Geomorphology and Climate

13

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.

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.

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

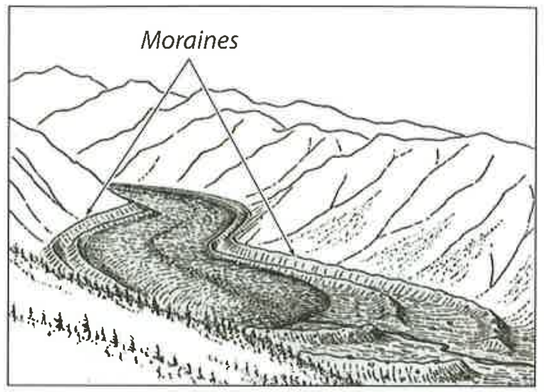
Some landforms are direct or indirect indicators of a changed climate. The challenge lies in dating the landforms and, in some cases, showing that the landforms result from changing climate and not changes in other factors such as tectonics. In much of the world, glacial and periglacial features are 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.

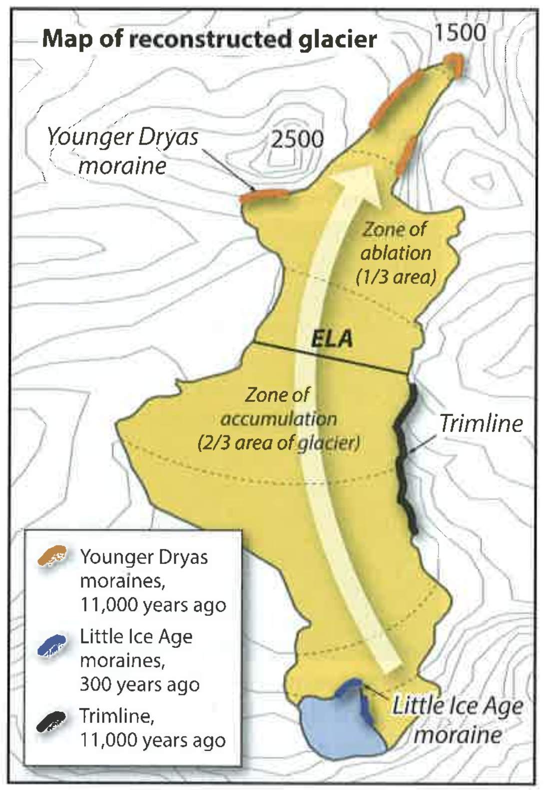
The most extreme interglacial periods, occurring about 130,000, 300,000, and 400,000 years ago were warm enough and/or long enough that large volumes of the Greenland and most likely the Antarctic ice sheets melted. This, along with expansion of the ocean water from warming (which for the warming of 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

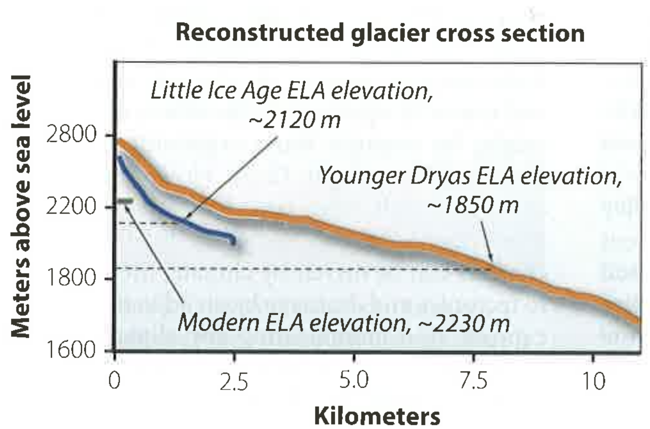


USGS

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).]

P. Bierman

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.

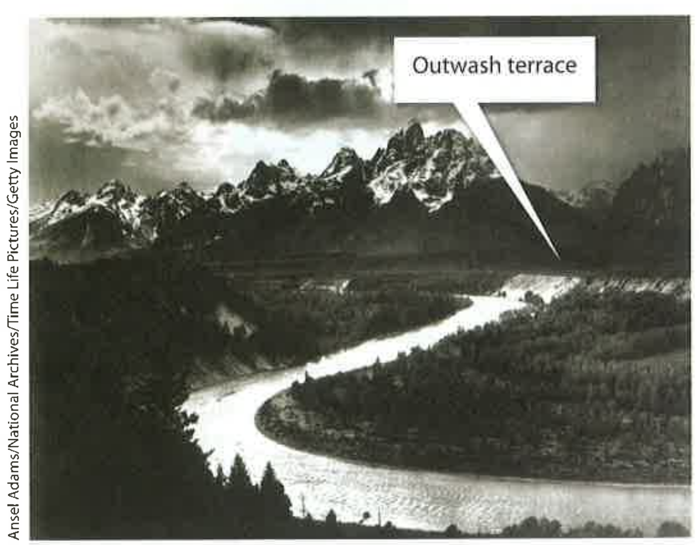
critical to any climatic interpretation of river terrace sequences and knickzones.

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.

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,

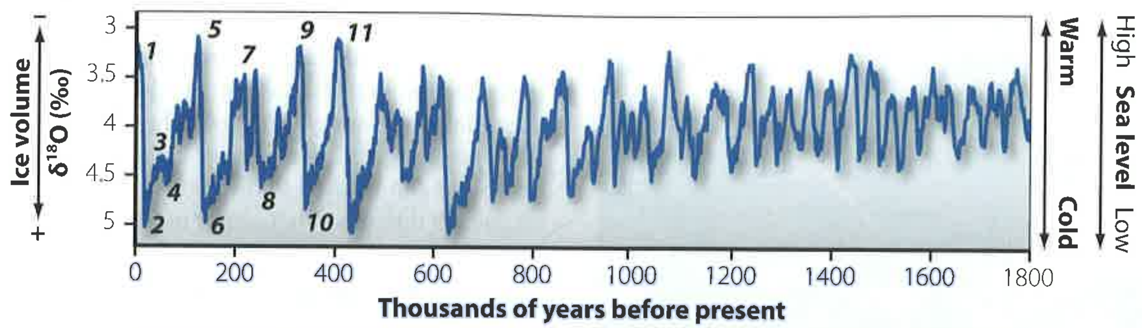


Ansel Adams/National Archives/Time Life Pictures/Getty Images

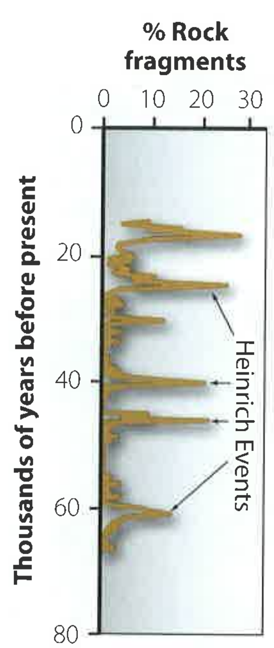
PHOTOGRAPH 13.3 Origin of River Terraces. Some flights of river terraces are climatic in origin. Here, along the Snake River, outwash terraces at Deadman's Bar are composed of material that came from the termini of outlet glaciers of the Southern Yellowstone–Absaroka Mountains ice cap. The upper terraces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

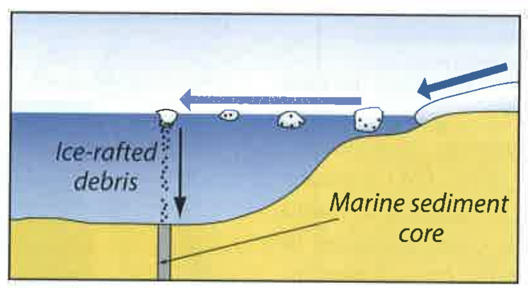
A. Noren

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.





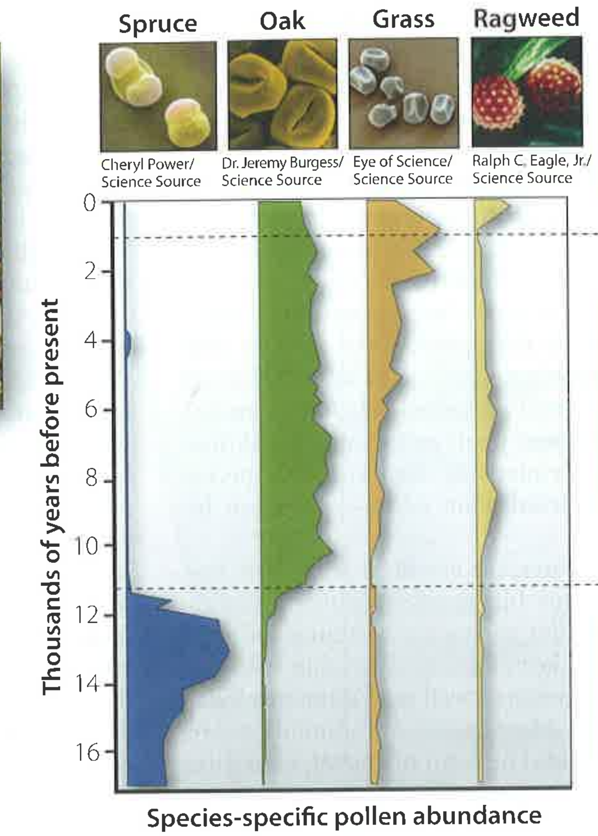
When glaciers calve and shed icebergs, those icebergs float out to sea; as they melt, they drop **ice-rafted debris (IRD)** onto the ocean floor. Marine cores sample this sediment and can be analyzed to determine how sediment grain size changes over time. The peaks in debris concentration identify times 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.

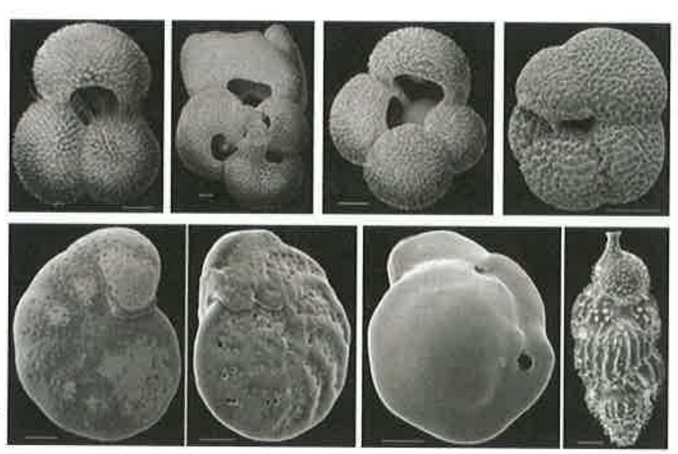
Pollen analysis is one of the most widely applied tools in paleoclimate research. The morphology or shape of pollen grains released by plants is unique for each genus [Figure 13.3] and pollen is durable and thus well-preserved. Qualitative pollen analysis is straightforward. If the pollen of cold-tolerant 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



R. M. Leckie

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.

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

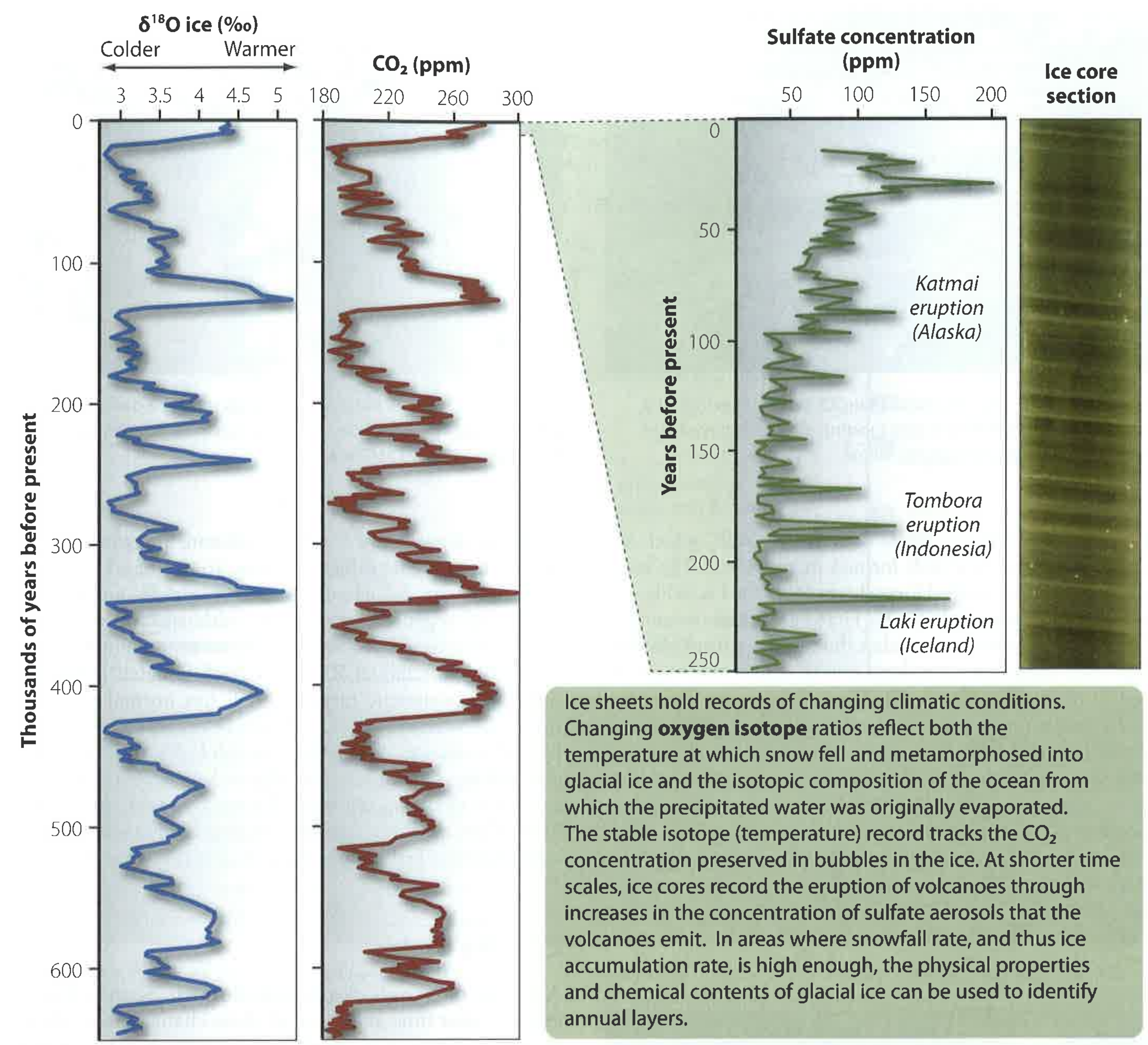


FIGURE 13.4 Climate Records Preserved in Glacial Ice. The stable oxygen isotope composition, CO_2 concentration, and sulfate concentration are used to interpret paleotemperature, atmospheric composition, and the history of volcanic eruptions. [Data from Siegenthaler et al. (2005).]

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.

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.

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.


M. Smith, T. Stevens

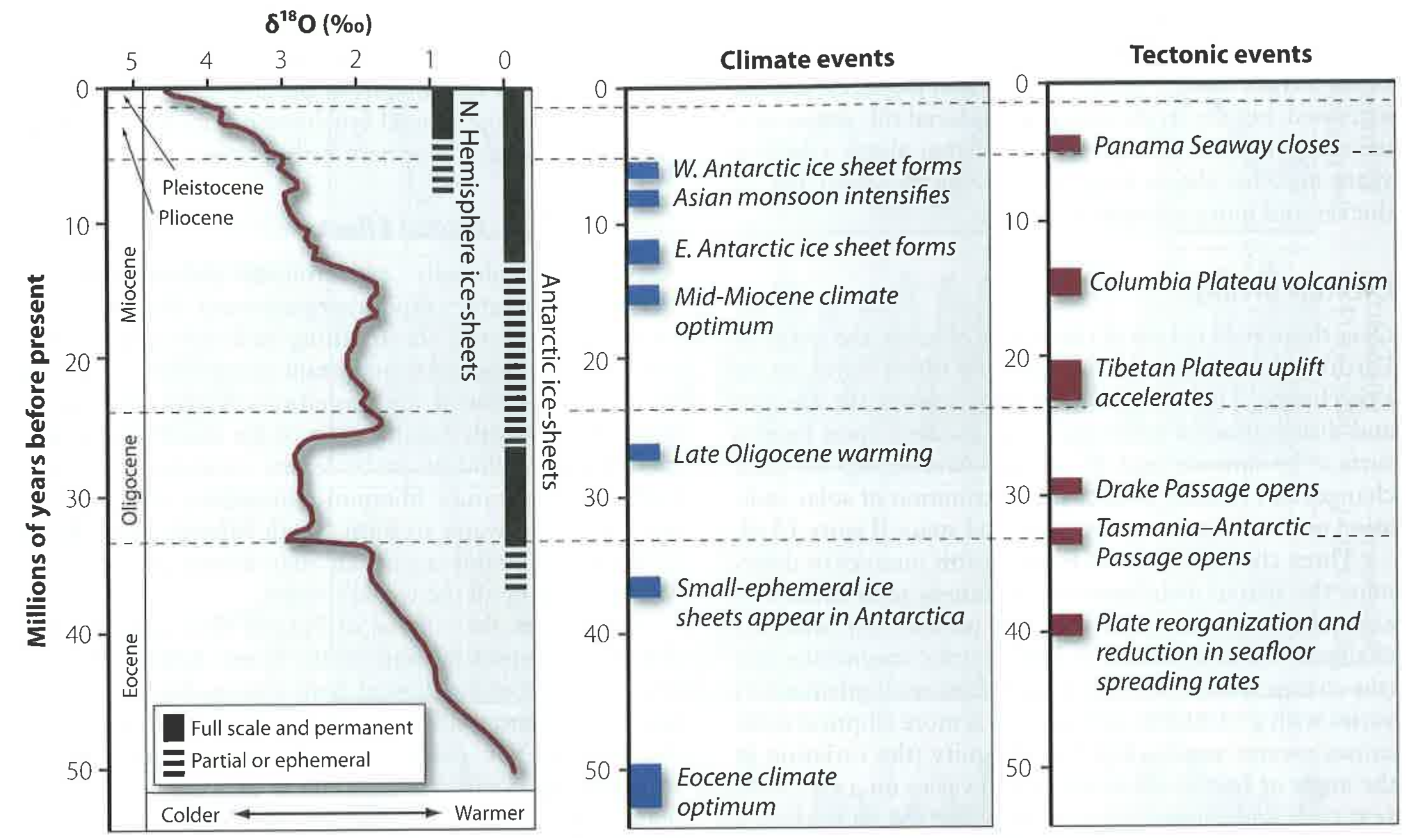
PHOTOGRAPH 13.8 Paleosol. Scientists sample a dark red paleosol (upper section) developed on tan loess (lower section) in the Loess Plateau of China.

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.

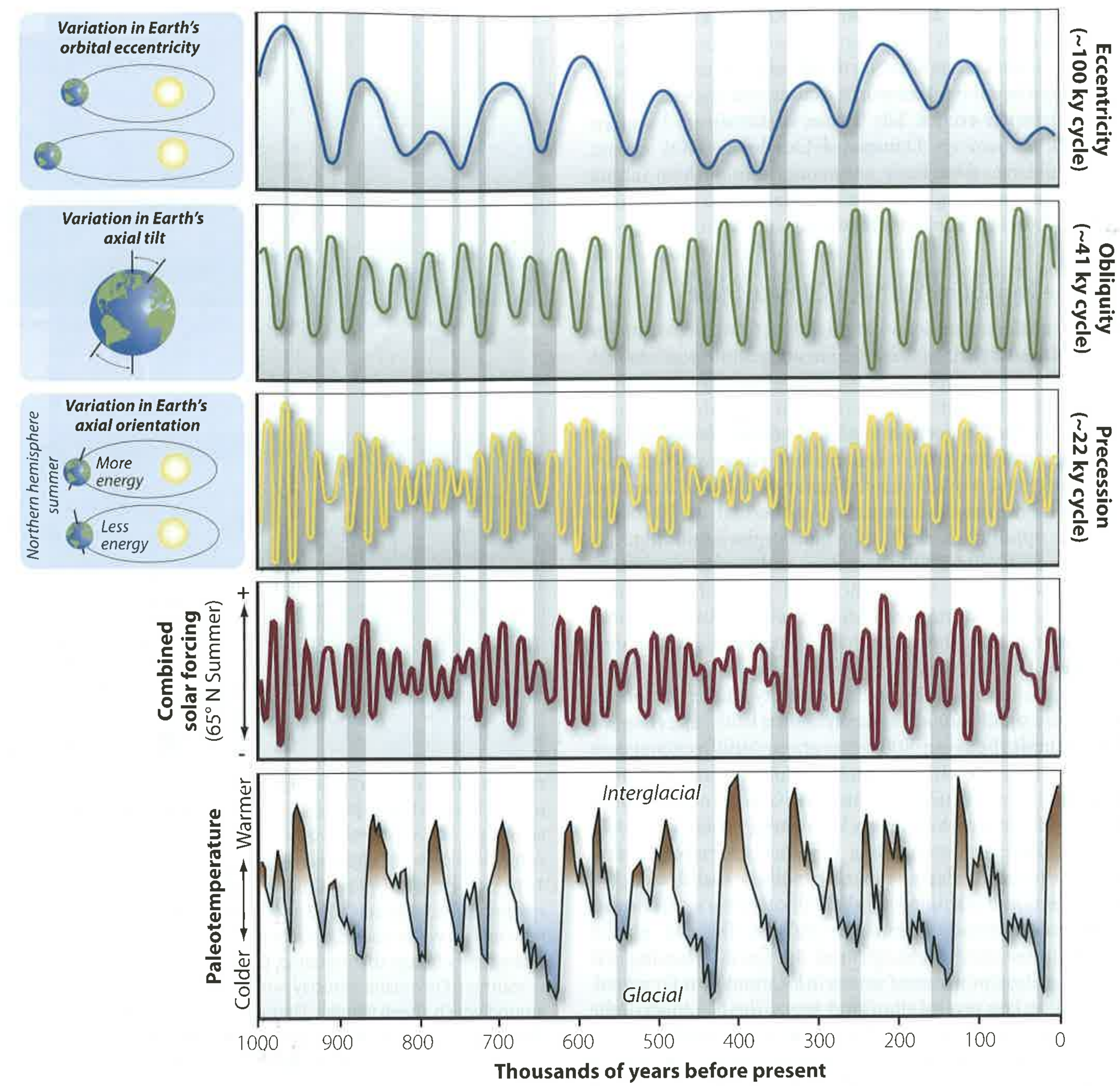
Understanding exactly how changes in solar radiation (specifically at high northern latitudes where ice sheets are born) translate into planet-wide glaciations, interglaciations, and rapid changes in paleotemperature has not come easily and has revealed numerous interactions, feedbacks, and resulting 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



The geometry of Earth's orbit around the sun varies in a regular and predictable fashion with three main orbital parameters (**eccentricity, obliquity,** and **precession**) changing on 100,000-, 41,000-, and 22,000-year cycles, respectively. Together, these orbital parameters act to change the seasonal distribution of solar radiation (energy) incident upon Earth. When little solar energy is incident on high northern latitudes, some of the snowpack survives the summer melt season and begins to grow an ice sheet. The solar forcing is smooth and symmetrical, but the paleotemperature record has many fewer variations of greater amplitude. This reflects ocean/atmosphere interactions and amplification of the solar forcing.

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

(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.

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**].

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.

Precipitation and Temperature

Linked changes in precipitation and temperature over time control the availability of water on the landscape and thus the frequency, magnitude, and rate of many surface processes. For example, increases in temperature at the end of the Pleistocene in North America's desert southwest reduced effective moisture and vegetation cover leading to the generation of less runoff. In some cases, this drying appears to have decreased sedimentation on alluvial and debris fans. In other cases, increasing aridity and the concomitant loss of soil-anchoring vegetation appears to have increased erosion and sediment transport from slopes. Unfortunately, at most arid-region study sites, dating of erosion and deposition episodes is imprecise enough (due to the lack of preserved organic carbon) that it is not possible to conclusively link climate changes and geomorphic response.

Climate also controls the intensity and spatial extent of large-scale weather patterns and storms. For example, **monsoons**, the seasonal reversal of winds and the resulting torrential summer rainstorms, are caused by a seasonal change in the location of the subtropical semipermanent high-pressure cells. Monsoon-associated precipitation tends to be best developed where elevated land masses heat up during the summer months, causing the air above them to warm, become less dense, rise, and pull moist air from a nearby ocean up steep topography where it cools, the moisture condenses, and torrential rains result. Such monsoons, including those in India, Australia, and southwestern North America, are geomorphically important, shaping channels, moving sediment off slopes, and carrying large sediment loads. During the last glacial maximum, monsoonal precipitation appears to have been less than current values, but monsoon strength increased in the early to middle Holocene. In the Himalaya, this heavy early Holocene precipitation caused channels to fill with alluvium, presumably because increased landslide frequency stripped soil off slopes more rapidly than the rivers below could remove the sediment.

A variety of information sources tell us that the frequency and intensity of hurricanes and coastal storms change over time. Instrumental and written records, including ships' logbooks, provide detailed information about storms that occurred during the past few hundred years. On a longer timescale, geologic records can be used to determine the timing and location of paleostorms that flood coastal regions and wash over barrier islands forcing salt water, sand, and marine organisms into coastal freshwater ponds and lagoons. In order to detect coastal paleostorms, geomorphologists analyze sediment from cores and outcrops. Coarse, sandy layers are indicative of storm surges, and the presence of saltwater organisms reveals flooding of coastal marshes and ponds by seawater. Such geologic detective work indicates that hurricanes were more frequent than today between 1000 and about 3400 years ago along the U.S. Gulf Coast and that storm intensity has increased over the past several hundred years along the coast of Maine.

Vegetation, Fire, and Geomorphic Response

Climate, vegetation, fire, and the rates and distribution of landscape-scale processes are tightly linked because climate, in particular moisture availability, temperature, and wind speed, determines the type and density of vegetation covering the landscape. Vegetation accelerates weathering and soil development through mechanical disruption, decomposition of leaf litter, the production of organic acids, and the acidification of regolith by CO_2 emitted during respiration (see Chapter 3). Roots also provide apparent cohesion, strengthening soil and holding it on hillslopes (see Chapter 5).

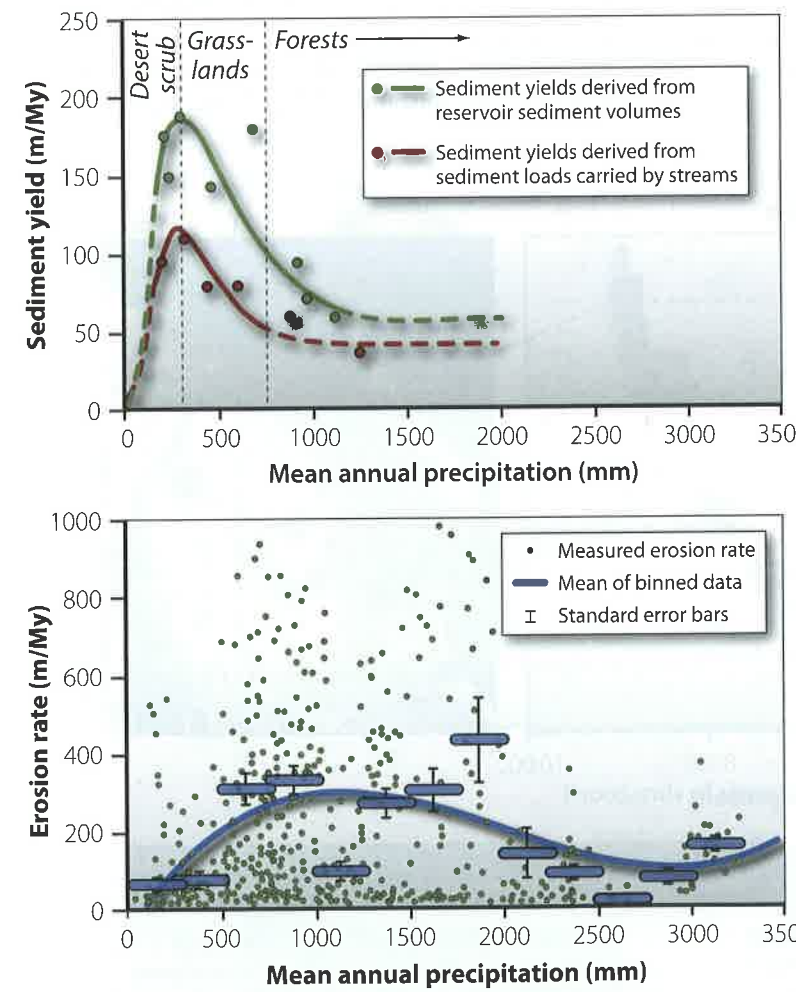
Fire is a geomorphic agent that can quickly remove vegetation, making slopes susceptible to erosion until vegetation regrows [**Photograph 13.11**]. Removing vegetation from the landscape reduces surface roughness and increases the speed and erosivity of overland flow. Fire kills trees and over time the roots from fire-killed trees rot, reducing effective soil strength. In some ecosystems, such as the chaparral of southern California, fire releases hydrophobic (water-repellent) compounds from the plants, coating the soil. This coating diminishes infiltration, increasing the risk of erosive runoff and the chance of generating debris flows when heavy rain strikes.

PHOTOGRAPH 13.11 Post-fire Erosion. After the great Yellowstone forest fires of 1988, a single thunderstorm that dumped rain on burned slopes caused rilling and gullying in Madison Canyon, Yellowstone National Park, Wyoming.

Fire can act as a direct geomorphic agent. The heat of fire rapidly expands the surface of rocks; yet the interior of rocks, especially those exposed to rapidly moving fires, remains cool due to rock's low thermal conductivity. This differential heating and expansion results in significant stresses within fire-heated boulders and outcrops. These stresses crack and erode exposed rock over time (see Photograph 3.6).

Climate determines the type and density of vegetation that can grow on a landscape and thus the fuel load; it also determines the moisture content of that fuel and its susceptibility to ignition during dry seasons. Climate strongly influences the frequency of fire and the resulting geomorphic effects, particularly the generation of debris flows and the deposition of debris flow material in fans at the base of slopes and its subsequent transport through river systems. Periods of aggradation in the past, dated with charcoal buried in debris flow sediment, correlate strongly to times when climate was warmer and drier, such as the Medieval Warm Period (900–1300 CE). Today's warming and drying climate in western North America, as well as heavy fuel loads due to a century of wildfire suppression, mean that the next century likely will see more fires and increased geomorphic activity, specifically hillslope erosion, debris flows, and fan aggradation in areas affected by fire (Photograph 13.11).

Sediment yield is the amount of sediment transported out of a drainage basin over time. It reflects both the rate at which the basin erodes and the efficiency of sediment transport mechanisms that remove sediment from the basin. Early measurements, such as those plotted in the classic Langbein-Schumm curve [**Figure 13.7**], showed maximum sediment yield from catchments located in semi-arid climates. This curve was created using data (mainly from the tectonically inactive central United States where drainage basins are underlain by sedimentary rocks) collected from



The Langbein-Schumm curve relates **sediment yield** (the amount of sediment leaving a watershed as determined by either suspended sediment load or reservoir sedimentation over a known time period) to mean annual precipitation. Langbein and Schumm found that semi-arid drainage basins had the highest sediment yields. The data for this compilation were collected from central North America and represent sediment transport during a period when people were altering land use through agriculture.

A compilation of over 1000 measurements of ^{10}Be made in river sediments around the world shows great variability in long-term erosion rates for similar amounts of precipitation (green dots are the data). On average, arid regions (<500 mm/yr of rainfall) had low rates of erosion, humid regions eroded more quickly, and erosion rates were lower in areas where mean annual precipitation was very high, perhaps reflecting the importance of vegetation stabilizing slopes.

FIGURE 13.7 Sediment Yield, Erosion, and Climate. Sediment yield data and cosmogenic estimates of long-term, drainage-basin scale erosion rates offer different views on how mean annual precipitation affects erosion and sediment yield. [Data from Langbein and Schumm (1958) and Portenga and Bierman (2011).]

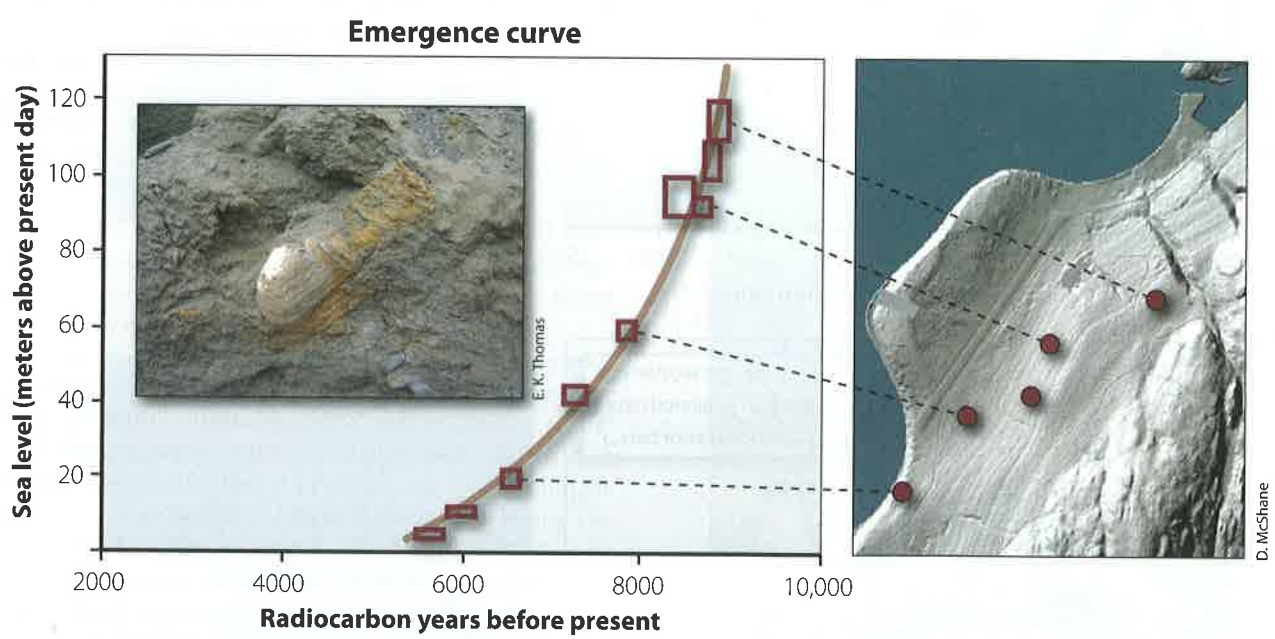
170 gauging stations as well as reservoir sedimentation data; thus, the sediment yield data reflect both human impact and a short (years to decades) integration time. One plausible explanation for the humped shape of the resulting curve is that arid climates are so dry that sediment yield was limited by rainfall and runoff (transport-limited) even though there was little if any vegetation to hold regolith in place. In humid climates, vegetation anchored sediment making the slopes supply-limited. In semi-arid climates, the sparse vegetation provided little erosion resistance and intense thunderstorms produced erosive runoff.

In contrast, a 2011 compilation of more than 1000 measurements of the cosmogenic nuclide ^{10}Be in fluvial sediment from around the world shows that long-term, basin-scale erosion rates vary up to two orders of magnitude even if the amount of precipitation is similar. The difference between the global isotopic estimates of erosion rate and the Langbein-Schumm compilation of predominantly central United States sediment yield data likely reflect a combination of the major influence of global variability in tectonics and lithology, as well as differing integration time and the influence of human impacts on short-term rates of sediment yield. Such a comparison points out the importance of understanding all of the variables that influence both the rates of sediment generation and the rates of sediment delivery from a watershed.

Base Level

Sea level, the base level for the world's major rivers, repeatedly fell more than 100 m between maximum interglacial and glacial stages, reflecting changes in the volume of ocean water as ice sheets advanced and retreated. Volumetric control on sea level is termed **eustatic** and includes not only sea-level fluctuations driven by the expansion and contraction of ice sheets and the amount of water they hold on the 1000-year to 10,000-year time scale but also the thermal expansion and contraction of ocean water driven by changes in global temperature. Streams respond to lower sea level by incising. Incision propagates upstream, rapidly removing unconsolidated alluvium and triggering the formation and migration of knickzones in bedrock channels. As sea level increases, base level rises, triggering aggradation, initially in areas nearest the sea, but then spreading upstream a limited distance like a wave.

Incision and aggradation can occur simultaneously on different parts of the landscape (see Figure 12.14) because the effects of base-level fall and the incision it causes take time to translate up a drainage network. Given enough time, the effects of a major base-level fall will eventually propagate up to the head of the channel network (and onto the hillslopes) through the processes of knickzone retreat, channel incision, and hillslope lowering by erosion.



As an ice sheet shrinks, the land begins to rebound as the load on Earth's mantle is reduced. At the land-sea interface, **wave-cut terraces** can form from large storms and beach sediments are deposited. If the material in terraces of known elevation can be dated (usually using ^{14}C dating of marine shells or driftwood), an **emergence curve** may be constructed that represents local sea level over time. Uplift can result in stacked flights of terraces rising above the modern shoreline.

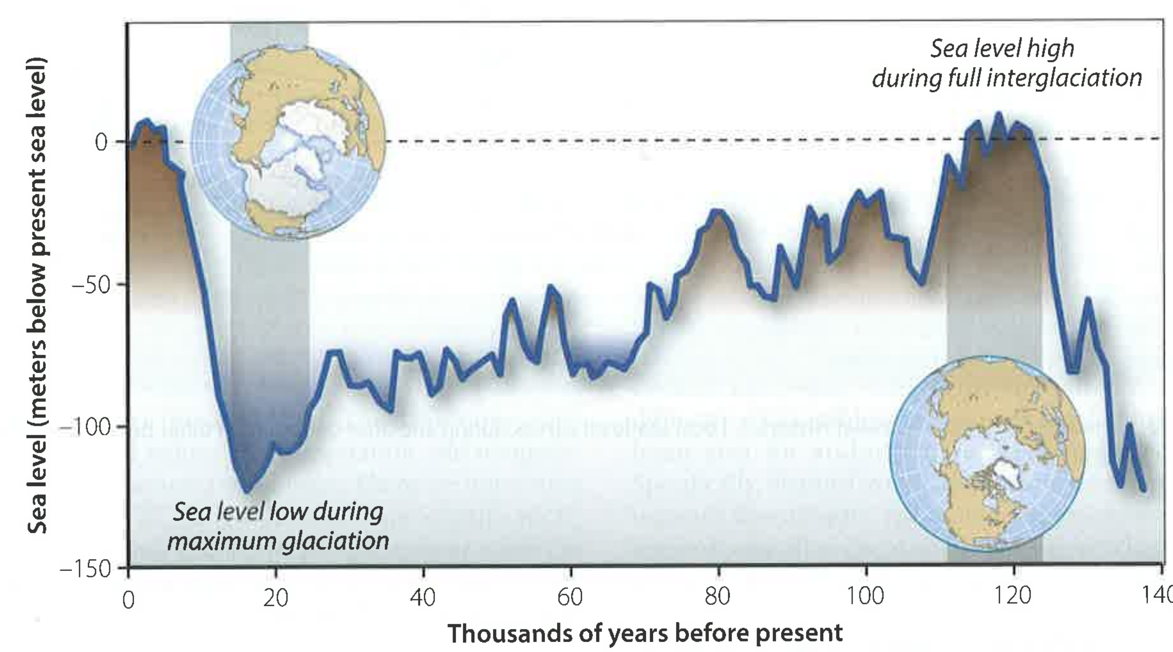
FIGURE 13.8 Emergence Curves. Local emergence curves describe the change in sea level at one place on Earth's surface reflecting both eustatic sea-level change and the isostatic response of the solid Earth. [Adapted from Ten Brink (1974).]

The passage of knickzones and the lowering of local base level will increase local gradients and erosion rates, changing the amount of sediment delivered to the channel network. In contrast, the effects of a base-level rise are limited to estuarine and lowland river systems where channels will flood and sediment will aggrade. Prime examples of such aggradation include the drowned landscape of the Maine coast and the distal end of lakes in previously glaciated areas where postglacial, isostatically driven tilting raises land closest to the ice sheet, drowning stream and river valleys farther away from the ice.

Paleo–sea levels are a challenge to determine because it is difficult to find a stable frame of reference as both sea level and land level change over time. At many sites around the world where the land is rising relative to the sea, geomorphologists have created **emergence curves** (graphs of land level over time) by dating shoreline features and measuring their elevation [**Figure 13.8**]. Dating usually relies on ^{14}C analysis of either driftwood found on raised beaches or fossils found in marine sediment underlying wave-cut terraces. Differences in the amount and timing of region-specific isostatic responses cause local sea-level histories to differ, sometimes markedly, from global **eustatic sea-level curves,** which reflect primarily the volume of water in the world's oceans [**Figure 13.9**].

Consider the northeast coast of North America, which for the most part was overrun by ice and isostatically depressed at the last glacial maximum. As the ice melted, and the volume and area of the Laurentide Ice Sheet diminished, the load on the crust decreased and the land rebounded. As long as the rate of local rebound exceeded the rate of eustatic sea-level rise from worldwide melting of glacial ice, land emerged from the sea [**Figure 13.10**]. However, when global sea-level rise (eustatic) later became faster than the local rate of isostatic rebound, then the land that initially emerged isostatically was resubmerged by eustatic sea-level rise. The land reemerged again above sea level as isostatic response, though waning, continued after the eustatic sea level stabilized about 8000 years ago, because all of the Northern Hemisphere ice sheets, except Greenland, had melted.

Such complex emergence scenarios are more likely if the ice readvances during periods of rapid eustatic sea-level rise. Indeed, this is the case in Greenland, where the ice margin in some places retreated perhaps kilometers in the mid-Holocene only to readvance in the last few thousand years. This readvance caused mantle displacement and isostatically driven crustal subsidence of a meter or two. This subsidence has left a clear stratigraphy in near-shore deposits. There, one finds saltwater-tolerant



As ice sheets grow, they lock up water that otherwise would fill the ocean basins. During glaciations, sea level falls proportional to the amount of water stored in ice caps and glaciers. During interglaciations, sea level rises as ice caps and glaciers shrink and meltwater returns to the oceans.

FIGURE 13.9 Glacial-Interglacial Eustatic Sea-Level Changes. During the past few million years (the Pleistocene), global, or eustatic, sea-level change reflects the amount of water held in ice sheets. The last glacial–interglacial cycle saw sea level change 130 m over 100,000 years.

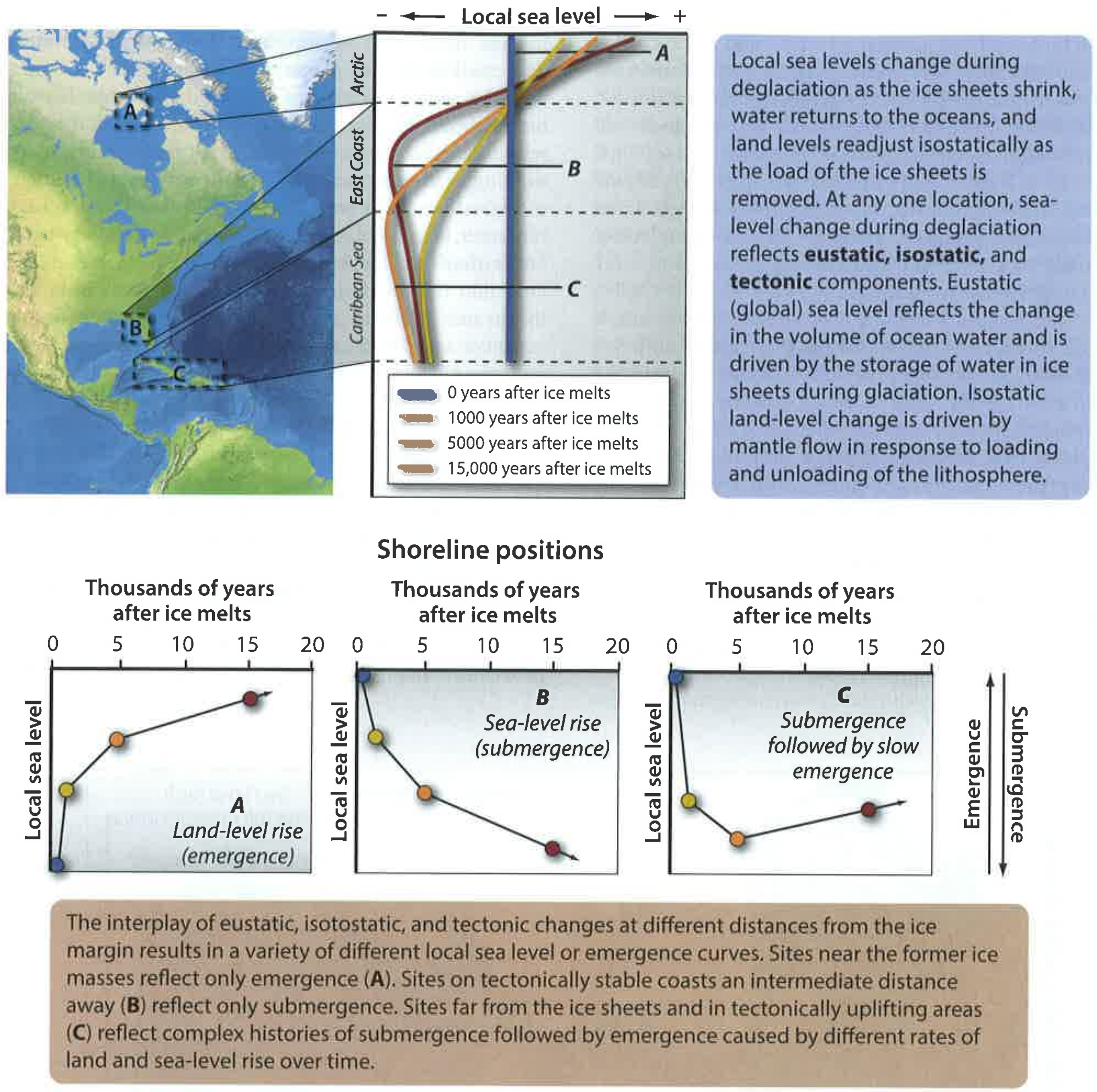


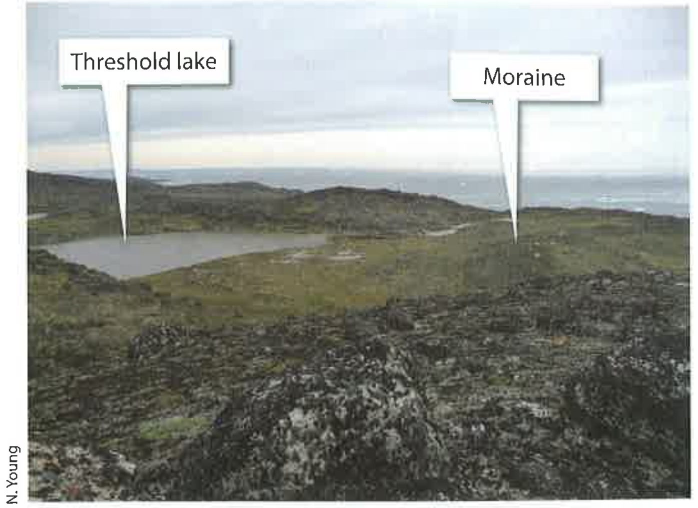
FIGURE 13.10 Variability of Local Sea-Level Histories. Local sea-level curves during and after deglaciation differ between locations. [Adapted from Clark et al. (1978).]

flora and fauna deposited over the remains of freshwater vegetation.

The advance of glaciers in response to small changes in climate can be detected not only from moraines but, if the geometry of the landscape is right, in the sediment of **threshold lakes.** A threshold lake typically receives sediment only from its watershed until a tongue of glacial ice crosses a topographic threshold and begins to deliver glacial sediment directly to the lake [**Photograph 13.12**]. When this happens, the character of the sediment changes as the concentration of fine, glacial sediment (glacial flour) increases manyfold. Dating the first and last occurrence of this glacial sediment can accurately delineate the timing of glacial advance and retreat.

Climatic Geomorphology

The relationship of some landforms to specific climate zones led to the development of **climatic geomorphology,** a school of thought that suggested climate primarily controlled the distribution and shape of many landforms. Taken at face value, such a suggestion makes sense because temperature, precipitation, and wind regimes



N. Young

PHOTOGRAPH 13.12 Threshold Lake. In Tasiussaq, west Greenland, is a lateral moraine with a threshold lake to the left that collected meltwater and sediment from the ice when the glacier deposited the moraine.

(their averages, ranges of variability, and frequency and intensity of variation) define climate and the resulting vegetation cover that shape landforms through the action of surficial processes. In practice, climatic control on landforms is not always clear-cut because some landforms occur in multiple but different climate zones, and the influence of tectonics, especially in terms of base-level control, can make it difficult to isolate the influence of climate.

Köppen Climate Classification

The most widely used classification of climate zones was proposed by climatologist Vladimir Köppen in 1884 and, although it has been revised several times since, still bears his name. Köppen classified the world's land areas into five major climate zones: equatorial, arid, warm temperate, snow, and polar [**Figure 13.11**]. The definitions were based on measured values for precipitation and temperature as well as vegetation assemblages. There are numerous climatic subzones in the Köppen climate classification, reflecting more specific climate phenomena or controls such as summer drought and proximity to the coast and coastal moisture.

The distribution of climate zones roughly follows latitude (see Chapter 1), although there are clearly relationships with elevation and location downwind of major mountain chains. Different climate zones are distinguished by a variety of characteristics. For example, areas in the dry (arid) climate zone have significant moisture deficits—evaporation exceeds precipitation. Cold temperatures characterize polar regions. In temperate regions, precipitation exceeds evaporation, leading to a positive moisture budget so that runoff is routinely generated.

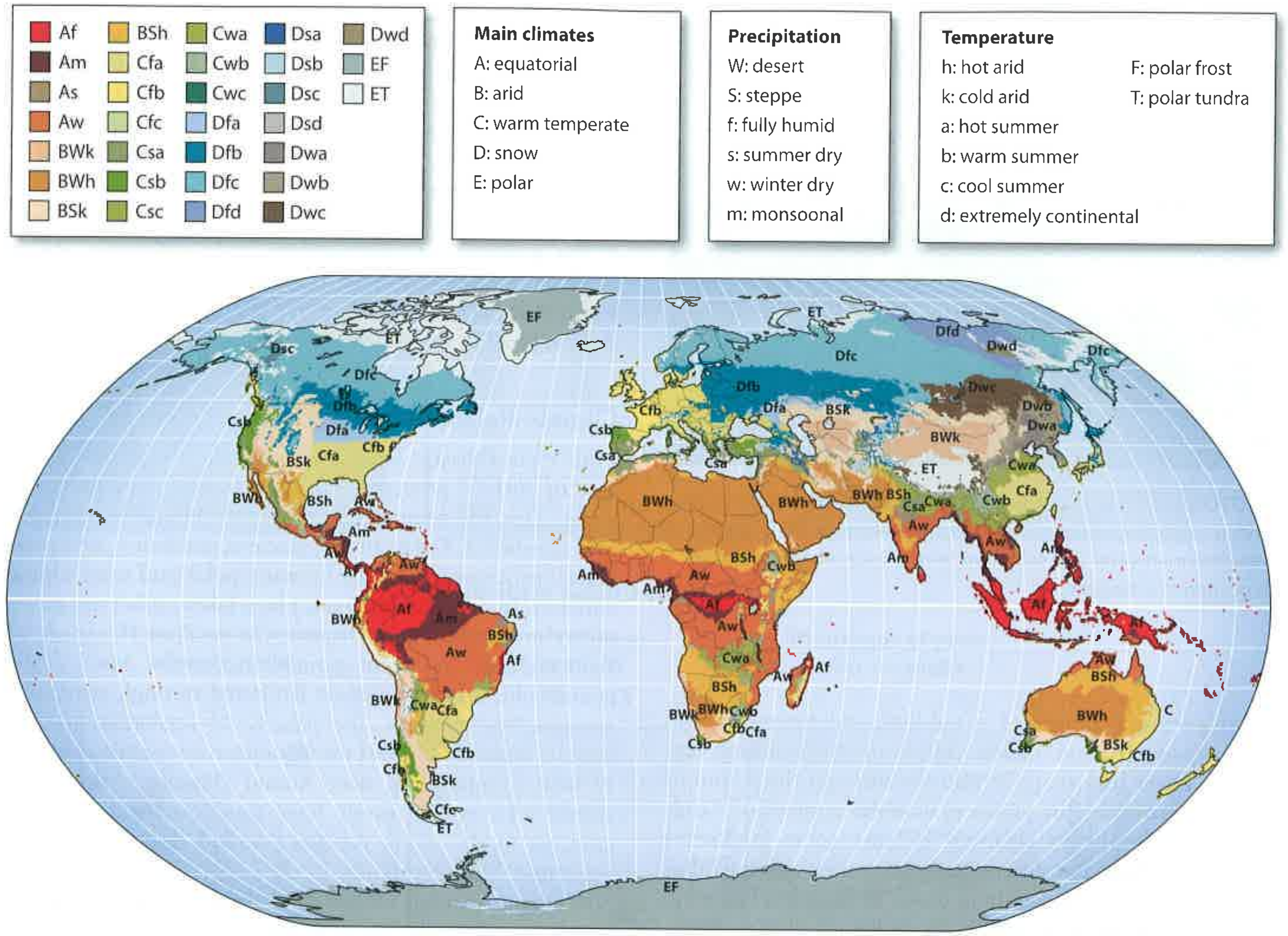
As climate changes, so does the location of climate zones and their boundaries. Paleoclimatological studies using pollen and macrofossils document past shifts in climate zones. Contemporary climatological data allow mapping of climate zone shifts over the past century while computer models, based on estimates of greenhouse gas–driven climate change, predict the location of climate zones well into the future. Shifting climate zones mean changes in the intensity and duration of hydrologic phenomena (such as rainfall and runoff) and thus changes in the rate and distribution of Earth surface processes including landsliding, sheetwash, and the generation of debris flows.

Climate-Related Landforms and Processes

Both characteristic landforms and the distribution and rate of surface processes are often related to climate. In some cases, the geomorphic effects of climate are obvious and significant. For example, glaciers, glaciation, and periglacial processes dominate in many polar and some alpine climates. In dry climate zones, playa lakes and desert pavements are common landforms and aeolian transport of sediment is an important geomorphic process. Much of the geomorphic effect of climate is filtered through vegetative systems. For example, the paucity of vegetation in arid regions does not support the significant root reinforcement of soils common in more humid climates. Arid-region slopes >15° to 20° rarely have significant soil cover—a contrast to the steep, yet soil-mantled slopes of heavily vegetated, humid uplands.

Climate strongly affects the behavior of hillslopes and rivers. For example, rivers in dry climate zones flow only intermittently. Without the roots of vegetation to stabilize their banks, arid-region channels often shift large distances laterally during rare but high-discharge floods. Bankfull flows are much rarer events in arid than in humid regions. Flows often decrease downstream in arid-region streams as water flowing downstream from higher, wetter headwaters infiltrates into the bed of the channel or into alluvial fan surfaces (see Figure 4.12). This infiltration results in very different relationships between channel dimensions and basin area for arid-region versus humid-region streams. Specifically, channel width, depth, and cross-sectional area increase downstream for humid-region channels (gaining streams) and often decrease for arid-region channels (losing streams) in which the volume of flow diminishes with distance from wetter highlands.

Slopes in arid regions are often weathering-limited and thus bare rock outcrops are common with little soil or regolith cover. This leads to rapid runoff and "flashy" hydrologic responses because infiltration rates are low and there is little if any shallow subsurface flow. In the wet tropics, warm temperatures and abundant moisture catalyze deep weathering that produces thick, clay-rich saprolites and lateritic soils—conditions conducive to gullying. In contrast, well-developed soils in temperate climates support large trees, the roots of which bind soil together



The Köppen system is the most widely applied classification of climate zones, which some geomorphologists believe can be associated with predictable sets of landforms and active surface processes. The system uses a three-letter system to classify zones according to a main climate, precipitation, and temperature. Polar regions are characterized by two letters showing whether they are tundra or ice. Climate zones are based on the annual amount and seasonal distribution of precipitation as well as the range and average of temperatures. The main climate zones reflect global wind patterns and the location of major mountain ranges.

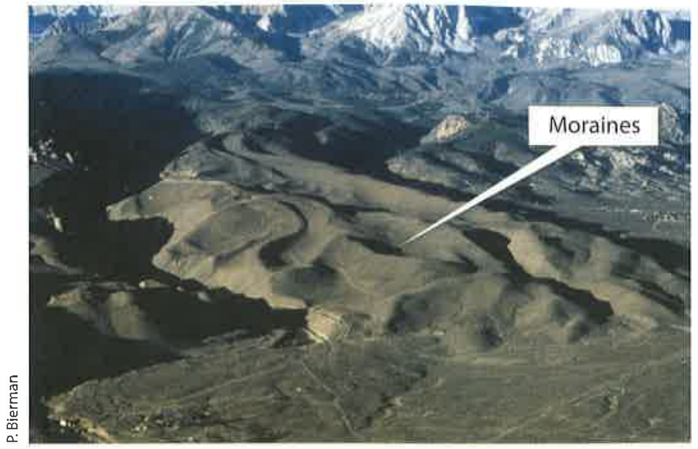
FIGURE 13.11 Köppen Climate Zones. Climate zones reflect the global distribution of precipitation and temperature.

on steep slopes and increase porosity when they die and rot; these roots thus help prevent mass movements until the exceptional storm strikes.

Relict Landforms

Regular and significant changes in Earth's climate over the past several million years have left behind **relict landforms**, those that that are no longer forming or being maintained by active geomorphic processes—for example, moraines on the warm, semi-arid floor of Owens Valley in southern California [**Photograph 13.13**] or in southern Connecticut, places where ice would never survive a summer today. Many geomorphologists devote much of their careers to interpreting the paleoclimatic significance of relict landforms.

Many relict landforms are either directly or indirectly related to the comings and goings of glaciers and ice sheets. These include moraines that mark the extent of former glaciers, glacial outwash deposits along what were once meltwater streams, and periglacial features that reflect the presence of now-vanished permafrost. In mid-latitudes, dunes stabilized by vegetation are common, the result of increased moisture, lower sediment supply, lighter winds, and more vegetation during the Holocene than the Pleistocene.



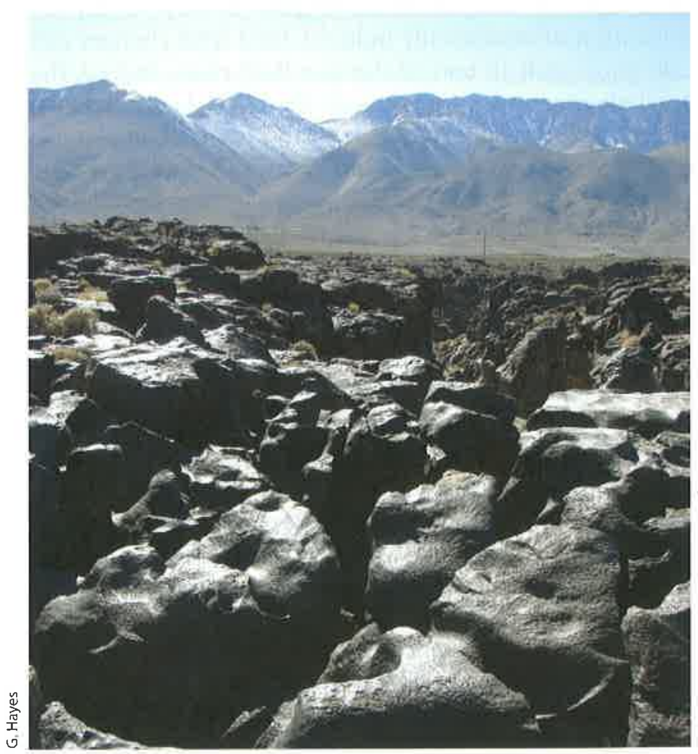
P. Bierman

PHOTOGRAPH 13.13 Moraines. Here, on the western side of semi-arid, temperate Owens Valley, California, are a set of moraines. The Sierra Nevada, from which the glaciers that deposited the moraines originated, are visible in the distance.

P. Bierman

PHOTOGRAPH 13.15 Pluvial Lakes. Series of pluvial (glacial period) wave-cut shorelines and terraces (extensive horizontal surfaces) in the Great Salt Lake Basin, Utah, record falling lake levels after the last glacial maximum.

In the desert southwest of North America, there are clear, relict geomorphic indicators of a much wetter time. For example, in the Mojave Desert, where today only tens of millimeters of rain fall annually, in the Pleistocene there were a series of **pluvial lakes** connected by rivers. The extent of these lakes is demarcated by now-dry beach ridges, fossil waterfalls [**Photograph 13.14**], and erosional shoreline notches cut into rock by pounding waves as well as terraces



G. Hayes

PHOTOGRAPH 13.14 Fossil Falls. Fossil falls in southeastern California have not always been dry. Waters from Pleistocene Owens Lake overflowed the basalt during the ice ages, polishing and fluting the rock surface.

deposited at the lake margin [**Photograph 13.15**]. Today's Great Salt Lake is just a shadow of its former self. During glacial times, its surface area was >10 times larger than today. That expanded lake, known as Lake Bonneville, left many distinct, relict shoreline features including spits and bars now sitting high and dry above desert valley bottoms. The drying of desert lakes after glacial times reflects a changing water balance across the landscape. In some cases, the change resulted primarily from decreased precipitation. In other cases, lake contraction was driven mostly by increased evaporation, the result of warmer air temperatures and less cloudiness.

Landscape Response to Climate

Landscapes respond to changing climate because the type, rate, and spatial distribution of active geomorphic processes, such as weathering, erosion, and sediment transport, depend to a large degree on climate.

Glacial–Interglacial Changes

Over the past few million years, the most important climate changes have been regular shifts between glacial and interglacial climates. Occurring on regular cycles taking many tens of thousands of years, the climate cooled slowly as ice sheets thickened and covered large areas in the Northern Hemisphere. Then, in comparatively rapid terminations lasting about 10,000 years, these ice sheets collapsed, melting back rapidly as the climate warmed. Large areas of Earth's surface were exposed from under ice, ecosystems were reestablished, and nonglacial surface processes began remolding the glacial landscape.

The effects of glaciation extended far beyond the borders of the ice sheets. Immediately adjacent to the ice, fierce **katabatic winds**, spawned by cold, dense air draining off the ice, whipped loess into the sky from barren outwash plains along raging meltwater streams. In many areas where today's temperatures are warm (such as the southern Appalachian Mountains and parts of Africa and Australia), there are distinctive clues indicating that the climate was once much colder, with mean annual air temperatures sufficiently cold (below freezing) to form ice wedge casts, patterned ground, rock glaciers, block fields, solifluction lobes, and even thermokarst and relict pingos (see Chapter 9)—all indicating a permafrost-dominated landscape [**Photograph 13.16**].

The advent of glaciation about 2.7 million years ago appears to have been geomorphically significant globally, even in places far away from ice margins and isolated from the effects of glacially induced rises and falls of sea level. Throughout the world, more and coarser sediment appears in the geologic record between 2 and 4 million years ago. Some think this change in sediment character and abundance reflects an increasingly variable climate that destabilized landscapes and removed readily erodible regolith generated by long periods of weathering; others suggest that a cooler, drier climate reduced vegetation cover, increasing erosion.

The effect of glaciation does not end when the ice melts away but continues as nonglacial surface processes modify glaciated landscapes. **Paraglacial sedimentation** refers to the material deposited as a result of the disequilibrium landscapes left after deglaciation. For example, steep cliffs

PHOTOGRAPH 13.16 Ice Wedge Cast. An ice wedge cast, evidence of permafrost conditions in New England, disturbs well-stratified outwash sand and gravel in Connecticut. The ice wedge likely formed just after deglaciation while the climate was still very cold.

that characterize deeply incised glacial valleys rapidly shed sediment after the buttressing ice melts away, generating large talus slopes at their base. Steep, unvegetated slopes of till erode rapidly after deglaciation, creating fans and colluvial deposits. Permeable glacial deposits, such as outwash gravel, are transmissive to groundwater while less-permeable, glacial lake sediment functions as an aquitard, limiting flow. When exposed on steep slopes, this massive, fine-grained lacustrine sediment often fails in large rotational landslides.

Isostatic Responses

Climate change causes surface processes to redistribute mass across Earth's surface. This redistribution of mass drives isostasy, a solid-Earth process, which then affects surface processes for millennia after glaciation has ceased. For example, the growth of large glaciers and pluvial lakes adds mass to the crust and the upper mantle beneath responds, slowly flowing away from the load. The lithosphere then deforms, downwarping beneath the load and upwarping with much lower amplitude away from the load. This long-wavelength upward movement raises land tens to hundreds of kilometers away from the load forming what is known as the **forebulge**. As an ice sheet melts or a pluvial lake dries, the mantle, although quite viscous, responds to the decreased load. The land beneath the load rises, rapidly at first, and then more slowly over time. Farther away from the load, the forebulge relaxes and the once uplifted land subsides [**Figure 13.12**]. Complete relaxation of isostatically induced, land-level changes can take thousands to tens of thousands of years; indeed, the forebulge formed in response to the LGM Laurentide Ice Sheet is still slowly lowering and causing the gradual inundation of coastal Maryland and Virginia. A similar isostatic response happens near the coast as sea level rises and falls, changing the load on the continental shelf.

The magnitude of postglacial uplift can be deduced in a variety of ways. In recently deglaciated areas, such as the coast of Greenland, continuously recording high-precision GPS measurements document both the deglacial trend (the land is rising over time as the ice sheet shrinks) and the annual oscillation as mass (ice and snow) accumulates in the winter and melts away in the summer. Raised shorelines in Scandinavia and the Hudson Bay region of Canada testify to the land rising out of the ocean after deglaciation (Figure 13.8).

Ancient deltas, formed where rivers and streams entered standing bodies of water, such as large, glacially dammed lakes, can also be used as paleo-level lines. Presuming such deltas were deposited synchronously, their present-day elevation indicates the differential amount of isostatically driven uplift. Where the ice was thicker, deltas rose more after the ice melted away and today have higher elevations than deltas closer to the former ice margin, where ice was thinner and isostatic depression was less.

Shoreline and delta locations can be combined into maps and the uplift amounts contoured. These **isobases**,

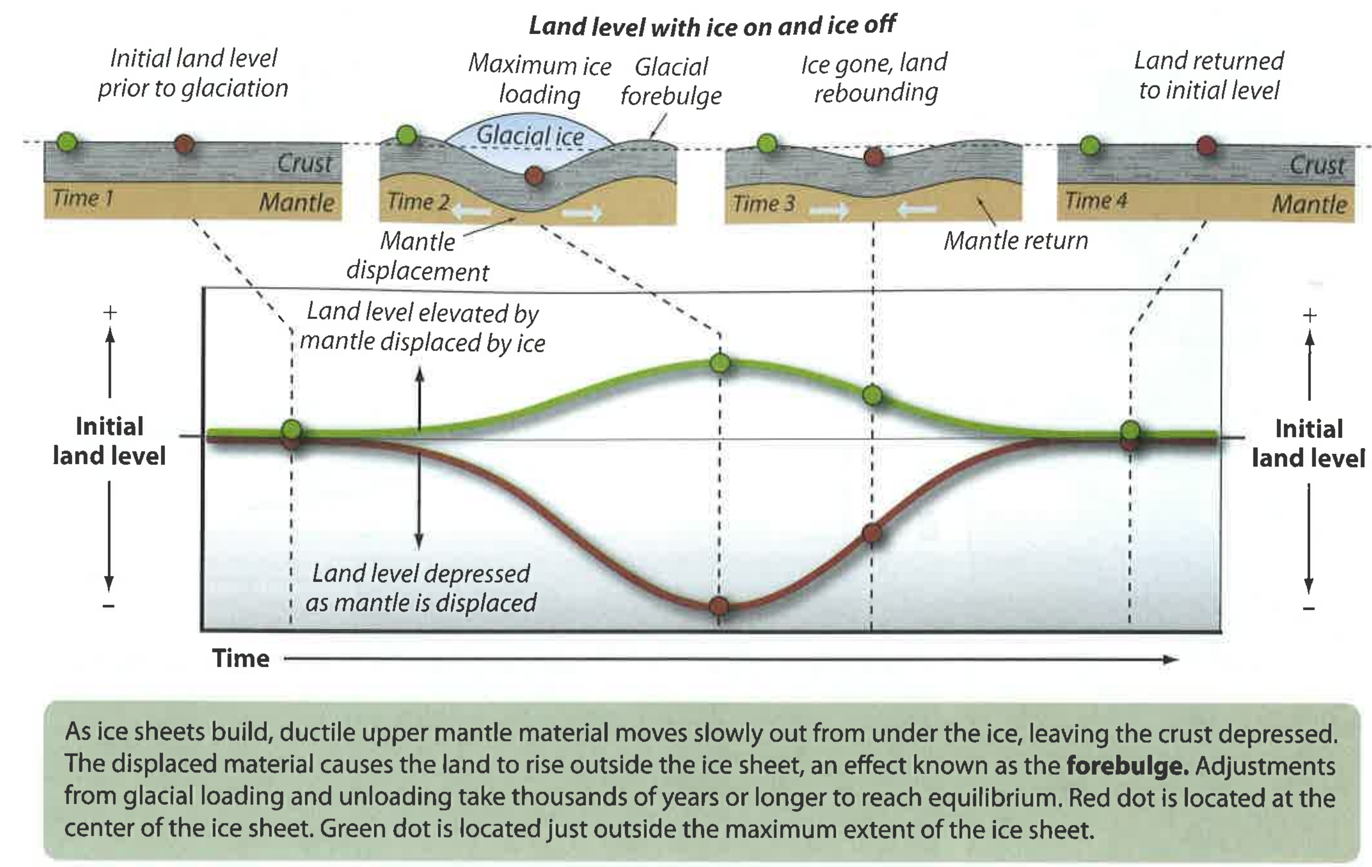


FIGURE 13.12 Land Levels Change in Response to Changing Ice Sheet Volume. During glaciation, land under the ice is isostatically depressed. Outside the ice margin, the land rises as mantle flow creates a forebulge.

or lines of equal uplift, can be used to infer the mass distribution of the now-vanished ice and the magnitude and orientation of the uplift field. Differential postglacial uplift can change the appearance of contemporary landscapes. For example, greater uplift in the northern Champlain Valley of Vermont and New York caused the Lake Champlain Basin to tilt southward after deglaciation, drowning river valleys in the southern part of the basin [**Figure 13.13**].

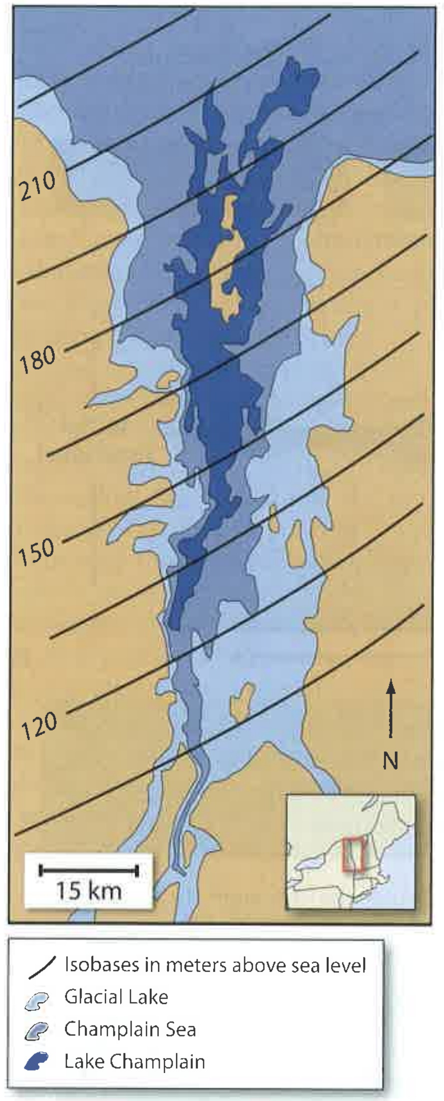
Climatic Control of Mountain Topography

One of the most striking examples of climate-landscape interactions on the scale of mountain ranges has been called the **glacial buzzsaw** [**Figure 13.14**]. In the Andes, the mountains that run along the west coast of South America, latitudinal trends in the **snowline** (the elevation above which snow survives through the summer, an approximation of the ELA) track variations in the height of the range. Such correspondence in this and other mountain ranges where peak heights rise a limited distance above the ELA has been interpreted as implying that the elevation of mountain ranges is kept in check by glacial erosion—the glacial buzzsaw cutting away at rock as valleys are excavated and cirque headwalls cut back into the range. In other words, glacial erosion is able to match the rate of rock uplift.

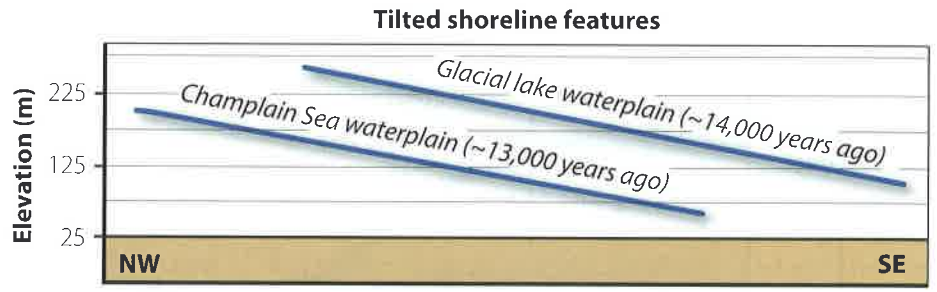
The amount of sediment shed from a continent can influence the behavior of a subduction zone and the shape of the resulting trench and mountain range. For example, the trench off the west coast of South America varies in depth depending on the amount of sedimentary fill originating from the uplifted western margin of South America. Where the climate is wetter, sediment yield is higher, more sediment loads the oceanic crust, and the trench is filled and bathymetrically shallow. Where the climate is very arid, less sediment is delivered to the trench, which as a result is less filled and deeper. Not only is the trench morphology affected by sediment yield but so is the morphology of the mountains. The Andes are wider where the climate is more arid. There are likely two reasons that the mountains are larger where it is drier. First, erosion rates are less in the hyperarid Atacama Desert than in wetter parts of the range. Second, the lack of soft, easily deformable sediment on the downgoing slab increases the shear stress between the downgoing plate and the oceanic crust; thus, the plate boundary, because it is stronger, can support the load of more massive mountains.

Climate Change Effects

Glacial-interglacial climate changes are large enough and long enough to cause considerable geomorphic effects



Isobases, or lines of equal uplift, are defined by measuring the elevation of shoreline features, here related to the glacial lake and later to marine waters that filled the Champlain Lowland between Vermont and New York. Paleoshoreline elevations (isobases, the water-surface plain, or level, of a now vanished body of water) increase to the northwest at a rate of about a meter per kilometer, reflecting differences in post-glacial rebound due to the thickness of the now-receded ice sheet.



These cross sections show shoreline features, such as beach ridges and deltas, that were once horizontal but are now at different elevations due to differential post-glacial uplift. These features, deposited at the same time in the glacial lake that once filled the Champlain Lowland, are now more than 100 m higher in the north than those farthest south. Not only is the paleoglacial lake water-surface plain tilted to the northwest but so are landforms related to the waterplain of the Champlain Sea that filled the Champlain Basin when eustatic sea level rose enough that ocean water, pouring through the Saint Lawrence River valley, flooded the Champlain Lowland replacing the glacial lake. Several thousand years of continued isostatic uplift then caused local sea-level fall and modern freshwater Lake Champlain was born about 10,000 years ago.

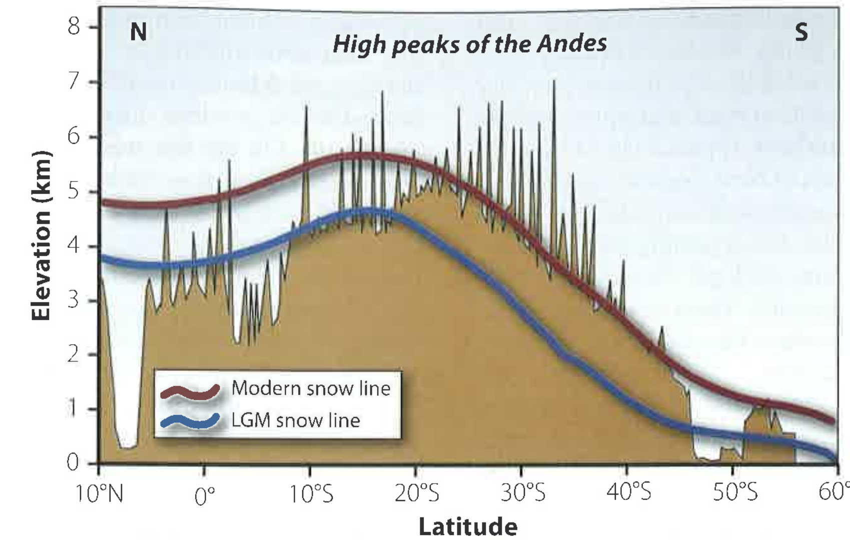
FIGURE 13.13 Uplift Triggered by Deglaciation. Mapping uplifted geomorphic features such as deltas deposited into now-vanished glacial lakes allows definition of isobases, contours of equal isostatic uplift. [Adapted from Chapman (1937).]

over much of Earth. However, geomorphic responses to changes in climate are not everywhere the same. Recent compilations of dozens of local temperature records indicate that temperature changes during and after the last glacial cycle were both broadly synchronous (peak of the last glacial maximum, ~22,000 years ago; peak of the Altithermal or the mid-Holocene warm period, ~7700 years ago) and similar in the Northern Hemisphere and Southern Hemisphere.

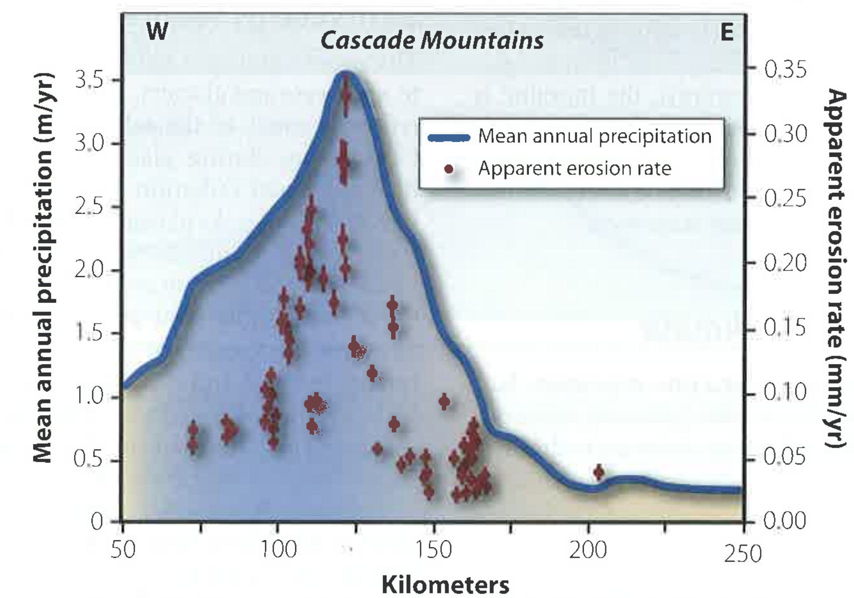
Precipitation changes were much more variable. Some parts of the world became wetter and others became drier as the climate changed. Consider the pluvial lake systems of the Sahara and Mojave deserts. In the Mojave Desert, pluvial lakes expanded during glacial times. As glaciation ended, the lakes dried as temperatures increased and cloudiness and precipitation decreased. In the Sahara Desert, glacial times were dry but lakes there filled with water during the early to mid-Holocene (10,000 to 5000 years ago) in response to shifting storm patterns and the location of moisture-laden winds. It was during the wetter mid-Holocene that channels apparent in space shuttle radar images, and verified in the field by archaeologists, carried water across the now-parched Sahara Desert (Photograph 13.10).

Not only does climate change over time but the frequency of extreme events, and their intensity, also changes. Understanding how the distribution and character of extreme events change is important because some geomorphic processes (such as mass movements, gully initiation, and fluvial sediment transport) are driven not by average conditions but by extreme conditions. In many landscapes, only long-lasting or high-intensity storms can provide enough energy (through rainfall, wind, or runoff) to cross stability thresholds and initiate significant geomorphic activity (see Chapter 1).

Consider, for example, how the transport of large cobbles requires a critical shear stress provided by a deep-enough



The overall topography of the South American high peaks tracks the last glacial maximum (LGM) **snow line** or ELA toward the equator and the modern snow line at higher latitudes. The correspondence between snow line and topography has been interpreted as erosive mountain glaciers controlling topography. The nearly 1 km displacement between the modern and LGM snow line reflects climate change (warming) between glacial and interglacial times.



Rock erosion rates, inferred from **U/He dating** of apatite crystals in granitic rocks, are highest on the west slope of the Cascade Mountains of western Washington where the most precipitation (up to 3.5 m/yr) falls. Erosion rates are lowest in the rain shadow on the east slope of the range where less than half a meter of precipitation falls annually.

FIGURE 13.14 Climate and Erosion. Active glaciers remove rock and thus keep the average elevation of mountain ranges similar to the elevation of the glacial equilibrium line altitude. Erosion is focused on the windward side of mountain ranges where precipitation is greatest and the glaciers are largest and most active. [Adapted from Porter (1977) and from Reiners et al. (2003).]

and fast-enough flood flow (see Figures 6.6 and 6.7), and how the triggering of landslides requires critically high pore pressure, provided by rainfall of sufficient intensity and duration (see Figure 4.1). Other examples include landslides in the forested southern Appalachian Mountains or the well-vegetated highlands of New Zealand. Such failures occur in large numbers and region-wide only when very large hurricanes or typhoons strike. Consequently, there is great interest in refining long-term, geologic records of paleo-hurricanes, floods, and monsoons. These events can be so geomorphically effective that they change the rate of erosion and thus sediment yield over time.

The effects of climate change on landscapes depend strongly on the changes in active surface processes that a changing climate causes. Consider the effect of rising and falling sea level driven by warming and then cooling climate and the concomitant contraction and expansion of ice sheets. As ice sheets grow and sea level falls, the effect on a major river system entering the ocean is likely to be incision as potential energy increases; this incision could leave strath terraces along the channel and/or cause a knickzone to move upstream. When sea level begins to rise, knickzone retreat slows and the river aggrades, perhaps burying the strath terraces in alluvium. The process and the resulting landforms are very different when sea level falls and when it rises, an example of hysteresis. Second, consider the advance of a glacier when climate cools or precipitation increases. Such an advance obliterates landforms created by previous glaciations and then deposits a moraine at the glacial maximum. When the climate warms, the moraine is abandoned and a series of recessional moraines may form. These moraines slowly degrade as surface processes redistribute and remove mass. Warming and cooling climate result in very different surface process responses.

Landscape Controls on Climate

Climate-landscape interaction is not a one-way street. Surface processes, and the landforms they generate, influence climate by changing the planet's energy balance, hydrologic cycle, and atmosphere.

Regional Climate

The combined effect of dominant wind direction and topography on local and regional climate, and thus on geomorphology, is often pronounced. Because air masses ascending mountain ranges are forced to rise and thus cool **adiabatically** (by expansion but without heat transfer), they become saturated with water vapor, triggering **orographic** precipitation [**Figure 13.15**]. As a result, **windward** (upwind) slopes are much wetter than **leeward** (downwind) slopes. Such hydrologic asymmetry increases erosion rates on the windward side of mountain ranges through a combination of increased runoff, higher stream power, and more frequent mass movements. Increased snowfall lowers snow lines, which result in large, active glaciers. For example, long-term (millions of years) erosion rate data from the Cascade Mountains of Washington State are well correlated to the modern distribution of precipitation (Figure 13.14). On the wet west side of the range, long-term erosion rates and thus rock uplift rates are high. Erosion rates and rock uplift rates are lower on the drier, less rapidly eroded east side. Over geologic time, this can result in the development of different geologic structures on either side of a range, as the loss of mass by erosional processes drives an isostatic response that draws rock upward in addition to or in place of the push of tectonics.

On longer timescales, the arrangement of continents has influenced climate and the distribution of climate-sensitive landforms (Figure 13.5). For example, as Antarctica pulled away from South America and Australia, the circum-Antarctic current developed, thermally isolating the continent from the rest of the Southern Ocean, allowing the landmass to cool, and likely helping to trigger the onset of continental-scale glaciation in Antarctica about 30 million years ago. The growth of Antarctic ice sheets lowered sea level as water was sequestered and also increased planetary albedo as the area of highly reflective ice and snow expanded. Thus, the climate-changing effect of establishing the Antarctic ice sheet was felt worldwide.

Earth's Energy Balance

The most significant landscape-climate linkage is related to ice sheets and glaciers. Ice and snow have high albedos, reflecting much of the solar energy delivered to their surfaces. Thus, during glaciations, a larger percentage of incoming solar radiation is returned to space than during interglaciations. As glacial ice begins to melt and saturate with water, its albedo decreases, more solar energy is absorbed (accelerating melting), and less is reflected. Once the ice and snow melt away, revealing rock and soil or allowing the growth of dark green vegetation, albedo is further reduced and even more solar energy is retained on and near Earth's surface. This albedo decrease further warms the planet in a positive **feedback loop.** Climate models suggest that if the Greenland Ice Sheet melts away due to climate warming driven by increased greenhouse gases, it would not form anew even if greenhouse gas concentrations were reduced to preindustrial levels. Similarly, loss of sea-ice cover on the Arctic Ocean, another effect of global warming, will reduce albedo as dark seawater absorbs more sunlight than white snow and ice.

Hydrologic Cycling

Feedbacks link biologic, geomorphic, and hydrologic systems. For example, removing trees from the Amazonian rain forest has dramatically reduced transpiration. The loss of transpiration has reduced atmospheric water content and thus the volume of precipitation in the Amazon Basin. Because Amazonia is a partially closed system, with

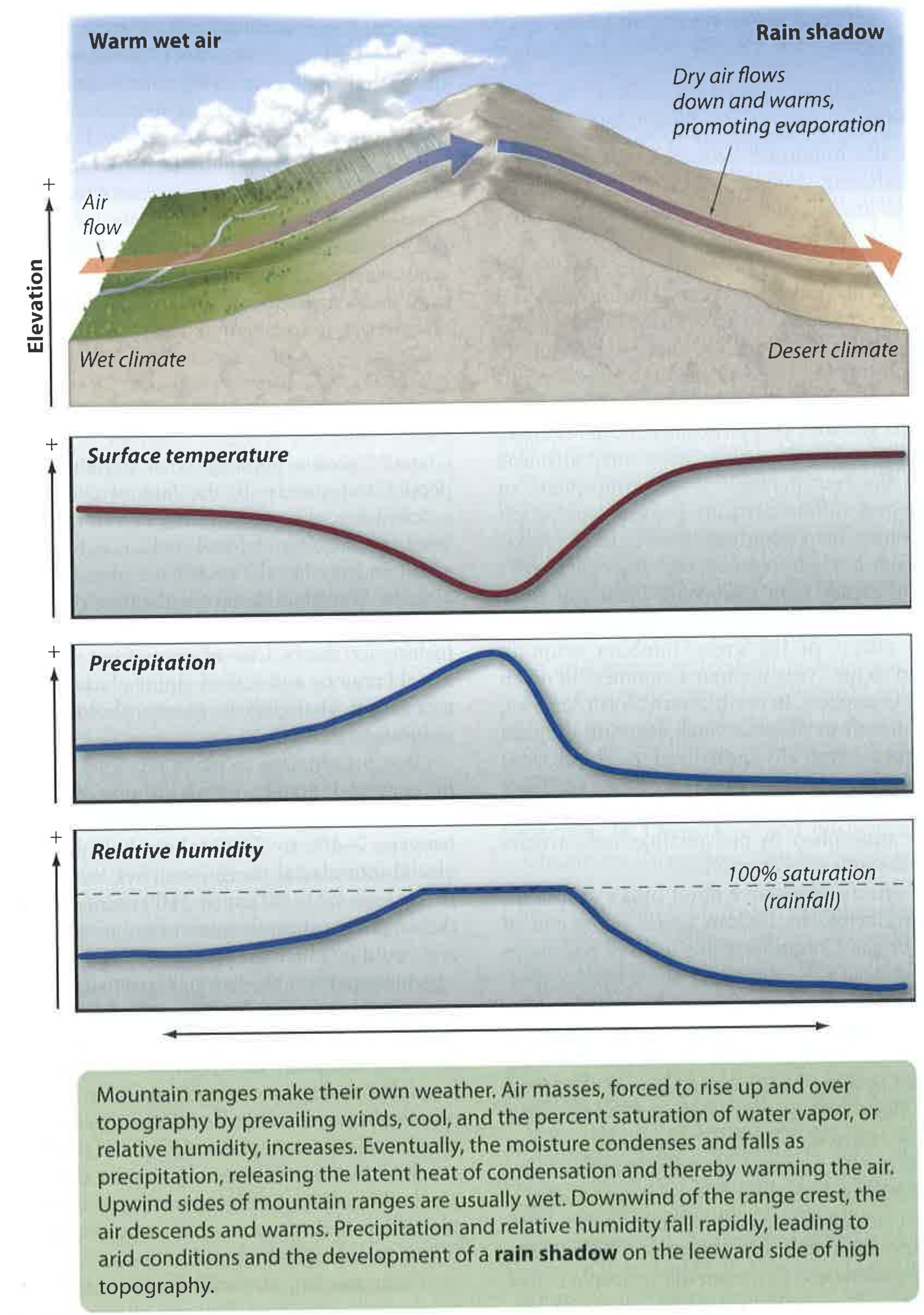


FIGURE 13.15 Orographic Effects on Climate. Mountains strongly affect local climate by forcing masses of air to rise and thus cool, leading to precipitation as water vapor in the rising air mass reaches saturation and condenses.

significant amounts of water recycling within the basin, it is unclear if sufficient water now exists to reestablish the full extent of the original rain forest. In some places, savanna encroachment is establishing a new regime for soil formation, bioturbation, and geomorphic activity.

Tectonically driven surface uplift of large areas, such as the Tibetan Plateau, can accentuate monsoonal circulation due to summer heating of the uplands and the air masses above them. Such rising air unleashes large volumes of precipitation in short periods, filling rivers in extreme runoff events and doing significant geomorphic work.

The Atmosphere

Two geomorphically important processes, volcanic emissions and the weathering of silicate rocks, alter atmospheric physical properties and chemistry. These important forcings affect climate on very different timescales.

Volcanic emissions can dramatically change climate over the short term, i.e., months to years. During large and explosive volcanic eruptions, such as those typical of subduction zone volcanism, particulate matter and sulfate aerosols (Figure 13.4) are injected into the upper atmosphere where they are carried around the world by high-level winds (the **jet stream**). This airborne material changes the energy balance of Earth's surface since small airborne particles reduce the transparency of the atmosphere to solar radiation, and sulfate aerosols provide nucleation sites for water vapor, increasing the rate of cloud formation. Because clouds are light-colored, they increase Earth's albedo, reflecting rather than adsorbing incoming solar radiation. Both effects tend to cool the planet as typified by the climatic effects of the great Tambora eruption (1815), which led to the "year without a summer" in much of the Northern Hemisphere. In northeastern North America, snow fell every month of the year. Such eruption-induced changes in climate are typically short-lived (on the order of months to years) because of the limited atmospheric residence time of both particulate matter and sulfate aerosols. Both are washed from the atmosphere by precipitation and particles settle out under the influence of gravity.

The climatic effects of massive flood basalt eruptions, such as those producing the Deccan Traps at the end of the Cretaceous or the Columbia Plateau flood basalts in the Miocene, last longer than those of short-lived, explosive eruptions of silicic volcanoes (Figure 13.5). Due to the volume of magma erupted, these long-duration, high-volume events can release large amounts of CO_2 gas, changing the chemistry and thus the radiation balance of the atmosphere. Because the processes that remove CO_2 from the atmosphere occur relatively slowly in comparison to precipitation scrubbing of particles, the greenhouse-induced climatic effect of high-volume eruptions is longer lasting.

Weathering of silicate minerals drives climate on much longer timescales than volcanic eruptions by slowly removing CO_2, a greenhouse gas, from the atmosphere (see Chapter 3). Such rock weathering is the source of dissolved loads in rivers, which carry cations and bicarbonate icons to the oceans where organisms make $CaCO_3$ shells from these components of carbonate minerals. Once locked up in mineral form, CO_2 is removed from the atmosphere and the greenhouse effect it causes is reduced.

Over the past 20 years, some scientists have argued that pulses of uplift, such as the growth of the Himalaya, have affected global climate by changing the global rate of rock weathering. They suggest that surface uplift steepens topographic gradients, increasing stream power, and causing incision. River incision leads to steeper slopes, more mass wasting, and more exposure of fresh, unweathered rock. If the global rock-weathering rate increases, this will draw down atmospheric CO_2 as more carbon is sequestered. Such weathering-catalyzed removal of CO_2 from the atmosphere is thought to have reduced the greenhouse effect over the Cenozoic era, cooling climate and thus exacerbating if not catalyzing glaciation in the Pliocene and Pleistocene (Figure 13.5). Removal of CO_2 by weathering reactions is a slow process and will not compensate, on human timescales, for the anthropogenic additions of CO_2 to the atmosphere.

Applications

Climatic geomorphology is of particular importance to people and society. In the face of continued, and likely accelerating, climate warming driven by emission of CO_2 from the burning of fossil fuels, it is important to understand and predict the response of landscapes to changing climate. Warming climate will cause geomorphic changes related to rising sea level in coastal regions as the result of melting ice sheets. Loss of permafrost will destabilize soils in cold regions and loss of alpine glaciers will reduce summer runoff, changing the geomorphology (and hydrology) of mountain channels.

It is informative to place previous climate changes in the context of predictions for the next century. Most climate models predict an increase in global average temperatures between 2–4°C by 2100 (about half the rise that typifies glacial-interglacial transitions). Sea levels are predicted to rise at least 20 to 60 cm by 2100 unless feedbacks increase the rate of ice sheet disintegration, in which case, sea-level rise could be much greater.

Analogs from the geologic past suggest that warming estimates from models are reasonable. Such analogs suggest that over the next hundreds to thousands of years, sea level will rise many meters as the Greenland and Antarctic ice sheets melt. For example, emission of CO_2 during the middle Miocene eruption of the Columbia River flood basalts raised atmospheric CO_2 levels only about half as much as human emissions have over the last century but resulted in a 3°C average warming of Earth. Because little water was locked up in polar ice caps, Miocene sea level was tens of meters higher than today.

Understanding the behavior of ice sheets and the speed at which they react to changing climate is critical, because the transfer of water from melting glaciers and ice sheets to the ocean will be one of the most devastating results of a warming climate. Accelerating sea-level rise will inundate low-lying areas, including many of the world's most important cities. Melting of the Greenland Ice Sheet in its entirety would raise global sea level by 6 to 7 m, enough to submerge many coastal cities, such as New York. Not only will cities be lost but deltas, estuaries, and their marshes, the nurseries for many important marine species and a buffer against storm surges, will disappear as rates of sea-level rise exceed rates of sedimentation and these coastal landforms vanish beneath the waves of a rising sea.

As both alpine glaciers and polar ice sheets melt, sea level will rise as it has before, most recently at the termination of the last glaciation, starting about 22,000 years ago. However, over the past several million years, sea level was rarely higher than it is today. Only during a few particularly warm or long interglaciations, such as marine oxygen isotope stages (MIS) 5e, 9, and 11, did the Greenland and/or the Antarctic ice sheets melt back more than they have during the Holocene. There is evidence, some of it ambiguous and still controversial, that during at least one interglaciation (MIS 11, about 420,000 years ago) sea level was as much as 12 m higher than it is today. Did this extreme sea-level rise reflect the disappearance of the Greenland Ice Sheet? Perhaps. There is evidence in and below the ice, including pollen indicative of boreal forests, for a much-reduced Greenland Ice Sheet during MIS 5e, 9, and/or 11.

Many feedbacks associated with global warming have the potential to amplify initial climate changes. For example, the melting of both sea ice and polar ice sheets reduces planetary albedo, changing Earth's energy balance. Other positive feedbacks include the release of methane (a powerful greenhouse gas) from melting permafrost as arctic soils warm. Large amounts of methane are also stored in sediment on the continental shelf in the form of **clathrates**, complex mixtures of ice and organic molecules unstable at temperatures above 4°C. Warming seawater will catalyze thermal decomposition of these clathrates, causing release of this methane and weakening of the sediment in which it is contained. Not only will such releases accentuate the greenhouse affect and climate warming but they could also trigger giant submarine landslides with the potential to generate devastating tsunamis.

On a warming planet, the distribution of atmospheric water vapor will change. The hydrologic cycle will become more active because at higher temperatures the atmosphere can hold more water (the saturation vapor pressure of water increases with temperature). Speeding the hydrologic cycle is likely to increase both the rate and frequency of precipitation, and thus runoff, in many areas. More water in a warmer atmosphere will likely lead to stronger and longer duration storms—conditions conducive to mass movements and debris flows. If discharge increases, streams are likely to incise and widen, changing their courses.

Models suggest that the extra water will not be evenly distributed. Some parts of the planet may get wetter, such as northeastern North America; other parts, such as Australia and the middle of North America, are predicted to become drier. In areas that become drier, aeolian activity is likely to increase. Once-stable dunes may reactivate as vegetation dies back. In a hotter and at least seasonally drier climate, the increased size and severity of wildfires will be an important driver of hillslope erosion and debris flow generation. Predicting the nature and extent of landscape response to climate change presents a challenge that draws on geomorphological insights and expertise.

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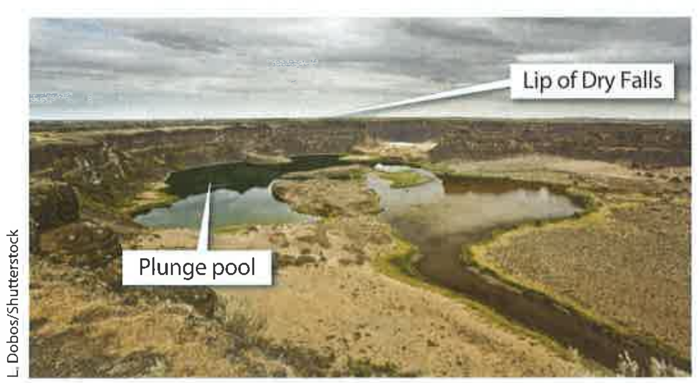
DIGGING DEEPER Do Climate-Driven Giant Floods Do Significant Geomorphic Work?

Many people think of climate change as the ups and downs of average global temperature over time and thus the slow comings and goings of glaciers or the gradual wetting and drying of land outside the glacial margin. But there is a long-standing argument that the geomorphic effects of climate change can be much more dynamic, even catastrophic. The first hints of sudden and massive climate-induced geomorphic change came about a century ago, the result of detailed fieldwork by one man in the semi-arid plains of eastern Washington.

Using horses and primitive automobiles, a young geologist, J Harlen Bretz, traversed the rugged, loess-covered basalt terrain of eastern Washington. He was fascinated by the deep canyons and large deposits of gravel and river sediment at locations inconsistent with the size and course of the Columbia River that flows through the area today. Massive dry waterfalls and plunge pools were fed by empty channels where the loess was stripped away and the basalt deeply potholed by long-vanished rivers **[Figure DD13.1]**. Current ripples made of gravel were up to 15 m high and filled valley bottoms. Immense boulders were deposited in broad fan-form features where the flow diverged. All of these features were evidence implying massive flows of water **[Figure DD13.2]**.

Bretz thought these features were best explained by a giant flood and coined the term **Channeled Scabland** to describe the landscape they left behind (Bretz, 1923). The flood he envisioned was so large, he termed it a "debacle." But, without being able to identify a source of the floodwaters and because his hypothesis invoked a catastrophic event for which there was no modern analog, Bretz's ideas were ridiculed for decades by most geologists who held to an overly restrictive interpretation of **uniformitarianism**—that processes operating at typical modern rates must be able to explain all past landscape features.

Then, in 1942, Pardee published a paper entitled "Unusual currents in glacial Lake Missoula," describing glacial Lake Missoula as the source of Bretz's flood (Pardee, 1942). There, ~2500 km^3 of water had been impounded by



L. Dobos/Shutterstock

FIGURE DD13.1 Dry Falls is the result of the great Missoula floods pouring across the Channeled Scablands of eastern Washington. Floodwaters moved left to right pouring over the basalt cliffs and eroding the deep plunge pools and sinuous channels, here filled with standing water.

(a)

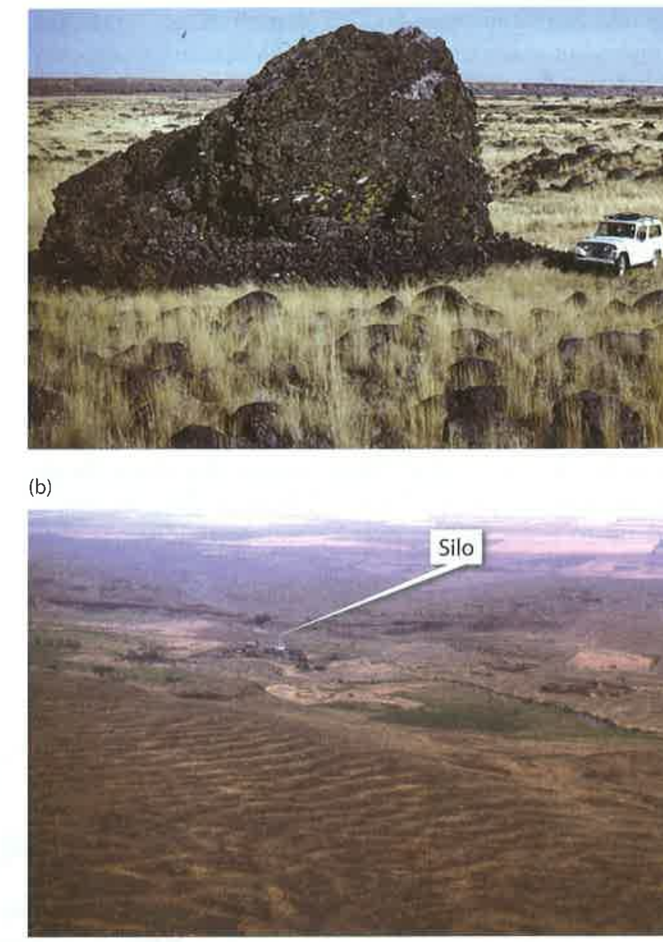


FIGURE DD13.2 Oversize geomorphic evidence shows the magnitude of the floods that swept over the Channeled Scablands. (a) Boulders many times the size of a field vehicle were moved kilometers by floodwaters onto the Ephrata fan in eastern Washington. (b) Giant current ripples cover the top of a massive gravel bar in eastern Washington. Roads and a grain silo provide scale. [From Baker (2009).]

DIGGING DEEPER Do Climate-Driven Giant Floods Do Significant Geomorphic Work? *(continued)*

a tongue of glacial ice that blocked the Clark Fork River. Distinctive megaripples on the ancient lakebed showed that the ice dam failed, rapidly draining the lake and providing a source for Bretz's flood. Analysis of maps and newly available air photographs sealed the case and allowed the routes of the floods to be determined in detail [**Figure DD13.3**]. Careful examination of backwater sediment, deposited in tributary valleys of the Columbia River, showed evidence that the scabland was flooded not once as Bretz had thought, but many times. In the late Pleistocene alone, over 40 massive floods poured down the Columbia as the ice dam repeatedly breached and reformed (Waitt, 1980).

The idea of massive floods as significant agents of geomorphic change has gained traction over the past several decades but there are still controversial flood hypotheses. For example, Shaw and Gilbert's analysis of the morphology and orientation of several North American drumlin fields (Shaw and Gilbert, 1990) suggested to them that some drumlins were shaped by massive subglacial floods. They posited that such floods likely resulted from the release of meltwater stored subglacially and that subglacial floods were a glacial-hydrologic response driven by a warming climate. Calculations suggest that these floods would have raised sea level >20 cm in days to weeks and up to a few meters over several years (Shaw, 1989). However, many researchers (e.g., Benn and Evans, 2010) question the subglacial flood hypothesis for the origin of drumlins, and the hypothesis remains untested (and may be untestable).

It is indisputable that large episodic floods of glacial meltwater have affected the world's oceans, perturbing circulation, and thus further changing climate (Clark et al., 2001). For example, abrupt changes in ocean water oxygen isotope composition after the last glacial maximum are consistent with large, episodic inputs of isotopically light (^{18}O-depleted) glacial meltwater [**Figure DD13.4**]. Floods of glacial meltwater into the North Atlantic likely caused the region-wide Younger Dryas cooling event from 12,800 to 11,500 years ago (Broecker, 2006). The climatic effects of major glacial-lake floods continued well into the Holocene. About 8400 years ago, as the last section of the Laurentide Ice sheet melted away, it released over 100,000 km^3 of water from proglacial Lake Agassiz into Hudson Strait (Clarke et al., 2004). For nearly 200 years after this discharge, the ice in cores collected from the Greenland Ice Sheet shows a 5-degree drop in average temperature over central Greenland and increased Northern Hemisphere dustiness from cooling and drying of the landscape.

Nearly 100 years after Bretz first suggested that megafloods have major geomorphic impacts, the data now show that megafloods were widespread around glacial margins, including such places as Siberia and Alaska. Bretz's thinking has also greatly influenced the interpretation of surface features on other planets. For example, using Earth as an analog, there is now consensus that the streamlined islands and channels on Mars formed billions of years ago in giant Martian outburst floods (Baker, 1978).

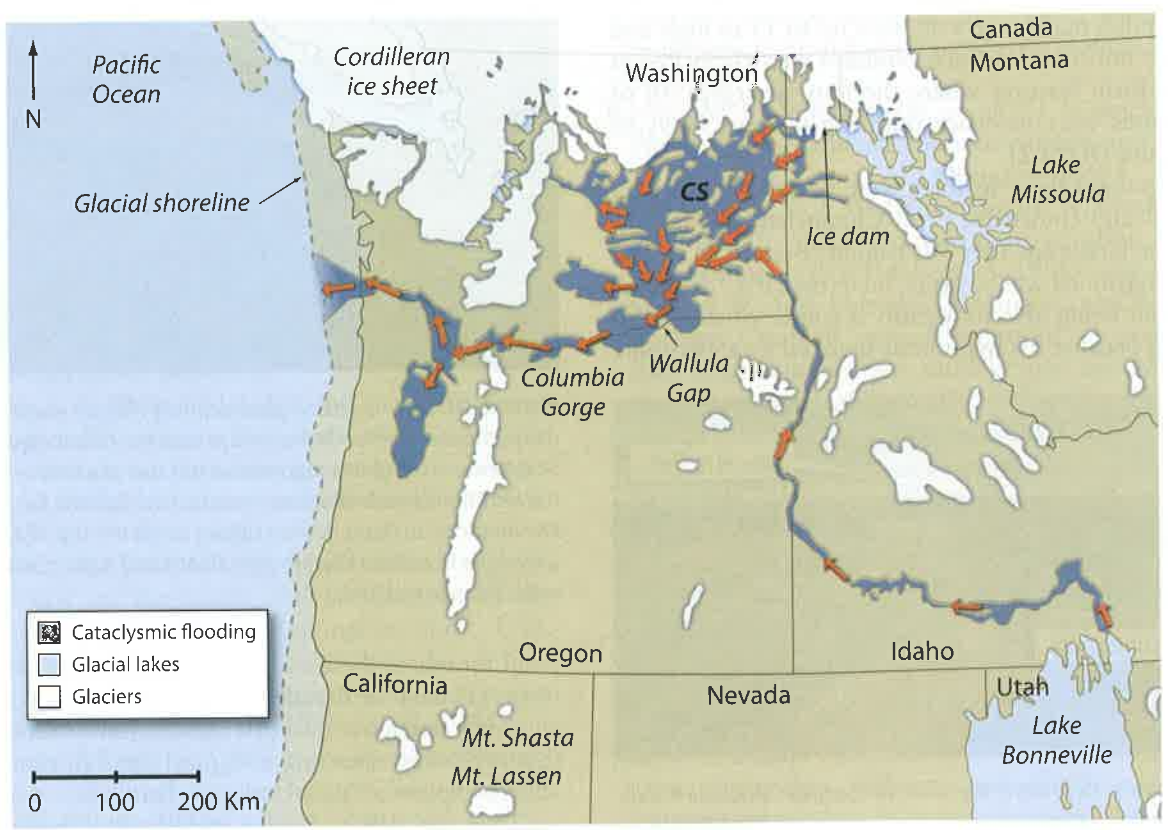


FIGURE DD13.3 The Channeled Scablands (CS) occupy much of eastern Washington State. This map shows ice-dammed Lake Missoula and pluvial Lake Bonneville in light blue, ice sheets and ice caps in white, and the scabland channels in dark blue with flow directions shown by red arrows. [From Baker (2009).]

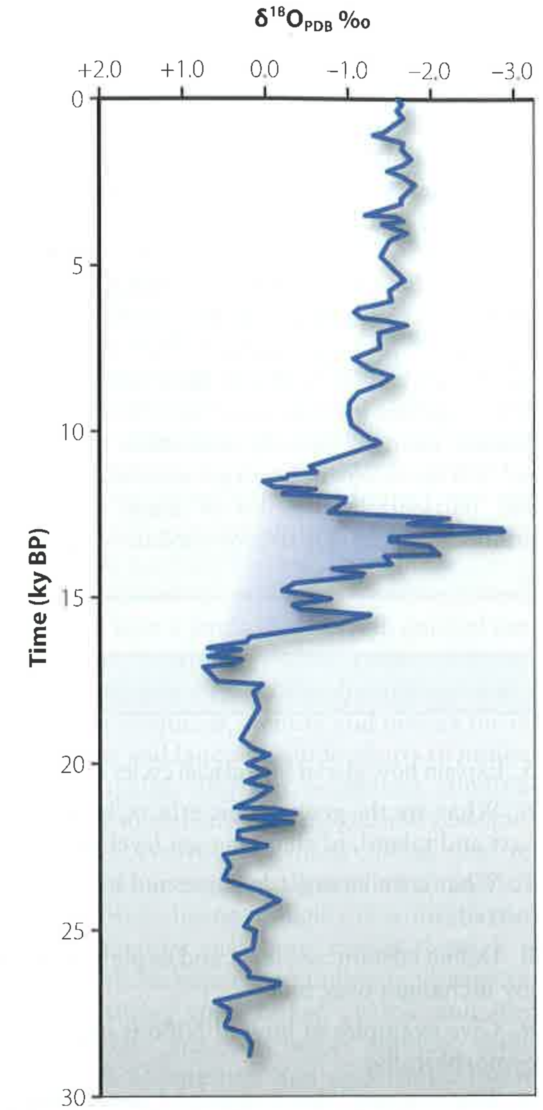


FIGURE DD13.4 Record of the stable oxygen isotopic composition of seawater in the Gulf of Mexico created by analyzing foraminifera, the fossils of small marine organisms isolated from marine sediment cores. The large isotopic excursions (to more depleted values) between 17,000 and 12,000 years ago (shaded area of the graph) are thought to represent pulses of glacial meltwater pouring down the Mississippi River. [From Shaw (1989).]

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WORKED PROBLEM

The Intergovernmental Panel on Climate Change as well as climate modelers have made predictions about how Earth's climate will change over the next 100 years. These predictions are based on assumptions about the amount of carbon dioxide and other greenhouse gases that humans will add to the atmosphere and take into account a variety of landscape, atmospheric, and oceanic feedback mechanisms.

Question: Consider the geomorphic impact of the following climate change predictions in terms of hillslope and river processes including the effects of vegetation.

1. As an environmental planner in the humid, temperate northeastern United States, describe the geomorphic impact of a warming climate in which winter snowpacks and river ice are reduced, the frequency and extent of summer drought is increased, mean annual

precipitation goes up, hurricane frequency and intensity increase, and the number of exceptionally hot days doubles.

2. As a dry-land rancher in central Australia where the climate is semi-arid, you face predictions of less annual rainfall, longer droughts, and higher summer temperatures. What changes are likely to happen to the landscape you farm?

Answer: In the humid northeast, climate change will cause a variety of impacts to both hillslopes and rivers. Thinner snowpacks and warmer winters will shift the timing of spring floods earlier and shorten their duration, and will increase the frequency of rain-on-snow flooding. Summer drought and more heat waves will reduce stream and river baseflow during summer months, impacting aquatic organisms and the ability of farmers to take water for irrigation. Heavier rainfalls and stronger hurricanes may cause more frequent and extensive landslides as precipitation intensity-duration thresholds for slope stability are crossed more often. Higher river flows during storms may move bedload and erode banks that would otherwise be stable, causing channel change. Changes in vegetation amount and character as the result of droughts could also influence hillslope and channel dynamics through the effect of roots on soil strength.

In the dry lands of central Australia, rising temperatures and decreased rainfall will reduce already scant vegetation, limiting the number of animals that can be grazed and potentially destabilizing hillslopes and valley bottoms when heavy rains do come. Baseflow and hyperheic flow in streams will diminish, reducing the number and density of riparian plants including trees that stabilize the banks. In the absence of root reinforcement from vegetation and the bioturbation caused by plants, these rains are likely to cause more surface erosion, creating rills and gullies as precipitation intensity exceeds infiltration rates. Eroded material will move downslope and accumulate on valley bottoms, particularly if flows in main channels are discontinuous and unable to move sediment far from its source.

KNOWLEDGE ASSESSMENT Chapter 13

1. Explain how climate differs from weather.
2. Provide two examples of how geomorphologists create records of paleostorms.
3. Define climatic geomorphology.
4. Give two examples of how ocean circulation affects terrestrial climate and explain how each works.
5. Predict and explain the effect that surface uplift of a mountain range will have on the distribution of precipitation across and downwind of the mountains.
6. Give an example of how lake sediments are used to document climate change over time.
7. Define "ice-rafted debris" and explain its origin.
8. Explain how oxygen isotopes can be used to infer paleoclimatic changes using two different geologic archives.
9. What is the relationship in ice cores between the concentration of CO_2 and paleotemperature?
10. Define loess and explain where you would be likely to find it.
11. Predict the response of climate if rock weathering rates increase.
12. Explain how volcanic eruptions can influence climate.
13. What is the physical explanation for how erosion of preglacial regolith may have changed the behavior of ice sheets and the cyclicity of glaciations?
14. What are the three primary Milankovitch cycles and together how are they thought to control climate change over time?
15. Explain how glacial-interglacial cycles affect sea level.
16. What are the geomorphic effects, both near the coast and inland, of changing sea level over time?
17. What is an emergence curve and how is it constructed?
18. Define eustatic sea level and explain how and why it changes over time.
19. Give examples of how El Niño is important geomorphically.
20. What caused lakes to form in arid southwestern North America during the last glaciation?
21. How did the elevation of Earth's surface change under and just outside of Pleistocene ice sheets when they were fully advanced?
22. Explain how geomorphic evidence can be used to deduce the magnitude of glacial isostatic response.
23. Explain how the location of glacial moraines is used to estimate the change in ELA over time.
24. Give an example of the linkage between biologic, geomorphic, and hydrologic systems.
25. Compare the magnitude of climate changes predicted for the next 100 years due to human-induced global warming with those that occurred between glacial and interglacial times.
26. Describe a positive feedback that takes place when the Earth warms.
27. Who was J Harlen Bretz and how did he revolutionize geologic thinking?