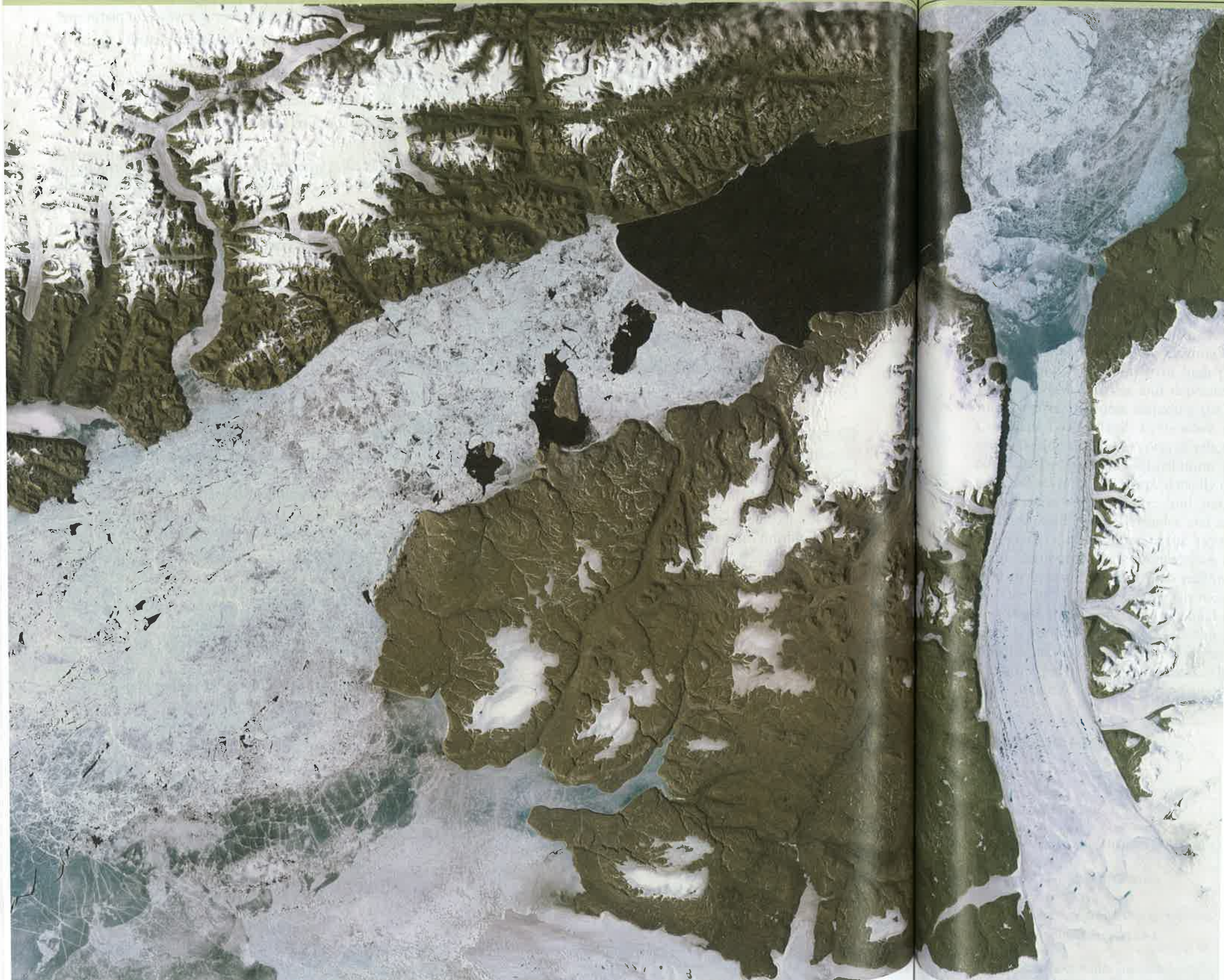
PART III



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Ice, Wind, and Fire

(Chapters 9 to 11)

Glaciers, wind, and volcanoes shape some of Earth's most spectacular landscapes. In Chapter 9, we consider landscapes where geomorphic processes are dominated by the presence of ice. Wind, the landforms it creates, and the sediments it transports are the subject of Chapter 10. In Chapter 11, we consider the geomorphic impact of volcanism on Earth's surface, both the constructional forms that result from eruptions and those that remain after erosion of volcanic landforms.

In this view, a narrow strait passes between Ellesmere Island (at top) and the northwestern Greenland coast. The Petermann Glacier to the right flows north in a narrow fiord from Greenland's central icecap and had been the largest floating glacier in the northern hemisphere until August 2010, when a large portion of it broke off. Landsat ETM + image of the Arctic is at midsummer, near the melt maximum for July 1999. At this season, the snowfields have generally melted away from shore areas, but icecaps and glaciers persist near shore.

Glacial and Periglacial Geomorphology

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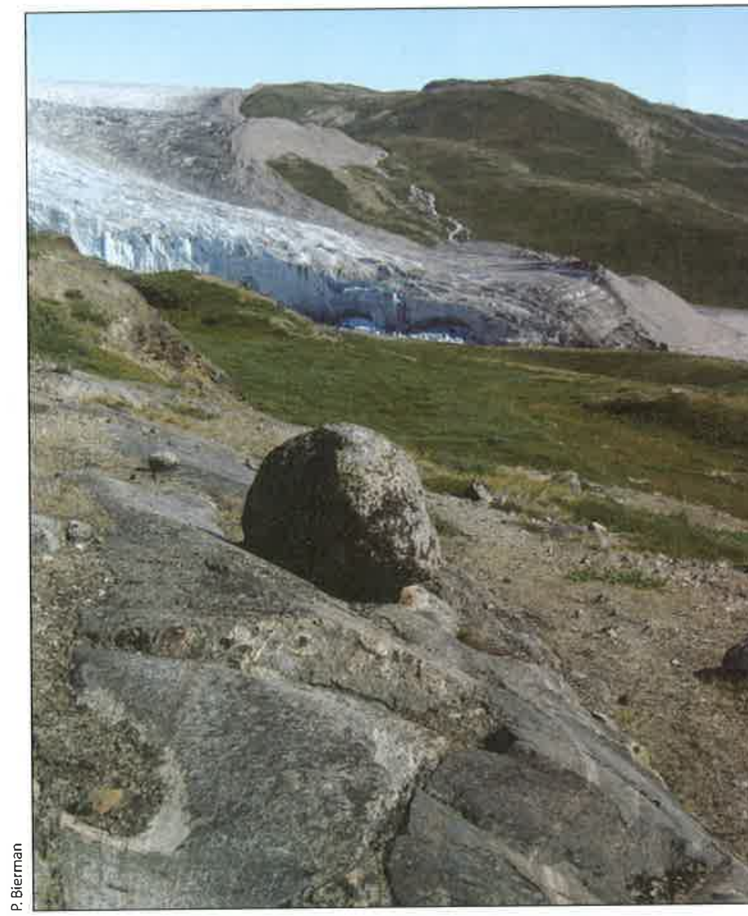

P. Bierman

Margin of the Greenland Ice Sheet about 40 km east of Kangerlussuaq in western Greenland. In the foreground, a granitic erratic boulder sits on glacially sculpted gneissic bedrock. In the distance, Russell Glacier issues from the southwestern margin of the ice sheet. Rocky, light-colored lateral moraines, probably deposited several hundred years ago during the Little Ice Age cold period, fringe the ice on the right along with small meltwater streams.

Introduction

Glaciation is a geomorphically powerful process. The comings and goings of glaciers at high latitudes and high altitudes not only affect the geomorphology of glaciated regions, but the growth and decay of ice sheets and glaciers also affect the rate and distribution of surface processes around the world by changing sea level, affecting global average weathering rates, and shaping topography.

Glaciation is geomorphically powerful enough to influence the hypsometry (elevation/area relationship) of mountain ranges by preferentially eroding high-elevation terrain and limiting the elevation of high peaks. Glaciers excavated the Great Lakes and deposited the sediment that forms marine features including the Grand Banks, Cape Cod, and Long Island. Even when glaciers have melted away, their legacy, in the form of relict or fossil landscapes, controls the pace and distribution of surface processes such as rockfall from steep, glacially carved valley walls.

Glaciers, rivers of ice, and **permafrost**, the frozen ground found at high latitudes and altitudes in areas that lack glaciers, create some of the most spectacular and dynamic geomorphic environments on our planet. Much of their dynamism results from several physical properties of ice. One, on Earth, most ice exists near its melting point, so that **phase transitions** between liquid and solid water [**Figure 9.1**] occur commonly and account for much of the geomorphic activity of glaciers and permafrost. Two, ice melts when the pressure on it increases. Three, because ice is a weak material near its freezing point, it readily deforms

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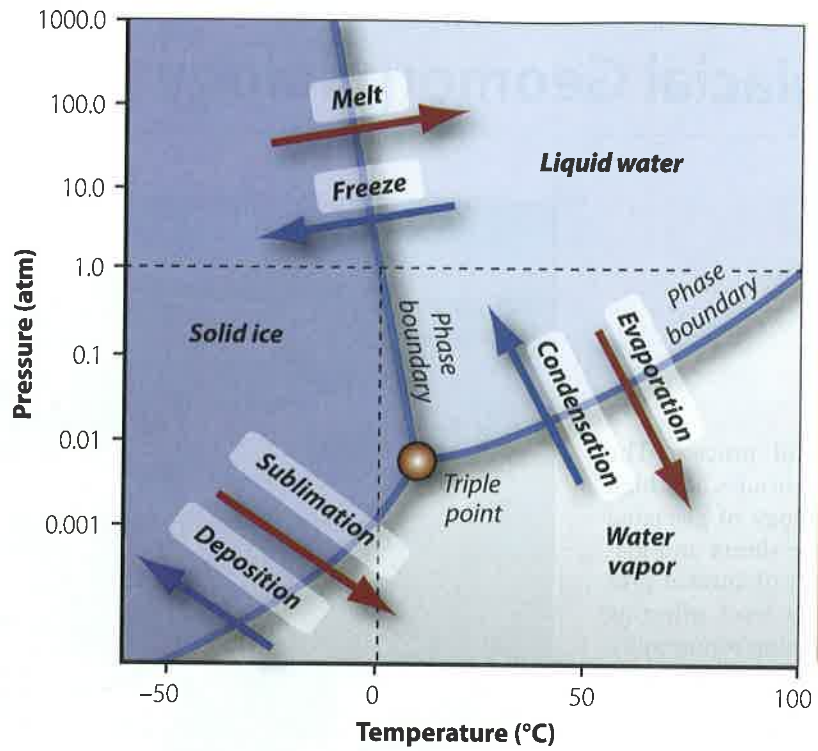


Water is unusual because, in the range of temperatures and pressures common on Earth's surface, it can exist in the vapor, solid, and liquid phases. The negative slope of the **phase boundary** between liquid water and ice means that as pressure increases, solid ice, kept at a steady temperature, will melt—a phenomenon referred to as **pressure melting.** The triple point is the pressure and temperature condition at which all three phases of water coexist.

FIGURE 9.1 Phase Diagram for Water. The diagram shows that all three phases of water (vapor, solid, liquid) exist at Earth surface temperatures and pressures.

under gravitational loads. This deformation allows glaciers and frozen soil to flow and transport large amounts of mass, both ice and sediment. This transport of mass shapes landscapes as does glacial **abrasion,** the grinding away of bedrock below the ice, and **quarrying,** the removal of blocks of rock aided by changes in water and overburden pressure.

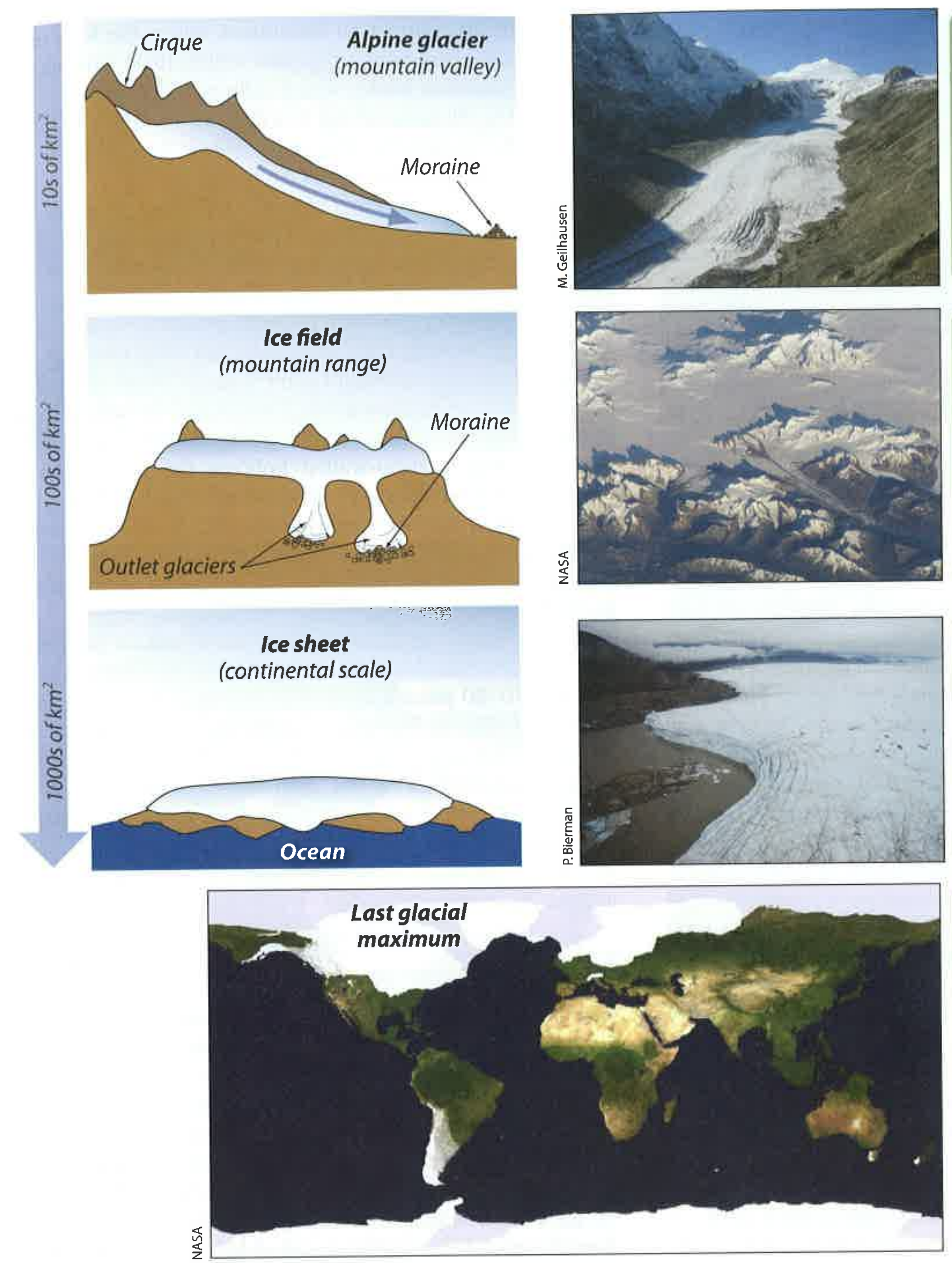
The erosive power of **alpine glaciers** creates steep cliffs and deep basins characteristic of once-glaciated mountain landscapes. **Continental ice sheets** have scoured whole regions, leaving behind landscapes of smoothed bedrock covered in some places with **till** (unsorted glacial sediment) and covered in other places with sand and gravel sorted and transported by water derived from melting glacial ice. The dramatic and deep **fjords** (deep linear troughs) of coastal Norway, Alaska, Greenland, southwestern New Zealand, and elsewhere owe much of their grandeur to glacial erosion focused in valleys.

Because some landforms and sediment assemblages can be unambiguously attributed to glacial processes, geomorphic mapping indicates the extent of now-vanished ice sheets and glaciers. For example, the distribution of glacial deposits and glacial landforms demarcates the margins of ice sheets that once covered much of the high latitudes and mountain ranges of the world [Figure 9.2]. Careful examination of the rock record indicates multiple episodes of glaciation stretching back at least into the Precambrian era, 2.9 billion years ago. Tillite, or glacial sediment turned to rock (lithified), is preserved in the rock record and in some cases covers striated and polished rock surfaces [Photograph 9.1].

Times when continental glaciation was widespread are thought to reflect both favorable atmospheric conditions (low levels of CO_2 and, consequently, less effective greenhouse processes), lowered inputs of solar radiation, and continental arrangements with large land areas near the poles (providing land at cold, high latitudes on which ice sheets can grow from snow that survives the summer melt season). The pacing of glaciation over hundreds of millions of years is controlled by plate tectonics, while solar radiation input, as dictated by cyclic changes in the shape of Earth's orbit, controls the advance and retreat of glaciers over shorter, 100,000-year timescales.

In the last 2.7 million years (the late **Pliocene** and **Pleistocene** epochs), much of Earth's surface, including northern Europe and North America, has been directly and indirectly affected by multiple episodes of continental and alpine glaciation. Significant Quaternary climate changes, which brought Earth in and out of dozens of glaciations and interglaciations in the past several million years, are recorded in deep-sea sediment cores (see Chapter 13), but the evidence for multiple glaciations is much more difficult to find and interpret unambiguously on land. Only the largest and latest glaciations have left consistent and easily recognizable geomorphic signatures on the landscape.

Geomorphic evidence for the most recent glaciation, which reached its maximum extent about 21,000 years ago, dominates the surficial geology of glaciated regions. By mapping moraines and other glacial sediments, we know the maximum extent of the Laurentide Ice Sheet that covered large parts of North America. In some places



Alpine glaciers are topographically constrained by the cirques in which they originate and the valley walls that confine them. Lower reaches of alpine glaciers are often bordered by moraines.

Ice fields occupy highlands and in many places bury existing topography. They are drained by outlet glaciers that transport ice to lower elevations where it melts and deposits moraines.

Ice sheets largely overwhelm topography, ice buries peaks, and the surface slope of the ice sheet controls ice flow direction. Ice sheets deposit moraines and other glacial sediments.

During the last 2.7 million years, ice sheets and glaciers have repeatedly covered high latitudes and high altitudes. Ice was most extensive during glaciations and least extensive during interglacial periods. At the last glacial maximum, ice covered 25% of Earth's landmass.

FIGURE 9.2 Bodies of Ice. Ice sheets, ice fields, and alpine glaciers differ in size, location, and their interaction with topography.

outside the most recent ice margin, one finds discontinuous and eroded remnants of material deposited and rock eroded during previous glaciations. This terrestrial geomorphologic record of older glaciations and, by inference a changed climate, is fragmentary. The extent of ice sheets before the last glaciation is poorly known in large part because glaciers usually destroy the evidence of prior glaciations by overrunning, eroding, and incorporating older glacial sediment as they advance.

Glacial and **periglacial** (low temperature but not glaciated) landscapes are distinct from landscapes shaped by fluvial processes. Periglacial environments are defined by both the geomorphic importance of the ice/water phase transition and the presence of seasonally snow-free ground. Much of the geomorphic activity in periglacial environments is induced by the freezing of water and thus climate, and in particular, temperature, is an important driver of periglacial surface processes. Expanses of **patterned ground** that exhibit striking geometric regularity typify periglacial zones and are the result of thermal contraction, the growth of ice lenses, and stirring of soils in periglacial terrain. Periglacial processes are active in many cold regions that were not covered

PHOTOGRAPH 9.1 Ancient Glacial Grooves. Glacial grooves of Permian age eroded into south Australian bedrock are exposed as overlying till (now solidified into a rock known as tillite) is removed by erosion (person shown for scale).

by glacial ice as well as in areas fringing many glacial margins. **Permafrost,** permanently frozen ground, is important in many periglacial environments. Glacial and periglacial environments are similar in a climatic and geographic sense, although flowing glaciers can extend to elevations below periglacial environments. Glacial and periglacial processes involve different percentages of ice and sediment but they share the importance of linked energy and mass balances.

Ice and glaciers are not limited to our planet. Ice covers oceans on Europa (a moon of Jupiter), as well as other moons of Jupiter, Saturn, and Neptune. Water ice is present in regolith covering shadowed craters on the Moon. Remote sensing data indicate that Mars has polar ice caps, which shrink and expand with the seasons. New data, gathered by ground-penetrating radar, have been interpreted to suggest that icy Martian glaciers, extending from mountains at lower latitudes, are preserved under blankets of rocky debris. It is likely that the debris blanket protects underlying ice from melting or subliming, not unlike many debris-covered glaciers on Earth.

Glaciers

Glaciers are persistent, flowing bodies of ice on the land surface that originate as accumulations of snow. Glaciers come in a wide variety of sizes and shapes (Figure 9.2). Largest are **ice sheets** [**Photograph 9.2a**], also known as **continental glaciers,** which in the past have covered extensive areas of the high latitudes including much of North America, northern Europe, and Antarctica. Ice sheets are large enough to influence climate by diverting storms, altering wind directions, and orographically enhancing precipitation. Once ice overtops topography, the slope of the ice surface itself drives glacial flow, allowing ice sheets to move over landscapes in their path. Such overtopping of topography was prevalent in the Northern Hemisphere where the Laurentide Ice Sheet formed over subdued shield topography. In contrast, in Antarctica, where the ice sheet formed over mountains of greater relief, there are many exposed mountain peaks, which influence the direction of ice flow. The Pleistocene-age Cordilleran Ice Sheet, which covered the mountains of western North America, had a similar topographic control of ice flow.

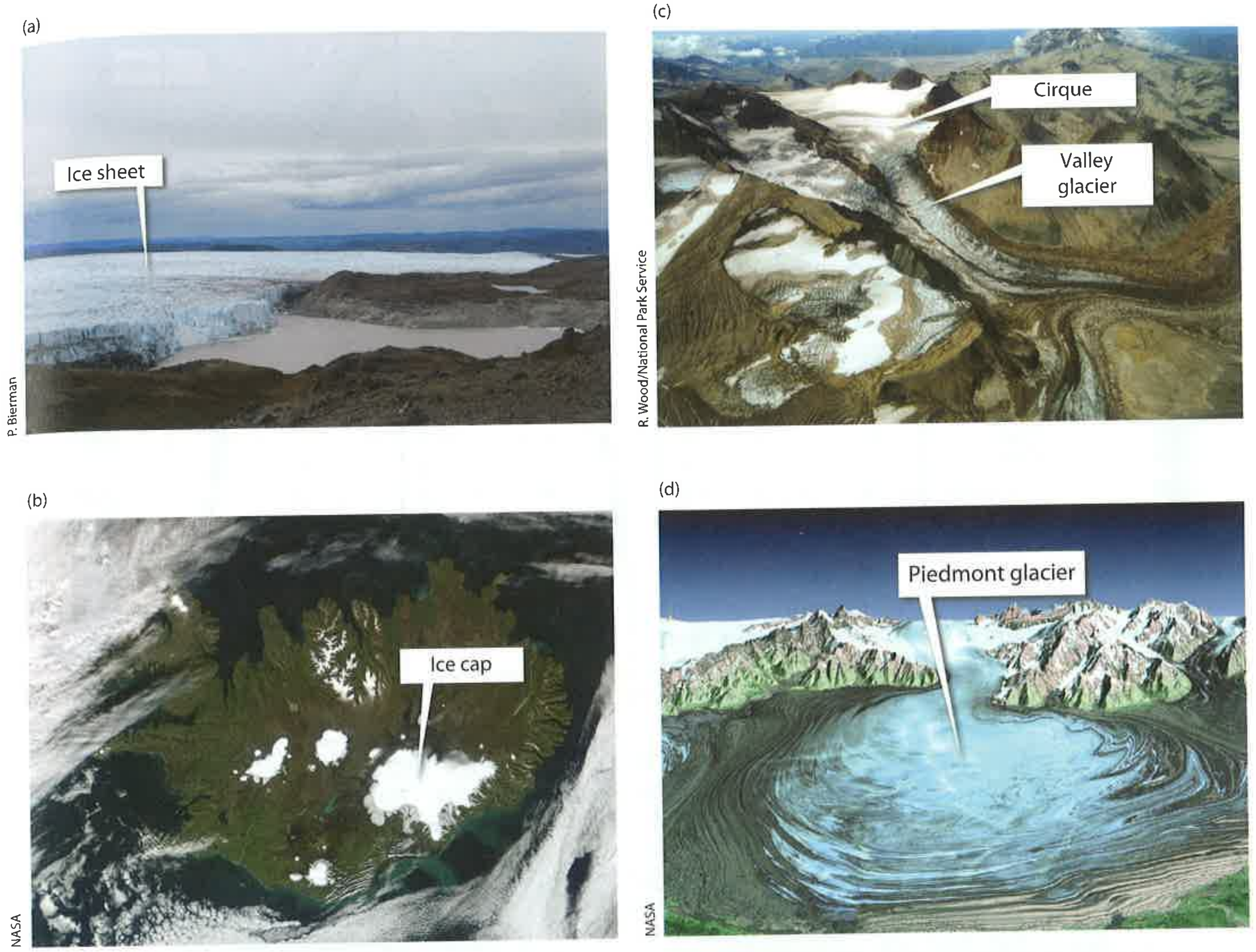
Ice caps ($<$50,000 km^2) are found on the high portions of mountain ranges or on elevated plateaus [**Photograph 9.2b**]. They are smaller than ice sheets and flow largely independent of topography, although the tongues of ice that drain ice caps are often confined to narrow alpine valleys. **Ice fields** are on the same scale as ice caps, but subglacial topography still exerts a significant control on the shape of the ice field and thus on the flow of ice. Ice masses that are confined to valleys are referred to as **valley glaciers** and typically originate in **cirques,** steep-walled hollows the glaciers have eroded into mountain sides [**Photograph 9.2c**]. **Alpine glaciers** may flow out of the mountains and onto lower gradient piedmonts where they are no longer constrained by valley walls [**Photograph 9.2d**]. Cirque glaciers are restricted to basin-shaped cirques high in the mountains.

Today, there are many alpine glaciers and ice caps but only the Greenland and Antarctic ice sheets remain; these are only 10–20 percent smaller in area than they were during glacial maxima but they contain much less ice. Since the last glacial maximum, parts of the Greenland Ice Sheet have lowered hundreds of meters, contributing enough meltwater to raise global sea level about 3 m. Thinning of the Antarctic Ice Sheet since the last glacial maximum has released enough water by itself to raise the world sea level by at least 10 m.

Glacier Mass Balance

Glaciers are best understood by considering their mass balance, which results from the **accumulation** of ice (mostly by snowfall) and its loss by **ablation** (mostly by melting or by calving of large chunks into standing water). You can think about glacier mass balance in the same way that the relationship between income and expenditures influences your bank balance and determines your financial stability.

Ice is added to glaciers by precipitation, both falling directly onto glacier surfaces and added by avalanches and wind carrying snow from above or from adjacent highlands. In high and mid-latitudes, away from monsoonal areas, most glacial ice originates as snow falling during the winter months. In monsoon-affected areas and on mountain peaks in tropics, glaciers accumulate snow during the wet season, which does not need to be winter. As the snow ages over the winter and loses its intricate crystal forms, it changes to larger, more rounded, and compact grains known to skiers as corn snow. Over the melt season, that corn snow further compacts, becoming **firn.** With time, and under the pressure of the overlying snowpack, the firn consolidates to glacial ice, gradually becoming denser and less porous [**Figure 9.3**]. In areas near freezing with high annual precipitation rates, this can happen in just a few



PHOTOGRAPH 9.2 Glacier Sizes and Shapes. Glaciers come in a wide variety of shapes and sizes. (a) The western margin of the Greenland Ice Sheet showing the ice margin and an ice-marginal lake it impounded. (b) Satellite view of Iceland showing several bright white ice caps. (c) Cirque and valley glaciers, Mount Katmai, Alaska. (d) The Malaspina Glacier, a piedmont glacier in southeastern Alaska, spreads into a wide lobe where it is no longer confined by valley walls. View created from a Landsat satellite image and SRTM digital elevation model.

years. In very cold areas with low precipitation rates, the transition to glacial ice can take several tens to a few hundred years. Crystal dimensions increase in glacial ice; older ice crystals can be more than 10 cm long.

Ice is lost from glaciers (ablates) in several ways. Mass leaves the glacier system as ice and snow melt and water drains away in **meltwater streams** (also known as outwash streams) [**Photograph 9.3**]. Some ice is also lost by sublimation, the direct transfer of water from the solid to vapor phase. Sublimation is driven by water vapor pressure gradients when wind moves dry (undersaturated) air over ice surfaces; thus, sublimation is of particular importance for cold glaciers in semiarid and arid regions. Ice can also be lost by **calving** of ice margins into bodies of water, including lakes and the ocean as well as by toppling failures at steep ice margins [**Photograph 9.4**].

Mass balance determines the fate of any particular glacier. If mass losses exceed mass gains, a glacier will thin and the position of the glacial margin will retreat upvalley or toward the center of the ice sheet. It is important to realize that when the position of the ice margin changes, the ice itself continues to flow downgradient; ice margin retreat only happens when the rate of ice ablation at the margin exceeds the rate of ice flow to the margin. If mass gains exceed mass losses, a glacier will grow and the position of its margin will advance.

Alpine glaciers respond rapidly to changes in climate that affect mass balances—they advance and retreat on yearly to decadal timescales. In contrast, large ice sheets and ice caps have greater inertia. Although ice sheets can respond quickly to climate change at their margins (by advancing or retreating) or in their accumulation zones

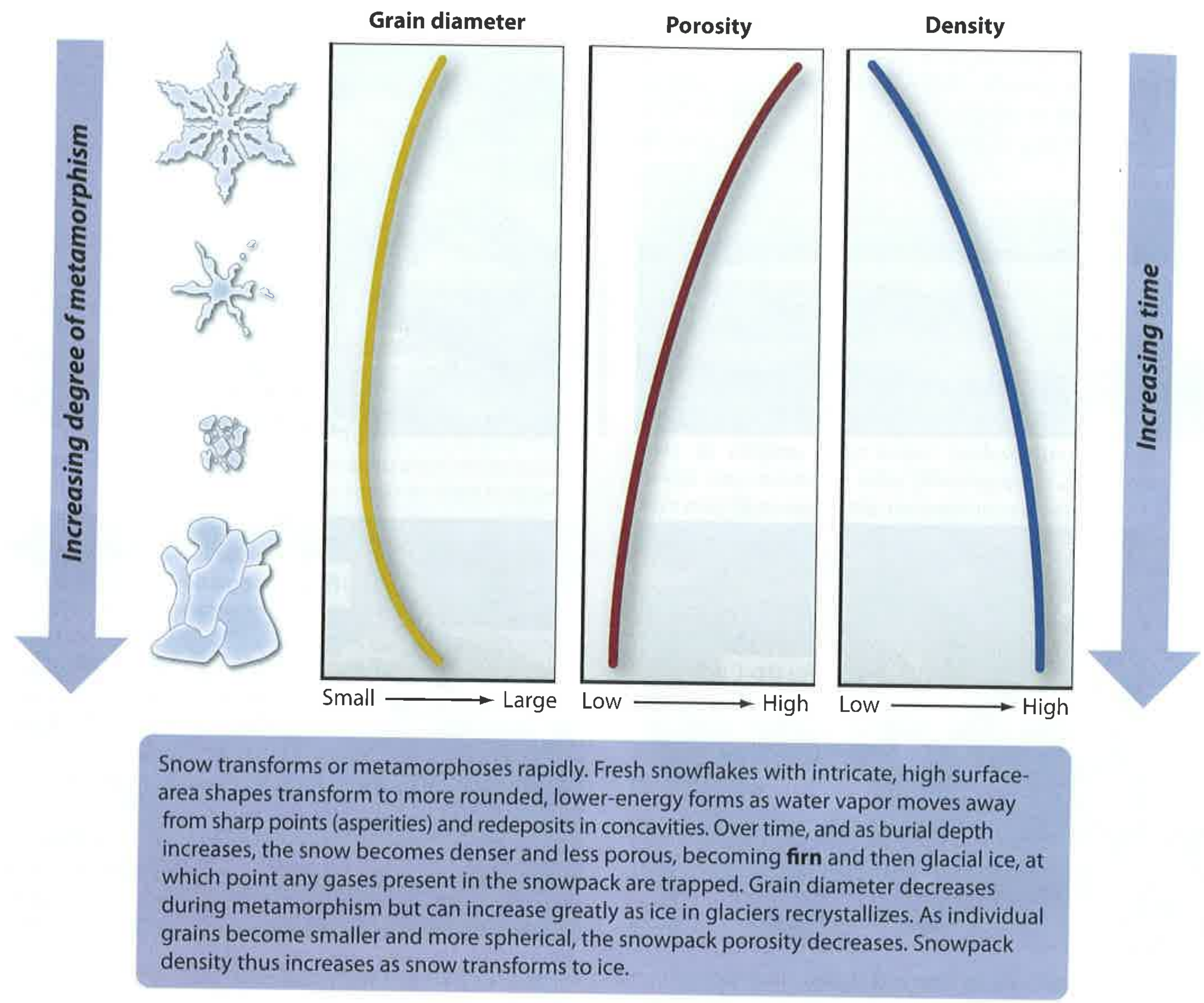
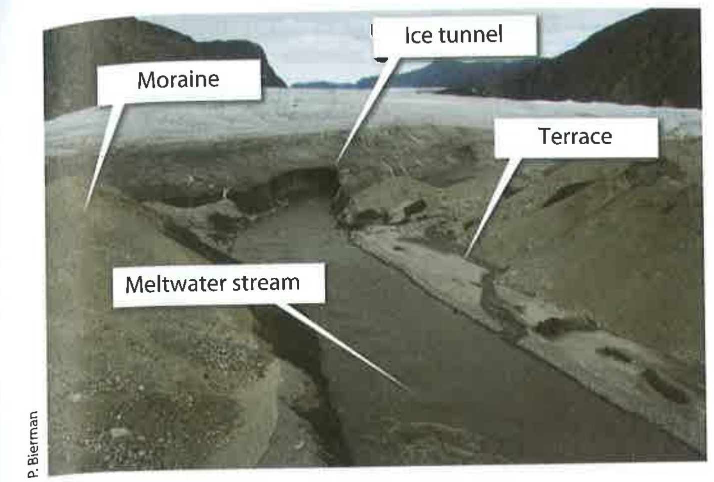


FIGURE 9.3 Snow Metamorphism. Over time, snow metamorphoses, transitioning from intricate forms to rounded grains with lower surface area. Grain diameter, surface area, and snowpack porosity and density change as these transitions continue.

(by thickening or thinning), ice sheets take centuries to respond fully to a new climate state and millennia to disappear in response to changing climate because there is so much ice to melt. For example, the Laurentide Ice Sheet that covered much of North America took more than 12,000 years to largely disappear after the last glacial maximum about 21,000 years ago. The Barnes and maybe the Penny Ice Cap on Baffin Island are thought by some to be the only surviving remnants of the Laurentide Ice Sheet.

Glacier Energy Balance

The energy balance of a glacier is complex but critical to determining whether it will advance or retreat. Energy is lost and gained from glaciers in numerous ways. Sunlight provides energy to the glacier surface, causing mass loss by sublimation (even as the ice remains frozen) and by warming snow and ice until they melt. When the glacier surface is covered by fresh, white, highly reflective snow, its **albedo** (reflectivity) is high and little energy from the Sun is absorbed directly. As snow gets covered in dust, or starts to melt, its albedo lowers, less light is reflected, and the snow absorbs more of the incoming solar energy, increasing the potential for melting. Long-wavelength, blackbody radiation from the ice surface removes thermal energy from the ice, and radiation from the atmosphere and clouds adds energy to the ice. Sensible heat is transferred to the ice by warm air masses and supplies some of the energy to drive the phase transition from ice to water. Large amounts of energy, in the form of latent heat, can be transferred to the glacier surface by the condensation of moisture. A small amount of geothermal heat is transferred from the rock below the glacier. For glaciers, energy



PHOTOGRAPH 9.3 Ice Sheet Margin. Ice tunnel from which large meltwater stream is emerging. Large accumulation of glacial sediment (a moraine) in the foreground was deposited by this outlet glacier draining the Greenland Ice Sheet margin east of Upernavik, west central Greenland. The flat river terrace on the right side of the channel is underlain by ice. The meltwater stream channel is between 50 and 70 m wide.

and mass balances are tightly linked because energy inputs control the melting, sublimation, and calving rates of ice [Figure 9.4].

Accumulation and Ablation of Glacial Ice

For mid- and high-latitude glaciers, spatial and temporal changes in glacier **accumulation** and **ablation** are seasonal and predictable. During the winter months, glaciers gain mass as temperature and incoming solar radiation are low and snowfall rates are high. During the summer months, increased solar radiation, minimal snowfall, and higher air temperatures increase ablation, decrease accumulation, and glaciers lose mass.

PHOTOGRAPH 9.4 Glacier Calving. Calving at Lamplugh Glacier, Glacier Bay National Park and Preserve, southeast Alaska. The ice cliff shown is ~30 m high.

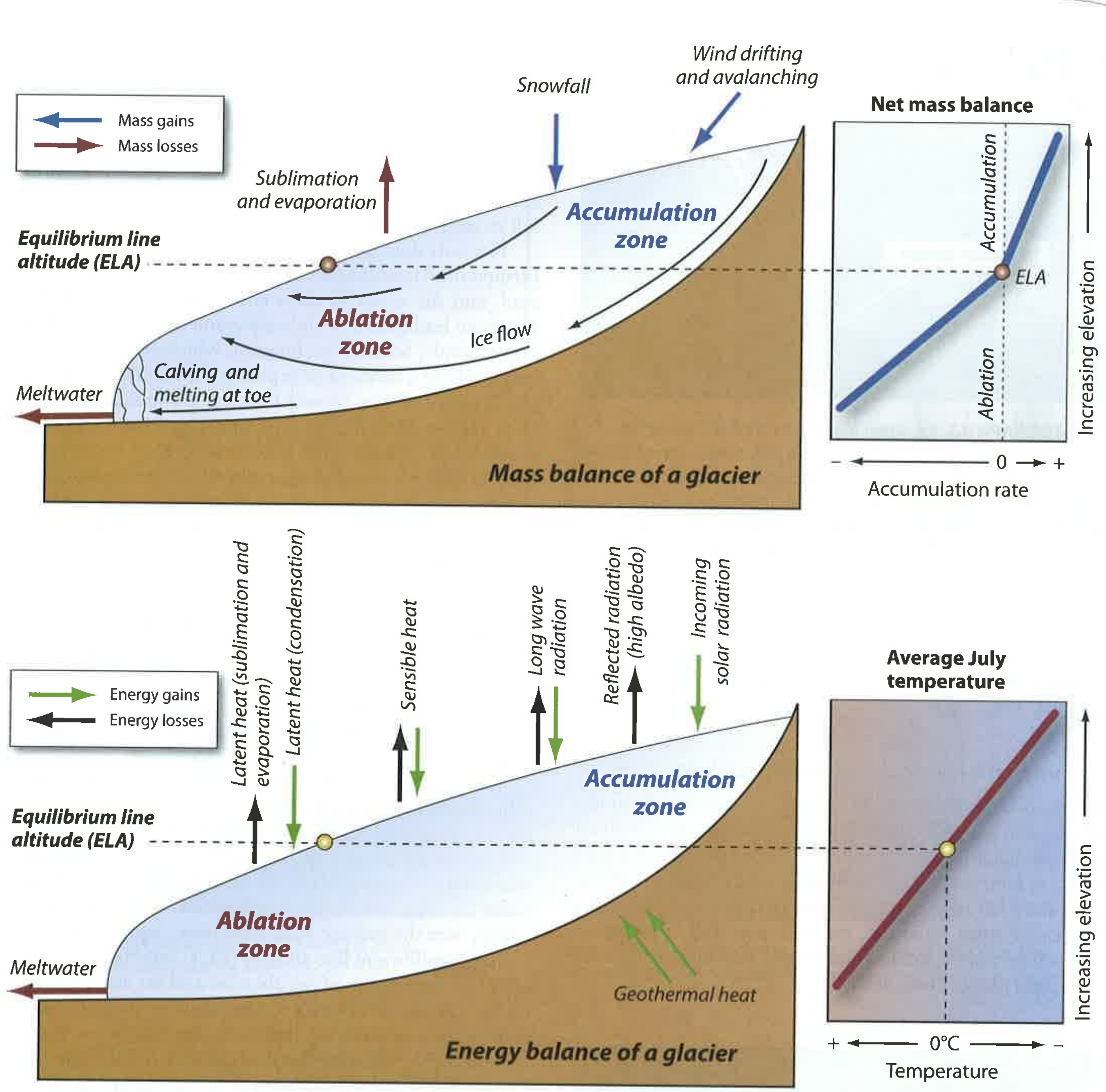
Rates of accumulation and ablation are also tied to elevation. Because air masses cool as they are forced to rise over higher elevation terrain, the average air temperature falls between 0.6 and 1°C for every 100 m of elevation gain. This change in temperature with elevation is termed the **lapse rate.** Thus, in their upper reaches, glaciers will be cooler than in their lower reaches, more precipitation will fall as snow and less snow will melt.

Not only does temperature decrease with elevation, but precipitation tends to increase as air masses gain elevation, cool, and the water in them condenses. This **orographic effect** can lead to extremely steep gradients in precipitation. For example, Seattle, Washington, which sits at sea level, gets less than a meter of precipitation annually; in contrast, the flanks of nearby Mount Rainier, at about 2 km elevation, receive almost 3 m. The upper reaches of glaciers usually receive more total precipitation than their lowest reaches and a greater proportion of the precipitation at higher elevations falls as snow. However, in many places, the winter mass balance is close to constant with altitude; there may actually be a decrease in snow accumulation with altitude on the upper portion of glaciers due to wind-driven snow transport downslope. The summer mass balance is typically much more strongly controlled by elevation, causing strong altitudinal control on most net (annual) mass-balance curves (Figure 9.4). This seasonal disparity in mass balance occurs because in the summer, the lower parts of glaciers tend to be above freezing and thus subjected to melt much more often than the upper parts.

Elevation-dependent energy inputs and mass-balance differences result in distinct zonation of glacial process and glacier morphology. One can demarcate areas of any glacier where net mass loss occurs (the **ablation zone**) and where net mass gain occurs (the **accumulation zone**). These zones are separated by the **equilibrium line** at an altitude where, over the average year, mass losses equal gains.

The **equilibrium line altitude (ELA)** reflects the boundary between the area of net ablation and net accumulation on the glacier surface and is controlled by climate. As climate warms or snowfall decreases, the ELA rises. As climate cools, or snowfall increases, the ELA falls. The ELA can be approximated in the field by measuring the elevation of the **firn line** (the boundary between melting glacial ice in the ablation zone and firn, transformed snow, in the accumulation zone) on a glacier at the end of the melt season in early fall.

There are significant positive and negative feedbacks and interactions that change the chance of glacier survival in a warming climate. Consider the feedbacks that promote the stability of alpine glaciers as climate changes. As the ELA rises, the ablation zone extends to higher altitudes, where the average temperature is lower, limiting melt. In alpine settings, the glacier itself is likely to become more topographically confined and thus better shaded as it shrinks within the valley it occupies. Ice sheets behave very differently. Decreasing accumulation or increasing ablation lessens the ice sheet thickness and lowers the elevation



Mass balance is critical to glacier persistence and size. In the **accumulation zone,** more frozen water is deposited (by snowfall, wind drift, and avalanching) than melts away. This excess mass flows as glacial ice into the **ablation zone,** where it leaves the glacial system by melting, sublimation, and calving. In the ablation zone, there is net mass loss. The **equilibrium line** separates the accumulation and ablation zones. Its elevation is similar to the end of summer snow line and the elevation where, in the northern hemisphere, the average July temperature is about 0°C. If the ELA does not change over time, the glacier is in steady state and its mass balance is stable. However, if the ELA rises, the glacier will lose mass and retreat; if the ELA lowers, the glacier will gain mass and advance.

FIGURE 9.4 Glacier Mass/Energy Balance. A glacier's mass balance is determined both by gains of mass from snowfall and loss of mass by ablation. Energy losses and gains by glaciers are largely a function of elevation because of the correlation between elevation and average annual temperature. Arrows in the glacier show the direction of ice flow.

of the ice sheet surface, raising the temperature in the accumulation zone, thereby melting more ice. As an ice sheet begins to shrink and as the average elevation in its accumulation zone lowers, orographically driven precipitation also diminishes and the rate of ice accumulation declines, a feedback that serves to further shrink the ice sheet.

Glacier Movement

Glacial ice moves in several ways as it responds to the shear stress induced by gravity (see eq. 1.3). All glaciers move as a result of internal ice deformation, a process referred to as **ice creep [Figure 9.5]**. Some glaciers also slide at the bed/ice interface, a process referred to as **basal sliding.** Other glaciers move on slowly deforming fluidized sediment (till) between bedrock and the ice. The flow of glacier ice is laminar (like most lava) rather than turbulent (like most water).

The rate of ice deformation or the **strain rate** ($\dot{\varepsilon}$) varies with shear stress ($\tau = \rho g h \sin\theta$) and can be approximated using an experimentally derived equation known as Glen's flow law with the form of

$$\dot{\varepsilon} = A\tau^n \qquad \textbf{eq. 9.1}$$

where A is a rate parameter that increases with temperature, indicating that warm ice deforms more rapidly than cold ice. Laboratory and field evidence indicate that the exponent, n, has a value of about 3 for ice. Because n is greater than unity, a small change in shear stress causes large changes in strain rate. The model, derived from deforming single ice crystals in the laboratory is consistent with the observation that ice behaves similarly to a plastic material, deforming with increasing rapidity in response to increasing stress **[Figure 9.6]**. Evidence for ice deformation is preserved both in ice cores and at ice margins, including folded and overturned layers of sediment in glacial ice.

Indeed, it is the rapidly increasing rate of deformation with τ that limits the height and surface slope of continental

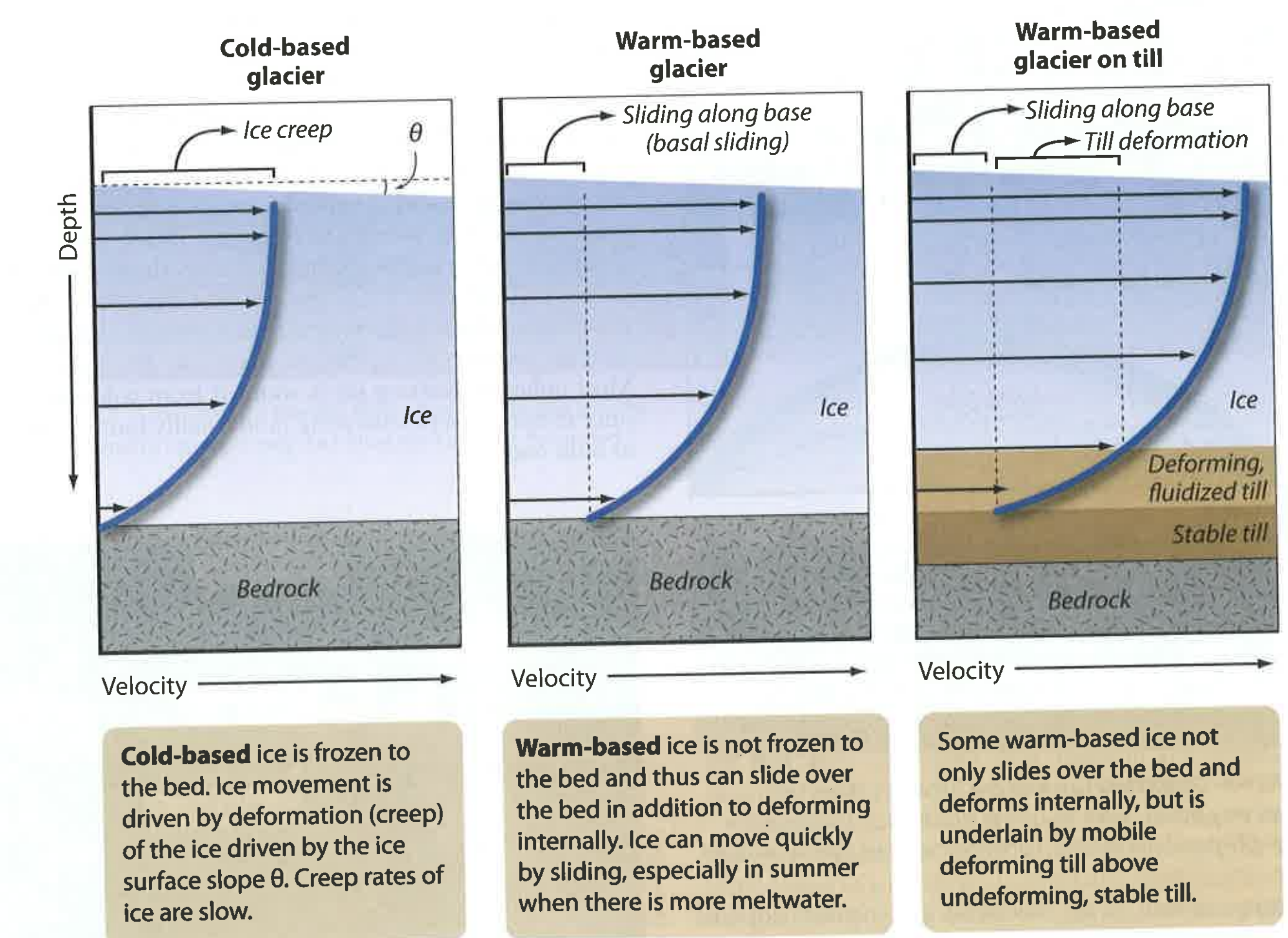


FIGURE 9.5 Glacier Velocity Profiles. Glaciers move as a result of different processes including ice deformation (creep), basal sliding, and deformation of a mobile bed. The relative contribution of each process results in velocity profiles that differ among glacier types.

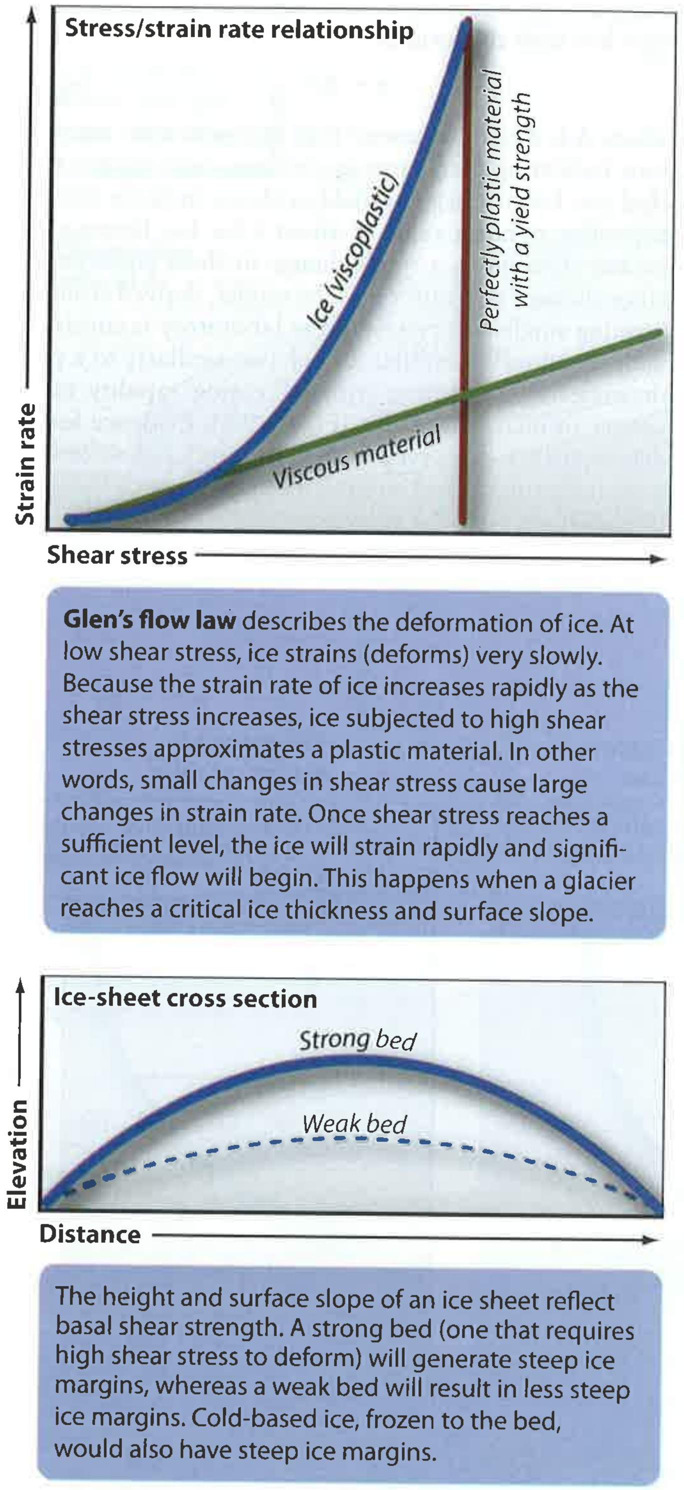


FIGURE 9.6 Glen's Flow Law and the Behavior of Ice. Ice deforms very slowly under low shear stresses. Ice deforms more rapidly at higher shear stresses, behaving more like a plastic material.

ice sheets as well as the thickness and surface slope of alpine glaciers—geomorphically important characteristics because they determine the area of the landscape covered and eroded by ice. To understand why ice sheets can get only so thick, recall eq. 1.3 and consider that increasing slope and increasing thickness both increase shear stress. As shear stress increases, ice, approximating a plastic material, deforms more quickly, moving more ice and limiting further increases in ice thickness.

Not only does the shear strength of ice matter to glacial flow speeds, the shear strength of the underlying bed, if it is deformable, also matters. In fact, the shear strength of the bed material underlying the ice largely controls the shape of the ice sheet. Weak beds, which have a low critical shear stress (τ) for deformation, result in gently sloping ice sheets. Such beds are common where ice overruns weathered material high in clay and silt or where subglacial water is prevalent. Strong beds that only deform under higher τ are found in areas of scoured rock with little basal water or deformable sediment and result in steep ice margins (Figure 9.6).

When glacial ice terminates in water, such as where ice sheets enter the ocean or alpine glaciers enter ice marginal lakes, ice thickness and water depth determine whether the ice will be grounded or whether it will float. Grounded ice maintains contact with the underlying lakebed or seabed, restricting its flow because resistance to movement (basal shear stresses) is relatively high. Floating ice is buoyed by the water below and can both flow and calve very rapidly because there is no resistance offered from the bed under the floating ice; thus, a floating ice margin can catalyze a rapid loss of ice, in some cases significantly drawing down the volume and elevation of the source glacier or ice sheet. If the floating ice does not melt quickly or if it gets pinned on islands offshore or in an embayment, it can become an **ice shelf**, a large, floating mass of glacial ice still connected to its source glacier. Most ice shelves are found around Antarctica [**Photograph 9.5**]. Ice shelves are connected to their source glaciers and lose mass by calving, sometimes in large, dramatic events. Grounded ice and ice shelves buttress ice upstream, reducing the flow rate. Most coherent floating ice is sourced from cold glaciers, since temperate ice is too weak (and usually too fractured) to hold together.

PHOTOGRAPH 9.5 Ice Shelves. Ice shelves, such as the Ross Ice Shelf, are common in Antarctica. Thin seasonal sea ice is breaking up the in the foreground. The thicker ice shelf is in the distance. The ice cliff is ~50 m tall.

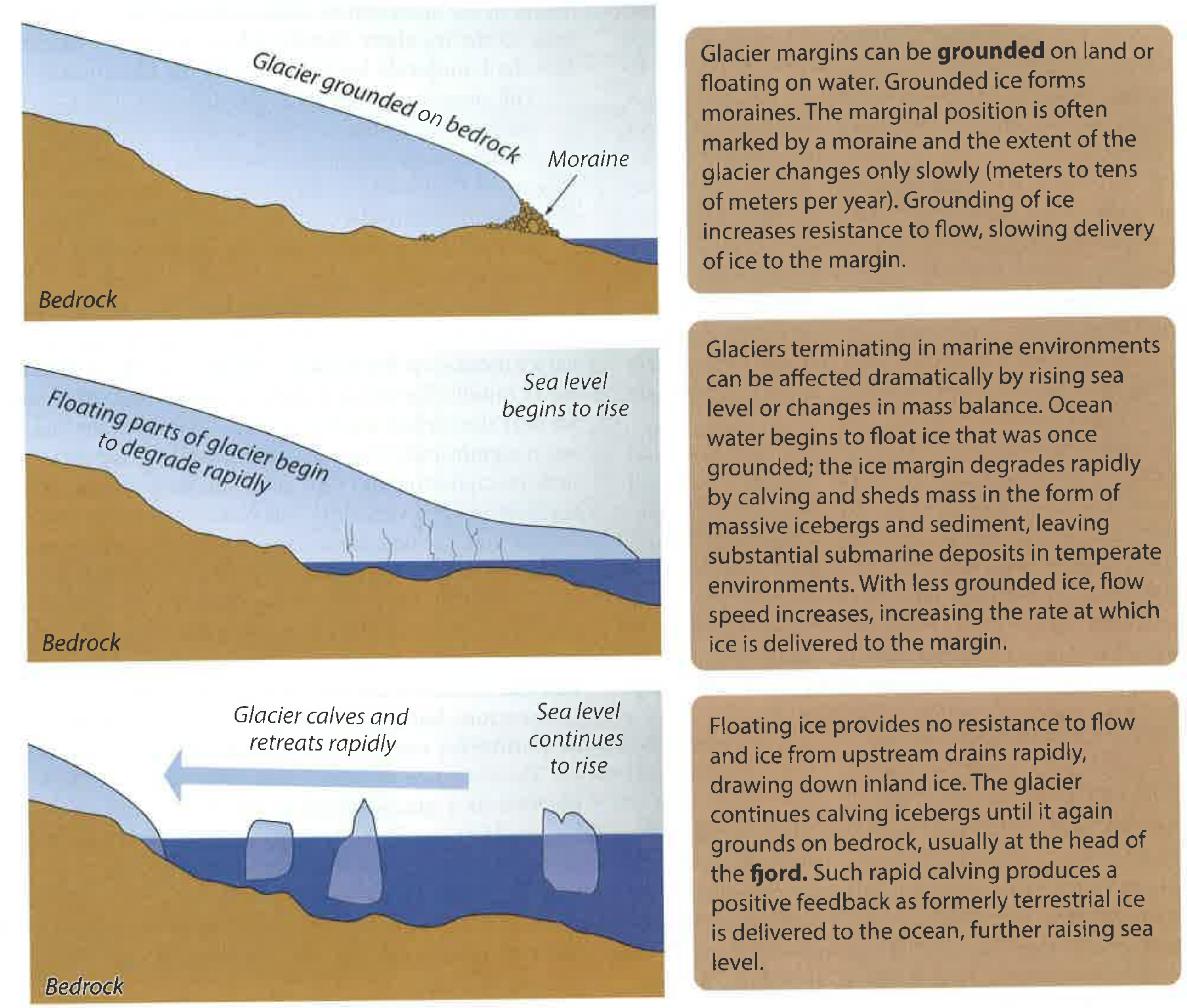


FIGURE 9.7 Grounded Ice Margin. Grounded ice margins tend to be stable and to retreat slowly. Sea-level rise can float grounded ice and result in rapid calving of once-stable ice masses—an important feedback that can hasten deglaciation.

Rising sea levels that floated glacier termini, increased glacier flow rates, and drew down ice sheets at the end of each glaciation contribute to the positive feedback loop that helps end glaciations. As sea level rises, many formerly grounded ice margins begin to float and calve rapidly, adding more mass to the ocean, further raising sea level and thus accelerating deglaciation [**Figure 9.7**]. The calving of the floating ice itself does not cause sea level to rise; however, calving can lead to drawdown of the grounded and land-based portions of the ice mass, which, as ice is transferred from land to sea and melts, increases the volume of ocean water, raising sea level.

The amount and distribution of subglacial water is a critical determinant of the rate of glacial sliding and subglacial erosion as well as the magnitude of **englacial** (within ice) and subglacial sediment transport. High subglacial water pressures reduce the effective normal stress on the bed, thus reducing the shear stress needed to initiate and maintain movement. Glaciers with abundant subglacial water and substantial subglacial sediment can slide rapidly on fluidized beds of saturated sediment, enhancing both the rate of sliding and of sediment deformation. Sediment is transported both along the bed and in water-filled tunnels within the ice (englacial drainages). Large amounts of subglacial water likely increase erosion rates by speeding flow, thus catalyzing rock abrasion and speeding the quarrying process (also known as **plucking**).

Glacial ice flows from the accumulation zone to the ablation zone (Figure 9.4). In the accumulation zone, the net vertical motion of ice is downward relative to the ice surface. In the ablation zone, the net motion of the ice within a glacier is upward, toward the surface. At the ELA, ice moves parallel to the glacier surface and perpendicular to the ELA. Everywhere, basal glacial ice moves parallel to the glacier bed.

A small percentage of the world's glaciers have been observed to **surge**, a phenomenon where ice flow is very rapid (meters to many tens of meters per day) for a short period of time followed by a longer, quiescent period of much slower flow. Because surges can only rarely be related to other events such as changes in climate or the hydrologic system, they are thought to represent instability inherent to the glacial system. During a surge, large amounts of deforming ice move toward the margin, which commonly advances and then stagnates after the surge. Copious amounts of meltwater are typically released during a glacial surge, consistent with the idea that surges are mediated by changes in bed hydrology, specifically hydraulic head increases that reduce resistance to basal slip. Surging glaciers are found around the world but are particularly common in southeastern Alaska and in areas underlain by soft, sedimentary rock.

In a somewhat morbid twist on glaciology and glacial flow paths, rangers in places like North Cascades National Park in Washington State, in the European Alps, and on Mount Rainier sometimes have the grim job of recovering bodies from the ablation zones of glaciers. By keeping careful records of who fell into crevasses, when they fell in, and where the crevasses were located, they can use a body's glacial journey to determine glacier flow lines and flow rates. Conversely, known glacial flow rates can help them figure out who is melting out of the ice.

Thermal Character of Glaciers

Glaciers can be characterized based on their thermal regime both at the bed and at the ice surface. **Cold-based glaciers** are largely frozen to the substrate and thus have little if any liquid water at their beds. **Warm-based glaciers** are not frozen to their bed and are characterized by moist bed conditions where liquid water is abundant, at least during the melt season.

Glacier surfaces can be characterized as polar, subpolar, and temperate. **Polar glaciers** remain below freezing and have no surface melt at any time of year. **Subpolar glaciers** have surface melt during the summer but most of the glacier remains at temperatures colder than the **pressure melting point** of ice, the temperature at which the ice/water phase transition occurs. Because the freezing point of water drops with increasing pressure, water under thick ice can remain liquid at temperatures a few degrees below 0°C (Figure 9.1). **Temperate glaciers,** with the exception of their near-surface zones during winter, have ice at the pressure melting point throughout.

Large ice sheets, such as in Antarctica, are warm-based at the center. There, ice is very thick, the pressure of the overlying ice has depressed the melting point of water, and the thick ice both insulates the base from the cold air above the ice sheet and traps geothermal heat. In Antarctica, radar remote sensing confirms the presence of liquid water at the bed of the ice sheet, revealing at least 150 lakes trapped beneath the ice. These lakes formed in sub-ice bedrock basins and can be quite large, hundreds of kilometers long with volumes up to several thousand cubic kilometers.

Water in the lakes comes from local pressure-melting at the base of the ice sheet. Similar lakes were likely present below the Laurentide Ice Sheet during the Pleistocene.

The situation in Greenland is different. The Greenland Ice Sheet is warm-based along much of its margin where mean-annual air temperatures are at or above freezing [Figure 9.8]. However, large areas of the Greenland Ice Sheet are frozen to the bed; the basal temperature is −9°C at the thickest spot, the summit, where several ice cores have been drilled.

The critical factor accounting for the difference between the Antarctic and Greenland ice sheets is the effect of vertical ice motion on the basal ice temperature. Ice in Greenland flows rapidly due to high snow accumulation rates, so cold ice near the surface is advected rapidly toward the bed, cooling it significantly. Central Antarctica, by contrast, receives little precipitation and thus accumulation rates are low and the flow of ice is very slow and vertical transport of ice does little to cool the bed. Hence, the bed is melting in most places under Antarctica but not under much of Greenland.

As ice thickness and temperature change over time, so do glacier bed conditions. Some areas of Greenland that were cold-based during glacial times are warm-based today and vice versa, reflecting not only changes in mean annual air temperature, but in ice thickness and the advection of colder or warmer ice toward the bed in the accumulation zone.

The presence or absence of significant liquid water at the bed of a glacier (cold-based versus warm-based ice) has implications for the speed of glacial flow, the efficiency of glacial erosion, and thus the geomorphology of glaciated landscapes. Cold-based ice, where water is not present at the bed, flows only due to internal deformation of the ice. At low temperatures, the rate of this internal ice deformation is low (eq. 9.1).

Regions where the ice is at the pressure melting point (warm-based) are subjected to **scouring**, the removal, by ice sheets, of both rock weathered during interglaciations and fresh rock from below the weathered zone. Much of the bare rock landscape of the Canadian Shield was scoured by warm-based ice. Exposed bedrock in these areas displays **p-forms,** features such as grooves and other streamlined landforms eroded into rock by flowing ice. In contrast, landscapes once occupied by cold-based ice retain intensively weathered rock as shown by the presence of deep saprolite, bedrock tors, and weathering pits in outcrop surfaces. The best clue that ice once covered these surfaces is the presence of isolated erratics left behind when the ice melted away.

Both field observations and very old cosmogenic nuclide exposure ages (Chapter 2) suggest that many now-deglaciated Arctic uplands in Baffin Island, Scandinavia, and Greenland were once covered by non-erosive, probably cold-based ice that was frozen to the bed and slowly moving relative to the fast-moving, wet, erosive, warm-based ice in the surrounding lowlands. Over time, such a contrast in thermal regime, and thus erosion efficiency, can erode large amounts of rock in valleys, deepening fjords while leaving upland surfaces relatively unaltered.

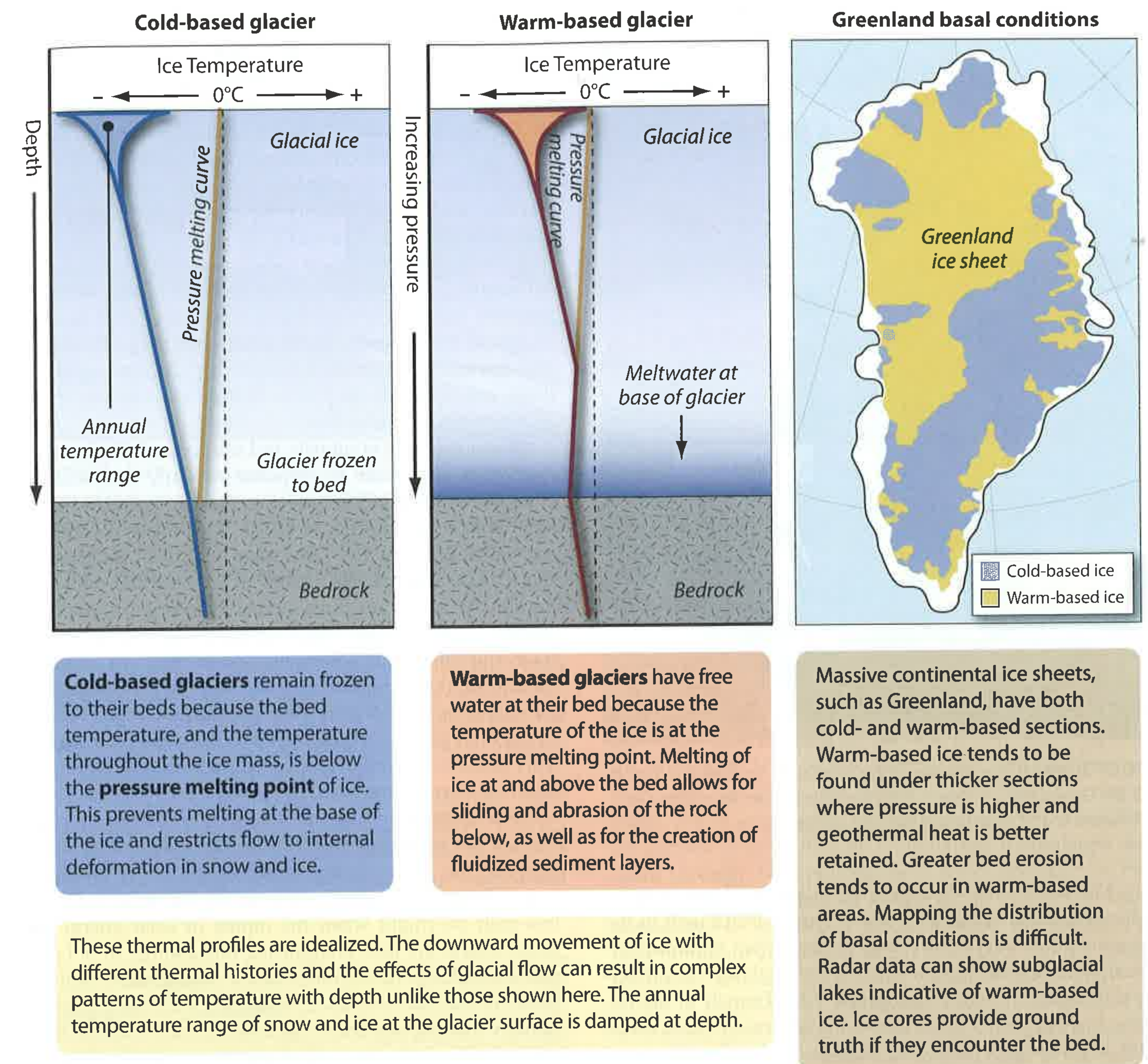


FIGURE 9.8 Cold-Based and Warm-Based Glaciers. Glaciers can be cold- or warm-based. Cold-based glaciers are frozen to their bed with ice temperatures below the pressure-melting point. Warm-based glaciers have basal temperatures at the pressure-melting point. Large ice sheets and ice caps with outlet glaciers have basal thermal regimes that differ over time and space.

Glacial Hydrology

In many ways, glacial hydrologic systems are similar to those found in karst terrain (Chapter 4), with the distinction, of course, that ice melts and carbonate rocks dissolve. Glacial ice, like dense limestone, has low intrinsic permeability, and water moves only slowly through pores in the ice. Both glacial and karst hydrology are dominated by pipe flow; most water and sediment moves through macropores such as fractures that become enlarged into tunnels. Clear evidence of pipe flow is provided by dye-tracer experiments that indicate subglacial water speeds up to 0.5 m/s, far faster than glacier flow (centimeters to meters per day). Radar and bore-hole observations of temperate glaciers further reveal large crevasses extending to within a few meters above the bed. When the summer melt season is in full swing, there is a well-connected plumbing system within most glaciers, connecting the top and sides with the bed and points in between and thus accounting for the mobility of meltwater.

K. Steffen

PHOTOGRAPH 9.6 Moulin. Large moulin on the western margin of the Greenland ice sheet brings meltwater from the surface to the interior and probably to the bed of the glacier.

Much like the disappearing rivers of karst terrain, glacial streams accumulate water from surface melt in the summer, flow over the ice, and plunge down **moulins**, vertical conduits, to become part of the englacial or subglacial drainage system [Photograph 9.6]. Tunnels in the ice, which would otherwise close as ice deforms around them, are kept open by advected heat as well as by frictional heating from turbulent water flow driven by head gradients within the ice-bound hydrologic system. Because water flows through glaciers in response to the pressure gradient (differences in hydraulic head), sediment and water can be transported "uphill" (with respect to the underlying terrain) and some sub-ice channels climb valley sides.

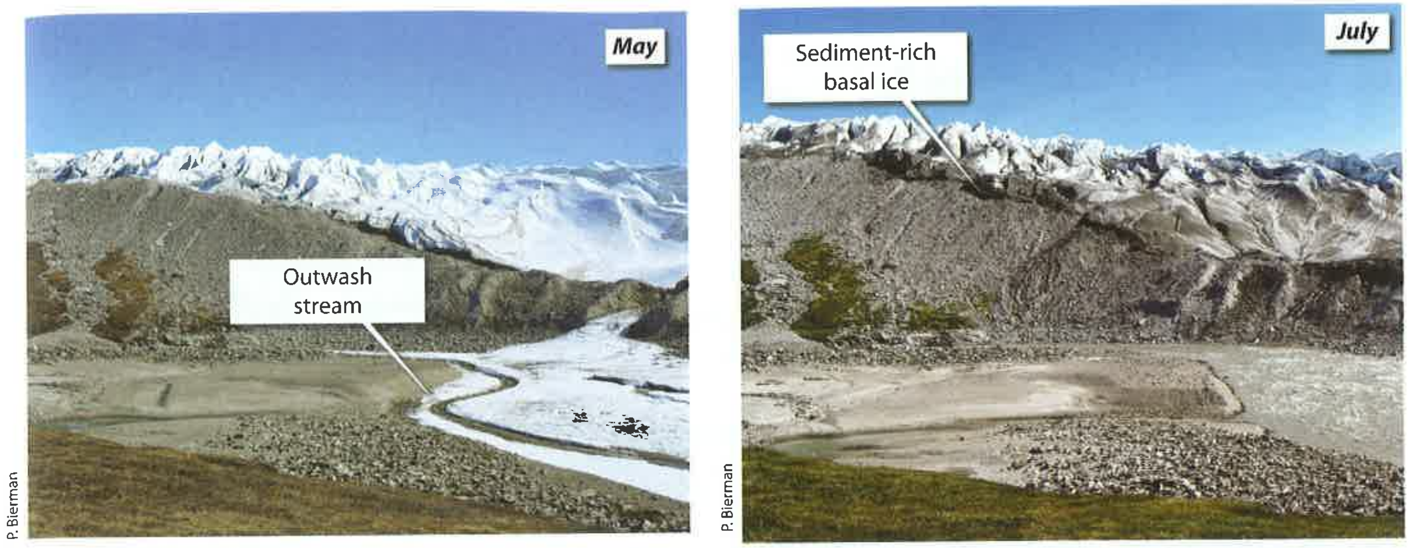
Both theory and field observations suggest a close connection between the character of the bed and the character of the sub-ice drainage network. Under most glaciers, water is present in a variety of places, including pore space in rock and sediment, films between rock and ice, cavities in the ice and rock, and large channels cut into the bed and in the ice. Where the substrate is hard and not deformable, such as over much of the Canadian Shield, drainage will organize into a few large, long-lived channels as flow concentrates. In contrast, where the substrate is soft, deformable, and saturated, there are many small, broadly distributed sub-ice channels that are ephemeral. Thus, it is no accident that water-sorted sediment deposits from former sub-ice channels are found in areas dominated by hard bedrock exposures rather than deformable sediments.

Subglacial drainage influences glacier dynamics, erosion, and sedimentation because glacial hydrology controls in large part the speed of glacier flow (consider also surging glaciers) and the propensity for glaciers to modify their beds. Warm-based ice moves more rapidly than cold-based ice because warm-based ice slides over the bed and over the deformable, wet sediment layer beneath. Subglacial water pressure reduces the effective normal force and thus the frictional resistance to shear at the bed (see Chapter 5). This increases the speed at which the bed deforms and the ice flows, at least in regions that previously were moving slowly or had limited basal water.

Water pressure in tunnels and sub-ice cavities fluctuates over time and space in response to daily and seasonal changes in meltwater production. For example, during the melt season, the speed at which ice moves toward the western margin of the Greenland Ice Sheet increases as more meltwater reaches the bed. It is likely that daily fluctuations in water pressure increase the rate of subglacial erosion by altering the stress regime on the bed, fracturing rock and entraining sediment, and perhaps by inducing **cavitation**, the formation and violent collapse of small gas bubbles that can fracture even the hardest rocks because of the high pressures this stress exerts on the rock surface. This process is similar to that contributing to erosion in bedrock river channels (Chapter 6).

Discharge from glaciers peaks in late summer, when the ice has warmed from long summer days and above-freezing air temperatures [Photograph 9.7]. Discharge from glaciers is usually lowest in the morning, reflecting less melt overnight when the inputs of solar energy and sensible heat are low, even during midsummer in polar regions. Discharge in streams sourced from glaciers typically peaks in the late afternoon, reflecting the effectiveness of midday melting and some lag time for meltwater to move toward the glacier margin. Early European explorers in central Asia, far from the glaciated peaks, learned the hard way about meltwater pulses that reached them out of phase with midday heating, often after they had camped near what seemed to be nearly dry streambeds. If you need to cross a glacial outwash stream during the midsummer melt season, it is best to do so in the morning.

The amount of meltwater reaching the bed of ice sheets and glaciers is a critical control on both glacial erosion and ice flow and may well be one of the most important ways by which climate change influences the distribution and stability of glaciers. For example, increasing summer season melt along the margin of the Greenland Ice Sheet has expanded the number and size of meltwater lakes on the surface of the ablation zone [Photograph 9.8]. Some of these lakes drain catastrophically through fractures in the ice and in no more than an hour or two deliver large amounts of water directly to the bed of the ice sheet.



P. Bierman

P. Bierman

PHOTOGRAPH 9.7 Seasonal Changes. Compare these two images of the same location on the western margin of the Greenland Ice Sheet where ice is flowing from left to right. In the left image, taken May 2011, the ice is still snow-covered and the outwash stream is so small that one could step over it. In July 2008, at the height of melt season, it was very different. The outwash stream was roaring and all of the snow had melted, exposing sediment-rich basal ice at the margin of the ice sheet. Field of view is about 1 km wide.

Presumably, drainage occurs when a threshold is reached, allowing the water pressure to wedge open fractures until the permeability of the ice suddenly increases. Added water at the bed increases the hydraulic head, decreases the effective normal force, and speeds glacier flow by lowering shear strength of the bed/ice interface.

Occasionally large floods, termed **jökulhlaups**, from the Icelandic words jökull (glacier) and hlaup (burst), issue from beneath ice sheets, glaciers, and glacial deposits that dam lakes. Large jökulhlaups can have exceptional discharges, on the order of 50,000 m^3/s, within the range of flow of the Amazon and Mississippi rivers. Such floods transport massive amounts of sediment and have buried outwash stream valleys beneath tens of meters of sand and gravel. In Iceland, jökulhlaups are frequent due to the locally high heat fluxes from the volcanic terrain that melt ice at the base of ice sheets and glaciers. Sub-ice eruptions that melt glacial ice can trigger catastrophic releases of water. Many of Iceland's outwash plains, termed **sandurs**, are formed in large part by jökulhlaups. Jökulhlaups also occur through the rapid release of water stored at the bed of the glacier or failure of glacial dams. Failures of lakes impounded by glaciers or glacial deposits are common in high mountain areas such as the Himalaya and the Andes of South America. Dangerous outburst floods may increase in frequency as global warming causes glaciers to retreat and generates more meltwater. The great Missoula floods that scoured eastern Washington State at the end of the Pleistocene (see Digging Deeper, Chapter 13) and even larger glacial floods in central Asia were jökulhlaups caused by the failure of massive ice dams.

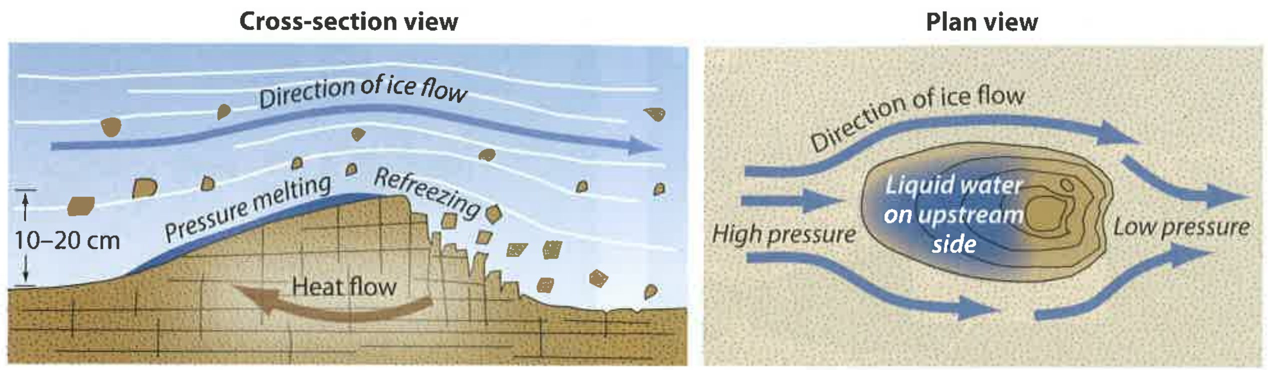
I. Joughin

PHOTOGRAPH 9.8 Meltwater Lake. Meltwater lake on the surface of the Greenland Ice Sheet near Ilulissat. Such lakes form each summer and can add large amounts of meltwater directly to the bed of the glacier when they drain catastrophically through fractures. Lake is several kilometers wide.

Subglacial Processes and Glacial Erosion

There are several processes by which glaciers erode rock and incorporate unconsolidated material. All are related through the distribution of pressure, water, and connected cavities at the base of the glacier, the interface between the ice and the bed.

The bed of a glacier is rough on a variety of scales generally due to bedrock topography. As warm-based glaciers slide over their beds, cavities can form on the down-ice side of bedrock bumps. **Regelation** can be an important means by which glacial ice moves around small-scale (less



Regelation occurs when ice flows over rock obstacles on the glacier bed. Pressure-melting of ice up-glacier of an obstacle releases water, which then travels down ice to areas of lower pressure and consequently refreezes. This refreezing may incorporate debris and enrich the remaining solution in elements such as calcium, which precipitates and forms calcium carbonate coatings on subglacial rock surfaces. Regelation facilitates the movement of warm-based glacial ice over rough bedrock beds.

FIGURE 9.9 Regelation. Regelation is an effective means of ice flow over rock at the base of glaciers as meltwater moves around small basal roughness elements, on the scale of centimeters to tens of centimeters, and refreezes on the down-ice side.

than tens of centimeters) bumps in the bed. Regelation describes the pressure-melting of ice upstream of an obstruction on the bed where pressures are higher, the movement of that water downgradient, the transfer of heat through the bedrock bump, and the subsequent refreezing of water (and attendant release of latent heat) downstream of the bump in the lower-pressure zone [**Figure 9.9**]. The refreezing process can incorporate debris into the ice as well concentrate dissolved elements in the remaining solution. As a result, some dissolved elements precipitate, such as calcium carbonate that forms rinds on subglacial bedrock outcrops [**Photograph 9.9**].

At a larger scale than regelation, roughness or bumps on the bed below the glacier produce cavities in the ice, changing the distribution of pressure on those bumps. The presence of large sub-ice cavities shifts the load of the ice to bedrock outcrops, especially during times when water pressure below the glacier is falling. Increased pressure on the outcrops can lead to fracturing of the rock. Removal of the fractured rock by entrainment in ice is accomplished as ice flows around and drags the loosened material. Some debris may also freeze onto the base of the glacier. Field evidence for bedrock **quarrying** is clear—large blocks of rock are conspicuously removed along joint planes from outcrops [**Photograph 9.10**]. Most quarrying occurs in low-pressure regions like the down-ice side of hills because large cavities form most readily in such places.

The effectiveness of glaciers as agents of erosion varies spatially, reflecting differences in the intensity and character of erosion processes such as abrasion, quarrying, and glaciofluvial activity as well as the character of the rock below the ice, particularly the spacing of joints. Subglacial and ice-marginal fluvial erosion can be significant in areas of copious meltwater, as evidenced by deep potholes and incised marginal channels once covered by ice but now revealed by deglaciation. Glacially mediated incision can cause erosion near the glacial margin; for example, valley deepening by glaciers may oversteepen adjacent valley walls, promoting mass movement by rockfall onto the ice surface—the source of much **supraglacial** (on top of the ice) debris (see Digging Deeper).

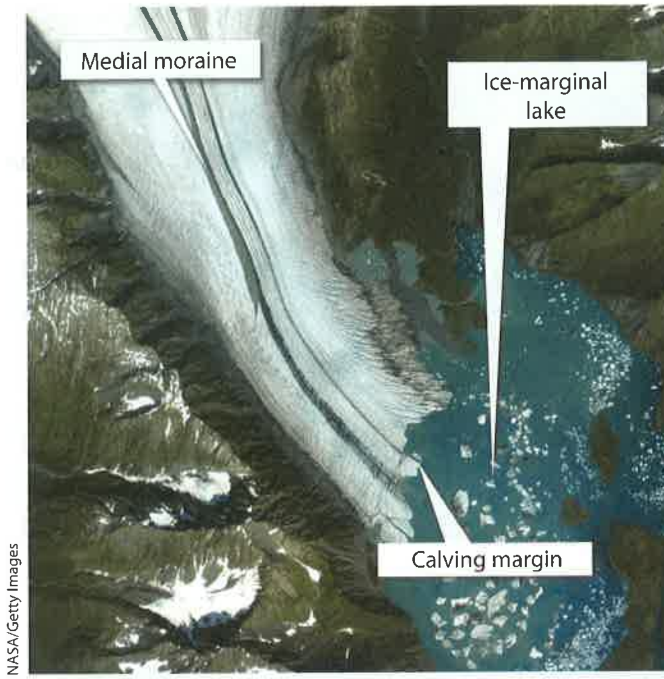
Much of what we know quantitatively about rates of glacial erosion has been inferred from measuring the volume

B. Hallet

PHOTOGRAPH 9.9 Subglacial Carbonate Precipitation. Subglacial calcium carbonate precipitates that grew under the ice of Tsanfleuron Glacier, Switzerland. Pocket knife for scale; blade points in direction of ice flow.

J. Briner

PHOTOGRAPH 9.10 Glacial Bedrock Quarrying. Jointed and glacially scoured bedrock on western Greenland is ready to be quarried by the next ice advance. This surface was covered by the Greenland Ice Sheet during the Little Ice Age.



NASA/Getty Images

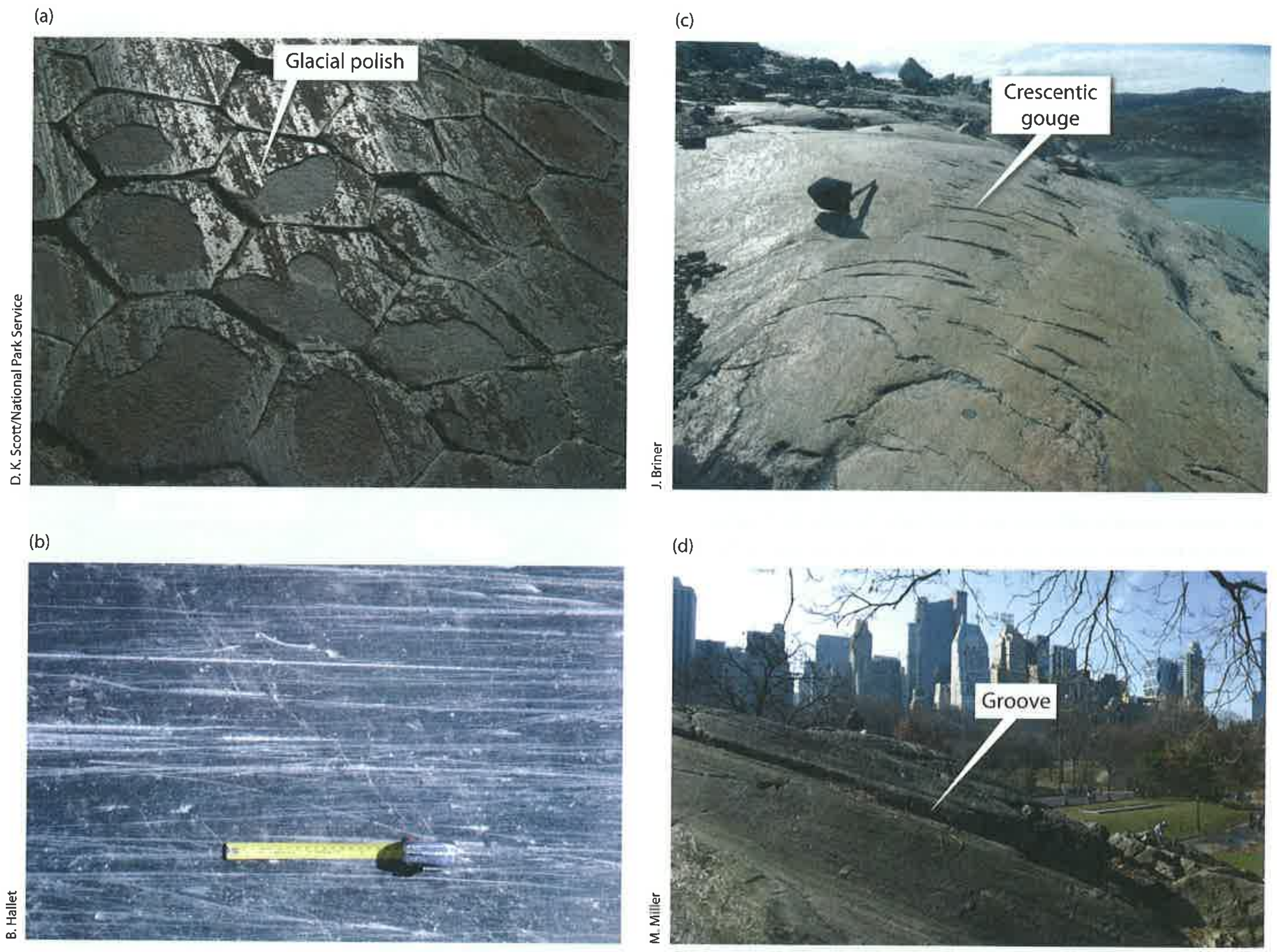
PHOTOGRAPH 9.11 Ice-Marginal Lake and Rock Flour. This ice-marginal lake, bordering the calving margin of Bear Glacier on the Kenai Peninsula of Alaska, is blue-green in this natural-color image because very fine rock flour, material abraded by glaciers off outcrops at the base of the ice, is suspended in the lake water. The color of the water depends on the size and abundance of suspended particles. Note the medial moraine.

of sediment carried by glaciers and deposited outside glacial margins and from sediment flux measurements made downstream of glacial margins. Recent work has used tunnels, bore holes, and other means of both observing active processes and placing instruments within and under active ice to better understand glacial erosion processes. For example, time-lapse cameras have been placed in sub-ice cavities, and well-characterized blocks of rock have been placed under flowing glacial ice to measure erosion rates directly over short periods of time.

Glaciers can be effective agents of erosion although rates of glacial erosion vary over time and space. Warm-based glaciers in areas with high precipitation rates (such as southeastern Alaska) erode rock at rates between 1 and 1000 mm/yr with the higher rates over short time frames of years to decades and the lower rates averaged over the last million years. Glaciers in areas with less precipitation erode more slowly, as little as 0.1 mm/yr. Cold-based ice, which is frozen to the bed, does little to erode and entrain subglacial material. Cold-based glaciers preserve the landscapes they cover.

Glacial abrasion produces some of the most distinctive and characteristic evidence of glaciation including **striations, grooves, glacial polish, loess,** and **rock flour.** Rocks of all sizes form tools in the ice that abrade the rock below as they move across the bed while the warm-based ice or mobile sediment layer above is slowly deforming. The deforming ice above moves these erosive tools as it shears. One result of this abrasion is **rock flour,** the finely ground, silt-sized, rock fragments that colors streams issuing from glacial margins a distinctive milky blue-green [**Photograph 9.11**]. Rock flour is important geomorphically and to society. It is the source of much of the fine sediment that, after being carried away from active ice margins by wind and later deposited, forms regionally extensive blankets of **loess** (fine, wind-blown sediment, derived from the German word for "loose"). Loess underlies much of the world's most productive farmland such as the Palouse of eastern Washington State, the midwestern United States, the Russian Steppe, and the Chinese Loess Plateau. Once deposited, loess can erode easily if vegetation removal exposes it to the direct erosive action of water or wind.

Other evidence for abrasion includes smooth **glacial polish** [**Photograph 9.12a**] that typifies fresh outcrops of rock emerging from under glacial ice. **Glacial striations** [**Photograph 9.12b**] are the fine grooves and scratches left on polished rock surfaces by the movement of debris-rich basal glacial ice or deforming basal sediment. **Crescentic gouges** (large semicircular cracks on the rock surface) or chattermarks also form from the pressure of large abrading clasts on the bedrock below the ice [**Photograph 9.12c**]. **Glacial grooves** of varying scales can be found on many outcrops aligned with the paleo ice-flow direction [**Photograph 9.12d**]. Grooves, striations, and gouges are routinely used as flow direction indicators for now-vanished ice sheets. These erosive forms are distinctive; some have survived burial over hundreds of millions of years to document ancient glaciations (such as the Permian glaciation of Australia, Photograph 9.1). Together, quarrying and abrasion can yield streamlined forms indicative of glacial erosion known as **roches moutonnées** (French for "rock sheep"),



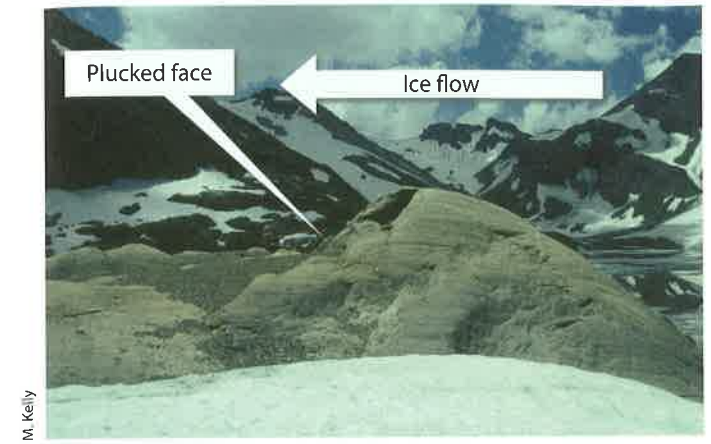
PHOTOGRAPH 9.12 Glacial Abrasion. Features of glacial abrasion come in scales from microns to meters. (a) Shiny glacial polish on columnar basalt column at Devil's Postpile, California, is the result of abrasion and polishing on the micron scale. Each basalt polygon is about 80 cm wide. (b) The surface of this limestone rock in Alberta, Canada, is striated and smoothed. Tape measure provides scale. (c) Crescentic gouges on polished bedrock covered by the Greenland Ice Sheet during the Little Ice Age. Ice flow was in the direction that the shovel handle is pointing, away from the photographer. (d) Glacial grooves in Manhattan Schist, Central Park, New York City, are a few tens of centimeters deep. Flow was parallel to the direction of the grooves.

with gently sloping, smooth, abraded up-ice sides and steeper, plucked down-ice sides [**Photograph 9.13**].

Glacial erosion modifies valleys into distinctive shapes (see Digging Deeper). Ice flowing out of cirques moves down V-shaped stream valleys, preferentially eroding and widening lower valley walls, thereby leaving behind U-shaped valleys [**Photograph 9.14**]. In drainage basins where small tributary glaciers flow into larger trunk glaciers in major valleys, these tributary glaciers are significantly less efficient at eroding rock than the larger trunk glaciers. When the ice melts away, the tributary valleys are left hanging—sometimes hundreds of meters above the floor of the main valley (Photograph 7.12b).

The most dramatic example of glacial erosion is the excavation of deep troughs, known as glacial valleys inland and **fjords** along coastlines where they are at least partially filled by seawater. The location of many fjords appears coincident with areas of increased bedrock fracture density (the result of tectonic forces in the past). Fjords occupy valleys first deepened by fluvial then glacial erosion. They are most common along wet, windward, maritime coastlines where large amounts of ice move through glacier systems, the result of high rates of precipitation and glacier activity. The physical properties of ice makes glaciers an efficient eroder of fractured rock, creating deep fjords and valleys by plucking jointed rock in tectonically inactive cratonic settings such as Norway and Greenland.

Fjords provide a good example of positive feedback. Ice concentrated in valleys is thicker, warmer, and thus flows faster than ice on adjacent highlands, increasing the efficiency of valley-bottom abrasion and quarrying processes. The thicker ice increases the pressure on the bed, lowering the pressure melting point and making it more likely that ice



PHOTOGRAPH 9.13 Roche Moutonnée. Roche moutonnée shows ice movement from right (smooth slope) to left (plucked face) near Lac d'Emosson in the western Alps, Switzerland (person to the left for scale). The roche moutonnée is several meters tall.

in the fjord is warm-based, further enhancing the efficiency of erosion. As erosion is concentrated in the troughs, ice there becomes increasingly efficient at erosion, further deepening the troughs, making a positive feedback loop. Some fjord troughs are so deep they extend well below even the lowest sea levels of the Quaternary, those coincident with past glacial maxima, when ice sheets were at their largest.

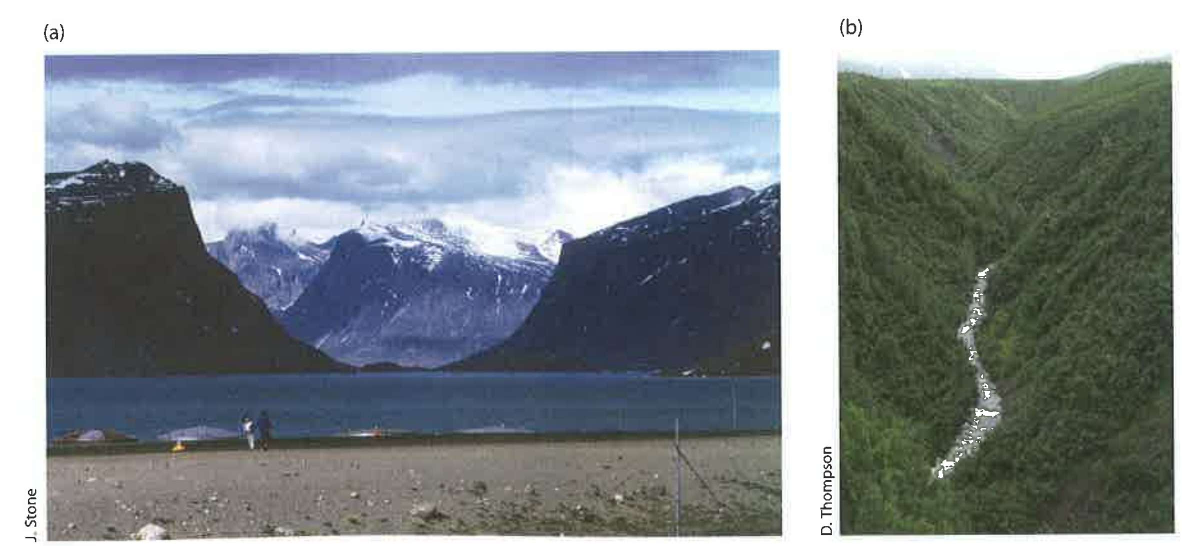
The limited relief above numerous glacial cirques high in many mountain ranges, the parallel along-range and cross-range trends in ELA and summit elevations, and the steep topography that such glacial activity creates led to the theory that glacial activity limits the elevation of mountain ranges by chewing away at cirque headwalls. The idea that glacial processes efficiently remove rock elevated some distance above the ELA, and can thus remove material at about the rate that tectonics and/or isostasy elevates it, has been informally termed the **glacial buzzsaw.** While DEM analyses suggest that the glacial buzzsaw mechanism appears relevant in mountain ranges around the world, including the Washington Cascades, the South American Andes, and the Sierra Nevada of California, the buzzsaw is not perfect; the elevation of some mountain peaks extends well above the snowline.

Glacial Sediment Transport and Deposition

The sediment carried by glaciers is transported within (**englacial**), on (**supraglacial**), and under (**subglacial**) the ice before it is eventually deposited on the landscape. The proportion of material transported in, on, and under ice differs among glaciers and depends at least in part upon the underlying landscapes they are eroding. Some glaciers, particularly alpine glaciers incised deeply into weak rocks, carry large amounts of rockfall material supraglacially because rockfall brings in large volumes of debris from above the glacier; indeed, so much material may cover the glacier from rockfall that ice is barely if at all visible [**Photograph 9.15**]. Conversely, large ice sheets that overwhelm topography carry little if any material on their surface. Rather, ice sheets carry most sediment either in the lowermost basal ice or below the glacier in mobile beds of sediment. During transport at or near the bed of a glacier, sediment is broken down or **comminuted** by interaction with nearby grains or the rocky bed below the ice.

Subglacial Sediments and Landforms

The deforming bed and abundant liquid water below warm-based glaciers create diagnostic sediments and landforms that can be preserved when glaciers and ice



PHOTOGRAPH 9.14 Glacially Eroded (U-shaped) and Fluvially Eroded (V-shaped) Valleys. (a) U-shaped glacial valley near Pangnirtung village on Baffin Island, Canada. (b) V-shaped fluvial valley in Cantwell, Alaska.

D. Thompson

PHOTOGRAPH 9.15 Glacier Ablation Zone. The ablation zone of the debris-covered Emmons Glacier in Mount Rainier National Park, Washington State. High debris load results from rapid weathering of weak volcanic rocks. Most of the debris cover is the result of a single rockfall event in 1963 from nearby Little Tahoma Peak. Recent retreat of this glacier has isolated a lateral moraine on the right, leaving a trimline (area cleared of weathered material by glacial abrasion, often a zone of little or no vegetation) on the cliff to the left. Field of view is several kilometers wide.

P. Bierman

PHOTOGRAPH 9.16 Till. Unsorted, consolidated lodgment till in northern Vermont holds a near-vertical slope. The till is exposed by a landslide where the Huntington River undercuts the bank. Width of view is about 10 m.

sheets vanish. The most ubiquitous sediment left by glaciers is **till** [**Photograph 9.16**], a specific type of **diamicton** (any unsorted, unstratified deposit). Till is deposited both from below the ice (**lodgment** or **basal till**) and from above the ice as ice melts away (**ablation till** or **melt-out till**). Lodgment till tends to be highly compacted (i.e., low pore space, high density) and may show preferred clast orientations or **fabric**, as well as some fissility, a tendency to split along preferred directions of breaking or weakness, the result of emplacement under the effective pressure of overlying ice (the weight of the ice minus the subglacial water pressure). Ablation till tends to be more heterogeneous and less dense because it was not compacted by the weight of overlying ice. All tills may contain pockets of better-sorted, water-washed material.

Till plains cover large areas that were occupied by continental ice sheets during the Pleistocene, including much of northern North America, Europe, and Asia. Formerly glaciated terrain usually contains erratics, boulders transported from afar on or in glacial ice that were stranded when the ice melted away (see chapter-opening photo). Erratics are often easily recognized because they are different from local rock types or have been weathered very differently than the bedrock on which they sit.

Ice-Marginal Sediments and Landforms

Ice margins are geologically complicated places where numerous different geomorphic and sedimentologic processes occur simultaneously. Thus, the texture and sorting of sediment deposited at and near ice margins varies greatly over short distances and ice-marginal sediment is often disturbed by faulting and folding related to the melting of buried ice blocks, overrunning by advancing ice, and/or by collapse after the ice melts away. Much of the heterogeneity in the ice-marginal environment is due to the interplay of sediment deposition directly from the ice, from meltwater, and from mass movements of sediment and sediment-laden ice. A variety of **ice-contact landforms** are created at and near ice margins. These are known by a variety of different names and are created by several different processes.

Adjacent to former ice margins, one often finds small depressions that are nearly circular in plan view. These **kettles** [**Photograph 9.17**] result from subsidence of glacial sediments where ice blocks, buried during rapid, ice-marginal sedimentation, melt out, leaving large depressions in the land surface that may fill with water if the water table intersects the depression.

If sediment fills voids in the ice and that ice later melts away, then the sediment will form a high point on the landscape. If the deposit is circular in plan form, it is referred to as a **kame** [**Photograph 9.18**]. Kames contain both sorted sediment, deposited by flowing water and unsorted material, either till deposited by ice or debris flow sediment originating from the ice surface. **Kame terraces** are benches of fluvial sediment deposited by ice-marginal streams; these streams once flowed between the ice and the valley walls. Where ice-contact deposits are elongated perpendicular

R. Strimbeck

PHOTOGRAPH 9.17 Kettle Pond. This kettle pond (~20 m wide) is forming as an ice block melts below a depositional terrace, downstream of the Hugh Miller Glacier, Alaska.

J. Briner

PHOTOGRAPH 9.19 Esker. A small, sinuous esker is all that is now left of a glacial tunnel once filled by meltwater. The esker was deposited beneath the western Greenland Ice Sheet during the Little Ice Age.

to the direction of ice flow, they have been interpreted as **crevasse fills,** the result of sediment-carrying meltwater filling open crevasses. Crevasse fills tend to have more complex patterns in plan view than kame terraces. In all of these deposits, the presence of ice is indicated by faulting and partial collapse caused by the ice beneath and alongside the deposited sediment melting away.

Eskers are elongate, sinuous ridges composed of water-laid, sorted sediment capped and underlain in many cases by till [**Photograph 9.19**]. Eskers are thought to be remnants of subglacial drainage systems, the traces of sand and gravel-filled tunnels that once drained meltwater and sediment near now-vanished ice margins. Eskers often terminate in sediment fans or deltas. Because eskers would be destroyed by significant ice movement, their length and presence implies an immobile or slowly moving zone of ice at the glacial margin, most likely an area where the ice was too thin to deform rapidly. Eskers may also occur where ice flow and subglacial water flow were parallel.

P. Bierman

PHOTOGRAPH 9.18 Kame. Eroding till and ice-contact sediment exposed in kame near North Adams, Massachusetts.

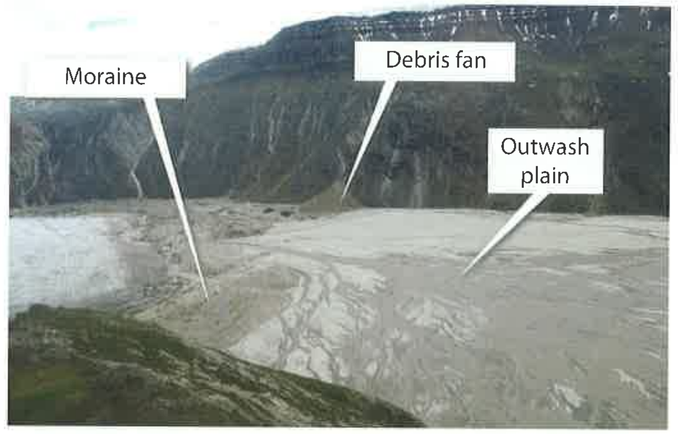
Eskers can disregard surface topography, which makes sense when one considers subglacial water flow does not respond to bed topography but rather hydraulic head gradients in a now-vanished ice sheet.

Glacially Related Sediments and Landforms

Glacially related sedimentation and landforms are not restricted to areas under or near the ice; the influence of glaciers, particularly continental ice sheets, can extend many kilometers from their margins. Glacial sediments are carried from the ice margin by water, wind, and floating ice.

Flowing meltwater transports sediment away from glacial margins generating **outwash plains.** These gently sloping surfaces, underlain by sand and gravel, head at former ice margins [**Photograph 9.20**]. Outwash-laden streams may occupy preexisting drainages, in which case they will deposit fill terraces that will usually be incised after glacial retreat when the sediment load declines. Outwash plains are often pitted with kettles, closed depressions reflecting the presence of now-melted ice blocks. Much of Cape Cod, in eastern Massachusetts, is a pitted outwash plain extending away from a moraine and indicating the former margin of the Laurentide Ice Sheet.

Glacially dammed lakes are common in hilly and mountainous terrain where tongues of ice disrupt drainages and impound both meltwater issuing from the glacier and rainwater. The water, unable to leave the drainage basin by the normal route, backs up until it is high enough to flow though a higher **spillway** (gap) and exit from the watershed. For example, a tongue of the Laurentide Ice Sheet blocked the northern Champlain Valley in Vermont, forcing water that would have flowed north into the St. Lawrence Seaway south into the Hudson River drainage. Some glacial lakes drain catastrophically over or under the ice itself. For example, the Channeled Scablands, deep and intricate drainage ways cut through loess and basalt



P. Bierman

PHOTOGRAPH 9.20 Outwash Plain. This outwash plain is fed by a meltwater stream (flowing from left to right) from the Greenland Ice Sheet margin east of Upernavik. Gray glacial moraines border the white ice on the left side of the image. A debris fan (paraglacial sedimentation) has formed below the cliffs, which expose flat-lying basalt over deformed gneiss. Glacially streamlined rock is visible in the foreground. Valley is about 2 km wide.

M. Kelly

PHOTOGRAPH 9.21 Ice-Contact Delta. An ancient ice-contact delta (at the right end of the lake) now stands isolated in the Holger Danskses Briller Valley, Greenland. During the last glaciation, ice filled the valley (today occupied by the lake) below the cliffs on the left of the image. The field of view in the foreground is several kilometers wide.

in eastern Washington State, are the result of failure of the ice dam that impounded glacial Lake Missoula (see Digging Deeper, Chapter 13). Evidence for similarly massive glacial floods has been found in Siberia, Tibet, the U.S. Midwest, Alaska, and in the English Channel.

Glacial lakes are also held behind dams formed by terminal and recessional moraines deposited during glacier retreat. In the Connecticut River Basin, 300-km-long glacial Lake Hitchcock was impounded by a dam of glacially derived sediment in southern Connecticut. The lake lasted for more than 5000 years. Water flowing into this and other ice-marginal lakes deposited deltas graded to paleolake levels. With the lakes long gone, these glacial-age deltas and their well-sorted sand and gravel are left high and dry above the present-day landscape [**Photograph 9.21**]. By measuring the elevation of abandoned delta topset/foreset contacts, and by mapping the distribution of fine-grained sediment deposited in the deeper sections of these glacially dammed lakes, one can reconstruct the extent and elevation of now-vanished water bodies. Glacial lake deltas are mined intensively for the gravel they contain. Such deposits of readily accessible sand and gravel are economically important in formerly glaciated regions.

Distinctively layered fine-grained deposits, indicative of quiet water with abundant sediment, are commonly laid down in glacier-fed and glacier-dammed lakes. Such deposits commonly display rhythmically bedded couplets of finer and coarser layers, termed **rhythmites** (Photograph 2.1). These couplets may represent annual layers known as **varves**, an assumption that can, in some places, be proven by other dating methods. During varve deposition, fine material settles out when the lake is iced over during the winter and sediment loads are diminished. The coarse material is deposited by currents during the summer when sediment loads are higher and fine material is kept in suspension by turbulence and higher-energy flows. Before the advent of radiocarbon dating, varved glacial sediments were meticulously counted, their thicknesses measured, and the results correlated between outcrops (much like tree rings, Chapter 2) in order to build composite sections and thus estimate the timing of glacial lake formation and the rate of ice-sheet retreat. Rhythmite and varve sequences deposited in proglacial lakes provide important paleoclimatic records. The fine sediment deposited at the bottom of such glacial lakes provided raw material for valley-bottom brickyards.

Glacial lake sediment commonly overlies till, indicating that the retreating ice margin directly bordered a lake (Photograph 2.1). In that case, sediment issuing from tunnels in the ice would be deposited below the water as a subaqueous fan. Indeed, some of the most productive aquifers in New England are in deep mountain valleys where glacial gravels are capped by glacial lake silt and clay. **Dropstones,** anomalously large rocks found isolated in the deposits of fine-grained glacial lake sediment, provide direct evidence for the delivery of sediment-laden, floating ice to water bordering ice margins. In ocean sediments, such dropstones are referred to as **ice-rafted debris** (IRD) and are used as an indication that glaciers were discharging icebergs directly to the ocean. The presence of isolated IRD deposits in marine sediments has been used to determine that limited glaciation began in the Northern Hemisphere almost 40 million years ago. Much more extensive deposits of IRD indicate that continental ice sheets, including the Laurentide, began forming and discharging icebergs to the ocean about 2.6 million years ago [**Photograph 9.22**].

International Ocean Drilling Program

PHOTOGRAPH 9.22 Dropstones. Dropstones (ice-rafted debris) in the core of sandy marine sediment collected by drilling off the coast of southern Greenland. The scale on the left side is in centimeters.

Glacial Landscapes, Landforms, and Deposits

Glaciers leave behind a characteristic, and in many cases diagnostic, set of depositional and erosional landforms, some of which are unique to glacial settings. Glacial landforms were first recognized in alpine settings and later this knowledge was applied to understanding the landforms of continental ice sheets.

Glacial landforms and sediments are formed and deposited under ice, at the ice margin, and distal from the ice [**Figure 9.10**]. It is important to note that most glacial sediment is deposited in the ablation zone where the ice is melting and releasing material entrained up-glacier. Landform complexes in alpine and continental glacial settings differ in type and scale; thus, we discuss them separately even though there are clear parallels in ice mechanics and process that cause alpine and continental ice to leave behind many similar landforms.

Landforms of Alpine Glaciers

Three closely related erosional landforms—the **cirque, arête,** and **tarn**—typify valleys in alpine glacial landscapes

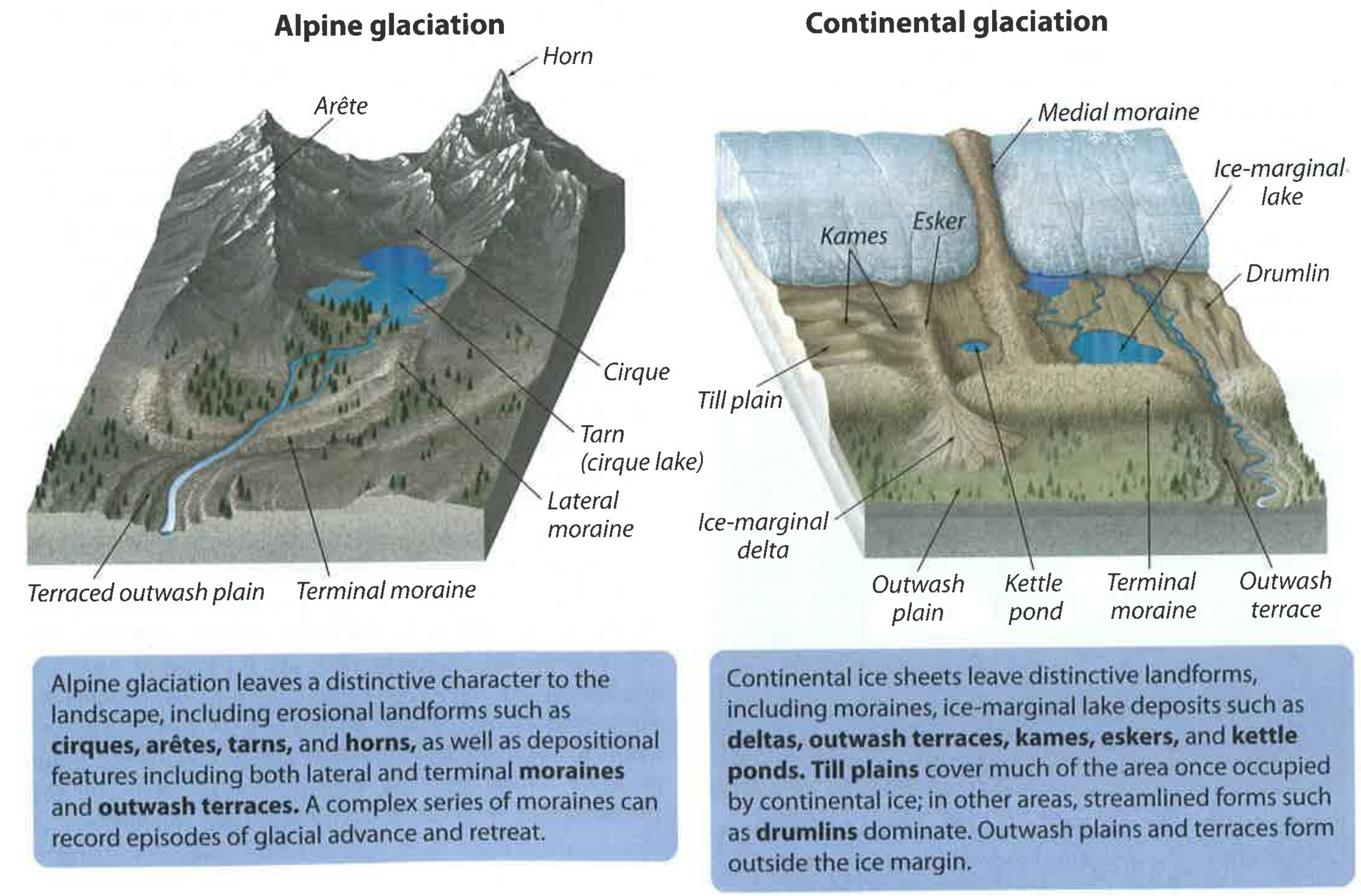
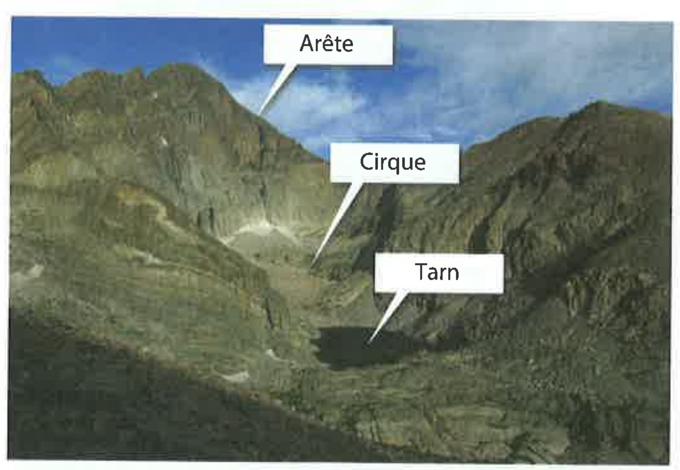


FIGURE 9.10 Alpine and Continental Glacial Landforms. Characteristic sets of landforms occur at and near retreating continental and alpine ice margins depending on regional topography and the abundance of sediment.

[Photograph 9.23]. Cirques are half bowl-shaped forms cut into mountainsides and backed by a steep **headwall,** the near-vertical face of rock defining the upper limit of the cirque. The processes creating the cirque forms we see today are complex. Basal erosion by ice-driven or meltwater-driven processes must keep up with (and originally exceed) the cirque retreat rate due to freeze thaw and mass wasting. Cirques are deepened by the flow of alpine glaciers but freeze-thaw loosening of blocks on the headwall and subsequent rockfall are also important erosion processes. Most cirques are multigenerational, having been occupied and modified by ice during many glaciations and weathered during the intervening interglacial periods. Many cirques, deepened into closed basins by flowing ice, now hold small lakes, known as tarns.

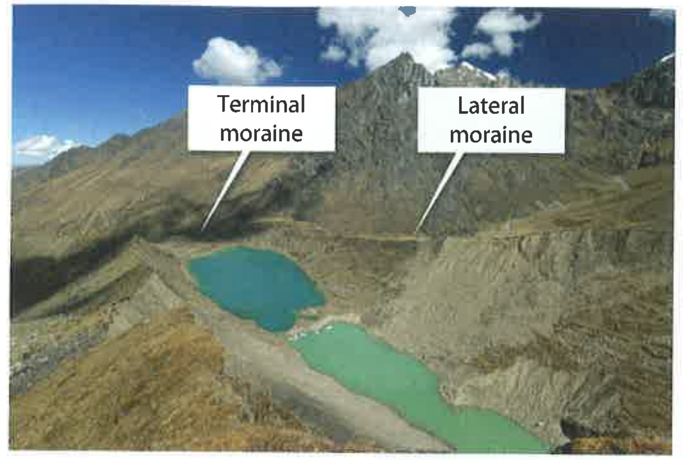
Arêtes are knife-edge ridges formed as two cirques or parallel U-shaped valleys erode toward one another, leaving only a steep-sided, narrow ridge of rock between. If headward erosion of cirques continues, the upland surrounded by headward-retreating cirques with intervening arêtes may be eroded into an isolated peak of rock known as a **horn.** Alpine valleys with roches moutonnées, rock steps, and irregular profiles, some partially filled with sediments, lie between cirques and the lower reaches of alpine valleys that are lined with lateral moraines.

Terminal moraines or **end moraines** are piles of debris that accumulate at the distal end of glaciers as sediment melts out in the ablation zone and as basal ice deforms and in some cases is thrust up from the bed. The size of a terminal moraine depends on the length of time the terminus remained at one location, the amount of debris in the glacier, and the flux of debris-bearing ice to the glacial margin. Terminal moraines are breached by outwash streams that carry away meltwater from the glacier. **Lateral moraines** form alongside the ablation zones of alpine glaciers as debris both rolls off and melts out of the ablating ice. **Medial moraines** are concentrations of debris where two tributary glaciers merge into one larger glacier (Photograph 9.11).



M. Miller

PHOTOGRAPH 9.23 Common Alpine Glacial Landforms. Cirque, tarn, and arête along the cirque headwall, Longs Peak, Rocky Mountain National Park, Colorado. Glacial features here are cut into metamorphic bedrock and granitic intrusions. Field of view is several kilometers wide.



J. Licciardi

PHOTOGRAPH 9.24 Glacial Moraines. Little Ice Age moraine complex deposited during the last 1000 years in Vilcabamba, Peru, damming small lakes. The moraine may still be ice-cored. Ice flowed from right to left.

Medial moraines can be preserved as low-standing linear ridges after the ice melts away, although often, they are not preserved at all.

Alpine moraines are built largely of till delivered to the glacier margin, both in basal ice and supraglacially [**Photograph 9.24**]. The presence of a moraine indicates stability of the ice margin for a period of time sufficient to deliver a morphologically significant pile of debris—rapidly melting ice does not leave well-defined moraines but rather a featureless **till plain.**

Geomorphologists use the presence and location of moraines to understand the extent of prior glaciations and thus to infer past changes in climate (both precipitation and temperature) over time. The outermost moraine reflects conditions most favorable to the advance of ice, high ice accumulation rates and/or cold temperatures. Progressively younger moraines, usually found inside the outermost moraine, provide evidence for later climates less favorable for glaciers. Rarely, alpine moraines may crosscut one another; in that case, crosscutting relationships can be used to determine the relative ages of the moraines. The alpine moraine record is incompletely preserved because younger advances of greater extent than older advances destroy older moraines and thus the geomorphic evidence of those previously expanded glaciers.

Various techniques, including relative weathering, soil development, carbon-14 dating of organic material, and cosmogenic nuclides are used to estimate the age of moraines. Because many alpine moraines are narrow, steep-sided features deposited at least in part on ice, they can erode rapidly, losing mass from their crests and confounding accurate dating by techniques, like cosmogenic nuclide measurement in boulders, that date surface exposure.

Landforms of Ice Sheets

Ice sheets create many of the same landforms as alpine glaciers, such as moraines and till plains, but they also

M. Groll

PHOTOGRAPH 9.25 Drumlin. Drumlin covered by a housing development in Friedrichshafen Raderach, Germany. The steep, up-ice side is upper right and the less-steep tail is lower left. Arrow shows ice flow direction.

create forms that are distinctive and not found in alpine settings. Mapping these landforms delineates the extent of now-vanished ice sheets and provides clues about their basal conditions.

Drumlins are streamlined, distinctively glacial landforms. These elongate hills are found in many low-relief areas formerly overridden by continental ice sheets [**Photograph 9.25**]. Famous drumlin fields include those in Wisconsin, upstate New York, and Ireland. Some drumlins are rock-cored, whereas others are made entirely of sediment including both unsorted till (common) and water-washed, well-sorted sand and gravel (less common). Drumlins range from hundreds of meters to as much as several kilometers in length, and are usually hundreds of meters in width and tens of meters in height. Drumlins are oriented in the direction of ice flow and are usually found clustered in drumlin fields where hundreds to thousands of individual drumlins may be preserved. They can be steep on the up-ice side and taper more gently in the down-ice side. The preferred orientation with glacial flow suggests that drumlins are sculpted as the deforming bed of the glacier interacts with materials below it—both rock and preexisting sediments.

When ice sheets cover large continental areas, they change the arrangement of drainages, both by blocking river outlets and by changing the slope of the land via transient isostatic changes in land-surface elevation. The topography of the ice sheet affects head gradients within the ice and therefore the flow direction of meltwater. Ice sheet control of surface drainage leads to a variety of landforms related to surface water hydrology. These include marginal lakes dammed by glacial ice and river channels cut into rock by ice-marginal drainages pinned against valley walls by ice. Such channels are often located above present-day valley bottoms and thus contain little or no water today. Streams in these large channels are far smaller than one would expect given the size of the valley that they occupy and are termed **underfit.**

In some areas, ice sheets and ice fields were not thick enough to cover the landscape completely. Mountain peaks and high plateaus standing above the ice as islands are known as **nunataks.** Ice-free nunataks provided **refugia,** places where plants and animals continued to live during glacial times. When glaciation waned, these refugia provided a source of seeds and animals to repopulate and revegetate the adjacent landscape.

Ice sheets, ice caps, and ice fields left moraines indicative of the their former extent; for example, Long Island and parts of Cape Cod are moraines of the Laurentide Ice Sheet. Such moraines are composed of a mixture of till and outwash sediment and tend to be broader and less steep than those of alpine glaciers. Similar to alpine glacial moraines, older moraines are often overrun and destroyed by later ice advances. Early workers, who studied glacial deposits on land, did not recognize that older moraines and till had been removed and thus believed that there were only four major terrestrial glaciations in the Pleistocene—based on the relative ages of continental ice-sheet deposits. Analysis of deep-sea sediment cores has shown convincingly that there were several dozen major glaciations—the terrestrial deposits of which have largely been erased by a few of the most extensive ice advances.

Geomorphic Effects of Glaciation and Paraglacial Processes

During each glaciation, global climate cooled, ice sheets expanded, and sea level fell as water once in the oceans was taken up and stored in the ice on land. The expansion of ice sheets and the general global cooling altered the operation of the climate and ocean system throughout the whole planet, not just on and around the glaciated regions. The rate of geomorphic processes and the distribution of landforms around the world responded to glacial/interglacial changes in temperature, precipitation, and the resulting distribution of plants and animals.

The legacy of glaciers on our planet is far reaching. For example, alpine glacial erosion likely controls not only the elevation of mountain ranges but, at least in part, the rate of rock uplift. In heavily glaciated mountain ranges, where erosion rates are high, isostatic compensation (Chapters 1 and 12) brings rock toward the surface, partially making up for mass loss by erosion. In glaciated regions, such as large portions of the Northern Hemisphere, pre-Quaternary regolith and deeply weathered soil profiles have been eroded and the landscape is a matrix of bare rock outcrops and till-covered slopes. Large moraine complexes, the distribution of glacial lake sediments, and the orientation of landforms shaped by flowing ice all indicate the large-scale geomorphic impact of continental-scale glaciation.

Changes in the mass of glaciers over time not only alter the land level under and near ice sheets by loading and then unloading the crust (**glacial isostasy**), but they change sea level by storing water on land as ice. When ice sheets expand, sea level drops as much as 130 m and thus so does the base level to which rivers flow around the planet. Even river systems and hillslopes far away from the ice sheets respond geomorphically to this change in a fundamental boundary condition. Falling sea level provides the impetus for river incision and the transport of eroded sediment away from the continents. For example, during times of lower sea level, Chesapeake Bay was a river valley and sediment from the Susquehanna River was deposited far away from the continents. The coming and going of glacial ice thus drives geomorphology and surface processes on a global scale.

The lasting geomorphic impact of glaciers on the postglacial landscape controls the distribution and rate of surface processes. Such **paraglacial** effects (consider them as glacial hangovers) result from direct landscape conditioning by glaciation and deglaciation. For example, recently deglaciated landscapes are typically unstable because they lack vegetation, have glacially oversteepened slopes, and are covered by large amounts of unconsolidated sediment deposited as the ice melted. Immediately after deglaciation, erosion and sediment transport rates are high and then fall off exponentially over many thousands of years. It appears that most landscapes take thousands to tens of thousands of years to readjust to interglacial conditions (longer than many interglacial episodes). Typical paraglacial features of formerly glaciated areas include sediment storage in rockfall talus, debris and alluvial fans, and valley-bottom terraces of outwash. Not only do these features contribute sediment during interglacials, but they also serve as major and easily reworked sources of sediment for incorporation during the next glacial advance. Thus, the long-lasting, landscape-scale memory of past glaciations influences postglacial landscape dynamics in most previously glaciated terrain.

Periglacial Environments and Landforms

About a quarter of Earth's terrestrial surface can be classified as **periglacial**—environments that are not permanently ice-covered but experience below freezing temperatures for much of the year and in which geomorphic processes are largely driven by the seasonal freezing and thawing of water.

Of particular importance in periglacial environments is the downslope movement of water-saturated regolith in thin sheets during the summer, when the upper layers of frozen ground thaw, lose much of their strength, and move downhill over the still-frozen substrate. Most deformation is concentrated at the boundary between frozen and thawed material. This process is called **solifluction**, a term first associated with any downslope movement of saturated regolith but now applied primarily to periglacial processes. Closely related is **gelifluction**, a type of solifluction that describes the downslope movement of regolith by the freezing process en masse, rather than movement along a specific plane in the subsurface (see Figure 5.3, second panel).

Many periglacial landscapes have never been glaciated. Relict permafrost features, left over from colder climates, have been found in places far from the most recent

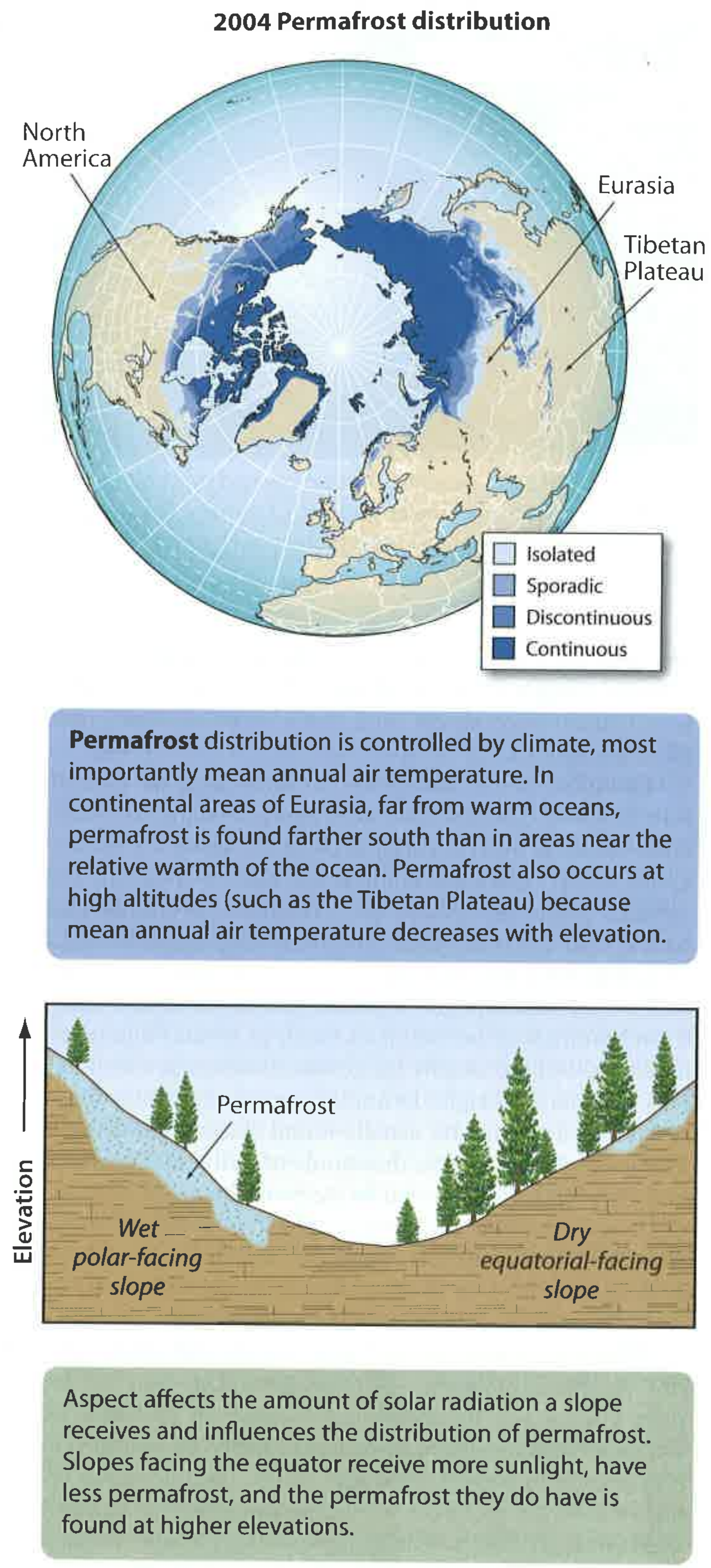


FIGURE 9.11 Permafrost Distribution. Permafrost occurrence is controlled by the combination of low temperature and low precipitation. Thus, permafrost occurs both at high latitudes and high altitudes.

Pleistocene glaciation including the highlands of South Africa and the southern Appalachian Mountains. As our global climate warms, both natural resource and environmental interests will increasingly focus on periglacial processes and landforms because thawing will disturb now-stable frozen soils and landscapes.

Permafrost

Integral to much of the periglacial environment is **permafrost,** ground that has remained frozen for at least two years and in many cases much longer. Because permafrost is the product of seasonally cold temperatures, usually mean annual air temperatures below −2°C, it is found at both high elevation (due to cooling of the atmosphere with elevation) and high latitude (the result of reduced annual solar input). For example, elevation-related permafrost is found at mid- and low latitudes on the summits of Colorado's Rocky Mountains at 3800 m above sea level, on the peak of tropical Mauna Kea in Hawaii at 4200 m above sea level, and even at the top of Tanzania's equatorial Mount Kilimanjaro (almost 6000 m above sea level). Latitude-related permafrost begins at about 50° N latitude in North America and continues northward. However, in Asia, the Tibetan Plateau is high enough that permafrost there is found below 30° N latitude [**Figure 9.11**]. In the Southern Hemisphere, there is scant permafrost outside of Antarctica because there is little landmass in the appropriate latitude range.

The upper decimeters to meters of most permafrost thaw every summer due to increased air temperatures and net solar radiation delivery to the ground surface [**Figure 9.12**]. The seasonally thawed thickness above permafrost is referred to as the **active layer.** It is often water-saturated as the frozen ground and ice lenses below are less permeable and act as an aquitard. The **permafrost table** defines the bottom of the active layer and indicates the uppermost extent of ground that does not thaw during the summer. **Taliks** (from the Russian tayat, meaning "melt") are areas that remain unfrozen—for example, the thawed region below a lake. Permafrost usually contains **ground ice,** both dispersed as pore fillings and concentrated in lenses. Ground ice is most common in the upper tens of meters of permafrost.

Microclimates and the annual distribution of precipitation control the establishment and survival of permafrost. Permafrost is most common where winters are dry and cold (resulting in thin snowpacks and more effective radiational cooling of the soil) and where summer temperatures are not much above freezing. Thick snowpacks reduce conductive

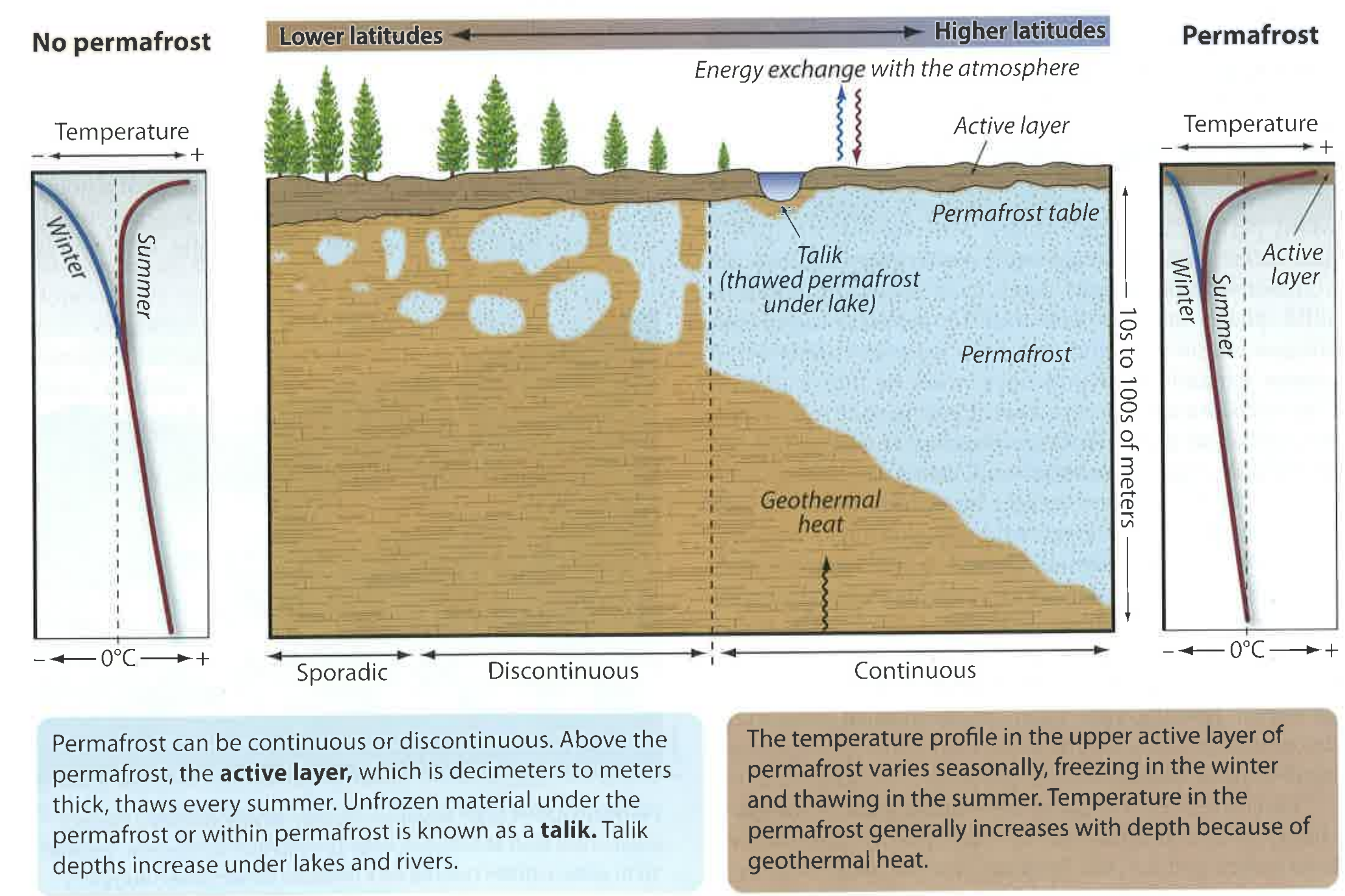


FIGURE 9.12 Permafrost Cross-Section and Temperature Profiles. Longitudinal cross-section through the arctic and subsurface temperature profiles showing the distribution of permafrost and seasonal change in temperature profile.

heat loss from the soil, and thereby prevent sufficient chilling to establish or maintain permafrost.

Permafrost is described by its continuity and thickness (some Siberian permafrost is over 1500 m thick, but the world average is closer to 100 m). **Continuous permafrost** implies that all of the ground is frozen, except for the seasonally thawed, active surface layer; any warmer microenvironments do not change the energy balance enough to keep parts of the surface thawed. **Discontinuous permafrost** is indicative of a landscape where some areas remain perennially thawed. **Sporadic permafrost** indicates that unfrozen ground is the rule—not the exception. In general, permafrost becomes more continuous at higher latitudes where mean annual air temperatures are lower.

Permafrost thickness is controlled by the loss of heat to the overlying atmosphere and the gain of heat geothermally from below. Although permafrost thickness is correlated with current mean annual air temperature, it is likely controlled by long-term average climate. Very thick permafrost is long-lived and implies long periods of cold temperatures over multiple glacial-interglacial cycles. It is likely that some permafrost, especially sporadic permafrost at lower latitudes, may be inherited from colder climates of the past. The presence of relict permafrost landforms, in areas such as midwestern North America and the southern British Isles, indicates that at some time in past, the climate was much colder with mean annual air temperature well below 0 °C (see Chapter 13).

Characteristic Periglacial Landforms and Processes

Many periglacial landforms and processes are distinctive and differ from their thawed counterparts because the volumetric expansion of water as it goes from liquid to solid phase has the potential to generate significant stresses within rock and soil if the water is confined and cannot expand either into open pores or into fractures. One such situation is top-down freezing that seals water in cracks and pores. Another situation is **segregation ice growth,** during which water and water vapor migration to growing ice lenses increases internal stresses in earth materials. Rapid freezing, cold temperatures, and frequent freeze-thaw cycles all appear conducive to shattering rock. Such **frost shattering** is common in periglacial environments (both arctic and alpine) and often leaves bedrock outcrops hidden under a mantle of angular, shattered rock fragments termed **felsenmeer** (German for sea of rocks) [Photograph 9.26]. Movement of subsurface water is also essential for frost heave (the predominant movement of material by ice growth).

Extensive areas of felsenmeer outside present-day periglacial limits reflect harsher climates during glacial times. Felsenmeer can be an integral part of many different permafrost landforms, including rubble streams oriented downslope, large low-gradient boulder fields, and **cryoplanation terraces** (flat surfaces cut into bedrock hillslopes).

B. Kimball/NH Natural Heritage Bureau

PHOTOGRAPH 9.26 Felsenmeer. Felsenmeer surface of broken rock and sheeted tors on Nelson Crag, northeast shoulder of Mount Washington in the White Mountains of New Hampshire.

These terraces have abrupt backing edges, and the terraces themselves are covered with broken rock. It is not possible to observe the processes that cut these terraces because they appear to form slowly over millennia; however, those studying cryoplanation terraces believe that they are cut by scarp retreat, the result of decades to centuries of **nivation,** the accumulation of snow in erosional depressions termed **nivation hollows** [Photograph 9.27]. In these hollows, both physical and chemical weathering increase because of repeated freeze-thaw cycles and the presence of moisture. Nivation hollows deepen as weathered rock is removed from the slope by mass movement processes.

If the upper, seasonally unfrozen, active layer of regolith and soil becomes saturated with water, mass movements will move material downslope. For example, solifluction,

K. Crozier

PHOTOGRAPH 9.27 Nivation Hollow. Nivation hollow (snowy area) in the Bald Mountains, near Jasper, Alberta, Canada, is about 10 m wide. Lighter colored rock (without lichen cover) suggests that a larger area around snow patch was snow-covered earlier in the summer or during previous summer seasons. Photo was taken in early August with snow cover at a minimum.

M. Miller

PHOTOGRAPH 9.28 Solifluction. Solifluction, shown in the Tien Shan Mountains, Kyrgyzstan, moves regolith downslope in lobes with downslope edges about a meter high.

the downslope movement of thawed soil and sediment in coherent masses, becomes most active in late spring and summer as the active layer deepens; the soil saturates from snowmelt, rainfall, and melting interstitial ice; the shear strength drops; and the soil weakens. The saturated mass of soil behaves like a viscous fluid moving slowly downslope and forming distinctive, steep-sided, lobate features [Photograph 9.28].

Solifluction is most common on slopes between 5 and 20 degrees and was critical in the development of permafrost-rich, valley-bottom muck deposits that contain Pleistocene megafaunal remains (extinct animals such as mammoths and giant sloths) in Alaska and Siberia. Many steeper slopes (>20°) are too well drained and thus the soils are too strong to flow, whereas the driving force on lower slopes (<5°) is insufficient to overcome the strength of even saturated soil masses. Shallow landslides are also common during the summer months in periglacial terrain. Many of these slides have slip surfaces at the base of the active layer, reflecting both a strength and hydrologic contrast with the underlying frozen material.

Karst-like topography can develop in landscapes where melting of permafrost ice masses produces solution-like features including closed depressions, subsurface streams, and blind valleys. Such **thermokarst** results when ground-ice melts and surficial materials collapse as supporting ice is removed. Thermokarst often initiates through disturbance of the permafrost surface. Some of these disturbances are natural, such as vegetation change, river-channel migration, or mass movements; others are human-induced, such as road building or off-road vehicle use. Disturbance changes the surface energy balance and thus can initiate thawing. If thawing then changes albedo or surface hydrology, melting can become a positive feedback process and the thermokarst can expand rapidly.

The most distinctive and common thermokarst features are thaw lakes [Photograph 9.29a] and thaw slumps [Photograph 9.29b]. **Thaw lakes** occupy shallow depressions that may have once been underlain by higher-than-average

(a)

NASA

(b)

W. B. Bowden

PHOTOGRAPH 9.29 Thermokarst Features. (a) Thermokarst thaw lakes on Alaska's North Slope. Lakes are 1 to 10 km long. (b) Thaw slump located on the Alaskan North Slope in the foothills of the Brooks Range.

concentrations of ground ice. Once thawing begins, the lower albedo of water in comparison to ice allows it to absorb more solar energy. Warm water thermally erodes the shorelines while waves generated on the lake by wind cause physical erosion. Over time, thaw lakes fill with organic and inorganic sediment. **Thaw slumps** are mass movements (Chapter 5) initiated by the melting of ground ice and are often the result of bank undercutting by streams and rivers. Weak, often cohesionless, material mobilized by thawing rapidly flows from thaw slumps, allowing retrogressive failures to continue. Thaw slump scarps can move rapidly back into the landscape and supply large amounts of sediment to river channels.

Patterned ground comes in many different forms (sorted circles, polygons, and stripes) and is common in periglacial terrain [Photograph 9.30]. Some patterned ground (circles and polygons) has coarse margins and finer-grained centers; other patterned ground is unsorted.

(a)

B. Hallet

(b)

B. Hallet

(c)

B. Hallet

PHOTOGRAPH 9.30 Periglacial Patterned Ground. (a) Sorted stripes near the summit of Mauna Kea, Hawaii, at an elevation of about 4300 m. The stripes are about 15 cm across. (b) Sorted circles, Spitsbergen, Norway. The sorted circles are about 2 m in diameter. This area is an uplifted gravel beach; freeze-thaw cycling in permafrost sorted these circles. (c) Polygons, in the Dry Valleys of Antarctica. These are large features as shown by the shadow of the helicopter for scale.

There are high-centered and low-centered polygons; high-centered polygons have water occupying their margins in the summer, whereas observations suggest that low-centered polygons are actively growing. Stone stripes running down hillslopes are defined by contrasting grain sizes.

A variety of processes have been suggested to explain the development of these features, in particular their sorting. Earlier hypotheses emphasized thrusting and heaving of particles by frost action and the ability of fine grains to fill voids more readily than large grains. Recent explanations center on slow convection in the active layer and preferential movement (and thus sorting) of different grain sizes [Figure 9.13]. Polygon formation is attributed to ice filling ground cracks resulting in ice wedges. When the wedges melt, the resulting voids are often filled by sand, leaving ice wedge casts as indicators of periglacial conditions in the past.

Rock glaciers, present in polar and alpine areas, share many similarities with alpine glaciers and resemble them in form. Unlike alpine glaciers, rock glaciers are completely

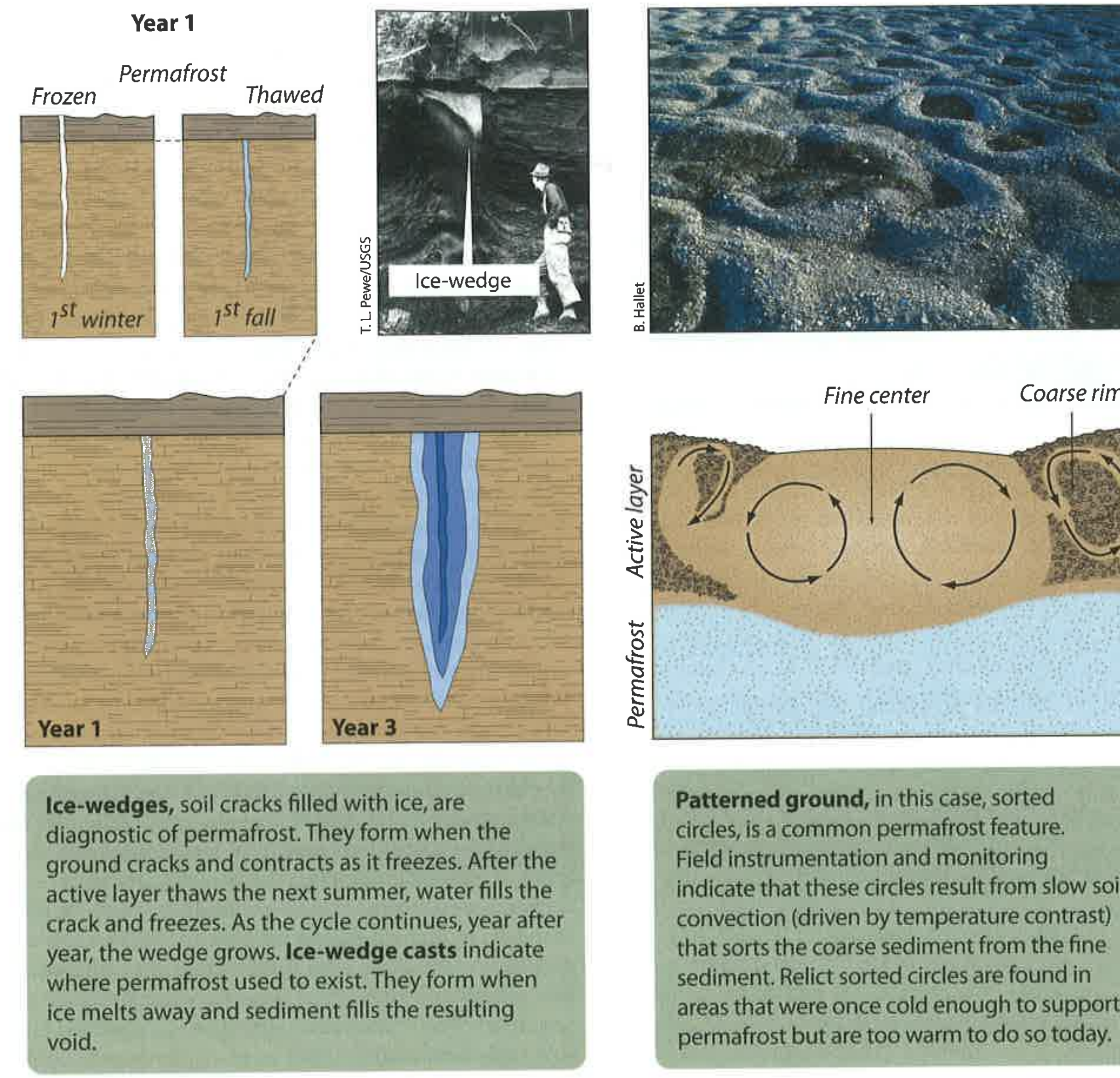


FIGURE 9.13 Patterned Ground Features. Cross sections showing processes thought to be active in the formation of sorted circles and ice wedges, common forms of patterned ground.

covered with rubble or regolith below their accumulation area and they move at slow rates of centimeters to a few meters per year. Both rock glaciers and alpine glaciers issue from cirques or cliffs, are elongate in plan form, have lobate margins with steep snouts, and move downslope deforming under the force of gravity. Rock glaciers contain ice, which facilitates their flow. This ice can be massive and layered (as has been found in the few rock glaciers cored so far) or it can be interstitial, filling the voids between rubble. The rock cover provides significant insulation, tempering thermal changes experienced by the ice below. Rock glacier genesis is controversial with some arguing that interstitial ice forms in place from percolation of surface water and others suggesting that rock glaciers are simply debris-covered glaciers with ice originating from snowfall in the accumulation zone. It is important to realize that while rock glaciers are a distinct landform, there may be more than one way in which they form (the problem of equifinality, in which the landform is not diagnostic of the process by which it formed).

Pingos (from Inuit, meaning "small hill") are one of the more unusual and distinctive periglacial landforms [Photograph 9.31]. Dome-shaped landforms tens to hundreds of meters wide and meters to tens of meters high, pingos are cored with massive ice that may be exposed at their tops. The surface layer of many pingos is extended (in tension) from the growing ice body below. The pingo-forming process illustrates well the distinctive nature of permafrost hydrology. Closed-system pingos (where water does not migrate in from the outside) are typically found in what used to be lake basins or boggy areas. After the lake drains, the surface freezes and the permafrost table

Paolo Koch / Science Source

PHOTOGRAPH 9.31 Pingo. Pingo covered by green vegetation and surrounded by river and thermokarst lakes in the Mackenzie River Delta, Canada.

rises, trapping pore water from below and forcing it to expand upward upon freezing. Open-system pingos result from artesian conditions that supply groundwater to the growing ice mass.

Rivers and soils in periglacial regions behave somewhat differently than their temperate-region cousins. Flows in periglacial rivers are highly seasonal—open water may exist for only a few weeks to a few months each year. Extreme runoff events, generated by snow melt, commonly recur every spring and create ice jams that force overbank flow. In addition to physical erosion of riverbanks, flowing water can thermally erode permafrost with which it comes in contact. Permafrost soils are typically very disturbed, or **turbated**, as they are stirred by the melting and freezing of ice and possibly by convection. In places, this cryoturbation is evidenced by folds and discontinuous soil horizons.

Applications

Many of our planet's great cities, much of the population of northern Europe, Asia, and North America, and many of the world's most productive agricultural regions (the loess belts) are located in or just beyond areas once covered by ice sheets. In these regions, understanding the distribution and character of glacially deposited sediments is important for many reasons. The distribution of glacial sediments can control slope stability; glacial lake and glacial marine clays are notorious for landsliding. Oversteepened slopes, the result of glacially induced valley deepening, are prime terrain for rockslides and rockfalls. Aquifers hosted in glacial outwash gravels provide reliable but easily polluted water supplies. Many till-covered uplands in New England, especially in areas where the till is clay- and silt-rich and highly compacted, have low permeability and poorly drained, hydric soils, making septic systems difficult to site. Glacially deposited sand and gravel are important resources for construction.

Mapping the distribution of glacial sediments can be important for solving significant environmental problems. In previously glaciated regions, the surface and subsurface distribution of earth materials, both permeable sand and gravel and less permeable silt and clay, are in large part determined by the geometry of now-vanished glaciers and their associated outwash plains and ice-marginal lakes. Understanding the resulting distribution of glacial sediments can be critical for determining the fate and transport of hazardous materials and the provision of clean drinking water. For example, contaminants released in a valley underlain by a thick sequence of impermeable glacial lake sediments are unlikely to affect an artesian drinking-water aquifer hosted in deeper, outwash-derived sand and gravel.

Understanding the behavior and particularly the mass balance of glaciers over time and into the future is critical as our planet warms. In the past century, most alpine glaciers have retreated, some dramatically. The Greenland Ice Sheet appears to be losing mass and contributing to sea-level rise at a rate greater than anticipated even a decade ago. The most direct impact of melting glaciers is the rise in global sea level as water is transferred from land-based glaciers to the ocean. Indeed, predicting sea-level rise has become the dominant glaciological issue of our time, in part because of the fear that rising sea level could lead to catastrophic positive feedbacks. For example, as sea level rises, the rate at which floating ice margins calve increases. Calving of floating ice does not itself increase sea level, but calving, by reducing buttressing, can increase the velocity of ice streams and thus can increase the flux of inland, terrestrial ice to the ocean. There is also concern that rising sea level could float and destabilize the massive Antarctic ice shelves, allowing more terrestrially based ice there to flow into the sea.

The geologic record clearly indicates that large ice sheets have come and gone over time, and, as a result, sea level has gone up and down. During some of the warmest and longest interglaciations of the past 2 million years, sea level was at least several meters, and perhaps as much as 10 m, higher than it is today. It is sobering to consider that the Greenland and Antarctic ice sheets together hold enough ice that, if melted, sea level would rise about 80 m (7 m from Greenland, 73 m from Antarctica). Most of the world's major cities, and much of its population, are within 80 m of sea level.

Glaciers function as natural water reservoirs, particularly alpine glaciers, in areas as diverse as South America, the Himalaya, and western North America. These glaciers store winter or wet season precipitation as ice and snow and release it slowly as runoff over the summer/dry season months. As the climate warms, many retreating glaciers are forecast to disappear entirely. With them will go reliable summer water supplies in many arid areas because water stored as ice in glaciers provides a buffer against both seasonal and longer-term drought. The storage function that glaciers provide can be replaced by artificial reservoirs, but only at great monetary and environmental costs. Many communities around the world will be severely affected by the loss of glacier water storage, which will reduce summer drinking water supplies and hydroelectric power generation. In the developing world, where funds for investment in water storage infrastructure are scarce, the impact will likely be most severe.

Areas underlain by permafrost have proven to be extremely sensitive to human impacts. Locally, such impacts include excavation, road building, and the presence of structures such as houses, which change the energy balance of the landscape, leading to melting and damaging ground movements.

Globally, our planet's warming climate has already begun to affect periglacial environments. Recent observational data of significant polar warming (at least several °C over the past decades) support the predictions of climate models, which indicate that future warming will be most severe at high latitudes. Warming is already melting permafrost in discontinuous permafrost areas, causing problems for roads, airstrips, and buildings. As arctic warming intensifies, geoengineering problems will become more common and widespread. Not only will the foundations of many structures become unstable, but thermokarst-catalyzed erosion, and the disruption it causes, will likely increase in both extent and severity. The active layer, which is responsible for much periglacial geomorphic activity, will become thicker and stay active for longer each year. Solifluction will become more common in permafrost regions during the summer months. Melting permafrost will release methane, an effective greenhouse gas, causing further warming.

Although we are now in an interglacial period during which the climate is relatively warm and the distribution of ice is restricted, geomorphic and societal effects of past glaciations remain important. In the future, understanding the response of periglacial landscapes to global warming will become increasingly relevant.

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DIGGING DEEPER How Much and Where Do Glaciers Erode?

How much do glaciers erode, modify, and shape the land? This deceptively simple question continues to intrigue geomorphologists and has fueled arguments running back to the nineteenth-century roots of the discipline when glaciers were initially recognized as very effective agents for transporting material.

> *For the moving of large masses of rock, the most powerful engines without doubt which nature employs are the glaciers. . . .*
>
> —Playfair (1802, p. 388)

In the 1840s, when Agassiz subsequently developed his idea of ancient ice ages based on recognition of the formerly greater extent of European glaciers, the issue of glacial erosion became central to a century-long controversy over whether glaciers could erode valleys into rock or simply entrained and transported the loose, weathered material above rock.

Many early geologists could not accept that ice, a soft material, could erode into much harder bedrock. How could something with low enough density to float in water carve deep valleys into solid rock? Although some argued that permanent snow protected landscapes from erosion (Bonney, 1902; Fairchild, 1905), others such as the German geologist Penck argued that glaciers did not protect but vigorously attacked the rock over which they flowed (Penck, 1905). Penck was not alone. Davis, an American geologist, argued that glacial action sculpted alpine terrain based on how hanging valleys, rock basins, and cirques only occurred in glaciated and formerly glaciated mountains (Davis, 1900).

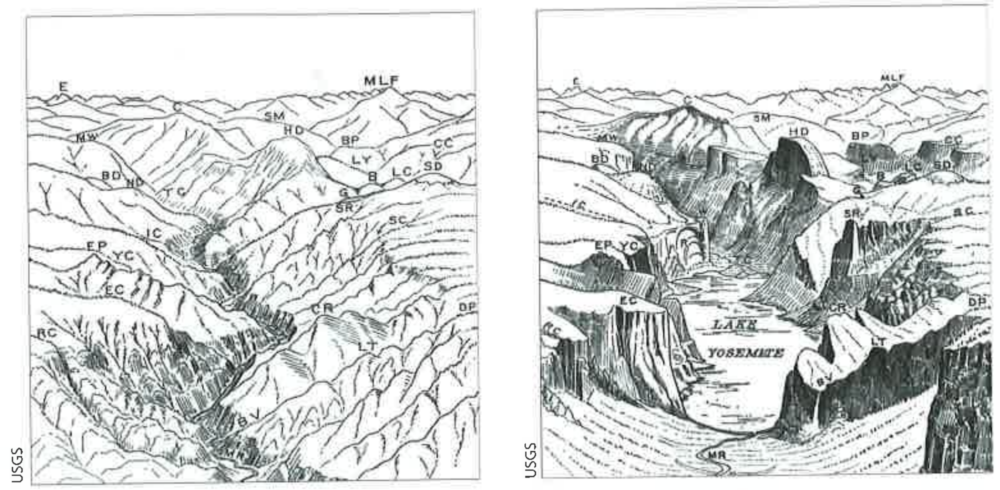

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FIGURE DD9.1 Yosemite Valley before and after glaciation, as depicted by Matthes in his conceptual illustration of the transformation of the incised preglacial valley of the Merced River into the glacially carved terrain of the modern Yosemite Valley, California. [From Matthes (1930).]

Although early arguments over glacial erosion centered on the European Alps, the spectacular topography of California's Yosemite Valley became a focus of such arguments when the first state geologist of California, Whitney, proposed that the steep-walled valley was a down-dropped, fault-bounded graben (Whitney, 1865). Decades later, U.S. Geological Survey physiographer Matthes compellingly argued that glaciers gouged out the previously incised valley of the Merced River to form Yosemite's modern topography (Matthes, 1930) [**Figure DD9.1**]. Numerous field studies went on to verify the association of U-shaped valleys with formerly glaciated landscapes and support the notion that glacial erosion transformed V-shaped river valleys into distinctive U-shaped cross-sectional profiles.

But how does this valley transformation happen? In the 1970s and 1980s, advances in understanding the mechanics of glacial erosion (Hallet, 1979 and 1981) and the development of ice-flow models that included basal sliding and realistic ice rheology paved the way for process-based modeling of glacial valley excavation. Building on these advances, Harbor et al. (1988) developed a numerical model of the development of U-shaped valleys by glacial erosion [**Figure DD9.2**]. The model assumes that glacial erosion is controlled by the basal sliding velocity of the ice, which they modeled as a function of the basal shear stress (Harbor et al., 1988; Harbor, 1992). Simulating the cross-section of glacial flow through an initially V-shaped valley, they found that the greatest erosion initially occurred on the lower valley walls until establishment of an equilibrium U-shaped profile that persisted as long as the valley continued to entrench and excavate a deeper trough [**Figure DD9.3**]. By calibrating the model to observations from modern glaciers, Harbor (1992) was able to show that the transformation from a V-shaped to U-shaped valley cross section could begin in a single glacial cycle.

But by how much did glaciers widen their valleys? It was difficult to tell in most cases because glaciers eroded evidence of the former, preglacial valley profile. Montgomery (2002) studied the unusual case of the Olympic Mountains, which are ringed by some glaciated and some unglaciated valleys. There, valley width, relief, and cross-sectional areas were similar for valleys with drainage areas <10 km^2, but glaciated valleys draining > 50 km^2 had 2 to 4 times the cross-sectional area and up to 500 m greater relief than unglaciated fluvial valleys [**Figure DD9.4**]. With increasing drainage basin area, glaciers excavate far more rock volume than do rivers flowing through fluvial valleys. Unlike river valleys, in which tributaries tend to seamlessly join larger channels at the same elevation, large glacial valleys erode far faster than small valleys, resulting in hanging valleys draining over steep valley walls in formerly glaciated terrain.

Excavation of deep glacial valleys is not spatially uniform, and depends on both the ice thickness and thermal regime of the glacier—whether warm-based or cold-based. Glaciers that are frozen to their bed are static at their base and no motion at the bed means little to no erosion. Glacial erosion is most aggressive where flow is fastest, which is facilitated by thick ice, a warm base where melting and meltwater promote basal sliding, and a steep ice-surface gradient to provide the driving stress. High debris content embedded in basal ice increases the rate of glacial abrasion. Alpine glaciers typically exhibit strong longitudinal patterns of glacial erosion in which the deepest glacial troughs are excavated near and below the glacial equilibrium

DIGGING DEEPER How Much and Where Do Glaciers Erode? *(continued)*

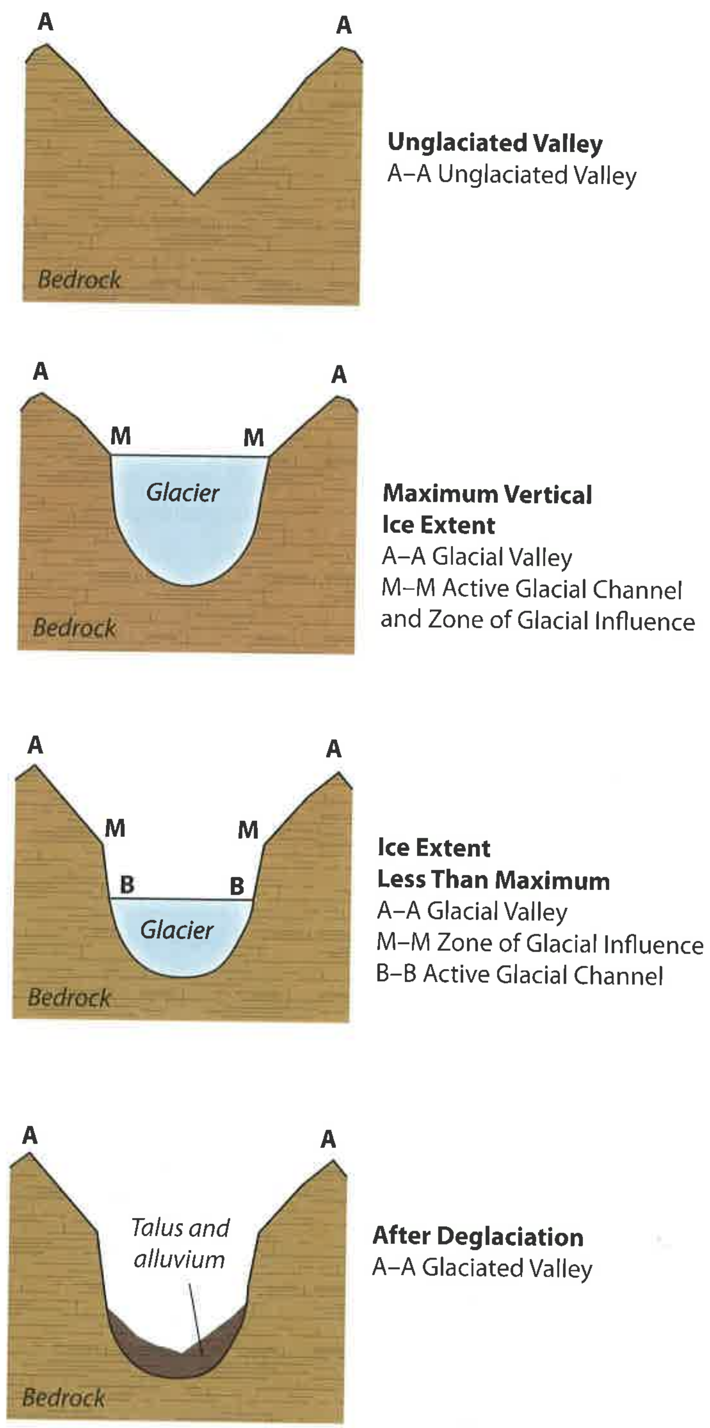

FIGURE DD9.2 Schematic illustration of the evolution of an unglaciated V-shaped valley into a U-shaped valley as a result of glacial erosion. [From Harbor (1992).]

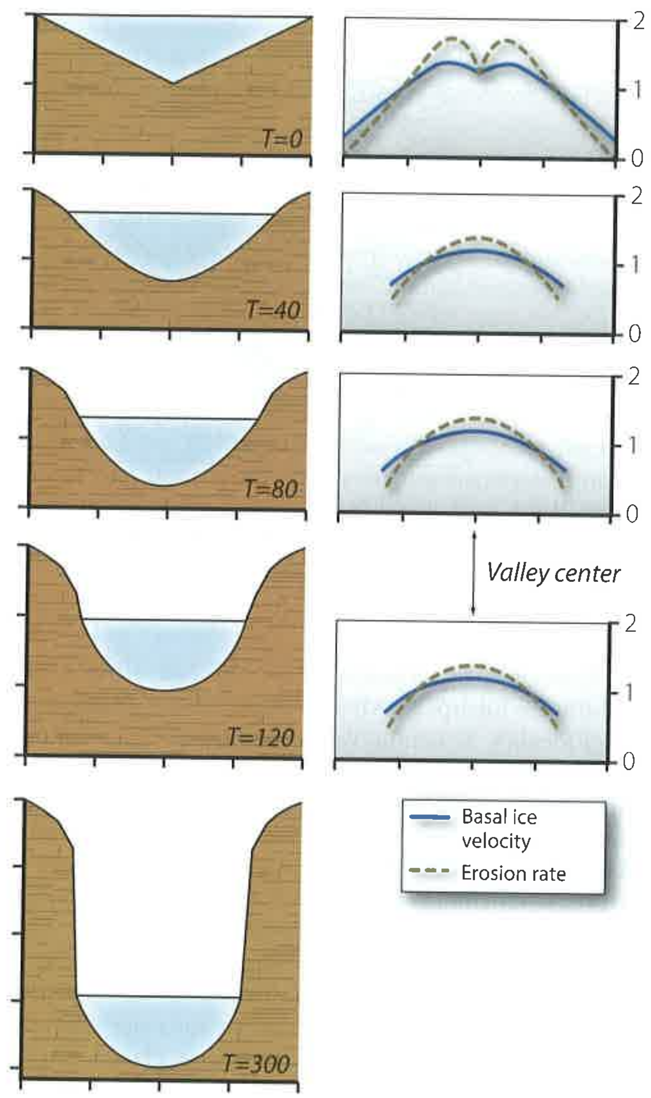

FIGURE DD9.3 Model results for simulation of valley development by glacial scour from an initially V-shaped valley cross section (T = 0) to a steady-state U-shaped cross section (T = 120). Plots are dimensionless and have no vertical exaggeration. Graphs to the right show the corresponding cross-valley profiles of basal ice velocity and erosion rate. [From Harbor (1992).]

line altitude (ELA). Ice sheets have uplands that are cold-based, and deep troughs on their margins where warm-based ice streams are much more effective in excavating terrain. The combination of concentrated erosion in deep ice streams and subglacial fluvial erosion also leads to deeply eroded valleys in the form of fjords and great lakes near ice-sheet margins.

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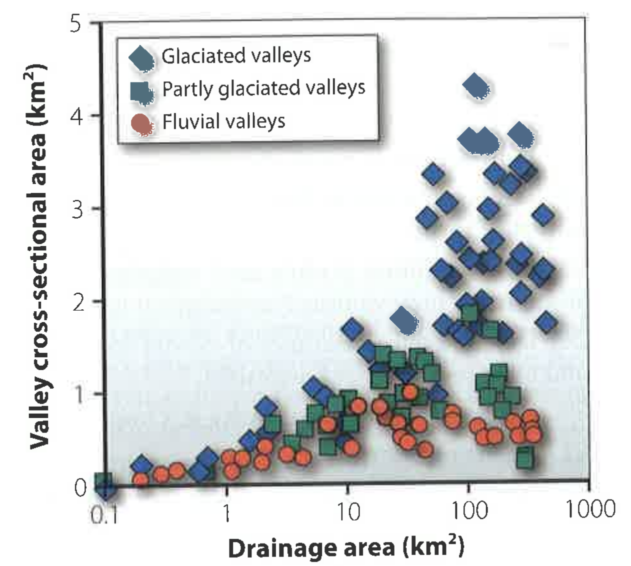

FIGURE DD9.4 Comparison of cross-sectional areas for valleys in the Olympic Peninsula, Washington State, showing the greater amount of rock excavated from valleys with major alpine glaciers (blue diamonds) than from partly glaciated valleys (green squares) and unglaciated fluvial valleys (red circles). [From Montgomery (2002).]

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WORKED PROBLEM

Question: The existence of glaciers implies a balance in mass and energy fluxes into and out of the glacial system. Using what you know about glaciers, create a qualitative model of a glacier considering both mass and energy balance; that is, the glacier is in dynamic steady state. The model, which can be in the form of an equation or a discussion, should show the ways and locations in which mass and energy are gained and lost from the glacier.

Answer: If the glacier is in a dynamic steady state, then both the energy and mass inputs and outputs must be in balance. In other words, the amount of energy gained by the glacier must equal that lost, and, similarly, mass gains and losses from the glacier must be in balance.

Glaciers gain mass from precipitation, blow-in of snow, and condensation of water vapor onto the ice surface. Mass can be added in both the accumulation and ablation zones but is lost primarily in the ablation zone. Glaciers lose mass by calving at their terminus, melting at the surface and on the bed, and sublimation from the ice surface.

Energy is lost and gained from glaciers through a number of different pathways. Sunlight adds energy to glacial surfaces and geothermal heat is added from the bed below. Latent and sensible heat can carry energy both to and from the glacier surface depending on the temperature contrast between the ice and atmosphere. Energy is lost from the ice surface by blackbody radiation.

KNOWLEDGE ASSESSMENT Chapter 9

1. Explain the three physical properties of ice that make it so dynamic near Earth's surface.
2. Compare the size, shape, and location of ice sheets, ice caps, ice fields, and alpine glaciers.
3. By what processes do glaciers lose mass?
4. Predict how a newly deposited snow crystal will change over time before eventually becoming glacier ice.
5. Make a sketch showing the energy balance of a glacier including all the major terms.
6. Explain glacial advance and retreat in terms of mass balance.
7. Predict the effect of elevation on glacial ice accumulation and ablation.
8. Give three reasons why the upper reaches of glaciers function as net accumulation zones.
9. Define the equilibrium line and suggest how you might approximate its location in the field.
10. Describe two ways in which glaciers move.
11. Write Glen's flow law, and describe how it is used.
12. Sketch the movement of ice in an alpine glacier using flow lines in a cross section.
13. Explain why floating and grounded ice margins behave differently.
14. Define cold-based and warm-based glaciers.
15. Explain how regelation works and how it can help glaciers erode rock.
16. List three observations that indicate glaciers are capable of abrading bedrock outcrops.
17. Explain how cirques, arêtes, and tarns are interrelated.
18. Explain the glacial buzzsaw hypothesis.
19. Describe glacial till.
20. What does an esker look like and how does one form?
21. Describe a drumlin and suggest where these landforms are likely to be found.
22. Explain why the nature and texture of ice-marginal sediment is so variable over short distances?
23. How are kettle ponds thought to form?
24. Define outwash and outwash plains and explain how they form.
25. What landform is often used to determine the elevation of now-vanished ice-marginal lakes?
26. Describe the appearance of varved sediment and explain how it was deposited.
27. Define IRD and explain why it is important.
28. Define periglacial.
29. What characteristics best define periglacial environments?
30. Explain the global distribution of permafrost.
31. What do relict permafrost features tell us about paleoclimate?
32. Consider the relationship between energy balance and the thickness, extent, and continuity of permafrost.
33. Define talik and active layer.
34. What is felsenmeer, how is it related to frost shattering, and where might you find it?
35. Define solifluction and discuss when and where it is most active.
36. Where is thermokarst found and what are some of its characteristic features?
37. Sketch the process by which ice wedges and ice wedge casts form.
38. Compare the similarities and differences between glaciers and rock glaciers.
39. What is a pingo and where would you go to find one?
40. How are glaciers, water supply, and climate change related?