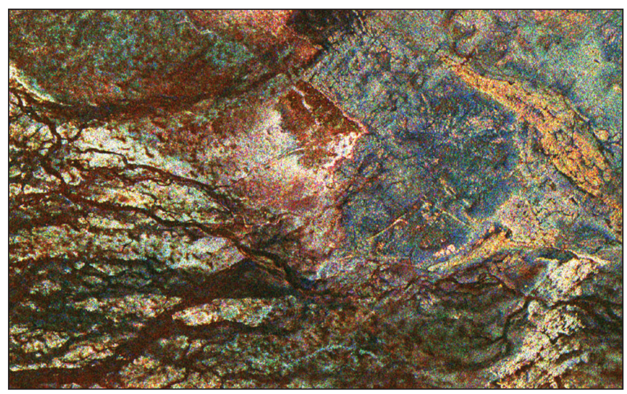
CHAPTER 9

THE CLIMATE RECORD: PALEOCLIMATES



Clues such as ancient river beds help scientists reconstruct what a region's climate must have been in the past. The Safsaf Oasis in the Sahara was clearly once a thriving river valley. An ancient river's widest channel runs from the lower left corner towards the center of the image. The image was taken by the Spaceborne Imaging Radar-C/X-band Synthetic Aperture Radar (SIR-C/X-SAR), which uses radar to penetrate the thin sand cover. The sensor flew aboard the space shuttle Endeavour, and took this image on April 16, 1994. [Courtesy NASA]

Case-in-Point

Late in the 13th century CE, the Anasazi people (also known as the Ancestral Pueblo) abandoned their homelands near the present-day Four Corners area of the American Southwest (where Arizona, New Mexico, Colorado, and Utah adjoin). At about the same time the Anasazi's neighbors to the north and northwest (in present-day Utah and western Colorado), the Fremont people, permanently left the region after 1000 years of habitation (Figure 9.1).

Why did these people leave their ancestral lands? The first archaeological studies of the ruins of the Anasazi

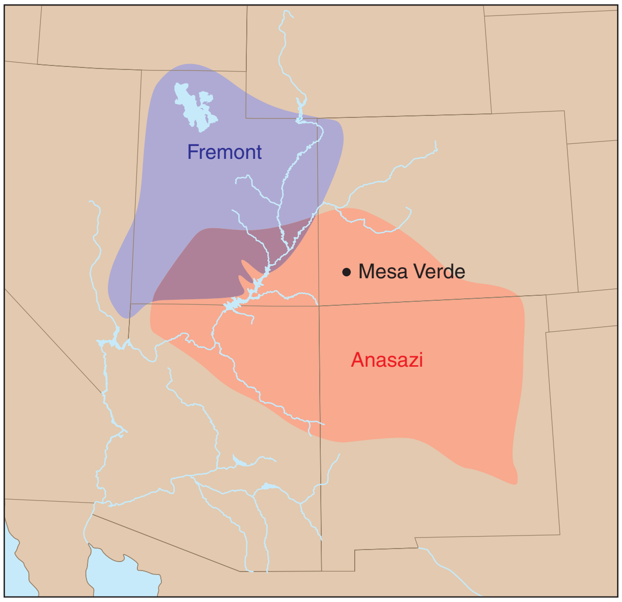


FIGURE 9.1
Map of the American Southwest showing the ancestral lands of the Anasazi and Fremont peoples.

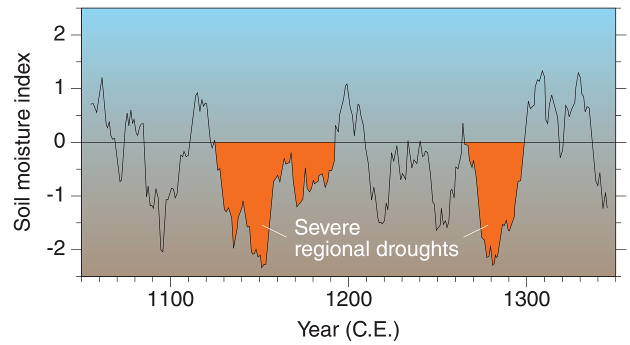


FIGURE 9.2
Drought record from the Four Corners Area of the Southwest derived from tree growth ring data. [Modified after Larry Benson of the U.S. Geological Survey and Edward R. Cook of Lamont-Doherty Earth Observatory]

cliff dwellings attributed abandonment to a single factor, either climate change or conflict. In the March-April 2008 issue of *American Scientist*, Timothy A. Kohler, an anthropologist at Washington State University, and his colleagues proposed a combination of causes: climate change, population growth, competition for resources, and conflict. This proposed explanation came from the *Village Ecodynamics Project* which focused on how the Anasazi interacted with their environment. The southwest Colorado study area covered 1816 km^2, and among the data gathered were proxy climate records from tree growth rings, pollen profiles, and macrofossils. Organic materials were dated using the radiocarbon method.

Kohler and colleagues noted that the Anasazi experienced two major cycles of population growth and decline: CE 600-920 and CE 920-1280. In both cycles, the drop in population was preceded by aggregation of the population in relatively dense clusters. Abandonment took place by the end of the second cycle.

The Anasazi were farmers and hunters who, in the 12th and 13th centuries, depended heavily on maize (corn). It was their principal source of carbohydrate calories and was fed to domesticated turkeys, their primary protein source. Droughts in the 1100s and 1200s caused shortfalls in the maize harvest and serious nutritional deficiencies. Some of these episodes constituted **megadroughts**, which are droughts that persist for multiple decades (Figure 9.2). Readily recognized in the tree growth ring records of the period, megadroughts occurred from CE 1135-1180 and CE 1276-1299. The first of these (perhaps the most severe of the past two millennia) forced the Anasazi to leave Chaco Canyon (New Mexico) and move to the cliffs of Mesa Verde, CO (Figure 9.3).

Between droughts, during the relatively wet periods, the population soared, making them all the more vulnerable when drought and food shortages returned. In response to food scarcity, people stored food in granaries hidden away in places difficult to access such as high cliffs and narrow ledges. And the repeated failures of the maize harvest forced the Anasazi to depend increasingly on hunting and gathering. Inevitably, competition for declining food sources led to conflict among neighbors; some sought shelter in cliff dwellings, but there is evidence that conflict culminated in violent confrontations and death.

To the north and northwest, similar stresses were impacting the Fremont people as megadroughts caused food shortages and civil strife during the 1200s. The

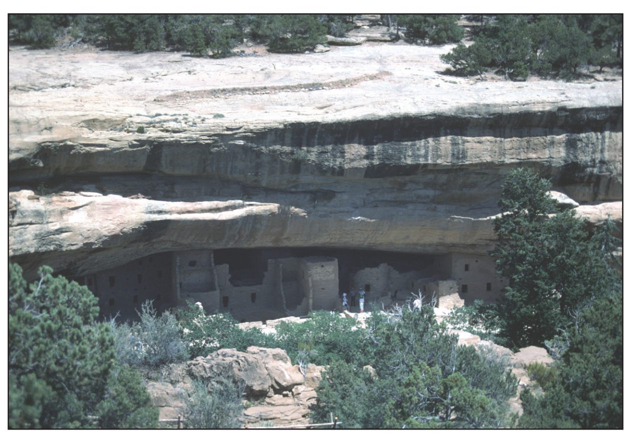


FIGURE 9.3
Cliff dwellings of Mesa Verde in southwestern Colorado.

Fremont people were somewhat more versatile than the Anasazi; depending on environmental conditions, they readily shifted between hunting, farming, and foraging. Nonetheless, individuals were forced to hide what little food was available and to build defensive structures on ridge tops. Eventually the Fremont people also had no alternative but to abandon the Southwest.

In summary, many factors combined to force the Anasazi and Fremont peoples to abandon their ancestral homelands. A climate change involving more frequent megadroughts was one of those factors. In fact, often climate is only one of many factors that in combination result in societal upheaval.

Driving Question: *How and why do scientists reconstruct the climate record prior to the instrument era?*

Extending the climate record as far back in time as possible aids our understanding of Earth's climate system, provides a valuable perspective on the present climate, and with an eye to the future, gives insight as to the nature of climate change. For example, reconstruction of the long-term mean annual Northern Hemisphere temperature indicates that the observed warming trend of the late 20th century was not equaled in magnitude during the past millennium. However, the reliable instrument-based climate record taken under standardized conditions extends back to only the 1870s. Climate information prior to the instrument era is drawn from a variety of climate-sensitive sources, including historical documents, tree growth rings, pollen profiles, coral, and deep-sea sediment cores.

Climate reconstruction often requires calibration of modern climate forcing with modern environmental response. That calibration is applied to ancient environmental response records (e.g., fossil flora and fauna in bedrock) to unlock information on the climate past. Reconstruction of past climates is the principal realm of the subfield of climatology known as **paleoclimatology**. The accuracy of climate reconstructions generally decreases with increasing time before present because of limitations, such as declining resolution of information, increasingly fragmented data, and problems with time control and correlation of data. Ideally, proxy climate data can be tested for validity by comparing the data with the instrument record for overlapping periods. The problem here is that instrument-derived data typically offer much finer resolution. Consequently, the climate of geologic time, spanning hundreds of millions of years, or longer, can be described in only very general terms. Resolution improves for the latter portion of the Pleistocene Ice Age and into the Holocene Epoch.

The primary purpose of this chapter is to survey the techniques and findings of paleoclimatology, beginning with a general overview of why and how past climates are reconstructed, followed by a brief summary of the major proxy climate data sources. We then summarize the climate record prior to the instrument era.

Reconstructing Past Climates: Why and How?

Why reconstruct climates of the past, that is, climates prior to the era of reliable instruments and standardized methods of observation? As with any mystery, the climatic past appeals to human curiosity and offers a challenge. But beyond satisfying an inquisitive nature, many practical reasons justify our delving into the climate past. For one, the process of climate reconstruction improves our understanding of environmental response to climate variability and climate change. A lengthier climate database encompasses a broader range of fluctuations in Earth's climate than is represented in the reliable instrument-based climate record, thereby aiding our study of the possible causes of climate variability and climate change. In addition, coupling the instrument-based and reconstructed climate records provides a valuable perspective on the present climate. And it appears reasonable to assume that *what has happened in the past could happen again in the future*.

The validity of climate reconstruction rests on a systematic analysis and synthesis of a complex and diverse array of data. Information on climate prior to the era of instrument records consists of a montage of contributions from many disciplines including palynology, paleontology, glaciology, and geology. This information consists of biotic and abiotic response records, that is, climate-sensitive time series. To be most useful in climate reconstruction, these heterogeneous data sets must be converted into records of compatible climate elements, such as mean annual temperature and precipitation.

Climate reconstruction requires identification of a link between quantitative climate forcing and environmental response. This link is formulated by calibrating the record of modern environmental response against modern instrument-derived climate data and then applying it to the ancient (fossil) record. This procedure assumes *methodological uniformitarianism*; that is, factors that are ecologically limiting today were limiting in the same way in the past. For example, Bergthórsson developed a calibration between historic records of the duration and extent of drift ice off the coast of Iceland with climatic data for a recent 150-year period. Applying this modern calibration to documentary records of drift ice conditions in the past, Bergthórsson was able to reconstruct the decadal mean annual temperature in Iceland dating back to almost CE 900 (Figure 9.4). The credibility of Bergthórsson's model is supported by the close association between relatively cold episodes and the occurrence of famine years in Iceland as documented by Thoroddsen.

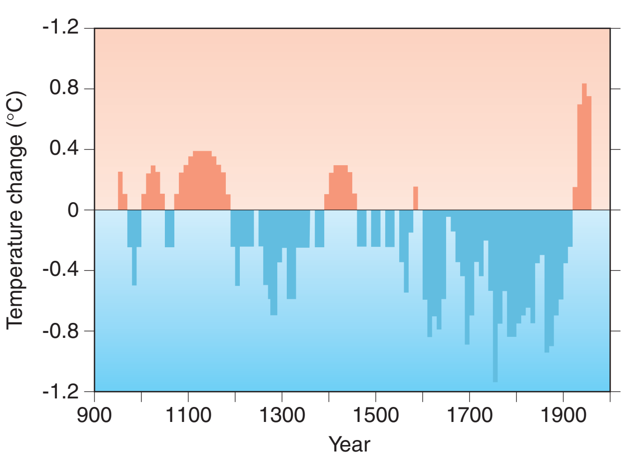


FIGURE 9.4
Reconstructed record of the decadal mean annual temperature for Iceland over the recent millennium based on historical accounts of the duration and extent of sea ice along the shore. [Data from Bergthórsson]

Proxy Climate Data Sources

For times and places where no instrument-derived record of climate exists, past climate information may be inferred from various sensors that substitute for actual weather instruments. These sensors of climate, known as **proxy climate data sources**, include historical documents, tree growth rings, pollen profiles, deep-sea sediment cores, speleothems, corals, and glacial ice cores. No one type of proxy alone is sufficient to enable scientists to reconstruct broad scale patterns of climate. For a summary of the overall limitations of proxy climate data sources, refer to this chapter's first Essay.

HISTORICAL DOCUMENTS

Under cautious scrutiny, certain historical documents archived in libraries and museums can yield a wealth of information on past climates. Personal diaries, almanacs, old newspapers, and mariner's log books may yield qualitative and some quantitative references to weather and climate. Other types of documents refer only indirectly to weather and climate but can be useful nonetheless. Records of success of grain harvests, quality of wine, or various phenological phenomena (such as dates of blooming of plants in spring) provide indirect indications of growing season weather. For example, in his book *Times of Feast, Times of Famine* (1971), Emmanuel Le Roy Ladurie relies heavily on vineyard records to reconstruct the climate of Western Europe during the Middle Ages. More recently, researchers at the University of Bern reconstructed summer weather/climate patterns for parts of Switzerland based on records of grape harvests dating as far back as the late 15th century. The growth of grapevines and the ripening of the fruit strongly respond to average April through August temperatures.

Caution must be exercised in inferring climate information from historical documents because many factors, in addition to weather and climate, usually influence such records. For example, aside from growing season weather, harvest dates in vineyards are affected by fluctuations in the wine market. Usually, the authors of such documents had no intention of chronicling the weather or climate and it is also important to bear in mind that people have long applied ingenuity to moderate the impact of climate—particularly extremes in climate. (Refer, for example, to the second Essay in Chapter 4 on agricultural freeze protection strategies.) Hence, climate information derived from written records of human activity is not always reliable and corroborating data from other independent sources are necessary to support these climatic inferences.

TREE GROWTH RINGS

Analysis of variations in the thickness and density of annual growth rings of certain tree species can yield detailed information on past climates. The study of tree growth rings for climate data is known as **dendroclimatology**. Andrew E. Douglass (1867-1962), a solar astronomer, pioneered this work in the American Southwest between 1894 and 1901 while at Lowell Observatory, a private non-profit research institution in Flagstaff, AZ. In 1937, he founded the Laboratory of Tree-Ring Research at the University of Arizona. Today, his successors at the Laboratory are reconstructing past climates using computers programmed with special statistical techniques.

At the onset of the growing season in spring, plant tissue located immediately beneath tree bark produces relatively large thin-walled wood cells, which give the wood a relatively light appearance. Wood cells produced in summer, however, are thick-walled, giving the wood a darker appearance. A year's growth of spring wood plus summer wood constitutes an annual growth ring, so counting the number of growth rings gives the age of the tree in years. Because the width of growth rings normally decreases as the tree ages, widths are usually expressed in terms of a *tree-growth index*, defined as the ratio of the actual tree growth-ring width to the width expected based on the tree's age. The index is relatively low in stressful growing seasons and high in favorable growing seasons.

Only trees living in subpolar terrestrial regions are useful in dendroclimatic reconstructions and primarily record conditions during the warm season. (Tropical species do not have well-defined seasonal growth rings.)Trees growing near the limits of their range are the most sensitive to climate variability so that their growth rings are the most reliable sensors of climate. A simple hollow drill is used to extract cores from living trees or cut timber (Figure 9.5). Usually cores are taken from many trees at one site, and tree-ring indexes are averaged. In western and southwestern North America, the primary locale for dendroclimatic research, scientists sample ponderosa pine, Douglas fir, and three closely related species of Bristlecone pine. Some of the longest tree ring records are obtained from Great Basin Bristlecone pine (*Pinus longaeva*) that grows at or near the tree line in mountains of Utah, Nevada, and eastern California (Figure 9.6). At up to almost 5000 years old, the Great Basin Bristlecone pine is thought to be one of the oldest known living tree species in the world. Bristlecone pine is characterized by heavy gnarled limbs, a spiked top, and in the case of the oldest specimens, only a narrow ribbon of living bark (cambium).

FIGURE 9.5
A simple hollow drill is used to extract a core from a living tree or cut timber. [Frome Hannes Grobe]

Typically, tree ring chronologies date back some 500 to 700 years but range up to 11,000 years in a few cases. By assiduous matching of tree growth ring records from living trees with those from timbers in prehistoric dwellings, detailed tree ring chronologies are extended back in time thousands of years. This matching technique is known as *cross-dating*.

Although other environmental factors (e.g., soil type, drainage conditions which is accounted for by sampling trees in the same area) can be important, the thickness and density of tree growth rings are especially sensitive to moisture stress and have been used to reconstruct lengthy drought chronologies.

FIGURE 9.6
The long-lived Bristlecone pine growing near the tree line in the White Mountains of California is a valuable source of tree growth ring data for dendroclimatic reconstructions. [Photograph courtesy of Mark A. Wilson, Department of Geology, The College of Wooster, Wooster, OH.]

Tree growth ring analysis has enabled researchers to extend the drought record across large portions of North America and many centuries into the past. In 1998, Edward R. Cook of Columbia University's Lamont-Doherty Earth Observatory and colleagues from Arizona and Arkansas reconstructed drought chronologies across the nation based upon annual tree ring data obtained from a network of 388 climatically sensitive tree ring sites. From these data, time series of summer (June through August) Palmer Drought Severity Index (PDSI) values were determined stretching back to 1700 at 155 grid points. These gridded tree ring chronologies were calibrated with PDSI instrument-based records from selected Historical Climatology Network stations commencing in the late 19th century. Researchers found that the 1930s drought (discussed in the Case-in-Point of Chapter 6) was the most severe drought to impact the nation since 1700.

By 2004, the drought record had been expanded to include 835 tree ring sites, primarily in the West, where precisely dated annual tree ring chronologies were obtained. The new grid covered most of North America with a latitude/longitude spacing of 2.5 degrees. In addition to the 286 grid point PDSI time series, annual contour maps of PDSI were constructed that span much of the continent. This work permitted extension of the spatial and temporal coverage of drought reconstruction not only into Canada and Mexico, but also back 2000 years. From this more extensive data set, researchers produced an online *North American Drought Atlas*. They identified several North American droughts that were even more severe than the 1930s drought. In addition to being more severe, some droughts persisted through several decades, considerably longer than those of the 20th century. A megadrought that occurred in the 16th century, along with another megadrought extending into the early 17th century, may have contributed to the disappearance of the Roanoke Colony on Roanoke Island in Dare County in what is now North Carolina. Intended to be the first English colony in America, the *Lost Colony* vanished in the 1580s.

Tree growth ring records have been used to reconstruct past variations in the flow of the Colorado River prior to 1896 when the gauge-based record began (Chapter 7). The flow of the Colorado River correlates with the mean annual precipitation in the river basin. In addition, the width of tree rings from living trees from many sites on the Colorado Plateau correlates with mean annual precipitation. Hence, the flow of the Colorado River directly correlates with the width of tree rings. Application of this correlation to the prehistoric tree ring record enables scientists to reconstruct the prehistoric flow of the Colorado River.

In 2006, Connie A. Woodhouse, a climatologist at the University of Arizona, and her colleagues used tree ring records to estimate the long-term natural mean annual flow of the Colorado River to be 14.6 MAF. They found that drought can occur suddenly and may persist from several years to a few decades. For example, during the drought of 1844-1848, the mean annual flow was only 9.6 MAF. Evidence was found that a 60-year *megadrought* impacted the Colorado Plateau during the 12th century. Analysis of tree growth rings enabled these scientists to extend the region's drought chronology back to CE 762. They relied on data from both living trees and cross-dating of ring patterns in tree trunks scattered throughout the upper Colorado River drainage basin, where dry conditions preserved them. About half way through the megadrought that lasted from CE 1118 to 1179, flow of the Colorado River remained below the long-term average for 13 consecutive years. By contrast, over the past century, consecutive years of below average flow numbered no more than 5 years.

POLLEN PROFILES

Ponds, peat bogs, marshes, and swamps are favorable sites for the accumulation and preservation of wind-borne pollen. *Pollen* is the tiny dust-like fertilizing component of a seed plant that is dispersed by the wind. Mixing with other sediments (clay, silt and organic particles), pollen grains settle and accumulate in low-lying depositional areas. Upward of 20,000 pollen grains may be mixed in a single cubic centimeter of pond mud. Assuming that the pollen is the product of nearby vegetation and that climate largely governs vegetation types, climate may be inferred from pollen. When climate changes, vegetation changes, and so too do the types of pollen delivered to depositional sites. Hence, changes in the abundance of pollen of different species at various depths within accumulated sediment may provide a record of past climatic regimes and climate change.

Scientists use a corer to extract a sediment column (core), then separate pollen from its host sediments, identify pollen species and frequency of occurrence of each species, and reconstruct the sequence of past changes in vegetation. From the climate requirements of the reconstructed vegetation (based on modern species distribution and modern climate), scientists decipher the sequence of past climate changes.

Pollen is a valuable source of information on the vegetation and climate of the late Pleistocene Ice Age and subsequent Holocene Epoch, especially over the past 15,000 years. Using sophisticated statistical techniques to calibrate climate and pollen, scientists have reconstructed remarkably detailed quantitative climate data. For example, a pollen record (profile) from Kirchner Marsh near Minneapolis, MN, yielded a reconstructed record of variations in July mean temperature and annual precipitation back to 12,000 years ago. Unfortunately, relatively few sites favor the accumulation and preservation of a continuous long-term pollen/climate record. In North America, most such records come from a few geographical areas, including western mountain valleys, the Great Lakes region, and interior New England.

DEEP-SEA SEDIMENT CORES

Cores extracted from sediments that blanket the ocean floor yield a continuous record of sedimentation dating back many hundreds of thousands of years and in some places, millions of years (Figure 9.7). Much of what we know about the climate of the Pleistocene Ice Age is based on analysis of the shell and skeletal remains of microscopic marine organisms found in deep-sea sediment cores. Identification of the environmental requirements of these organisms, plus oxygen isotope analysis of their remains, enables scientists to distinguish between relatively cold and warm climatic episodes of the past.

With **oxygen isotope analysis**, scientists use a special property of water to reconstruct large-scale climate fluctuations of the Pleistocene Ice Age. A water molecule (H_2O) is composed of one of two stable isotopes of oxygen, ^{16}O or ^{18}O. (*Isotopes* consist of atoms that are chemically identical, with the same number of protons but different numbers of neutrons in the nucleus.) In the Earth system, the lighter isotope (^{16}O) is much more abundant than the heavier isotope (^{18}O); only one ^{18}O exists for every thousand or so ^{16}O. Nonetheless, small but significant variations occur in the ratio of light oxygen to heavy oxygen circulating in the global water cycle, which have important implications for past fluctuations in glacial ice volume.

On average at a particular temperature, water molecules containing the lighter ^{16}O isotope move slightly faster than water molecules containing the heavier ^{18}O isotope and therefore evaporate more readily. Hence, water vapor is enriched with light oxygen compared to heavy oxygen and the amount of ^{16}O compared to ^{18}O is greater in cloud particles and precipitation versus liquid water on Earth's surface. When it rains or snows, the ^{16}O returns to the ocean, replenishing the ocean's supply of light oxygen and maintaining a relatively constant average ratio of light to heavy oxygen. However, geographical variations in the oxygen isotope ratio of seawater arise because of differences in precipitation amounts and evaporation rates. Seawater has more ^{18}O at subtropical latitudes, where evaporation exceeds precipitation, and less in middle latitudes, where rainfall is greater.

A B C

FIGURE 9.7
Sediment cores extracted from beneath the ocean floor provide valuable information on the geologic and climatic past. (A) A hollow pipe lined with plastic tubing and coupled to a weight at the top is lowered over the side of a ship. When within about 8 m of the bottom, the corer free-falls into the sediment as a piston suctions sediment into the tube. The coring device is recovered and the sediment core is removed and split lengthwise for analysis. (B) USGS crew on the research vessel *G.K. Gilbert* collect a 20-ft sediment core, using an electric coring system and hydraulic crane. [USGS] (C) Core racks holding a total of 72,000 m of sediment cores at the Deep-Sea Sample Repository of Lamont-Doherty Earth Observatory in Palisades, NY. [Courtesy of the Lamont-Doherty Earth Observatory]

During a climatic episode that favors the formation or growth of a glacier, snow that accumulates on land converts to ice. Heavy water molecules condense and precipitate slightly more readily than light water molecules. Moisture plumes moving from the tropics to high latitudes lose heavy oxygen along the way, so snow falling at high latitudes has less ^{18}O than rain falling in the tropics. The result is that growing glacial ice sheets sequester more and more light oxygen, while ocean water has less and less. With a shift to an interglacial climate, ice sheets shrink and meltwater rich in ^{16}O drains back into the ocean, increasing the ratio of ^{16}O to ^{18}O.

Organic sediments that accumulate on the ocean floor record fluctuations in the oxygen isotope ratio of seawater. Marine organisms, such as foraminifera living in the sunlit surface waters, build their shells from calcium carbonate ($CaCO_3$) that is dissolved in seawater. Shells formed during warmer interglacial climatic episodes contain more light oxygen than those formed during colder glacial climatic episodes. When these organisms die, their shells settle to the ocean floor and mix with other marine sediments. With specially outfitted deep-sea drilling ships, scientists extract cores from an undisturbed sequence of ocean bottom sediments. In the laboratory, the core is split open, and shells are extracted and analyzed for their oxygen isotope ratio. The youngest sediments are at the top of the core and the oldest sediments at the bottom. Variations in oxygen isotope ratio document changes in the planet's glacial ice volume, a measure of past changes in temperature. The proportion of light to heavy oxygen in ocean water decreases with increasing glacial ice volume.

Oxygen isotope analysis of deep-sea sediment cores indicates that the Pleistocene Ice Age (1.7 million to 10,500 years ago) was punctuated by numerous abrupt changes between glacial and interglacial climatic episodes. Oxygen isotope analysis has also been applied to ice layers within cores extracted from the Greenland and Antarctic ice sheets. These analyses confirm the abrupt change behavior of climate dating back hundreds of thousands of years.

SPELEOTHEMS

A **speleothem**, also called *dripstone*, is a calcite ($CaCO_3$) deposit in a limestone cave or cavern that can yield high-resolution records of past temperature and rainfall. A speleothem forms when calcite precipitates from groundwater that seeps into a cave and can either build downward from the roof of the cave creating a stalactite, or grow upward from the floor of the cave to form a stalagmite.

Climate is reconstructed using oxygen isotope analysis from samples of calcite extracted from a speleothem. As noted above, scientists measure the ratio of two isotopes of oxygen from the sample: ^{16}O (light oxygen) and ^{18}O (heavy oxygen). The ratio of these isotopes is sensitive to temperature and rainfall. The age of the sample is derived independently using the ratio of uranium-234 to thorium-230, a very precise radiometric technique that determines the age of a speleothem sample with a resolution ranging from a year to decades. A potential radiometric dating technique, involving the decay of uranium to lead, may extend speleothem-derived climate records back 600,000 years. The Case-in-Point that opens Chapter 7 summarizes how a speleothem from a cave in northern China was used to reconstruct monsoon rainfall over a recent 1800-year period.

CORALS

A **coral reef** is among the most spectacular biological features of the ocean. Primarily built by carbonate-secreting colonial animals, these slow growing structures can be centuries old, and may be so large that they are visible from space (e.g., Great Barrier Reef along Australia's East Coast). In many parts of the tropical ocean, reefs stand hundreds of meters above the sea bottom and can extend hundreds of kilometers along the shoreline or cap extinct undersea volcanoes. They consist of thin veneers of living organisms growing on older layers of dead coral (limestone) or volcanic rock, and are bound together by layers of calcareous algae.

Worldwide, coral reefs provide shelter and food for up to 9 million species of marine life (about one-third of all known marine species) including fish, invertebrates, and plants. Coral reefs are among Earth's most productive habitats, ranking second to rain forests in biodiversity; most are in the tropical Pacific and Indian Oceans between about 30 degrees N and 30 degrees S. Reef-building corals prefer waters with an average annual temperature of 23 °C to 25 °C (73 °F to 77 °F) and most corals cannot tolerate prolonged exposure to high or low temperatures or to large fluctuations in temperature. For this reason, even small changes in sea-surface temperatures—such as associated with large-scale climate change—threaten coral reefs (Chapter 12). Corals also require clear water and are endangered by sediment runoff from land, oil spills, and other forms of water pollution. Excess nutrient input washed from the land can stimulate the growth of algae on the surface of coral reefs, smothering coral polyps (a process known as *eutrophication*).

Each type of coral animal builds a characteristic structure that is conspicuous on reef surfaces. Some corals (e.g., brain corals) form robust compact structures; others build delicate, complex branching forms. Many corals have growth rings, much like trees, that can be used to reconstruct past variations in maritime conditions and climate in the tropics and subtropics. Through oxygen isotope analysis and measurement of chemical species ratios, scientists acquire climate data including information on SST and past El Niño and La Niña events. Corals provide a resolution typically measured in months (weeks in rare cases) and the time range is up to about 400 years.

Edward R. Cook and colleagues at the Lamont-Doherty Earth Observatory relied on climate data inferred from coral and tree growth rings to demonstrate a link between exceptionally long-lived La Niña episodes in the tropical Pacific Ocean and persistent droughts in the American Southwest. Analysis of coral from the central Pacific revealed that SST were much lower than normal from 1855 to 1863. This SST anomaly was interpreted to be a manifestation of an exceptionally long-lived La Niña and also coincided with the driest decade in Texas since 1700 based on tree growth ring studies. The drought that began in 1855 along the edge of the Great Plains was more intense than the Dust Bowl drought of the 1930s. Other widespread droughts that corresponded to persistent La Niña conditions occurred in the Southwest from 1703 to 1709 and 1818 to 1824.

GLACIAL ICE CORES

To better understand Earth's climate system and climate change as a basis for predicting the climate future, scientists are collecting and analyzing proxy climate data from the Pleistocene Ice Age. Ice cores extracted from the Antarctic and Greenland ice sheets are important sources of information on climate change, as well as the chemical composition of air during the Pleistocene Ice Age (Figure 9.8).

In 1988, Soviet and French scientists reported on their analysis of a 2200 m (7200 ft) ice core extracted at Vostok station on the East Antarctic ice sheet. The ice core spanned 160,000 years. Oxygen isotope analysis yielded a temperature record, and chemical analysis of trapped air bubbles revealed trends in the greenhouse gases carbon dioxide and methane. During the summers of 1991-93, two independent scientific teams, one American and the other European, drilled into the thickest portion of the Greenland ice sheet. The two drill sites were located within 30 km (19 mi) of each other, about 650 km (404 mi) north of the Arctic Circle. Both cores were about 3000 m (9840 ft) in length and spanned a time interval of roughly 200,000 years. In the mid-1990s, drilling at Vostok recovered a 3100 m (10,170 ft) ice core spanning the past 425,000 years. In 2004, the *European Project for Ice Coring in Antarctica (EPICA)* extracted an ice core from East Antarctica representing a time interval of 740,000 years. By 2008, the recovered ice core encompassed about 800,000 years.

To derive air temperature from glacial ice cores, scientists use either oxygen isotope analysis (described above) or analysis of the ratio of deuterium to hydrogen in ice. The water molecule is composed of two different isotopes of hydrogen: ^{1}H and ^{2}H. ^{1}H consists of one proton and no neutrons whereas ^{2}H (also called deuterium or D) has one proton and one neutron. Isotopic ratios are compared to that of *standard mean ocean water (SMOW)*. Compared to SMOW, glacial ice cores contain slightly less of the heavier oxygen and deuterium isotopes. As the temperature falls, there is comparatively less and less ^{18}O and D.

In addition to reconstructing temperature, glacial ice cores are also analyzed for changes in the composition of the atmosphere, especially greenhouse gases. The present level of atmospheric carbon dioxide is about 27% higher than the highest levels detected in air bubbles trapped in glacial ice cores dating back 650,000 years. In addition, the reconstructed temperature record closely parallels the concentration of the greenhouse gases carbon dioxide and methane. This prompts the question as to whether greenhouse gas concentration drives climate or climate drives greenhouse gas concentration. It appears that with a slow change in atmospheric CO_2 concentration

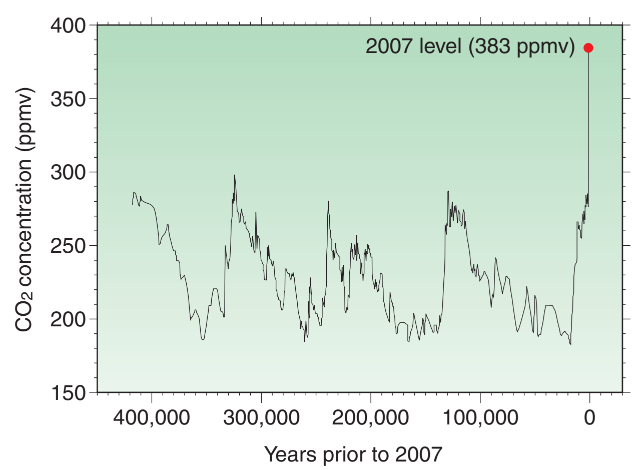


FIGURE 9.8
The concentration of atmospheric carbon dioxide in parts per million by volume (ppmv) from about 425 thousand years ago to 2007 based on glacial ice core analysis from the Vostok station in Antarctica. [Scripps Institution of Oceanography, University of California, San Diego]

over a long period, temperature drives CO_2 concentration by altering the global carbon cycle. On the other hand, with a rapid change in atmospheric CO_2 concentration over a short period (e.g., since the beginning of the Industrial Revolution), CO_2 appears to drive temperature.

Where snow accumulation rates are relatively high, glacial ice cores can be used to resolve short-term climate fluctuations of decades or less. As noted later in this chapter, cores extracted from the Greenland ice sheet reveal major abrupt oscillations in climate over periods ranging from millennia down to decades that are superimposed on much longer-term gradual climate cycles (i.e., Milankovitch cycles, Chapter 11).

STRATIGRAPHY AND GEOMORPHOLOGY

Bedrock, particularly sedimentary strata, is a widespread source of generalized information on environmental conditions of the geologic past, including climate. As noted in Chapter 1, internal and surface geological processes continually shape and reshape Earth's lithosphere. These same processes produce a multitude of rock types that compose the crust. Rocks, in turn, are made up of one or more minerals. (A *mineral* is a naturally occurring inorganic solid characterized by an orderly internal arrangement of atoms with fixed physical and chemical properties). Rocks composing the crust are classified as igneous, sedimentary, or metamorphic based on the general environmental conditions in which the rock formed.

Cooling and crystallization of hot molten *magma* produces **igneous rock**. Magma originates in the lower portion of the lithosphere or upper mantle, and migrates upward towards the Earth's surface (Figure 1.12). Magma may remain within the crust and cool slowly, forming coarse-grained igneous rock such as granite, or it may spew onto Earth's surface as *lava* through vents or fractures in bedrock and solidify rapidly, forming fine-grained igneous rock such as basalt or glassy material such as obsidian. Igneous rock-forming processes (e.g., volcanism) affect the chemical composition of the atmosphere and may contribute to climate change (Chapter 11).

Sedimentary rock may be composed of any one or a combination of compacted and cemented fragments of rock and mineral grains, partially decomposed remains of plants and animals (e.g. shells, skeletons), or minerals precipitated from solution. Sediments form as rocks undergo weathering (physical disintegration and chemical decomposition) when exposed to rain, atmospheric gases, and fluctuating temperatures at or near Earth's surface. Sediments are washed into rivers that transport them to the sea, and other standing bodies of water, where they settle out of suspension, accumulate on the bottom, and eventually compact into layers of solid sedimentary rock. Sediments are also transported and deposited by wind, glaciers, and icebergs at sea. Most sedimentary rocks have a granular texture; they are composed of individual grains that are compressed or cemented together (e.g., sandstone, shale), although some consist of precipitated minerals and are crystalline (e.g., limestone, rock salt).

Unless disturbed by tectonic forces, sedimentary rock occurs as a sequence of layers (strata) having a horizontal orientation (at least initially) with the youngest layer at the top of the sequence and the oldest layer at the bottom (Figure 9.9). By interpreting the environmental conditions represented by each layer (based on factors such as mineral composition, grain size distribution, fossil fauna or flora), geoscientists can infer environmental conditions (including climate) and change in general terms through hundreds of millions of years. In many cases, however, sediment deposition was not continuous, and portions of the environmental record are missing because of erosion or non-deposition of sediment. A hiatus in the sedimentary record may span many millions of years. So, like a history book with missing pages, this environment (and climate) record is incomplete.

Consider a few examples of climate inferences drawn from sedimentary strata. Layers of rock salt indicate a warm, dry climate that favored net evaporation of sea water and crystallization of salt. *Till* is the general term for sediments deposited by the direct action of advancing glacial ice without the influence of running water. Sediments are angular and occur in a broad size range (poorly sorted). Through lithification, till becomes *tillite*, a sedimentary rock that is evidence of ancient ice ages and glacial climates.

FIGURE 9.9
Exposure of fossil-rich sedimentary bedrock (limestone) around 300 million years old at a road cut in northwestern Missouri.

Like many sedimentary rocks, **metamorphic rock** is derived from other rocks. A rock is metamorphosed (changed in form) when exposed to high pressure, intense heat, and chemically active fluids—conditions that exist in geologically active mountain belts. Like igneous rocks, metamorphic rocks are crystalline; that is, they are composed of crystals that interlock like the pieces of a jigsaw puzzle. Marble is a common metamorphic rock formed by the metamorphism of limestone ($CaCO_3$) and quartzite is a very durable metamorphic rock formed by metamorphism of sandstone (mostly SiO_2).

Most of the bedrock composing Earth's crust is igneous with some metamorphic rock locally or regionally. In many places, thick layers of sedimentary rock and unconsolidated sediments overlie crystalline igneous and metamorphic rocks. Unconsolidated sediments include soils (on land) and clay, silt, sand, or gravel (on land and ocean bottom). Deposits of sediment vary widely in thickness, from a thin veneer to thousands of meters.

Geomorphology is the scientific study of landforms on Earth's surface and the processes responsible for their formation. In some cases, climate plays a major role in these processes such that some generalized information can be inferred from landforms regarding past climates. As noted later in this chapter, many different landforms are the consequence of glacial activity such that a distinction can be made between climate regimes that favor the thickening and expansion of glaciers (*glacial climates*) and climate regimes that favor the melting of glaciers (*interglacial climates*).

In another example of climate reconstruction based on geomorphology, large fields of active (mobilized) sand dunes usually occur where the climate is arid. According to the *United States Geological Survey (USGS)*, moisture balance and the amount of vegetation cover are more important than wind in controlling whether dunes are stable or active. Aerial photographs from 1936 and 1938 identified mobilized dunes in parts of North Dakota, Colorado, Kansas, Oklahoma, Texas, and New Mexico. From analysis of photos and climate data, scientists concluded that loss of vegetation cover (and its anchoring ability) because of higher temperatures and reduced rainfall was responsible for dune mobilization during the 1930s Dust Bowl.

VARVES

A **varve** is a thin layer (lamina) of sediment deposited annually in a body of still water, usually a lake fed by a stream. The stream may or may not drain a melting glacier, but the climate must be sufficiently cold for ice to cover the lake surface during winter. In the case of a glacial meltwater stream, rapid melting of glacial ice during the warm summer months delivers to the water body relatively coarse-grained, light-colored sediment (sand or silt). This sediment settles to the bottom and constitutes the summer band of a varve. In winter, lower temperatures slow the discharge of meltwater and ice covers the lake. Suspended sediment slowly settles out of the still water to the lake bottom. Consequently, the summer layer grades upward to the thinner winter layer consisting of very fine sediment (clay), often dark in color because of the presence of organic particles. Hence, a single varve consists of one light band and one dark band.

If undisturbed, a sequence of varves may provide a high-resolution record of variations in the annual mass budget of a glacier. A coring device is used to extract a sediment core from the lake bottom. From a reconstructed mass budget record, past climate information may be inferred.

Climates of Geologic Time

For convenience of study, the geologic past plus its climate record is subdivided based on the **geologic time scale**, a standard division of Earth history into eons, eras, periods, and epochs based on large-scale geological events (Figure 9.10). Throughout most of the approximately 4.5 billion years that constitute *geologic time*, information on climate is unreliable and descriptions of climate are highly generalized. Accounts of early climates often are suspect because of lengthy gaps in the proxy climate records and difficulties in determining the timing of events and correlating events that occurred in widely separated locations. As pointed out earlier in this chapter, in many areas lengthy episodes of erosion or non-deposition of sediment, sometimes spanning hundreds of millions of years, mean no bedrock for reconstruction of past climates. Furthermore, *plate tectonics* complicates climate reconstruction efforts that focus on periods of hundreds of millions of years (Figure 9.11). Nonetheless, the available evidence supports some conclusions regarding the climate over geologic time.

Geologic evidence points to an interval of extreme climate fluctuations about 570 million years ago, corresponding to the transition between Proterozoic and Phanerozoic Eons. In southern Africa, along Namibia's Skeleton Coast (South Atlantic Ocean), rock layers that formed in tropical seas directly overlie glacial deposits. Geoscientists interpret these rock sequences as indicating abrupt changes in climate between extreme cold and

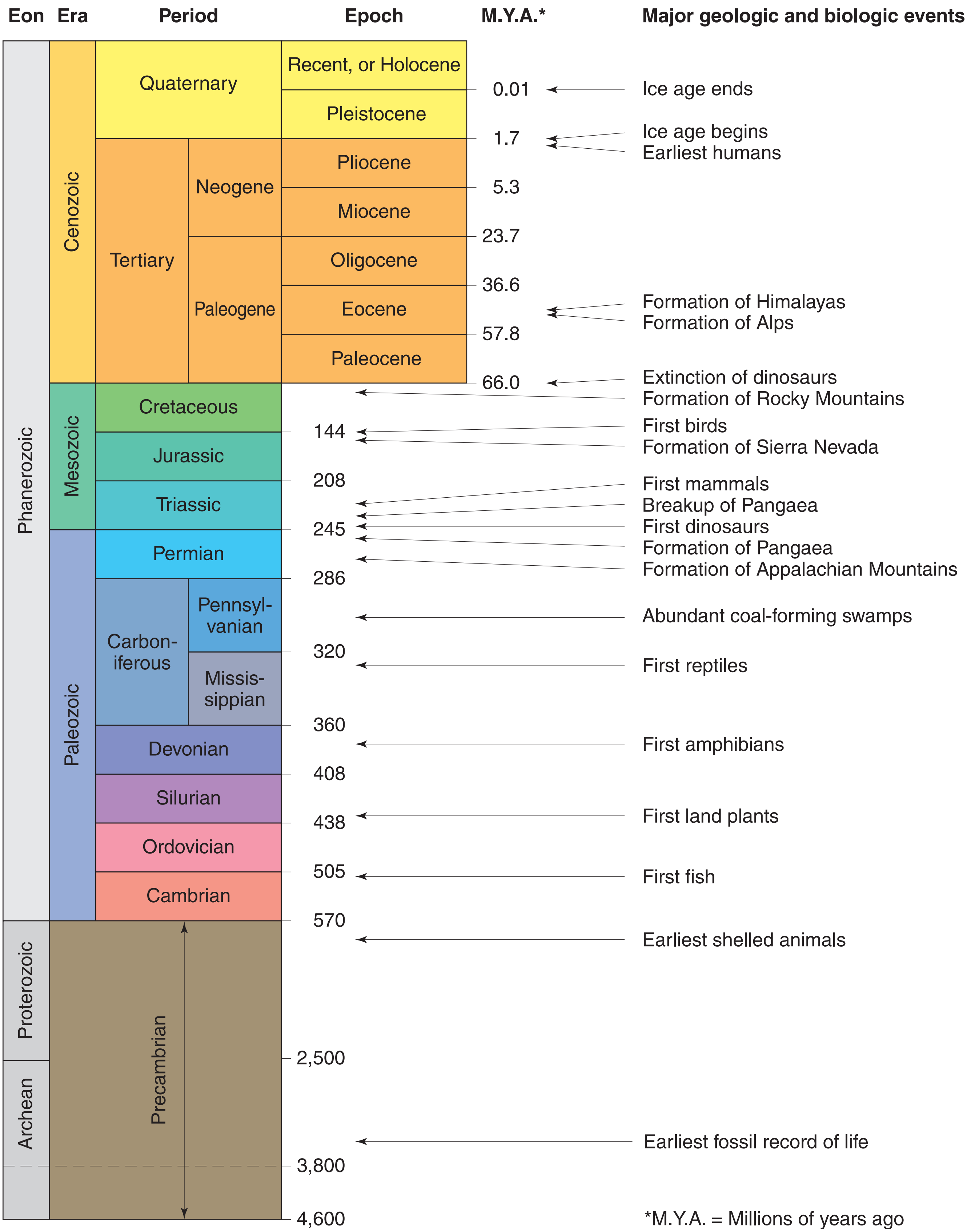


FIGURE 9.10
Geologic time scale.

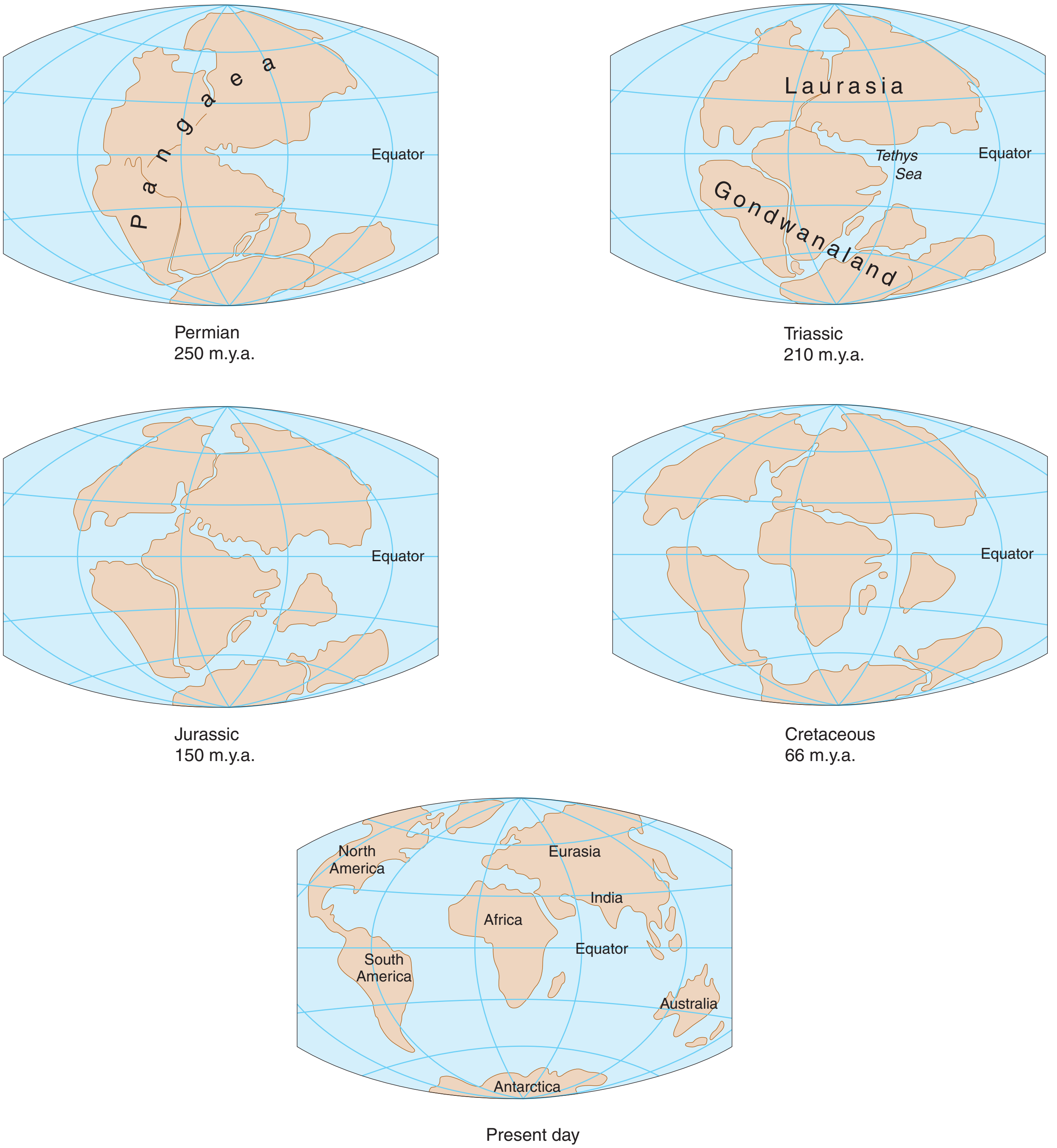


FIGURE 9.11
About 200 million years ago, the super-continent Pangaea began to split apart into separate continents that slowly drifted apart. Ocean basins opened and eventually the continents reached their present positions. Continental drift, a manifestation of plate tectonics, is responsible for climate changes operating over hundreds of millions of years. [U.S. Geological Survey]

tropical heat. During as many as four cold episodes, each lasting perhaps 10 million years, the continents were encased in glacial ice and the ocean froze to a depth of more than 1000 m (3300 ft). At the close of each cold episode, temperatures rose rapidly, and within only a few centuries, all the ice melted.

As noted in the second Essay of Chapter 1, five major mass extinctions occurred over the past 550 million years. Drastic changes in Earth's environment during each of these episodes eliminated up to 50% or more of all plant and animal species then in existence. Four of the five mass extinctions (at the close of the Ordovician, Devonian, Permian, and Triassic periods) were linked to a combination of chemical and circulation changes in the ocean, coupled with global warming due to an enhanced greenhouse effect.

The Mesozoic Era, from about 245 million to 70 million years ago, was characterized by a generally warm Earth free of large glacial ice sheets. At the boundary between the Triassic and Jurassic periods, the global mean temperature rose perhaps 3 to 4 Celsius degrees (5.5 to 7 Fahrenheit degrees), contributing to a major extinction or displacement of many animal and plant species. At peak warming during the Cretaceous Period, the global mean temperature was perhaps 6 to 8 Celsius degrees (11 to 14 Fahrenheit degrees) higher than now. Subtropical plants and animals lived as far north as 60 degrees N and dinosaurs roamed what is now the North Slope of Alaska.

A fifth mass extinction at the end of the Cretaceous period, about 65 million years ago, was likely caused by an asteroid impact that threw huge quantities of dust into the atmosphere, blocking sunlight, and causing cooling that contributed to the demise of the dinosaurs. An explosion in the population of mammals followed the extinction of the dinosaurs.

Great fluctuations in climate have characterized the present Cenozoic Era. About 55 million years ago, near the transition between the Paleocene and Eocene Epochs, the concentration of atmospheric greenhouse gases (either methane or carbon dioxide) increased dramatically. Scientists propose that these gases were released to ocean waters during massive intrusions of basaltic magma into carbonaceous marine sediments during tectonic events related to the opening of the North Atlantic Ocean. Some scientists propose that methane was released from gas hydrates. The gases escaped the ocean waters to the atmosphere, enhancing the greenhouse effect, and causing an already warm planet to become even warmer. Over a period of only a few thousand years, global temperatures climbed 5 to 10 Celsius degrees (9 to 18 Fahrenheit degrees).

By 40 million years ago, however, Earth's climate began shifting toward colder, drier, and more variable conditions, setting the stage for the Pleistocene Ice Age—an epoch of numerous major glacial advances and recessions that commenced approximately 1.7 million years ago. According to W.F. Ruddiman of the University of Virginia and J.E. Kutzbach of the University of Wisconsin-Madison, mountain building may explain this change in Earth's climate, specifically the rise of the Colorado Plateau, Tibetan Plateau, and Himalayan Mountains. Prominent mountain ranges influence the geographical distribution of clouds and precipitation and can alter the planetary-scale circulation (Chapters 5 and 6). Furthermore, mountain building may alter the global carbon cycle. Enhanced weathering of bedrock exposed in mountain ranges sequesters more atmospheric carbon dioxide in sediments thereby weakening the natural greenhouse effect.

In the American West, the region from the California Sierras to the Rockies, known as the Colorado Plateau, has an average elevation of 1500 to 2500 m (5000 to 8200 ft). Although mountain building began about 40 million years ago, about half of the total uplift took place between 10 and 5 million years ago. The Tibetan Plateau and Himalayan Mountains of southern Asia cover an area of more than 2 million km^2 (0.8 million mi^2) and have an average elevation of more than 4500 m (14,700 ft). About half of total Himalayan uplift took place over the past 10 million years. The plateaus diverted the planetary-scale westerlies into a more meridional pattern, increasing the north-south exchange of air masses and altering the climate over a broad region of the globe. Also, seasonal heating and cooling of the plateaus cause low pressure to develop in summer and high pressure in winter, enhancing the monsoon circulation over southern Asia (Chapter 7).

In 2008, Dennis V. Kent, a geoscientist at Rutgers University, and colleagues proposed an alternate tectonics-based explanation for the cooling trend that ultimately led to the formation of Antarctica's ice sheets and the Pleistocene Ice Age. Following the break-up of the supercontinent Pangaea, the tectonic plate carrying the Indian subcontinent moved northward and about 50 million years ago slammed into Asia. Prior to this collision, a million-year period of volcanic activity along Asia's southern border generated some 4 million km^3 of basaltic lava derived from carbonate-rich sediments on the sea floor. As a consequence, the level of atmospheric CO_2 increased to more than 1000 ppmv and temperatures rose worldwide. The collision, however, cut off the supply of carbonate sediments and, at the same time, weathering and erosion of rock exposed on the Indian subcontinent caused a gradual decline in atmospheric CO_2 to perhaps 300 ppmv by 30 million years ago. According to Kent and colleagues, the consequent weakening of the greenhouse effect would have been responsible for long-term cooling.

Climates of the Pleistocene Ice Age

Climate varies over a broad spectrum of time scales so that viewing the climate record of the past two million years in progressively narrower time frames is useful. Such an approach helps to resolve climate change into more detailed fluctuations, especially over the recent past. During the past two million years, plate tectonics has not been a major factor in climate change. For example, assuming a spreading rate of 10 cm (3.9 in.) per year, in two million years the Atlantic basin would spread a total distance of 200 km (124 mi), not very significant in a broad climatic sense. For all practical purposes, over the past two million years mountain ranges, continents, and ocean basins were essentially as they are today.

Compared to the climate that prevailed through most of geologic time, the climate of the last two million years was also unusual in favoring the development of huge glacial ice sheets (although evidence also exists of ice ages earlier in geologic time). During much of Earth's history, the average global temperature may have been 10 Celsius degrees (18 Fahrenheit degrees) higher than it was over the past two million years.

HISTORICAL PERSPECTIVE

The Swiss naturalist Louis Agassiz (1807-1873) is credited with championing the **glacial theory** during the mid-1800s. Previously, the prevailing view—as espoused by the Scottish geologist Charles Lyell (1797-1875)—attributed erratic boulders observed in Europe to rafting by icebergs during a massive flood. *Erratics* are large boulders that differ geologically from the local bedrock, implying transport from elsewhere—in some cases over distances of hundreds of kilometers. Interestingly, Swiss peasants long thought that alpine glaciers were once more extensive and interpreted erratics, polished, striated (scratched), and grooved rock surfaces, and other landscape features as indicators of ancient glaciation. At the invitation of Jean de Charpentier (1786-1855), a German-Swiss geologist and early proponent of the glacial theory, Agassiz had an opportunity to examine such evidence near Bex, Switzerland, in the summer of 1836. Agassiz quickly became convinced of the glacial origins of the landscape features and set out to develop a glacial theory.

At scientific meetings and in his writings, Agassiz argued that some time in the past (a period he called an *ice age*), large glacial ice sheets spread over Europe as far south as the Mediterranean and over much of northern Asia and North America. At first, his ideas met with considerable skepticism by the scientific establishment. However, after 1840, as more geologists had an opportunity to examine the field evidence, an increasing number of prominent scientists converted to the glacial theory. By 1865, the theory had gained widespread acceptance on both sides of the Atlantic. A major reason for the initial reluctance to accept the glacial theory was the fact that most geologists were unfamiliar with glaciers, having never observed them or their impacts in the field. Furthermore, scientific expeditions did not discover that Greenland's glaciers formed an ice sheet until 1852 and the size of the Antarctic ice sheets was not determined until the late 19th century.

During the 20th century much field and laboratory work was directed at working out the chronology and causes of the climate and glacial fluctuations of the Pleistocene Ice Age. In the early 1880s, the eminent geologist Thomas C. Chamberlin (1843-1928) proposed that the Ice Age had involved multiple advances and recessions of glacial ice sheets. Chamberlin's insight on the Ice Age came from his many years of field work in formerly glaciated southeastern Wisconsin. Believing that an ice sheet had advanced and receded over North America four times, geologists subdivided the Pleistocene Epoch into four glacial stages: from oldest to youngest, designated the Nebraskan, Kansan, Illinoian, and Wisconsinan, named for the states where the best evidence for glacial advance is present. European geologists had identified corresponding glaciations in the Alps that they called the Günz, Mindel, Riss, and Würm. Analysis of deep-sea sediment cores, however, reveals perhaps 12 to 15 major advances and recessions of glacial ice at periodic intervals. Nonetheless, division of the Pleistocene into the four glacial stages is still used today.

More than three decades ago, the *Climate Long-Range Investigation, Mapping and Prediction (CLIMAP)* project completed a global-scale reconstruction of sea-surface temperatures during the last glacial maximum. Although lacking many of the climate reconstruction techniques available today, this international, multidisciplinary enterprise was the first to reconstruct changes in SST by latitude and note the greater sensitivity of high latitudes to climate change. Today, an expanded and improved version of CLIMAP, known as *Multiproxy Approach for the Reconstruction of the Glacial Ocean Surface (MARGO)* is underway. Using a variety of temperature proxies and greater spatial resolution of data, an international team intends to develop an updated reconstruction of SST and sea-ice extent during the last glacial maximum.

GLACIERS

A **glacier** is a mass of ice that flows internally under the influence of gravity (Figure 9.12). Glacial ice is the second largest reservoir in the global water cycle and the single largest freshwater reservoir on Earth (Chapter 5). Like rivers of liquid water, glaciers flow toward the ocean as part of the runoff component of the global water cycle. Some glaciers partially or completely melt prior to reaching the sea and the meltwater supplies rivers, lakes, and groundwater. Other glaciers enter the ocean directly where they break up into icebergs that float out to sea and eventually melt. At present, glacial ice covers about 10% of Earth's land area. Major ice sheets, up to 3 km (1.5 mi) thick, cover most of Antarctica and Greenland while much smaller glaciers occupy some mountain valleys at lower latitudes and even at the equator. At times during the Pleistocene Ice Age, glacial ice may have covered as much as 30% of the planet's land area, but through most of Earth's 4.6 billion years, little glacial ice existed.

Epochs of extensive glaciation occurred prior to the Pleistocene Ice Age although supporting evidence is scarce because of subsequent erosion. Late in the Paleozoic Era, during the Caboniferous and Permian periods (360 to 250 million years ago), widespread glaciation occurred in what is now southern Africa, South America, India, Australia, and Antarctica. During the Ordovician and Silurian periods of the Paleozoic Era (450 to 420 million years ago), glaciation affected part of South America. During the early Paleozoic (500 million years ago), glaciers developed in Africa. At the end of the Precambrian (570 million years ago), glaciation occurred in various continents of the Northern Hemisphere. Glaciation may have occurred several times during the Precambrian, including the Huronian Ice Age, about 2.4 to 2.1 billion years ago.

FIGURE 9.12
A glacier is a mass of ice that flows internally under the influence of gravity. Climate change can affect the energy (heat) and mass balance causing the glacier to expand or shrink. This glacial ice stream is located on Fitz Roy Mountain in southern Patagonia, near the Chilean border. [Photo courtesy of Paul Sager, Professor Emeritus, University of Wisconsin-Green Bay.]

Ultimately, large-scale climate change is responsible for variations in Earth's glacial ice cover. And now changes in glacial ice or sea ice cover, in turn, may be influencing or hastening climate change.

Snow is the raw material for the formation of glacial ice. A glacier only develops where the climate is such that snowfall exceeds snowmelt over a succession of many years. Accumulating snow compacts under its own weight and gradually transforms into tiny granular spheres of ice (called *firn*) which eventually re-crystallize into solid ice. As snow converts to firn and then ice, its physical properties change. For one, density increases from an average value of about 0.1 g per cm^3 for fresh-fallen snow to about 0.9 g per cm^3 for ice. Spherical granules of ice pack more closely than plate-like snowflakes, increasing the density. As more and more snow accumulates, confining pressure increases and firn granules re-crystallize into fewer but larger crystals that interlock like the pieces of a jigsaw puzzle. The pressure at a depth of about 80 m (260 ft) is sufficient to convert densely packed snow and firn into essentially impermeable ice thereby trapping air bubbles within.

A distinction is made between temperate and polar glaciers. In a **temperate glacier**, the internal ice temperature rises to the *pressure-melting point* sometime during the year (most likely by late summer); the confining pressure slightly depresses the melting point of ice. Meltwater produced in a temperate glacier accelerates the transition of snow to ice and increases the flow rate of the glacier by reducing frictional resistance. This water accelerates compaction of the ice mass and seeps into open spaces where it refreezes, further increasing the density. Most mountain glaciers such as those in the Rockies are temperate. On the other hand, ice temperature within a **polar glacier** remains below the pressure melting point throughout the year so that little or no meltwater is generated. Conversion of snow to ice and the internal motion of ice are extremely slow processes in a polar glacier. The Antarctic ice sheets are polar glaciers.

The greater supply of meltwater in temperate versus polar glaciers largely explains why temperate glaciers form more rapidly than polar glaciers, but also have a shorter life expectancy. The greater time constant of a polar glacier means that it responds more slowly to climate change.

CLIMATE AND GLACIERS

The climate system governs the exchange of both heat energy and ice mass between a glacier and its environment. For a glacier to exist, the relationship between heat input and heat output must maintain water in the solid phase for an extended period of time. Heat flows between a glacier and its surroundings chiefly through radiation (solar and terrestrial infrared), conduction, and phase changes of water at the glacier-atmosphere interface. If radiational heating exceeds radiational cooling, then ice temperature rises; if radiational cooling exceeds radiational heating, ice temperature falls. If the air temperature were higher than the ice surface temperature, then heat would be conducted to the ice; if the air temperature were lower than the ice surface temperature, then heat would be conducted from the ice to the atmosphere. Deposition or condensation of water vapor onto the ice surface and freezing of water are phase changes that release latent heat to the ice (warming). Evaporation of water, sublimation of ice, and melting of ice absorb latent heat from the ice (cooling). In addition, geothermal heat may be conducted to the ice from the underlying bedrock.

Ice mass is added to a glacier principally through precipitation and avalanches (in mountain valleys). Ice mass is removed from a glacier primarily via melting (including drainage of meltwater away from the glacier), sublimation, and wind erosion of snow. As a glacier enters the ocean or a large lake, ice bergs break off at its leading edge; this process of ice loss is known as *calving*. Recall from the first Essay in Chapter 3 that shrinkage of the ice cap on Mount Kilimanjaro is primarily due to enhanced sublimation coupled with dry conditions and reduced snowfall rather than a warming trend. The general term for all processes that result in a loss of glacial ice mass is **ablation**.

The difference between ice mass gain (accumulation) and ice mass loss (ablation) over the course of a year (usually measured at the end of the melt season in late summer) is the glacier's **mass balance**, that is,

Mass balance = Mass gain – Mass loss.

If a glacier gains more ice mass than it loses, its mass balance is positive and the glacier thickens and expands. If a glacier loses more ice than it gains, its mass balance is negative and the glacier thins and shrinks. A **glacial climate** favors a positive mass balance so that new glaciers form and existing glaciers advance. An **interglacial climate** favors a negative mass balance so that glaciers fail to form or existing glaciers retreat and eventually may completely waste away.

During the Pleistocene Ice Age, the climate shifted numerous times between glacial climates and interglacial climates. The square wave in Figure 9.13A models the behavior of climate during the Pleistocene; typically, the climate shifts abruptly between interglacial and glacial climatic episodes. The sinusoidal wave in Figure 9.13B models the response of glacial ice volume to large-scale shifts between glacial and interglacial climates. During a glacial climatic episode, ice volume increases slowly at first. But after an ice sheet reaches a critical size, it begins to influence atmospheric conditions in ways that promote preservation of ice or growth of the ice sheet (an example of *positive feedback*). The surface of a snow covered ice sheet strongly reflects sunlight. Less solar radiation is absorbed (converted to heat) lowering the temperature and reducing the amount of melting, especially in summer. But as the growing ice sheet expands into lower latitudes, its growth rate slows in response to the more intense solar radiation. With a shift to an interglacial climatic episode, the ice sheet slowly begins to shrink. As the ice warms to its pressure melting point, melting accelerates. The rate of shrinkage slows, however, as the ice sheet melts back into higher latitudes where solar radiation is less intense and air temperatures are lower.

Comparison of Figure 9.13A with Figure 9.13B reveals that abundant glacial ice may exist during an interglacial climatic episode. The unusually great amount of latent heat of fusion of ice is another reason for the stability of a glacial ice sheet. The geologic record confirms

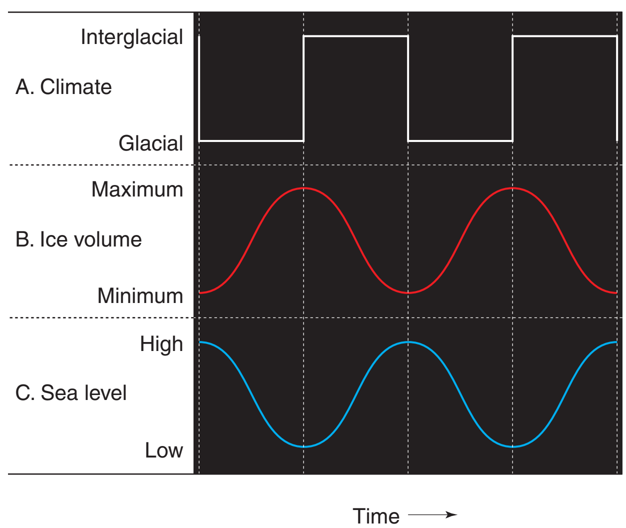


FIGURE 9.13
A graphical model showing (A) the variation of climate between glacial and interglacial episodes, (B) the response of glacial ice volume, and (C) the response of mean sea level. [Modified after R.A. Bryson, W.M. Wendland, and J.M. Moran, University of Wisconsin-Madison]

that climate change tends to be abrupt whereas ice sheets respond sluggishly to changes in climate; that is, it takes time for huge masses of ice to grow or melt away. On the other hand, small glaciers have relatively short time constants, are less stable, and melt faster.

Variations in the volume of glacial ice on Earth have important implications for mean sea level. As noted in Chapter 5, the total amount of water on the planet is essentially constant, a reasonable assumption throughout at least recent geologic time. Almost all the water locked in glaciers is drawn from the ocean as part of the global water cycle, so as the volume of glacial ice increases, the volume of water in the ocean basins decreases and sea level falls. Conversely, as glaciers melt, the volume of water in the ocean basins increases and sea level rises. This is an example of **eustatic sea-level change**, a worldwide fluctuation in sea level. Figure 9.13C models the response of mean sea level to changes in glacial ice volume. Note that the model eustatic sea-level curve is 180 degrees out of phase with the glacial ice volume curve.

Ocean water temperature also affects sea level, because water expands when heated and contracts when cooled. During the last glacial maximum (about 20,000 to 18,000 years ago) when Earth's glacial ice cover was about as extensive as anytime during the Pleistocene, mean sea level was 113-135 m (370-443 ft) lower than it is today. Roughly 10 m (33 ft) of this decline in sea level was due to greater water density in response to lower temperatures. The drop in sea level exposed portions of the continental shelf, including a land bridge linking Siberia and North America. If all the glacial ice currently on the planet were to melt, sea level would likely rise some 70 m (230 ft).

GLACIERS AND LANDSCAPES

The varied landscapes of the formerly glaciated regions of Canada, northern United States, and northwestern Europe are part of the legacy of the Pleistocene Ice Age. Lakes (including the Great Lakes), marshes, and swamps developed in lowlands excavated by advancing lobes of glacial ice. Meltwater streams draining shrinking glaciers deposited extensive layers of sand and gravel (called *outwash*) over broad areas. The fertile soils of the grain belt developed from dusty sediment, called *loess*, derived from sediment deposited first by glacial meltwater, and then transported and deposited by Ice Age winds. For more on loess and climate, refer to this chapter's second Essay.

During major glacial climatic episodes of the Pleistocene Epoch, the Laurentide ice sheet formed over what is now east-central Canada and spread westward to the Rocky Mountains, eastward to the Atlantic, and southward over the northern tier states of the United States (Figure 9.14). At about the same time, mountain

FIGURE 9.14
Glacial ice cover over North America about 20,000 to 18,000 years ago, the time of the last glacial maximum.

FIGURE 9.15
This glacial erratic near Washburn Observatory on the University of Wisconsin-Madison campus is called Chamberlin Rock, named for Thomas C. Chamberlin (1843-1928), a highly regarded glacial geologist.

glaciers in the Rockies coalesced into the Cordilleran ice sheet, and a relatively thin ice sheet spread over the Arctic Archipelago. In northwest Europe, the Fennoscandian ice sheet, much smaller than the Laurentide, covered what are now the United Kingdom, Norway, Sweden, and Finland. The Laurentide and Finnoscandian ice sheets thinned and retreated, and may even have disappeared entirely, during relatively mild interglacial climatic episodes, which typically lasted about 10,000 years. Throughout these interglacials, glacial ice cover persisted over most of Antarctica and Greenland as it still does today.

The way a glacier impacts a landscape depends to a large extent on whether the glacier is advancing (positive mass balance) or retreating (negative mass balance). During a glacial climatic episode, advancing glaciers can be formidable agents of erosion. Glaciers are more competent than rivers of water, and can incorporate and transport much larger sediments. Glacial ice transported and dumped huge angular boulders, some the size of cars, which today are strewn over parts of Wisconsin, Michigan, New York, and New England (Figure 9.15). Armed with rock debris at their base, advancing lobes of glacial ice scratch and scrape underlying surfaces to sculpture the landscape. Broad U-shaped mountain valleys and grooves excavated in bedrock attest to the great erosive power of glacial ice (Figure 9.16). Elsewhere, glaciers plastered broad areas with poorly sorted sediment (known as *ground moraines*).

During an interglacial climatic episode, glaciers stagnate, melt, and recede. Sediments are released from ice during melting and may be transported and deposited by meltwater streams. Many streams that drain melting glaciers are choked with sediment. Sediment deposition creates various landforms, including deltas, eskers, and kames, all of which are composed of mostly sand and gravel. An *esker* is a long steep-sided sinuous ridge composed of sediment deposited by a meltwater stream flowing through a tunnel within the ice (Figure 9.17). A *kame* is a cone-shaped hill of sediment formed when a meltwater stream flowing on the surface of a mass of stagnant ice plunges into a crevasse. The sediment accumulates at the base of the crevasse similar to the sand at the bottom of an hourglass (Figure 9.18). The presence of any of these glacial landforms today points to a time in the past when the climate was quite different from what it is today.

FIGURE 9.16
Armed with rock debris at its base, an advancing lobe of glacial ice sculptured this exposure of igneous bedrock into a streamlined form.

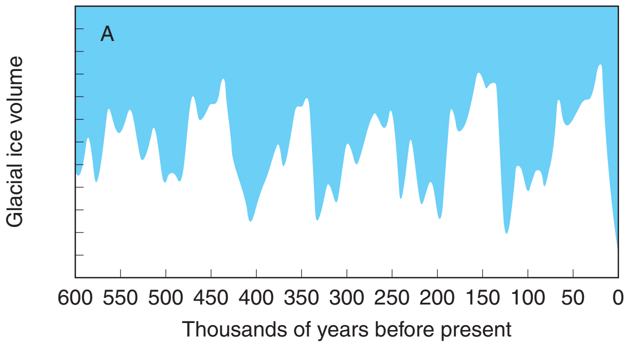
FIGURE 9.17
Vertical cross-section through a tunnel in an accumulation of sand and gravel deposited by a meltwater stream flowing through a mass of stagnant glacial ice. Viewed from above, the landform, known as an esker, is a long narrow steep-sided sinuous ridge that winds tens of kilometers through the countryside.

FIGURE 9.18
A cone-shaped hill of sand and gravel, known as a kame, formed when a meltwater stream plunged into a crevasse in a mass of stagnant glacial ice.

CHRONOLOGY AND TEMPERATURE TRENDS

During the Pleistocene, glacial ice sheets advanced and receded numerous times. The leading edge of a glacial ice sheet consisted of a series of lobes that sometimes moved independently of its neighboring lobes. Glacial lobes act like huge erasers as they advance across the landscape, modifying, and even eradicating, much of the evidence from prior glacial advances and recessions. Hence, geoscientists know the most about the more recent major fluctuation of Earth's glacial ice cover because it left the last imprint. To find out more about earlier glaciations, scientists rely on other techniques.

Oxygen isotope analysis of deep-sea sediment cores shows numerous fluctuations between major glacial and interglacial climatic episodes over the past 600,000 years (Figure 9.19A). Shifting focus to the past 160,000 years, resolution of the climate record improves. The temperature curve in Figure 9.19B is based on analysis of an ice core extracted from the Antarctic ice sheet at Vostok. A relatively mild interglacial episode, referred to as the *Eemian*, began about 127,000 years ago and persisted for nearly 7000 years. In some localities, temperatures may have been 1 to 2 Celsius degrees (2 to 4 Fahrenheit degrees) higher than during the warmest portion of the present interglacial. The Eemian interglacial was followed by



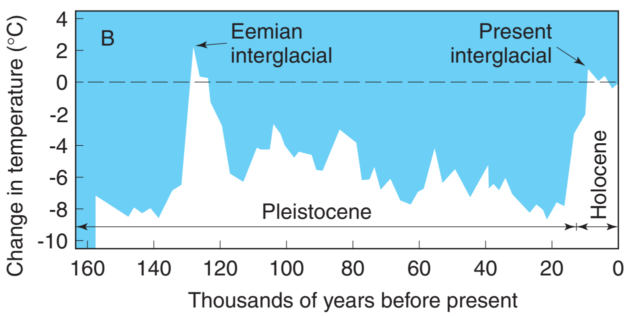


FIGURE 9.19
Reconstructed records of (A) the variation in global glacial ice volume over the past 600,000 years based on analysis of oxygen isotope ratio of shells in deep-sea sediment cores, and (B) temperature variation over the past 160,000 years based on oxygen isotope analysis of an ice core extracted from the Antarctic ice sheet at Vostok and expressed as a departure in Celsius degrees from the 1900 mean global temperature. [Compiled by R.S. Bradley and J.A. Eddy from J. Jousel et al., *Nature* 329(1987):403-408 and reported in *EarthQuest* 5, No. 1 (1991).]

numerous fluctuations between glacial and interglacial climatic episodes.

The last major glacial climatic episode began about 27,000 years ago and reached its peak about 20,000 to 18,000 years ago (Figure 9.14). At that time, glacial ice 3 km (5.1 mi) thick covered most of New England and the Great Lakes states, reaching as far south as the Ohio River Valley and Long Island, NY. About 18,000 years ago, the large-scale climate shifted to dominantly interglacial. Although occasionally interrupted by relatively brief shifts back to glacial climatic regimes, the Laurentide ice sheet gradually melted back, until it finally withdrew from what is now the coterminous northern United States about 10,500 years ago (Figure 9.20). Not until 5500 years ago had most of the residual Laurentide ice in northern Canada melted.

During glacial climatic episodes, temperatures were lower than today, but the magnitude of cooling was not the same everywhere. A variety of geologic evidence indicates that during the Pleistocene, temperature fluctuations between major glacial and interglacial climatic episodes typically amounted to as much as 5 Celsius degrees (9 Fahrenheit degrees) in the tropics, 6 to 8 Celsius degrees (11 to 14 Fahrenheit degrees) at middle latitudes, and 10 Celsius degrees (18 Fahrenheit degrees) or more at high latitudes. An increase in the magnitude of a temperature change with increasing latitude is known as **polar amplification**, indicating that polar areas are more sensitive to climate change.

Glacial ice cores from both Greenland and Antarctica clearly reveal an approximately 100,000 year Ice Age cycle consisting of cold glacial climatic episodes (e.g., the Wisconsinan stage) sandwiched between mild interglacial climatic episodes of approximately 10,800-year average duration. Perhaps 16 of these long-term cycles operated over the span of the Pleistocene Epoch. Evidence from deep-sea sediment cores indicates that regular variations in Earth-Sun geometry (the Milankovitch cycles) drive this approximately 100,000-year glacial/interglacial cycle (Chapter 11).

Greenland and Antarctic ice-core records correlate well both in terms of magnitude of temperature change and the timing of events suggesting that the 100,000-year Ice Age cycles were globally synchronous. However, comparison of the Greenland and Antarctic ice core data over the most recent Ice Age cycle (i.e., from about 142,000 years ago to 10,500 years ago) reveals marked differences between the Southern and Northern Hemispheres. Whereas the Antarctic record is reasonably smooth and "calm," the Greenland record shows numerous abrupt and drastic flip-flops between glacial and interglacial climatic episodes. Temperatures changed as much as 7 Celsius degrees (12.6 Fahrenheit degrees) over periods of decades or less (in some cases in only 3 years.) These abrupt temperature

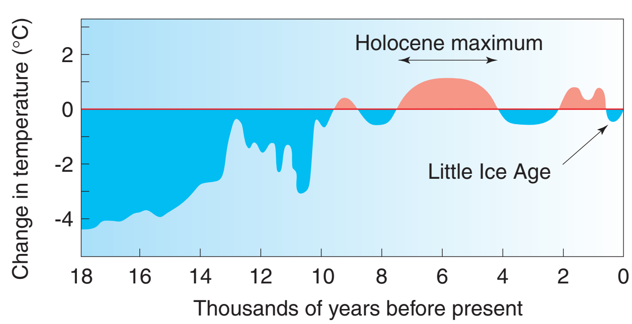


FIGURE 9.20
Reconstructed temperature variation over the past 18,000 years based on analysis of a variety of proxy climate indicators and expressed as a departure in Celsius degrees from the 1900 mean global temperature. [Compiled by R.S. Bradley and J.A. Eddy based on J.T. Houghton et al. (eds.), *Climate Change: The IPCC Assessment*, Cambridge University Press, U.K., 1990, and reported in *EarthQuest* 5, No. 1 (1991).]

changes, having two basic periodic components of 2000 to 3000 years and 7000 to 12,000 years, occurred during the Wisconsinan stage of the Pleistocene Epoch but not during the subsequent Holocene Epoch. The periods of these temperature fluctuations are much shorter than those of the Milankovitch cycles and hence are probably unrelated to changes in Earth-Sun geometry.

The most likely explanation for these short-term changes in temperature is the alternate weakening and strengthening of the ocean's meridional overturning circulation (Chapter 4). This may explain, for example, the occurrence of the relatively cool episode from 11,000 to 10,000 years ago, known as the **Younger Dryas** (named for the polar wildflower *Dryas octopetala* that reappeared in portions of Europe at the time). The return of glacial climatic conditions triggered short-lived re-advances of remnant ice sheets in North America, Scotland, and Scandinavia.

The Younger Dryas began abruptly when glacial ice lobes disrupted drainage patterns, diverting meltwater from the Mississippi River into the St. Lawrence River and North Atlantic. With this input of fresh water, North Atlantic surface waters became less saline and eventually were not sufficiently dense to sink and form North Atlantic Deep Water (NADW). This weakened the meridional overturning circulation, which in turn diminished the warm water flowing into the central and northern North Atlantic, causing a marked cooling of the surrounding lands. The Younger Dryas ended just as abruptly as it began when the input of fresh water into the North Atlantic decreased and formation of NADW resumed. The geographic pattern of the Younger Dryas climatic impacts (e.g., little in western North America and only a muted response in the Antarctic ice core record) suggests that the Younger Dryas was not part of the larger ice age variability driven by the Milankovitch cycles. Rather, the Younger Dryas was a regional short-term climate fluctuation linked to changes in the Atlantic meridional overturning circulation.

Atlantic Ocean circulation changes that may have triggered the Younger Dryas likely also occurred at other times. Scientists have interpreted certain layers of lithogenous sediment in cores extracted from the floor of the North Atlantic as materials released during melting of fleets of icebergs. These icebergs surged or slid off glaciated North America and floated out onto the North Atlantic every 2000 to 3000 years as the climate flip-flopped between warm and cold episodes. Melting icebergs freshened North Atlantic surface waters thereby weakening the meridional overturning circulation. With the return of colder conditions and fewer icebergs, freshening of the surface waters ceased, the water became salty again due to wind-driven evaporation, and the thermohaline circulation strengthened. After two or three of these events, an even greater discharge of icebergs occurred at intervals of 7000 to 12,000 years.

The smaller, shorter, more frequent occurrences are called *Dansgaard-Oeschger* or *D-O events*, named for the Danish paleoclimatologist Willi Dansgaard and the Swiss climatologist Hans Oeschger (1927-1998). They are also referred to as "flickers" because of their relatively short period. Each D-O event consisted of abrupt warming (almost to interglacial levels) followed by gradual cooling. The Greenland ice core record contains evidence of some 23 Dansgaard-Oeschger events between 110,000 and 15,000 years ago. The larger, longer, and less frequent episodes, known as *Heinrich events*, are discussed in the second Essay of Chapter 8. Flickers and Heinrich events occurred during both glacial and interglacial regimes and are evident in the temperature record reconstructed from Greenland ice cores.

Climates of the Holocene

Glacial ice finally withdrew from the North American Great Lakes region about 10,500 years ago ushering in the present interglacial, the **Holocene Epoch**, when civilization and agriculture developed. Although the Laurentide ice sheet was melting and disappeared almost entirely 5500 years ago, the Holocene was an epoch of spatially and temporally variable temperature and precipitation. Cores extracted from the Greenland ice sheet and sediment cores taken from the bottom of the North Atlantic Ocean reveal that the overall post-glacial warming trend was interrupted by abrupt millennial-scale fluctuations in climate.

Post-glacial warming during the Holocene gave way to a notably cold episode about 8200 years ago. The likely reason for this abrupt climate change was a sudden influx of fresh water into the North Atlantic when the ice dam holding back Lake Agassiz burst. Lake Agassiz was centered over south-central Canada and contained the meltwaters of the shrinking Laurentide ice sheet (a volume of water greater than the present-day Great Lakes combined). The arrival of fresh water in the North Atlantic suppressed the meridonal overturning circulaton causing regional cooling. A similar mechanism was responsible for the Younger Dryas described earlier.

Significant cooling also occurred from about 3100 to 2400 years ago. On the other hand, at times during the mid-Holocene (classically known as the *Hypsithermal*), mean annual global temperature was perhaps 1 Celsius degree (2 Fahrenheit degrees) higher than it was in 1900, the warmest in more than 110,000 years, that is, since the Eemian interglacial. A pollen-based climate reconstruction indicates that 6000 years ago, July mean temperatures were about 2 Celsius degrees (3.6 Fahrenheit degrees) higher than now over most of Europe.

In 2001, V.J. Polyak and Y. Asmerom of the University of New Mexico reported on their analysis of speleothems (stalagmites) recovered from Carlsbad Cavern and Hidden Cave in the Guadalupe Mountains of southeast New Mexico. The speleothems span a 4000-year interval through the late Holocene and provide annual resolution. In this area, the mid-Holocene (about 5600 years ago) was characterized by drier conditions than at present. From 4000 to 3000 years ago, the climate was similar to present, but from 3000 to 800 years ago, it became cooler and wetter. Subsequently, the present-day climate prevailed with the exception of a wetter period from 440 to 290 years ago.

Across North Africa, one of the most dramatic environmental changes of the past 10,000 years began during the mid-Holocene. Recall from the Case-in-Point in Chapter 5 that a long-term decline in rainfall caused North Africa to transition from a green savanna to the world's largest warm desert, the Sahara.

Climates of the Recent Millennium

A generalized Northern Hemisphere temperature curve for a recent 1000- year period, derived mostly from historical documents, is shown as Figure 9.21. The most notable features of this record are (1) the Medieval Warm

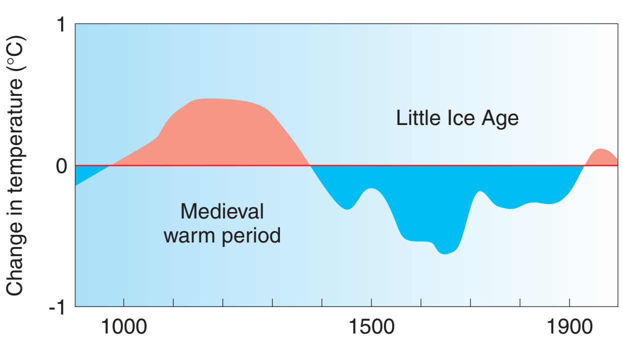


FIGURE 9.21
Reconstructed temperature variation of the Northern Hemisphere over the past 1000 years based on analysis of historical documents and expressed as a departure in Celsius degrees from the 1900 mean global temperature. [Adapted from J.T. Houghton et al. (eds.), *Climate Change: The IPCC Assessment*, Cambridge University Press, U.K., 1990, p. 202.]

Period from about CE 950 to 1250 and (2) the cooling that followed, from about CE 1400 to 1900, known as the Little Ice Age. The Medieval Warm Period and the Little Ice Age were not episodes of sustained warming and cooling, respectively. On the contrary, climate reconstructions based on tree growth rings, sediment cores, and glacial ice cores plus historical documents point to significant decadal-scale fluctuations in temperature and precipitation. The first Norse settlements in Greenland were established during the Medieval Warm Period and failed during the Little Ice Age. (For details on the Norse settlements in Greenland, refer to the second Essay in Chapter 2.)

MEDIEVAL WARM PERIOD

In 1965, the British climatologist Hubert H. Lamb (1913-1997) was the first to characterize the High Medieval (CE 1100 to 1200) as an episode of relatively mild winters and warm dry summers in Western Europe. During this relatively stable climate, harvests were generally bountiful and vineyards thrived in the British Isles. Apparently, a westerly (zonal) circulation pattern prevailed during winter whereas high pressure systems dominated during summer. Lamb estimated that average temperatures were 1 to 2 Celsius degrees (2 to 4 Fahrenheit degrees) higher than normal, based on anecdotal information plus proxy climate data from Western Europe. However, he had no data that would support a global-scale warming trend during Medieval time.

According to paleoclimatologist R.S. Bradley of the University of Massachusetts and colleagues, establishing the magnitude and spatial and temporal extent of warming during the Medieval Warm Period is problematic partially because of scarce observational data (especially from the tropics and Southern Hemisphere) and because Medieval time can encompass a relatively broad interval from CE 500 to 1500. Bradley notes that during the High Medieval, temperatures were higher than during the subsequent Little Ice Age. Large-scale climate reconstructions show that High Medieval mean temperatures were about the same as in the 1901 to 1970 period. During the following three decades, the mean temperature rose by about 0.35 Celsius degree (0.63 Fahrenheit degree). Bradley concludes that the available evidence does not support the notion that the High Medieval was equally as warm or warmer than the late 20th century.

While the Medieval Warm Period is usually defined in terms of positive temperature anomalies, drought was more notable in some regions of the globe. In 2004, Edward R. Cook of Lamont-Doherty Earth Observatory and colleagues reported evidence of long-term "elevated aridity" in the American West during the period CE 900 to 1300. This overlaps with the Medieval Warm Period and is based on a drought index derived from tree growth ring records spanning 600 to 1200 years. As discussed in the Case-in-Point of this chapter, a succession of megadroughts in the 12th and 13th centuries was likely a major reason why the Anasazi and Fremont peoples abandoned their homelands in the American Southwest.

In a 2006 report, V. Sridhar of the University of Nebraska and colleagues found that sand dunes, covering 7500 km^2 of the Nebraska Sand Hills, were last mobilized (indicating arid conditions) about 800 to 1000 years ago, during the early phase of the Medieval Warm Period. Today, these dunes are anchored in place (stabilized) by grasses that are supported by moisture transported in spring and early summer from the Gulf of Mexico on south to southeast winds. Reconstruction of prevailing winds based on dune characteristics indicates that the dunes were last mobilized when the prevailing spring-summer winds had shifted to a southwest direction. This flow descended from the Rockies and Colorado Plateau, undergoing adiabatic compressional warming and drying.

LITTLE ICE AGE

Cooling that heralded the Little Ice Age may have begun as early as CE 1200 in Greenland and the Arctic but closer to CE 1300 at lower latitudes. (François Matthes (1874-1948), a Dutch-American glacial geologist, was first to use the term *Little Ice Age* in 1939.) Based on climatic inferences drawn from tree growth rings, glacial ice cores, and historical documents, scientists characterize the Little Ice Age as a multi-century interval when the climate was more variable than the climate before (the Medieval Warm Period) or since. The climates of Europe and many other regions of the world were volatile, abruptly shifting from one extreme to another. In Europe, bitter cold winters, hot summers, drought, and torrential rains punctuated decades of mild winters and warm summers. Overall, the mean annual global temperature during the Little Ice Age was perhaps 0.5 Celsius degree (0.9 Fahrenheit degree) lower than it was in 1900.

Sea-ice cover expanded, mountain glaciers advanced, the growing season became shorter, and erratic harvests caused food scarcity and considerable hardship for many people. The unstable climate of the Little Ice Age contributed to economic, political, and social unrest in Europe. Interestingly, the Little Ice Age also coincided with the Renaissance, the Age of Discovery, the Enlightenment, the American and French Revolutions, and the Industrial Revolution.

Severe winters characterized the 1430s, culminating in the European-wide famine of 1433-1438. By 1440, almost all wine growing had ceased in England. An unusually cold interval from 1590 to 1610 was apparently synchronous at many localities around the globe. Perhaps the coldest periods of the Little Ice Age prevailed from 1670-1710 and during the 1810s. From about 1680 to 1730, the growing season in England was about five weeks shorter than at present and the average number of days with snow on the ground was 20 to 30, compared to today's 2 to 10 days. Recall the discussion of 1816, *the year without a summer*, in the Case-in-Point of Chapter 2. From the 17th century into the mid 19th century, mountain glaciers advanced well beyond their present-day limits in the Alps, Scandinavia, Iceland, Alaska, China, the southern Andes, and New Zealand. In the 19th century, glaciers also thickened and expanded in the Himalayas and Caucasus Mountains.

Conclusions

Our understanding of Earth's climate system benefits from in depth analysis of the climate record. Unfortunately, the reliable instrument-based climate record worldwide is limited to not much more than 140 years. To improve on this situation, scientists have devised many techniques that rely on climate proxy data to reconstruct the climate prior to the instrument era. Sources of climate proxy data range from low-resolution sedimentary strata to high-resolution tree growth rings. Central to this approach is our ability to calibrate climate forcing with environmental response.

A climate record based on proxy climate data may be compromised by many factors. Proxy climate data sources vary in resolution, time control, and spatial continuity. Consequently, our description of climate history becomes more fragmented, generalized, and uncertain with increasing time before present. Nonetheless, the reconstructed climate record provides a valuable perspective on present climate and prospects for future climate, keeping in mind the axiom, *if it has happened before, it could happen again*.

The portion of the climate record that is derived from measurements by weather instruments usually merits more confidence than the reconstructed climate record. However, many factors influence the integrity of the instrument-based climate record. For example, upgrades in instrument technology, relocation of weather stations from urban to rural sites, and the use of instrument shelters impact the climate record. The next chapter takes a closer look at the instrument-based climate record with special emphasis on what it reveals about recent trends in climate.

Basic Understandings

- Scientists reconstruct climates of the past for many reasons. It appeals to curiosity, adds to our understanding of how the environment responds to climate variability and climate change, aids our study of the nature and causes of climate variability and climate change, and provides a unique perspective on the present climate.
- For information on climate prior to the era of reliable instrument-based records, scientists rely on climate inferences drawn from historical documents, and geological/biological evidence such as bedrock, fossils, pollen, tree growth rings, glacial ice cores, and deep-sea sediment cores. These are known as proxy climate data sources.
- Climate reconstruction requires identification of a link between quantitative climate forcing and environmental response. Such modern calibrations are then applied to the ancient or fossil environmental response record to reveal the past climate.
- Caution is advised when attempting to infer climate information from historical documents because many factors in addition to weather and climate can influence the content of such documents.
- The width of tree growth rings normally decreases as the tree ages. Hence, widths are usually expressed in terms of a tree-growth index, that is, the ratio of the actual tree growth ring width to the width expected based on the tree's age. The index is relatively low in stressful growing seasons and relatively high in favorable growing seasons.
- Assuming that the fossil pollen recovered from a sediment core is the product of nearby vegetation and that climate largely governs vegetation types, climate and climate change can be inferred from analysis of pollen profiles.
- Cores extracted from sediments that blanket almost all the ocean floor yield a continuous record of sedimentation dating back hundreds of thousands to millions of years.
- Oxygen isotope analysis of calcium carbonate shells extracted from deep-sea sediment cores is used to reconstruct large-scale climate fluctuations of the Pleistocene Ice Age. Variations in oxygen isotope ratio ($^{16}O/^{18}O$) correspond to changes in Earth's glacial ice volume and hence, past changes in climate. The proportion of light to heavy oxygen in ocean water decreases with increasing glacial ice volume.
- The ratio of light oxygen to heavy oxygen in a speleothem, a cave deposit precipitated from groundwater, is sensitive to temperature and rainfall.
- Most corals cannot tolerate sea-surface temperatures outside of a relatively narrow range. They can be used as high resolution sensors of SST, El Niño, and La Niña.
- Scientists derive air temperature from glacial ice cores using either oxygen isotope analysis or analysis of the ratio of deuterium (D) to hydrogen in ice. As the temperature falls, there is less and less ^{18}O or D.
- For convenience of study, the geologic past and its climate record is subdivided using the geologic time scale, a standard division of Earth history into eons, eras, periods, and epochs based on large-scale geological events.
- Plate tectonics complicate climate reconstruction of periods spanning hundreds of millions of years. In the context of geologic time, topography and the geographical distribution of continents and the ocean are controls of climate change.
- Geologic evidence points to an interval of extreme climate fluctuations beginning about 570 million years ago, corresponding to the transition between the Precambrian and Phanerozoic Eons. During as many as four cold episodes, each lasting perhaps 10 million years, Earth was encased in glacial ice. At the close of each cold episode, temperatures rose rapidly, and within only a few centuries, all the ice melted. Relatively warm conditions appear to have persisted through much of the Mesozoic Era, from about 245 to 70 million years ago.
- By 40 million years ago, Earth's climate began shifting toward colder, drier, and more variable conditions. Scientists have implicated tectonic forces and the building of the Colorado Plateau, Tibetan Plateau, and Himalayan Mountains as the principal causes of this climate change.
- During the Pleistocene Ice Age that began about 1.7 million years ago, the climate shifted numerous times between glacial climatic episodes (favoring expansion of glaciers) and interglacial climatic episodes (favoring shrinkage of glaciers).

- Climate change during the Pleistocene was geographically non-uniform in magnitude; cooling was greatest at high latitudes and least in the tropics. This latitudinal variation in the magnitude of temperature change is known as polar amplification.
- A glacier is a mass of ice that flows internally under the influence of gravity. Presently, glacial ice covers about 10% of Earth's land area; at times during the Pleistocene Ice Age, ice cover appears to have increased to about 30% of the planet's land area.
- Earth's climate system governs the exchange of both heat energy and ice mass between a glacier and its surroundings. Heat flows between a glacier and its environment via radiation, conduction, and phase changes of water. Ice mass is added to a glacier via precipitation and avalanches. Ice mass is removed from a glacier primarily through melting, sublimation, and wind erosion of snow.
- A glacial climate favors a positive mass balance so that new glaciers form and existing ones advance. An interglacial climate favors a negative mass balance so that glaciers do not form and existing glaciers retreat.
- During glacial climatic episodes, glaciers advance and can be formidable agents of erosion, resulting in features such as bedrock grooves and striations, drumlins, and erratics. However, during an interglacial climatic episode, glaciers stagnate, melt, and recede. Sediment liberated during ice melt is transported and deposited by meltwater streams, building a variety of landforms such as deltas, eskers, and kames.
- Weakening of the meridional overturning circulation in the Atlantic Ocean caused an episode of cooling in the lands surrounding the North Atlantic from about 10,000 to 11,000 years ago. This cool period is known as the Younger Dryas.
- The last major glacial climatic episode began about 27,000 years ago and reached its peak 20,000 to 18,000 years ago when the glacial ice cover over North America was as extensive as it had ever been.
- The Holocene was an epoch of spatially and temporally variable temperature and precipitation.
- A generalized temperature curve for the past 1000 years, derived mostly from historical documents, reveals that the Medieval Warm Period lasted from about CE 950 to 1250 and was followed by the Little Ice Age, from about CE 1400 to 1900. The Medieval Warm Period and the Little Ice Age were not episodes of sustained warming or cooling, respectively. On the contrary, sediment cores and glacial ice cores, as well as historical documents point to significant decadal-scale fluctuations in temperature and precipitation.

Enduring Ideas

- One of the principal reasons why scientists attempt to reconstruct the climate record prior to the instrument era is to gain some perspective on the present climate and how climate might change in the future.
- Proxy climate data sources include historical documents, bedrock, fossil plants and animals, pollen, tree growth rings, glacial ice cores, deep-sea sediment cores, and varves.
- In the context of the hundreds of millions of years that constitute geological time, topographic relief and the geographical distribution of continents and the ocean are important controls of climate change.
- During the Pleistocene Ice Age, the climate shifted numerous times between glacial climatic episodes and interglacial climatic episodes with major implications for the landscape.
- The Holocene was an epoch of spatially and temporally variable temperature and precipitation.

Review

1. Of what value is reconstructing the climate record for the time prior to the era of weather instruments?
2. What is the significance of methodological uniformitarianism in reconstructing the climate past?
3. What two basic assumptions are made when analyzing fossil pollen for the purpose of reconstructing past climates?
4. How does the proportion of light oxygen (^{16}O) to heavy oxygen (^{18}O) in ocean water change as the volume of glacial ice on Earth increases?
5. How are past changes in the chemical composition of air determined from analysis of glacial ice cores?
6. Summarize the principal reasons why the climate during the Paleozoic Era is described in generalized terms.
7. What was a major reason for the slow pace of widespread acceptance of the glacial theory during the 19th century?
8. Compare the physical properties of temperate glaciers to those of polar glaciers.
9. Distinguish between glacial climates and interglacial climates in terms of glacial mass balance.
10. What evidence exists for the abrupt change behavior of glacial/interglacial climates?

Critical Thinking

1. Explain why trees growing near the limits of their range are the most sensitive to climate change and variability.
2. What is the advantage of cross-dating in climate reconstructions based on analysis of tree growth rings?
3. How does plate tectonics affect the reconstruction of climates of geologic time?
4. By about 40 million years ago, Earth's climate began to shift toward colder, drier, and more variable conditions. What might explain this climate change?
5. What is the relationship between large-scale shifts between glacial and interglacial climatic episodes and mean sea level?
6. Explain how it is possible for a great amount of glacial ice to cover the land even though the climate is interglacial.
7. How is the impact of a glacier on the landscape dependent on the sign of the glacier's mass balance?
8. Speculate on the significance of polar amplification for future climate change.
9. How might ocean circulation have played a key role in the occurrence of the relatively cool Younger Dryas?
10. Speculate on any possible parallels between what happened in the American West during the Medieval Warm Period and future warming in the American West.

ESSAY: Limitations of Proxy Climate Data

Reconstructing past climates can be challenging. Because the relationship between climate forcing mechanisms and environmental response can be complex, derivation of a reliable modern calibration between the two can be problematic. Even if the modern climate/environment calibration were reasonable, its application to the documentary or other proxy records may yield erroneous or misleading information. The older the period under study, the more fragmented the proxy record becomes, the sampling interval usually lengthens, resolution of events becomes poorer, and correlating events in widely separated regions becomes more challenging.

The diversity of proxy climate data is a major reason why climate reconstruction becomes less reliable, especially as the record becomes more fragmented. Although climate is the principal ecological control, it is not the only factor affecting ancient biotic or abiotic records. For example, local topography or drainage may influence the tree growth ring record. Furthermore, the relationship between climate forcing mechanisms and environmental response (even when non-climate factors are insignificant), is not always straight forward. Combinations of many climatic elements may be responsible for the environmental response.

Another complication that arises in efforts to reconstruct past climates is *lag response time*. According to *LeChatelier's principle*, a system responds to a change in its environment by shifting in such a way as to relieve the new stress, but different components of ecosystems may take different lengths of time to come into the new equilibrium. Hence, a modern calibration between climate forcing mechanisms and biotic/abiotic response is appropriately applied to the fossil record only when it is a reasonable assumption that an equilibrium condition actually existed.

Problems may also arise from an incomplete understanding of the frequency response mechanism of a fossil climate sensor. For example, some sensors, such as tree growth rings, do not respond significantly to high frequency climate variability but do respond to gradual systematic climate shifts. That is, tree growth rings do not record a single dry summer, but will contain the signature of a succession of several dry summers (drought). A distinction must be drawn between climate data reconstructed from high-resolution sensors (e.g., annually laminated ice cores) that can delineate dominant atmospheric circulation patterns versus low resolution sensors (e.g, sedimentary strata) that primarily mirror long-term average conditions.

If data are available from a high-resolution climate sensor, caution must then be exercised in extrapolating reconstructed climate data from one region to another. Recall from Chapter 2 that climate anomalies are geographically nonuniform in both sign (direction) and magnitude. In response to dominant atmospheric circulation patterns, average values of climatic elements (e.g., temperature, precipitation) calculated for specific weeks, months, or years over broad geographical areas typically show regions of positive anomalies and regions of negative anomalies. Hence, for example, unless corroborated by independent evidence, extrapolating a speleothem-derived temperature record from Missouri to Montana cannot be justified. Even if data are available to generate broad spatial patterns of climate anomalies, the data may not be synchronous. Radiometric dating techniques used to estimate the age of some material or object have many sources of error and generally magnify their effects the further back in time they are applied.

With increasing time before present, the decline in time control and resolution of climate detail means that the most reasonable large-scale climate reconstruction patterns represent lengthy periods of time. During the Holocene, available resolution permits a description of climate in increments of 1000 years. During the Pleistocene Ice Age, climate is generally described over periods of 10,000-yr increments. And in the Paleozoic Era, climate is described in highly generalized terms over hundreds of millions of years.

ESSAY: Pleistocene Climate and Loess Deposition

One of the most important legacies of the Pleistocene Ice Age is *loess* (Figure 1), unconsolidated fine sediment deposited by the wind across broad areas of the central United States, where it is the parent material for some of the most productive soils in the world, helping make possible the grain belt of North America. In the central U.S., loess owes its origin primarily to wind erosion of glacial outwash floodplain deposits. Hence, loess is intimately associated in time and place with continental-scale glacial ice sheets. Meltwater generation is key to floodplain sediment deposition so that loess was deposited primarily during interglacial climate episodes. Loess is also a valuable source of information on past climates including prevailing wind direction and strength (based on trends in loess thickness and particle size) and moisture balance.

FIGURE 1
An exposure of late Pleistocene loess near the Missouri River in Kansas City, MO. Loess consists of unconsolidated sediment, mostly silt-sized, transported by winds from the floodplains of rivers and streams that drained melting glacial ice.

Loess consists of accumulations of unconsolidated, moderately well sorted, and locally thinly layered blankets of sediment. Although the dominant (40% to 50%) grain size is silt, ranging from 0.015 mm to 0.05 mm in diameter, deposits generally include significant clay (5% to 30%) and sand (5% to 30%) fractions. Loess has a texture similar to flour or talcum powder, is porous and permeable, typically yellow to buff in color, and exposed in nearly vertical cliffs (up to 70 degrees). Ancient buried soils (*paleosols*) are often found between layers of loess representing periods of non-loess deposition. Layers of glacial *till* (sediments deposited by the direct action of glacial ice) alternating with layers of loess indicate alternating advances and recession of glacial ice during a rapidly changing climate regime.

Besides the central U.S., loess is widespread in Alaska, China, Russia, northern Europe, and Argentina. In the central U.S., late Pleistocene loess occurs in two convergent belts (Figure 2): one deposit thins eastward from eastern Colorado, Nebraska and Kansas through Iowa, Missouri, Wisconsin, Illinois, and Indiana into southern Ohio. The second belt is east of the Mississippi River and south of the Ohio River from Mississippi through western Tennessee and Kentucky.

FIGURE 2
Major areas of loess distribution (orange) across North America. [Modified after the U.S. Geological Survey]

Loess is a wind (*eolian*) deposit. In the Great Plains and South, the bulk of loess was derived via wind erosion of glacial outwash floodplains and channel bar deposits. Loess thickness decreases systematically with distance from source floodplains, suggesting that loess was a river deposit. However, loess is just as thick on hilltops as in valley bottoms, which confirms deposition by wind. In Nebraska and adjacent regions of Colorado and Kansas, wind erosion of floodplain deposits may have been secondary in importance to wind erosion of the Sand Hills of west-central Nebraska. According to the *United States Geological Survey (USGS)*, erosion of silt-rich poorly consolidated bedrock may also have contributed to loess deposition in the Great Plains. Furthermore, in the narrow periglacial zone along the southern edge of the Laurentide ice sheet, frost heaving and churning of soils probably were minor sources of silt for subsequent re-deposition as loess. The periglacial zone consisted of tree-less tundra underlain sporadically by permanently frozen ground (*permafrost*).

The relationship between loess deposition and Pleistocene climate fluctuations follows from the model presented in Figure 9.13. Elsewhere in this chapter, we noted that a lag time exists between climate forcing and the response of glacial ice volume so that considerable amounts of glacial ice persist long after the climate has shifted from glacial to interglacial. Therefore, the discharge of meltwater also peaks sometime after such a shift. Oxygen isotope analysis of planktonic foraminifera in deep-sea sediment cores extracted from the Gulf of Mexico confirms this lag time for the final disintegration of the Laurentide ice sheet. The influx of Laurentide meltwater reached dramatic proportions by about 15,000 years ago—at least 3000 years after the major shift from a glacial climate episode to a dominantly interglacial climate episode. The discharge of meltwater into the Gulf of Mexico then tapered off rapidly after 13,500 years ago, when the principal drainage shifted away from the Mississippi River to the St. Lawrence River. Furthermore, the most rapid rate of eustatic sea level rise apparently began thousands of years after the last Laurentide glacial maximum.

Fluctuations in climate affect the volume of glacial ice and thereby the discharge of meltwater into rivers and streams draining the ice sheet. Meltwater discharge, in turn, controls floodplain sedimentation and the amount of sediment (alluvium) available for wind erosion and re-deposition elsewhere as loess. (Typically, streams that drain melting glaciers are choked with fine sediment.) If, as suggested by our climate/glacier model (Figure 9.13), meltwater discharge peaks some time after the shift from a glacial to an interglacial climate episode, more alluvium would be available for loess generation during interglacial climate episodes. Regardless of climate, however, as long as glacial ice was present on the North American continent, there was some seasonal ablation so that some loess deposition took place during glacial climate episodes as well. This conclusion is supported by radiocarbon dating of organic matter in loess deposits and *optically stimulated luminescence (OSL)*, a technique that dates the last time the sediment was exposed to sunlight (Chapter 7).