C H A P T E R 1 3

Glacial and Periglacial Landscapes

Vertical aerial photograph of a part of Malaspina Glacier, Alaska, shows intricately deformed moraines that have been compressed into tight folds. Stagnant ice zone below forested ridges (red color) is pockmarked with kettles.

GLACIERS

Glaciers as Part of the Cryosphere

A person living in New York City who wished to see a glacier firsthand would have to travel some 2500 km west to the Rocky Mountains or a nearly equal distance north into eastern Canada to find even a few small examples. Yet, about 20,000 years ago thick glacier ice covered the entire landscape between western Canada and New York City. The southern margin of the ice followed an irregular line from the Montana Rockies across the Great Plains to central Illinois, then eastward beyond New York City to the now-submerged continental shelf beyond southern New England. If people inhabited North America at that time, they could not have lived at the site of downtown New York because Manhattan Island lay buried beneath the glacier.

Similarly, Stone Age people, known to have been living in Europe since long before the culmination of the last glacial age, were driven southward by another vast ice sheet that spread across northern Europe and stretched from the British Isles in the west far eastward into the region of the Soviet Union.

Not only during the last glaciation, but also during numerous earlier ones, glaciers formed where none exist today and caused dramatic changes in the Earth's surface environments. However, glacial ages have not typified all of Earth history; instead, they seem to have been clustered in intervals that were separated by many millions of years. Such intervals are now thought to be due to the shifting of lithospheric plates that move continents from warm latitudes into colder ones. These crustal movements cause mountain ranges to be uplifted, thereby changing the paths of ocean and wind currents and permitting glaciers to form and expand in size. The succession of glacial and interglacial ages within these longer episodes seems to be controlled mainly by slight but important changes in the position of the Earth relative to its primary source of energy, the Sun. These changes affect the amount and seasonal distribution of radiation striking any point on the Earth's surface (Chapter 20). Relatively small changes in energy input apparently can lead to substantial changes in the amount and distribution of glacier ice. This in turn can affect surface conditions over much of the planet, for as ice sheets form and grow in subpolar latitudes, world sea level falls and the paths and moisture content of wind systems change, causing a shift of the Earth's climatic zones.

Glaciers constitute an important part of the cryosphere, that portion of the planet where temperatures are so low that water exists primarily in the frozen state. Other components of the cryosphere include snow, perennially frozen ground, and sea, lake, and river ice. Most glacier ice on the Earth resides in the polar regions, above the Arctic and Antarctic circles. In these regions sea ice forms a vast sheet over the polar seas, but its extent fluctuates seasonally. Because it is so thin (generally 3 m or less), it is not as important volumetrically as the ice contained in large ice sheets, but it has an important effect on global climate because of its highly reflective surface.

Large fluctuations also occur in the areal extent and volume of glaciers, but on a longer time scale. Widespread deposits and glacially eroded terrain beyond existing glaciers point to changing climatic conditions on the Earth, for glaciers and other

FIGURE 13.1 Glaciers in eastern Greenland spill from cirques carved in mountain flanks and enter deep ice-eroded valleys that constrain them and channel their flow.

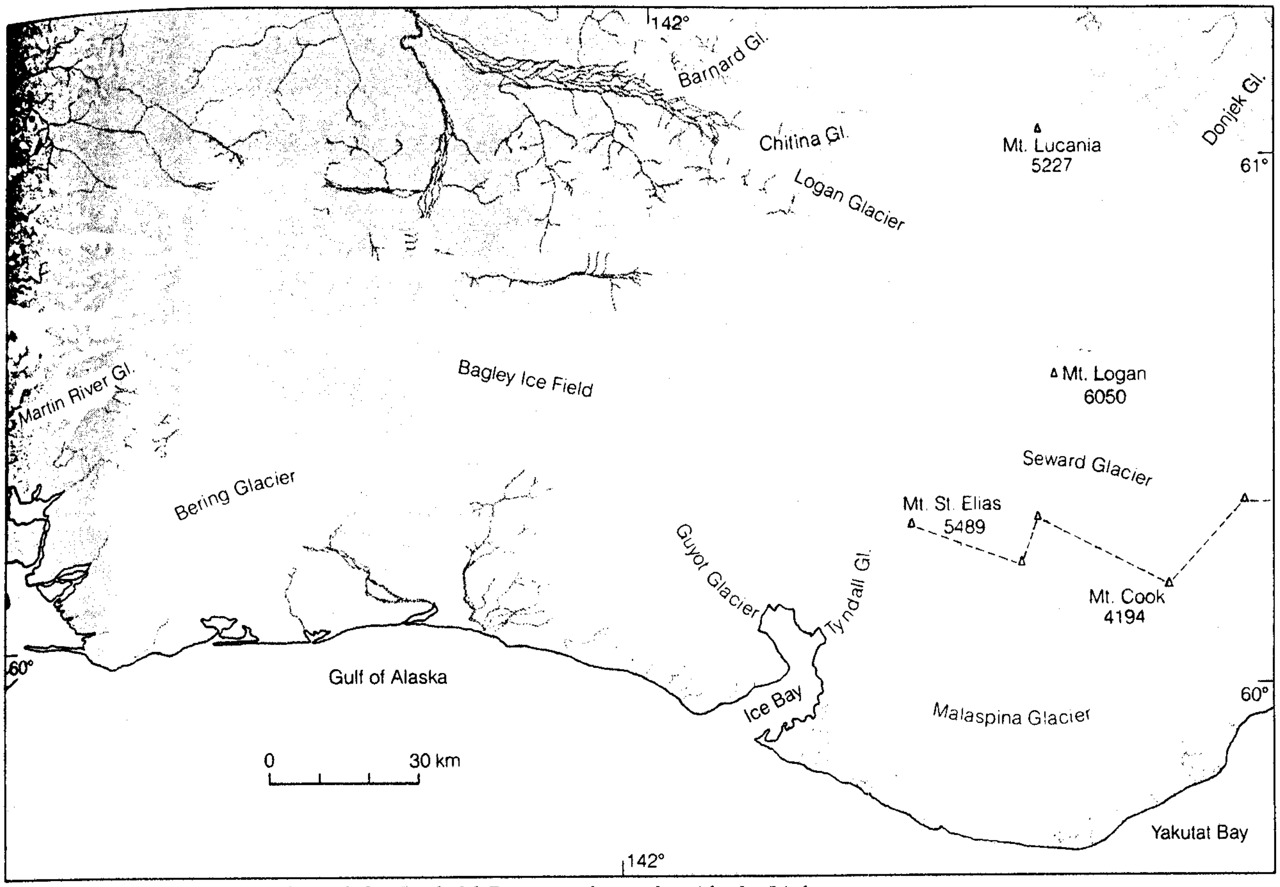


FIGURE 13.2 Glacier complex of the Icefield Ranges along the Alaska/Yukon border. Malaspina Glacier and Bering Glacier are broad piedmont lobes fed by ice draining from a large intermontane icefield. The base of each lies far below sea level. If they were to recede, long deep fjords would appear and extend far inland from the present coast.

forms of ice are very sensitive to climate. Therefore, a study of their past distribution can provide important information about global changes of climate over millions of years.

Forms of Glaciers

Defined simply, a ***glacier*** is *a body of ice, consisting largely of recrystallized snow, that shows evidence of downslope or outward movement due to the pull of gravity*. On the basis of form and extent, several classes of glaciers can be distinguished (Table 13.1). The smaller types are confined by surrounding topography that determines their shape and direction of movement. The smallest glaciers mostly occupy protected hollows or depressions on the sides of mountains. Larger glaciers spread downward onto valley floors where their shapes are controlled mainly by the structure and erosional pattern of the bedrock landscape on which they lie (Fig. 13.1). Most of the world's glaciers are no more than about 1–2 km long and have areas of less than several square kilometers, but some large valley glaciers in the high mountains of Alaska and central Asia reach lengths of tens of kilometers.

Still larger glaciers spread beyond the confining walls of mountain valleys and onto gentle slopes where topography exerts little control on their form. Malaspina Glacier and Bering Glacier, the two largest glaciers of this type in North America, occupy lowlands along the Gulf of Alaska and are fed by extensive intermontane ice fields in the coastal ranges of Alaska and Yukon Territory (Fig. 13.2). Ice caps of various sizes cover mountain highlands or lower-lying lands at high latitude, and display generally radial outward flow. One well-studied example is the small tropical Quelccaya Ice Cap of the Peruvian Andes which covers 70 km^2 and lies at altitudes of 4950–5645 m. Vatnajökul, a far larger ice cap in Iceland, measures

TABLE 13.1 *Principal Types of Glaciers, Classified According to Form*

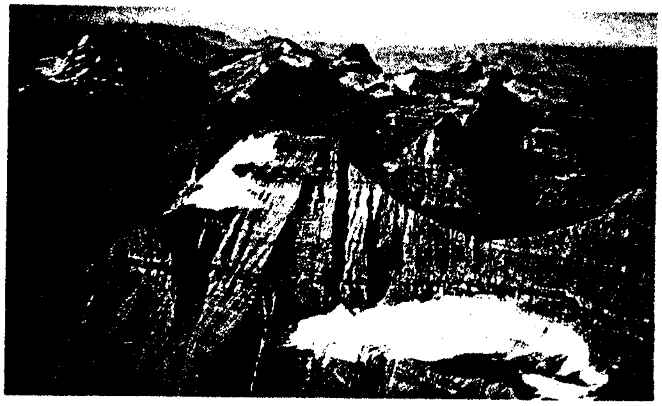
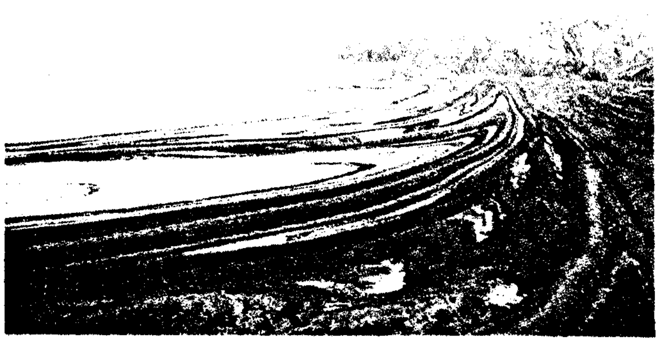
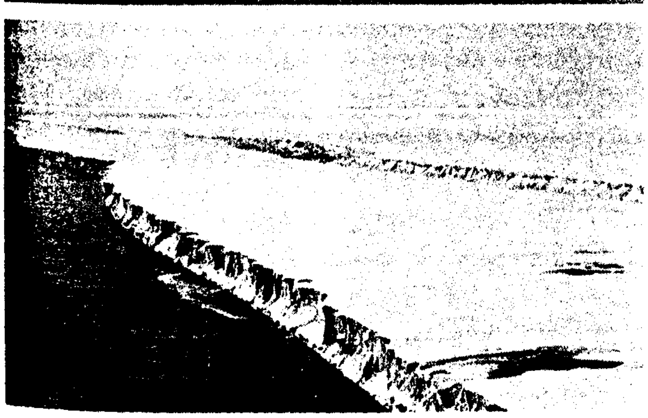
	Glacier Type	Characteristics
	Cirque glacier	Occupies bowl-shaped depression on the side of a mountain (*Sexton Glacier, Glacier National Park, Montana*)
	Valley glacier	Flows from cirque(s) onto and along floor of valley (*Trimble Glacier, Alaska Range*)
	Fjord glacier	Occupies a submerged coastal valley and its base lies below sea level. May have steep terminus that recedes rapidly by frontal calving. (*Muir Glacier, Glacier Bay, Alaska*)
	Piedmont glacier	Terminates on piedmont slopes beyond confining mountain valleys and is fed by one or more large valley glaciers. (*Malaspina Glacier, Alaska*)

TABLE 13.1 (Continued)

	Glacier Type	Characteristics
	Ice cap	Dome-shaped body of ice and snow that covers mountain highlands, or lower-lying lands at high latitudes, and displays generally radial outward flow. (*South Patagonian Ice Cap, Chile and Argentina*)
	Ice field	Extensive area of ice in a mountainous region that consists of many interconnected alpine glaciers. Lacks domal shape of ice caps. Its flow is strongly controlled by underlying topography. (*Juneau Icefield, Alaska*)
	Ice sheet	Continent-sized masses of ice thick enough to flow under their own weight and which overwhelm nearly all land within their margins. (*West Antarctic Ice Sheet*)
	Ice shelf	Thick glacier ice that floats on the sea and commonly is located in coastal embayments. (*Pine Island Ice Shelf, Antarctica*)

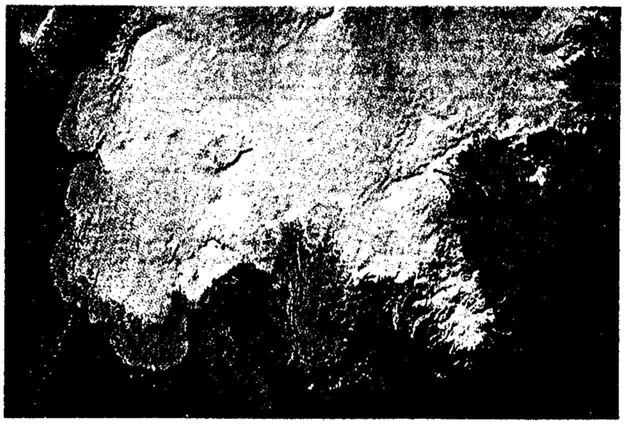


FIGURE 13.3 Vertical satellite image of Vatnajökul ice cap near the southeast coast of Iceland. The firn limit separates new white snow of the accumulation area from darker ice of the ablation area. Moraine bands on the glacier surface are oriented in the direction of ice flow. Dark bands of volcanic ash, locally deformed by flow, cross the ablation zone in several places. Braided streams have built extensive outwash plains beyond the glacier front.

about 100 by 140 km across and terminates close to sea level (Fig. 13.3).

Ice sheets are the largest glaciers on the Earth. These continent-sized masses of ice overwhelm nearly all the land surface within their margins. Modern ice sheets are confined to Greenland and Antarctica, and collectively comprise about 95 percent of all glacier ice on our planet. During glacial ages, as we shall see later, ice sheets also covered extensive portions of North America and Eurasia. The Greenland Ice Sheet, which has an area approximately equal to that of the United States west of the Rocky Mountains, reaches such a great thickness (some 3000 m) that the crust of the Earth beneath much of it has been depressed below sea level by its weight. Were the glacier suddenly to melt away, the island of Greenland would have the unusual form of an elongated ring of land enclosing an extensive arm of the sea.

On many maps, Antarctica appears to be covered by a single vast glacier, but in reality it consists of two large ice sheets that meet along the lofty Transantarctic Mountains (Fig. 13.4). The East Antarctic Ice Sheet is the larger of the two and covers the continent of Antarctica. It is the only truly polar ice sheet on Earth, for the North Pole lies at the center of the deep Arctic Ocean which is covered only by a thin layer of sea ice. Because of its ice sheet, Antarctica has the highest average altitude and the lowest average temperature of all the continents. The smaller West Antarctic Ice Sheet overlies numerous islands of the Antarctic archipelago. Like its more massive neighbor, *terrestrial* portions rest on land that rises above sea level while *marine* portions cover land lying below sea level. Measurements of ice thickness obtained by sending radio waves through the glacier where they then bounce off rocks at its base show that Antarctic ice reaches a thickness of 3600 m or more. The combined estimated volume of the two ice sheets of close to 24×10^6 km^3 would be sufficient to raise world sea level by nearly 60 m if the ice were to waste away entirely (Fig. 13.5).

Ice caps typically have a relatively simple geometry and constitute a single broad dome. Large ice sheets, on the other hand, may be more complex and consist of several domes from which ice flows radially to the ice margin or to broad interdome saddles where the ice flow diverges downslope. The location of ice domes and ice saddles determines the flow path of ice within an ice sheet, but such features do not necessarily remain fixed. Instead, they may shift position with time as an ice sheet grows or shrinks in size.

Ice shelves occur at several places along the margins of the Greenland and Antarctic Ice Sheets, as well as locally in the Canadian Arctic islands. They are mainly located in large coastal embayments (Fig. 13.4), are attached to land on one side, and their seaward margin generally forms a steep ice cliff rising as much as 50 m above sea level. The largest ice shelves extend hundreds of kilometers seaward from the coastline and can reach a thickness of at least 1000 m. They are nourished by ice streams flowing off the land, as well as by direct snowfall on their surface.

Large tabular icebergs produced by marginal breakup of floating ice shelves can form "ice islands;" some, discovered drifting in the Arctic Ocean, have been used as remote stations by polar scientists. Recently, studies have been made regarding the feasibility of towing massive tabular icebergs from Antarctic waters to arid countries of the Middle East where they could provide large quantities of fresh water for agricultural use. Although the bergs would have to be towed very slowly northward across warm equatorial waters where substantial melting would occur, calculations suggest that sufficient ice would remain at the end of a trip to make such a venture feasible. Some day desert countries may be growing abundant food crops irrigated by meltwater that originated as snowflakes falling in central Antarctica tens of thousands of years ago!

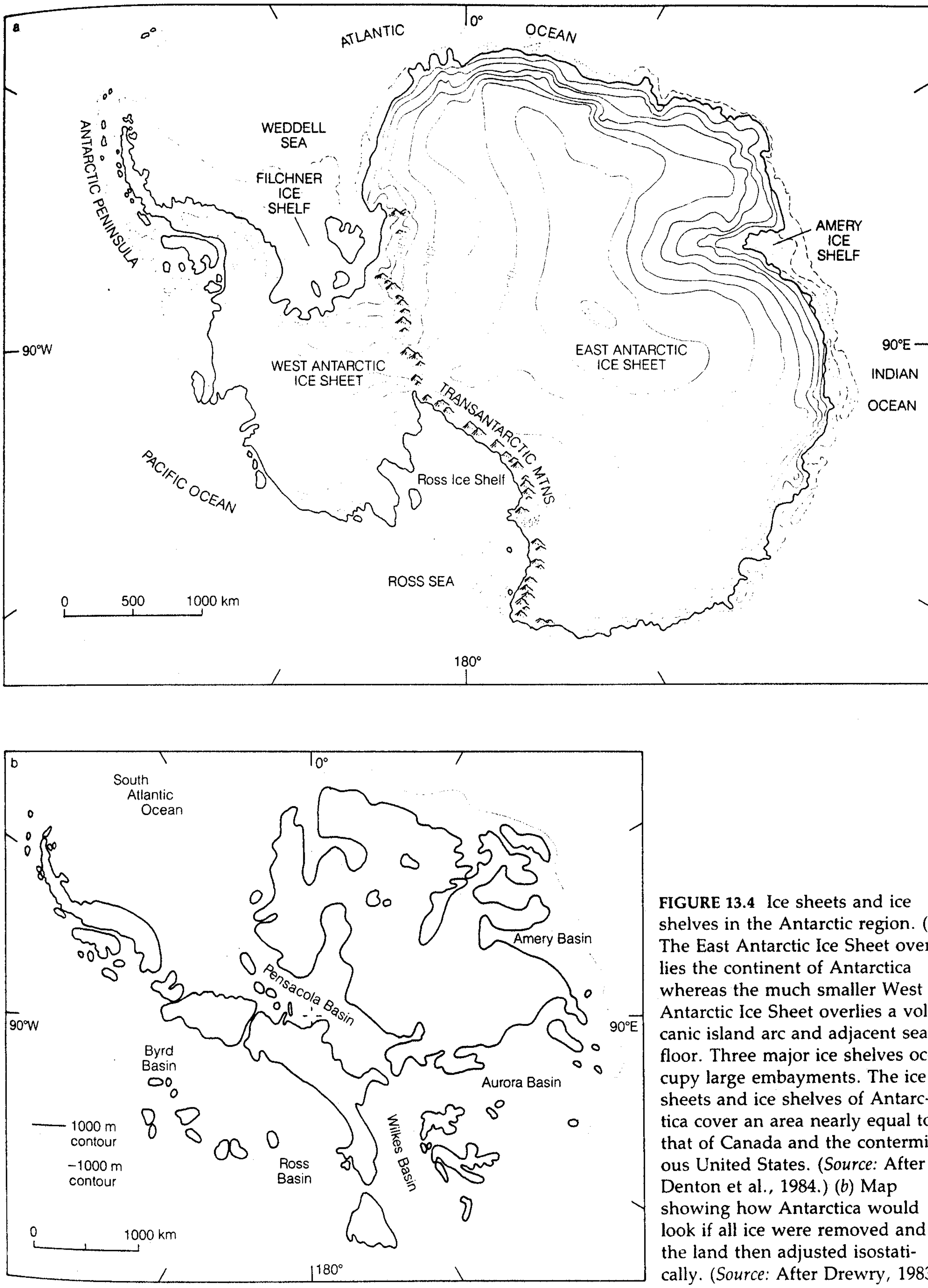


FIGURE 13.4 Ice sheets and ice shelves in the Antarctic region. (*a*) The East Antarctic Ice Sheet overlies the continent of Antarctica whereas the much smaller West Antarctic Ice Sheet overlies a volcanic island arc and adjacent sea floor. Three major ice shelves occupy large embayments. The ice sheets and ice shelves of Antarctica cover an area nearly equal to that of Canada and the conterminous United States. (*Source:* After Denton et al., 1984.) (*b*) Map showing how Antarctica would look if all ice were removed and the land then adjusted isostatically. (*Source:* After Drewry, 1983.)

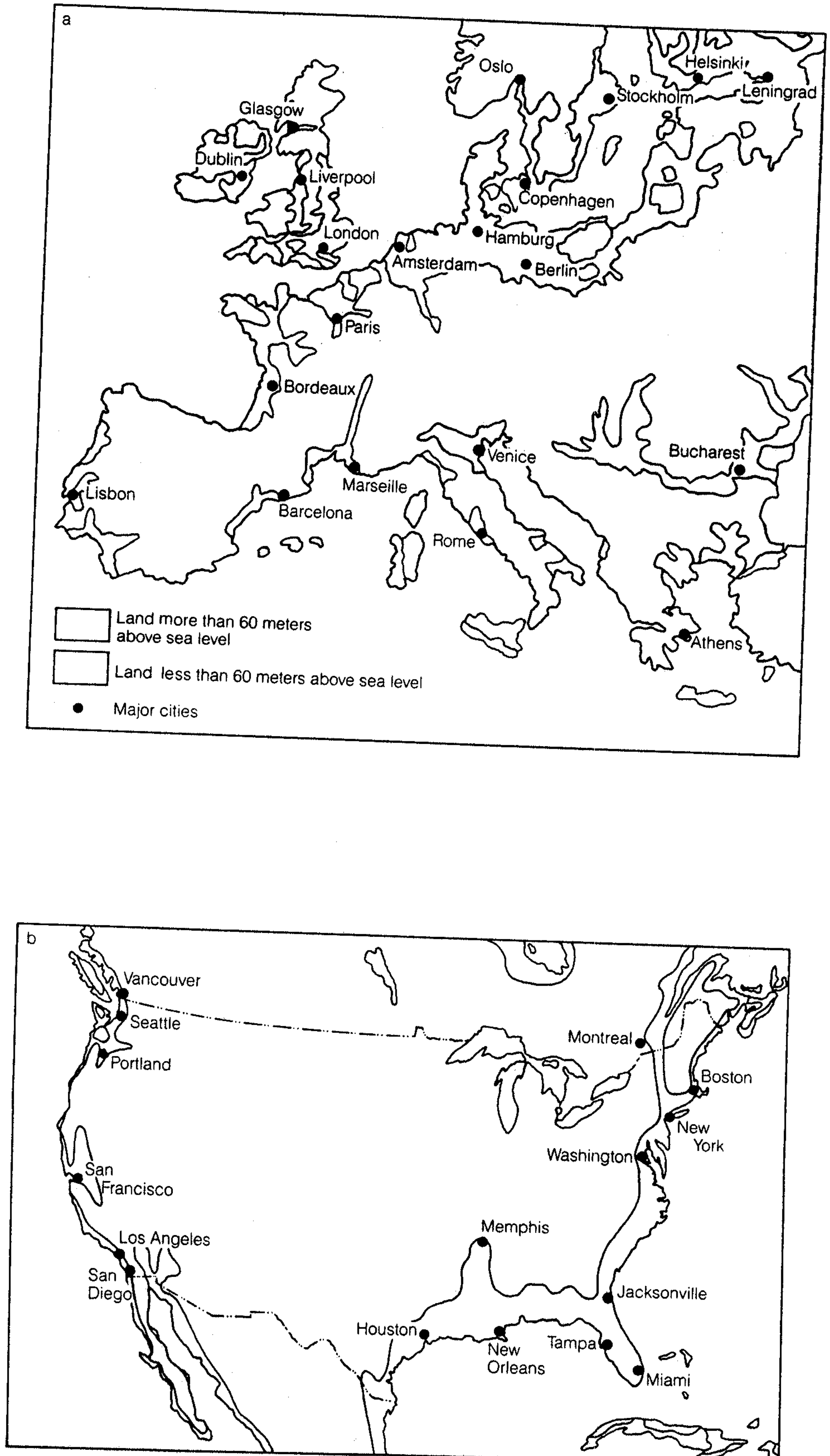


FIGURE 13.5 Parts of (*a*) coastal Europe and (*b*) central North America that would be flooded if all the glacier ice on Antartica were to melt. The volume of ice involved is nearly equivalent to a 60 m rise of world sea level. Many major cities of western Europe and the United States lie below the 60 m contour and would therefore be totally or partially submerged. Melting of all the additional ice on the earth would cause the oceans to rise about another 6 m.

Temperatures in Glaciers

Except for a thin surface layer that is chilled below freezing each winter, the ice throughout many glaciers is at the ***pressure melting point,*** *the temperature at which ice can melt at a particular pressure* (Figs. 3.6 and 13.6). Under such conditions meltwater and ice can exist together in equilibrium. ***Temperate*** or ***warm glaciers,*** *in which the ice is at the pressure melting point throughout,* are found mainly in low and middle latitudes. At high altitudes and latitudes, where the mean annual air temperature lies below freezing, ice temperature drops below the pressure melting point and little or no seasonal melting occurs. *Glaciers whose ice remains below the pressure melting point* are termed ***polar*** or ***cold glaciers. Subpolar glaciers,*** an intermediate type, may *have surface temperature at the freezing point in summer, but temperatures beneath the upper meter or two of ice remain below freezing.* Some large glaciers that originate in high mountains may contain polar or subpolar ice in their higher parts but have temperate ice in their lower parts. Warm basal ice may also occur in the deepest parts of large thick ice sheets where very high pressure at the glacier bed allows the ice to reach the pressure melting point.

Ice temperature is very important in controlling the way glaciers move and their rate of movement. Meltwater at the base of temperate glaciers acts as a lubricant and permits the ice to slide across its bed. By contrast, polar glaciers are so cold they are frozen to their bed. The motion they display does not involve basal sliding, and their rate of movement is greatly reduced.

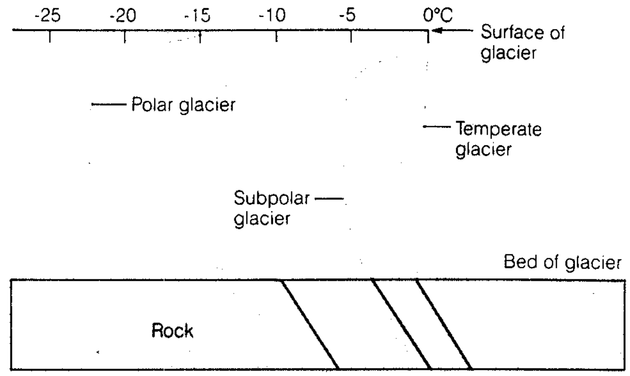


FIGURE 13.6 Temperature profiles for polar, subpolar, and temperate glaciers. Ice in temperate glaciers is at the pressure melting point from surface to bed, whereas in polar glaciers the temperature remains below freezing and the ice is frozen to its bed. In subpolar glaciers only a thin surface zone may seasonally reach the melting point. (*Source:* After Meier, 1964.)

Snowline

The two chief requirements for the existence of glaciers are adequate snowfall and low temperatures, both of which depend on climate. Such requirements are fulfilled at high latitudes and at high altitudes. These conditions also are more frequently met in moist coastal regions than in the dry interiors of continents. It is not surprising, therefore, that existing ice sheets lie in high latitudes and are surrounded by marine waters. Most of the world's numerous smaller glaciers are found 1) in moist coastal mountain systems, such as the high cordillera of northwestern North America; 2) on polar and subpolar islands like Spitsbergen and Iceland; and 3) in the rugged mountain ranges of central Asia, which are so high that very low temperatures offset the prevailing aridity and allow glaciers to develop.

Glaciers can only form at or above the ***snowline,*** *the lower limit of perennial snow.* Because its altitude is controlled mainly by temperature and precipitation, the snowline rises from near sea level in polar latitudes to altitudes of about 5000–6000 m in the tropics, and it also rises inland from moist coastal regions toward the drier interiors of large islands and continents (Fig. 13.7). Where high coastal peaks intercept moist air traveling onshore, resulting in strong climatic contrasts on opposite sides of mountain ranges, the snowline rises inland with a steep gradient.

Mass Balance

The mass of a glacier is constantly changing as the weather varies from season to season and, on longer time scales, as local and global climates change. These ongoing environmental changes cause fluctuations in the amount of snow added to the glacier surface, and in the amount of snow and ice lost by melting. These, in turn, determine the ***mass balance*** of the glacier, which is *a measure of the change in total mass during a year.*

Mass balance is measured in terms of ***accumulation,*** *the addition of mass to the glacier,* and of ***ablation,*** which is *the loss of mass to the glacier.* Accumulation occurs mainly as snowfall, whereas ablation takes place mainly through melting. In the case of subpolar and polar glaciers, however, ablation may result from evaporation of meltwater from the ice surface, or from direct vaporization without the ice passing through a liquid phase (a process called *sublimation*). Ablation may also involve melting at the base of ice shelves, or the breaking off of bergs into the sea or into lakes marginal to the ice.

If, during a year, more mass is added to a glacier than is lost, then the result is a positive mass balance (Fig. 13.8). By contrast, if more mass is lost than gained, the glacier experiences a negative mass balance. If the balance is mainly positive over a period of years, it means that the glacier is increasing in mass. Accordingly, the front, or *terminus*, of the glacier is likely to advance as the glacier grows. A succession of predominantly negative years normally leads to retreat of the terminus. Alternatively, if no net change in mass occurs, the glacier is in a balanced state. If this condi-

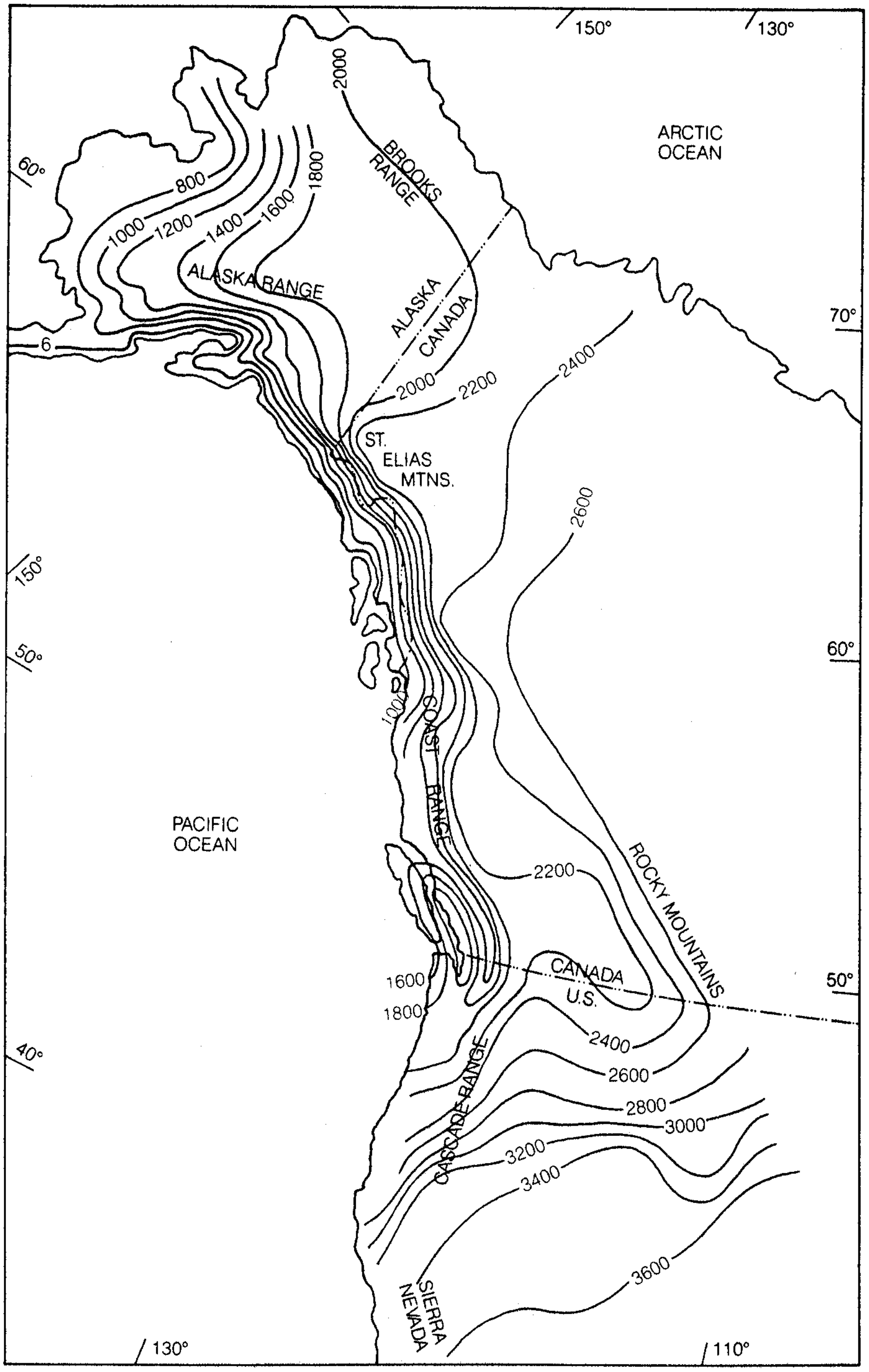


FIGURE 13.7 Contours, in hundreds of meters, show the regional pattern of glacier equilibrium-line altitudes throughout northwestern North America at the end of the 1961 balance year. The surface defined by the contours rises steeply inland from the Pacific coast in response to increasingly drier climate, and also from north to south in response to rising mean annual temperatures. (*Source:* After Meier and Post, 1962.)

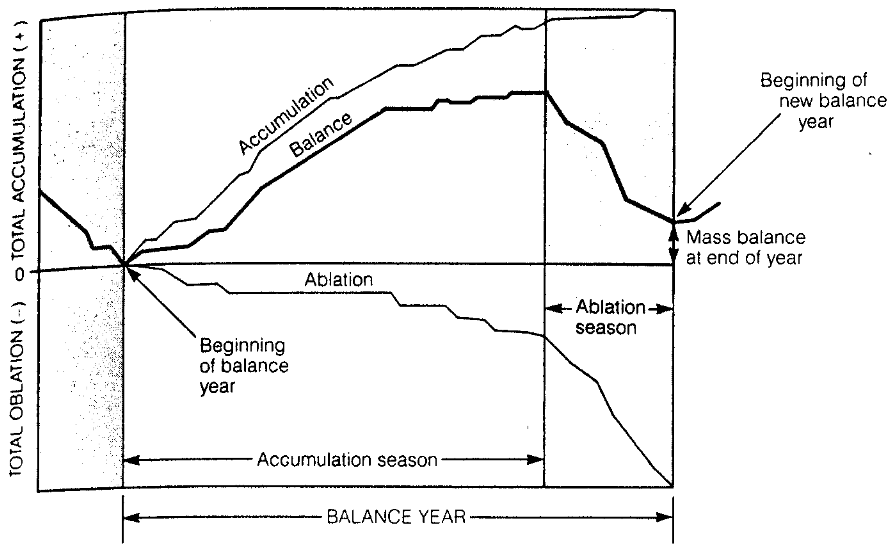


FIGURE 13.8 Diagram showing how accumulation and ablation determine glacier mass balance (heavy line) over the course of a balance year. The balance curve rises during the accumulation season as mass is added to the glacier, then falls during the ablation season as mass is lost. Mass balance at the end of the balance year reflects the difference between mass gain and mass loss. (*Source:* After International Commission of Snow and Ice, 1969.)

tion persists, the terminus is likely to remain relatively stationary.

If a mountain glacier is viewed at the end of the summer ablation season, two zones are generally visible on its surface. An upper zone, the ***accumulation area,*** is *that part of a glacier covered by remnants of the previous winter's snowfall and is an area of net gain in mass.* Below it lies the ***ablation area,*** *a region of net loss characterized by a dark-toned surface of bare ice and old snow from which the previous winter's snowcover has largely melted away* (Figs. 13.1 and 13.3). If a glacier is close to a balanced condition, then on average about two thirds of its total area lies in the accumulation area. The ***equilibrium line*** separates the accumulation area from the ablation area and, therefore, *marks the level on the glacier where net loss equals net gain.* On temperate glaciers it coincides with the lower limit of fresh snow at the end of the summer (the snowline). When a balanced condition exists, the equilibrium line lies approximately midway in altitude between the terminus and the head of the glacier.

Being very sensitive to climate, the equilibrium line fluctuates in altitude from year to year and is higher in warm dry years than in cold wet years (Fig. 13.9). Its position during a glacial age is many hundreds of meters lower than during an interglacial age, which helps explain the great contrast in the areal extent of glaciers at such times.

Response of the Glacier Terminus

Although measurement of the mass balance of a glacier may provide an excellent indication of its current "state of health," observations of its marginal fluctuations are not as good an indicator, for there normally is a lag in the time it takes for the terminus to respond to a change in climate. The lag reflects the time it takes for the effects of an increase or decrease in accumulation rate above the equilibrium line to be transferred through ice flow to the glacier terminus. The length of the response lag depends on the size and flow characteristics of a glacier, and will be longer for large glaciers than for small ones, and longer for cold glaciers than for warm ones. For glaciers of modest size in temperate latitudes (like those in the European Alps), response lags may range from several years to a decade or more. This partly explains why in any area having glaciers of different sizes, fluctuations of glacier margins may not be synchronous.

Conversion of Snow to Glacier Ice

Glacier ice is basically a metamorphic rock, for it consists of interlocking crystals of the mineral ice and has been deformed by flow due to the weight of overlying snow and ice. Newly fallen snow consists of hexagonal crystals we know as snowflakes.

It is very porous, having a density less than a tenth that of water. Air easily penetrates the pore spaces where the delicate points of snowflakes gradually disappear due to evaporation. The resulting water vapor then condenses, mainly in constricted places near the centers of ice crystals. In this way, snowflakes gradually become smaller, rounder, and thicker, and the pore spaces between them disappear (Fig. 13.10). The entire mass of snow takes on the granular appearance that we associate with old snowdrifts at the end of winter. In the process, snow is transformed from a loose sediment into a more cohesive mass. *Snow that survives a year or more of ablation and achieves a density that is transitional between snow and glacier ice* is called *firn*. Ultimately, firn passes into true glacier ice when it becomes so dense that it is no longer permeable to air. Although now a rock, such ice has a far lower melting point than any other naturally occurring rock, and its density of about 0.9 g/cm^3 means that it will float in water.

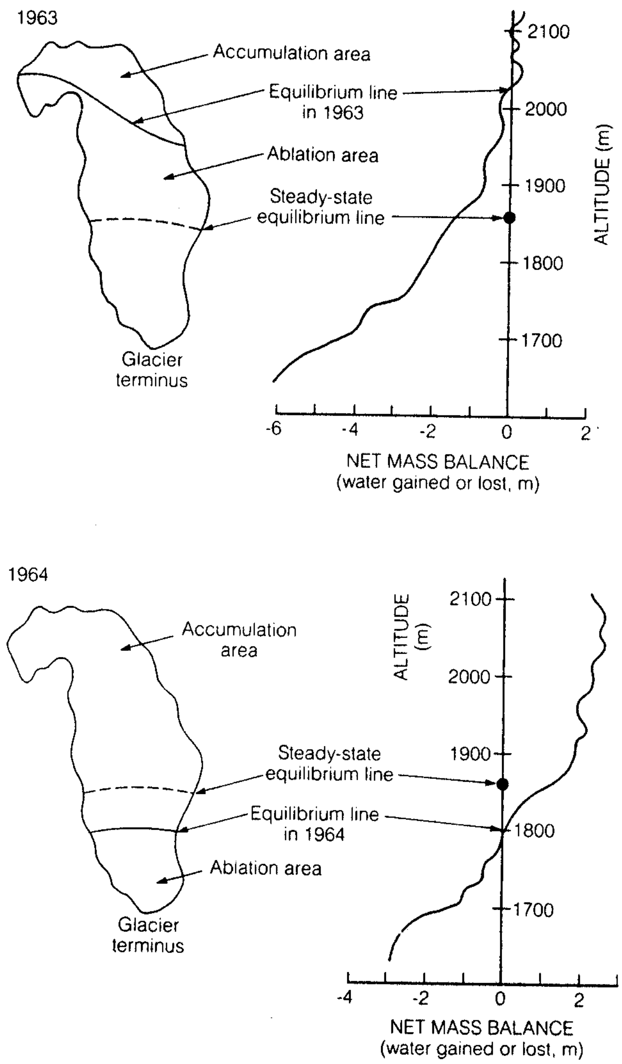


FIGURE 13.9 Maps of South Cascade Glacier in Washington State at end of 1963 and 1964 balance years, showing the position of the equilibrium line relative to the position it would have under a balanced (steady-state) condition. Curves show values of mass balance as a function of altitude. During 1963, a negative balance year, the glacier lost mass and the equilibrium line was high (2025 m). In 1964, a positive balance year, the glacier gained mass and the equilibrium line was low (1800 m). (*Source:* After Meier and Tangborn, 1965.)

Movement of Glaciers

Glacier Flow

Newly formed ice in a glacier consists of a myriad of randomly oriented small interlocking crystals having a texture like that of igneous rocks (Fig. 13.11). At some point, depending on the steepness of the surface on which it is lying and on the surrounding temperature, a mass of compacted snow and ice begins to deform and flow downslope under the pull of gravity. The flow takes place mainly through movement within individual ice crystals. Crystals in the accumulating glacier are subjected to higher and higher stress as the weight of the overlying snow and ice increases. Under this stress, deformation (*creep*) takes place along internal planes in an ice crystal in much the same way that playing cards in a deck slide past one another if the deck is pushed from one end. As movement proceeds, differential pressures between crystals cause some to grow at the expense of others, and the resulting larger crystals end up having a similar orientation. This leads to increased efficiency of flow, for the internal creep planes of all crystals now are approximately parallel (Fig. 13.12).

Importance of Water

In temperate glaciers, ice flow is enhanced by the presence of water, especially at the glacier bed. Measurement of the progressive deformation and downglacier displacement of a borehole drilled to the bed of a glacier shows that only a part of the observable surface motion is due to internal creep. An important component of movement results from sliding of the glacier across its bed (Fig. 13.13). In some glaciers basal sliding may contribute up to 90 percent of the total measured velocity.

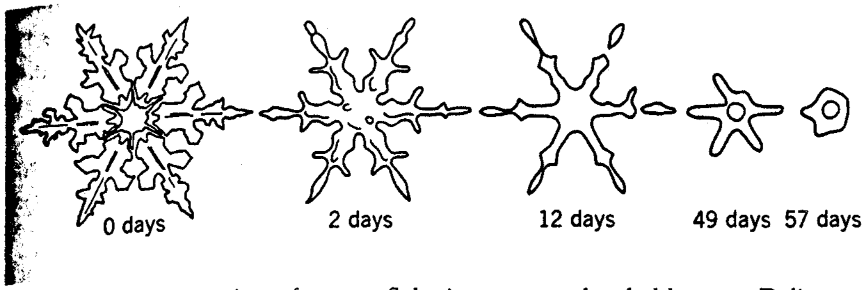


FIGURE 13.10 Conversion of a snowflake into a granule of old snow. Delicate points of a snowflake disappear through melting and evaporation. The resulting water refreezes and vapor condenses near the center of the crystal, making it denser. (*Source:* After Bader et al., 1939.)

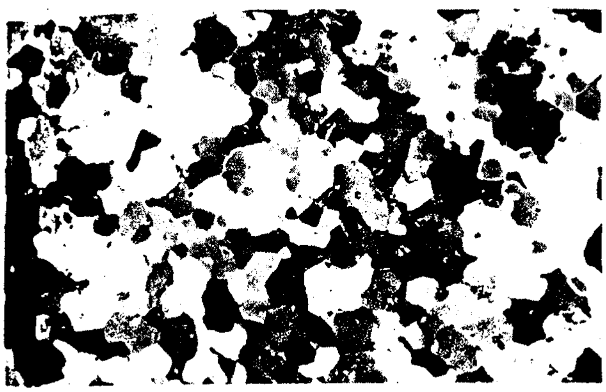


FIGURE 13.11 Thin section of glacier ice showing interlocking structure of ice crystals. An average crystal in this view has a diameter of about 3 mm.

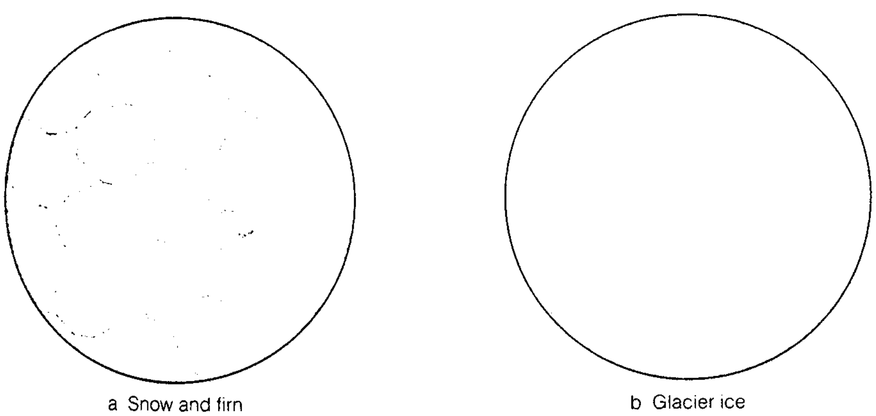


FIGURE 13.12 Arrangement of ice crystals in snow, firn, and glacier ice. (*a*) Random arrangement of crystals in snow and firn. Crystal planes along which creep could occur are not parallel. (*b*) Oriented crystals of glacier ice. Crystals are arranged with creep planes parallel, enhancing the ease of internal flow.

Crevasses

The surface portion of a glacier, having little weight upon it, is brittle. Where a glacier flows over an abruptly steepened slope, such as a bedrock cliff, the surface ice is subjected to tension and it cracks. The cracks open up and form ***crevasses,*** which are *deep, gaping fissures in the upper surface of a glacier* (Fig. 13.26). Although they are often difficult and dangerous to cross, crevasses are

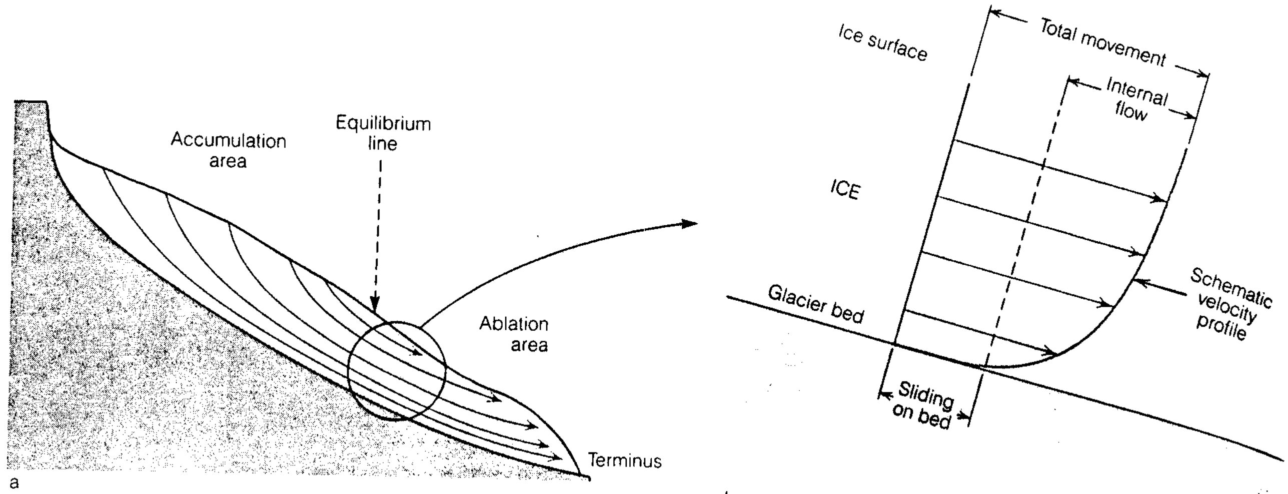


FIGURE 13.13 Flow of ice within a glacier. (*a*) Snow accumulating above the equilibrium line is compacted and flows downward and toward the terminus. The flow lines emerge at the surface below the equilibrium line in the ablation area. (*b*) In a vertical velocity profile through a temperate glacier a portion of the total observed movement is due to internal flow within the ice (with velocity increasing upward from the bed), whereas part is due to sliding of the glacier along its bed, lubricated by a film of meltwater.

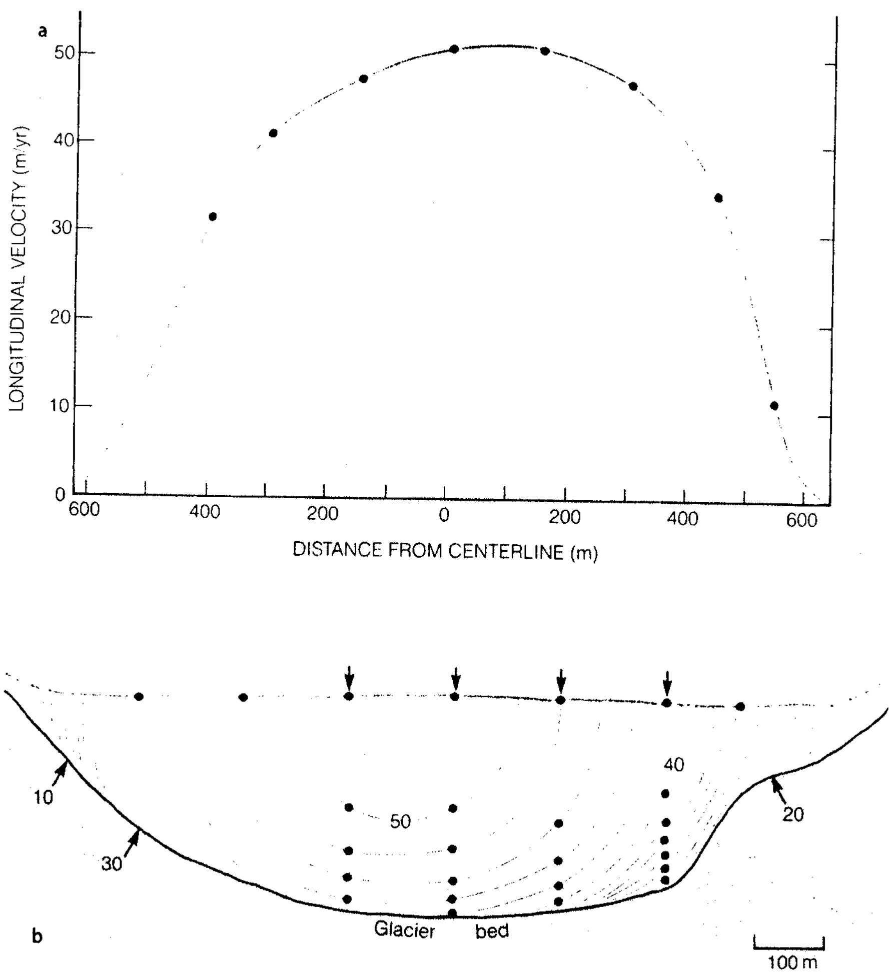


FIGURE 13.14 Velocity of flow within a valley glacier. (*Source:* After Raymond, 1971.) (*a*) Map showing down-glacier velocity profile across Athabasca Glacier in the Canadian Rocky Mountains. Velocity is highest at the center of the glacier and decreases rapidly toward margins. (*b*) Vertical section through Athabasca Glacier at right angles to the direction of flow showing velocity distribution (m/yr). Arrows mark positions of boreholes. Velocity decreases toward bed and margins.

rarely as much as 50 m deep. At greater depths internal flow prevents crevasses from forming. Because it cracks at the surface, yet flows at depth, we can liken a glacier to the upper layers of the Earth itself, which consist of a surface zone that cracks and fractures (the lithosphere) and a deeper zone (the asthenosphere) that can flow very slowly.

Rates of Flow

The surface velocity of a valley glacier can be measured by surveying from the sides of the valley, at intervals of time, a line of markers or targets extending across the glacier (Fig. 13.14*a*). Results show that surface ice in the central part of the glacier moves faster than ice at the sides, similar to the velocity distribution in a river. The reduced rates of flow toward the margins are due to frictional drag against the valley walls. A similar reduction in flow rate toward the bed is observed in a vertical profile of velocity, obtained by drilling a borehole through a glacier and measuring the angle of the hole with an inclinometer lowered into it. If measurements at different depths are repeated after a year, the annual flow as a function of depth can be determined (Fig. 13.14*b*).

If a glacier is in a balanced condition, then the ice passing through any transverse vertical section in the accumulation area must equal the amount of snow added to the surface upglacier. At the same time, ice passing through any transverse section in the ablation area must equal the amount of ice lost between that section and the terminus. Therefore, the ice flowing through any cross section must steadily increase downglacier toward the equilibrium line, then decrease downglacier away from the equilibrium line. This means that ice-flow velocity should be highest near the equilibrium line.

High rates of flow are observed where a glacier moves over an abrupt cliff to form a steep *icefall*. However, flow velocities in most glaciers range from only a few centimeters to a few meters a day, or about the same slow rate that groundwater percolates through crustal rocks. Probably hundreds of years have elapsed since ice now exposed at the terminus of a very long glacier fell as snow near the top of its accumulation area.

Directions of Flow Within a Glacier

Although snow continues to pile up in the accumulation area each year, while melting removes snow and ice from the ablation area, the surface profile of a glacier does not change much because ice is transferred from the accumulation area to the ablation area. If the altitude of the surface is to remain relatively unchanged, then ice flow cannot parallel the surface. Instead, the movement must be downward in the accumulation area, where mass is being added, and upward in the ablation area, where mass is being lost (Fig. 13.13). Crystals of ice that enter the glacier near its head, therefore, have a long path to follow before they emerge near the terminus, whereas those falling closer to the equilibrium line may travel only a short distance through the glacier before reaching the surface again.

Measurements show that marginal parts of the ice sheets in Antarctica are flowing at rates of about 50 m/yr (about 15 cm/day). Ice streams within the body of the ice sheets follow large valleys in the underlying bedrock and are not always visible at the surface. Such ice streams, somewhat analogous to well-defined currents in the oceans, are flowing as much as ten times faster than the ice around them.

Nonstable Behavior

Most glaciers slowly expand or contract in size as the climate fluctuates, but certain glaciers change dimensions rapidly and in a way that is either unrelated, or only secondarily related, to climatic change.

Glacier Surges

From time to time certain glaciers experience ***surges,*** which are *unusually rapid rates of movement marked by dramatic changes in glacier flow and form.* Such events appear mainly to affect temperate valley glaciers, although sectors of ice caps have been observed to surge and it is believed that portions of ice sheets may also be capable of surging. When a surge occurs, a glacier seems to go berserk. Before a surge, the lower part of the glacier often consists of stagnant ice. As the surge begins the boundary between active ice in the upper glacier and stagnant ice below moves rapidly downglacier and a chaos of crevasses and ice pinnacles forms. Although the terminus does not always advance, in some cases advances of up to several kilometers have been observed. Rates of movement as great as 100 times those of normal glaciers and averaging as much as 6 km a year have been measured during surges. Most surges run their course within a year or two, after which the lower part of the

a

b

FIGURE 13.15 Glaciers experiencing normal flow and surge flow. (*a*) Banded moraines of Barnard Glacier, Alaska, are oriented parallel to ice-flow direction, indicating stable flow. (*b*) Moraines of Tweedsmuir Glacier, Alaska, are contorted by periodic surges of tributary ice streams.

glacier will slowly revert to a stagnant condition before the next surge begins, generally a decade or more in the future.

A surging glacier can frequently be identified from the air or on satellite images because bands of rock debris on its surface tend to be intricately folded, in contrast to the generally parallel pattern of debris bands on nonsurging glaciers (Fig. 13.15). The reasons for glacier surging are not fully understood. It is generally believed that as water accumulates beneath a glacier over a period of years high pressures are generated within the water that lead to widespread separation of the ice from its bed. The resulting effect is similar to what happens when a rapidly moving automobile encounters a wet street during a rainstorm: The weight of the car places the water layer beneath the tires under so much pressure that they are floated off the wet pavement and the vehicle slides along out of control. The high velocities observed during glacier surges, therefore, probably result from greatly enhanced basal sliding rather than from any significant change in internal flow rate.

Calving Glaciers

When Captain George Vancouver sailed up the coast of southeastern Alaska in the late eighteenth century he plotted on his charts the positions of a number of large ice streams that issued from the high coastal mountains and terminated near the rocky shore. Today, in their place, one finds long open ***fjords.*** Such *deep glacially carved valleys submerged by the sea* extend many tens of kilometers back into the mountains. The dramatic recession of these fjord glaciers during the past century and a half at rates far in excess of typical glacier retreat rates on land is due to frontal ***calving,*** a process that involves *the progressive breaking off of icebergs from a glacier that terminates in deep water* (Fig. 13.16). Although the base of a fjord glacier may lie far below sea level along much of its length, its terminus can remain stable as long as it is "grounded" against a shoal. However, if the glacier's mass balance becomes negative, the front will recede into deeper water and calving can proceed. Once it commences, calving may continue rapidly and irreversibly until the glacier front once again becomes grounded, generally near the head of the fjord. Calving glaciers may undergo extensive retreat while nearby glaciers that rest on dry land fluctuate through only short distances in response to climatic variations. The final breakup and disappearance of the interior parts of the great ice sheets of eastern North America and Europe during the last glaciation very likely involved the retreat of calving margins that rapidly cut back into the ice sheets and caused them to collapse.

While most of the large fjord glaciers of coastal Alaska have retreated well back into their source regions during recent decades, an exception is Columbia Glacier which has remained relatively stationary with its terminus near the point of its greatest recent advance. During the early 1980s, signs of imminent retreat were detected along its front as calving increased and the terminus began to recede into deeper water. Calculations indicate that the rate of recession is likely to increase dramatically as the terminus calves, releasing a vast number of icebergs. Many of the bergs are likely to drift across nearby shipping lanes where large oil tankers enter and leave the port of Valdez at the southern end of the Alaska Pipeline, thereby creating potential hazards to navigation. When retreat is

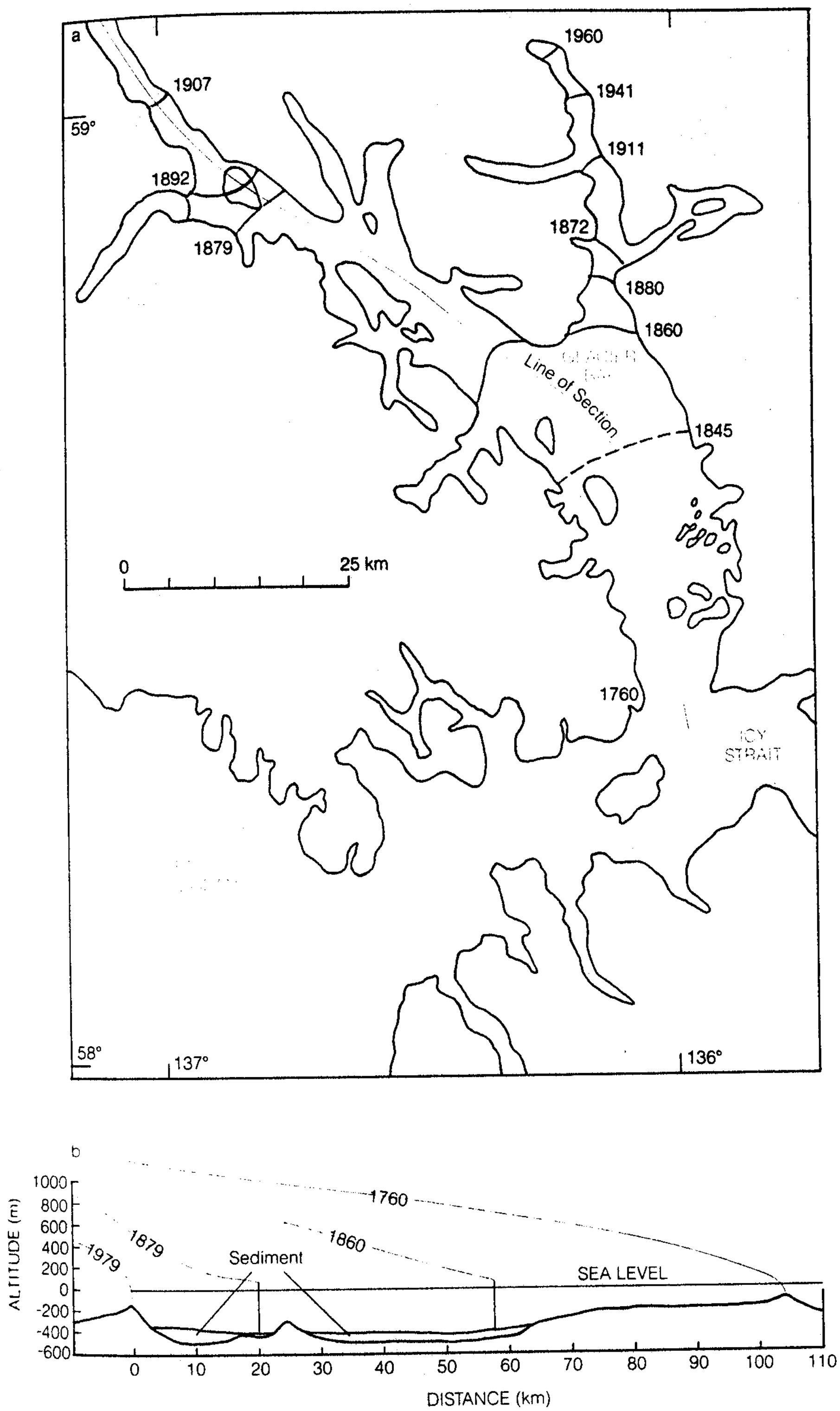


FIGURE 13.16 Rapid glacier recession in Glacier Bay, Alaska, resulting from frontal calving. (*a*) Map of Glacier Bay showing historically observed positions of the ice front between 1760 and 1960. Over these two centuries, the glacier retreated about 100 km into the upper reaches of its fjord system. (*b*) Section along length of Glacier Bay showing four successive profiles. Note steep vertical front of calving terminus and sediment deposited on fjord bottom since recession of the glacier. (*Source:* After Brown et al., 1982.)

complete, perhaps late in the next century, a newly exposed long fjord system will extend as an arm of the sea far back into the coastal mountains.

GLACIATION

Landscapes in Canada, northern United States, and northern Europe differ from those somewhat farther south, a principal reason being that these northern regions have been glaciated. ***Glaciation,*** defined as the *modification of the land surface by the action of glacier ice,* occurred so recently that weathering, mass-wasting, and erosion by running water have not had time to alter the landscape appreciably. Except for a cover of vegetation, the appearance of these glaciated landscapes has remained nearly unchanged since they emerged from beneath the ice. Like the geologic work of other surface processes, glaciation involves erosion, transport, and deposition of sediment.

Glacial Erosion and Sculpture

In changing the surface of the land over which it moves, a glacier acts collectively like a plow, a file, and a sled. As a plow it scrapes up weathered rock and soil and plucks out blocks of bedrock; as a file it rasps away firm rock; and as a sled it carries away the load of sediment acquired by plowing and filing, along with additional rock debris fallen onto it from adjacent slopes.

Small-Scale Erosional Features

The base of a temperate glacier is studded with rock particles of various sizes. The fragments move with the flowing ice across underlying bedrock and produce ***glacial striations,*** which are *long subparallel scratches inscribed on a rock surface by rock debris embedded in the base of a glacier,* as well as larger grooves that also are aligned in the direction of ice flow (Fig. 13.17). Striations are abraded as well on the moving rock fragments themselves. In places, fine particles of sand and silt in the basal ice act like sandpaper, and polish the rock until it has a smooth reflective surface. At the same time the basal ice drags at the bedrock, breaking off blocks (usually along joints or fractures) and quarrying them out. Blocks are mainly removed on the downglacier sides of hillocks, whereas on the upglacier sides abrasion and polishing of the rock is dominant. The asymmetry of the resulting landforms clearly indicates the direction in which the glacier was moving (Fig. 13.18).

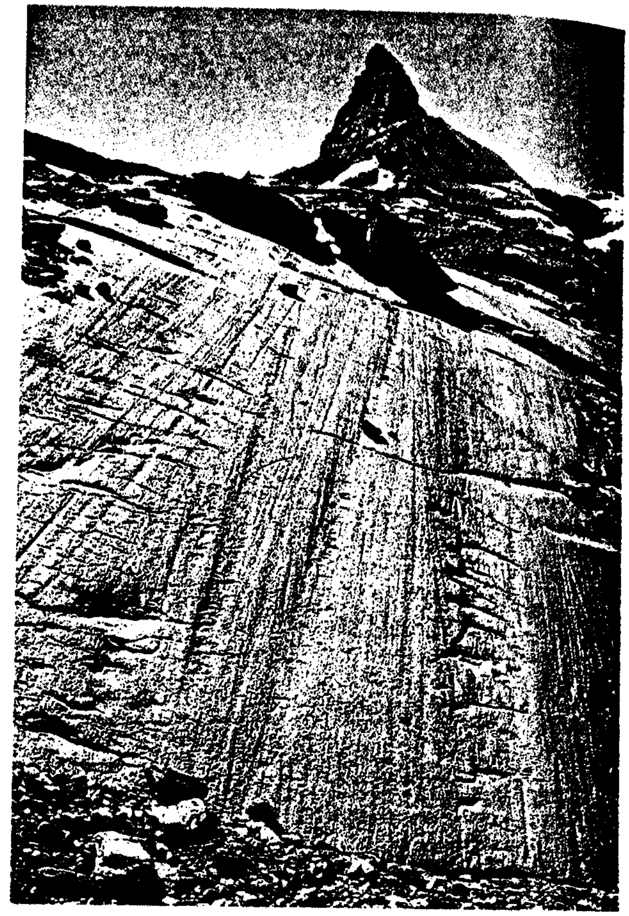


FIGURE 13.17 Recently deglaciated bedrock surface beyond Findelen Glacier, Swiss Alps. Debris carried at the base of the glacier produced grooves, striations, and polish on bedrock as the ice flowed forward in the direction of the Matterhorn. Crescentic marks on the rock surface represent fractures produced by the impact of large stones against the rock surface, which slopes gently in the up-glacier direction.

Landforms of Glaciated Mountains

Cirques and Related Forms. Most of the world's high mountains owe their scenic grandeur to sculpture by present and former glaciers. These mountains bear a distinctive suite of landforms attributable to glacial erosion that are lacking in nonglaciated uplands. Among the most characteristic is the ***cirque,*** *a bowl-shaped hollow on a mountainside, open downstream and bounded upstream by a steep slope (headwall), and excavated mainly by frost-wedging and by glacial plucking and abrasion* (Fig. 13.19). The floors of many cirques are rock basins. Some contain small lakes, called *tarns,* ponded behind a bedrock threshold at the edge of the cirque.

A cirque probably begins to form beneath a large

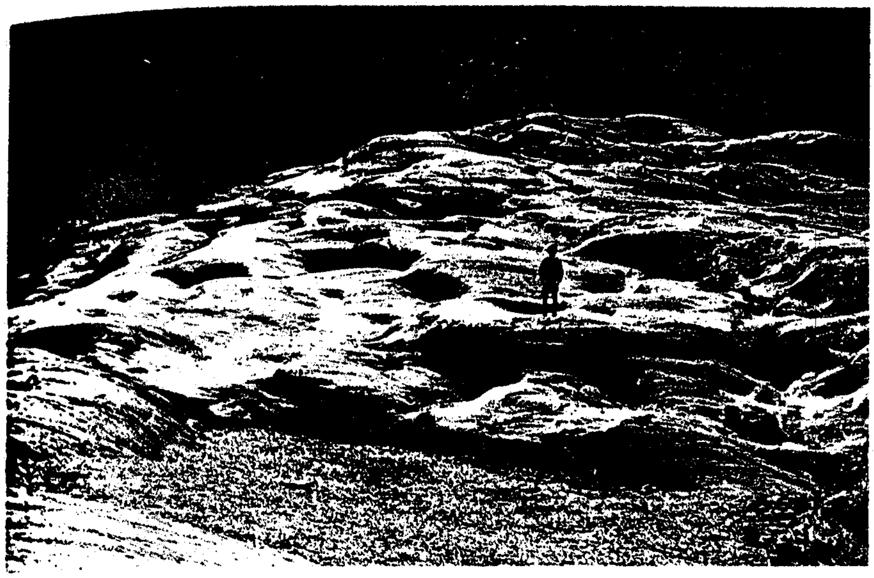


FIGURE 13.18 Asymmetrical glacially sculptured bedforms in front of Franz Josef Glacier in New Zealand's Southern Alps. The glacier flowed from right to left. Up-glacier slopes are smooth and polished. Scarps facing downvalley result from the plucking of bedrock blocks by flowing ice.

snowbank or snowfield just above the snowline. Meltwater infiltrating rock openings beneath the snow refreezes and expands, disrupting the rock and dislodging fragments. Small rock particles are then carried away by snowmelt runoff during periods of thaw. This activity gradually creates a depression in the land and enlarges it. As the snowbank turns into a glacier, plucking helps to enlarge the cirque still more and abrasion at the bed will further deepen it.

Most cirques owe their form to repeated episodes of glaciation. During interglacial periods, rock fragments dislodged from headwalls of ice-free cirques by frost action accumulate as taluses, which are transported away when glaciers reform during the next glaciation. Through successive glacial and interglacial ages cirques on opposite sides of mountain crests expand headward, creating a characteristic assemblage of features for which we use names given to them by Alpine mountaineers (Fig. 13.20).

Glacial Valleys. Glaciated valleys differ from ordinary stream valleys in several ways. Their chief characteristics, not all of which are present in every case, include a cross profile that is troughlike (U-shaped) and a floor that lies below the floors of tributary valleys, from which streams often descend as waterfalls or cascades (Fig. 13.20). The tributaries are referred to as *hanging valleys.* They "hang" above the floor of the main valley

FIGURE 13.19 Cirque carved in the side of the Aiguille Noire in the Mont Blanc massif of the northern Italian Alps. During glacial ages, ice spilled over the threshold of the cirque as an icefall and joined a large glacier in the main valley below.

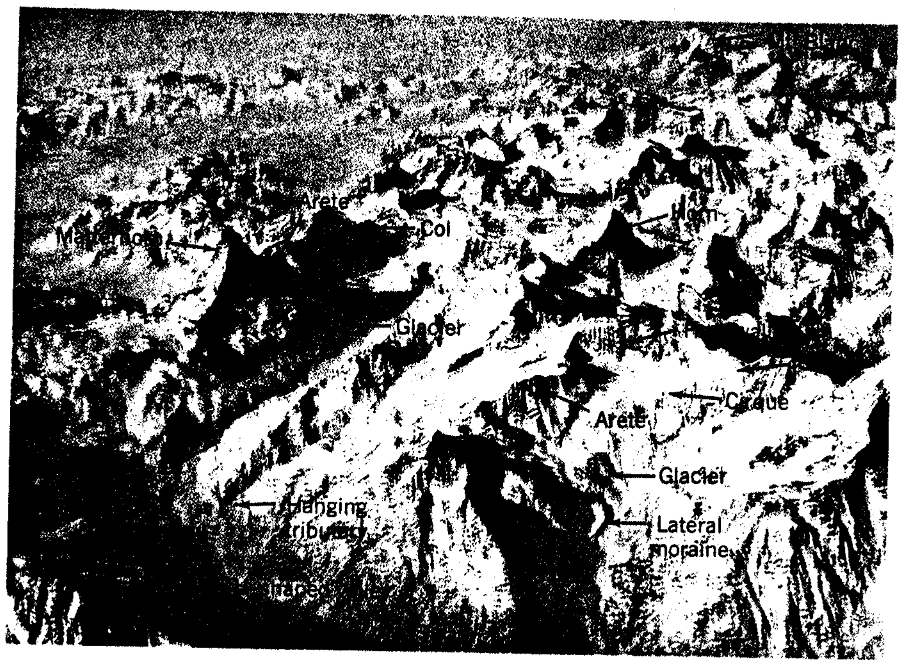


FIGURE 13.20 Photograph of central Swiss Alps near Zermatt showing alpine glacial landforms. Major U-shaped valleys head in cirques and have hanging tributary valleys. The sharp-pointed Matterhorn (middle distance) is a classic **horn,** *a sharp-pointed peak bounded by the intersecting walls of three or more cirques.* Many peaks are connected by an **arête,** *a jagged, knife edge ridge created where glaciers have eroded back into the ridge,* or by a **col,** *a gap or pass in a mountain crest where the headwalls of two cirques intersect.*

because tributary ice streams can merge with the glacier in the main valley well above its base. Unlike a river, in which water from tributaries quickly mixes with the main stream, tributary glaciers retain their identity and flow beside, or are inset within, the main valley glacier. Both the main valley and the tributary valleys are shaped by erosion at the sides as well as the base of the glacier. The long profile of a glaciated valley floor also may be marked by steplike irregularities and shallow basins. These are related to the spacing of joints in the rock, which influences the ease of glacial plucking, or to changes in rock type along the valley. Finally, the valley is likely to head in a cirque or group of cirques.

Some valleys are glaciated from head to mouth. Others are glaciated only in their headward parts; downstream their form has been shaped largely by mass-wasting processes and stream erosion. The resulting contrasts in landscape form make it relatively easy to distinguish glaciated from nonglaciated valley segments on topographic maps, on aerial photographs, or on images taken by satellites orbiting 100 km or more above the Earth's surface (Fig. 13.21).

FIGURE 13.21 Vertical satellite image of high mountains bordering the Indus River in northern Pakistan. Glaciated valley segments (G) with typical U-shaped cross profile and cirques (C) at their head contrast with nonglaciated V-shaped valleys (V) downstream.

Fjords. Long, deep fjords are common features along the mountainous west-facing coasts of Norway, Alaska, British Columbia, Chile, and New Zealand (Fig. 13.22). The form and depth of many fjords imply glacial erosion of 300 m or more. Sognefjord in Norway reaches a depth of 1300 m, yet near its seaward end it shallows to only about 150 m. The characteristic overdeepening of fjord floors far below sea level is due mainly to deep glacial erosion resulting from thick and fast-moving ice. The frequent linear geometric arrangement of fjord systems suggests that geologic structures in the bedrock exert a strong control on glacial erosion. Unlike streams, which cease to erode when they reach the sea, glaciers can erode their beds far below sea level. For example, a large coastal glacier 300 m thick, and having a specific gravity close to 0.9, can continue to erode its bed until it is in water about 270 m deep, whereupon it begins to float and basal erosion ceases.

Landforms Associated with Ice Caps and Ice Sheets

Abrasional Features. Landscapes glaciated by ice sheets display the small-scale erosional features typical of most glaciated terrain. Striations, especially, have been helpful to geologists in reconstructing the flow lines of long-vanished northern ice sheets. Erosional features indicative of basal sliding are common in the central zones of former ice sheets where the ice was between 3 and 4 km thick, indicating that basal ice was at the pressure melting point. However, in a peripheral zone, evidence of glacial erosion is often less obvious, leading to the conclusion that thinner ice there may have been very cold and largely frozen to its bed.

Where ice sheets overwhelmed mountainous terrain, as in the cordillera of northwestern North America, the upper limit of glaciation can frequently be seen where smooth, abraded mountain slopes pass abruptly upward into rugged frost-shattered peaks and mountain crests. In such terrain, some divides between adjacent drainages are broad and smooth and show evidence of glacial abrasion and plucking where they were overridden by ice (Fig. 13.23).

Streamlined Forms. In many areas that lie near the outer edge of former ice sheets, the land surface has been molded into smooth, nearly parallel ridges that range up to many kilometers in length. These forms resemble the streamlined bodies of supersonic airplanes and offer minimum resistance to glacier ice flowing over and around them. The best-known variety of streamline forms is the ***drumlin,*** *a streamlined hill consisting of glacially deposited sediment and elongated parallel with the direction of ice flow* (Fig. 13.24). Not all such landforms consist

FIGURE 13.22 Icebergs break away from the calving front of a fjord glacier in southwest Greenland. This fjord was filled by an arm of the sea as the glacier retreated back rapidly by frontal calving.

FIGURE 13.23 High pass in northern Cascade Range, Washington that was overriden by ice sheet flowing south from Canada during the last glacial age. Ice over the adjacent valleys was at least 1500 m thick. Sharp-crested summits to the left and in the distance stood above the ice sheet as **nuntaks,** *areas of ice-free land rising above the surface of a glacier.*

FIGURE 13.24 A field of drumlins near Snare Lake, Saskatchewan, Canada. The higher, wider, and blunter side of each drumlin points toward the direction from which the ice flowed (from the northeast). Some of the hills have tapered, streamlined tails.

of contemporaneous ice-laid sediment. Some are cored by preexisting glacial deposits, and these too are drumlins. Others are shaped by glacial erosion of bedrock; even though they have the form of a drumlin they are not true drumlins, but *rock drumlins.* These landforms also owe their streamlined shape to molding by flowing ice, and their long axes lie parallel to the direction of flow of the glacier that produced them.

Lake Basins. The margins of ice sheets typically are lobate, a result of control of ice flow by subglacial topography. In the north-central United States, for example, the irregular southern margin of the last continental ice sheet marked the terminus of lobes that flowed along former drainage courses and deepened them into basins that now contain the Great Lakes. These lakes and other large ones in central Canada lie along the boundary between resistant metamorphic rocks and more erodible sedimentary rocks that overlie them. As the spreading ice sheet rose against and crossed the low northward-facing sedimentary escarpment, glacial erosion created an arcuate series of basins that filled with meltwater as the ice receded. These basins still retain large lakes today. Analogous lake basins are found inside glacial limits within and near many mountain ranges. The famous Alpine lakes of Switzerland and northern Italy, like those of the southern Andes and New Zealand, are glacially deepened basins surrounded by glacial deposits.

Glacial Transport

A glacier differs from a stream in the way in which it carries its load of rock particles. Part of its load can be carried at its sides and even on its surface. A glacier can carry much larger pieces of rock and it can transport large and small pieces side by side without segregating them according to size and density into a bed load and a suspended load. Because of these differences, deposits made directly from a glacier are neither sorted nor stratified.

The load of a glacier typically is concentrated at its base and sides because these are the areas where glacier and bedrock are in contact and where abrasion and plucking are effective. Much of the rock material on the surface of valley glaciers arrived there by rockfalls from adjacent cliffs.

A good deal of the load in the base of a glacier consists of very fine sand and silt. The particles are mostly fresh and unweathered. As revealed under the microscope, they have jagged, angular surfaces (Fig. 5.20*a*). These *fine rock particles, the products of glacial crushing and grinding,* are referred to as ***rock flour.*** They differ from the more rounded and chemically weathered particles found in sediments of nonglaciated areas.

Much of the transfer of sediment from a glacier to the ground occurs by release of particles as the surrounding ice melts. Therefore, glacial deposition takes place below the equilibrium line where melting is dominant.

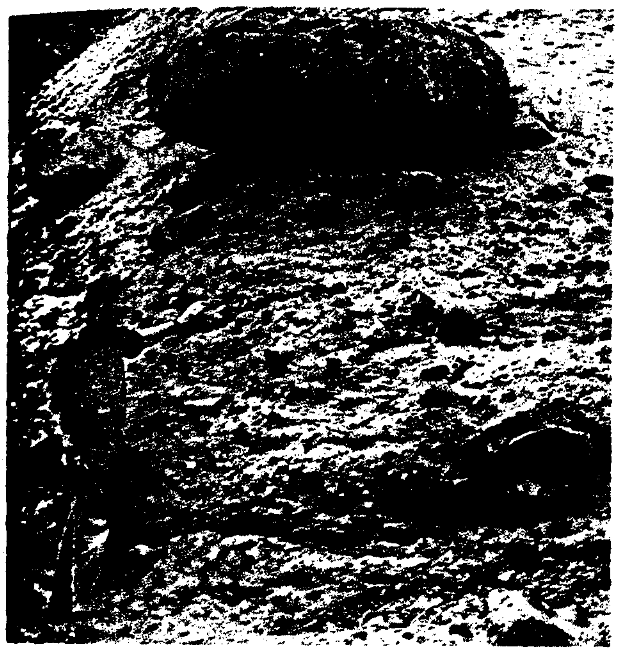
Glacial Deposits

Drift, Till, and Stratified Drift

Sediment deposited directly by glaciers or indirectly by meltwater in streams, in lakes, and in the sea together constitute ***glacial drift,*** or simply ***drift.*** The term *drift* dates from the early nineteenth century when it was vaguely conjectured that all such deposits had been "drifted" to their resting places during the biblical flood of Noah or by some other ancient body of water. Included within drift are several kinds of sediment that form a gradational series ranging from nonsorted to sorted types.

At one end of the range is ***till,*** which is *nonsorted drift deposited directly from ice.* The name was given by Scottish farmers long before the origin of the sediment was understood. The constituent rock particles in till are not sorted according to size or density, but lie just as they were released from the ice (Fig. 13.25*a*).

Probably most till is plastered onto the ground, bit by bit, from the base of flowing ice in the ablation area (Fig. 13.26). We say "probably" because as yet the process has not been directly observed at the glacier bed. Most tills are a random mixture of rock fragments consisting of a *matrix* of fine-grained sediment surrounding larger stones of various sizes. The till matrix consists largely of sand and silt particles derived by abrasion of the glacier bed and from reworking of preexisting fine-grained sediments. Pebbles and larger rock fragments in till often have faceted surfaces, the result of abrasion, and some are striated (Fig. 13.25*b*). Both the stones and the coarser matrix grains in till tend to have their longest axis aligned in the direction of ice flow.



a

b

FIGURE 13.25 Glacial deposits. (*a*) Bouldery till in an end moraine of an alpine glacier in the eastern Cascade Range showing wide range in grain size and lack of sorting. (*b*) Striated boulder embedded in till on a land surface that was exposed by recent retreat of Ruitor Glacier in the Italian Alps.

Glacial marine drift, which closely resembles till, is *sediment deposited in the sea from floating ice shelves or bergs.* As an iceberg or the base of an ice shelf melts, the contained sediment is released and falls to the seafloor. Unlike till, the elongate rock particles in glacial marine drift are randomly oriented, for they settle through water rather than being deposited by flowing ice. Furthermore, such drift may contain the remains of marine organisms, still in growth position.

Where many icebergs are generated and the resulting sedimentation rate is high, a continuous layer of glacial marine drift will accumulate. However, where bergs are less plentiful and fine-grained sediment is settling out continuously from suspension, stones dropped from passing bergs will plunge into marine sediments on the seafloor, deforming them. Such *dropstones* are diagnostic features of glacial marine environments, as well as ice-marginal lake environments where bergs are produced from a calving glacier front.

A number of other nonglacial sediments resemble till and are easily confused with it. This has sometimes led to misinterpretations about the former extent and age of glaciations. For example, colluvium, mudflow sediments, landslide deposits, and some very poorly sorted alluvium at first glance look very much like till. However, the presence of faceted and striated stones, rock particles of distant origin, and an underlying grooved or striated rock pavement help to identify till.

Stratified drift, by contrast, is *drift that is both sorted and stratified.* It is not deposited by glacier ice, but by meltwater emanating from the ice. Stratified drift ranges from coarse, very poorly sorted sandy gravels that are transitional into till, to fine-grained, well-sorted silts and clays deposited in quiet-water environments.

Deposits of Active Ice

Moraines. In actively flowing glaciers, sediment transported by the ice is plastered onto the ground as till or is released by melting at the glacier margin where it either accumulates as a moraine or is reworked by meltwater and transported beyond the terminus. *Widespread drift with a relatively smooth surface topography consisting of gently undulating knolls and shallow, closed depressions* is known as ***ground moraine.*** Most commonly it consists of till that blankets the landscape and that may reach a thickness of 10 m or more. ***End moraines,*** on the other hand, are *ridge-like accumulations of drift deposited along the margin of a glacier.* The terminal part is a *terminal moraine,* while the lateral part is a *lateral moraine,* but both are normally part of a single continuous landform (Fig. 13.27). End moraines can form by bulldozing action of the glacier front, by slumping of loose surface debris off the glacier margin as the ice melts, by repeated plastering of drift from basal ice onto the ground, or by streams of meltwater that build up deposits of stratified drift at the glacier margin. They range in height from a few to hundreds of meters. The great height and thickness of some lateral moraines are due to repeated accretion of drift upon them during successive ice advances. Buried land surfaces, marked by soils and organic remains, can sometimes be seen exposed in lateral moraines, and provide evidence of their composite character.

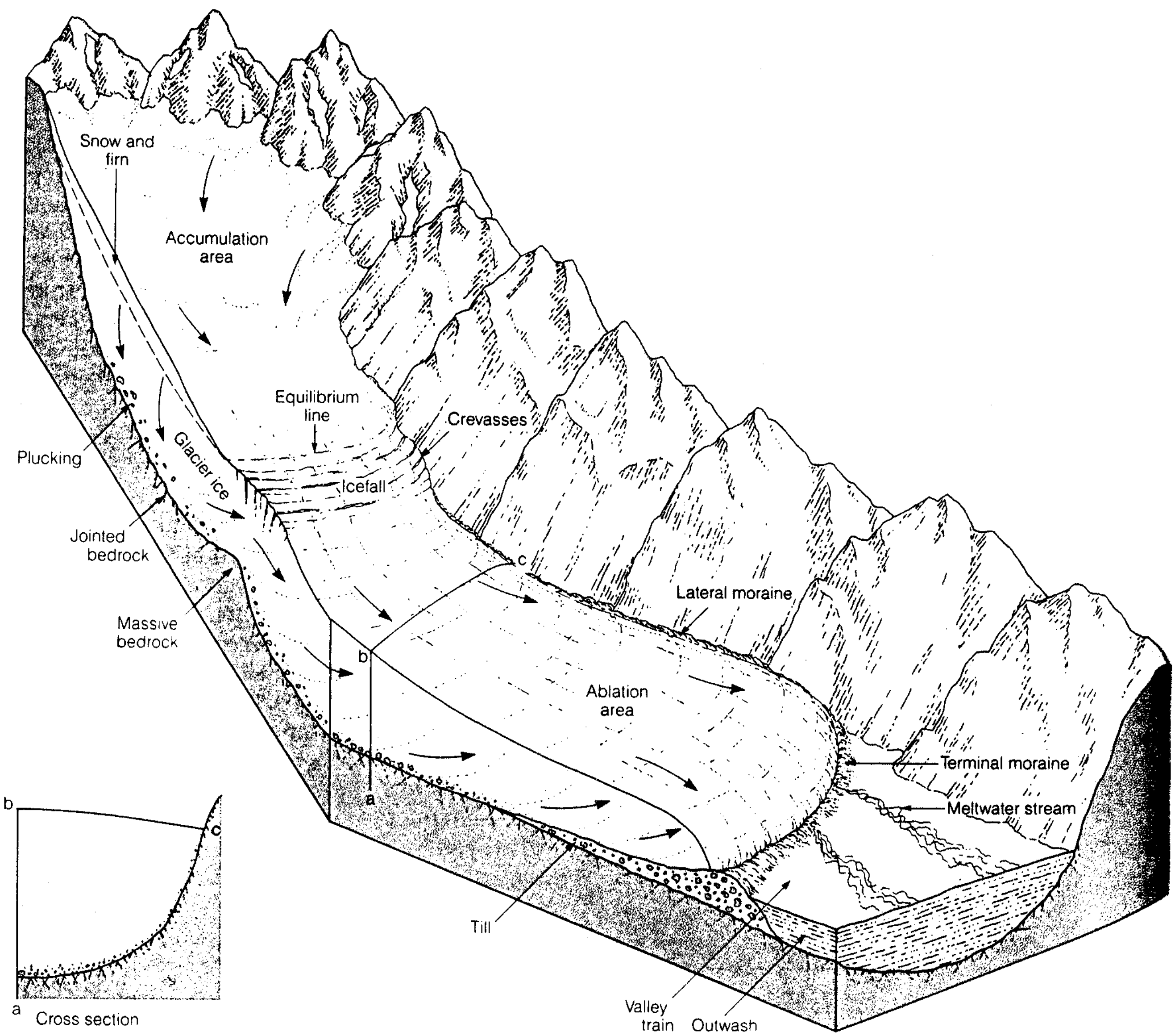


FIGURE 13.26 Main features of a valley glacier and its deposits. The glacier has been cut away along its center line; only half is shown. Crevasses form where the glacier passes over a steeper slope at its bed. Length of arrows are proportional to velocity of flow.

FIGURE 13.27 Sharp, arcuate end moraines mark the successive position of the margin of an ice-age valley glacier in the eastern Sierra Nevada, California. The glacier originated in cirque basins beneath the distant high peaks and flowed down steep tributary valleys. Several tributary ice streams joined to form a single ice tongue that terminated on gentle ground beyond the mouth of the valley.

Erratics and Boulder Trains. Some of the boulders and smaller rock fragments in till are the same kind of rock as the bedrock on which the till was deposited, but many are of other kinds, having been brought from greater distances. *A glacially deposited rock or rock fragment whose composition differs from that of the bedrock beneath it* is an ***erratic*** (Latin for "wanderer"). The presence of foreign stones on the land surface was one of the earliest recognized proofs of former glaciation. Many erratics form part of a body of drift, but others lie isolated on the ground. Some erratics are enormous, and have estimated weights of thousands of tons (Fig. 13.28). Such boulders are far larger than can be transported by an ordinary stream of water.

In areas that have been glaciated by ice sheets, erratics derived from some distinctive bedrock source are often so plentiful and easily identified that their distribution can be plotted on a map. The resulting plot may have a fanlike shape, spreading out from the area of outcrop and reflecting the diverging pattern of ice flow. Such *a group of erratics which are spread out fanwise* is a ***boulder train,*** so named in the nineteenth century when rock particles of all sizes were called boulders. In Canada, boulder trains have been used to prospect for mineral deposits in regions where the bedrock is obscured by glacial drift.

Outwash. *Stratified drift deposited by streams of meltwater* as they flow away from the glacier margin is called ***outwash*** ("washed out" beyond the ice). Such streams typically have a braided pattern because of the large sediment load they are moving, and they have sedimentary characteristics like those of many nonglacial braided streams. If the

FIGURE 13.28 Large erratic boulder of granite embedded in an end moraine of the last glaciation on Tierra del Fuego in southernmost Chile. The local bedrock is sedimentary. The nearest possible source area for the boulder lies in the high Cordillera Darwin to the south, on the opposite side of a deep fjord system.

streams are free to swing back and forth widely beyond the glacier terminus, they can build *a body of outwash that forms a broad plain*, an ***outwash plain,*** like that lying beyond the front of Vatnajökul, a large ice cap on Iceland (Fig. 13.3). By contrast, meltwater streams confined by valley walls will build a ***valley train,*** *a body of outwash that partly fills a valley.*

When a glacier retreats, the sediment load supplied to the meltwater stream is greatly reduced and the underloaded stream is therefore able to cut down into its valley train to produce *outwash terraces.* Series of terraces are common in valleys that have experienced repeated glaciations (Fig. 13.29). Generally, each major terrace can be traced upstream to an end moraine or former ice limit.

Deposits of Stagnant Ice

When rapid ablation greatly reduces ice thickness in the terminal zone of a large glacier, movement may virtually cease. Sediment carried by meltwater flowing over or beside the nearly motionless stagnant ice is deposited as stratified drift, which slumps and collapses as the supporting ice slowly melts away. This *stratified sediment deposited in contact with supporting ice* is called ***ice-contact stratified drift.*** It is recognized by abrupt changes in grain size, by distorted, offset, and irregular stratification, and by extremely uneven surface form. Bodies of ice-contact stratified drift are classified according to their shape (Fig. 13.30). Some extensive end-moraine systems of former ice sheets consist of broad belts of ***kettle-and-kame topography,*** *an extremely uneven terrain resulting from wastage of debris-mantled stagnant ice and underlain by ice-contact stratified drift* (Fig. 13.30). Many of the multitude of lakes that dot the land surface in the states of Michigan and Wisconsin occupy kettles in such terrain.

Reconstructing Former Glaciers

The landforms and sediments left by a former mountain glacier can be used to reconstruct its geometry. From the surface form, the altitude of the equilibrium line can also be estimated, thereby providing information about the climatic environment in which the glacier existed. A terminal moraine or the head of a valley train marks the downvalley limit of the glacier, while lateral moraines, erratics, and the upslope change from glacially sculptured rock surfaces to unglaciated frost-shattered slopes above permit the level of its upper surface to be approximated. The headward portions of the glacier are delimited by cirques. Measured altitudes of such features can be used to reconstruct the glacier's topography (Fig. 13.31). If one then assumes that the accumulation area occupied two thirds of the total area of the glacier, as is the case for most present glaciers in a balanced state, then the altitude of the equilibrium line can be calculated. Using such techniques, it has been possible to map the distribution of former glaciers in many parts of the world and determine the regional pattern of glacier equilibrium lines during and since the last glaciation.

FIGURE 13.29 Outwash terraces rise above meandering Cave stream on South Island, New Zealand.

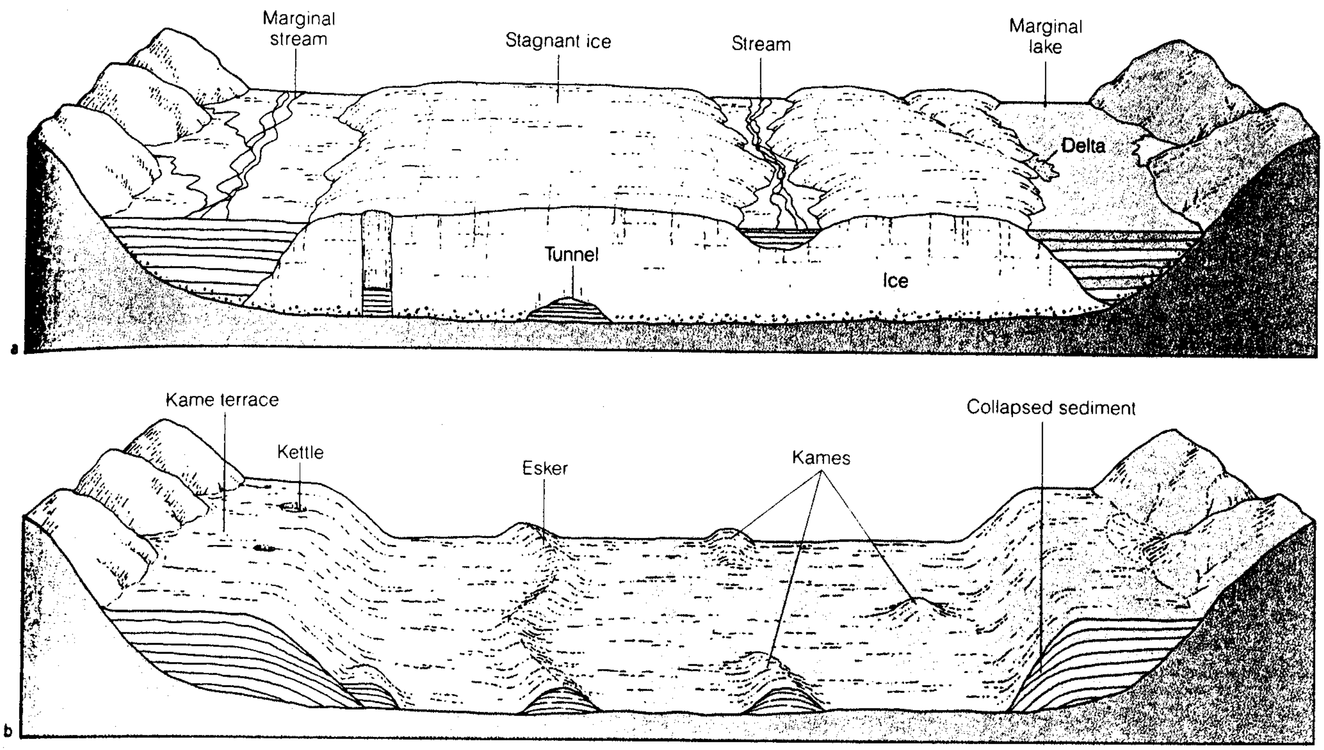


FIGURE 13.30 Origin of ice-contact stratified draft associated with stagnant-ice terrain. (*a*) Nearly motionless melting ice furnished temporary retaining walls for bodies of sediment deposited chiefly by meltwater streams and in meltwater lakes. (*b*) As ice melts, bodies of sediment slump, creating a kettle-and-kame topography. Resulting landforms include a **kame,** *a body of ice-contact stratified drift in the form of a knoll or hummock;* a **kame terrace,** *a terracelike body of ice-contact stratified drift along the side of a valley;* a **kettle,** *a basin in glacial drift, created by melting-out of a mass of underlying ice;* and an **esker,** *a long, narrow ridge, commonly sinuous, composed of stratified drift.*

Reconstruction of continental ice sheets is a more difficult exercise, for although end moraines that mark ice-sheet limits can be mapped on land, significant portions of former ice sheets terminated beyond present coastlines on the submerged continental shelves where detailed information is difficult to obtain. Furthermore, the large ice sheets covered virtually all the terrain within their margins, and little land projected above the ice surface. Consequently, within most of the region that was covered by ice there is no way to measure former ice thickness directly from geologic features. Instead, ice thickness and glacier profiles have been calculated numerically on the basis of ice-flow properties derived from laboratory and field measurements of existing glaciers; the results, however, are only approximations. Therefore, we do not yet have an accurate measure of the area, thickness, topography, or volume of former ice sheets. Nevertheless, a good deal is known about directions of ice flow, for much of the land surface over which the glaciers moved is abundantly marked with striations, grooves, drumlins, and other aligned ice-flow indicators.

PERIGLACIAL LANDSCAPES

Land areas beyond the limit of glaciers where low temperature and frost action are important factors in determining landscape characteristics are spoken of as ***periglacial zones.*** Modern periglacial conditions are most widespread in polar and subpolar regions and at high altitudes. However, during past glacial ages such zones extended far beyond their present limits and into now-temperate latitudes. Geolo-

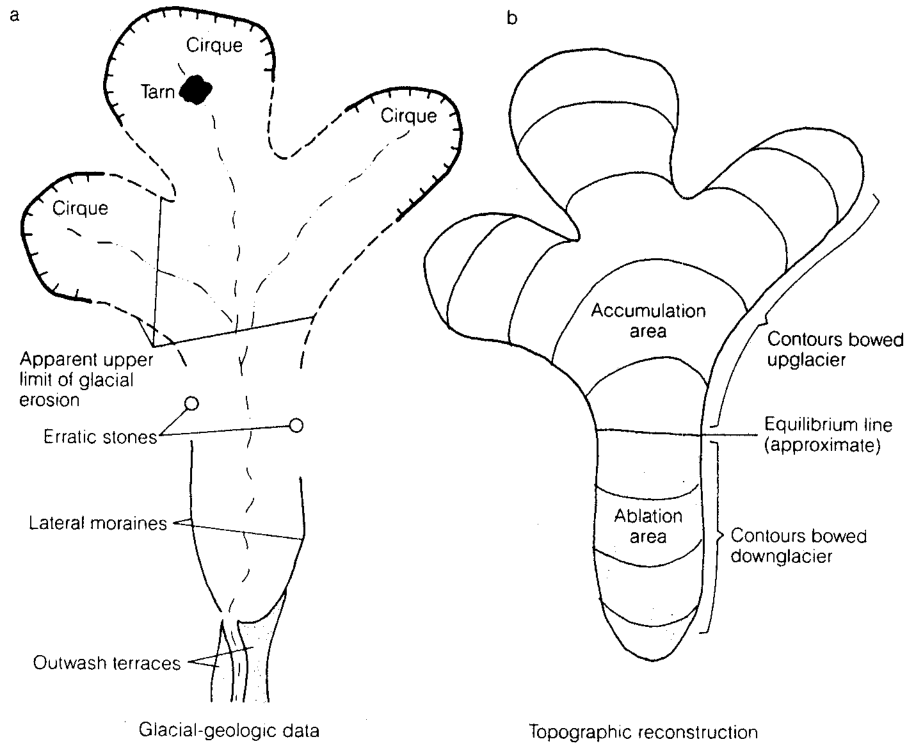


FIGURE 13.31 Method of reconstruction of a former glacier from geologic data. (*a*) Glacier-marginal features (end moraines, upper limits of erratics and ice-eroded bedrock, cirque headwalls) are used to determine the areal extent of the glacier. (*b*) Topography of the glacier is reconstructed using altitude of ice-marginal features as a basis for drawing contours. The equilibrium line is assumed to lie at the contour above which two-thirds of the total area of the glacier lies. Contours above the equilibrium line are bowed up-glacier; those below are bowed down-glacier. Contours are drawn perpendicular to the general trend of ice-flow indicators, such as striations.

gists have discovered and mapped the distribution of relict periglacial features throughout much of central Europe and across northern United States in zones that were marginal to the great Pleistocene ice sheets. Active and relict periglacial features have also been found widely in many other areas, including Alaska, Canada, Siberia, northern and western China, Patagonia, and the ice-free areas of Antarctica.

Today periglacial conditions are found over more than 25 percent of the Earth's land areas. Despite their vast areal distribution, periglacial landscapes remained little known until recently due to their remoteness and their low number of inhabitants. However, many of these regions have been receiving increasing attention because of their important energy resources and mineral wealth. Long pipeline systems in Alaska, northern Canada, and Siberia now carry newly discovered oil and natural gas across periglacial landscapes. New settlements are springing up in areas underlain by thick frozen ground. This has created an urgent need for information about periglacial processes and environmental conditions that has spurred field and laboratory research on cold-climate phenomena.

Permafrost

A common feature of periglacial regions is perennially frozen ground, generally known as ***permafrost.*** It is defined as *sediment, soil, or even bedrock that remains continually at a temperature below 0° C for an extended time* (from two years to tens of thousands of years). Such conditions exist mainly in the circumpolar zones of each hemisphere, as well as at high altitudes. The largest areas of permafrost occur in northern North America, northern Asia, and in the high, cold Qinghai-Xizang Plateau of western China (Fig. 13.32). It also has been found on many high mountain ranges, even including some lofty summits in tropical and subtropical latitudes. The southern limit of continuous permafrost in the Northern Hemisphere (as opposed to discontinuous patches) generally lies where the annual air temperature is between −5 and −10° C.

Most permafrost is believed to have originated during the last glacial age or earlier glacial ages. Remains of wooly mammoth and other extinct Pleistocene animals, which have been found well preserved in frozen ground, indicate that permafrost existed at the time of their death.

The depth to which permafrost reaches depends

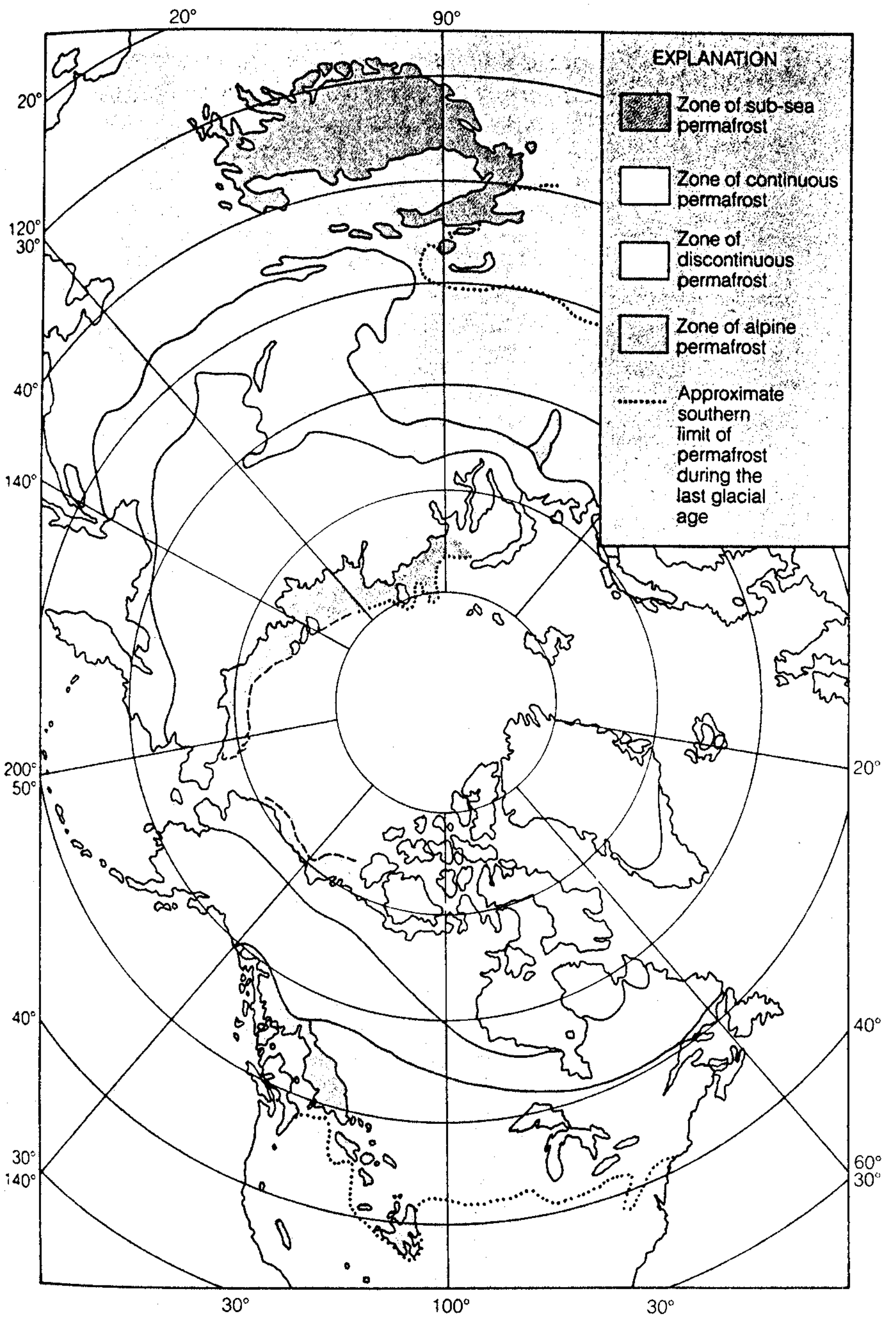


FIGURE 13.32 Distribution of permafrost in the Northern Hemisphere. Continuous permafrost lies mainly north of the 60th parallel and is most widespread in Siberia and arctic Canada. Extensive alpine permafrost underlies the high, cold plateau region of central Asia. Smaller isolated bodies occur in high mountains of the western United States and Canada. (After Péwé, 1983.)

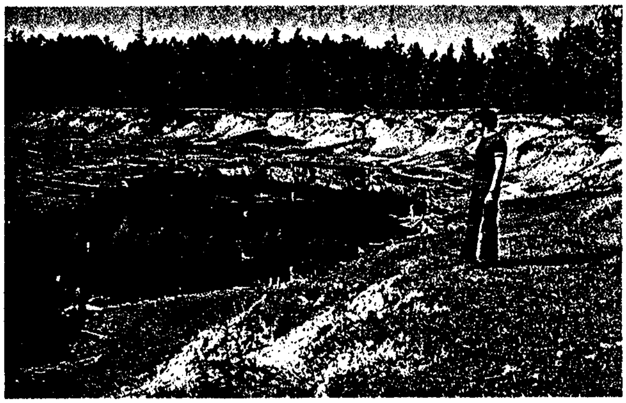


FIGURE 13.33 Thaw lake in thermokarst terrain near Yakutsk in northern Siberia where the ground is underlain by thick permafrost.

not only on the average air temperature but also on the rate at which heat flows upward from the Earth's interior and on how long the ground has remained continuously frozen. The maximum reported depth of permafrost is about 1400–1500 m in Siberia. Thicknesses of about 1000 m have been reported in the Canadian Arctic, and of at least 600 m in northern Alaska. These areas of very thick permafrost all occur in high latitudes outside the areas of former ice sheets. The ice sheets would have insulated the ground surface and, where thick enough, actually caused ground temperatures beneath them to rise to the pressure melting point. On the other hand, nonglaciated open ground unprotected from subfreezing air temperatures could have become frozen to great depths during prolonged cold periods.

A thin surface layer of ground that thaws in summer and refreezes in winter is known as the ***active layer.*** Within it the thawed ground in summer tends to become very unstable and subject to movement. The permafrost beneath, however, is capable of supporting large loads without being deformed. Many of the landscape features we associate with periglacial regions reflect movement of rock particles within the active layer during freezing and thawing.

Permafrost presents unique problems for people living on it. If a building is constructed directly on the surface, the warm temperatures developed inside when the house is heated are likely to thaw the underlying permafrost making the ground unstable. Arctic inhabitants learned long ago that they must place the floors of their buildings on pilings above the surface so that cold air can circulate freely beneath, thereby keeping the ground frozen.

Wherever tundra vegetation that forms a continuous cover over some permafrost landscapes is ruptured, melting can begin. In nature, this may lead to the collapse of the ground and formation of *thaw lakes.* The resulting topography resembles karst terrain (Chapter 10) and is known as *thermokarst* (Fig. 13.33). Thawing can also commence through human activity, and the results can be environmentally disastrous. Large wheeled or tracked vehicles crossing arctic tundra can quickly rupture it. The water-filled linear depressions that result from thawing can remain as features of the landscape for many decades.

Periglacial Landforms

The seasonal freezing and thawing action in the active layer disrupts the mineral soil and promotes differential sorting of surface sediments into varied surface patterns. *More-or-less symmetrical patterned forms due to frost action* are collectively known as ***patterned ground.*** They include such features as circles, polygons, nets, and stripes, which are descriptive terms for the resulting patterns (Fig. 13.34). While such features are common in ground underlain by permafrost, many also occur in areas that only experience seasonal frost.

Ice-wedge polygons, which are *large polygonal features that form by contraction and cracking of frozen ground,* are diagnostic of continuous permafrost. Water that enters the cracks expands as it freezes so that after many repeated cycles a typically wedge-shaped body of ice is produced (Fig. 13.35*a*). Some polygons reach diameters of 150 m (Fig. 13.35*b*) and may extend to depths of several meters. If the climate changes, causing the ice to thaw, sediment may fill the resulting cavity.

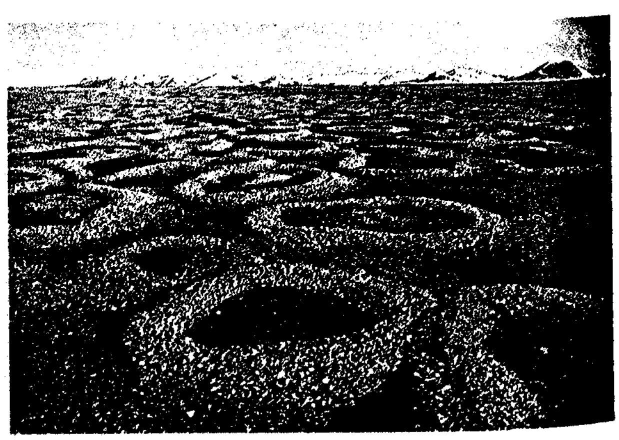


FIGURE 13.34 Sorted circles, about 4 m in diameter, form a striking pattern on the barren landscape at Brogerhalvoya in western Spitsbergen.

identification of such geologic features beyond the limits of former ice sheets in Europe and North America provides evidence of glacial-age permafrost in areas that now lie 1000 km or more from the nearest contemporary permafrost (Fig. 13.35*c*).

Other typical periglacial landforms, characteristic of at least discontinuous permafrost, include ***pingos*** which are *large, generally conical ice-cored mounds that commonly reach heights of 30–50 m and are formed from freezing of water within the permafrost* (Fig. 13.36*a*). Pingos are common features in coastal regions of arctic Alaska and Canada, and from a distance resemble small volcanoes. The crest is sometimes ruptured and collapsed due to exposure and melting of the lenslike ice core. Remains of collapsed pingos also have been found in former periglacial zones beyond the limits of Pleistocene ice sheets.

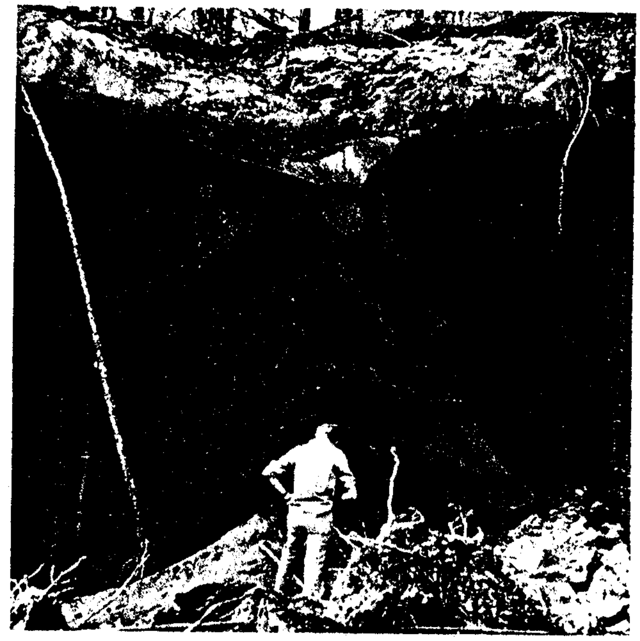
A ***rock glacier,*** another characteristic periglacial landform, is *a glacierlike tongue or lobe of angular rock debris containing interstitial ice or buried glacier ice that moves downslope in a manner similar to glaciers* (Fig. 13.36*b*). Active rock glaciers may reach thicknesses of 50 m or more and advance at rates of up to about 5 m/yr. They are especially common in high interior mountain ranges, such as the Swiss Alps, the Rocky Mountains, and the northern Andes of Argentina.

Gelifluction, *the slow downslope movement of saturated sediment associated with frost action in cold-climate regions,* produces lobes and sheets of debris that creep slowly down hillslopes (Fig. 13.36*c*). Although measured rates of movement are low, generally less than 10 cm/yr, gelifluction is so widespread on arctic landscapes that it constitutes a highly important agent of mass transport.

THE GLACIAL AGES

History of the Concept

As early as 1821 European scientists began to recognize features characteristic of glaciation in places far from any existing glaciers. They drew the then remarkable conclusion that glaciers must once have covered wide regions. Consciously or unconsciously, they were applying the principle of uniformity. The concept of a glacial age with widespread effects was first proposed in 1837 by Louis Agassiz, a Swiss scientist who achieved considerable fame through his hypothesis. Although at first many regarded the idea as outrageous, gradually, through the work of many geologists, the



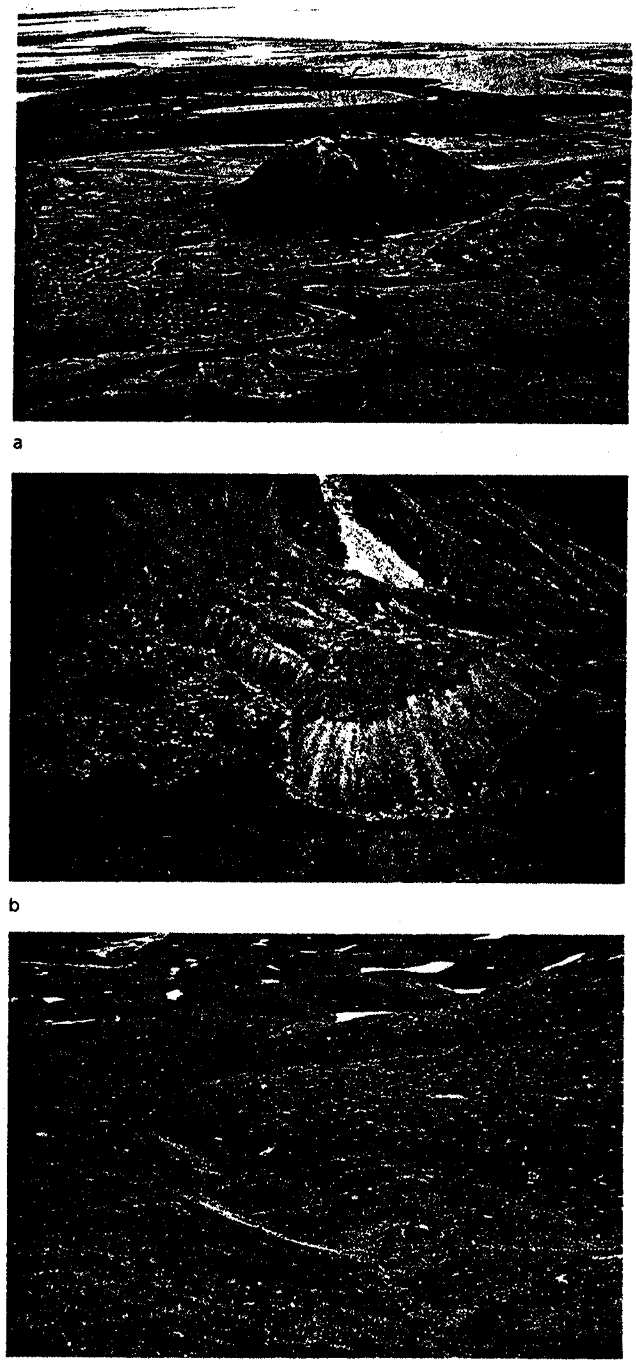
a

b

c

FIGURE 13.35 Modern and ancient ice-wedge polygons. (*a*) Active ice wedge exposed in bank along the Aldan River in northern Siberia. Growth of the wedge has caused deformation of frozen river silts as its thickness has increased. (*b*) Ice-wedges form distinctive polygonal patterned ground enclosing shallow lakes near the shore of the Arctic Ocean. (*c*) Polygonal ice-wedge casts exposed during road construction near Rawlings, Wyoming.

concept gained widespread acceptance. Today we have a basic understanding of the nature of the glacial ages, although important questions remain unanswered.



a

b

c

FIGURE 13.36 Common periglacial features. (*a*) Ibyuk Pingo rises steeply above surface of Mackenzie Delta in northwestern Canada. (*b*) Active rock glacier, fed by rockfall from steep slopes above, advances across valley margin in the Alaska Range. (*c*) Gelifluction sheets creeping slowly downslope on mountain (to right) In Italian Alps have overridden moraines on valley floor.

Extent of Former Glaciers

During the second half of the nineteenth century geologists mapped the distribution of glacial drift and other characteristic features in order to determine the extent of former glaciers. Their studies focused on the regions of North America and Europe that were most accessible and which became the "classic" regions of glacial-geologic studies. Only after the middle of the present century were extensive surveys of more remote lands undertaken. Despite uncertainty regarding the local extent of ice in many areas, the global distribution of former glaciers is reasonably well known (Fig. 13.37). On a global scale, the areas of former glaciation add up to an impressive total of more than 44 million km^2, or about 29 percent of the entire land area of the Earth. Today, for comparison, only about 10 percent of the world's land area is covered with glacier ice. Of this area, 84 percent lies in the Antarctic.

Directions of Flow

In most mountainous regions, former glaciers flowed down existing valleys which determined their direction of movement. The larger ice tongues terminated in adjacent lowlands or spread out to form lobate ice fronts similar in form to the modern Bering and Malaspina glaciers of Alaska. In glaciated continental interiors, streamlined landforms and small-scale abrasion features show that the great ice sheets spread outward from their source regions in a radial pattern. Flow directions are determined also by tracing erratics of conspicuous rock types to their places of origin (Fig. 5.30). For example, native copper found as far south as Missouri has been traced to rock outcrops on the south shore of Lake Superior.

Lowering of Sea Level

Whenever large glaciers formed on the land, the moisture needed to produce and sustain them was derived primarily from the oceans. As a result, sea level was lowered in proportion to the volume of ice on land. During each glacial age, world sea level probably fell by 100 m or more (Fig. 13.38), thereby causing large expanses of the shallow continental shelves to emerge as dry land. At such times the Atlantic coast of the United States between New York and Florida lay about 150 km east of its present position. Based on fossils dredged from the seafloor on the continental shelf, the emergent coastal plain was forested with spruce and pine and its animal population included mammoths, mastodons, and other now-extinct mammals. At the same time, lowering of sea level joined Britain to France where the English Channel now lies, and North America and Asia formed a continuous landmass across what is now the Bering Strait. These and other land connections allowed plants and animals, including humans, to pass freely between land areas that now are separated by ocean waters.

Depression of the Crust

The weight of the massive ice sheets caused the crust of the Earth to subside beneath them, an effect described further in Chapter 16. The contrast between the density of crustal rocks (about 2.7 g/cm^3) and glacier ice (about 0.9 g/cm^3) means that an ice sheet 3 km thick might cause the crust to subside by as much as 1 km. The Hudson Bay region of central Canada, which formerly lay beneath the central part of an ice sheet at least 3 km thick, is still rising in response to removal of the ice load (Fig. 13.39). As a result, Hudson Bay is becoming progressively shallower and its area is diminishing in size as the water is slowly decanted. A similar change is affecting the floor of the Baltic Sea and surrounding lands. This region lies near the center of the former Scandinavian Ice Sheet and continues to rise as the crust adjusts to deglaciation.

FIGURE 13.37 Areas of the Northern Hemisphere that were covered by glaciers during the last glacial age. Arrows show general directions of ice flow. Coastlines are shown as they were at that time, when world sea level was about 100 m lower than present. Sea ice is shown covering the Arctic Ocean and extended south into the North Atlantic. Some scientists postulate that thick ice shelves, rather than sea ice, covered these portions of the ocean. The extent of former glacier ice over the Barents and Kara seas, as well as in parts of northern North America, is controversial.

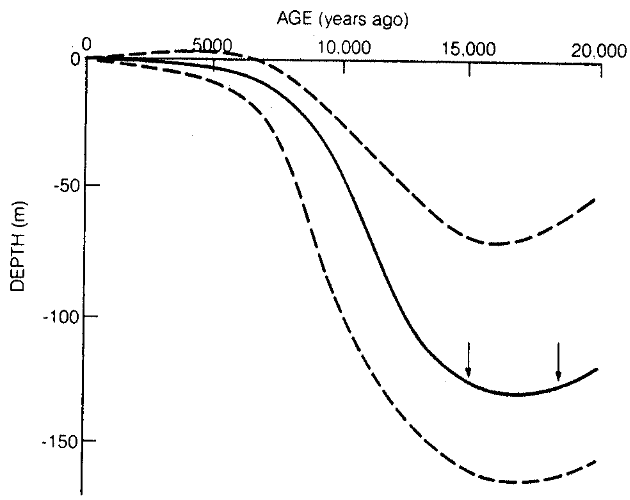


FIGURE 13.38 Curve showing estimates of the mean position of world sea level over the last 20,000 years. Estimates for maximum lowering range from about 75 to 165 m, reflecting considerable uncertainty. The time of lowest sea level (arrows) is also uncertain. The upper part of the curve, spanning the last 10,000 years, is better known. Dashed lines mark the limits within which most data lie.

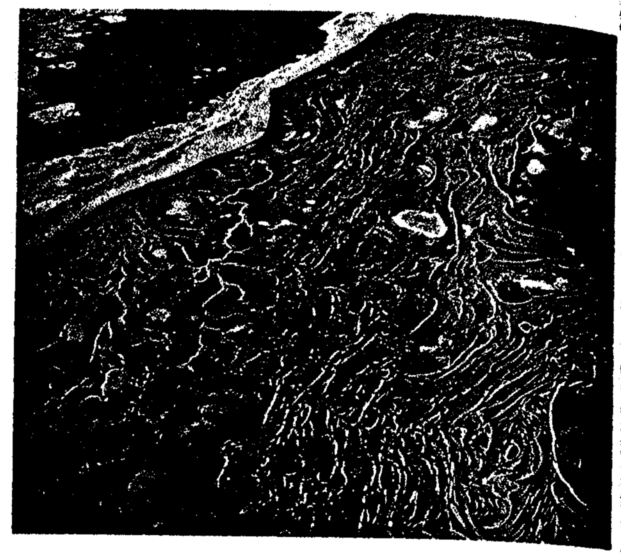


FIGURE 13.39 Emerged beaches on the southwest shore of Hudson Bay, Canada. They were formed shortly after the continental ice sheet disappeared from this region and were progressively uplifted as the land rose isostatically in postglacial time. The highest beaches in this region lie 250–300 m above the present level of the bay (upper left corner).

Repeated Glaciation

Most of the glacial drift seen at the land surface is fresh and little weathered. From that observation it was realized very early that the glaciation must be geologically recent. However, in the middle nineteenth century geologists began to find exposures showing that the layer of comparatively fresh surface drift overlies another layer whose upper part is chemically weathered. This led to the realization that there had been two glaciations separated by a long enough interval of time to produce surface weathering to a depth of a meter or more. Before the beginning of the twentieth century, evidence had been found of not merely two but three or more glaciations, during each of which ice covered approximately the same geographic areas.

Radiometric dating indicates that the latest glaciation culminated between about 20,000 and 14,000 years ago. Older drifts are not as well dated, but the earliest evidence of middle-latitude ice-sheet glaciations reaches back several million years. The time that embraced the majority of these glaciations is called the Pleistocene Epoch (Table 7.3). Although evidence now shows that some high-latitude glaciers developed even earlier during the Cenozoic Period, the Pleistocene was an interval when glaciers were especially widespread on the Earth (Chapter 20).

Evidence of still older glacial ages appears repeatedly in the rock record throughout at least the last half of Earth history. The evidence includes occurrences of ***tillite,*** *till that has been converted to solid rock,* in strata of several different ages (Fig. 5.12). Such evidence tells us that the Earth's climate has fluctuated repeatedly, causing glaciers to form and later disappear and creating important environmental changes throughout the world.

SUMMARY

1. Glaciers consist of ice which has been transformed from snow by compaction, recrystallization, and flow. They form part of the cryosphere, the part of the planet where water exists primarily in the frozen state.
2. The major types of glaciers, based on their ge-

ometry, are cirque glaciers, valley glaciers, ice caps, ice sheets, and ice shelves.

3. Ice in temperate glaciers is at the pressure melting point and water exists at the glacier bed. Polar glaciers consist of ice that is below the pressure melting point and are frozen to their bed. In subpolar glaciers, a thin surface zone reaches the melting point in summer, but the ice beneath is below freezing.
4. Glaciers depend for their survival on low temperature and adequate precipitation. They bear a close relationship to the snowline which is low in polar regions and rises to high altitudes in the tropics.
5. The mass balance of a glacier is measured in terms of accumulation and ablation. The equilibrium line separates the accumulation area from the ablation area and marks the level on the glacier where net gain is balanced by net loss.
6. Glaciers flow under their own weight. Their surface zone is brittle; however, at greater depth internal flow occurs.
7. The motion of temperate glaciers includes both internal flow and sliding along the bed. In polar glaciers, which are frozen to their bed, motion is much slower and involves only internal flow.
8. Surges involve extremely rapid flow, probably related to excess water at the glacier bed. Frontal calving can lead to rapid recession of glacier margins that recede into deep water.
9. Glaciers erode rock by quarrying and abrasion. They transport the waste and deposit it as drift.
10. Mountain glaciers convert stream valleys into U-shaped troughs with hanging tributaries and with cirques at their heads. Mountain areas that project above glaciers are shaped by frost action into angular landforms that contrast with glacially smoothed terrain downslope.
11. Fjords are excavated far below sea level by thick, fast-flowing ice streams in coastal regions.
12. Flow directions of ice sheets are inferred from striations, grooves, and drumlins.
13. The load, carried chiefly in the base and sides of a glacier, includes rock fragments of all sizes, from fine rock flour to large boulders.
14. Till is deposited directly by glaciers, while glacial marine drift is deposited on the seafloor from floating glacier ice. Stratified drift is deposited by meltwater. It includes outwash deposited as outwash plains or valley trains beyond the ice margin, and ice-contact stratified drift deposited upon or against stagnant ice.
15. Ground moraine is built up beneath a glacier, whereas end moraines (both terminal and lateral) form at the glacier margins.
16. Permafrost is widespread in periglacial regions. Where continuous, it may reach a thickness of 1000 m or more, but toward its limit it becomes patchy and discontinuous. In summer, a thin active layer at the surface thaws and may become unstable.
17. Patterned ground results from repeated freezing and thawing of the ground. Distinctive relict periglacial features permit mapping the extent of former permafrost conditions.
18. During glacial ages huge ice sheets repeatedly covered northern North America and Europe, eroding bedrock and spreading drift over the outer parts of the glaciated regions. As the ice sheets grew, world sea level fell and the crustal rocks beneath them subsided due to the added weight.

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