Geomorphology and Climate

13

D. Thompson

Racetrack Playa, an ephemeral lake in Death Valley National Park, California, is known for its enigmatic moving rocks and the tracks they leave behind. One theory suggests that wind moves the rocks over mud slickened by rain, but calculations indicate that only the most extreme winds could overcome the friction of the playa mud. Other theories suggest that wind could move rocks if they are locked in a sheet of ice that buoys them and provides more surface area for wind to act upon.

Introduction

Climate, the long-term average of day-to-day weather, varies dramatically over the surface of our planet and can be described in terms of the temporal distribution and variability in the amount of precipitation, the speed of the wind, the range of temperature, and the relative humidity. Climate strongly influences geomorphology through its effects on the rate and character of atmospheric and surface processes, such as the volume, duration, and type of precipitation (rainfall versus snowfall), runoff, and flood flows, as well as the distribution of vegetation. Together, these climate characteristics set the dimensions of channels, the nature of soils, and the pace and temporal variability of fluvial and hillslope processes, such as sediment transport by mass movements.

Earth's climate is not steady but changes over a wide variety of timescales. On the timescale of plate tectonics, millions of years, large-scale climate zonation reflects the arrangement of continents and resulting oceanic and atmospheric circulation as well as changing rates of volcanism, carbon dioxide (CO_2) emission, carbon sequestration in sediment, and thus atmospheric CO_2 content. On intermediate timescales, thousands of years, changes in Earth's orbit alter the seasonal distribution of incoming solar radiation. Over years to decades, the ash and aerosols that volcanic eruptions spew into the atmosphere cool the planet.

IN THIS CHAPTER

Climate zones have been mapped worldwide, and some landforms, such as moraines and arid-region playa lakes, are directly related to specific climatic conditions. Other landforms, such as sand dunes (see Chapter 10), occur in a wide variety of climatic settings and thus are not climatically diagnostic [**Photograph 13.1**]. For example, sand dunes are common in hyperarid regions, such as the Skeleton Coast of Namibia, because there is little vegetation to anchor noncohesive sand. But, dunes are also found along many humid-temperate beaches and along glacial margins (see Photograph 10.4). The link here is not climate but the ready availability of sand, unsecured by the roots of vegetation.

Geologists use a variety of continuous climate archives, including sediment and ice cores, to decipher the timing and magnitude of past climate changes. Studies of such environmental records have revealed that over the past 2.7 million years, glacial-interglacial cycles have become pronounced, shifting global climate dramatically and repeatedly on 10^4 to 10^5 year timescales. **Relict landforms**, such as glacial cirques, permafrost ice wedge casts, shorelines above modern lake levels, and vegetated dune fields, provide geomorphic evidence for such climatic change. The **last glacial maximum (LGM)** was the peak of the most recent glacial expansion, about 22,000 years ago. At the LGM, Earth's surface was, on average, about 5°C colder than during the warmest time of the **Holocene** (the last 11,700 years). In general, high latitudes cooled the most; the tropics cooled the least.

(a)

National Park Service

(b)

L. Galuzzi

PHOTOGRAPH 13.1 Sand Dunes. Sand dunes can form in a variety of climates because the major control on their presence is a source of mobile sand. (a) Partially vegetated coastal sand dunes along the beach at Channel Islands National Park, Santa Barbara and Ventura counties, California. (b) Sand dunes in Tadrart Acacus, a hyperarid desert area in western Libya, part of the Sahara Desert. Rock outcrops are heavily coated in dark, shiny rock varnish.

The ocean and its currents influence climate dynamics and thus the pace, type, and distribution of geomorphically important Earth surface processes. Usually the oceans buffer the planet from rapid change, but in some cases, oceans can amplify small changes when modes of ocean circulation and heat transport change suddenly. For example, large amounts of glacial meltwater, released as late Pleistocene ice sheets melted away, appear to have capped the North Atlantic with buoyant fresh water and temporarily altered (perhaps for centuries) heat transport patterns in the ocean.

Climate, through its control on the growth and shrinkage of ice sheets over the past several million years, has repeatedly taken sea level up and down by more than 100 m. Thus, rivers flowing to the sea have experienced numerous base-level rises and falls over time (see Figure 7.9). The history of these base-level changes is written in terrace sequences and knickzones throughout the world but can be difficult to decipher because climate, tectonics, and other effects on sediment supply can also catalyze terrace formation (see Figure 7.10). The continental shelves owe their planar form in part to beveling by waves, as the seas repeatedly transgressed and then regressed, and in part to the deposition of sediments during periods of high sea level (see Chapter 8).

There are feedbacks and interactions between solid-Earth processes and climate that affect geomorphology. Surface uplift changes airflow patterns and induces orographic precipitation. Heavy rainfall on the windward slopes of mountain ranges speeds erosion there, causing an isostatic response that moves rock more quickly toward the surface on the windward slopes than on the leeward slopes. Loading of the crust and mantle by ice sheets and greatly expanded pluvial lakes during glacial times results in isostatic compensation. Earth's surface sags under the load only to rebound after the ice retreats or the lakes dry. Glacial erosion is capable of limiting the height of mountain ranges.

Not only does climate affect surface processes, but the rate and distribution of surface processes affect climate. Geomorphic influences on climate include the consumption of CO_2 through weathering of fresh minerals exposed by erosion (see Chapter 3); changes in the global energy balance driven by the extent and **albedo** (reflectivity) of large glaciers, sea ice, and snow-covered land (see Chapter 9); and the recycling of water on a massive scale by plant transpiration (see Chapter 1).

In this chapter, we examine the relationships between geomorphology and climate. First, we consider various geologic and instrumental records documenting changes in climatic variables that drive geomorphic response over a wide variety of temporal and spatial scales. Then we examine the variability and influence of climate on geomorphic processes before discussing the distribution of resulting climate-sensitive landforms over time and space. Finally, we consider both how landforms respond to changes in climate and how the landscape itself can drive climate change.

Records of a Changing Climate

Geomorphologists have explored and interpreted a variety of natural archives to understand and quantify geomorphic changes and the changes in climate that drive them over a variety of timescales. Such archives include lake and marine sediments, glacial ice, and terrestrial sediments such as loess. Some archives, such as lake cores, preserve more or less continuous sedimentation while other archives, such as paleoflood deposits, record individual events. Archives differ in their degree of time-averaging; for example, soils integrate the record of climatically driven pedogenic processes over thousands to tens of thousands of years while a debris flow deposit may result from a single, exceptional rainstorm. As new data reveal the magnitude and effects of human-induced climate change, it becomes increasingly important to understand both the natural range of climate variability as well as past geomorphic responses to climate change.

Landform Records of Climate Change

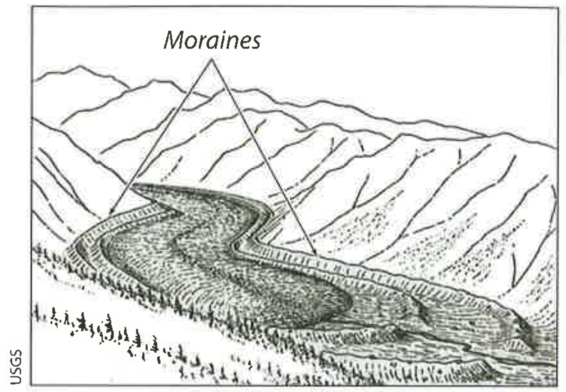
Some landforms are direct or indirect indicators of a changed climate. The challenge lies in dating the landforms and, in some cases, showing that the landforms result from changing climate and not changes in other factors such as tectonics. In much of the world, glacial and periglacial features are the landforms most indicative of a changed climate because they directly reflect changes in temperature and/or precipitation. Periglacial features require mean annual temperatures below freezing. Glacial landforms are less diagnostic climate indicators, because the glaciers that form such landforms respond to changes in both precipitation and temperature.

In alpine glacial systems, moraines far down valley from the limit of present-day ice indicate very different glacial mass balances in the past, either suggesting increased snow and ice accumulation and/or decreased **ablation** (ice loss) as the result of changing climate. The presence of now ice-free cirques, such as those on Mount Washington in New Hampshire, indicate that in the past, conditions were once sufficient to support alpine glaciers. If alpine moraines can be mapped, then equilibrium line altitudes can be estimated using a variety of methods, including the empirical **accumulation area ratio** technique (which relies on the observation that on average two-thirds of an alpine glacier's area lies in the accumulation zone [**Figure 13.1**]). Such data indicate that equilibrium line altitudes lowered during the LGM on average about 1000 m but lowering amounts varied greatly around the world because of localized conditions affecting ice accumulation and ablation rates. Continental glacial deposits (such as till and outwash) are broadly indicative of a cooler climate. Direct interpretation of the continental glacial record in terms of temperature and precipitation at a specific location is not possible because many different variables (e.g., bed materials, bed thermal status, and climate patterns) affect the size of ice sheets (see Chapter 9).

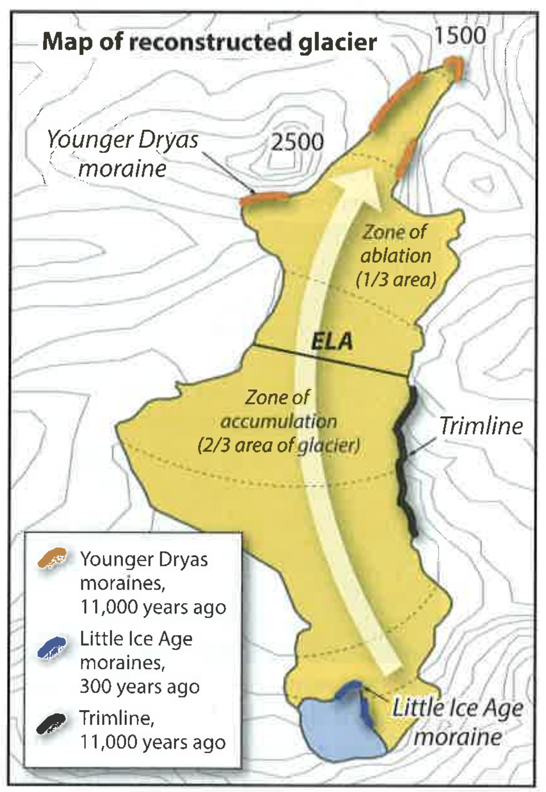
Climate-induced changes in sea level have left a significant mark on the landscape. Drowned valleys and deep estuaries along many coastlines (such as eastern North America) are evidence for postglacial sea-level rise. Similarly, many barrier islands (see Chapter 8) were born when sea level was lower and are today marching shoreward as sea level continues to rise. During periods of lower sea level, large portions of the continental shelves were exposed. This exposure is particularly important in terms of human and animal migration and archaeology. Lowered sea level exposed land and narrowed open water crossings, such as the Torres Strait and Bering Strait, during glacial times, facilitating migrations between continents and the peopling of both Australia and the Americas.

The most extreme interglacial periods, occurring about 130,000, 300,000, and 400,000 years ago were warm enough and/or long enough that large volumes of the Greenland and most likely the Antarctic ice sheets melted. This, along with expansion of the ocean water from warming (which for the warming of the past century accounts for about 40 percent of sea-level rise), increased the volume of the oceans so that that global sea level was as much as 12 m higher than it is today. During such sea-level high stands, waves cut shoreline platforms that today are preserved above sea level as marine terraces [**Photograph 13.2**].

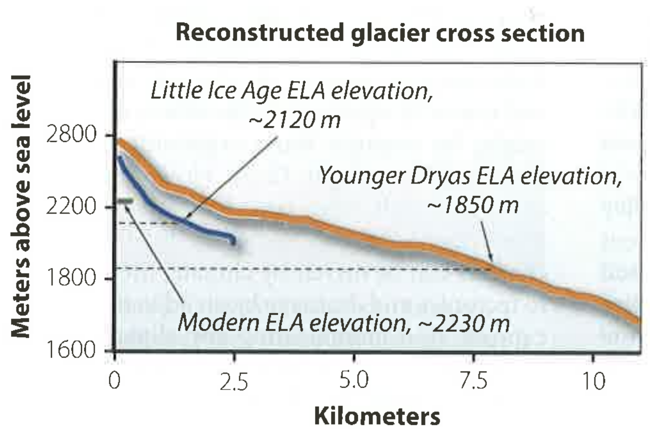
Fluvial features, including river terraces and knickzones (where river gradient steepens abruptly), are more difficult to interpret as climatically significant landforms. Some terraces, such as those made up of glacial outwash and traceable upstream to moraines, are clearly climatic in origin, for example those originating from ice cap outlet glaciers [**Photograph 13.3**]. However, terrace formation can also result from base-level changes and changes in river discharge and sediment loading. Although these changes can be driven by climate, they can also be related to tectonics and drainage basin adjustments such as stream capture, thus complicating any climatic interpretation of river terraces (see Chapters 7 and 12). Knickzones can form in response to climate change, expressed as lowered sea level when ice sheets grow, but they can also reflect active faulting as well as structural and lithologic discontinuities. Understanding landform setting and context are



Alpine **moraines** can be mapped in the field and used to define the down-valley extent of now-vanished glaciers. Moraines, which are built by deposition of material from melting ice, are found only in the ablation zone. In the accumulation zone, ice extent is defined by the extent of glacially polished rock and trimlines where weathered rock has been removed by glacial erosion.

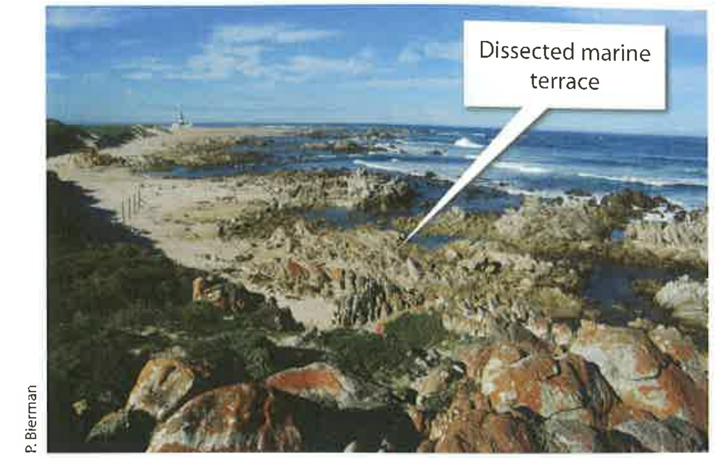


Using a topographic map and the location of mapped moraine segments and trimlines, one can sketch the outline of the glacier that once filled the valley. Empirical studies have shown that, on average, about 2/3 of an alpine glacier's surface area lies in the **accumulation zone** and 1/3 lies in the **ablation zone.** Using this **accumulation area ratio** (AAR) of 2/3, one can define the former equilibrium line altitude or **ELA,** the boundary between the accumulation and ablation zones. The surface of the reconstructed glacier can be contoured. In the accumulation area, the contours are convex up glacier because ice flow is convergent. In the ablation zone, the contours are convex down glacier because the flow is divergent. You can create a cross section of the vanished glacier using the contour map-based glacier reconstruction.



During the Younger Dryas cold period between about 12,800 and 11,500 years ago, the ELA in the Alps fell almost 400 m below its elevation today. During the Little Ice Age (between 1300 and 1850 CE), the ELA dropped 110 m from today's elevation of 2230 m. Assuming that ELA changes reflect only cooling, and considering a lapse rate of 1° C per 100 m elevation, the Little Ice Age was about a degree cooler than today. Younger Dryas times were about 4° C cooler.

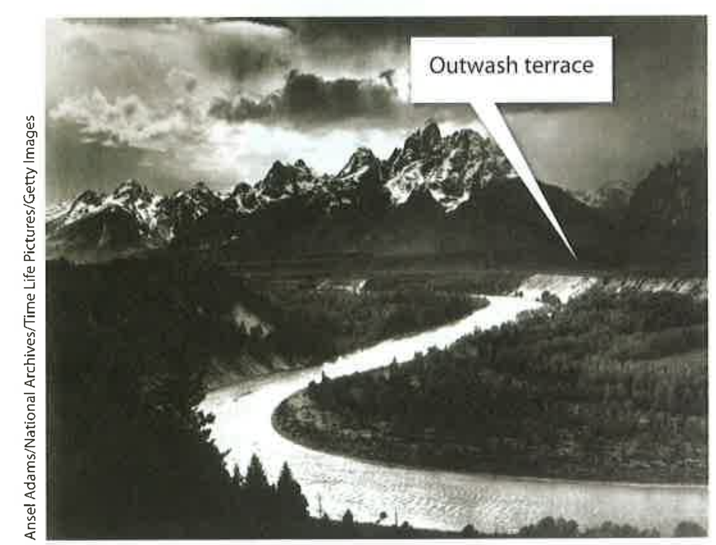
FIGURE 13.1 Reconstructing Vanished Glaciers. Using the accumulation area ratio (AAR) method, the equilibrium line altitude (ELA) of now-vanished glaciers can be reconstructed from a topographic map and field mapping of lateral and terminal moraines as well as trimlines. [Adapted from Sailer et al. (1999).]



PHOTOGRAPH 13.2 Dissected Marine Terrace. Determining paleo–sea levels in tectonically stable continental regions is a challenge because uplift rates are very low and thus older marine terraces remain close to present-day sea level. Here, along the southeastern coast of South Africa, is a dissected marine terrace that was likely cut about 400,000 years ago during MIS (marine isotope stage) 11, when global sea level was higher than today.

critical to any climatic interpretation of river terrace sequences and knickzones.

Soils developed on landforms preserve evidence of changes in climate. For example, relict calcic (Bk) horizons (accumulations of calcium carbonate in soil B horizons found in areas where no calcium carbonate is being deposited in soils today) directly indicate a change in the ratio of precipitation to evaporation (see Chapter 3). Other indicators of climate change are less direct. Many plants create microscopic structures made of opal (amorphous silica) and known as **phytoliths.** Phytoliths can be preserved in soil A horizons and separated for analysis. Because the shape of phytoliths differs depending on plant type, they can be useful for documenting the type of vegetation that once covered a site; for example, phytoliths can be used to distinguish between grass and forest cover, a difference that can be used to infer climate change.



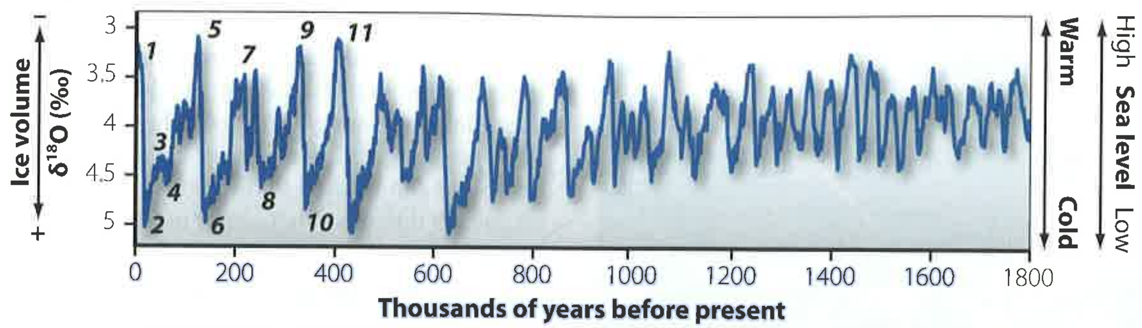
PHOTOGRAPH 13.3 Origin of River Terraces. Some flights of river terraces are climatic in origin. Here, along the Snake River, outwash terraces at Deadman's Bar are composed of material that came from the termini of outlet glaciers of the Southern Yellowstone–Absaroka Mountains ice cap. The upper terraces are of last glacial maximum age. The lowest terraces are postglacial.

Lake and Marine Sediment

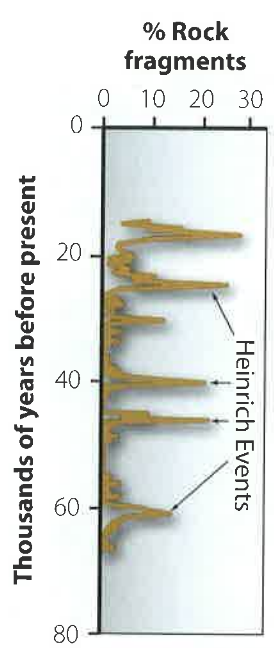
The stratigraphy and composition of marine and lake sediment cores can be used to decipher changes in climate and differences in geomorphic processes and their rates over time.

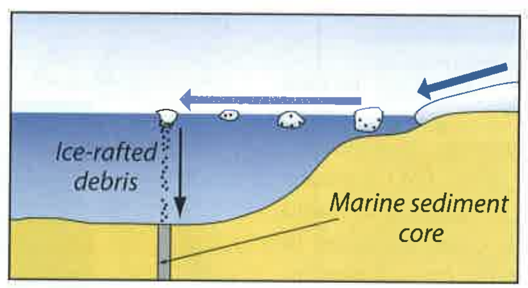
Some sediment shows a clear cyclicity in the color or grain size of the material it contains. For example, cores from postglacial, humid-temperate zone ponds are typically dominated by gyttja (from the Swedish, for "mud" or "ooze"), the dark, fine-grained, organic-rich muck that squeezes between your toes when you walk in for a swim. Cores from such ponds reveal thin bands of gray or tan sand and silt [**Photograph 13.4**], interpreted as storm deposits resulting from flooding-induced erosion in the pond's watershed. In marine cores from the North Atlantic,

PHOTOGRAPH 13.4 Paleoflood Layer. Core section with centimeter scale on the right side showing gray, sandy, paleoflood layer sandwiched between organic-rich pond sediment. The sand was deposited when a large storm hit the basin, causing runoff and sediment transport. Core was taken from Chapel Pond in northern Vermont.



Marine sediment cores hold information about the history of climate, including the extent of glaciation (ice volume) and ocean temperature. By measuring stable oxygen isotope ratios in foraminifera, single-celled, $CaCO_3$-rich organisms, we can infer the long-term behavior of the glacial system because the formation of large ice masses preferentially sequestered large amounts of ^{16}O, removing it from the oceans. Odd-number marine isotope stages indicate small ice masses, warm climate, and thus high sea level, whereas even-number marine isotope stages indicate large ice masses, cold climate, and lowered sea level. The pacing of glaciations changed from 100,000-year cycles to 40,000-year cycles about a million years ago.





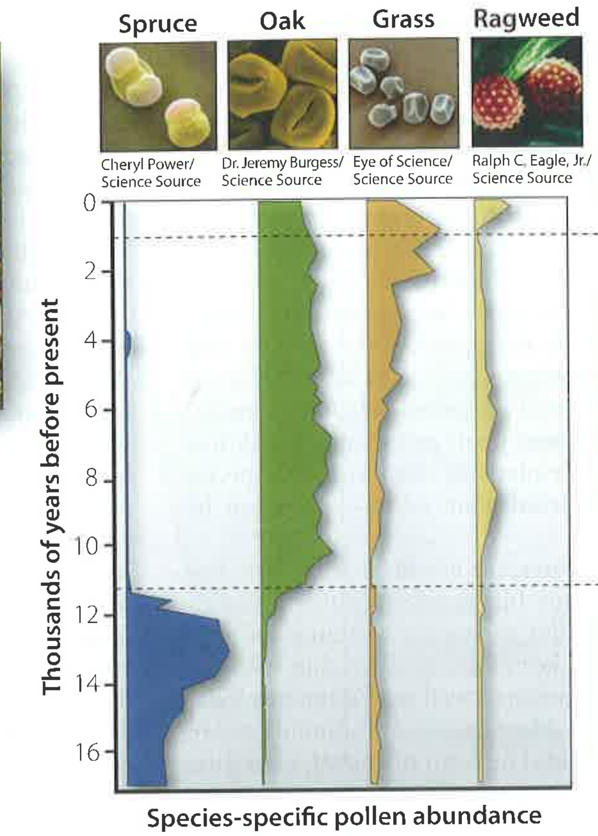
When glaciers calve and shed icebergs, those icebergs float out to sea; as they melt, they drop **ice-rafted debris (IRD)** onto the ocean floor. Marine cores sample this sediment and can be analyzed to determine how sediment grain size changes over time. The peaks in debris concentration identify times of rapid calving and high rates of iceberg production. High concentations of ice-rafted debris in marine sediment cores are known as **Heinrich events.**

IRD in core section

FIGURE 13.2 Marine Sediment Record of Changing Quaternary Climate. Stable oxygen isotope ratios of marine foraminifera ($CaCO_3$ shells) reflect the amount of water stored in the planet's glaciers and the temperature of the oceans. Ocean sediment cores also reveal periods when ice-rafted debris dumped from icebergs was abundant. [Data from Lisiecki and Ramyo (2005).]

grain size and stratigraphic analyses reveal isolated accumulations of large stones embedded in a matrix of poorly sorted debris [**Figure 13.2**]. Known as **ice-rafted debris** (**IRD**), such material is the smoking gun of glaciation in marine cores and indicates where and when icebergs, calved from tidewater glaciers, dropped loads of previously ice-bound sediment. In glacial lake sediments, changes over time in the thickness and grain size of annual silt, clay, and sand layers (varves) can be interpreted both in terms of sediment delivery from changing summer climate (seasonal runoff) and ice retreat (distance from the paleo–ice margin). Warm summers and nearby ice result in thick varves composed of coarse sediment. During cold winters and when the ice margin is farther away, varve layers are thinner and composed of finer-grained sediment.

Lake and marine sediments can be analyzed using a variety of techniques; chemical, physical, and isotopic data are used to infer changes in climate from established



Ragweed colonizes disturbed land after European settlement.

During the wetter and warmer Holocene, oak and grass pollen become more prevalent.

Spruce pollen dominates in cold and dry glacial periods.

In arid regions, packrats scavenge vegetation, return it to their nests, and urinate on it, creating a resinous material called **amber-rat** that preserves characteristic macrofossils for identification by paleoclimatologists. These accumulations are termed **middens.**

Pollen grains are preserved in geologic archives such as lake mud that can be cored. The cores can be dated and the pollen can be identified to the species level by its characteristic shapes. Counting pollen grains of different types allows reconstruction of the vegetation that used to grow near the sample site and illustrates how that vegetation assemblage changed over time. This example is from Nelson Lake, Illinois.

FIGURE 13.3 Vegetation Responds to Climate Change. Animals, such as packrats, collect vegetation and store it in their middens. Pollen, preserved in lake sediments, can be identified to determine ancient vegetation assemblages.

relationships between climate and the measured parameter. For example, the inorganic chemical composition and stable oxygen isotope ratios of preserved marine organisms can be used to estimate ocean temperature. In other cases, **pigments** (organic molecules extracted from sediment) are analyzed because their composition relates to water temperature. Many sediments preserve biologic debris, a rich source of information about climate change. In some cases, analysis of **pollen** and **macrofossils** reveals the species of plants in local or regional vegetation communities; such fossils are usually well-preserved in lake sediments because oxygen levels are low enough to prevent substantial decay.

Pollen analysis is one of the most widely applied tools in paleoclimate research. The morphology or shape of pollen grains released by plants is unique for each genus [**Figure 13.3**] and pollen is durable and thus well-preserved. Qualitative pollen analysis is straightforward. If the pollen of cold-tolerant species such as spruce is found in sediments and spruce trees are not found near the pond or lake today, one can infer climate was likely colder in the past. Finding the pollen of species like oak and pine indicates a warm climate, and ragweed pollen increases after western settlement and land clearance. Determining how much temperature change is represented by a change in the species composition of fossil pollen requires comparison with modern analogs in different climatic and geographic zones. Because some pollen, like pine, can travel tens to hundreds of kilometers, pollen records of climate are more regional than local.

Macrofossils (primarily plant parts such as cones, leaves, and needles) are often recovered from the same lake cores used for pollen analyses and are useful because they provide information about only the biologic communities within the watershed of the lake. Insect parts preserved in sediment also tell a climate story. For example, distinctive pieces of chitinous beetles' exoskeletons preserve well in lake sediment. Because different beetle species live in different climate zones, and because each species of beetle has distinctive skeletal parts, the distribution of beetle pieces found in prehistoric sediment can be used to reconstruct climates of the past.

In arid regions, where lakes are scarce and ephemeral, the content of packrat **middens** (nesting sites in caves and rock crevices) is used to infer the paleodistribution of vegetation (Figure 13.3). Packrats (small desert rodents) gather bits of vegetation from the area around their middens and then urinate on the debris. Over decades and centuries, the fossilized plant debris and urine build up into a solid mass of well-preserved organic material known as **amber-rat.** Former plant communities and thus paleoclimate can be determined by analyzing the species distribution of the preserved plant parts—which can be dated by ^{14}C.

In marine sediment cores, the chemical and stable isotopic character of various biota, in particular **foraminifera**, single-celled, calcium carbonate secreting animals **[Photograph 13.5]**, have been used to determine both the paleotemperature of the ocean as well as to estimate global ice volume (Figure 13.2). Most species of foraminifera are small, the size of fine sand. The ratio of Ca/Mg in calcite, the mineral that makes up the shells of these tiny marine creatures, changes with water temperature, providing a paleothermometer for ocean water.

Both ocean water temperature and the amount of water locked up in ice caps, ice sheets, and glaciers determine the oxygen isotope ratio of the shells secreted by marine organisms. $^{18}O/^{16}O$ works as a paleotemperature indicator because lower $^{18}O/^{16}O$ ratios are correlated with lower ocean water temperatures. Global ice volume estimates work because $H_2{}^{16}O$ has a slightly higher vapor pressure than $H_2{}^{18}O$, and thus $H_2{}^{16}O$ is preferentially evaporated from the ocean. As glaciers grow on land and store the isotopically lighter water from the ocean in their ice, the $^{18}O/^{16}O$ ratio in seawater increases. Foraminifera incorporate some of this oxygen into their $CaCO_3$ shells, die, drop to the ocean floor, and are preserved in sediments. The ratio of stable oxygen isotopes in the shells, ^{18}O versus ^{16}O, can be interpreted as a paleothermometer (if calculations are made to account for the loss of ocean water to ice-sheet storage) and as a measure of global ice volume (if other temperature records are used to correct the measured isotope ratio for temperature dependence).

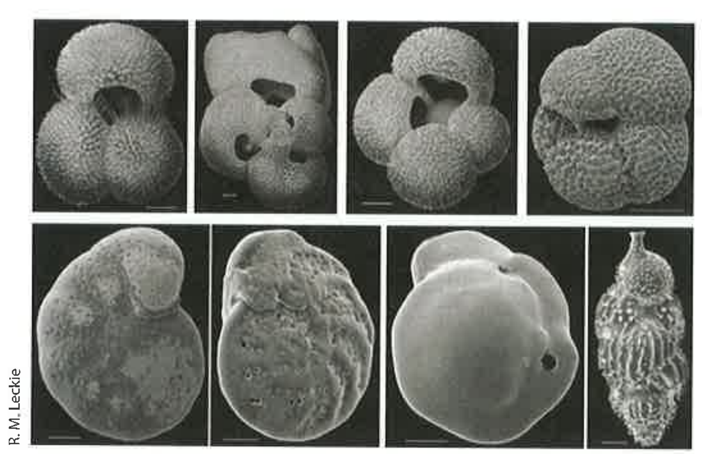
The cyclical nature of marine oxygen isotope changes is clear, and the major peaks and troughs in the curve (Figure 13.2) are identified as numbered **marine oxygen isotope stages (MIS)**. Even-numbered stages are times when climate was cooler than average, ice sheets were larger, and sea level was lower. Odd-numbered stages are times when climate was warmer than average, ice volume was less, and sea level was higher. Dating of these stages was first done by tuning (stretching) marine records to match predictions of warm and cool times deduced from calculations of Earth's changing orbit and thus the latitudinal distribution of solar radiation. More recently, the MIS timescale has been updated by other, more direct dating approaches; this updating has largely validated the original orbitally tuned timescale.

Ice Cores

Ice, which can be cored and collected from ice sheets as well as smaller glaciers, preserves a detailed record of climate in cold regions, both in the Arctic and Antarctic as well as in high mountains **[Figure 13.4]**. Ice cores contain the water that fell as snow on the glacier, along with dust and fragments of volcanic ash as well as chemical aerosols and gases (such as CO_2), which become trapped as the snow consolidates into impermeable ice containing discrete bubbles (see Figure 9.3), closing off exchange with the atmosphere. The concentration of CO_2, CH_4, and other gases trapped in bubbles in the ice tells us the past composition of Earth's atmosphere. From analysis of glacial ice cores, we now know conclusively that the concentration of these greenhouse gasses has varied closely with temperature over at least the past 800,000 years. Layering in ice cores allows counting of annual bands much like tree rings and so the upper portion of ice cores can be dated precisely. Such layering can persist for thousands to tens of thousands of years in ice where accumulation rates are high and ice deformation rates are low.

Much of what we know about paleoclimate has come from ice cores, in particular measuring the stable oxygen isotope composition of the ice. For example, the $^{18}O/^{16}O$ ratio of glacial ice can be used to infer the air temperature (which in polar regions was as much as 15°C colder than today at the last glacial maximum). As air cools, the $^{18}O/^{16}O$ ratio of precipitation becomes more depleted; there is less ^{18}O.

Ice cores preserve the chemistry of the snow and they are archives of dust fall. Changing acidity levels in the ice are related to volcanic eruptions because such eruptions emit gases, such as SO_2, that react with water to form acid (Figure 13.4). Sometimes, volcanic ash is found in an ice core. If tephrochronology can be used to identify the ash and if the ash has been dated isotopically, then the age of the ice layer in which the ash was found can be determined. Such ash dates are used to cross-check layer counting and to extend ice-core chronologies beyond the age (below the depth) at which discrete layers are preserved.



PHOTOGRAPH 13.5 Planktonic and Benthic Foraminifera Used in Paleoclimatic Studies. The upper row illustrates planktonic, near-surface dwelling species. The lower row illustrates benthic, deep water foraminifera. Grey bar under each foraminifera is 100 μm long.

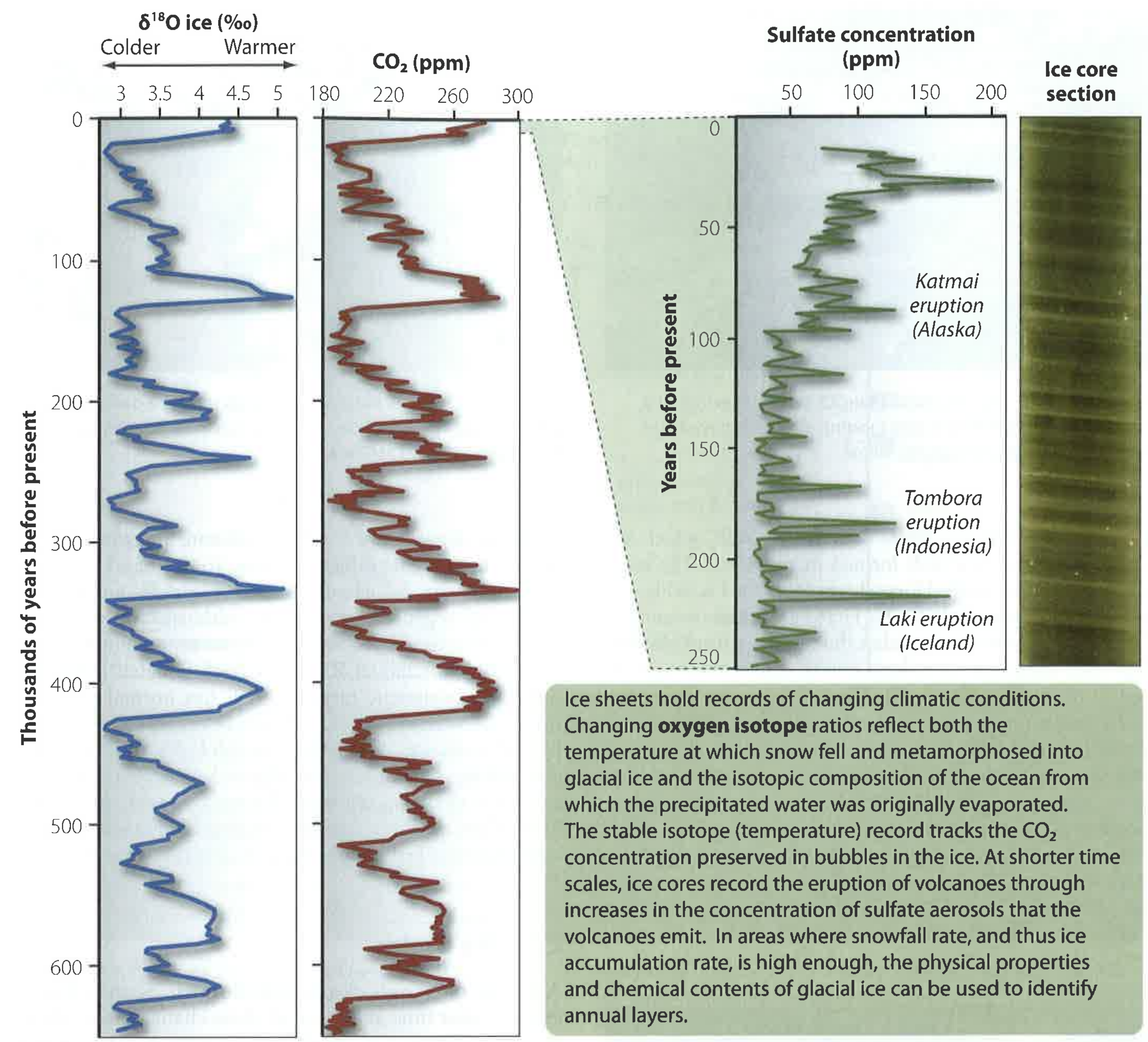


FIGURE 13.4 Climate Records Preserved in Glacial Ice. The stable oxygen isotope composition, CO_2 concentration, and sulfate concentration are used to interpret paleotemperature, atmospheric composition, and the history of volcanic eruptions. [Data from Siegenthaler et al. (2005).]

Windblown Terrestrial Sediment

Loess, silt-sized wind-blown dust, preserves a record of both sediment availability and windiness, both of which are related to climate (see Chapter 10). Loess deposits are found downwind of significant sediment sources; for example, much fine sediment was sourced from the broad outwash plains that bordered glaciers and ice sheets **[Photograph 13.6]**. The middle of North America has deep and fertile loess deposits left behind when strong winds scoured fine sediment from massive outwash plains

J. Licciardi

PHOTOGRAPH 13.6 Outwash Plain. Outwash plain (braided stream) from which wind is transporting dust (which could be deposited as loess) in Kagbeni, Nepal.

M. Smith, T. Stevens

PHOTOGRAPH 13.8 Paleosol. Scientists sample a dark red paleosol (upper section) developed on tan loess (lower section) in the Loess Plateau of China.

of the Laurentide Ice Sheet. Rich prairie soils, which are today intensively farmed, formed in this loess. The most famous and best-studied loess sheets cover the Loess Plateau of China [Photograph 13.7]. Here, loess that originated from deserts to the west, rather than directly from glaciers, has been deposited for at least several million years, and is >100 m thick in places. The Chinese loess sequence contains many **paleosols** or buried soils [Photograph 13.8]. Each paleosol indicates a period of soil formation and stability when the climate was relatively warm and wet in the source and deposition areas; therefore, dustiness was reduced, loess deposition rates were low, and biologic activity on the plateau was high. Each loess layer indicates a time when climate was colder and dustier in the source area and therefore loess deposition rates increased, exceeding rates of soil development.

Jim Richardson/National Geographic Stock

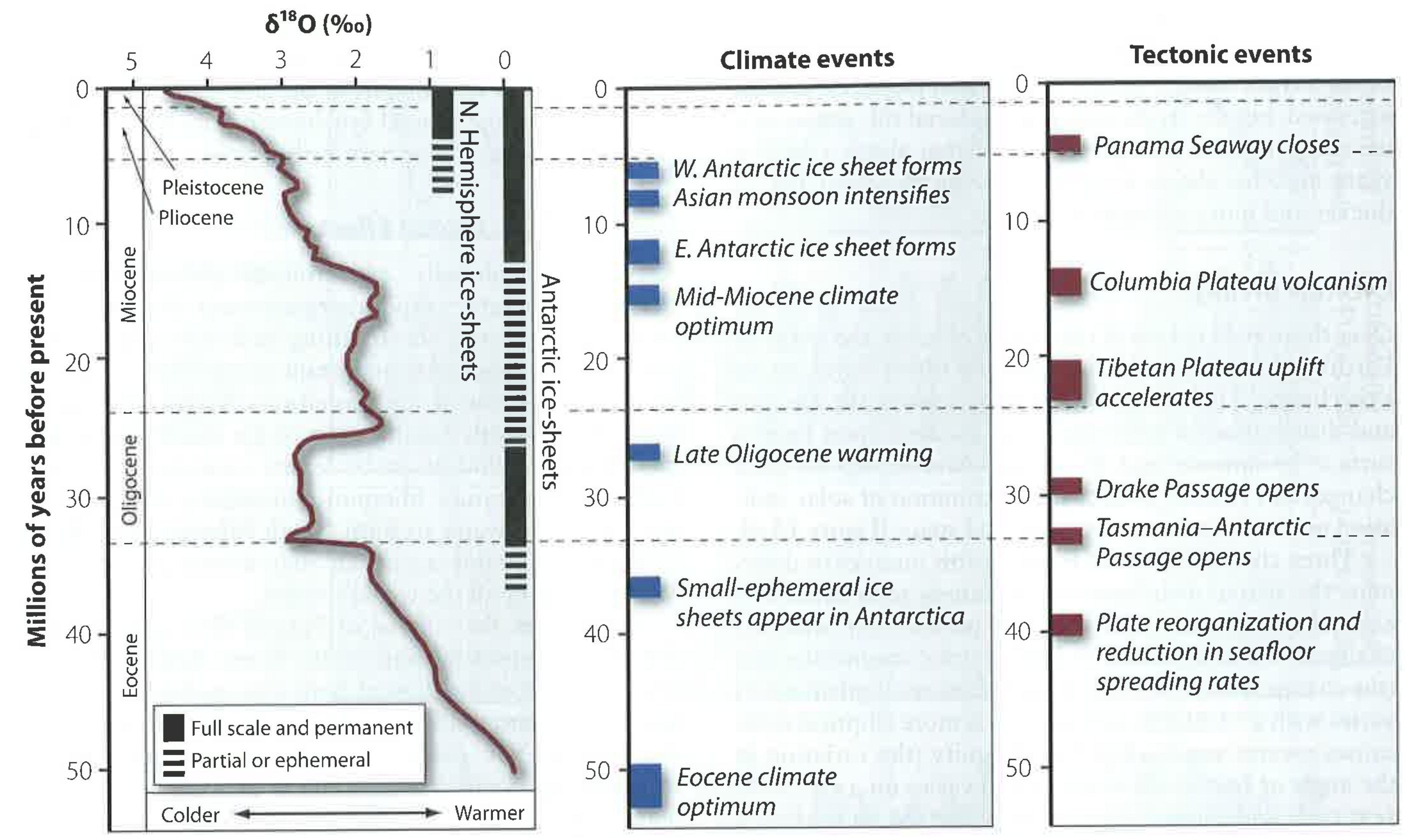
PHOTOGRAPH 13.7 Loess Plateau. The Loess Plateau of China is covered by tens to hundreds of meters of loess deposited over the Quaternary period. Deforestation and intensive land use have caused significant soil erosion. Here, trees are planted in an attempt to stabilize the easily eroded loess.

Loess deposits are a valuable climate and geomorphic archive but they are difficult to date. Young loess (<40,000 years old) can be dated using radiocarbon analysis of associated organic matter. Loess with ages up to a few hundred thousand years can be dated using luminescence methods (see Chapter 2). Older loess is dated primarily using paleomagnetic methods (if it has normal polarity, the loess was deposited <780,000 years ago; reversed polarity indicates that the loess is older). With such imprecise dating, older loess provides a less well-constrained record of changing regional climate over time. In Alaska and other volcanic regions, tephrochronology of ash found in loess beds can be used to date loess deposits.

Climate Cycles

A variety of records clearly indicates that Earth's climate changes over time and some of those changes are cyclical (Figures 13.2 and 13.4). Over the timescale relevant for most geomorphology, the last few million years of Earth history, climatic changes have periodically plunged our world into glaciations that spread ice over much of the Northern Hemisphere and some of the Southern Hemisphere. On the timescale of decades, climate variability associated with a variety of large-scale atmospheric and ocean processes has had significant geomorphic effects.

Earth's climate largely reflects an energy balance between incoming solar short-wave radiation and outgoing long-wave radiation. Thus, the dominant explanation for longer-term climate variability considers changes in the seasonal distribution of incoming solar radiation on Earth's surface due to predictable changes in Earth's orbit. Variations in atmospheric and oceanic heat and water transport can influence climate on shorter timescales. The brightness of the Sun may also change over decadal to millennial timescales, possibly affecting climate.



Oxygen isotope evidence from marine sediment cores indicates that over the last 50 million years, Earth's climate has grown significantly colder (leftmost diagram). On the right side, panels show both tectonic events that influence climate and major climate events of the last 50 million years. Over this timescale, cooling was driven by the changing position of continents, such as Antarctica moving toward the South Pole and the opening of the Drake Passage which isolated Antarctica from the warm waters of the southern Atlantic and Pacific oceans. This isolation allowed glaciation to begin in Antarctica long before it did in Greenland. Feedbacks between the solid Earth, the hydrosphere, and the cryosphere (such as increasing albedo as the area of ice sheets increased) continued this cooling trend.

FIGURE 13.5 Climate and Tectonics. Over the past 50 million years, Earth's climate has changed because of tectonic events, including uplift, opening of ocean passages, and large volcanic eruptions. [Adapted from Zachos et al. (2010).]

Glacial Cycles

The most recent period of Earth history during which glaciation was common began in earnest about 2.7 million years ago, although large amounts of ice began to cover parts of Antarctica several tens of millions of years earlier [Figure 13.5]. Most of Greenland remained ice free until the early Pleistocene (~2.7 million years ago), although it appears some ice began to accumulate in Greenland during the Pliocene (7 million years ago), and there are hints of ice in Greenland (evidence from marine sediment records of ice-rafted dropstones) as early as 38 million years ago.

Early glaciations, identified by isotopic excursions in deep-sea sediment records, were fast paced and symmetrical with cooling and warming occurring in a regular, smooth pattern on a ~40,000-year cycle (Figure 13.2). About a million years ago, something changed abruptly. The cycles became much longer (~100,000 years) and asymmetric, with slow cooling and ice expansion, and then very rapid warming and quick demise of extensive ice sheets; these rapid warmings are referred to as **terminations**.

There is no consensus regarding why the cyclicity changed. Some suggest that the change in cyclicity was the result of decreasing atmospheric CO_2 and the concurrent cooling of the planet, the result of which was larger and less responsive ice sheets. Another explanation considers evolving ice sheet interactions with previously weathered regolith. The regolith idea suggests that after the soft, extensively weathered, pre-Quaternary regolith was stripped away, Northern Hemisphere ice sheets began flowing over hard, crystalline bedrock, which increased basal shear stresses ($\tau = \rho g h \sin\theta$, see Chapter 9); thus, ice sheets thickened and thereafter responded more slowly to orbital forcing (hence the jump from 40,000-year to 100,000-year cycles). This inference is consistent with data showing that after the first million years of extensive glaciation, the ocean oxygen isotope record suggests that global ice volume increased, but the terrestrial record (glacial till) shows that ice extent did not change. It appears that about a million years ago, ice sheets went from being thin and flat to thicker and more voluminous.

Orbital Forcing

Over thousands to tens of thousands of years, the shape of Earth's orbit around the Sun and the tilt of Earth on its axis change. These orbital variations control the amount and distribution of solar radiation incident upon Earth's surface by latitude and by season. Accounting for such changes can reliably predict the distribution of solar radiation on Earth's surface over time and space [Figure 13.6].

Three characteristics of Earth's orbit interact to determine the spatial distribution of incoming **solar radiation;** each characteristic has a different period over which it changes. The **eccentricity** of Earth's orbit around the Sun (the change from a more circular to a more elliptical orbit) varies with a ~100,000-year period. A more elliptical orbit causes greater seasonality. The **obliquity** (the variation in the angle of Earth's tilt as it rotates) varies on a ~41,000-year cycle and changes seasonality. When the tilt is greater, there is more contrast between summer and winter temperatures. Cool summers better preserve snow from the previous winter and thus initiate ice sheet growth. The **precession of the equinoxes,** or the variation in the direction of Earth's rotation axis as it orbits the Sun, also changes seasonality but with ~22,000-year cyclicity. Together, these three orbital characteristics define **Milankovitch cycles,** named in honor of Milutin Milankovitch (1879–1958), a Serbian engineer who made the first detailed calculations indicating that such cycles could control Earth's climate. Although orbital variations change the seasonal distribution of insolation for a specific time and place on Earth's surface up to 20 percent, they do little to change total annual insolation for Earth as a whole, which varies over time by no more than 0.3 percent.

Understanding exactly how changes in solar radiation (specifically at high northern latitudes where ice sheets are born) translate into planet-wide glaciations, interglaciations, and rapid changes in paleotemperature has not come easily and has revealed numerous interactions, feedbacks, and resulting amplifications between the solid Earth, the atmosphere, and Earth-surface processes. Glaciations begin when cool summers allow some snow at high Northern Hemisphere latitudes (nominally 65°) to survive the melt season and thus begin to build up into glacial ice. Once this buildup begins, the ice and snow have greater albedo (reflectivity) than the forest or tundra they covered, and thus reflect more of the incoming solar radiation. This positive feedback further cools the planet, encouraging expansion of the nascent ice sheet. The effect of greenhouse forcing (changes in the atmospheric concentration of CO_2, CH_4, and water vapor resulting from glaciation and its effects) is responsible for the general synchroneity of climate change in the Southern and Northern hemispheres.

Local Events—Global Effects

Abrupt and globally synchronous paleotemperature changes likely reflect rapid reorganization of ocean circulation patterns and the resulting redistribution of heat around the planet. An important component of global oceanic circulation is the formation of cold, salty, dense water in the North Atlantic, part of the ocean's **thermohaline circulation,** flow driven by density contrasts that depend both on temperature (thermo) and salinity (haline). The sinking of this water to form **North Atlantic Deep Water** (NADW) maintains a current that moves 20 times the average flow of all the world's rivers.

Changes in the volume of NADW flow can alter the delivery of tropical warmth to the North Atlantic through small changes in the flow of both intermediate depth and surface currents, perhaps amplified by changes in sea ice distribution. For example, diminished northward heat transport by ocean currents (likely because of a sudden outburst of cold, fresh water into the North Atlantic, see Digging Deeper) contributed to the rapid cooling of the North Atlantic, Europe, and eastern North America known as the **Younger Dryas** cold episode 12,800 to 11,500 years ago. The name Dryas refers to an alpine plant that appeared at lower altitudes during this chilly time period. Sea ice cover expanded significantly at this time, likely driving a large temperature change around the North Atlantic because ocean-atmosphere heat exchange was suppressed. Evidence from Greenland ice cores suggests that average temperatures there plummeted about 15°C within a few decades and that the millennium-long cold spell caused glacial margins to advance and changed plant communities around the North Atlantic.

Similar flooding of the North Atlantic by fresh, less-dense glacial meltwater appears to have altered ocean heat flow on at least six occasions during the last glacial period, rapidly changing climate at least regionally. These episodes of flooding, called **Heinrich events** (Figure 13.2), were caused by Northern Hemisphere ice sheets episodically discharging large volumes of fresh water and icebergs to the North Atlantic. Evidence for Heinrich events comes from the abrupt appearance of coarse, terrestrial sediment in ocean sediment cores collected far from land; such terrestrial material could only have been brought far out into the ocean by icebergs. Heinrich events seem to have been particularly frequent when the Laurentide Ice Sheet was partially extended. In this configuration, small changes in glacier extent forced meltwater discharges to alternate between the Mississippi River drainage (greater

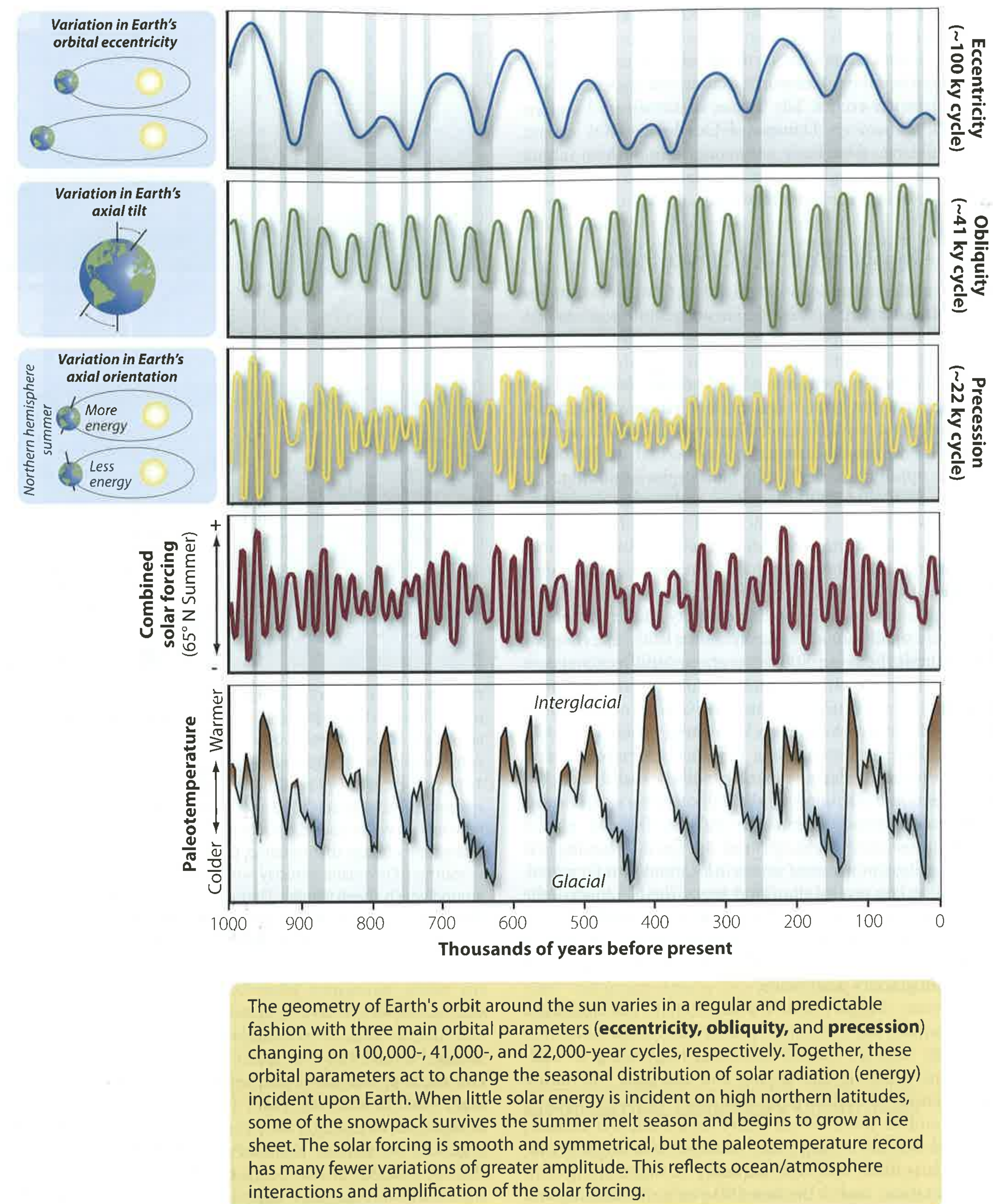


FIGURE 13.6 Orbital Forcing of Climate Change. The geometry of Earth's orbit changes the seasonal distribution of radiation incident on Earth's surface over time.

ice extent) and the Saint Lawrence River drainage (lesser ice extent).

Significant changes in North Atlantic climate, recorded as changes in isotope ratios in ice cores, are associated with some Heinrich events. The causes of these rapid climate changes, known as **Dansgaard–Oeschger** (DO) events, remain elusive, since there are many more of them (about 20 during the last glacial period) than there are massive iceberg discharges (Heinrich events).

Climate Variability Within a Climate State

Within a glacial or an interglacial period, there can be variability sufficient to drive significant geomorphic change. Such variability can occur on the scale of decades, centuries, and millennia.

During the Holocene interglacial period (the last 12,000 years), climate has been in general much more stable than it was during the previous glacial interval. The Holocene was in general warmest during the **Altithermal** or **Holocene climatic optimum** (several millennia in the mid-Holocene, between 8000 and 5000 years ago) and coolest during the **Neoglacial** (the last several thousand years). The distribution of moisture during the Holocene changed as wind belts and storm tracks shifted latitude in response to changing Northern Hemisphere summer insolation.

Despite the relative climatic stability of the Holocene, there were significant, climatically driven landscape changes. For example, between 8000 and about 5000 years ago, in response to a slightly warmer mid-Holocene climate, various lines of evidence indicate the Greenland Ice Sheet retreated inland, perhaps many kilometers, behind its present margin. The most compelling evidence for retreat comes from radiocarbon dates of marine mollusk shells reworked by the ice and included in the till of historic moraines. These mid-Holocene age clams indicate that the glacier margin must have been located up fjord, allowing the mollusks to grow in marine waters in locations now far inland. During the last several thousand years, the Neoglacial, the ice tongues then readvanced, scooping up the mollusks and delivering them to the glacial margin mixed with glacial sediment. Such Neoglacial advances are also typical of mountain glaciers worldwide.

Different changes occurred away from the glacial margin. For example, the woodland-prairie ecosystem boundary in North America moved eastward, responding to mid-Holocene warming and drying and changing the nature and intensity of soil-forming processes. In New England, spruce and fir trees, which thrive in a cool, wet climate, replaced warmth-loving pine trees over the last few millennia. It is likely that fire frequency changed along with the vegetation, and if the last 100 years of warming is a clue to past biotic-landscape interactions, the tree line was probably higher, stabilizing steep mountain slopes in the warmer mid-Holocene.

More recently, and on a shorter timescale (centuries), the North Atlantic climate warmed sufficiently (perhaps a degree Celsius on average) to allow the Vikings to settle Greenland and Newfoundland during the **Medieval Warm Period** (900–1300 CE) but soon after, the climate cooled, slipping into what has been termed the **Little Ice Age**, several hundred years (1300–1850 CE) of temperatures cool enough that the canals of Holland froze regularly in winter, the Norse settlements were abandoned, glaciers advanced in the Alps destroying buildings that were in their way, and crops failed repeatedly in the cool, wet weather setting the stage for extensive famine [**Photograph 13.9**]. Other societally important climate changes include drying of the Sahara Desert and the advance of sand dunes over streambeds and settlements about 5000 years ago [**Photograph 13.10**].

(a)

Ice Sports, c. 1610 (panel), Avercamp, Hendrick (1585–1634)/Mauritshuis, The Hague, The Netherlands/ The Bridgeman Art Library

(b)

P. Bierman

PHOTOGRAPH 13.9 Climate Change Over the Past Millennium. Climate change has affected society's interaction with the landscape. (a) The painting *Ice Skating Near a Village*, by Hendrick Avercamp, shows a Dutch scene during the Little Ice Age, about 1610, depicting colder winters. Several hundred years of cold temperatures between 1300 and 1850 allowed ice to form regularly on water bodies, such as rivers and lakes, that today rarely if ever freeze. (b) Brattahlid, Erik the Red's Viking farmstead in southern Greenland, is today nothing more than foundations around which sheep wander. During the Medieval Optimum, Viking society thrived here but the cooling that led into the Little Ice Age ended Viking settlement on Greenland about 1300 CE.

R. Blom/NASA/JPL-Caltech

PHOTOGRAPH 13.10 Sahara Desert. Today the Sahara Desert is dry and sandy but in the early to mid-Holocene, before 5000 years ago, it was much moister. In 1982, imaging radar flown on the space shuttle Columbia detected a large network of river channels buried beneath just a few meters of sand. The radar swath in gray shows the channels. The background image in orange shows the sand-sheet surface. Fieldwork confirmed the presence of stream-rounded pebbles (upper left) and Neolithic artifacts (lower right) left by people attracted to the water resources near the channels.

Short-Term Climate Changes

Repeated and persistent oscillations in patterns of rainfall and runoff on yearly to decadal timescales affect the tempo, distribution, and intensity of surface processes. On the basis of barometric pressure comparisons and sea-surface temperature distributions, a number of climate patterns characterized by short-term variability have been identified. The best-known climate patterns are the **North Atlantic Oscillation** (NAO), the **Pacific Decadal Oscillation** (PDO), and the **El Niño–Southern Oscillation** (ENSO). Despite the name "oscillation," variability of the PDO and NAO are indistinguishable from random noise with the time between different phases lasting from weeks to decades. In contrast, ENSO is a true oscillation, varying with a period of 2–7 years.

ENSO events (defined by sea-surface temperatures and atmospheric pressure changes in the equatorial Pacific) can be geomorphically significant. The phase during which sea-surface temperatures are high in the eastern Pacific is defined as an **El Niño** event. El Niño events are associated with heavy rains, floods, and landslides along the western coast of South America and the southwestern coast of North America—precipitation that greens the Mojave Desert as long-dormant seeds sprout after soaking rains. On the west coast of the United States, in California, El Niño brings storminess that causes large ocean waves, high river flows, and warm average temperatures. Such changes increase beach erosion, the frequency of debris flows and landslides, and the number of rain-on-snow flooding events. In contrast, during El Niño years, hurricanes and the landslides, floods, and coastal erosion they cause are suppressed along the Gulf Coast and eastern North America. Coral records from the tropics show that ENSO variability existed throughout the Holocene and the last glacial period. Analysis of varves from glacial Lake Hitchcock, which occupied the Connecticut River Valley for >4000 years in the Late Pleistocene, shows distinct changes in the amount of runoff and thus possibly rainfall at the 3–5 year timescale, consistent with ENSO climate oscillations at the end of the last glacial period.

The NAO, defined as the pressure gradient between the Icelandic Low and the Azores High, is related to the intensity of storms affecting the North Atlantic, including northern North America and Europe. Evidence for long-term NAO variations was identified by tracking the frequency of storm-induced inorganic sediment layers in New England lake cores dated by ^{14}C measurements on macrofossils (leaves, cones, and twigs; Photograph 13.4). Geomorphically effective changes in runoff and sediment transport (storm sediment layers) were attributed to the NAO because they matched in phase, frequency, and timing layers of ice containing high levels of sea salt recovered by ice coring from the center of the Greenland Ice Sheet. This increase in sea salt in the glacial ice was attributed to strong, salt-laden ocean winds blowing inland to central Greenland. The cause of NAO changes is not understood but may relate to small changes in solar output, which in climate models tend to produce an NAO-like response. Such changes may also explain the pattern of temperature and precipitation changes during the Little Ice Age cooling.

Geomorphic Boundary Conditions

Changing climate directly affects almost all geomorphically important variables including the frequency, intensity, and duration of precipitation, temperature, and both local and global base levels to which rivers and streams are graded. These changes are linked through the intensity of the global hydrologic cycle, the activity of which increases with temperature and with the temperature gradient between the poles and the equator.