spans of decades to years. Changes away from the North Atlantic were notably smaller, and in the far south the changes appear to see-saw (southern warming with northern cooling). The “shape” of the climate records is interesting: northern records typically show abrupt warming, gradual cooling, abrupt cooling, near-stability or slight gradual warming, and then they repeat (see Figure 6.9).

The most recent interglacial interval has lasted slightly more than 10,000 years. Generally warm conditions have prevailed compared with the average of the last 0.9 m.y. However, important changes have been observed. These changes include broad warming and then cooling in only 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 4 (temperature and precipitation history) and 5 (rates of Arctic climate change).

3.5 Chronology

In any discussion of past climate periods, we must use a time scale understandable to all readers. Beyond the historical period, then, we must use 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 Earth and one for much more recent Earth history, approximately the past 2.6 m.y. (the Quaternary Period). For the longer period of Earth history, we use the terminology and time scale adopted by the International Commission on Stratigraphy (Ogg, 2004). This time scale is well established and has been widely accepted throughout the geologic community. The Quaternary Period is the youngest geologic period in this time scale, and constitutes the past approximately 2.6 m.y. (http://www.stratigraphy.org/gssp.htm; Jansen et al., 2007) (Figure 3.10). The Quaternary Period is of particular interest in this report, because this time interval is characterized by dramatic changes—between glacial and interglacial—in climate.

FIGURE 3.10 NEAR HERE

Some problems are associated with the use of time scales within the Quaternary Period.

These problems are common to all geologic dating, but they 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. That is, a particular geologic event is recorded earlier at one geographic location and later at another.

A good example of time-transgression is the most recent deglaciation of mid-continent North America, the retreat of the *Laurentide Ice Sheet*. Although this retreat marked a major shift in a climate state, from a glacial period to an interglacial period, by its very nature it 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 about 22.6 ka, but it was still present in what is now northern Illinois until after about 15.1 ka, and was still in Wisconsin and Michigan until after about 12.9 ka (Johnson et al., 1997) (radiocarbon ages were converted using the algorithm of Fairbanks et al., 2005), and in north-central Labrador until about 6 ka (Dyke and Prest, 1987). 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 (biomes) as a result of climate change is not an instantaneous process throughout a wide geographic region. Thus, many records of climate change that reflect changes in plant communities will take place at different times in a region as taxa within that community migrate.

Another difficulty is not with the geologic records themselves but with the terms used in different regions to describe them. 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, used by 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 report, we have chosen to use the stages of the oxygen isotope record from foraminifers in deep-sea cores as our terminology for discussing different intervals of time within the Quaternary Period. The identification of glacial-interglacial changes in deep-sea cores, and the naming of stages for them, began with a landmark report by Emiliani (1955). The oxygen isotope composition of carbonate in foraminifer skeletons in the ocean shifts as climate shifts from glacial to interglacial states (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 changes 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 variety of geologic records of climate change show the same glacial-interglacial cycles that can be compared and correlated with the deep-sea record. These geologic records include glacial records (e.g., Booth et al., 2004; Andrews and Dyke, 2007), 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 foraminifers, other records of climate change such as pollen from 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 time scales that have been developed for the oxygen isotope record are important to understand. The mostly widely used time scales 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. The timing closely matched the known periodicities in 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. Several highly similar time scales have been developed using this approach. The most commonly cited are the SPECMAP studies of Imbrie et al. (1984) and Martinson et al. (1987) (Figure 3.11), and the more recent work of Lisiecki and Raymo (2005).

FIGURE 3.11 NEAR HERE

However, there are disadvantages to using the astronomically tuned oxygen isotope records. Very few deep-sea cores are dated directly, except in the upper parts that 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 time scales in the Quaternary Period. Challenges to this assumption are 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. Additional assumptions, including that response is proportional to forcing, are inherent in tuning.

Recognizing the assumptions inherent in the SPECMAP time scale, we use this time scale and the marine oxygen isotope stage terminology in this report for four reasons:

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 compared with "Eemian," in Europe. 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.

3.6 Synopsis

Earth's climate is a complex, interrelated system of air, water, ice, land surface, and living things responding to the Sun's energy. Scientific understanding of this system has been increasing rapidly, and the broad outline is now quite well known, although many details remain obscure and further discoveries are guaranteed.

The climate system can be forced to change, but it also varies internally without external forcing. Both forced and unforced variations interact with various feedback processes that may either amplify or reduce the resulting climate change, often with interesting patterns in space and time.

Changes in the energy emitted by the Sun, the amount of that energy reaching Earth, the amount of that energy reflected by Earth, and the greenhouse effect of the atmosphere are important in controlling global climate. Changes in continental positions, ocean currents, wind patterns, clouds, vegetation, ice, and more affect regional climates as well as contribute to the global picture. The Sun has brightened slowly for billions of years, and its brightness shows very small fluctuations measured in years to centuries. Features of Earth's orbit change the latitudinal and seasonal distribution of sunshine, and they have a small effect on total sunshine reaching the planet over tens of thousands of years. Great tectonic forces in the Earth rearrange continents and promote or reduce volcanic activity and growth of mountain ranges. All three affect greenhouse-gas concentrations and other features of the climate over millions of years or longer, and they interact with changes in the biosphere in response to biological evolution. And, these general statements omit many interesting and increasingly well-understood features of the system.

Many deposits of the Earth system—muds and cave formations and tree rings and ice layers and many more—have characteristics that reflect the climate at the time of formation, that are preserved after formation, and that reveal their age of formation. Careful consideration of these deposits underlies paleoclimatology, the study of past climates. Varied investigative techniques focus on physical, chemical, isotopic, and biological indicators, and they provide surprisingly complete histories of changes in time and space.

This report especially focuses on the last tens of millions of years. This interval has been characterized by slow cooling, leading from a largely ice-free world to ice-age cycling in response to orbital changes. Both the cooling trend and the ice-age cycling were punctuated occasionally by abrupt shifts. The last approximately 10,000 years have been a reduced-ice interglacial during the ice-age cycling, but they have experienced a variety of climate changes linked to changing volcanism, ocean currents, solar output, and—recently evident—human perturbation.

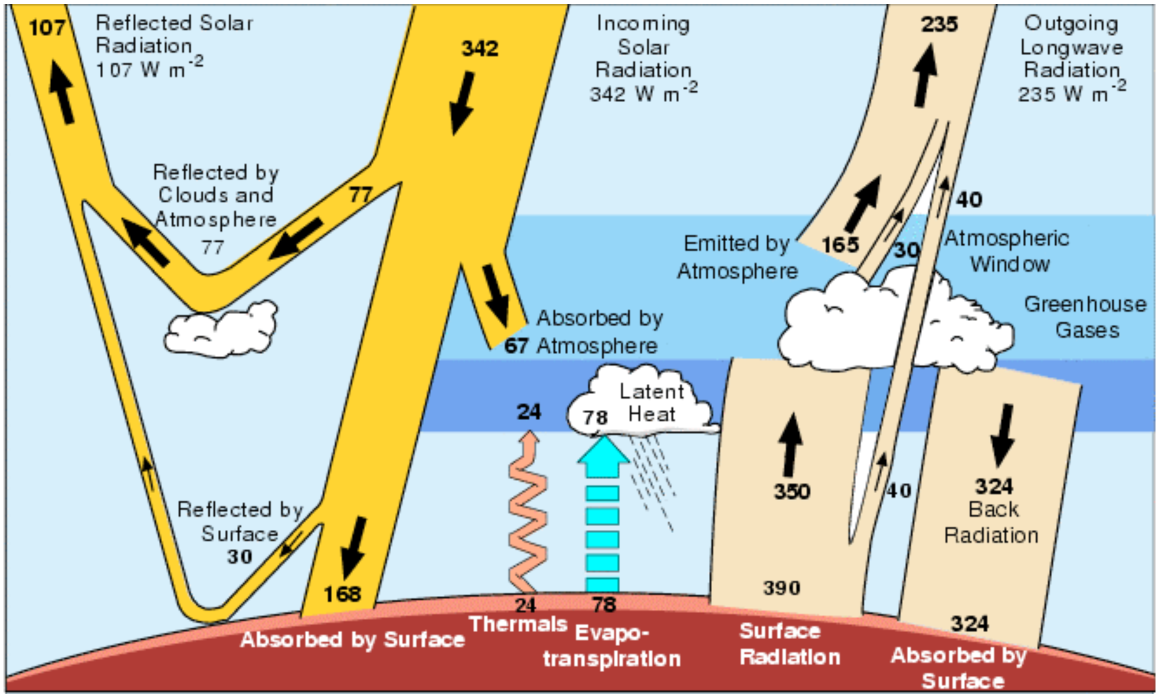

Figure 3.1 Earth's energy budget is a balance between incoming and outgoing radiation. [Numbers are in watts per square meter of the Earth's surface, and some estimates may be uncertain by as much as 20%.] Incoming shortwave radiation from the Sun entering Earth's atmosphere [342 W/m^2] may be reflected 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/m^2) of reflected sunlight plus 235 W/m^2 of infrared [long-wave] radiation) equal to the 342 W/m^2 of incident sunlight (Kiehl and Trenberth, 1997).

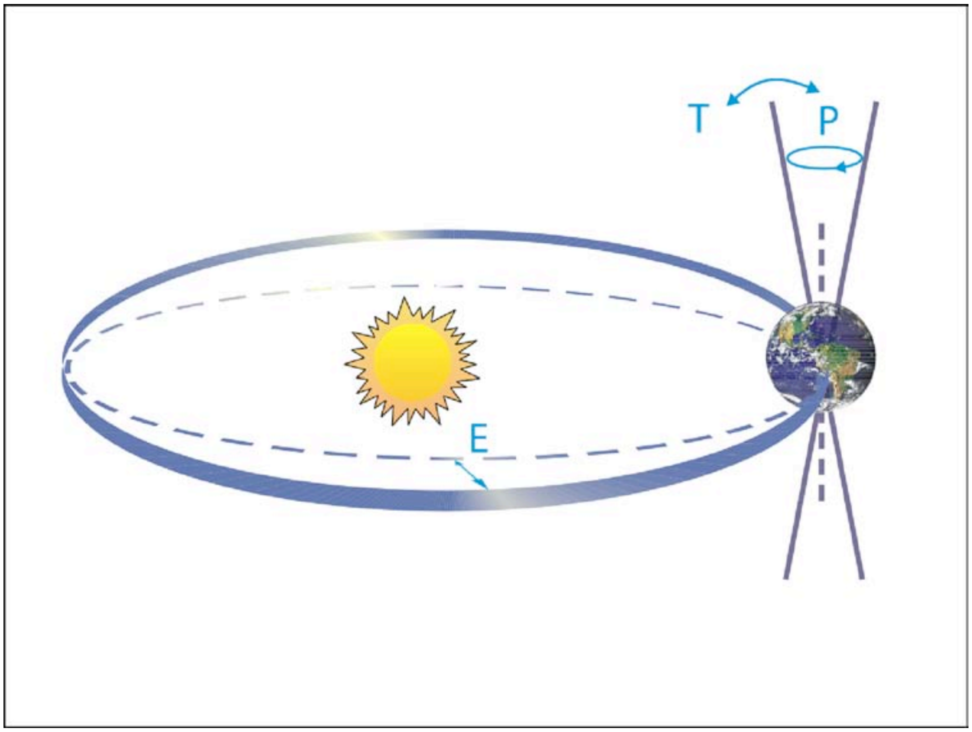

Figure 3.2 Earth's orbital variations (Milankovitch cycles) control the amount of sunlight received (insolation) at a given place on Earth's surface (Rahmstorf and Schellnhuber, 2006; Jansen et al., 2007). E, variation in the eccentricity of the orbit (owing to variations in the minor axis of the ellipse) with an approximate 100 k.y. periodicity; P, precession, changes in the direction of the axis tilt at a given point of the orbit, which has an approximate 19 to 23 k.y. periodicity; T, changes in the tilt (obliquity) of Earth's axis, which has and approximate 41 k.y. periodicity.

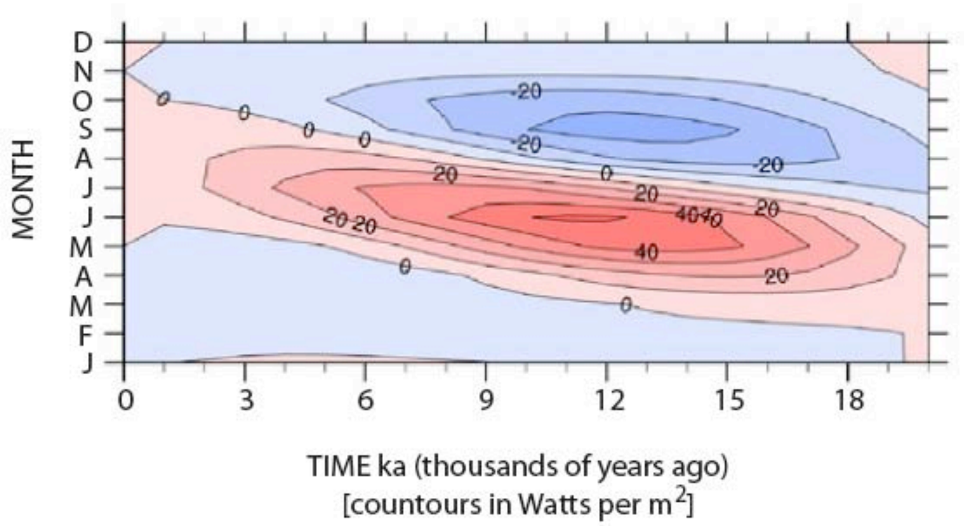

Figure 3.3. Milankovitch-driven monthly insolation anomalies (deviations from present), 20–0 ka at 60°N. Y axis, calendar months. Contours and numbers depict a history of insolation values. Contours in watts per square meter (W/m^2) (data from Berger and Loutre, 1992). Midsummer **i**nsolation values at 11 ka exceeded 40 W/m^2, whereas current values are less than 10 W/m^2.

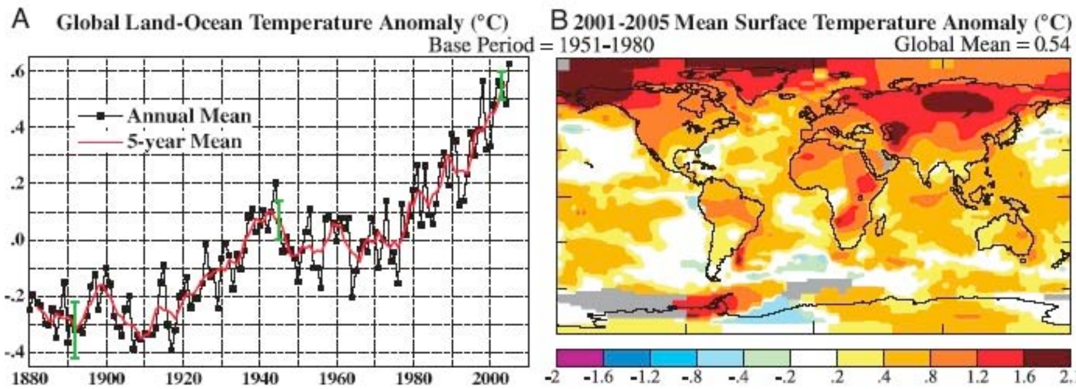

Figure 3.4 Mean surface temperature anomalies for Earth relative to 1951–1980. Panel A, the global average. Panel B, temperature anomalies 2000–2005. High northern latitudes show the largest anomalies for this time period (Hansen et al., 2006).

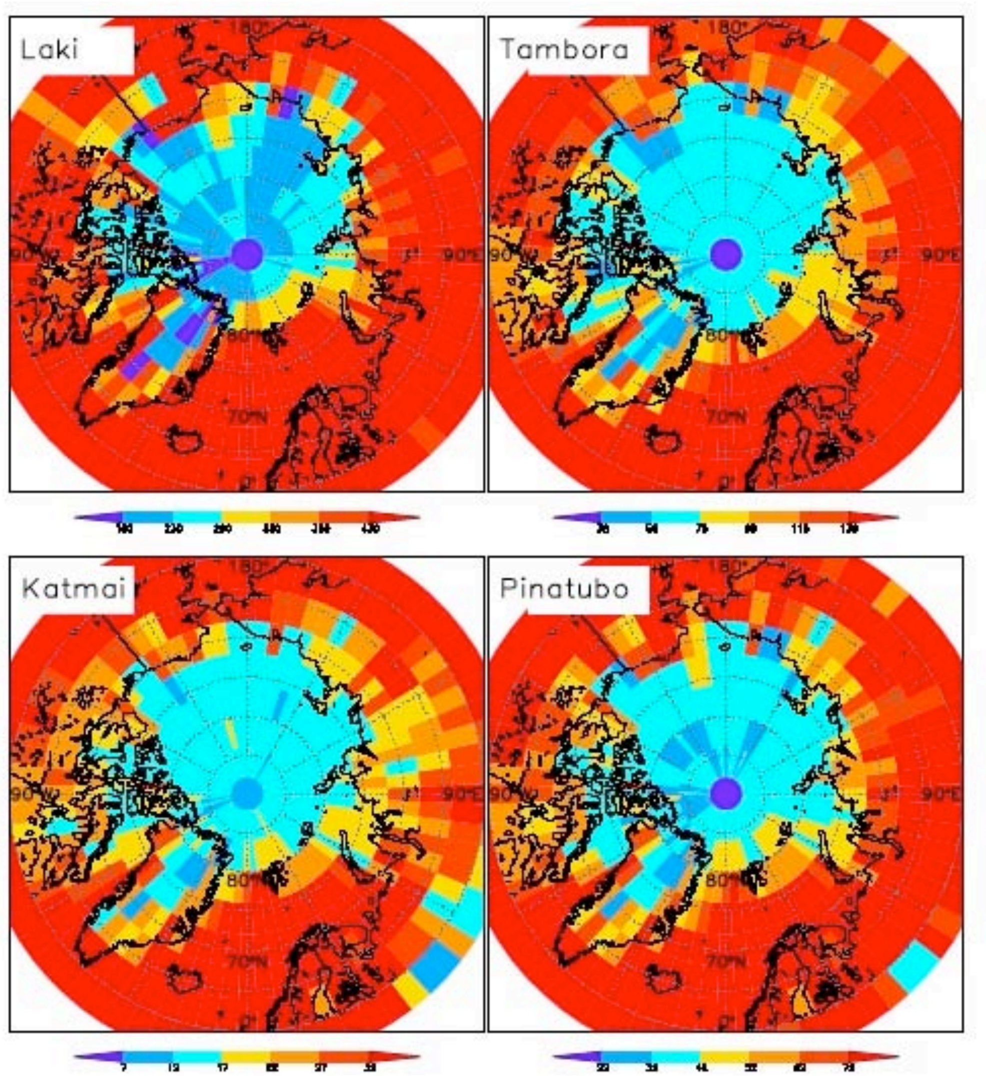

Figure 3.5 Simulated spatial distribution of volcanic sulfate aerosols (kg/km^2) produced by the Laki (1783), Katmai (1912), Tambora (1815), and Pinatubo (1991) eruptions in the Arctic (region shown, 66°–82°N. and 50°–35°W.). Blue, smaller than average deposits; yellow, orange, and red, increasingly larger than average deposits (from Gao et al., 2007). Volcanic evidence derived from 44 ice cores; analysis used the NASA Goddard Institute for Space Studies (GISS) ModelE climate model.

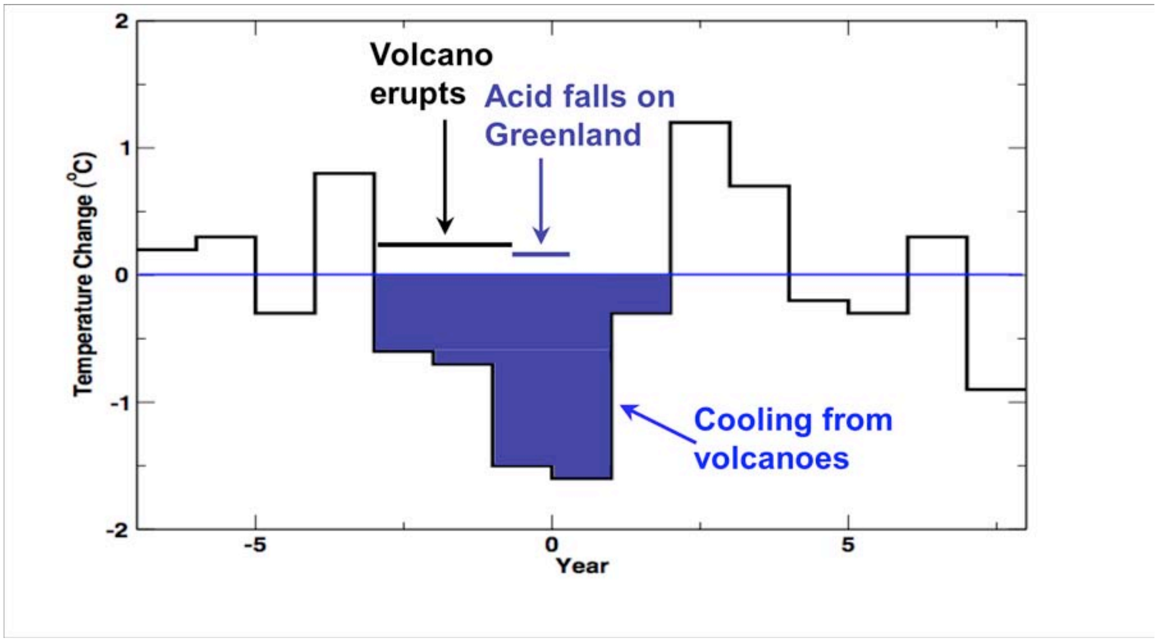

Figure 3.6 Temperature response (derived from stable isotopes) in Greenland snow to large volcanic eruptions reconstructed from the GISP2 ice core. (modified from Stuiver et al., 1995).

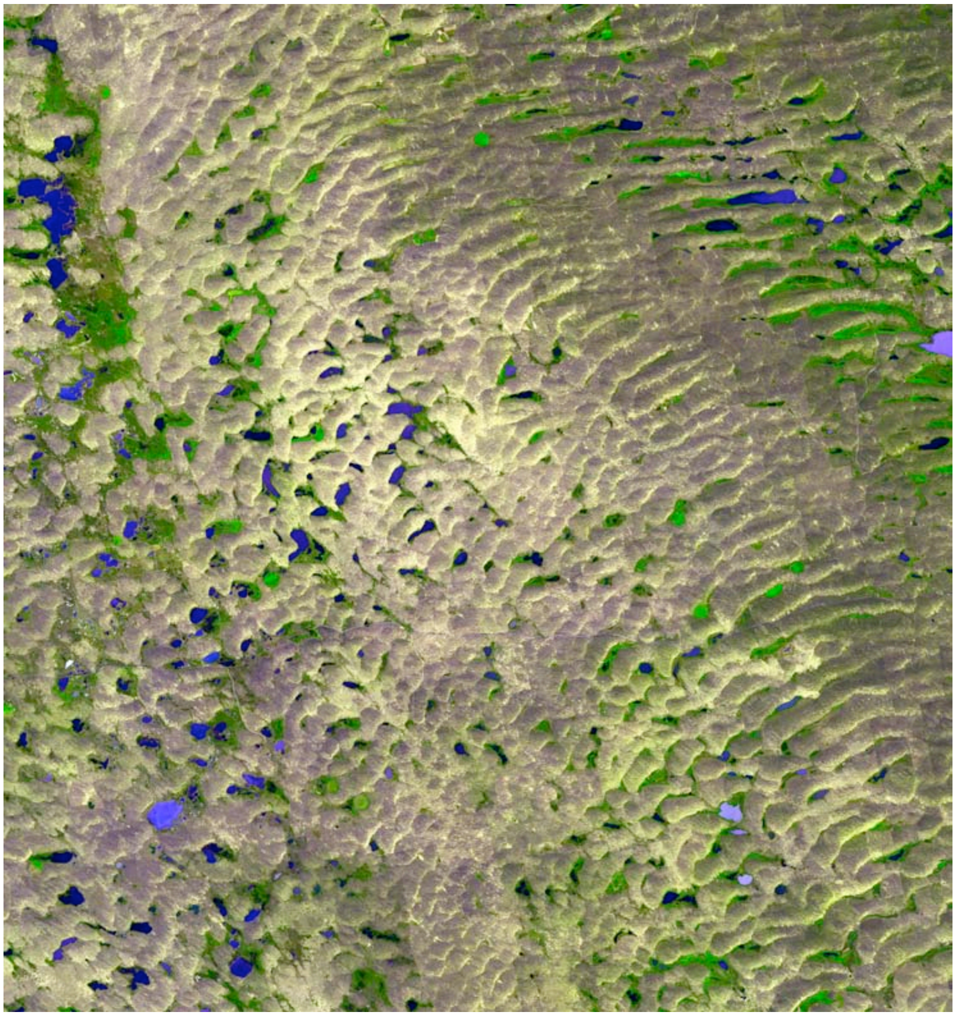

Figure 3.7 The Sand Hills of western Nebraska. The Sand Hills cover 51,400 km^2 (about a quarter of the state) and are the largest sand-dune deposit in the United States. They derive from Pleistocene glacial outwash eroded from the Rocky Mountains and now stabilized by vegetation. The hills are characterized by crowded crescent-shaped (barchan) dunes, general absence of drainage, and numerous tiny lakes filling the closed depressions between dunes. (Photo credit: NASA/GSFC/METI/ERSDAC/JAROS, and U.S./Japan ASTER Science Team. 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° N. and 102.2° W.)

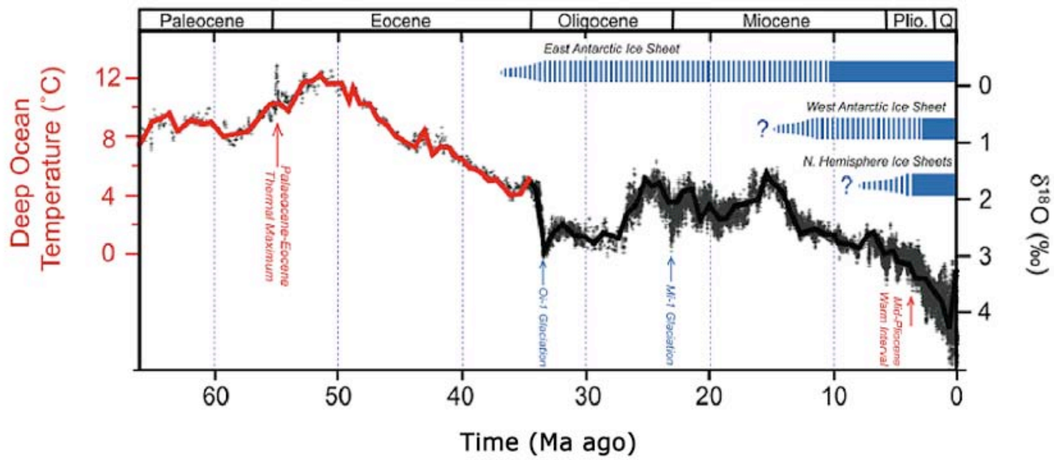

Figure 3.8. Global compilation of more than 40 deep sea benthic $\delta^{18}O$ isotopic records taken from Zachos et al. (2001), updated with high-resolution Eocene through Miocene records from Billups et al. (2002), Bohaty and Zachos (2003), and Lear et al. (2004). Dashed blue bars, times when glaciers came and went or were smaller than now; solid blue bars, ice sheets of modern size or larger. (Figure and text modified from IPCC Chapter 6, Paleoclimate, Jansen et al., 2007.)

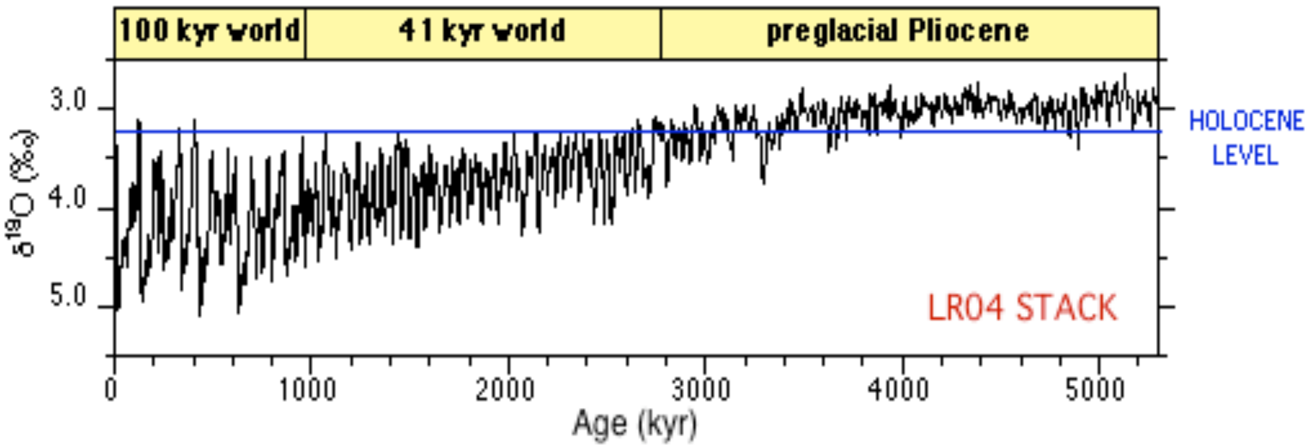

Figure 3.9. Composite stack of 57 benthic oxygen isotope records (a proxy for temperature) from a globally distributed network of marine sediment cores**.** This foraminifer $\delta^{18}O$ record indicates low-magnitude climate changes from about 5.3–2.7 Ma, when the amplitude of the foraminifer $\delta^{18}O$ signal increased markedly (data from Lisiecki and Raymo (2005) and associated website)

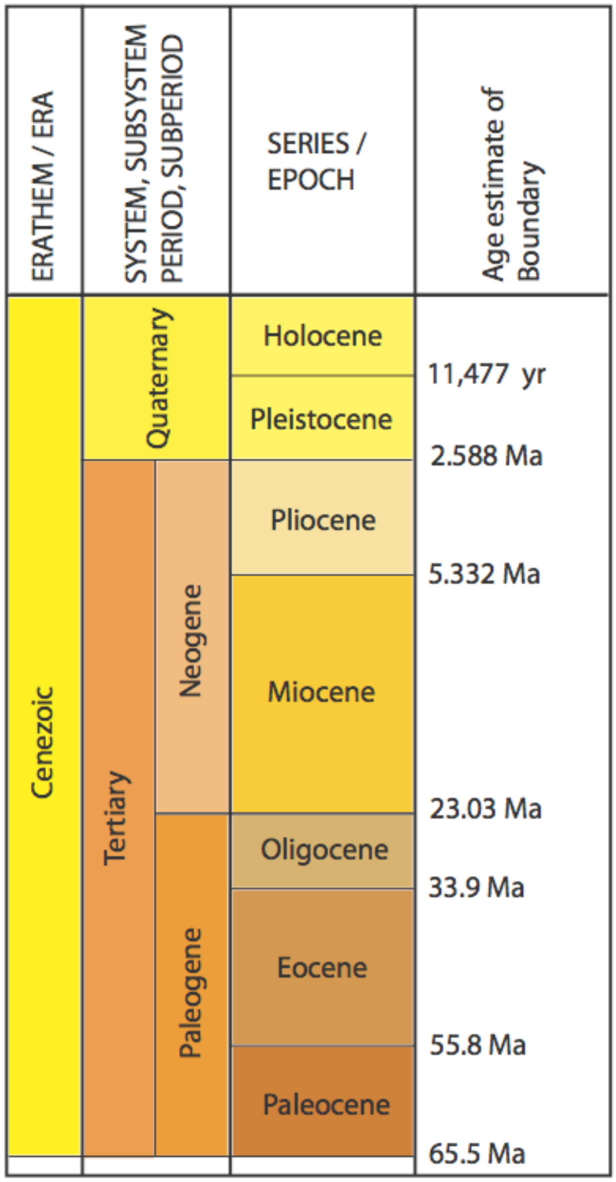

Figure 3.10. Cenozoic time periods as used in this report (modified from Ogg and 2004)

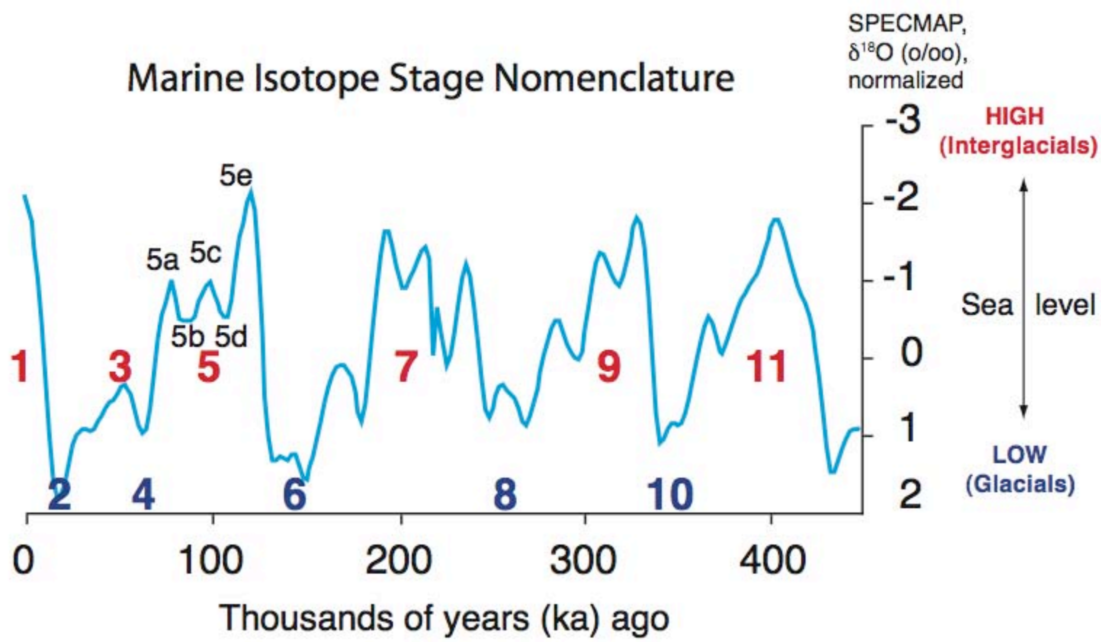

Figure 3.11. Marine isotope stage (MIS) nomenclature and chronology used in this report (after Imbrie et al., 1984; Martinson et al., 1987). Red numbers, interglacial intervals; blue numbers, glacial intervals.

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