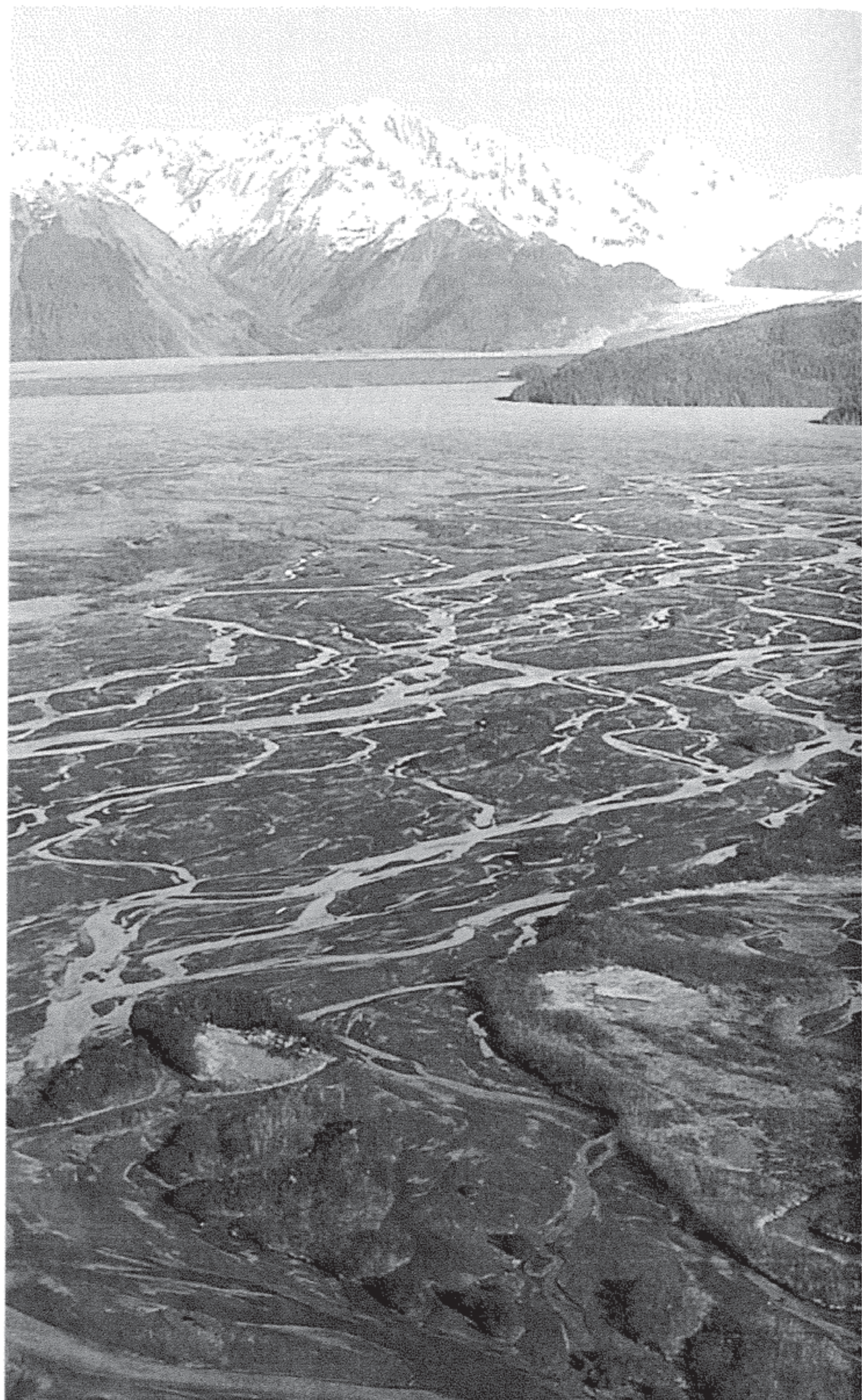
CHAPTER

10

# Glacial Erosion, Deposition, and Landforms





## Introduction

We are now ready to consider the geomorphic significance of glacier mechanics. It would be ideal if we could directly correlate geomorphic features with specific ice processes, but we seem to be far from such sophistication. The reasons for this are several: (1) the conceptual models of glacier mechanics are greatly oversimplified and based on theory rather than extensive observation; (2) much erosion and deposition takes place at the base of glaciers where it is rarely feasible to document the actual process; and (3) we are just beginning to understand the feedback mechanics that function between the basal ice and the geologic framework. For example, we know that glaciers have the ability to erode because we can see the results of that erosion after the ice disappears. What we do not understand is precisely how the erosive process itself is modified by remolding of the subglacial framework. For example, does removal of certain obstructions by erosion decrease the subsurface roughness enough to facilitate rapid sliding velocities, as suggested by Kamb (1970), and might this in turn accelerate further erosion?

Until we can resolve these deficiencies in our understanding, studying the process-feature relationship is a bit like traveling a one-way street in the wrong direction—we must interpret the process from the character of its results. Nonetheless, glaciers have formed a variety of deposits and a myriad of features that demand attention. It is not completely satisfying to reconstruct their origin without a wealth of solid observations, but this type of deductive reasoning has been a part of geology for a long time and is not necessarily incorrect. Furthermore, investigations of features and deposits can provide tangible clues about their origins that become invaluable in deciding how and where to study glaciers in the future. It seems inevitable, however, that new techniques for investigating glacier mechanics will demand continuous modification and testing of our present interpretations of features and deposits.

## Erosional Processes and Features

### Minor Subglacial Features

Glacial erosion is accomplished primarily by two processes, a scraping action called **abrasion** and a dislodgement or lifting action called **quarrying** or **plucking.** Because ice is not a hard mineral (1.5 on Mohs' scale at 0° C), it cannot abrade most solid rock material unless it utilizes fragments of rock carried as load in the ice as grinding tools. Therefore, the efficiency of abrasion and the features it produces depend on the character and concentration of the debris being dragged along the base of the ice and, of course, on the properties of the bedrock being overridden. Various models have been proposed to describe the rate of glacial abrasion (e.g., Boulton 1974; Hallet 1981; Shoemaker 1988) and most utilize the wear law expressed by

$$A_v = kF_nV_p$$

where $A_v$ is the volume of rock removed from a single fragment per unit time, $F_n$ is the effective contact force exerted normal to the bed surface by the fragment, $V_p$ is the velocity of the fragment parallel to the bed, and $k$ is a coefficient dependent on shape of the fragment and the relative hardness between the fragment and bed material. The parameters cited as governing the effective contact force (the force applied by the particle onto the bedrock) represent a primary difference between the various abrasion models. For example, Boulton (1974) suggests that the contact force is dependent on the weight of the overlying ice which presses the clast to the bed. However, glacial ice greater than 22 m thick should flow around and under the abrading rock fragments, and thus the contact force should be independent of ice overburden pressure. This assumption led Hallet (1981) to propose a model in which the contact forces are derived from viscous drag exerted on the fragments by ice flowing toward the bed. Vertical ice motion results from extending flow on the stoss side of bed irregularities and from the replacement of basal ice melted by frictional or geothermal heat.

The rate of abrasion, estimated by various methods, usually ranges from 0.06 to 5 mm a year, but it may be considerably higher under thick, high-velocity glaciers. Boulton (1974) reports that the abrasion of a marble plate inserted beneath the Glacière d'Argentiere in France proceeded at a rate up to 36 mm a year where the ice was 100 m thick and moving at 250 m a year.

Small erosional features produced by abrasion are usually in the form of linear scratches or crescentic marks that show the relationship between the size and composition of the abrading particles and the resistance of the underlying bed. Very fine particles in sufficient abundance produce a smoothly polished surface composed of microscopic scratches (fig. 10.1). As the grain size of the load increases, the scratches become larger in a transitional sequence from polish to striations to grooves and furrows.

**Striations** like those shown in figure 10.1 have been noted in every glaciated region of the world. They are best preserved on fine-grained rocks that have not been deeply weathered and on smooth bedding planes that dip gently away from the direction of ice movement. Striations and larger linear features can also form in unconsolidated material, such as till or loess, if it is highly compacted (Westgate 1968). Striations are only millimeters deep and are most likely eroded by sand grains or by jutting edges of larger particles carried in the basal ice. They tend to be continuous for only relatively short distances, probably because the sliding clast itself produces a carpet of plowed debris at its forward edge. When sufficient debris accumulates, the scratching particle will ride over the material, thereby interrupting its contact with the solid bedrock until it reaches a fresh surface downstream from the debris cover (Boulton 1974). This effect, however, can be eliminated if circulating subglacial water removes the fines (Vivian 1970). The discontinuous nature of striae may also be related to the propensity of abrading fragments to rotate; the more rapid the rotation, the less likely the fragments will create deep and continuous striations (Iverson 1991a). It is also known that the entire scratching phenomenon becomes ineffective unless particles within the ice mass continuously move downward to replace the original abrasive grains. Those grains become smoothed during the abrasive action, and failure to replenish the basal ice with new particles having sharp edges and corners will cause abrasion to end (Boulton 1979).

Increasing grain size in the load produces grooves and furrows larger than striations. Normally grooves are up to 1 to 2 m deep and 50 to 110 m long, but under the proper controlling factors they may achieve giant proportions. For example, H.T.U. Smith (1948) described grooves in the MacKenzie River valley of Canada that are 30 m deep, 100 m wide, and several kilometers long. These are not the product of a single boulder but possibly represent gouging by a pocket of boulders solidly frozen together (Embleton and King 1975a). There may, in fact, be a limit to the size of boulder that can act in the grooving process, because large fragments, having too much surface contact with the underlying rock, will force the ice to flow over and around the boulder instead of carrying it as part of the basal load.

In addition to scratches of all types, a group of small features, generally referred to as *friction cracks* or *chattermarks,* are formed by chipping or grinding of the underlying rock surface (Harris 1943). Most workers believe these features result when ice flow is temporarily retarded in its forward motion and then is suddenly released. This produces a jerky flow component commonly referred to as slip-stick movement. The various cracks and marks are usually lunate in form, 10 to 12 cm long and 10 to 25 mm deep, and perpendicular to the direction of ice flow as determined by other criteria. Chattermarks are sometimes present on surfaces of minerals within glacial deposits (Folk 1975; Gravenor 1982). These are indicative of the grinding action within a glacier, although chemical etching may produce similar features.

The features of abrasion are thought to reflect the direction of ice movement. Remember, however, that other processes such as mudflows or snowslides can form striations, and even floating ice blocks can cause them in nonresistant materials (Dionne 1974). Ice will diverge and converge over an irregular bedrock topography, and basal ice may be moving in different directions than surficial ice (Engelhardt et al. 1978). Thus, minor abrasion features have somewhat limited value as indicators unless a large number of measurements are obtained and treated for statistical significance.

Quarrying differs from abrasion in that the functional success of the process depends less on the type of load being transported than on the properties of the underlying rock. In fact, fractures must exist in the bedrock if plucking is to operate at all. Intense shattering of rock in preparation for plucking probably requires some form of pressure release (Lewis 1954; Glen and Lewis 1961), crushing (Boulton 1974), freezing and thawing, or cyclic water-pressure fluctuations (Iverson 1991b). These processes, which weaken the internal cohesion of the bedrock, may occur subglacially (Sugden and John 1976; Anderson et al. 1982) or, with the exception of water-pressure fluctuations, in association with periglacial conditions prior to the arrival of the glacier.

Plucking has two basic requirements. First, the ice must exert a shear force on the loosened particles, and second, this force must exceed the resistance caused by friction when the particle is dragged over the residual

(A)

**FIGURE 10.1**

(A) Glacial polish on columnar basalts, Mammoth Lakes, California.
(B) Striations on outcrop of limestone in south central New York state.

(B)

bedrock (Boulton 1974). Boulton suggests that where ice flows over unconsolidated sediments, plucking is a relatively straightforward process related totally to shear stress.

In tightly lithified materials the mechanics are more complex. In some cases, the plucking process may be related to the cyclic opening and closing of cavities beneath the ice caused by obstructions to flow and fluctuations of ice thickness or velocity. When cavities are open, free water may freeze to the fragmented particles within the shattered bedrock mass. If the cavity is subsequently closed, the glacier incorporates the ice frozen to the shattered particles into its basal layer and the forward motion of the glacier plucks some rock fragments away from the surface. It is important to recognize, however, that frictional resistance to plucking increases rapidly as the cavity is closed because it is directly proportional to the normal pressure exerted by the overlying ice. As Boulton suggests, plucking is probably most effective when the normal pressure is sufficient to incorporate the loose particles into the basal ice but not great enough to inhibit their forward movement by increasing the frictional resistance. Exactly when this condition prevails is not well understood and probably varies from glacier to glacier. In fact, Iverson (1991b) suggests that the closure of subglacial cavities is not required for plucking to occur; instead, he suggests that the quarrying process is caused by water-pressure fluctuations at the base of the glacier. Based on theoretical and numerical studies, he found that rapid water-pressure drops in cavities located downstream of a bedrock obstruction transfer some of the weight of the ice from the subglacial water to the bedrock (fig. 10.2). In addition, the lateral support on the lee side of the bed obstruction by the water in the cavity is reduced (Iverson 1991b). The result is a change in the principal stresses on the bedrock and the growth of preexisting cracks, producing a shattered bedrock zone. In this model, decreases in water pressure lead to rock shattering, while increases presumably promote plucking because high water pressures may (1) result in pressure-release freezing between the ice and the glacier bed and enhance drag on the subglacial materials, (2) decrease the normal stress on the glacier bed, reducing frictional resistance, and (3) accelerate basal sliding, thereby increasing the shear stress on the shattered bedrock surfaces (Iverson 1991b). Iverson's model is intriguing in that it provides a unified mechanism for both the shattering and plucking of rock from the glacial bed.

The evidence of plucking action in glaciated landscapes is usually found in erosional features somewhat larger than those produced by abrasion, but commonly the two erosive processes are closely associated in the same landforms. Where abrasion is dominant, the landscape may be indented with smoothly curved elongate surfaces whose long axes are subparallel to the direction of ice flow. Some of these surfaces are distinctly higher at one end, and they taper laterally and longitudinally until they blend into the surrounding ground level, producing a unique teardrop or raindrop shape which Flint (1971) describes as a "whaleback form." Some whaleback forms may be related to streamlined depositional features, such as drumlins (discussed later in this chapter), in that their shape represents the minimum resistance to flowing ice. The composition of streamlined forms seems to be of little consequence, however, as they can be entirely bedrock, entirely sediment, or a combination of the two. Whalebacks, therefore, may be merely a transitional form in a range of streamlined features from pure bedrock to pure drift (Flint 1971).

When plucking is a significant factor, whalebacks develop a pronounced asymmetry, having a gently sloping upstream surface and a steep rock face on the down-ice side of the feature. Such a form, commonly called *roche moutonnée,* is the result of abrasion on the upstream slope and intense quarrying at the position of the steep, downstream face. The development of this form is due to irregular spacings of fractures within the bedrock. Where joints are widely spaced, abrasion is the dominant process; closely spaced jointing, on the other hand, facilitates plucking and more rapid erosion (fig. 10.3). Flint (1971) objects to the term "roche moutonnée" because of its wide misuse and suggests that *stoss and lee topography* better describes the forms developed by the combination of abrasion and plucking.

Subglacial erosion is much more complex than our simple analysis of abrasion and plucking suggests. Many subglacial processes are involved in fracturing of the underlying bedrock (Gray 1982; Drewry 1986), and these directly affect the intensity of plucking and the character of ice-scoured topography (Gordon 1981). Flow of basal ice around obstacles often reaches extreme velocities. This fast flow, known as *ice streaming,* produces large, elongate erosional forms (Boulton 1979; Goldthwait 1979). Water may also be involved in subglacial erosion. For example, in carbonate regions subglacial meltwater may differentially dissolve the bedrock surface (Hallet 1976a, 1976b). In addition, the

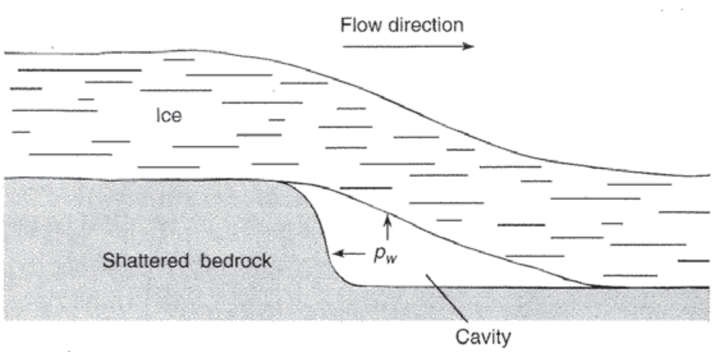



**FIGURE 10.2**
Schematic diagram of a subglacial cavity formed downstream of a bedrock obstruction. Localized bedrock shattering may be produced by a rapid decrease in water pressure ($p_w$) within the cavity which transfers some of the weight of the ice from the subglacial water to the bedrock and removes the lateral support of the water on the lee side of the obstruction. Subsequently, plucking may occur as subglacial water pressure increases.

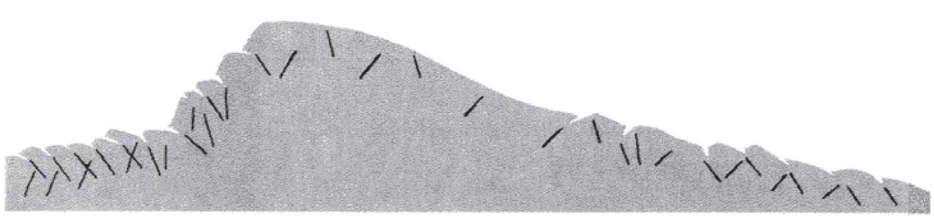


**FIGURE 10.3**
Relationship between joint spacing and roche moutonnée development in Yosemite Valley.
(After Matthes 1930)

catastrophic release of subglacial meltwaters can erode the bedrock into a myriad of forms (Dahl 1965; Rains et al. 1993; Sharpe and Shaw 1989), as can the fluid mobilization of subglacial till (Gjessing 1967; Gray 1982).

## Cirques

The striking landscapes found in the uplands of glaciated mountains have been sculpted primarily by the erosive action of ice contained in cirques. The term "cirque" was first used in the early 1800s to describe the collecting basins for valley glaciers in the Pyrenees, and locally the feature has been given a variety of names including *cwm, corrie, kar,* and *botn.* A **cirque** by any name is still a deep erosional recess with steep and shattered walls that is usually located at the head of a mountain valley. It is normally semicircular in plan view, often being described as an amphitheater, and it is floored by a distinct rock basin where the surface has been smoothed by abrasion. As figure 10.4 shows, the bowl-shaped rock basins commonly contain lakes, called *tarns,* that are dammed by a convex-up rock lip

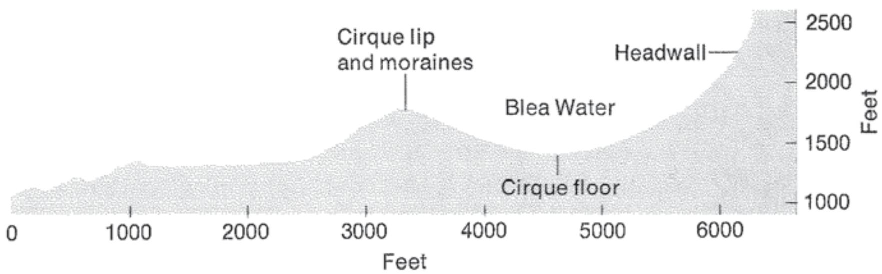



**FIGURE 10.4**
Long profile through Blea Water corrie, a cirque.
(Lewis 1960)

**FIGURE 10.5**
Cirques developed in the Uinta Mountains, Utah.

that stands as a threshold boundary between the cirque floor and the downstream part of the valley. The cirque lip is often capped by small moraines that contribute to the damming effect. The rock basins can be of spectacular dimensions. For example, the rock basin floor of Blea Water corrie in England is 96 m lower than the rock lip (Lewis 1960).

Cirques range in size from shallow depressions to monstrous cavities that are kilometers wide and several thousand meters high along the rear wall. Their dimensions and geomorphic form depend not only on the type of rocks into which they are cut, being larger and more perfectly developed in igneous or high-rank metamorphic rocks, but also on the rock structures (Olyphant 1981), the preglacial relief, and the time span of the formative glaciation. Most maturely developed cirques seem to possess a reasonably consistent geometry when length to height ratios are compared, indicating that cirques probably attain some equilibrium form related to the processes of formation.

Cirques are often preferentially oriented according to the direction of solar radiation and the prevailing winds (Graf 1976), and their elevation is probably (but not necessarily) related to the snowline at the time of their formation (Porter 1977; Trenhaile 1977). Thus, although most cirques originate in the headward reaches of stream valleys, any hollow, regardless of its origin, that stands at the proper elevation and has the ideal orientation may progressively accumulate snow and finally become a maturely developed cirque like those in figure 10.5.

The significance of cirque processes in the development of Alpine scenery is that cirque expansion by continued erosion gradually eliminates the preglacial upland surface. As a number of cirques grow headwardly and laterally, they progressively consume much of the intervening upland region and leave as its only vestiges spectacular **horns** and **arêtes,** the features so indicative of mountain glaciation (fig.10.6). With prolonged headward erosion, adjacent cirques may merge, forming *col* depressions in the knife-edged arêtes. The erosive work in cirques suggests that Alpine topography can be classified according to stages in its development, and such hypotheses are not uncommon (see Embleton and King 1975a). Our purpose here, however, is to understand the processes that create the Alpine topography, not the sequence of its development. Therefore, we must examine the origin of cirques and the erosive mechanics that function within their boundaries. Of utmost importance are the processes that scour the basin in the cirque floor and those that cause the recession of the cirque wall.

Cirques result from two separate groups of processes: (1) mechanical weathering and mass wasting, and (2) erosion by cirque glaciers. The development of a cirque (fig. 10.7) begins with a patch of firn that fills a small depression and stands near the regional snowline. In the ablation season, meltwater released during the day percolates into fractures of the bedrock beneath the firn bank and refreezes there at night. The repeated pressure associated with freezing and thawing presumably wedges out particles of rock that are moved slowly downslope by creep and by water flowing at the base of the firn. The combined processes are commonly referred to as nivation (Matthes 1900). Actually, *nivation* refers to a set of geomorphic processes, including chemical weathering (Thorn 1976), each of which may function more effectively under different controlling factors (Thorn and Hall 1980). Nonetheless, the shape of the original depression is gradually deepened and widened, and eventually it approaches a semicircular form that can logically be called a nivation cirque. The nivation hypothesis also contends that with continued accumulation, the firn changes into true glacier ice, and thereafter erosion by cirque glaciers rather than nivation becomes the dominant process in the development of the cirque. Exactly when this transition occurs is not clear, but it is virtually impossible for rock basins to be formed by physical nivation processes alone because they cannot carry particles upslope to the cirque lip.

**Erosion by Cirque Glaciers** Observations made in tunnels excavated into cirque glaciers indicate that such glaciers move by a process known as rotational sliding, in which the ice slides over the arcuate bedrock floor, rotating at the same time around a horizontal axis. The ice exposed in the tunnels displays recognizable yearly accumulation layers that are separated by marked ablation surfaces, giving the entire glacier a banded stratigraphy (Grove 1960). Some of the ablation zones are laden with debris that fell onto the ice surface near the cirque headwall. As figure 10.8 shows, these layers originally dip downglacier at the angle of the ice surface, and with time each layer is incorporated into the ice mass as more snow accumulates in the headwall area. As the ice moves, however, the layers are reoriented so that near the equilibrium line they are almost horizontal, while close to the terminus they dip steeply upglacier. The deformation of the layers, combined with changes in the position of stakes in the tunnels and on the surface (McCall 1952, 1960), makes it clear that flowlines are moving downward

**FIGURE 10.6**
Schwan Glacier in the Chugach Mountains of Alaska. Note dark lobes on glacier surface where rockslide avalanches have moved onto the ice. Medial moraines are displayed as long linear dark bands. Horns and arêtes are shown in mountain uplands.

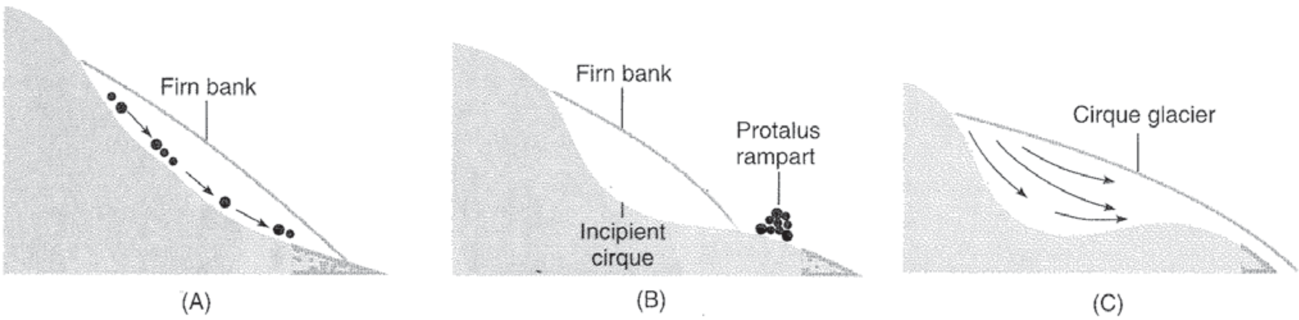



**FIGURE 10.7**
Stages of cirque development. (A) Nivation beneath firn bank. (B) Nivation cirque. (C) Cirque with fully developed cirque glacier.

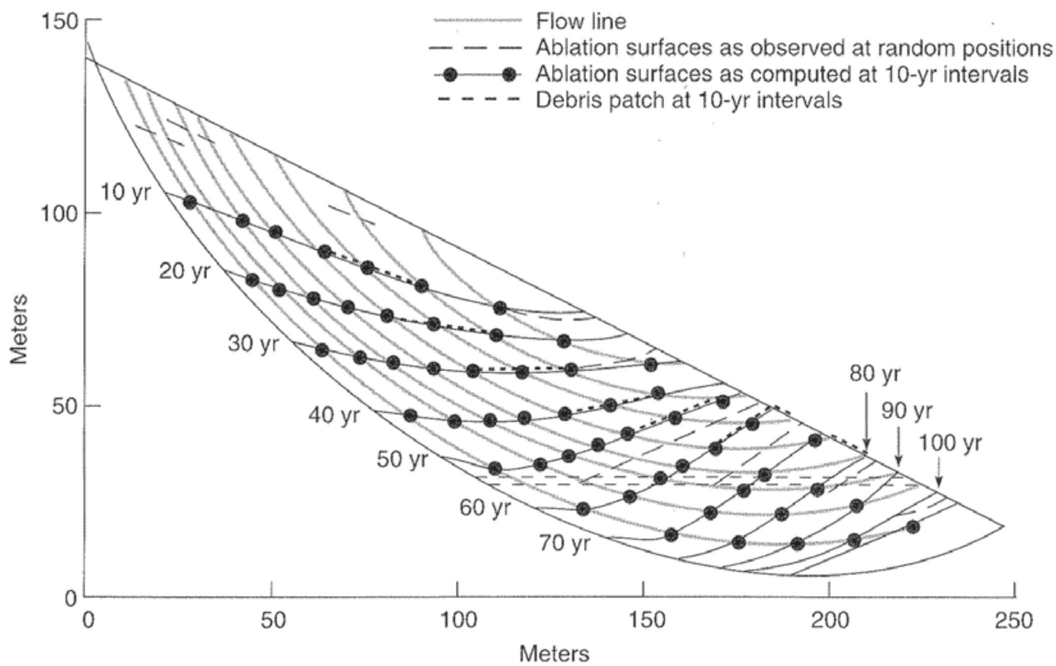



**FIGURE 10.8**
Long section through a cirque glacier in Norway showing ablation surfaces and debris patches at 10-year intervals. Note rotation of ablation surfaces in the down-ice direction.

near the headwall, parallel to the surface at the firn line, and upward near the terminus.

The mechanism of rotational sliding offers an appealing explanation for the scouring of the bowl-shaped depression in the cirque floor, for this process should be capable of carrying the products of abrasion or frost wedging upslope and over the cirque lip. However, not all cirque glaciers exhibit rotational sliding as the dominant flow mechanism.

The rate of cirque erosion and growth seems to vary with the geologic and climatic setting. In polar and subpolar regions, erosion rates have been estimated as 8–76 mm/1000 years (Anderson 1978), and plucking seems to be the dominant erosional process. In temperate settings where abrasion is more important, the rate has been estimated as 95–165 mm/1000 years but could have been much greater during full glacial conditions (Reheis 1975). Actually, the rate of cirque growth produced by erosion is difficult to ascertain because the time span over which erosion acted is not precisely known, and the influence of preglacial topography is not usually considered (Olyphant 1981b). Regardless, because the length to depth ratio of cirques in widespread locations is normally 2:1, headwall retreat probably occurs at a faster rate than deepening of the cirque floor (Gordon 1977; Olyphant 1981a).

**Headwall Erosion** The creation of Alpine topography requires removal of large amounts of bedrock from the head and side walls of the cirque floor. The surfaces of the cirque walls are assumed to be prepared for erosion by severe frost shattering of the wall rock. In addition, joints often develop parallel to the wall faces when pressure is released by the removal of outer rock layers (Glen and Lewis 1961). These *dilatation joints* aid in the fracturing process by providing avenues for percolating water and by isolating rock material into discrete units.

The actual processes of frost shattering are not so simple as they might first appear and, like many geomorphic processes, seem to be accepted more on faith than on solid evidence. Historically, frost shattering was attributed to the volumetric expansion of freezing water and, from laboratory experiments (Battle 1960), was thought to be significant only when the temperature falls rapidly to between −5° C and −10° C. Shattering by volumetric expansion requires that the water in a crack be frozen from the top of the opening downward to the bottom. An ice plug must form first at the top of the fracture to produce the closed system that will allow pressure from ice growth to exceed the tensile strength of most rocks.

Since W. D. Johnson descended into a deep bergschrund in 1904, many workers have investigated the role played by bergschrunds in the mechanics of headwall retreat. A **bergschrund** is a crevasse-like opening near the headwall that separates actively moving ice of the glacier from nonactive ice frozen to the headwall. Johnson (1904) suggested that surface meltwater gains access to the base of the headwall by percolating down the bergschrund and thus produces extensive frost shattering when water that permeates rock fractures is alternately frozen and melted. The significance of frost shattering has been questioned, however, as actual measurements of temperature fluctuations in bergschrunds (Battle 1960) suggest that the variations are not severe enough to produce fracturing by the expansion of freezing water in the ice-sealed cracks. More detailed analysis of freeze-thaw mechanics, now indicates that frost shattering involves more than the simple volumetric expansion of freezing water. Walder and Hallet (1985) argue that at temperatures below 0° C, unfrozen water will migrate to ice bodies in small, preexisting cracks in the bedrock and subsequently freeze. As the water is added to the ice bodies, pressures increase enough to allow the cracks to propagate. Unlike frost shattering by the volumetric expansion of water, shattering by this mechanism is greatest during sustained temperatures ranging from −4° C to −15° C and is enhanced by slow rates of cooling (Walder and Hallet 1985).

In light of the above analysis, frost shattering associated with the movement of water down the bergschrund is plausible (Hooke 1991). Others have argued, however, that the steep headwall morphology of cirques is indicative of erosion below the depths reached by the bergschrund, and therefore the primary mechanism of headwall retreat is by plucking. Hooke (1991) suggests that the flow of rain and meltwater down the bergschrund creates rapid water-pressure variations beneath the glacier in the exact location where erosion is needed to maintain a steep headwall (fig. 10.9). He argues that rock shattering is initiated along the headwall by abrupt decreases in subglacial water pressure, while rapid or more gradual pressure increases lead to plucking of the shattered rock. This model of erosion is similar to that proposed by Iverson (1991b) and discussed earlier in this chapter.

It is important to note, however, that some cirque glaciers do not have bergschrunds. Thus, the above hypotheses of cirque formation might not apply to all cases. Another possible explanation was presented by S. E. White (1976a). He suggests that hydration may exert almost as much pressure as that produced by ice expansion at −22° C. As minerals adsorb water, expansion and contraction of this water occur in response to temperature fluctuations in the range of freezing and thawing. The shattering seen at the base of the headwall might then take place beneath a glacier without the stringent conditions necessary to generate forces by ice growth. The hydration-shattering proposal is thought-provoking and deserves careful investigation. The process not only can explain shattering where temperature fluctuations are minimal, but it also dispels the problems associated with the bergschrund hypothesis. Nonetheless, experimental evidence suggests that hydration by itself may not be as effective as frost action in promoting disintegration (Fahey 1983). Washburn (1980) reminds us that most large products of disintegration are found in periglacial environments, the point being that hydration shattering should operate in any climatic zone as long as water is available. Clearly, more work needs to be done before the mechanics of headwall erosion are completely understood.

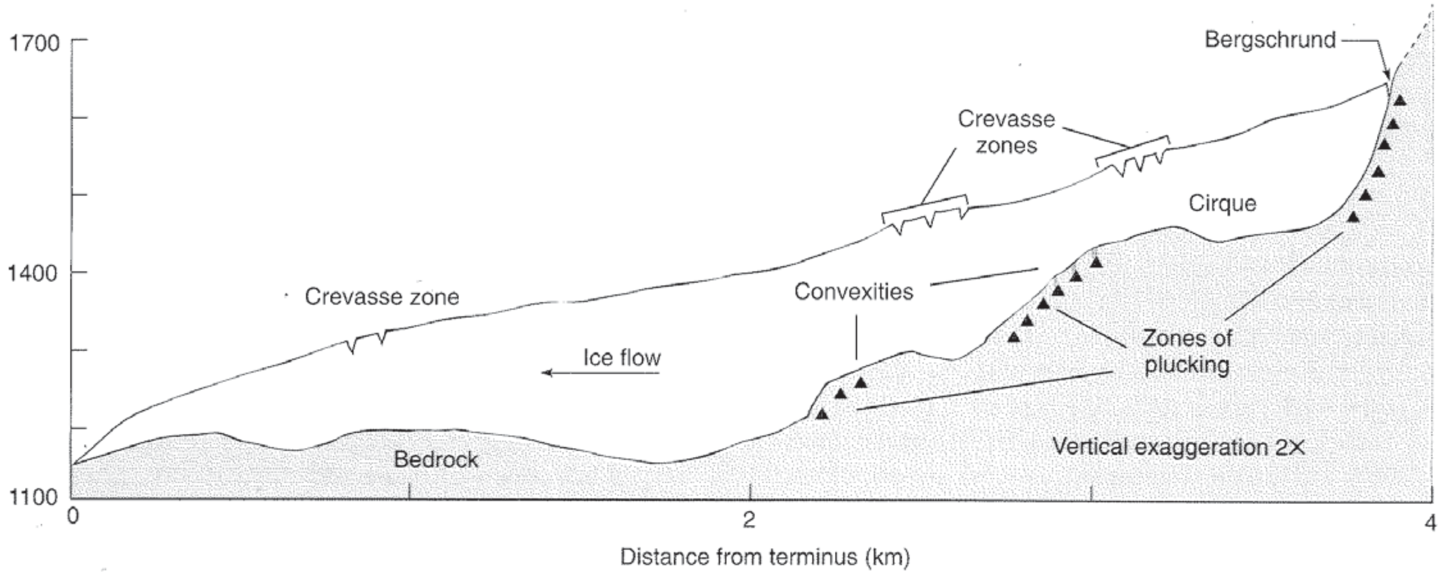



**FIGURE 10.9**

Longitudinal profile of Storglaciaren, Sweden, showing the relationship between zones of plucking and points of water input along crevasse zones and the bergschrund. The downward flow of water through these openings may lead to rapid subglacial water-pressure variations and localized plucking.

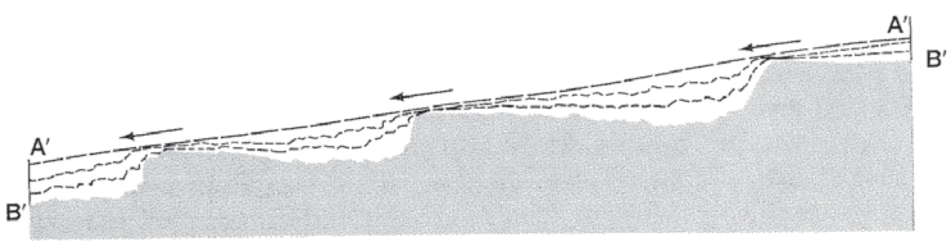
(Hooke 1991)

## Glacial Troughs

Glacial erosion is not limited to the cirque environment, for ice passing over the cirque lip can also remold the preglacial valley topography into a characteristic glaciated form. The ability of ice to remove rock protuberances tends to produce valleys with steep, nearly vertical, sides and relatively wide, flat bottoms (fig. 10.10). The transformation of a V-shaped river valley into a U-shaped glacial valley has been explained mechanically by A. Johnson (1970), and is shown in figure 10.11. Assuming ice to be pseudoplastic and using Nye's (1965) analysis of ice behavior in a triangular form, Johnson suggests that dead regions (areas of no flow or shear stress) should exist in a glacier that invades a mountain stream valley. The dead regions should be present in the ice at the top of the valley sides and along the valley bottom. Erosion would be negligible in these zones but appreciable along other parts of the valley sides where shear stress and velocity would be high. As differential erosion causes the sides to bow outward, the dead regions are progressively removed because the shear stress and velocity distribution change as a more parabolic cross-profile is produced. Eventually, erosion affects all parts of the valley sides, resulting in the characteristic U-shaped glacial valley. The temporal changes in the pattern of valley floor and side erosion, envisioned by Johnson, have been supported by more recent numerical studies (Harbor et al. 1988; Harbor 1992).

Although the assumption has been questioned (Harbor 1990), a parabolic cross-profile probably aids glacial movement because it exerts the minimum resistance to glacier flow (Flint 1971). It may also allow for the maximum efficiency of glacial erosion (Hirano and Aniya 1988, 1989). The precise width and depth dimensions, however, are likely to depend on the intensity of the glacial dynamics (Graf 1970) and the properties of the geologic framework (Augustinus 1992). The formation of the cross-sectional shape occurs by both lateral and vertical erosion of the preexisting valley. Whether the parabolic form is derived predominantly by widening or predominantly by deepening depends on the properties of the rocks and the ice, as well as the extent of preglacial weathering and the amount of load available within the ice to be utilized as cutting tools. In any case, the erosion leads to the truncation of rock spurs jutting into the valley and the formation of **hanging valleys.** These occur particularly where trunk valleys carry more ice and are more extensively eroded than their tributaries.

(A)




(B)

**FIGURE 10.10**

(A) U-shaped cross-profile of glaciated valley and a hanging valley. View up Yosemite Valley from vicinity of Artist Point with El Capitan at left, the Cathedral Rocks and the Bridalveil Falls at right. Yosemite National Park, Mariposa County, Calif.
(B) Longitudinal section of glacial valley in Yosemite National Park illustrating staircase profile. *AA* is preglacial valley floor; *BB* is present valley floor. Broken lines represent intermediate stages of development.
(After Matthes 1930)

Glaciated troughs are also characterized by uniquely irregular longitudinal profiles that essentially represent a series of interconnected basins and steps; the diagram in figure 10.10B shows this typical profile in Yosemite National Park. The basins may contain lakes, a series of which are sometimes called *paternoster lakes*. Immediately downvalley from the basins holding the lakes are rock bars or steps that commonly show

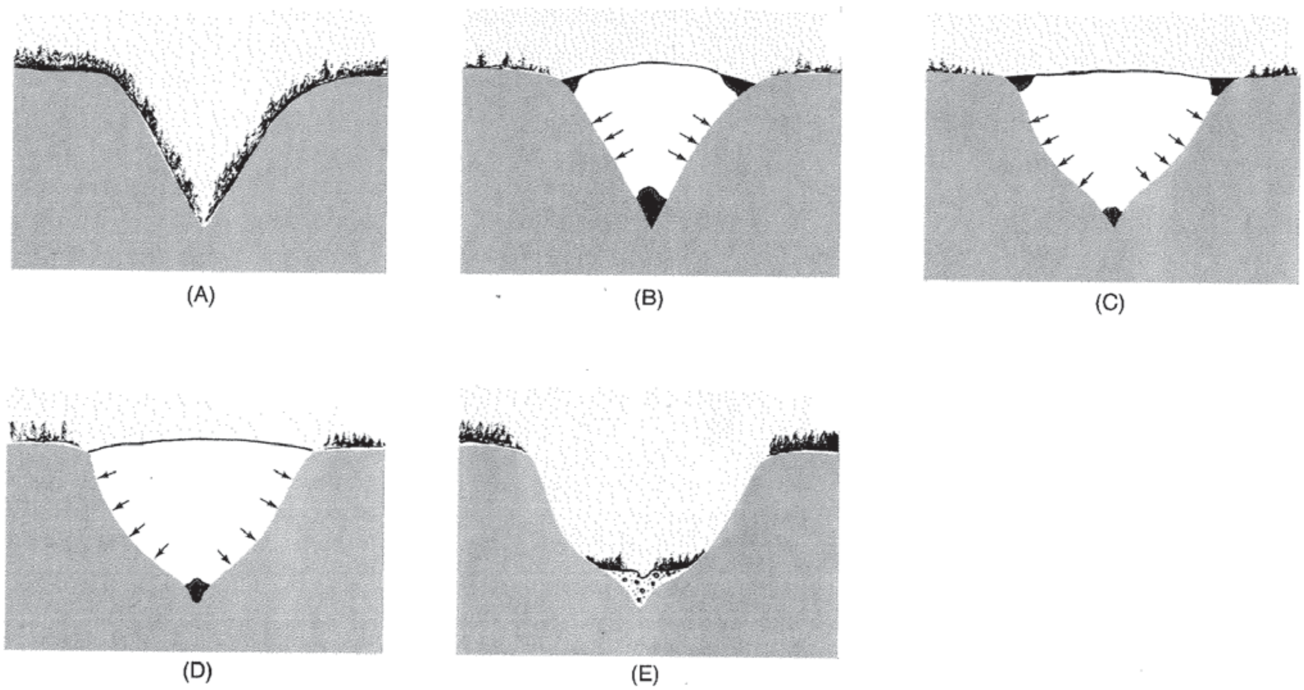



**FIGURE 10.11**

Possible sequence of events leading from V-shaped mountain canyon to U-shaped glacial valley. (A) V-shaped mountain canyon. (B) V-shaped canyon visited by glacier. (C) Glacier erodes sides of canyon. (D) "Dead" regions disappear and entire side of canyon is rasped by rock-studded glacier. (E) Glacier disappears, leaving U-shaped valley.
(Johnson 1970)

the effects of intense abrasional smoothing on their relatively flat surfaces. At the distal end of the step, a sharp break in slope occurs where plucking has produced a steep scarp that connects the rock bar to the next lower basin or step.

The staircase profile has intrigued geologists for years and has been the subject of considerable speculation. Hypotheses for its origin include (1) variation of rock structures, especially spacing of joints, causing differential erosion, (2) preparation of weak zones in the rock by preglacial weathering, (3) irregularities in the preexisting valley topography, and (4) increased erosive power at the confluence of tributary valleys with the main valley (see Bakker 1965). It is possible that rock bars producing a steplike profile require no special conditions for their formation. Theoretically, rock bars can form where no obviously harder rock exists, and, conversely, zones of resistant rocks do not necessarily evolve into rock bars under glacial erosion.

The abrasion prevalent on the treadlike steps and the quarrying at the position of the steep scarps led Lewis (1947) to note that each step resembles a large roche moutonnée and so probably has a similar origin. He concluded that the mechanics probably worked best beneath thin glaciers where meltwater could easily penetrate to the glacier floor and shatter the rock by refreezing. Although freeze-thaw mechanics might be a reasonable accomplice in the formation of staircase profiles, we cannot ignore the fact that development of the associated basins probably required some differential ice movement capable of flowing uphill and carrying load in the process. Such a requirement does not encourage the belief that the ice was thin; in fact, it is common knowledge that many valley glaciers have thickened to the point of overtopping their divides and spreading into adjacent valleys. In addition, as discussed earlier, the freeze-thaw component in rock shattering is somewhat suspect because we do not completely understand the thermal conditions at the base of glaciers. What we are left with, then, is the problem that while plucking is necessary to form the scarps, the rock shattering required for this process to function is probably not a result of frost action. Furthermore, some scarps form where no regional joint system is present to facilitate the shattering process.

Boulton (1974) suggests an alternative hypothesis of shattering that deserves our consideration because it not only resolves the above problem but also relates

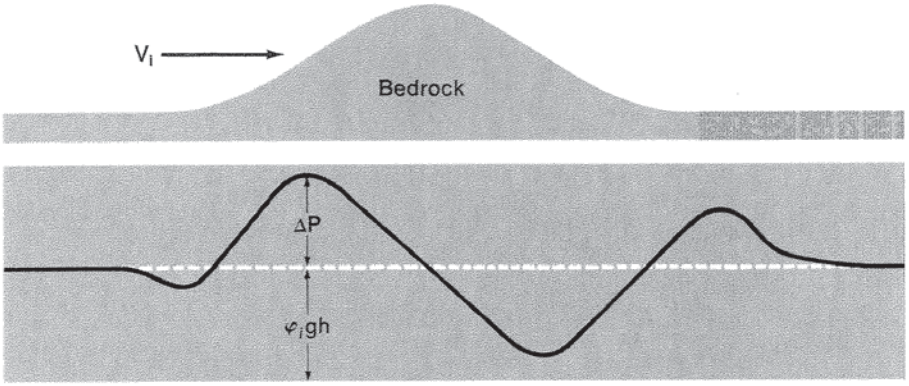



**FIGURE 10.12**

Schematic view of normal pressure distribution at glacier bed as ice flows over a bedrock obstruction. $\Delta P$ represents the increase or decrease of normal pressure produced by a bedrock obstruction. The normal pressure where no obstruction exists is represented by the dashed horizontal line and has a value equal to $\rho_i gh$.
(Boulton 1974)

directly to glacier mechanics. When a glacier moves across a horizontal bedrock surface, the effective normal pressure will be $\rho gh—w_p$ where $\rho$, $g$, $h$, and $w_p$ are, respectively, density, gravity, ice thickness and water pressure at the bedrock interface. The dashed line in figure 10.12 represents the normal pressure on the horizontal bed and is estimated by $\rho_i gh$ (where $\rho_i = \rho$). Since $\Delta P$ in the figure represents the increase or decrease of normal pressure produced by a bedrock obstruction, the total normal pressure at any point is $\rho_i gh + \Delta P$. Therefore, if the bed is irregular, the normal pressure will fluctuate so that it will be higher than average on the upglacier side of a bedrock obstruction and lower than average on the lee side. Boulton shows that the shear stresses included in the bedrock will be greatest where the normal pressure is lowest, that is, down-ice from the crest of the obstruction. The exact position and absolute magnitude of the maximum shear stress depend on whether cavitation occurs downglacier from the bedrock knob. Furthermore, Boulton points out that the shear strength of the rocks will be least in the downglacier position where the shear stress is greatest, creating the ideal situation for rock failure at that locale. Assuming this analysis is correct, the inevitable conclusion is that with the proper bedrock configuration even hard unfractured rocks may be crushed beneath the ice, and the rocks already jointed can be intensely shattered. Preglacial irregularities in the long profile, structural weakness, and lithologic variations all will be accentuated by this process because shattering will prepare those zones for the plucking action that produces the scarps in the staircase profiles.

Preglacial irregularities may also be accentuated by rock shattering and plucking processes associated with rapid subglacial water-pressure fluctuations. Hooke (1991) notes that crevasses often develop over convexities in the glacier bed (see fig. 10.9). Openings at the base of these crevasses allow rain and meltwater to enter the subglacial drainage system, and this input of water can cause rapid subglacial water-pressure variations that are much greater at the base of the convexities than in other localities. Based on these observations, Hooke (1991) proposed a model similar to the one he devised for cirque headwall erosion (discussed earlier) in which subglacial water-pressure changes lead to both rock shattering and localized plucking. These processes enhance erosion at the base of the initial convexities and can amplify preexisting valley floor morphology into a typical staircase profile.

A special type of glacial trough exists mainly in high-latitude coastal regions that are underlain by resistant rocks, so that the general land surface stands at considerable elevation above the nearby ocean. These troughs are called *fiords*, and they differ from other types only because they are partially submerged by the ocean. The inundated bottoms of fiords have the same variety of topographic elements, both erosional and depositional, that exist in a normal continental glaciated valley (Holtedahl 1967). Their history may include components of both glacial and fluvial processes, and they may be partly controlled by tectonic and lithologic factors. For these reasons, it is unwise to make sweeping generalizations about their origin.

Perhaps the most salient property of fiords is that part of their development took place when the ice was physically beneath the ocean (Crary 1966). Flint (1971) reminds us that a glacier 1000 m thick with a density of 0.9 will remain in contact with its bed and be fully capable of erosion at water depths up to 900 m. Even at greater depths, when the snout begins to float, high topographic irregularities of the valley might still be eroded (Crary 1966). The water depths in fiords, several hundred meters in many and greater than 1000 m in some, are well beyond that which can be attributed to a postglacial rise in sea level (see Flint 1971), adding credence to the suggestion that much fiord erosion was accomplished in a submarine environment.

## Deposits and Depositional Features

Before glaciers were recognized as viable geomorphic agents, deposits containing boulders that obviously came from a distant source were called *drift*. This term arose because elimination of known processes led to the belief that the anomalous boulders reached their site of deposition by riding on top of floating ice. After glaciers were recognized as the transporting vehicles, the term "drift," or **glacial drift,** was retained and expanded to include all deposits associated with glaciation. Drift covers approximately 8 percent of Earth's surface above sea level and almost 25 percent of the North American continent. The thickness of this cover varies greatly. In New England, for example, only a thin layer of drift (< 20 m) covers most upland areas, but drift may be several hundred meters thick in buried valleys. The drift in the central United States generally varies from 10 to 60 m thick, but once again these are average values. In some places the drift is merely a thin mantle on top of bedrock, and in other regions, such as parts of Michigan, it exceeds 200 m in thickness. The exact volume of drift deposited depends on the time span of glacier activity and the thermal regime of the ice, but with high velocities and loads, as much as 30 m of drift can be accumulated in less than 10 years (Flint 1971).

Through the years geologists have been intrigued with the amazing variety of glacial drift and the complex interrelationships that exist between the different types. This complexity arises for several reasons: (1) drift may be deposited from mediums that contain vastly different amounts of water; (2) deposition occurs beneath, within, or on top of the ice, at the glacier margins, in bodies of standing water, or in fluvial settings far from the glacier, the debris being transported there by streams rising in the ice mass itself; (3) the depositional sites and environments and the drift composition all change with time because glaciers themselves are not constant in their properties or fixed in their position; and (4) the glacier may be active or stagnant. Because of these complicating realities, any discussion of glacial drift and the associated depositional features is difficult to organize in a way that is entirely satisfactory. Our approach here is to briefly examine the varieties of glacial drift and then look at the depositional features according to the environment in which they originate. We will also attempt to relate the morphology of features and their sedimentary properties to the dynamics of the system. Throughout this discussion it is important to recognize the distinction between the sedimentological character of drift and the morphology of features that result from its deposition. Morphological terms such as moraines and kames do not imply a particular drift type. Many features with similar morphology are composed of a number of drift varieties, especially where the environment of deposition is subject to repeated change.

### Drift Types

Over the years glacial deposits have normally been divided into two categories based on their sedimentary characteristics, primarily the presence or absence of layers and the degree of sorting in the deposit. In our discussion, we will follow this typical division and separate drift into stratified and nonstratified types.

**Nonstratified Drift** The term "till" usually connotes material that has been transported and deposited by the ice itself, a process often indicated by striations or microscopic fractures on the grains (Krinsley and Donahue 1968). Till typically has no discernible stratification and is characterized by a mass of unsorted debris that contains angular particles composed of a wide variety of rock types. Many examples justify this description of till. Absence of layering and poor sorting (often characterized by an almost universal bimodal size distribution) seem to be the most reliably consistent properties. The bimodality observed in most tills is probably due to the differences in grains produced by abrasion and those derived by plucking. In an excellent review, Goldthwait (1971, p. 4) points out that till is probably more variable than any other sediment that is described by a single name.

Any of the above identifying criteria in our definition may be missing at a particular till locality as a result of varying transporting and depositing mechanics and heterogeneity in the rocks over which the ice has passed. For example, many clasts in till have a subangular pentagonal or triangular shape, but these forms may be significantly altered by rounding during transportation. This is especially true of particles that have been transported subglacially. In that environment, parent debris is rounded by attrition of edges and corners (Boulton 1978), although the degree of rounding is partly dependent on rock type (Holmes 1960; Vagners 1966), the original clast shape (Drake 1968, 1974), and the distance of transport. In addition, glaciers that override older stream deposits may incorporate boulders that have already been rounded, resulting in a till that is notably less angular than one would expect. Figure 10.13 shows till that contains rounded boulders.

**FIGURE 10.13**
Late Wisconsinan (Pinedale) till in Rock Creek valley, Beartooth Mountains, Mont. Note rounded boulders in till.

Overall particle size tends to be reduced by attrition in subglacially transported till (Mills 1977; Boulton 1978). Each particular source rock tends to produce a characteristic texture depending on how easily its clasts can be crushed and how far they have been transported (Mills 1977; Dreimanis and Vagners 1971; Humlum 1985). Thus, the overall angularity and texture of till depend on the lithologic heterogeneity of the source rocks as well as the distance of transport from the location of their outcrop.

The unpredictability of till increases still more because the material is carried in different parts of the glacier, and each transport subenvironment produces till with different characteristics. For example, in valley glaciers considerable debris may be shed onto the surface from the valley sides and be transported as supraglacial load. The resulting supraglacial tills have a texture dominated by coarse, angular clasts because the particles are not crushed during transport and fines tend to be washed away as the surface ablates. In contrast, englacial load, generally carried in the basal ice layers, can be deposited directly beneath the glacier under considerable pressure exerted by the overlying ice. In this case, the debris, referred to as **basal till** or **lodgement till,** is compact and contains a higher percentage of fine-grained sediment. In addition to being deposited at the base of the glacier, englacial load can rise within the ice along shear planes or along the normal upward flow lines in the terminal zone. This material finally emerges as supraglacial debris when ablation of the surface ice releases the contained particles. This debris, or **ablation till,** will be emplaced on the glacier bed as the ice beneath it gradually disappears. Ablation tills, in comparison to lodgement tills, are usually less consolidated and dense.

Ablation till forms as one of two types, flow-till (Hartshorn 1958; Boulton 1968) and melt-out till (Boulton 1970b). **Flow-tills** result when englacial debris is released at the surface, and ablation of the underlying ice destabilizes the material, allowing it to move downslope. Generally, the highly mobile debris moves as a sediment flow, but it may move by other mass wasting processes such as creep or semiplastic sliding. It is important to recognize that flow-tills are diamictons (poorly sorted, terrigenous sediments) that have been remobilized, transported, and redeposited at sites removed from the original position at which the debris is released from the ice. Therefore, flow-tills are often difficult to distinguish from sediments deposited by geomorphic agents in nonglacial environments, leading some geomorphologists to refer to flow-tills as secondary glacial deposits (Coates 1991).

**Melt-out till** forms by the downward movement and accumulation of debris during ablation. Ice covered by a thin veneer of debris ablates more rapidly than clean ice because it has a lower albedo and is able to transfer more solar radiation to heat. However, when the thickness of the surface cover reaches about 3 cm, the ablation process is severely retarded, and if the surface layer exceeds 1 or 2 m, ablation ceases. Further increases in the thickness of the supraglacial cover can occur only if more debris moves to the site from higher levels of the ice surface (via flow), or if the ice melts beneath the layer already concentrated at the surface. Melt-out till represents the in situ accumulation of sediment (Boulton 1971, 1972b) and rarely exceed 3 m in thickness. The total supraglacial layer can be much greater, however, if flow-till continues to accumulate on the surface. Little is known about the rate of accumulation of ablation tills, but Mickelson (1973) has calculated that the accumulation of basal melt-out till of the Burroughs Glacier in Alaska ranges from 0.5 to 2.8 cm a year.

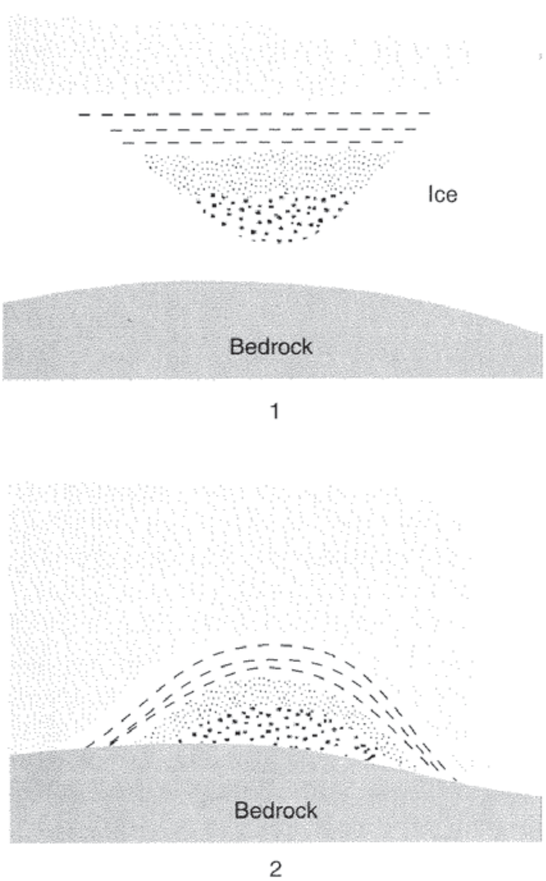
**Stratified Drift** The second major category of glacial drift is characterized by sediment that was transported by moving water before its final deposition, thereby acquiring a degree of stratification not normally seen in tills. One common form of such drift is often referred to as **fluvioglacial** because running water is involved in its origin, even though the water may not always be confined in discrete channels. Fluvioglacial deposits can also be distinguished from till because they are usually sorted and the clasts contained in the mass are more rounded. However, the demarcation between some types of fluvioglacial deposits and thoroughly washed ablation tills is a matter of degree rather than substance. Exactly where the line between the two is drawn is somewhat arbitrary. Highly saturated flow-till, for example, might move in a nearly fluvial manner.

The layering and sorting in a fluvioglacial deposit depends on precisely where it is formed with respect to the ice that provides the transporting meltwater. Sorting is also partly a function of the energy possessed by the meltwater, the distance of transport, and the continuity of the sorting process. Because the discharge of meltwater is notoriously inconstant, varying drastically with time of day, local climate, and the characteristics of the ice, significant differences in the sedimentology of fluvioglacial deposits can be noted over short distances. These are especially evident where deposits are formed in contact with the ice and the free circulation of meltwater is restricted (see Shaw 1972 and N. Smith 1985 for sedimentary characteristics). If debris is transported away from the glacier terminus, the sedimentary characteristics tend to vary more regularly (N. Smith 1985).

We should stress again that the prime requisite of stratified drift is transport by water, much of which is released from melting ice. However, this puts no constraints on the environment in which the sediment is deposited. Fluvioglacial debris can come to rest in stream channels, floodplains, lakes, ocean floors (Rust 1977), deltaic plains, or in any other place where sediment-laden running water loses its transporting energy. In some cases sedimentologic properties in the deposits, such as the degree of roundness and mean grain size, can help identify the depositional environment (King and Buckley 1968), but normally there are too many variables in the system to rely on these criteria alone (R. J. Price 1973). Detailed field study is almost always necessary to reach a firm conclusion about the environment of deposition.

Stratified deposits that originate in contact with the ice often contain interbedded bodies of ablation till, and the particles tend to be less well rounded and sorted because of the limited distance of transport. The characteristics of such a deposit often vary from the bottom of the sequence to the top because the environment of deposition repeatedly changes with time, especially if the ice is stagnating. In any deposit, then, a particular layer may be superseded vertically by one with different sedimentary properties that reflect a new depositional environment. The stratigraphy is complicated because much of the drift is physically supported by ice during its deposition, and when the ice dissipates, the support is removed and the sediment collapses. Such a process, as figure 10.14 shows, leads to flexures in some of the layers, minor faults, and beds dipping at angles well beyond the angle of repose for such material. Overall, the ice-contact

(A)




(B)

**FIGURE 10.14**

(A) Gravel pit cut in kame, south central New York, shows stratification in kame deposits. (B) Structures and deformation of strata in ice-contact stratified drift develop as ice melts and debris collapses and is lowered onto bedrock floor.

setting at places produces an interconnected maze of stratified and nonstratified drift in which every conceivable process and environment is possible and probably has been present at some time during the depositional history (fig. 10.15). Moreover, the manner in which the ice ablates influences the resulting deposit. Drift produced while the ice is still active may be quite different, especially in distribution, from that derived from a large mass of stagnant ice that is simply downwasting as it melts and is not influenced by internal glacial movement.

Sediment deposited beyond the terminal margin of the ice is formed in the proglacial environment and is often referred to as outwash. **Outwash** is usually well sorted and normally consists of rounded sand and gravel representing bedload carried and deposited in stream channels. Silt and clay are usually carried as suspended load and are commonly removed from the system unless, as in the lower Mississippi River valley, the transport distance is so

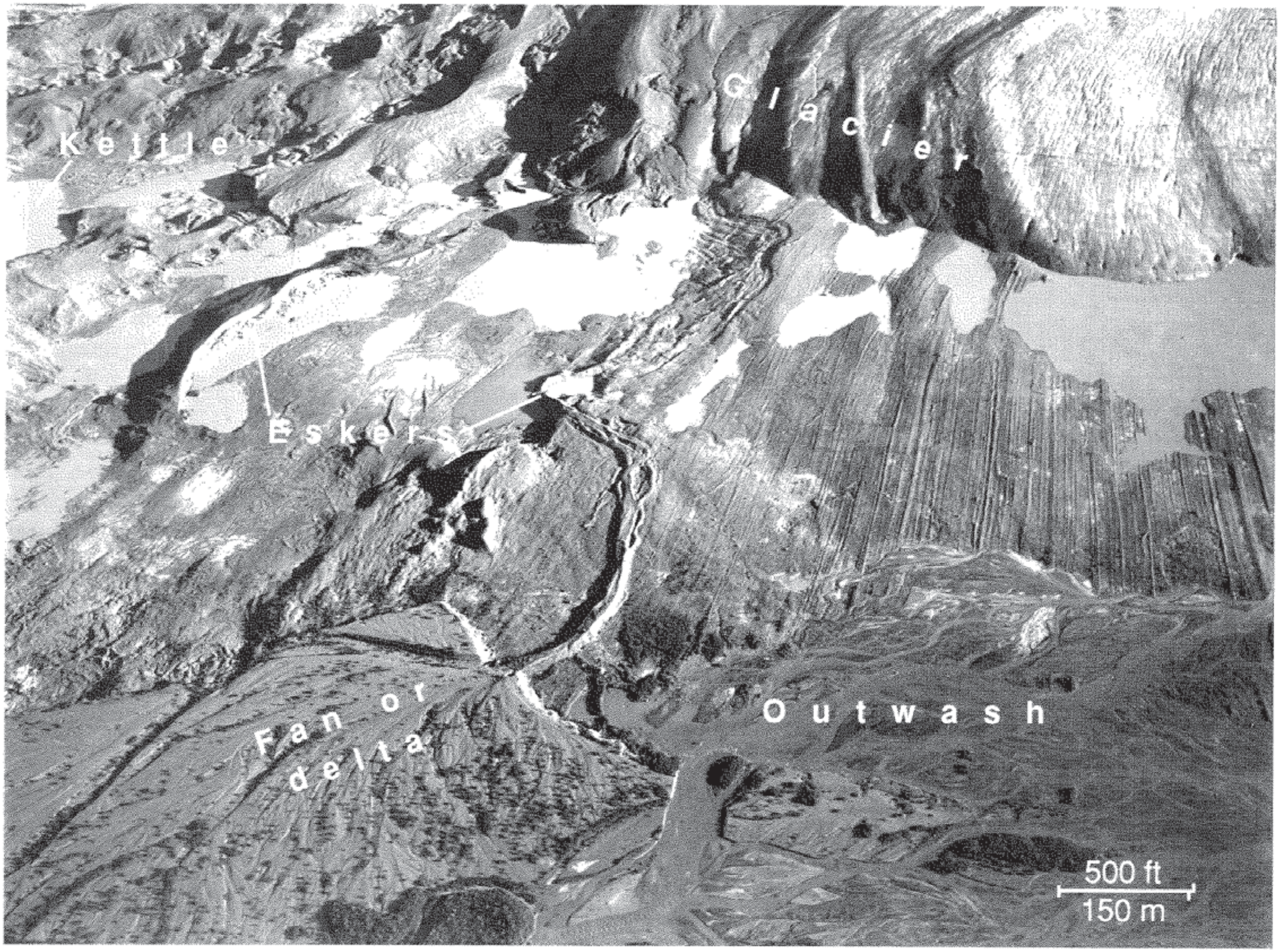



FIGURE 10.15
Marginal ice-contact zone at the terminus of Woodworth Glacier, Alaska.

great that some of the outwash is silty in texture. Streams transporting outwash do not usually head at the glacier terminus but begin on top of, beneath, or within the ice, well upglacier from the margin. Proglacial features and deposits often can be traced into and through the maze of ice-contact deposits, increasing the complexity of the depositional sequence developed near the ice margin.

## The Depositional Framework

Before considering what geomorphic features might be produced during deposition, we must first establish a realistic framework within which we can give some order to the subtle variations of depositional features associated with glaciation. This is done in table 10.1. From our brief look at the nature of drift, it is obvious that both stratified and nonstratified deposits can be formed in contact with the ice. They differ only in whether the ice alone was the primary transporting and depositing agent or whether a stage of water transport intervened between the release of particles from the ice and their final deposition. Thus, the **ice-contact environment** must be considered as one of the major settings of our depositional framework. The ice-contact environment can be subdivided geographically into **marginal** and **interior zones** depending on where the drift was originally deposited (table 10.1).

**TABLE 10.1** The depositional framework and associated features.

| Setting | Features | Type of Drift |
|---|---|---|
| Ice contact | | |
| Marginal | End moraines | Till and fluvioglacial |
| | Kames and kame terraces | Fluvioglacial |
| | Kettle holes | Fluvioglacial |
| | Eskers | Fluvioglacial |
| Interior | Medial and interlobate moraines | Till |
| | Ground moraines | Lodgement till |
| | Fluted surfaces | Lodgement till |
| | Drumlins | Lodgement till |
| Proglacial environment | Sandar | Fluvioglacial (outwash) |
| | Kettled sandar[a] | Fluvioglacial (outwash) |

[a]May merge with marginal environment.

Features formed in the ice-contact environment can be composed of either stratified or nonstratified drift (till or fluvioglacial sediment) or a combination of both types. The region beyond the terminal edge of the glacier is classified as the **proglacial environment.** In contrast to the ice-contact setting, features formed there are composed almost entirely of stratified drift. These sediments may be deposited directly by proglacial streams, or they may be deposited by other geomorphic agents operating in front of the glacier, such as those associated with proglacial lakes (e.g., see Teller 1987).

The distinction between marginal and interior zones in the ice-contact environment is problematical for several reasons. First, glacial margins migrate forward and backward with time according to the glacier's mass budget. An active glacier with a negative mass balance should have a terminal margin that is progressively receding toward its source. Marginal features will be formed in regions that were interior when the ice was at its greatest extent, and deposits of the two zones may be complexly intertwined. Second, the processes operating near the contact between the marginal and interior zones are dependent on the characteristics of the subglacial environment. For example, the outer 2 to 3 km of ice sheets are commonly frozen to the underlying surface even though farther upglacier the basal ice may be temperate. Where marginal ice is frozen to the bed and high pore water pressure exists, thrusting along preexisting planes of weakness in the substrate can inject large blocks of subglacial material into the ice (Moran et al. 1980). These thrusted blocks tend to concentrate where ice margins rest on aquifers and on the upslope edges of upland areas. Proceeding upglacier toward the interior zone, the ice is free to slide because the ice-bed contact is unfrozen. Thrust blocks in the transitional area between the marginal and interior zone are smaller and smoothed over by the sliding ice. In the true interior zone, no thrusting will occur and the subglacial terrain is characterized by streamlined forms (Moran et al. 1980). Clearly, the position of the frozen ice–thawed zone contact may migrate with time, changing the subglacial environment. Thus, in any given glaciated area, the inner portion of the marginal zone may be transitional into the interior zone rather than marked by a well-defined contact between the two.

Regardless of the problems associated with precise boundary locations, the depositional framework fits our purposes because the suite of features found in each zone is a direct reflection of the genetic processes involved. The utility of the classification is shown in figure 10.16. In this region of Wisconsin the terminal moraine, marking the ice-contact marginal zone, is characterized by a maze of small hills and depressions (kames and kettles) that distinguish deposits formed near the boundary between active and stagnating ice. North of the moraine the relatively low, flat area is underlain by material that was deposited beneath the active glacier (ground moraine); it represents the ice-contact interior zone. South of the moraine the proglacial zone is marked as a plain composed of debris (outwash) transported and deposited by meltwater streams heading within or on top of the ice.

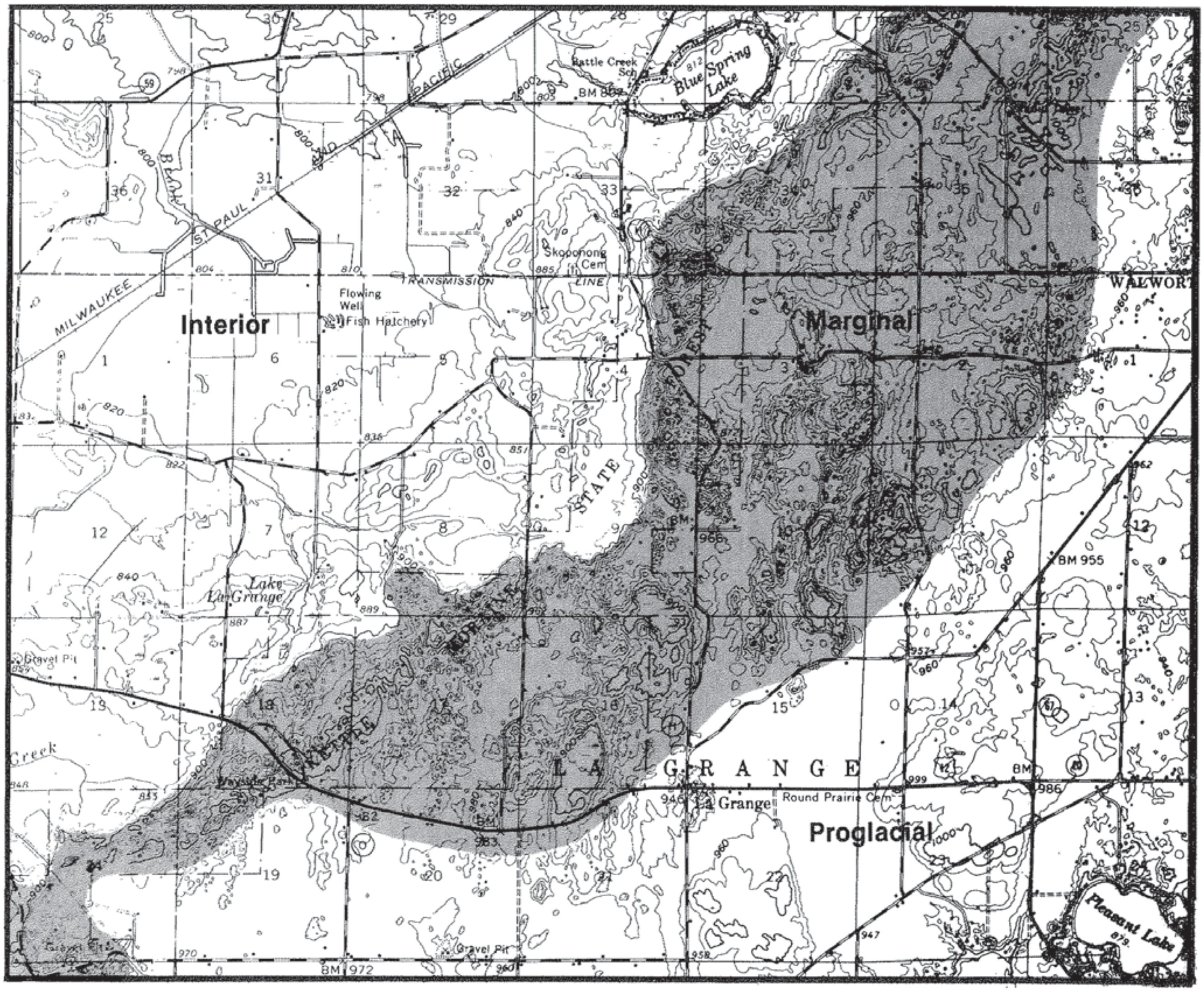



FIGURE 10.16

Map view of the depositional framework after ice has disappeared. Part of the Whitewater, Wis., quadrangle (U.S.G.S. 15′). Marginal zone contains terminal moraine and ice-contact stratified features, and interior zone is characterized by ground moraine. Proglacial zone is a large outwash plain or sandur.

## Marginal Ice-Contact Features

**Moraines** The term "moraine" originated several hundred years ago as a local name for ridges of debris found at the edges of glaciers in the French Alps. Since then, many definitions have appeared, but for our purposes we can think of a **moraine** as a depositional feature whose form is independent of subjacent topography and is constructed by the accumulation of drift, most of which is ice-deposited (Flint 1971). A precise morphological definition is not possible because, as table 10.2 shows, moraines take many different forms and have a variety of dimensions. The term "moraine" is not synonymous with "till," as many have suggested, but refers to a suite of topographic forms on which the only restriction is that they must be composed of drift.

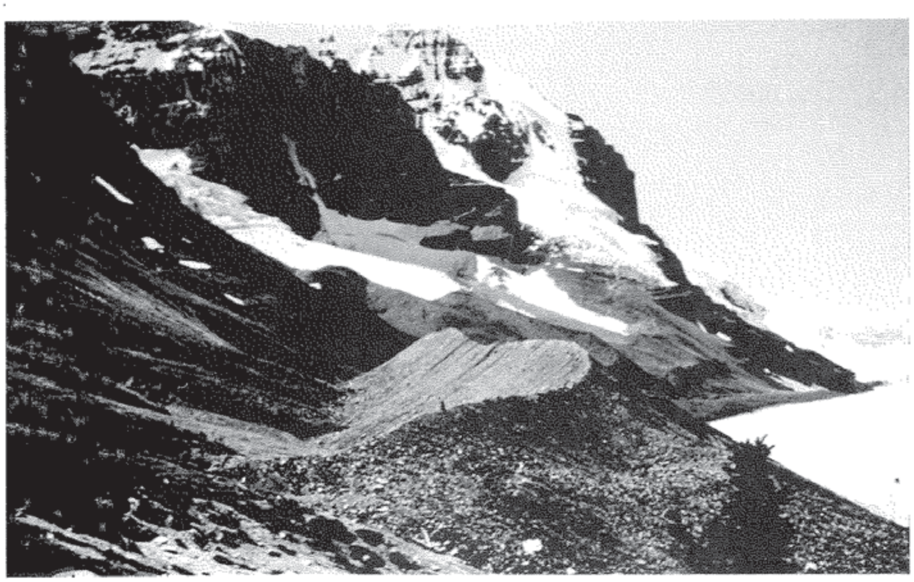
The most spectacular moraines develop at or near the edges of active glaciers and so are designated as **end moraines.** The end moraine constructed at the downstream edge of the ice at the farthest point of advance is called a **terminal moraine.** In valley glacier systems, it merges imperceptibly into **lateral moraines** on both sides of the valley because ablating ice deposits debris along the glacier's lateral extremities as well as at its terminus (fig. 10.17). Ice sheets form terminal moraines that tend to be long (often hundreds of kilometers), linear, topographic highs marking the forward boundary of the ice, but lateral moraines are absent because there is no side to the ice sheet. Where ice sheet movement was distinctly lobate, however, end moraines called *interlobate moraines* (Flint 1971) may develop along the junction of two lobes.

**TABLE 10.2** Moraine types.

| | |
|---|---|
| **End Moraines** | Moraines produced at front or sides of an actively flowing glacier.[a] |
| Terminal moraines | Mark the farthest advance of an important glacial episode. |
| Lateral moraines | Deposited at or near the side margin of a mountain glacier. |
| Recessional moraines | Formed at glacier front during temporary halt or readvance of ice in a period of general recession. |
| **Ground Moraine** | Gently rolling surface formed of debris released from beneath the ice. |
| **Interior and Minor Varieties** | |
| Washboard moraines | Small, parallel ridges oriented transverse to direction of ice movement. Also called moraine ridges or cross-valley moraines. |
| Interlobate moraines | Formed at junction of two ice lobes. |
| Medial moraines | Elongate ridge developed at junction of two coalescing valley glaciers. |
| Rogen moraines | Large sequence of ridges transverse to ice flow. Formed in the interior zone. |

[a]Lateral moraines may be excluded by some geomorphologists because end moraines are commonly considered only as topographic features developed at the front of a glacier.

(A)



(B)

**FIGURE 10.17**

(A) Terminal moraine of Pinedale glaciation, East Rosebud valley near Roscoe, Mont. Low ridge above road in left center of photo is lateral moraine merging with terminal zone. (B) Recently formed lateral moraine in Rocky Mountains of Canada.

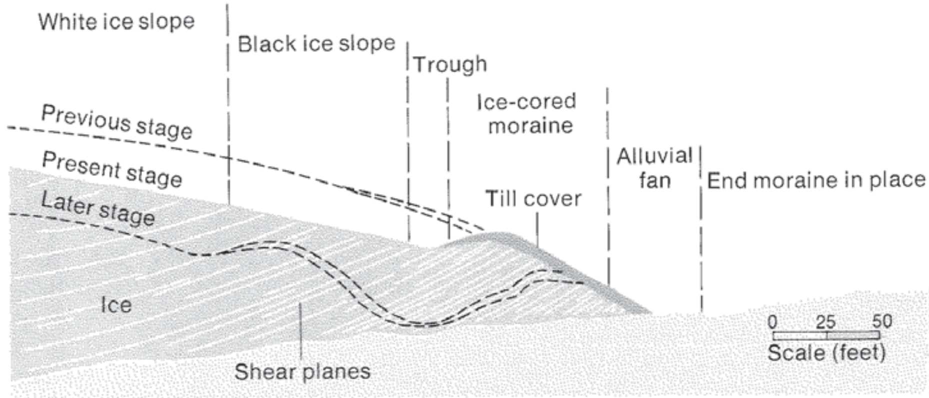



**FIGURE 10.18**
Diagram showing retreating margin of Barnes ice cap. Material moves up inclined shear planes and accumulates as till cover on core of ice.
(Goldthwait 1951)

Ideally, end moraines assume a rather narrow, ridgelike shape, but actually the form and size depend directly on the amount of glacial load, the mass budget, and the volume of meltwater circulating in the system. Temperate valley glaciers tend to build higher and more massive terminal moraines (some reaching 300 m in height) because these glaciers have higher flow velocities, loads, and total budgets. Ice sheets seem to generate moraines that are less dramatic in size, seldom exceeding 50 m in height. Moraine ridges formed in the marginal zone are developed by three processes known as *dumping, squeezing,* and *pushing* (R. J. Price 1973). Creation of an end moraine by dumping requires that englacial or subglacial debris be transported to and released on the ice surface. The sediment cover induces differential ablation and is often associated with subsequent flow movements of the material. Debris can be moved surfaceward along shear planes (Goldthwait 1951, fig. 10.18) or along the normal flow lines found in active glaciers (Boulton 1967). Regardless, the evolution of morainal topography occurs in a zone where debris dumped at the surface triggers a complex sequence of ablation and flow. Because active glacier fronts shift downvalley or upvalley according to the glacial budget, the position of dumped sediment moves, and the morainal topography may exist in a belt rather than a single ridge.

Moraines formed by stagnant ice in the marginal zone are not always characterized by distinct ridges that have developed transverse to the direction of ice flow. These moraines, commonly called *disintegration moraines* (Gravenor and Kupsch 1959), have local relief up to 70 m and develop from the release of supraglacial drift in the lower part of the ablation zone. When the ice in glacier margins stagnates, it often breaks into isolated blocks of wasting ice covered by ablation till (fig. 10.19). Although the depositional environment is ice-contact marginal, the morainic topography gradually develops as the glacier surface down-wastes over dissipating ice cores. This ice-cored moraine is different from ridges or linear belts formed from active ice that oscillates back and forth. In the stagnant marginal zones, flow till, melt-out till, and fluvioglacial deposits can all coexist on the wasting surfaces, and because the ice cores melt at different rates, they may become mixed within the chaotic surficial expression (fig. 10.20). Although disintegration moraines evolve in a way somewhat similar to dump moraines, they differ in that debris is not necessarily transported to the surface before the ablation process begins.

In contrast to moraines formed by the dumping process, moraine ridges are also developed by the squeezing of drift originally deposited beneath the ice. Ridges generated by squeezing are usually smaller than dumped moraines, normally standing less than 10 m high, although they can be higher where the subglacial till is highly saturated. Theoretically, the process involves the response of water-soaked lodgement till to pressure exerted by the weight of the overlying glacier. The till will move from under the ice and emerge along the ice front, or it will be squeezed into zones of low pressure within the ice, represented by crevasse openings (R. J. Price 1970; Mickelson and Berkson 1974). In either case the ridges seem to be typified by (1) steeper distal slopes than proximal slopes, (2) a plan view outline consisting of linked arcuate segments that parallel the ice front, and (3) till with pebbles oriented perpendicular (or nearly so) to the ridge crests, though occasionally oriented parallel to the trend of the ridge (Mickelson and Berkson 1974).

A third mechanism capable of forming a moraine ridge is the collision of advancing ice with older deposits, which deform them into a ridgelike feature. Such end moraines, called *push moraines,* are only partly composed of sediment carried by the ice and may include blocks of older rocks of a nonglacial origin (Kaye 1964b; Andrews 1980; Humlum 1985). Push moraines are distinguished by their internal structure. Individual layers, sheared away from the ground lying in front of the ice, are intensely deformed into large (sometimes overturned) folds and faults of all kinds

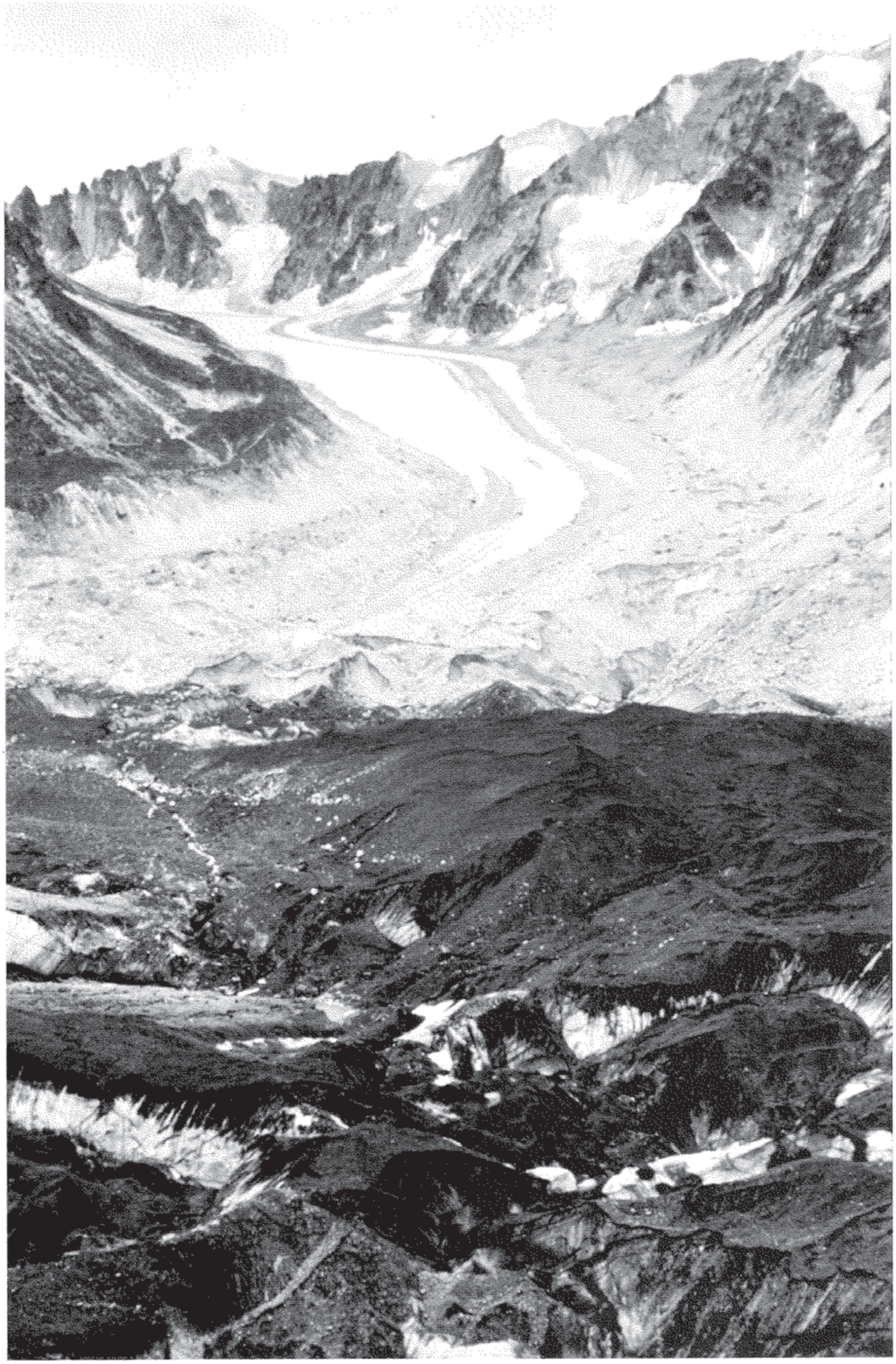


FIGURE 10.19
Ice-cored moraine, Yanert Glacier, Alaska.

**FIGURE 10.20**
Hummocky topography in terminal moraine. East Rosebud valley at front of Beartooth Mountains near Roscoe, Mont.

(Mills and Wells 1974). Thrusting, with plates as thick as 30 m, is not uncommon and may occur as imbricate slices or even as an underthrust phenomenon (Kaye 1964a, 1964b).

**Stratified Marginal Features** As pointed out earlier, many glaciers are characterized by marginal zones of thin, stagnating ice. Within these zones a suite of genetically related ice-contact features is developed which is composed predominantly of stratified drift and is morphologically distinct from the moraines just described. These features form by deposition of drift (1) where water flows through openings in and beneath the ice, or in ice-surface channels, (2) in spaces between the ice and the bedrock of the valley sides, and (3) where disseminated sediment is passively concentrated by melting of the encasing ice.

Many of the channelways for the flowing water originate when stagnating ice breaks into individual segments along planes of structural weakness in the ice, and so the features are genetically related to many of the ice-disintegration forms described by Gravenor and Kupsch (1959), Stalker (1960), Clayton (1967), Parizek (1969), and many others. Our discussion, however, is restricted to those features that consist primarily of fluvioglacial drift, even though ablation till may be a minor ingredient. Their morphology is entirely constructional, although they can be affected by slumping as the ice walls that support the drift during its deposition are melted away.

*Kames and Kettles* *Kames* are moundlike hills of layered sand and gravel that vary in size from minor swells to conical protuberances standing up to 50 m high and extending 400 m along their base. Kame material can accumulate at the ice-substratum interface and also in cavities located within stagnating ice or on its surface. Englacial or supraglacial debris can be lowered onto the ground surface as the ice dissipates (Cook 1946; Holmes 1947). The moundlike shape results only if accumulated debris was originally isolated in nonlinear forms (see fig. 10.14), because sediment deposited in linear openings is ultimately converted to ridges rather than mounds. In addition, some kames form when debris collects as fans or small deltas built against the ice or outward from the ice, with the apex resting at the stagnant margin. In either case, melting of the supporting ice allows the drift to collapse into a kame, a process evidenced by slumped strata within the deposits.

Kames are only one of many forms with essentially the same origin. They are transitional into eskers or minor ridges and circular forms that Parizek (1969) calls ice-contact rings and ridges, or into other features whose names utilize the term "kame" as a descriptive adjective, such as kame delta, kame moraine, and so on. Perhaps the most common of the latter type of feature is a *kame terrace.* Kame terraces originate from drift deposited in narrow lakes or stream channels between the valley side and the lateral edge of the stagnating ice. When the supportive ice disappears, the inner edge of the deposit collapses into the terrace scarp. Kame terraces differ from normal river terraces in that they are restricted in their longitudinal extent and are usually narrow. The tread may slope gently into the valley, and the surface may be dimpled by kettle holes (McKenzie 1969).

*Kettle holes* are circular depressions that are formed in a variety of ways (Fuller 1914) but most by the burial and subsequent melting of isolated blocks of ice in stratified drift. The gradual ablation of the ice leads to a gentle downward flexing of the sediment layers as they settle over the dissipating mass. Some kettles are almost 50 m deep and up to 13 km in diameter (Flint 1971), but these giants are exceptions to the normal kettle size of less than 8 m deep and 2 km wide. Kettles usually form in association with kames and other related features, producing an irregular surface described as kame-and-kettle topography, but similar surfaces can develop in the absence of either or both of these features.

*Eskers* The term "esker," evidently stemming from the Gaelic word for "crooked" or "winding," has been applied to a wide variety of ridged ice-contact features (see fig. 10.15). *Eskers* range in shape from the single, narrow, sinuous ridge that is the classic form to a complexly intertwined maze of branching and joining ridges (Huddart and Lister 1981; Shreve 1985; Gorrell and Shaw 1991). Eskers seem to form most commonly in stagnating margins of large ice sheets where the underlying surface is broad and relatively flat. In addition, they are generally, but not always, associated with temperate ice and an abundant supply of meltwater (Fitzsimons 1991). The ridges are not necessarily continuous but may consist of crudely connected linear segments. In broad valleys eskers usually parallel the slope of the valley floor, but this is not always the case. Some ridges, for example, ascend the valley sides, transect the divide, and descend the flanks of the adjacent valley (Shreve 1985).

Eskers show dramatic inconsistencies in dimension. They range from 2 m to more than 200 m in height, from several meters to as much as 3 km in width, and from tens of meters up to 500 km in length (if gaps are considered in the total distance). In cross-profile, they usually have rather sharp crests and steeply inclined sides (up to 30°), but broad eskers can maintain rather flat upper surfaces or may be pitted by kettle development. R. J. Price (1973) suggests that the height and width of eskers may be directly related to their overall length, longer ridges being proportionately higher and wider than shorter ones.

It seems certain that eskers result from sediment accumulation in a variety of openings, such as (1) ice channel fillings (e.g., crevasse fillings, fillings between stagnant blocks), (2) tunnels beneath or within the ice, (3) supraglacial channels, and even, in rare cases (4) narrow longitudinal embayments of the ice front (Cheel 1982). The subglacial origin is made possible by meltwater that descends from the ice surface to the base through fractures and holes in the ice. Along the subglacial surface a network of interconnected tunnels passes water and sediment toward the terminus, and normal fluvial deposition fills or partially fills the openings. Filling probably takes place during the last stage of deglaciation when internal ice flow is minimal or stagnant (Shulmeister 1989; Shreve 1985). In addition to forming in subglacial channels, field studies and photographic analyses of esker development by the Casement Glacier Alaska (Price 1966; Petrie and Price 1966) and the Breidamerkurjokull glacier in Iceland (Price 1969) provide rather conclusive evidence that englacial and supraglacial deposits can be transformed into eskers. In those areas, some of the eskers are definitely ice-cored and were measurably lowered in elevation during the period 1948 to 1963 by wastage of the buried ice. Presumably the final melting of the ice will rest the esker on the original subglacial floor.

## Interior Ice-Contact Features

Behind the marginal zone in the interior portion of a glacier, the predominant features are deposited at the base of the ice. There pressure from the overlying ice either spreads till rather evenly across the ground surface or molds water-soaked till already in that position into distinct morphologic shapes. Supraglacial drift is somewhat rare in the interior zone, but where two valley glaciers join, the debris dragged along their lateral edges may coalesce into *medial moraines.* Although these moraines appear as striking linear belts on the surface of the ice (see fig. 10.6), they are superficial in that the deposits are shallow. Not all of these linear

**FIGURE 10.21**
Flat till and loess plain in southern Illinois. Gently undulating topography on ground moraine has been smoothed by influx of younger loess.

features represent former ice margins. Some may have an interior origin (Small et al. 1979) and sediment sources that are both surficial and englacial (Eyles and Rogerson 1978). Medial moraines are rarely preserved on the ground surface because they are let down in the middle portions of valleys where meltwater streams are likely to destroy them. The dominant interior ice-contact features are those deposited beneath the ice such as ground moraines, drumlins, and fluted surfaces.

**Interior Moraine** In contrast to end moraines, **ground moraine** is distinguished by its apparent lack of topographic expression. It is accepted as a moraine despite its low relief and a complete absence of transverse ridges (Flint 1971) because its surface expression is independent of the topography it covers. Ground moraine occupies much of the surface covered by major Pleistocene ice sheets in North America and Europe. The moraines usually exist as smoothly undulating plains, like that in figure 10.21, seldom exceeding 10 m in total relief. They range in size from small areas interspersed among younger marginal features to regions covering thousands of square kilometers behind the terminal moraine.

Although in most cases the primary building material of ground moraine is lodgement till, the two terms are not synonymous. The deposits from which ground moraine is constructed also include ablation till and interbeds of fluvioglacial deposits that originate in the same glacial advance (R. P. Kirkby 1969; Boulton 1972a, 1972b) or in more than one episode of glaciation.

In addition to producing the rather featureless ground moraine, most, if not all, of the processes that function at the glacier front can create moraine ridges subglacially (Sugden and John 1976). For example, large transverse ridges, called *Rogen moraines* or *ribbed moraines*, are 10 to 30 m high, > 1 km long, and tend to develop as a series of separate hills spaced 100 to 300 m apart. They probably form by shearing or thrusting behind the glacier front, usually in broad depressions of older till sheets or in local bedrock valleys. Significantly, the ridges often occur in direct association with flutes or drumlins; therefore, it is reasonable to assume that they all have a common origin. This association suggests that the ridges reflect an interaction of basal debris, pore water pressure, and ice temperature and that the Rogen system develops where transverse variations in stress exist at the glacier bed (Sugden and John 1976).

The mechanics associated with Rogen moraines are remarkably similar to the thrusting phenomena proposed by Moran and his colleagues (1980). As discussed earlier, the thrusted blocks seem to occur along the inner part of the marginal zone and often assume a morainic topography. This topography may be subdued, however, where temperate ice slides over the blocks.

The pushing and squeezing phenomena may also relate to smaller varieties of subglacial moraines. Within the marginal zone the locus of deposition shifts periodically, giving rise to ridges, mounds, and depressions of varying size. The ridges in the sequence, usually small and composed of till, commonly parallel the orientation of the ice front. They have been given many names, such as *cross-valley moraines* (Andrews 1963; Andrews and Smithson 1966) and *washboard moraines* or *moraine ridges* (R. J. Price 1970), and several theories have been suggested for their origin (Elson 1968). The ridges, instead of exhibiting a regular undulation of ridge crests and intervening sags, are usually segmented or interwoven with many deposits of stratified drift, giving the entire topography a chaotic expression.

**Fluted Surfaces and Drumlins** As indicated above, the monotony of ground moraine topography is sometimes broken by ridges or elongate hills that manifest the mechanics functioning at the base of glaciers, especially in temperate ice sheets. The most common features developed in the subglacial environment are fluted surfaces and drumlins. Flint (1971) refers to them as streamlined molded forms but others (Sugden and John 1976) consider them as special types of moraine ridges that form parallel to the direction of ice flow.

**Fluted surfaces** consist of narrow, regularly spaced, ridges. The ridges are normally less than 5 m high and several hundred meters long, although individual ridges may be considerably larger. Small ridges are usually composed of till, and their origin may be related to pressure-squeezing of saturated debris into longitudinal cavities at the base of the ice, a process similar to that forming some of the minor moraine ridges. There is no question that till is mobilized and arranged in ridges where cavities open on the lee side of large boulders (Hoppe and Schytt 1953; Sharp 1985). In fact, Boulton (1976) defines flutes in a genetic sense as being formed when deformable subglacial material is intruded into ice tunnels on the lee side of boulders or other ridge obstructions. This occurs because unloading in the lee of obstructions sets up a pressure gradient in the till that causes it to flow into the cavity.

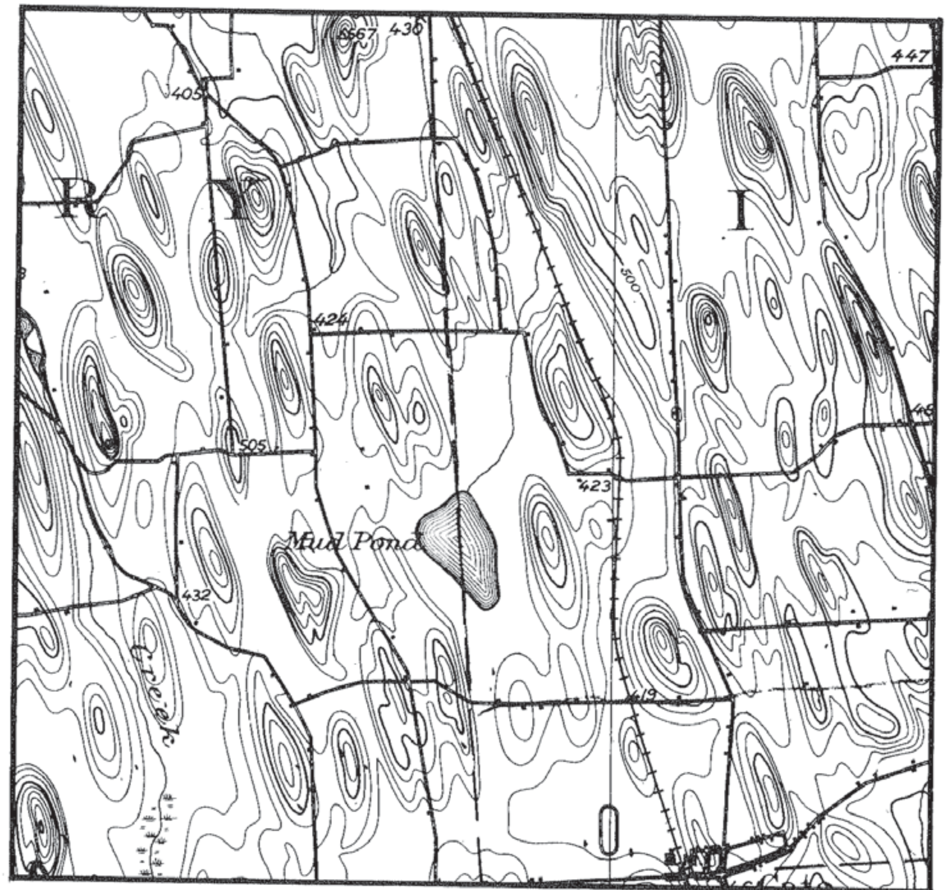
Some larger ridges are composed of material other than till (Lemkke 1958; Gravenor and Meneley 1958), and the surfaces between adjacent ridges are often noticeably grooved. These characteristics have led many workers to believe that the origin of fluted surfaces may be related not only to the deposition of ridges but to erosion of grooves into the subglacial materials, or a combination of the two processes. Gravenor and Meneley (1958), for example, suggest that alternating high- and low-pressure zones in the basal ice produce the unique fluted surface. In their model, grooving occurs in the material beneath the high-pressure zones, and debris eroded from there is moved not only downglacier but also upward into regions of low pressure. Boulton (1976) rejects the notion that fluted surfaces require periodic pressure distributions in the ice and suggests that the features represent postdepositional deformation of preexisting materials. As such, they are neither erosional or depositional.

**Drumlins** also are elongated parallel to the direction of ice flow, their long axes deviating only slightly from the average trend of the glacier movement (fig. 10.22A). They have been described as having a plan view shape that is similar to leminscate loop (Chorley 1959) or an ellipsoid (Reed et al. 1962) and in long profile a form that Flint (1971) likens to an inverted bowl of a spoon (fig. 10.22B). The exact shape, however, is probably variable enough that no particular model will fit all drumlins. In any case, drumlins are higher and wider near their rear edges and they narrow and thin downstream until they merge imperceptibly with the surrounding surface. Drumlins average in size from 1 to 2 km in length and from 400 to 600 m in width, and stand anywhere from 5 to 50 m high; individuals can be smaller or larger. Their length/width ratio seems to be reasonably consistent, ranging from 2 to 3.5 (Reed et al. 1962; Vernon 1966; Trenhaile 1971), although length/width ratios exceeding 6 have been measured (Boyce and Eyles 1991).

The formation of drumlins has received considerable attention during the last century, partly, as Menzies and Rose (1989) point out, because an understanding of their development provides insights into the processes operating at the base of the glacier. Nevertheless, their origin is not fully understood and remains an issue of controversy.

**FIGURE 10.22**

(A) A portion of the drumlin field located near Weedsport, N.Y. From the northeast quarter of the Weedsport, N.Y., quadrangle (U.S.G.S. 15′). Contour interval 20 feet.
(B) Drumlin shown in longitudinal profile near Rochester, New York.




(A)

(B)

Any hypothesis concerning the origin of drumlins must consider both their sedimentologic and spatial character. In terms of sedimentology, drumlins display a variable internal composition. Many are fabricated entirely from clay-rich till, but others have obvious cores of solid rock or preexisting drift that may or may not be stratified. Geographically, drumlins are located in only a small number of glaciated areas, but where they exist, they rarely occur as individuals but instead cluster together in fields that are commonly wider than most morainal belts (Gravenor 1953). The density of drumlins within a field seems to be inversely related to size; that is, very dense clusters are composed of relatively small drumlins (Doornkamp and King 1971). Some fields may contain as many as 10,000 individuals (Sharp 1985). In addition, many studies show a strong probability that drumlins within any field are spaced in a nonrandom manner; that is, spacing between neighboring individuals is somewhat regular (Reed et al. 1962; Vernon 1966; Smalley and Unwin 1968; Trenhaile 1971). Drumlin fields also seem to be located in zones close to but behind the terminal moraines that mark the limit of a particular glaciation.

The models used to explain drumlin genesis fall into four main groups: (1) drumlins are erosional features developed when moving ice streamlines preexisting drift or rock (Gravenor 1953; Kupsch 1955; Lemke 1958; Whittecar and Michelson 1979; Bouchard 1989; Aylsworth and Shilts 1989; Habbe 1992); (2) drumlins are depositional features formed when a moving glacier deposits till and, as a result of spatial differences in debris rheology caused by dilatancy, pore water dissipation, localized freezing, or grain size, locally molds the material (and perhaps preexisting sediment) into streamlined forms (Smalley and Unwin 1968; Boulton 1979, 1987; Aylsworth and Shilts 1989; Menzies 1989; Boyce and Nicholas 1991; Hanvey 1992); (3) drumlins are depositional features derived from the infilling of cavities carved into the basal ice by meltwater (Shaw et al. 1989; Shaw 1983, 1985; Shaw and Kvill 1984; Sharpe 1987; Hanvey 1992), and (4) drumlins are erosional remnants produced by the catastrophic release and flow of subglacial meltwater (Shaw and Sharp 1987; Shaw et al. 1989; Shaw 1989). There is reasonable evidence to support each of these theories and we should probably accept a multiple origin for drumlins, as Embleton and King (1975a) suggest, and not pay undue attention to hypotheses that utilize one process to the exclusion of all others.

## Proglacial Features

The large volume of water released from glaciers carries with it a tremendous quantity of sediment that is deposited in a number of environments beyond the margin of the ice. This debris usually accumulates in stream channels and associated floodplains that, because of their continuous lateral shifting, spread the sediment into a large plain called a **sandur** (from Icelandic; plural *sandar* or *sandurs*). Downstream from the sandar, the meltwater streams may empty into bodies of standing water and construct deltas, beaches, and other geomorphic features from the fluvioglacial sediment. These forms do not differ appreciably from their counterparts developed in normal fluvial, lacustrine, or marine environments and so will not be discussed here. Sandar are somewhat analogous to alluvial fans, but they are unique in that the hydrology of the streams that form them is controlled by intense seasonal variations in the melting of ice. In addition, because each glacial advance develops its own related sandur, the surface of which is generally located at a different elevation, the distribution and age of the alluvial surfaces are extremely helpful in unraveling the Pleistocene history of a glacier. Thus, the origin of sandar and the geomorphic criteria for recognizing their form in ancient settings deserve our attention.

The study of sandar began in the nineteenth century in Europe and North America. They are now recognized as consisting of two primary types (Krigstrom 1962). A **valley sandur,** which originates within well-defined valleys, is created by one main river and its anabranches; the entire system rarely occupies the total valley bottom at any given time. In the United States, valley sandar have been called **valley trains** and are usually associated with individual mountain glaciers (fig. 10.23). The second type of sandur, called a **plain sandur,** differs in that it develops with no lateral constraints but represents the form of a massive plain. In North America, these are often referred to as **outwash plains** and are usually associated with large ice sheets.

Most sandar are composed predominantly of gravel, although the deposits may include lenses of sand. A general decrease in particle size is sometimes, although not always, apparent in the downstream direction (Fahnestock 1963; Church 1972). The deflation of very fine material on active sandar that are

**FIGURE 10.23**
Valley sandur in front of the Scott Glacier, Alaska.

unvegetated helps to produce a coarse-grained surface and simultaneously produces loess (see chapter 8).

In general, the long profiles of sandar are similar to river profiles, being concave-up in form and expressible as a simple exponential function (Church 1972). The concavity, however, may not be perfect because of the presence of linear segments similar to those characterizing alluvial fans. Cross-profiles tend to be convex-up, but the exact form may be irregular or may slope continuously in one direction. In addition, the cross-profile shape seems to depend on where it is measured in relation to the ice margin.

Krigstrom (1962) has recognized on sandar three distinct zones relative to the ice front, called proximal, intermediate, and distal zones, each of which has different surficial characteristics. The *proximal zone,* closest to the ice, is usually traversed by only a few main rivers that flow in well-defined entrenched channels. These rivers may originate beneath or within the ice (Gustavson and Boothroyd 1987) or pass continuously onto the ice mass itself. In some localities, the stratified drift extends onto and buries extensive areas of stagnating ice. As the ice subsequently melts, kettle holes form and the proximal sandur takes on a rough, pitted configuration. These kettled sandar (or pitted outwash plains) are difficult to place in our framework classification because they form in the marginal ice-contact environment, but they are continuous with the surface of the proglacial sandur and develop with the same original mechanics (described in Price 1973). In addition, on many sandar the proximal surface stands well above the elevation of rivers emerging from the ice. Several hypotheses have been suggested to explain why rivers in the proximal zone are so entrenched in their own deposits: (1) the rivers may be regrading in response to hydrologic and load characteristics (Fahnestock 1969), parameters that vary significantly in surging glaciers (M. Sharp 1988); (2) the sediment is supraglacially deposited and left elevated as the ice front recedes and rivers emerge at a continuously lowered level; (3) modest incision may be normal in the proximal zone, with the sandur surface simply representing the high flow level; or (4) the channels may be downcutting as the ice margin is uplifted by glacitectonic processes. It is conceivable that each of these interpretations is correct.

In the *intermediate zone,* the channels become wide and shallow and distinctly braided, and the entire depositional network shifts its position rapidly from side to side. This active lateral migration leaves a maze of abandoned channels with a relief of one or two meters impressed on the surface topography of the plain. Commonly the main channel is aggraded to a higher level than the smaller channels, facilitating rapid changes in the position of the river. Downstream the system changes gradually into the *distal zone,* where channels become so shallow that the rivers may merge into a single sheet of water during high flow. The flow here commonly feeds deltaic growth when the river enters a body of standing water. However, the sandur may extend itself downstream by growing over the rear portion of the delta, while the deltaic front simultaneously progrades (Church 1972).

Sandar originate from the combined effects of a large sediment supply and the high floods associated with melting ice. Most of the abundant load is derived from older drift, morainal deposits, and the continual delivery of new debris to the ablation zone and its release from the ablating ice. In some cases, however, the debris may be transported to the proglacial environment by meltwater flowing in subglacial tunnels (Gustavson and Boothroyd 1987). The greatest fluvioglacial work occurs near the ice margin where floods are produced by summer melting or as sudden releases of lake water dammed within the ice called *jokulhlaups* (from an Icelandic word pronounced "yokel-lawp"). These floods are characterized by a rapid and drastic increase in discharge (Church 1972; Waitt 1980; Booth and Hallet 1993).

The bulk of aggradation on a sandur takes place during high flow events as channel fills, sandur levee deposits, and overbank sedimentation. Overbank deposits are more prevalent in the intermediate zone where channels are shallower and interchannel reaches are covered more frequently by floods. Although high flow does initiate pronounced channel scouring, the amount of aggradation during the peak and waning stages of a flood simply obliterates the scour channels. Thus, aggradation may be rather rapid. For example, Fahnestock (1963) measured a net elevation gain of 0.36 m in a two-year period on the sandur produced by the Emmons Glacier (Washington).

In general, then, sandar can be considered as transport surfaces that aggrade during high flows but are probably eroded and changed in form when discharge and load are at normal volumes. The seasonal variations in load and discharge may also be accompanied by changes in the river pattern (Fahnestock 1963). Therefore, the ultimate size and properties of a sandur are probably related to a quasi-equilibrium condition established by the balance between meltwater volumes and the quantity and size of the sediment made available for transportation. The surface will always be a montage of flood sediments, but the exact topography will change incessantly.

## Summary

In this chapter we examined the landforms developed by the process of glacial erosion and deposition. Erosional features range in size from minor embellishments of exposed bedrock to major forms that dominate the landscape. Minor features such as striations, grooves, roche moutonnées, and friction cracks are a function of the subglacial mechanics that control abrasion and plucking. Major erosional landforms develop in two environments. In mountain uplands, the expansion of cirques produces features such as arêtes and horns that give glaciated mountains their characteristically rugged appearance. Cirques are created by nivation and rotational sliding of cirque glaciers. Cirques increase in size by erosional retreat of their walls which is facilitated by frost shattering or possibly by hydration shattering. Glaciated valleys are created by large-scale abrasion and plucking that cause the dominant staircase longitudinal profile and the U-shaped cross-profile.

Deposits associated with glaciation, called drift, consist of either stratified or nonstratified material. Stratification requires that some sediment be transported by meltwater after the debris is released from the ice. Nonstratified drift is deposited directly by the ice. Certain types of depositional landforms tend to accumulate in particular geographic positions with respect to the ice front. In the marginal zone, moraines and stagnant ice features are most common. In the interior zone behind the ice margin, ground moraine, fluted surfaces, and drumlins are most conspicuous. Downstream from the glacial margin, in the proglacial zone, all drift is fluvioglacial and usually accumulates in the form of large plains called sandar. Processes operating in each of these depositional environments have been discussed with regard to how they might generate the landforms developed in the specific regions.

## Suggested Readings

The following references provide greater detail concerning the concepts discussed in this chapter.

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