CHAPTER

9

History and Evolution of Earth's Climate

9.1 PAST IS PROLOGUE[1]

The seasons come and go in an orderly cycle, and plants, animals, and humans adapt to this regular rhythm. Though anomalies and extreme weather events occur, it is natural to think of climate as a constant influence to which life has adapted. In the grossest sense, this view of climate is correct. Life has existed on Earth for at least 3.5 billion years, and the climate has been hospitable enough over that great span of time for life to continue. If we look in more detail at the climates of the past by sifting through the evidence recorded by nature, we find that climate is not so invariable and passive as it may seem. Past variations in climate are especially interesting for the climatologist, since they provide clues to the inner workings of the climate system that are difficult to infer in any other way. If past variations in climate can be understood thoroughly, then our chances of anticipating how climate will evolve in the future are greatly increased.

The history of Earth's climate is long and varied, and humans did not experience most of this history, since we appeared only at the last instant of geologic time. Human ingenuity led to the definition and measurement of temperature, pressure, and other climatic variables only during the last several centuries. Climate variables that may be directly measured using modern instruments constitute what we call the instrumental record. In addition to the instrumental record, we have historical data that stretch back farther in time, but are less quantitative. Such records include grape harvests in France, wheat harvests in ancient Egypt, and many bits of climate information contained in written accounts. A large amount of the

[1]"We all were sea-swallowed, though some cast again,
And, by that destiny, to perform an act whereof what's past is prologue, what to come,
In yours and my discharge." Antonio in William Shakespeare's *The Tempest*, Act 2, Scene 1.

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latter type of information, going back several thousand years, is contained in the ancient Chinese literature. Most of what we know of the deeper history of Earth's climate, before the invention of writing, has come from deducing climate variations from information left behind by various natural recording systems. These recording systems include physical, biological, and chemical information contained in geological information, lake and ocean sediments, terrestrial sediments, ice sheets, and tree rings. Information of this type, which we may call paleoclimatic data, can be used to derive time series of climatic information for many thousands of years into the past.

9.2 THE INSTRUMENTAL RECORD

Much of the early development and use of the thermometer took place in Florence in the mid-seventeenth century; however, the usefulness of early measurements was limited by the lack of a standard scale and calibration. In 1742, Anders Celsius invented the *Celsius temperature scale* (called the *centigrade scale* until 1948), but regular measurements of air temperature came into fashion considerably after its acceptance. Torricelli invented the barometer in 1644. It came more quickly into widespread use, because of the greater ease of its calibration and the perceived relationship between pressure and weather changes, which gives it a predictive capability. For a while, possessing a barometer was a status symbol. The first known reference to rain gauges is Indian literature from the fourth century BCE.

Temperature records as long as 150–200 years are available for only a few locations. Manley (1974) constructed a temperature record for central England going back to 1659. Other temperature time series include those for: Berlin, Germany beginning in 1700; de Bilt, Netherlands in 1706; Germantown, Pennsylvania in 1731; Milan, Italy in 1740; and Stockholm, Sweden in 1756. Sufficient measurements to define the hemispheric or global mean surface air temperature are available for only about the last century. Most of the variance of temperature is associated with seasonal and latitudinal variations, or weather. Any climatic trends are a small signal among much larger magnitude variations, particularly within the rather short period for which the instrumental record is available.

Because of concern over human-induced climate change, considerable effort has been devoted to developing estimates of global surface temperature based on the instrumental record and evaluating temporal trends in those estimates over the past century (Hansen et al., 2010; Morice et al., 2012). To obtain consistent records it is necessary to account for changes in instruments and their locations and surroundings. The land-based record has to be corrected for urbanization. Over the oceans, sea surface temperature measurements changed over time from buckets, to ship intake manifolds to drifting

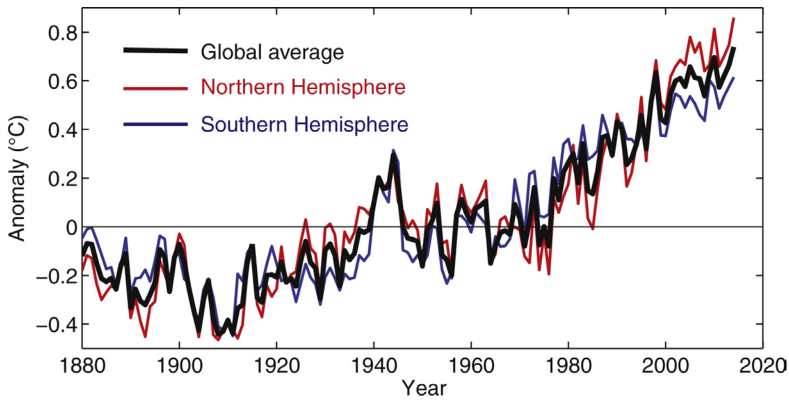

FIGURE 9.1 **Time series of annual mean temperature anomalies from 1880 to 2014 for the Global, Northern Hemisphere and Southern Hemisphere means.** *Data are from the NOAA global surface temperature anomaly data set (reference period 1901–2000).*

buoys, and each of these has distinct biases that need to be accounted for. The instrumental temperature record of global mean land and ocean surface temperature shows generally increasing temperatures since the 1880s, with a warming of about 0.3°C culminating in about 1940 (Fig. 9.1). After 1940, global mean surface temperature appears to have remained roughly constant until about 1970 when it began an increase of nearly 0.8°C until the present. After 1998, the temperature anomalies in the two hemispheres appear to diverge, with warming continuing in the Northern Hemisphere, but slowing in the Southern Hemisphere. It is believed that the change in the warming rate after about 1998 is related to the shifts in the Pacific Decadal Oscillation (PDO) and the Atlantic Multidecadal Oscillation (AMO) shown in Figs 8.16 and 8.17 that occurred at about that time after the large 1998 El Niño event. The change in the rate of warming in about 1998 could be attributed to changes in the natural modes of variability of the coupled ocean–atmosphere system, or could be part of the global-warming trend.

The spatial structure of warming can also be estimated from instrumental records, but it is more uncertain because some regions of the globe were very sparsely measured until the advent of global satellite measurements in about 1979, and profiling drifting buoys in about 2000. Figure 9.2 shows maps of the standard deviation of annual mean temperature and its trend over the periods 1880–2014 and 1979–2014. Not every feature in Fig. 9.2 is necessarily real, because of changes in spatial sampling and instrumentation over time. Nonetheless, some features of the standard deviation and trend seem robust. The standard deviation of annual mean temperature is greatest over the high latitude land areas, and less over the oceans. The variation over land is contributed by both the summer and winter seasons, but a bit more by the winter season when the variations

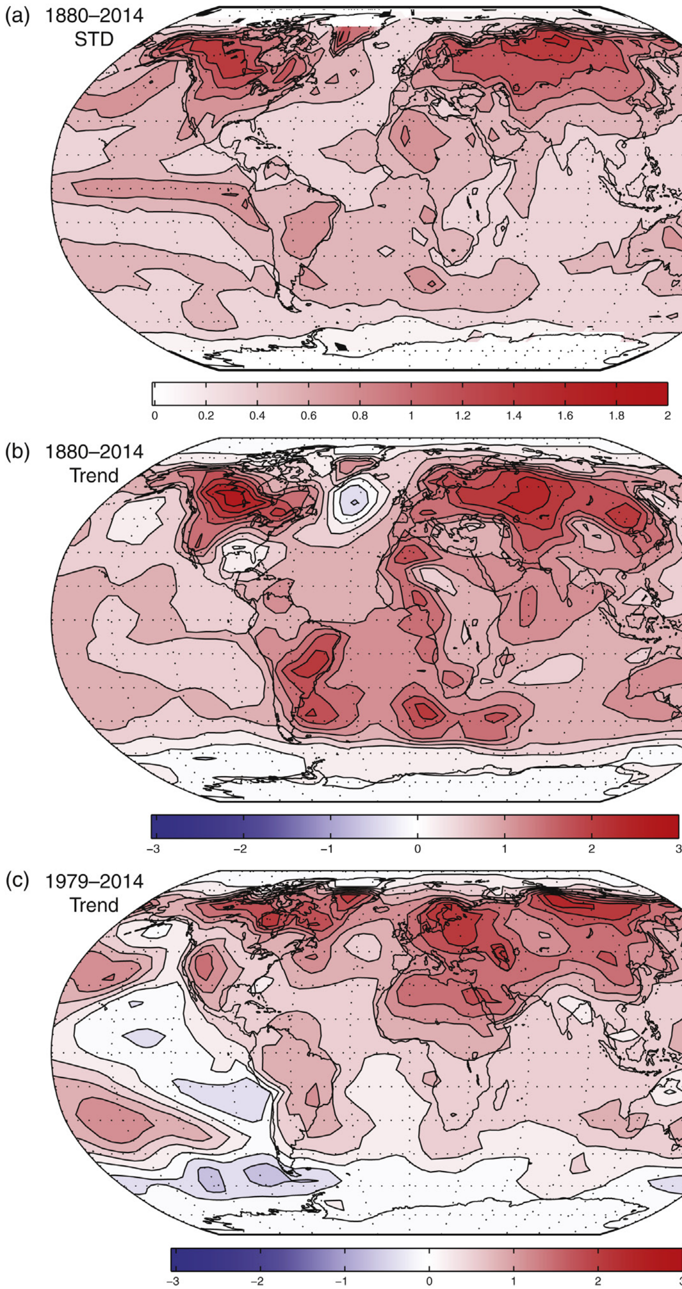

FIGURE 9.2 (a) Standard deviation of annual mean temperature, (b) linear trend over the periods from 1880 to 2014, and (c) linear trend over the period from 1979 to 2014 on the basis of the NOAA combined land and ocean surface temperature data set. Data in polar regions are not sufficient to compute these statistics. Contour interval is 0.2°C for standard deviation and 0.3°C for trend over time.

are larger. The large standard deviation in the equatorial eastern Pacific is associated with ENSO. The trends also show the largest warming over the large continents of the Northern Hemisphere, and generally smaller trends over the oceans. In the North Atlantic and the Far Southern Ocean, the trends are slightly negative, or at least much smaller than the global mean. These are regions where deep mixing occurs, so the effective heat capacity of the ocean is very large.

For contrast, one can consider the trends since 1979, when satellite measurements first started to give global coverage. For this period the standard deviation of annual means has a similar structure (not shown), but the trend structure is significantly different. The greatest warming is also over the northern continents for this shorter record. A stronger cooling over the Southern Ocean is evident, and the contrast between the warming in the Northern Hemisphere and Southern Ocean cooling is greater. In the Pacific Ocean, an east–west asymmetry appears with cooling in the east Pacific and warming in the west Pacific. An asymmetry is also suggested over mid-latitude North America, with greater warming in the Southwest than in the Northeast. Some part of the trend since 1979 is undoubtedly associated with natural variability of the climate system such as ocean–atmosphere coupling in the Pacific, since the changes in the Pacific look like a drift toward more La Niña conditions, but it is also possible that human-induced warming is affecting the natural modes of variability. Some of the dominant modes of natural variability like the PDO and the AMO were shown to have significant trends between 1980 and 2014 in Figs 8.16 and 8.17. A theorem of statistical mechanics says that change first appears in the natural modes of variability of a system.

9.3 THE HISTORICAL RECORD

The historical record consists of written or oral accounts of past events such as agricultural data (records of grape harvest dates in France, wheat yields in Egypt, cherry blossoming dates in Japan, etc.), weather diaries, diaries of a few intellectuals with an interest in weather (e.g., Aristotle, Tycho Brahe, and Johannes Kepler), ship logs, and ancient writings. Other more quantitative information might be about river levels and flooding. The flood levels for the Nile have been recorded for millennia, with especially careful records since 622 CE. Historical climate information is sometimes subjective, often sketchy, and limited to a few geographic areas. Often extreme events find their way into historical accounts, and these may not be representative of the climate of the time.

Although evaluation of the climatic information in ancient art and literature requires great care and skill, much useful and well-supported information can be deduced from historical accounts. Interesting examples include written and archeological evidence of cities that once flourished and then disappeared because of changes in the environment, including the climate. In some cases, it is clear that these changes were in part natural and in part human induced. An example is the city of Ephesus, an ancient Greek city in what is now Turkey. During the fourth century BCE, Ephesus was a major economic power and a center of learning, with an important port and a thriving local agriculture. An amphitheater in Ephesus had a seating capacity of 24,000. Before the city developed, the surrounding hills were covered with oak trees. As the city grew, these hills were cleared and given over to pasture and to cultivation of wheat. At the same time, the climate appears to have become more arid. The combination of changing land use and climate ultimately led to the demise of the city. Erosion led to the filling of the harbor with silt from the surrounding hills. By the ninth century the harbor was unusable, despite dredging the harbor and moving the port. Ephesus was left out of the economic life of the Mediterranean and fell into ruin.

Climatic information can also be contained in art. Rock paintings in the Sahara Desert dating from about 7000 years ago show pictures of the hippopotamus and grazing animals that can no longer exist there. This suggests that the central and southern Sahara were much wetter during the warm epoch following the end of the last glaciation. Paleoclimatic evidence, such as hippopotamus bones dating from the same period and evidence of much higher lake levels, confirms the inference derived from the paintings.

9.4 NATURAL RECORDING SYSTEMS: THE PALEOCLIMATIC RECORD

Another very important class of information on past climates exists for which the recording does not require a human attendant, and from which information for thousands or millions of years into the past can be obtained. This information is cataloged in various types of "natural recording" systems. These give continuous time histories that go back a few million years and are especially good for the last 100,000 years. The sources of paleoclimatic data are outlined in Table 9.1.

The data with the most accurate time chronologies come from analysis of annual tree rings (dendrochronology) or annual layers in lake sediments. The width and structure of the tree rings give some information on the climatic conditions when the tree ring was formed. By correlating tree-ring characteristics with contemporaneous instrumental data on temperature and precipitation, a transfer function can be developed to convert tree-ring characteristics into weather information. Once this transfer function is established and verified, tree-ring data can be used to estimate

TABLE 9.1 Characteristics of Some Paleoclimatic Data Sources

Proxy data source	Variable measured	Continuity of evidence	Potential geographical coverage	Period open to study (year BP)	Minimum sampling interval (year)	Usual dating accuracy (year)	Climatic inference
Layered ice cores	Oxygen isotope concentration, thickness (short cores)	Continuous	Antarctica, Greenland	10,000	1–10	±1–100	Temperature, accumulation
	Oxygen isotope concentration (long cores)	Continuous	Antarctica, Greenland	100,000+	Variable	Variable	Temperature
Tree rings	Ring-width anomaly, density, isotopic composition	Continuous	Midlatitude and high-latitude continents	1,000 (common) 8,000 (rare)	1	±1	Temperature, runoff, precipitation, soil moisture
Fossil pollen	Pollen-type concentration (varved core)	Continuous	Midlatitude continents	12,000	1–10	±10	Temperature, precipitation, soil moisture
	Pollen-type concentration (normal core)	Continuous	50°S to 70°N	12,000 (common) 200,000 (rare)	200	±5%	Temperature, precipitation, soil moisture
Mountain glaciers	Terminal positions	Episodic	45°S to 70°N	40,000	–	±5%	Extent of mountain glaciers
Ice sheets	Terminal positions	Episodic	Midlatitude to high latitudes	25,000 (common) 1,000,000 (rare)	—	Variable	Area of ice sheets
Ancient soils	Soil type	Episodic	Lower and mid-latitudes	1,000,000	200	±5%	Temperature, precipitation, drainage

(*Continued*)

TABLE 9.1 Characteristics of Some Paleoclimatic Data Sources (*cont.*)

Proxy data source	Variable measured	Continuity of evidence	Potential geographical coverage	Period open to study (year BP)	Minimum sampling interval (year)	Usual dating accuracy (year)	Climatic inference
Closed-basin lakes	Lake level	Episodic	Midlatitudes	50,000	1–100 (variable)	±5% ±1%	Evaporation, runoff, precipitation, temperature
Lake sediments	Varve thickness	Continuous	Midlatitudes	5,000	1	±5%	Temperature, precipitation
Ocean sediments (common deep-sea cores, 2–5 cm/1000 year)	Ash and sand accumulation rates	Continuous	Global ocean (outside red clay areas)	200,000	500+		Wind direction
	Fossil plankton composition	Continuous	Global ocean (outside red clay areas)	200,000	500+	±5%	Sea-surface temperature, surface salinity, sea-ice extent
	Isotopic composition of planktonic fossils; benthic fossils; mineralogic composition	Continuous	Global ocean (above $CaCO_3$ compensation level)	200,000	500+	±5%	Surface temperature, global ice volume; bottom temperature and bottom water flux; bottom water chemistry
Rare cores, >10 cm/1000 year	As mentioned previously	Continuous	Along continental margins	10,000+	20	±5%	As mentioned previously
Cores, <2 cm/1000 year	As mentioned previously	Continuous	Global ocean	1,000,000+	1000+	±5%	As mentioned previously
Marine shorelines	Coastal features, reef growth	Episodic	Stable coasts, oceanic islands	400,000	–	±5%	Sea level, ice volume

From National Research Council, 1975.

characteristics of the climate on an annual basis for thousands of years into the past, well before instrumental data became available. A reconstruction of precipitation in Iowa based on a 300-year-old tree correctly shows the "dust bowl" period of the 1930s and the lesser drought of the 1950s. It also shows four decades of dryness comparable to the 1930s that occurred prior to the beginning of the instrumental record. Lake sediments from Africa clearly show evidence of changing precipitation and the climate of the surrounding area can be inferred from pollen spores and other biological indicators in the sediment.

The paleoclimatic data source yielding the longest continuous record is the ocean sediment core. Ocean sediments are deposited over time, so that a core drilled into the sea bottom contains a time history of the environment at the time the layers in the core were formed. The time resolution attainable with sediment cores is determined by the sedimentation rate and the degree of stirring of the most recent sediment by bottom-dwelling animal life such as worms, a process called *bioturbation*. Ocean cores measure the history of the ocean ecology as laid down in the sediment. The sediment includes organic matter and the shells of tiny sea creatures (foraminifera, coccoliths, etc.). Each sea creature has a particular niche in the ecology of the ocean. The relative abundance of some species is related to sea surface temperature (SST), so that relative abundance can be used to estimate SST in the past. In addition, the relative abundance of oxygen isotopes in deep-sea sediment cores provides an indication of the global mass of water tied up in terrestrial ice. Because lighter isotopes (^{16}O) are more readily evaporated, they are more likely to be removed from the ocean and incorporated in ice sheets. Therefore, ice ages are marked by a relatively rich mixture of heavy isotopes (^{18}O) in ocean waters. These higher ^{18}O ratios are imprinted in the shells of sea creatures and are collected in sediments on the ocean bottom. The oxygen isotope abundance in ocean sediment can thus be used to estimate the global volume of ocean water that is tied up in continental ice sheets and mountain glaciers. The anomalies in ^{18}O are expressed as the fractional deviation from a reference oxygen isotope ratio, usually that of standard mean ocean water (9.1).

$$\delta^{18}O = \frac{\left.\frac{^{18}O}{^{16}O}\right)_{Sample} - \left.\frac{^{18}O}{^{16}O}\right)_{Standard}}{\left.\frac{^{18}O}{^{16}O}\right)_{Standard}} \times 1000 \tag{9.1}$$

Figure 9.3 shows a time history of $\delta^{18}O$ of the ocean as a function of time into the past, which has been pieced together from ocean sediment cores by Lisiecki and Raymo (2005). It shows that 5 million years ago the ocean

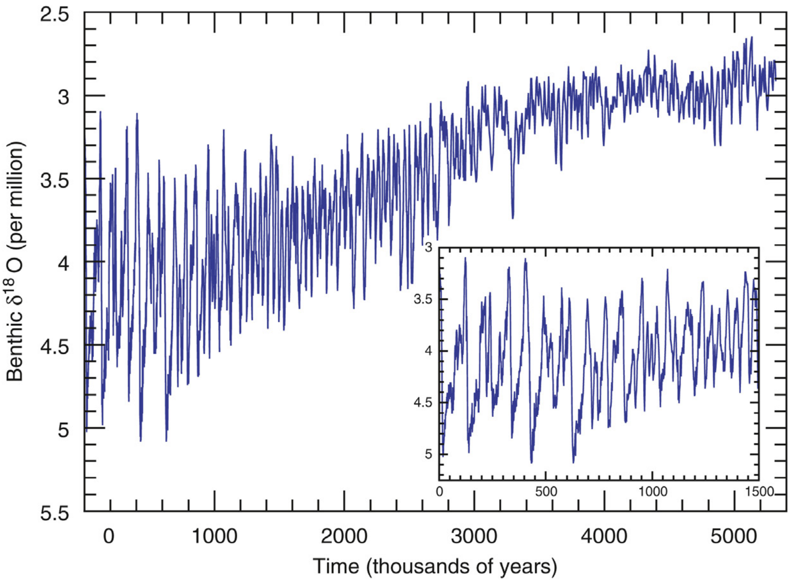

FIGURE 9.3 Ocean oxygen isotope record from ocean cores for the past 5 million years, present to the left and past to the right. Large values indicate large volumes of ice on land. Ordinate is plotted reversed so that up means a warm climate with less land ice and down means a glacial age. Inset shows the same record for the past 1.5 million years. *(Data from Lisiecki and Raymo 2005).*

was relatively light isotopically, indicating a relatively small amount of land ice, and the variations were small, indicating a relatively stable land ice volume. About 3 million years ago, the δ^{18}O began to increase and its variability also increased, indicating more ice and larger variations of ice volume. Until about a million years ago the variations had a characteristic time scale of about 40,000 years, but after that the variations grew even larger and took on a characteristic time scale of about 100,000 years (Fig. 9.4). The periodicities in this time series will be discussed further in relation to the orbital parameter theory of ice ages in Chapter 13.

Similar stratigraphic evidence can be collected in ice cores. Anomalies in stable isotopes such as δ^{18}O and δD [2] indicate temperature anomalies when the ice fell as snow. Water vapor containing heavier isotopes is more likely to condense. If the water in the ice is isotopically light, this means most of the water vapor containing the heavier isotope has condensed out before it reached the ice sheet and the air was anomalously cold. Since saturation vapor pressure depends on temperature,

[2] D = Deuterium = ^{2}H

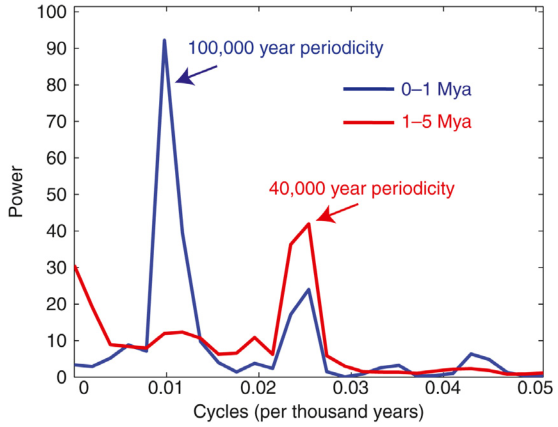

FIGURE 9.4 **Power spectra of the variance of the oxygen isotope record in Fig. 9.4 for the past million years (blue) and for the period from 1 million years ago to 5 million years ago (red).** Note how the strong 100,000-year variations in global ice volume appear only in the past million years, but that the strong variations at 40,000 years appear throughout the record.

the reduction in heavy isotope is related to the amount of cooling since the vapor evaporated. Most of the evaporation occurs in the subtropical oceans, where the temperature is less variable. Aerosols trapped in ice give evidence of past atmospheric dust loading and aerosol chemistry. Air bubbles in ice provide information on the gaseous composition of the atmosphere at the time the bubbles were formed. In the upper parts of ice sheets, annual layers can be identified. Deeper into the ice, and thus farther back in time, the ice becomes compacted and stretched out as the ice flows away from the accumulation regions toward the regions where the ice sheet melts or breaks off into icebergs at a marine boundary. Thus, the ability to resolve rapid changes decreases with depth and age in the core.

9.5 A BRIEF SURVEY OF EARTH'S CLIMATE HISTORY

9.5.1 Early Earth

Earth is believed to have formed nearly 5 billion years ago by the accretion of solid materials and gases that formed the solar nebula. The modern atmospheres of Earth, Mars, and Venus have less abundance of noble gases (e.g., argon, neon) than is typically present in stellar nebula, so it is assumed that any primordial atmosphere collected from the gases in the solar nebula was removed early in solar system history. The primordial atmosphere could have been removed during the collisions of large planetesimals with Earth or been swept away by the more intense solar wind of the early sun. Thus, the modern atmosphere is a secondary one, which has resulted from the release of gases that were mechanically or chemically trapped inside the solid Earth during its formation and were released slowly over time. The process of releasing gases from the interior of a planet may be called *outgassing*, and it continues on Earth today, most obviously in the form of volcanic eruptions. The gases released are primarily water vapor, carbon dioxide, and nitrogen.

We can hypothesize that immediately after the removal of their primordial atmospheres, each of the inner planets consisted of an essentially bare ball of rock, although some gases may have continued to be collected from space during and after the Sun's T-Tauri phase. Assuming that the release of heat from within the planets was negligible, their surface temperatures would have equilibrated at the emission temperature of a sphere with the albedo of bare rock and with a solar constant appropriate to the distance of the respective planet from the Sun. Thus Venus, because of its greater proximity to the Sun, started out at a higher temperature than Earth. Being farther from the Sun, Mars would have started at a lower temperature than Earth. With time, the temperatures of the three planets would gradually increase, because of the greenhouse effect of the water vapor and carbon dioxide that would steadily build up in their atmospheres through outgassing.

On Venus, the surface temperature stayed well above the condensation point for water during its evolution, so that all of the water stayed in the atmosphere. Eventually much of the hydrogen in the atmosphere escaped to space, and a thick atmosphere of primarily carbon dioxide remains. This thick atmosphere gives a very high surface temperature of about 700 K. The process that led to these conditions on Venus is often termed the *runaway greenhouse effect*, but it is also possible that oceans did not form because Venus simply received less water during its formation.

On Mars, it has been speculated that the temperature mostly stayed below the freezing point for water so that the greenhouse effect may never have significantly influenced the surface temperatures there. Because of its relatively low temperature, the partial pressure of water reaches the freezing point before substantial amounts of water vapor can accumulate in the atmosphere. Therefore, the greenhouse effect on Mars may never have really gotten established, or been established only for a time, and most of the outgassed water could have frozen on the surface. A considerable amount of frozen carbon dioxide may also exist on Mars. An alternative to this view of Mars is suggested by some surface terrain with erosional features that appear to have been followed by a running fluid, presumably water, very early in Mars' history.

The distance of Earth from the Sun is such that outgassed water vapor condensed into oceans and the surface temperature remained near the triple point of water, where liquid water, water vapor, and ice can exist simultaneously. The collection of outgassed water vapor in the oceans stabilized the climate and provided the environment that is appropriate for the development of life as we know it. Once the condensation point was reached on Earth, any additional water vapor that was outgassed went into the oceans and the infrared opacity of the atmosphere stopped increasing. The carbon dioxide was dissolved into the ocean and eventually reached equilibrium with carbonate rocks. At this point, the atmosphere was composed mostly of molecular nitrogen. Within about a billion years after the formation of Earth, life developed, leading to green plant photosynthesis, which produces molecular oxygen in the atmosphere. With the development of an oxygen-rich atmosphere came the stratospheric ozone layer, which by protecting the surface from harmful ultraviolet-B radiation, allowed life to emerge from the water and occupy the land surface.

Thus Earth stayed cool enough to avoid the runaway greenhouse effect that occurred on Venus, largely because of its favorable distance from the Sun, which allowed the oceans to form. Life developed quickly in the conditions provided by early Earth, and temperatures favorable for the life forms we are most familiar with ($0 < T < 42°C$) have been maintained at least in some places on Earth, ever since. Theories for the life cycles of stars suggest that, early in Earth history, the solar constant would have been as much as 30% less than its current value. This suggests that some compensating changes in atmospheric radiative properties may have occurred over time to keep the surface temperature in a favorable range for life to progress.

9.5.2 The Last Billion Years

The continents have been drifting as part of lithospheric plates for the last several billion years. Geologists have attempted to reconstruct the positions of the continents for the past 600 million years or so, but little is known about the positions of the continents before that. The continents reached their modern positions about 14 million years ago. Little is known about climate back farther than about a billion years, except for fossil evidence that life existed then and that liquid water was present on the surface (Fig. 9.5).

The climate was cold enough for large ice sheets to form during at least three periods over the last billion years. Evidence indicates ice sheets existed several times during the past billion years, with more extensive evidence for glaciation during the late Paleozoic 300 Mya[3], and fairly detailed

[3]Mya = millions of years ago

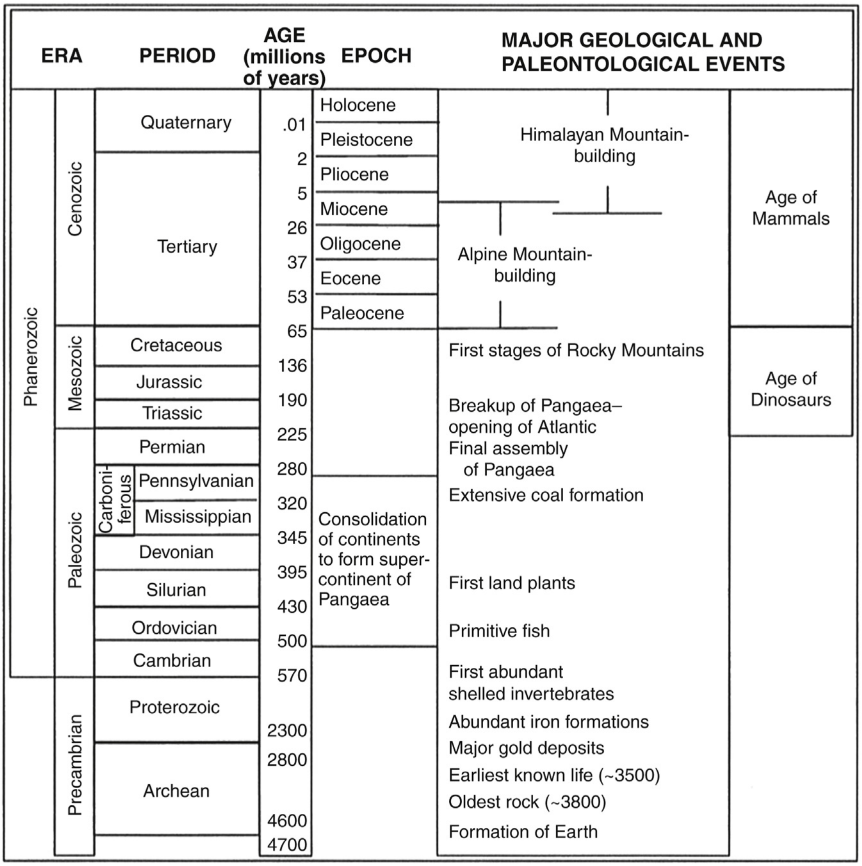

FIGURE 9.5 **The major geologic ages.** *From Crowley (1983).*

documentation for the late Cenozoic glaciation of the last several million years or so, which we are still experiencing.

Evidence suggests that during the Neoproterozoic glaciations about 720 Mya and 635 Mya continental ice reached the sea at low latitudes and the ocean may have been entirely frozen at the surface, which may be termed "Snowball Earth" conditions. At that time the total solar irradiance was about 95% of its current value and greenhouse gases would have needed to be much higher than today to maintain today's climate (Pierrehumbert et al., 2011). The continents were arranged in a single large landmass centered near the equator, which was slowly breaking apart. It is hypothesized that a cool episode was initiated, perhaps by efficient removal of carbon dioxide by weathering from a tropical wet continent, which then led to a full glaciation of Earth through ice–albedo feedback as the polar sea ice expanded to the equator.

On long time scales of thousands to millions of years, weathering and volcanism regulate the CO_2 content of the atmosphere. CO_2 in the atmosphere becomes dissolved in rain to form carbonic acid. When this carbonic acid falls on fresh rock, the rock is dissolved and the carbon flows in streams to the ocean where it is buried as carbonate rock. This carbon is returned to CO_2 when the rock is exposed to heat and pressure within the solid Earth to form metamorphic minerals. The CO_2 is then returned to the atmosphere as outgassing from volcanic activity. Weathering requires liquid water containing dissolved CO_2, so when the surface climate is frozen, weathering is suppressed and outgassed CO_2 can build up in the atmosphere. If the surface is warm and wet, the atmosphere has high concentrations of CO_2, and rock is exposed, weathering proceeds very rapidly. The carbon cycle can thus act as a stabilizing mechanism that eventually drives very cold or very hot climates back toward a more moderate climate.

Other evidence of a frozen ocean is found in iron formations indicating an anoxic ocean, and in carbon isotopic evidence. Carbon that is outgassed from the solid Earth has $\delta^{13}C \approx -6‰$. Photosynthetic life much prefers the lighter ^{12}C isotope of carbon so that organic carbon from marine photosynthetic life is about $-25‰$ lighter than the carbon pool from which it is made. Sequestration of lighter isotopes in organic carbon thus increase the $\delta^{13}C$ of the ocean water. During the Neoproterozoic glaciations the $\delta^{13}C$ in sediments approached the value for outgassed carbon, indicating suppression of photosynthetic life. On top of the glacial layers are thick layers of carbonate rock with $\delta^{13}C \approx -6‰$, indicating the deposition by chemical processes of large amounts of CO_2.

It is estimated that the Earth remained largely ice covered for about 10 million years. The relatively simple life forms present then could have survived in small refugia, such as near geothermal activity, or where wind and sun kept the ocean ice-free. During the snow ball period, outgassing of carbon would continue, while removal of CO_2 by weathering would be suppressed by the cold, dry climate. CO_2 would thus increase in the atmosphere until its greenhouse effect became large enough to overcome the albedo effect of global ice cover. That might require 300 times as much CO_2 as is currently in the atmosphere. Once the ice began to melt, ice–albedo feedback would result in a rapid warming and melting of all the surface ice. At that point, the planet would be very warm, a Hothouse Climate with very high CO_2. Weathering would be very rapid due to the high CO_2 content in a warm climate, and the cap carbonate layer over the glacial sediments would be formed rapidly due to enhanced weathering. This layer has the $\delta^{13}C$ of outgassed CO_2, indicating that the source of the cap carbonate layer was all the CO_2 that had outgassed during the snowball phase.

The late Paleozoic glaciations about 300 Mya appear to have occurred in southern Africa, South America, and Australia, at a time when all these continents were bunched together into a single, large, land mass in high southern latitudes. The glaciated areas were attached to Antarctica and were not far from the South Pole. About 300 million years ago, the continents moved apart and began to drift slowly toward their present positions.

The most studied of the nonglacial climates that occurred between these major glacial ages is the most recent one, the middle Cretaceous, spanning the period from about 120 to 90 million years ago. The continents were in a configuration very different from that seen today. North America and Europe were close together, as were South America and Africa, so that the Atlantic Ocean did not yet exist. Africa had not yet joined Europe, India was a large island in the mid-latitudes of the Southern Hemisphere, and a shallow tropical sea, called the Tethys Ocean, extended from Central America to Indochina (Fig. 9.6). Australia was in middle to high latitudes and still attached to Antarctica. Very little land ice existed during this period, so that the sea level was about 100 m higher than now. This greater mass of ocean water flooded about 20% of the continental areas that are now above water, including large portions of western Europe, northern Africa, and North America.

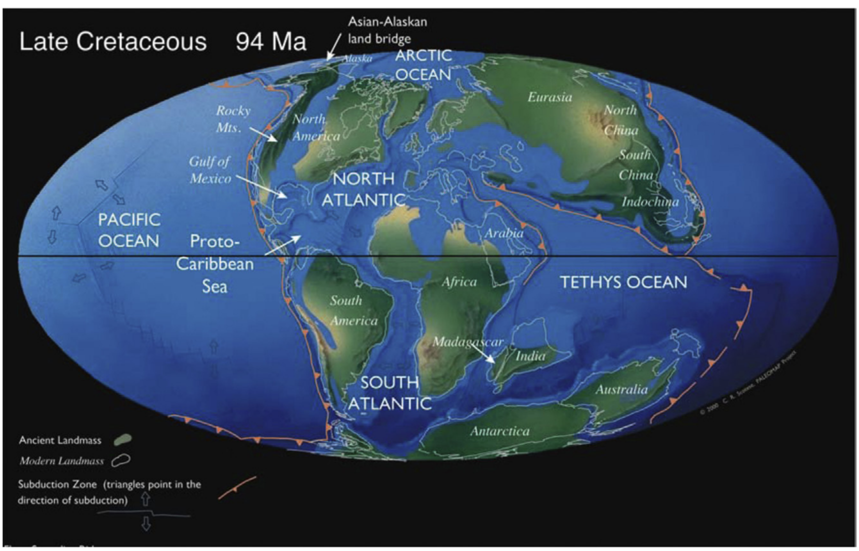

FIGURE 9.6 **Paleomap of the continental positions and sea level during the Cretaceous period 94 Mya.** *Late Cretaceous paleogeographic map by C.R. Scotese, PALEOMAP Project (Scotese, 2001).*

Abundant fossil evidence exists to suggest that the middle Cretaceous was substantially warmer than the present. Plant habitats moved up to 15° of latitude poleward of their current positions, as did the ranges of many animals. Sedimentary evidence also suggests that coal and other deposits indicating warm, moist climate formed during this period in lands that were then above the Arctic Circle, and dinosaurs ranged there also. Isotopic evidence taken from bottom-dwelling organisms suggests that the deep ocean temperature was between 15°C and 20°C. Such warm temperatures in the deep ocean are inconsistent with the presence of much marine ice in high latitudes. Annual mean Arctic temperature is estimated to have been 10–15°C, compared to current temperatures of around −11°C, a difference of around 25°C.

Unusually large amounts of carbon in the form of coal and oil deposits were laid down during the Cretaceous. Many of the major oil deposits were formed during this period. This high rate of coal and oil formation indicates that the cycling of carbon was very different than it is now. A change in the cycling of carbon might logically be associated with the apparently different oceanic circulation, deep ocean temperatures, and the great expanse of relatively warm, shallow seas. The warmth of this period might have been associated with enhanced levels of carbon dioxide in the atmosphere, which might have been up to five times higher during the Cretaceous. Even with these much higher levels of CO_2, climate models have difficulty keeping temperatures warm enough in the centers of northern continents, where paleoclimatic indicators suggest that freezing did not occur in winter.

The Cretaceous ended with a bang about 65 million years ago. About 75% of the total number of living species became extinct, some of them very abruptly. The dinosaurs were among several species that disappeared entirely, and the ascent of mammals as the dominant group of large vertebrates followed and persists today. The so-called *K–T boundary*, marking the end of the Cretaceous (K) and the beginning of the Tertiary (T) periods, is identified in sedimentary records by a well-defined layer of clay deposits. This layer contains anomalously large amounts of iridium. The iridium content of Earth's crust is lower than the amount in the stuff from which the solar system was formed, because much of Earth's iridium was carried to the core with molten iron during Earth's formative stages. Comets and meteors retain their cosmic abundance of iridium, and the clay layer with its high iridium content could have come from a comet or meteor with a diameter of about 10 km. This iridium anomaly at the K–T boundary can be found over a large portion of the globe and is a rare event in Earth history. The evidence is strong that a large meteorite or comet collided with Earth about 65 million years ago, and evidence of a likely impact site has been found near the Yucatan Peninsula.

The impact of such a large bolide with Earth would be a spectacular event with very serious environmental consequences. The shock would heat the atmosphere in the vicinity of the impact, which would produce large amounts of nitrogen oxides. In the stratosphere, nitrogen oxides would lead to a loss of ozone. The nitrogen oxides would likely leave the atmosphere as highly acidic rains. The impact of the projectile and its penetration of Earth's crust would cast fine particles high into the atmosphere and even into low orbits, where they might persist for several weeks or months. This pall of dust could block out the sunlight needed by photosynthetic organisms and would lead to a cooling of the surface after the heat of the impact had dissipated. All of these effects – shock heating, fires, acid rain, ozone loss, dark and cold – would be stressful to many species and may have been the cause of the rapid species extinctions at the end of the Cretaceous.

9.5.3 The Last 50 Million Years

Changes over the last 50 million years seem to be related mostly to the movement of the continents away from Antarctica. During this period, South America and Australia moved northward from the edge of Antarctica to their present positions as a result of sea floor spreading and movement of the continental plates. As the Drake Passage between South America and Antarctica was opened, the Southern Ocean circulation became circumpolar, and the temperature of the surface and deep waters gradually cooled by more than 10°C. Glaciers developed over this period on Antarctica, and the east Antarctic ice sheet formed about 14 million years ago.

Pollen from ocean cores shows that cool temperate forests existed on the Antarctic continent until 20 million years ago. Ice volume over Antarctica increased to reach its present value about 5 million years ago, and the current Northern Hemisphere polar ice sheets first appeared about 3 million years ago. The overall trend in the last 50–100 million years was for the climate to cool from the warm climate of the middle Cretaceous to our present Quaternary ice age.

9.5.4 The Last 2 Million Years

The last 1.8 million years constitute the Quaternary period, the most recent geologic age and the one during which *Homo sapiens* developed. This period is characterized by the presence of a large amount of land ice, which varied from the amount we have today to much larger amounts during periods of glacier advance. The advance and retreat of this land ice can be inferred from glacial deposits and from the ratios of oxygen isotopes in ocean sediment cores (Fig. 9.3). The last 700,000 years were marked by wide swings that indicate a large shift in the amount of land ice present. These swings had a characteristic interval of about 100,000 years between succeeding periods of maximum glaciation. Glaciers advanced and retreated in both hemispheres simultaneously. Prior to about 700,000 years ago the swings were more frequent and less extreme. The dominant period of the glacial–interglacial swings in the early part of the record is about 40,000 years (Fig. 9.4). We will see in Chapter 12 that some parameters of Earth's orbit vary with these periods.

9.5.5 The Last 150,000 Years

About 125,000 years ago the land ice on Earth reached a minimum amount, comparable to today's interglacial conditions. In the intervening period, the land ice gradually increased, while undergoing some minor oscillations, until the ice volume reached a maximum about 20,000 years ago. The most recent and one of the more abrupt swings in global ice volume occurred in the last 20,000 years, between the last glacial maximum and the current interglacial period (Fig. 9.3). The extent of ice cover at the last glacial maximum can be inferred from a variety of paleoclimatic evidence, including the physical evidence left behind by the huge ice sheets and the $\delta^{18}O$ in ocean cores. A synthesis of this information indicates a giant ice sheet covering much of northern North America, and another large ice sheet in northern Eurasia, which were 3–4 km thick (Fig. 9.7). The weight of these ice sheets was so great that the supporting crust was depressed by nearly 1 km. The plastic deformation of the crust under these massive ice sheets may have been one reason for the rapid removal of ice at the end of each major glacial maximum during the past 700,000 years. As the crust yields under the ice, the altitude of the ice is lowered, thus bringing the ice surface to higher ambient air temperature. Also, glacial lakes may form in the depression made in the crust by the ice sheet, and seawater may flow into the depression and assist in the removal and melting of ice. The crust near the locations of the maximum thickness of the major ice sheets is still rebounding from the removal of the ice sheets. Because so much water was tied up in these large continental ice sheets, sea level was about 120 m lower 20,000 years ago than at present. The $\delta^{18}O$ records and modeling studies suggest that this was about the maximum amount of water that could be moved to continental ice sheets, given the continental positions and climate regime of the late Quaternary. Because sea level was so much lower, the coastlines of the continents were changed. The Bering Sea became a land bridge so that early Americans could walk from Siberia to North America, and Southeast Asia extended to Indonesia (Fig. 9.7).

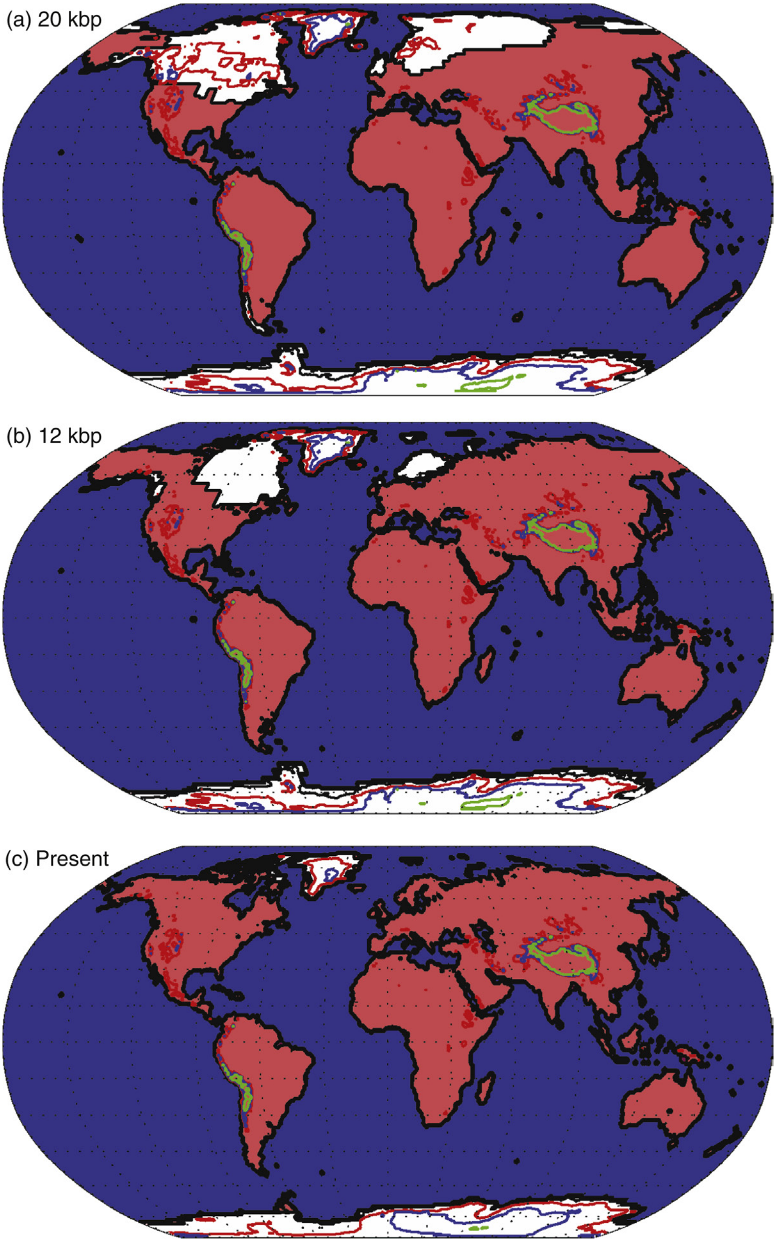

FIGURE 9.7 Maps of ice sheets and elevation above sea level for (a) 20 kbp, (b) 12 kbp and (c) present. White indicates perennial ice sheet positions. Red, blue, and green contours represent 2000, 3000, and 4000 m elevations, respectively. *Data from Peltier (1994), archived by the NOAA/NGDC Paleoclimatology Program, Boulder, Co.*

The high-latitude oceans also experienced major changes during the last glacial maximum. The Gulf Stream turned more sharply eastward at about 45°N and warm currents did not extend northward into the Greenland and Norwegian Seas as at present. It is likely that during winter the sea ice extended equatorward of 50°N, and snow cover on land extended ever further equatorward, so that a vast area of ice-covered surface extended from about 45°N to the pole in the Northern Hemisphere during winter. A simultaneous glacier advance was experienced in the Southern Hemisphere, where sea ice around Antarctica was greatly expanded, and mountain glaciers covered parts of Australia, Africa, and South America. During this time, tropical continental regions appear to have been drier than at present, but snowlines on tropical mountains dropped about 1000 m. Mid-latitude continental regions near the ice sheets were also drier for the most part, and windblown dust deposits in these areas indicate dry windy conditions at the equatorward margins of the great ice sheets.

Careful analysis of ice cores from the Greenland and Antarctic ice sheets can reveal much about changes in the climate system over the last major glacial–interglacial cycle. The composition of the atmosphere in the past can be inferred from air bubbles trapped in an ice core. Figure 9.8 shows 800,000-year records from an Antarctic ice core of temperature anomalies estimated from Deuterium abundances in the ice, atmospheric carbon dioxide (CO_2); and methane (CH_4) concentrations measured from bubbles trapped in the ice, and dust in the ice core. The temperature, CO_2, and CH_4 records closely track each other, and the dust is greatest during the coldest periods. About 20,000 years ago, during the most recent glacial maximum, CO_2 concentrations were about 190 ppmv, compared to about 275 ppmv during the warm periods, and CH_4 was about 350 ppbv compared to 700 ppbv during the warm periods. These large changes of atmospheric carbon dioxide and methane between glacial and nonglacial times indicate major changes in the cycling of carbon between ocean, atmosphere, and land, which accompanied the growth and decay of ice sheets. In Chapter 13, we will see how CO_2, CH_4, and other greenhouse gases have increased dramatically in the past century due to human activities, after remaining stable for thousands of years.

Ice cores also contain information on past variations in the amount and chemical composition of aerosols deposited in high latitudes. Non-sea-salt-sulfate and terrestrial calcium are much higher during glacial than interglacial ages. After correcting for the effects of changing accumulation rates, these changes indicate that sulfate concentration in the atmosphere was 20–40% higher during the glacial maxima. The increased atmospheric sulfate is thought to be related to increased production of sulfur-bearing gases by life in the ocean, most probably dimethyl sulfide gas.

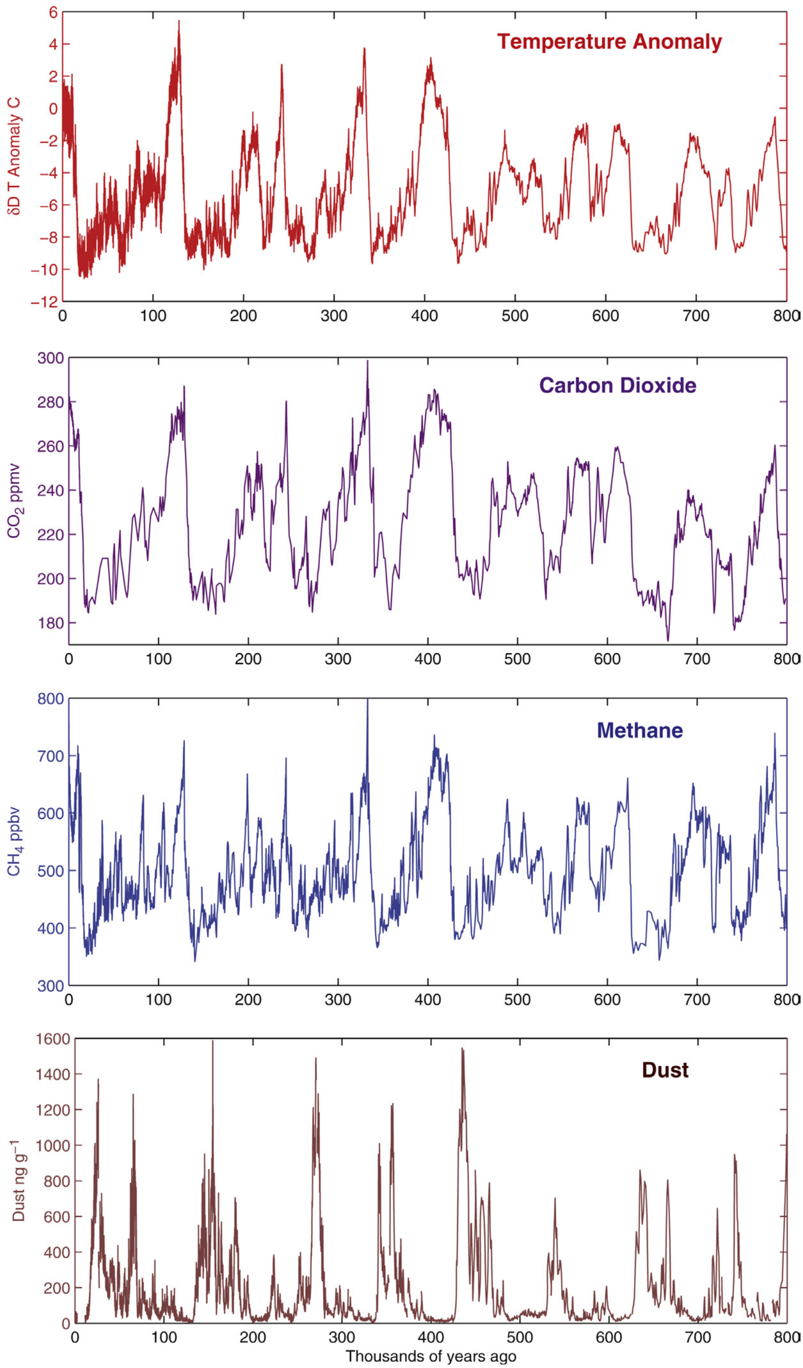

FIGURE 9.8 Estimates of temperature, CO_2, CH_4, and dust from ice cores drilled at Dome C in Antarctica plotted as functions of time before present for the past 800,000 years. *Data from NOAA/NCDC Paleoclimatology Program, Boulder CO, USA; Jouzel et al. (2007), Lambert et al. (2008), Loulergue et al. (2008), and Luthi et al. (2008).*

Variations in atmospheric carbon dioxide of the magnitude observed must be related to changes in the gross biological productivity of the oceans through photosynthesis. An increase in productivity during the ice age would reduce the partial pressure of CO_2 in the water and thereby reduce the CO_2 content of the atmosphere, which is constrained by the ocean concentration on these time scales. If it is assumed that the sulfate is produced biogenetically through dimethyl sulfide (DMS) emissions by phytoplankton, then the sulfate increase is consistent with an increase of gross ocean productivity during the ice age, assuming that the rate at which organisms in the ocean produce DMS rises and falls with gross productivity. Greater production of sulfate may also have increased the number of cloud condensation nuclei, which may have increased the reflection of radiation by clouds and aided in maintaining the cold climate. Comparison of CO_2 from ice cores with temperature estimates from many sources has indicated that the CO_2 concentration lags behind the temperature, with particularly clear evidence during the deglaciation following the Last Glacial Maximum (Clark et al., 2012). This supports the idea that during the Pleistocene the timing of glacial–interglacial transitions were set by orbital parameter variations that changed the temperature and ice volume, and that atmospheric CO_2 responded to the temperature change and thus forms a positive feedback on climate changes arising from other causes.

As discussed in Chapter 7, at the present time much of the intermediate deep water in the global ocean is formed in the North Atlantic Ocean. Because it is formed by cooling water that has spent some time near the surface where nutrients are efficiently consumed, north Atlantic deep water (NADW) is generally deficient in nutrients. In contrast, the source of deep water formed around Antarctica is water that has spent considerable time at intermediate depths collecting nutrients and then rises to the surface in the Southern Ocean. Therefore, Antarctic deep water is comparatively nutrient rich. Because of the different nature of North Atlantic and Antarctic deep water formation, a substantial gradient in deep-water nutrient content exists between the North and South Atlantic and between the Atlantic and the Pacific oceans.

Evidence from ocean sediment cores suggests that the contribution of the North Atlantic to deep-water formation was greatly diminished during the glacial maximum of about 20,000 years ago. Associated with the low nutrient content of NADW are relatively high levels of $\delta^{13}C$, a measure of the ratio of ^{13}C to ^{12}C, and low levels of the cadmium to calcium ratio, Cd/Ca. A time history of past levels of $\delta^{13}C$ and Cd/Ca in deep waters can be obtained from analysis of sediment cores. Records of $\delta^{13}C$ and Cd/Ca from ocean sediments indicate that the rate of deep-water formation in the North Atlantic Ocean was greatly reduced during the last glacial maximum, and nutrients in the deep ocean were more uniformly mixed.

The formation of deep water in the North Atlantic and the associated heat and water transports that were suppressed during the last glacial maximum increased again about 14,000 years ago at a time when the $\delta^{18}O$ in ocean cores indicates that the land ice volume began to decrease rapidly. The proxy data indicate that deep-water formation in the north Atlantic ceased again for a period between about 13,000 and 11,000 years ago. During the same interval, fossil data indicate that the climate in Europe returned nearly to glacial conditions after having warmed considerably during the period from 15,000 to 13,000 years ago. The cold event is known as the *Younger Dryas*, because pollen data indicate that forests that had recently developed in Europe during the aborted warming following the ice age were suddenly replaced again by arctic shrubs, herbs, and grasses, including the herbaceous plant *Dryas octopetella*. It ended abruptly about 11,000 years ago with the return to the interglacial conditions of today. Although the event is strongest in the North Atlantic region, it is seen in many fossil records around the world.

The Younger Dryas event is very evident in time series of $\delta^{18}O$ in ice cores from Greenland (Fig. 9.9). $\delta^{18}O$ in ice cores is a proxy for the air temperature in the vicinity of the ice sheet. When the temperature near the ice sheet is low, it is more likely that the heavier isotope of oxygen would have condensed out before reaching the ice sheet, because of its lower saturation vapor pressure, so lower $\delta^{18}O$ in glacial ice is

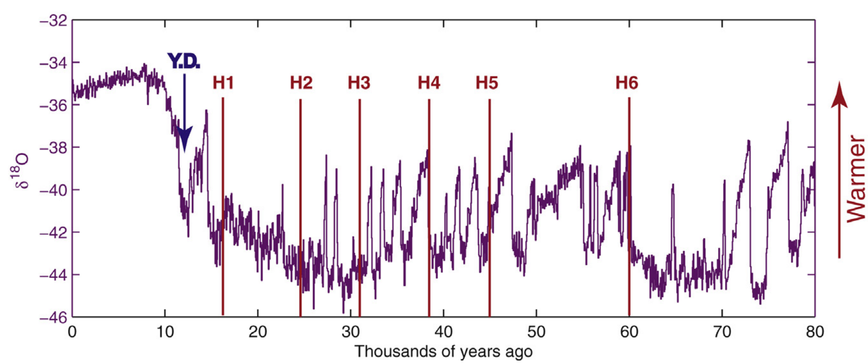

FIGURE 9.9 **$\delta^{18}O$ from the NGRIP ice core from Greenland for the past 80,000 years (Andersen et al., 2004).** The Younger Dryas event is marked with a blue arrow. The six Heinrich events of exceptional ice rafted debris in Atlantic Ocean sediment cores are shown with red lines (Hemming, 2004).

associated with colder temperatures when the snow fell. Greenland ice cores indicate a local cooling of about 6°C during the Younger Dryas cold event.

Figure 9.9 shows other interesting features of Greenland temperature during the ice ages prior to the current interglacial. Several events similar to the warm event prior to the Younger Dryas cooling occur in the record. A rapid warming of about half way to interglacial conditions occurs, followed by a period of slower cooling back to near glacial maximum temperatures. These events are called Dansgaard–Oeschger (D–O) events after their discoverers, Willi Dansgaard and Hans Oeschger, and they occur about 25 times during the previous glacial period. Particularly strong and long-lasting events are seen at about 38,000 and 48,000 years ago. Temperature changes over Greenland during D–O events were 9–16°C. Signals corresponding to D–O events in Greenland are also recorded in Antarctic ice cores, but with cooling in Antarctica corresponding to the warm events in Greenland, indicating a bipolar seesaw (Barbante et al., 2006). The cooling in Antarctica is much more gradual than the warming in Greenland and lags behind by about 200 years, suggesting that the two hemispheres are connected through the circulation of the Atlantic Ocean.

Also shown in Fig. 9.9 are six Heinrich events. Heinrich events are identified in ocean sediments from the North Atlantic, and are characterized by large amounts of ice-rafted debris. Large amounts of sand from continental rocks indicate that a large number of icebergs from continental glaciers had floated out into the Atlantic and melted, dropping sand and stones into the sediment. Heinrich events tend to occur during times when Greenland is cold, but their relation to D–O events is unclear.

It has been hypothesized that the brief shutdown of the deep thermohaline circulation of the North Atlantic during the Younger Dryas event was associated with the melting of the North American Laurentide ice sheet. The ice sheet melted most rapidly from its southern flank at first and the majority of this melted water flowed down the Mississippi River draining into the Gulf of Mexico. Several great meltwater lakes formed in the depression of Earth's surface left by the retreating ice sheet, and one of these occurred in what is now southern Manitoba. This paleo-lake has been named Lake Agassiz after the geologist, Louis Agassiz, who in 1837 was an early and ardent proponent of the idea that Earth had undergone an ice age. As the ice sheet retreated farther north, a channel was opened that allowed the meltwater lake to drain eastward down the St. Lawrence River to the North Atlantic Ocean. This diversion is dated at about 12,000 years ago from land evidence and by $\delta^{18}O$ records in sediment cores from the Gulf of Mexico, which went from isotopically light to heavy as the low $\delta^{18}O$ meltwater was suddenly diverted away (Fig. 9.10e). The supply of freshwater to the North Atlantic via the St. Lawrence reduced the salinity of the surface ocean waters. Since high salinities are critical to attaining the densities required to form deep water, the supply of freshwater was sufficient to cut off the thermohaline circulation of the North Atlantic. With the thermohaline circulation went its associated heat transport, resulting in large local cooling of the climate. The thermohaline circulation restarted about 11,000 years ago when the meltwater was again directed down the Mississippi, and has continued to operate as the climate warmed to today's interglacial conditions.

The Younger Dryas cool period occurred after the aborted warming of the Bölling-Alleröd from 14.7 to 12.7 kya[4], which followed the Oldest Dryas cold period from 18–15 kya. These temperature variations that are mostly referenced to North Atlantic climate have been shown to have global connections. Figure 9.10 shows a set of climatic indicators from 25 kya to present. Figure 9.10a,b shows temperature and ice accumulation records from Greenland indicating a warming to near present conditions about 14.7 kya (vertical line), which then dropped back to near full glacial conditions during the Younger Dryas event (shading). Figure 9.10c shows the record of temperature over Antarctica, which started warming about 18 kya, then cooled again after Greenland warmed about 14.7 kya, then warmed through the entire Younger Dryas event. These events are characteristic of what is called the *bipolar seesaw* between Greenland and Antarctica, which are connected through the deep circulation of the Atlantic Ocean. During the glacial maximum the Atlantic Meridional Overturning Circulation (AMOC) was suppressed and more of the deep water was formed by the enhanced sea ice extent in the Southern Ocean. As the Earth started to warm around 18 kya, the AMOC remained suppressed until it started again abruptly about 14.7 kya and the north then warmed rapidly. The SST in the Caribbean Sea also increased about 3°C at this time, as is recorded in the sediments of the Cariaco Basin (Fig. 9.10d). Then, however, increased freshwater flooded the Gulf of Mexico in several pulses about 17 and 13 kya, with the second of these marking the beginning of the Younger Dryas and a return to near glacial temperatures. It is hypothesized that the flood of freshwater from the melting ice sheets (Fig. 9.10e) suppressed the AMOC during the Younger Dryas, which then ended very abruptly with a change to near modern temperatures about 12.7 kya. Again the warming in the North Atlantic lagged behind that in the Antarctic, which changed more gradually, since the Antarctic is less dependent on the northward extension of the AMOC than is Greenland and the North Atlantic Ocean. The pattern of changes around the Atlantic

[4]kya = thousands of years ago

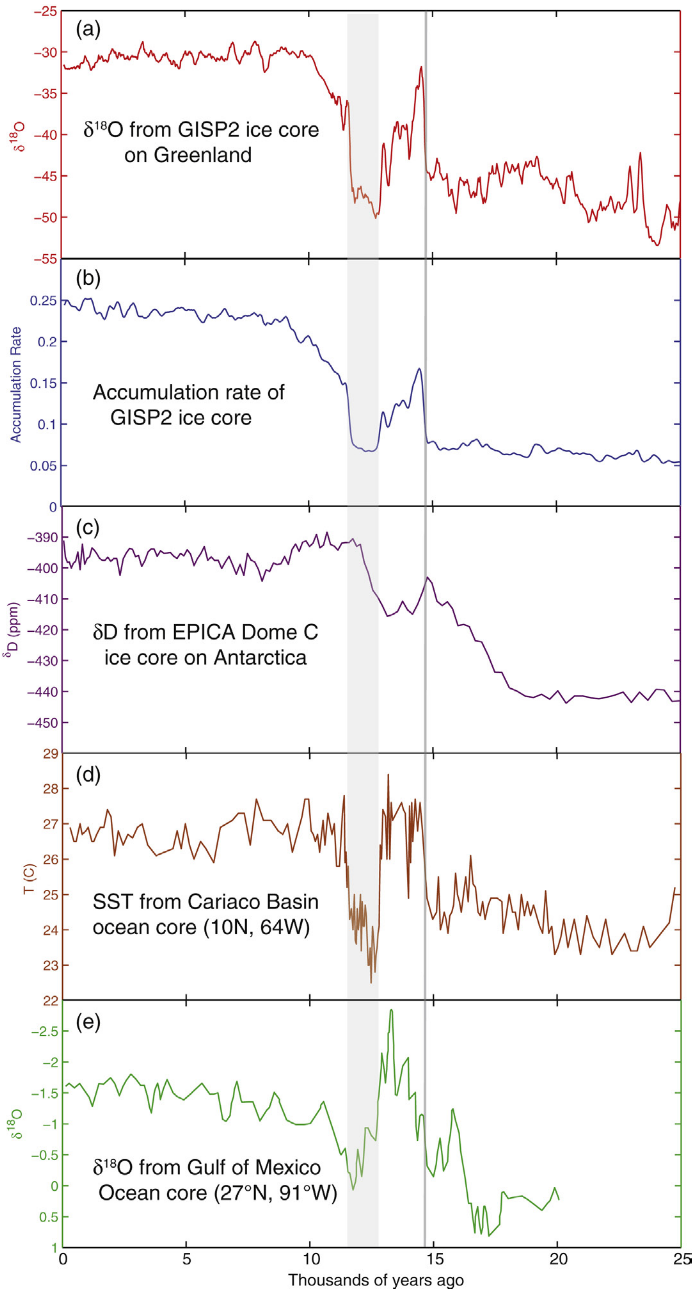

FIGURE 9.10 Five panels of the deglaciation showing the global nature of the Younger Dryas event (shading) and the Bölling warm event (vertical line) in time series from 25,000 years ago to the present. (a) δ^{18}O from GISP2 Greenland ice core indicates temperature near the top of Greenland. (b) Accumulation rate of ice for the GISP2 ice core (Alley, 2000). (c) δD from the EPICA-C ice core indicating temperature near the top of Antarctica (Augustin et al., 2004). (d) SST from an ocean sediment core near Venezuela (Lea et al., 2003). (e) δ^{18}O from an ocean sediment core in the Gulf of Mexico, indicating freshwater flux from North America and global land ice amount (Leventer et al., 1982).

Ocean indicate that the AMOC started and stopped during the deglaciation following the last glacial maximum.

9.5.6 The Last 10,000 Years

The global climate warmed following the last glacial maximum. Melting of the great ice sheets took about 7000 years from about 14 kya to about 7 kya, with considerable variation in the dates of final melting for different ice sheets. The climate has been comparatively steady for the past 8000 years or so (Figs 9.9 and 9.10). The early Holocene from about 10,000–5,500 years ago was a time when Northern Hemisphere summer climates were warmer than today's. This period was also characterized by remarkable changes in the hydrology of the monsoonal climates of Africa and Asia. Evidence from closed-basin lakes indicates that the period from about 9,000–6,000 years ago was the wettest in northern Africa in the last 25,000 years. It was during this interval that the hippopotamus, the crocodile, and a variety of hoofed animals occupied regions in the Sahara and in Saudi Arabia that are now some of the most arid on Earth. Sediment cores from the Arabian Sea indicate that monsoonal winds were stronger during the early Holocene than at present, and sediments from the Indus and Ganges rivers indicate that precipitation over India was greater than now. All of this evidence is suggestive of stronger summer monsoons during the early Holocene, which has been related to changes in Earth's orbital parameters, as will be discussed in Chapter 12.

Minor advances of mountain glaciers culminated about 5300, 2800, and 150–600 years ago. The last of these falls within the Little Ice Age, circa 1250–1850 CE. Between these cold periods there were relatively warm periods, the best known being around 900–1200 CE and during the present century. During the former period, the Norsemen populated Iceland, Greenland, and North America. The Norse Greenland colonies were abandoned at the onset of the Little Ice Age. Fields in coastal Greenland that had grown crops became perennially frozen. This seemingly drastic change, compared to what has occurred elsewhere, may be because climate changes appear to have larger amplitudes at high latitudes, or because of the sensitivity of Greenland climate to the strength of the deep circulation in the Atlantic Ocean.

9.6 USES OF PALEOCLIMATIC DATA

Paleoclimatic data are very useful in developing a scientific understanding of climate change that will be essential to anticipating the nature of future climate changes associated with human activities and natural processes. The description of past climates derived from paleoclimatic data is useful for three reasons: It gives us perspective on the range of climate changes that are possible and likely, it provides clues about how the climate system works, and it provides a data set for testing theories and models of how climate changes.

Paleoclimatic data give us a perspective on what constitutes a "significant" global climate change. During a full-glacial episode like the one 20,000 years ago, the global mean surface temperature was about 5°C colder than now. Therefore, 1°C in global mean surface temperature represents about 20% of the temperature change between glacial and interglacial conditions. The current climate is very near the warmest that has been observed in the last million years, so we do not have very good analog information for really warm world climates. The warm period about 9000–6000 years ago was nearly the warmest in the last million years and was probably not as much as 1°C warmer than today. To get climates that were substantially warmer than today one must return to the early Pliocene, about 5 million years ago, or to the Cretaceous period, about 65 million years ago, when continental positions and the configuration of the world ocean were very different from today. The dominant time scales for natural climate changes during the current geologic epoch, the Holocene, seem to be in the tens of thousands to hundreds of thousands of years. Although climate changes on a variety of time scales, the largest variations appear to occur on longer time scales. Nonetheless, the record also contains evidence of rapid changes of substantial magnitude on time scales of tens to hundreds of years, such as those associated with the Younger Dryas event and the Little Ice Age.

Paleoclimatic data can provide clues about how the climate system works and how the observed variations were produced. The thick ice sheets over the continents during glacial maxima suggest that continental ice sheets may play an active role in climate change, through their albedo, topography, and effects on ocean circulation and biogeochemical cycles. Temperature changes during an ice age seem to be largest at high latitudes and rather small near the equator. Why is this and what does it tell us about the mechanisms of climate change? The tropical land regions appear to have been drier on average during glacial maxima than during warm periods. What are the reasons for these hydrological and vegetation changes and do they play an active role in climate variability? Air trapped in bubbles in glacier ice shows that the carbon dioxide content of the atmosphere was lower (~ 190 ppmv) during the last ice age than 200 years ago before the industrial revolution (~ 280 ppmv). This suggests that global ocean productivity was enhanced during the ice age. Does this information indicate a role for biology in determining climate variability?

Paleoclimatic data provide evidence for testing theories and models of global climate change. Time series of global ice volume can be compared to time series of insolation variations associated with cycles in Earth's orbital parameters. Can we explain the relationship between Earth's orbital parameters and global ice volume? If we put inferred distributions of SST, surface ice, vegetation, and atmospheric composition into our best climate models, do they produce a consistent climate in balance with these conditions? How important are the changes in atmospheric composition for explaining glacial–interglacial cycles of climate change? In Chapter 10 we examine the question of how sensitive the climate is to ice sheets and atmospheric composition, and in Chapter 12 we examine the role of orbital parameter variations and other natural climate forcings in determining climate variability.

EXERCISES

1. Discuss the difficulties with interpreting a curve such as Fig. 9.1 or extrapolating it into the future.
2. One of the earliest examples of primitive humans, *Australopithecus*, lived in Africa between about 4 and 1 million years ago. Hominids began using stone tools about 2 million years ago. Place *Australopithecus* and tool use in their proper place in Fig. 9.5. For what fraction of Earth history have hominids been around?
3. How would the differences of atmospheric CO_2 and CH_4 during the last glacial maximum and the present preindustrial era have contributed to the differences in the climates between then and now?
4. Estimate the rate of freshwater mass production (kg s^{-1}) by the melting of the North American ice sheet, assuming that its melting took 500 years and that it consisted of 10^{19} kg of water. How does this compare with the rate of deep-water formation in the North Atlantic, which is estimated to be 1.5–2 $\times$ 10^{10} kg s^{-1}?
5. Estimate by how much the freshwater flux calculated in problem 5 would decrease the salinity of the surface waters of the far north Atlantic if the flow of the 35% salinity water from the south is 2 $\times$ 10^{10} kg s^{-1}?
6. What fraction of the North American ice sheet at its maximum would need to melt in 100 years in order that the flow of freshwater over this period would decrease the salinity by 2%? How would such a burst of freshwater flux affect the formation of deep water? What would be the response of the climate in the north Atlantic region? How would the rate of melting respond to the climate changes? Can you imagine how the interactions between the melting rate, deep-water formation, and climate variations might cause an oscillation during the decline of the ice sheet?

7. What do you think might have caused the D–O events during glacial times? Why is the warming more rapid than the cooling?
8. Why do you think more dust was deposited in Antarctica during glacial times than during interglacial times?
9. Explain why the North Atlantic and Antarctic glacial age indicators show a bipolar seesaw.