GLACIAL FIGURES

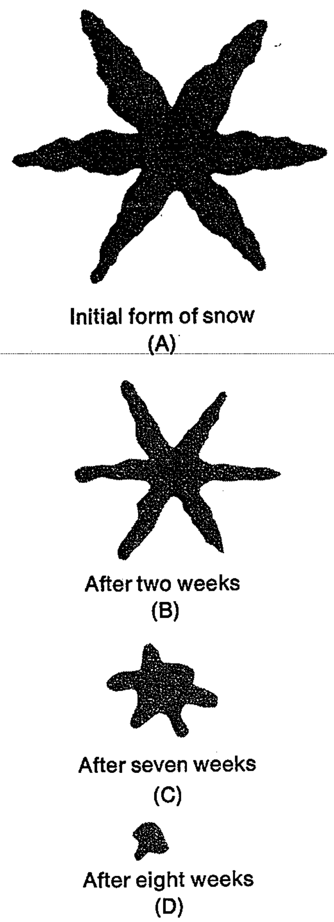

Figure 9.1
Changes in the shape of snow crystal in the transition to firn.

TABLE 9.1 Increasing Density of Snow During Transition to Ice.

Materials	Density (gm/cc)
New snow	0.05–0.07
Firn	0.4–0.8
Glacier ice	0.85–0.9

glacier ice (table 9.1). The rate of densification beyond this point is much slower. In addition, the mechanics involved in transforming firn to ice differs from earlier processes of densification. This transformation produces enlarged ice grains (in some cases up to 10 cm long) by recrystallization (fig. 9.2). Gradually, pore space is eliminated by crystal growth or by freezing of the downward-permeating meltwater until the density is approximately 0.8 to 0.85 gm/cc. The only air remaining is trapped as bubbles within the crystal. For all practical purposes, the creation of ice is complete at this stage, even though the bubbles continue to be slowly expelled by compaction, and the density may increase to about 0.9 gm/cc.

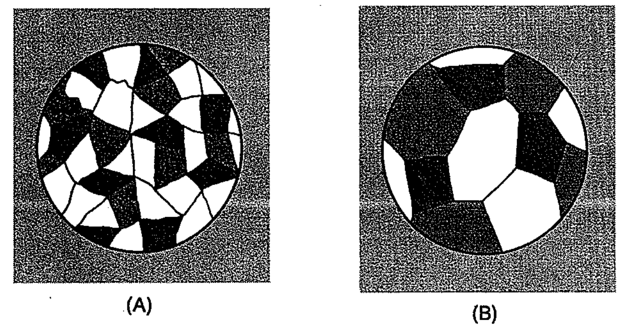

Figure 9.2
Increase in grain size caused by recrystallization process. (A) Original texture of water-soaked snow. (B) Texture after application of stress.

The rate of density increase and crystal growth is closely related to the temperature of firn and the rate of increased load on the individual grains. Firn in temperate regions will transform into ice faster and at a shallower depth than it will in polar regions. This fact is probably best demonstrated by comparing the depth-density relationship exhibited by glaciers in temperate and polar regions (Behrendt 1965). In the Antarctic, a 0.85 gm/cc density is not achieved until at least 80 m of firn is accumulated. In contrast, only 13 m is needed to change snow to ice on the temperate Upper Seward Glacier in Alaska. Because accumulation rates are greater in glaciers of temperate regions, it follows that the greater depth to true ice in the Antarctic ice sheet represents the monumentally greater time needed to create glaciers in dry, cold climates.

Glacier ice is not preserved intact after its creation but is susceptible to further changes with continued increase in stress. For example, experimental work shows that the random orientation in polycrystalline ice cannot be maintained at higher stresses. When deformation of the polycrystalline mass exceeds several percent, the process known as recrystallization begins. Theoretically, this process should align the *c*-axis of the ice crystal within the vertical plane of a glacier, altering what is known as the *ice fabric.* Changes in fabric (*c*-axis orientation) with increasing depth and shear stress have been noted in a number of glaciers (Gow and Williamson 1976; Russell-Head and Budd 1979; Hooke and Hudlestone 1980, 1981). In many cases, the fabric changes from a weakly oriented *c*-axis in fine-grained, near-surface ice to coarser ice with a broad, single maximum fabric at some lower level. At still greater depths, the fabric usually develops multiple orientations. Finally, in some, but not all, glaciers the fabric returns to a single maximum orientation at the glacier base (Hooke and Hudleston 1980).

The depth at which fabric transition occurs is highly variable and depends on a complex interaction of density, impurities, temperature, and so on. In fact, a gen-

TABLE 9.2 Morphological, Dynamic, and Thermal Classification of Glaciers.

Morphological classification	
Major types	
Cirque glaciers	Flowing ice streams restricted to amphitheater-shaped depressions in valley headlands
Valley glaciers	Streams of ice that flow downvalley well beyond the cirque
Ice sheets	Broad, flowing ice masses that are not confined to valleys; ice accumulates to massive thickness in high-latitude continental areas or broad uplands
Intermediate Types	
Mountain ice sheets	Valley glaciers that enlarge to form ice sheets that bury all but the highest Alpine peaks
Piedmont glaciers	Glaciers that discharge ice onto broad lowlands located along the base of mountains
Dynamic classification	
Active glaciers	Glaciers characterized by high rates of ice movement from accumulation zones to their terminus
Passive glaciers	Glacier exhibiting low rates of ice movement from accumulation zones to ablation zones
Dead glaciers	Glacier possesses no discernible internal ice flow
Thermal classification	
Temperate glaciers	Glacier in which the ice is at or near its pressure-melting point throughout the ice mass
Polar glaciers	
Subpolar glaciers	Glacier in which the ice mass is generally below the pressure-melting point except for summer melting of the upper few meters
High-polar glaciers	Glacier characterized by an ice mass that is below the pressure-melting point at all times

eral model describing the formation of ice fabric has yet to be developed (Alley 1992). Nevertheless, the *c*-axis orientation appears to be related to the overall scheme of glacier movement because the orientation of basal crystal planes is probably parallel to the direction of ice flow (Rigsby 1960).

Glaciers have been classified on the basis of many salient properties, but as with other physical phenomena, the appropriateness of the classification depends on the prime purpose of the groupings. Ahlmann (1948) suggested morphological, dynamic, and thermal criteria for classifying glaciers (table 9.2). The morphological classification, which is based on glacier size and the environment of its growth, is most commonly employed, and the different categories, such as valley glaciers and ice sheets, are undoubtedly familiar. Although Ahlmann (1948) recognized many morphological subdivisions, Flint (1971) suggested that glaciers can be placed in three broad categories—*cirque glaciers, valley glaciers,* and *ice sheets*—and two intermediate categories—*piedmont glaciers* and *mountain ice sheets.* In this abbreviated form the classification is quite useful, especially in a descriptive sense.

For our process orientation, however, two other classifications are probably more useful. The dynamic classification is based on the observed activities of glaciers and consists of three main groups: *active, passive,* and *dead* glaciers. Each type reflects a different balance between losses and gains of ice and probably somewhat different thermal properties. Active glaciers, like the one in figure 9.3, are characterized by the continuous movement of ice from their accumulation zones to their terminus. The movement may occur in response to normal snow accumulation, or it can be generated by avalanches or ice falls that provide the impulse for the forward motion. As you might expect, a glacier is passive when its movement is minimal. Dead ice has no discernible internal movement.

The thermal classification is based primarily on the temperature of the ice. Glacial behavior is directly influenced by ice temperature, and therefore the thermal classification should lend itself to process analysis. Two types of glaciers exist in this classification: *temperate glaciers* and *polar* or *cold glaciers.* In *temperate glaciers* the ice throughout the entire mass is at its pressure-melting point (the temperature and pressure conditions at which ice begins to melt), although the upper 10 m may freeze in the winter. Meltwater seems to be present in abundant amounts within or beneath the ice mass and, in contact with the underlying rock, often facilitates slippage of the ice over the bed. This causes velocity and erosive action to be generally greater in temperate glaciers than in other types. Near the snout meltwater may emerge as basal streams, or it may be temporarily dammed within the ice; both situations promote extensive fluvial removal of debris from the terminus of the glacier.

Polar glaciers were subdivided by Ahlmann into *subpolar* and *high-polar* types. In *subpolar* glaciers the

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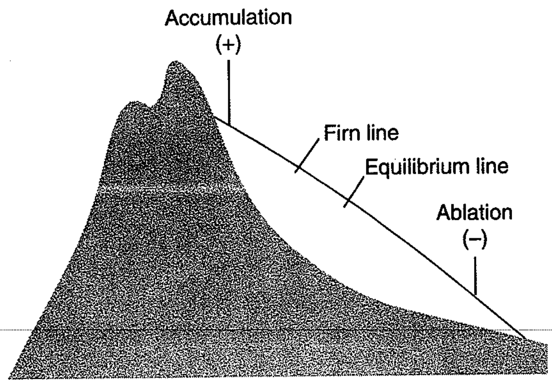

Figure 9.5
Zones of accumulation and ablation on a glacier as determined by budget analyses.

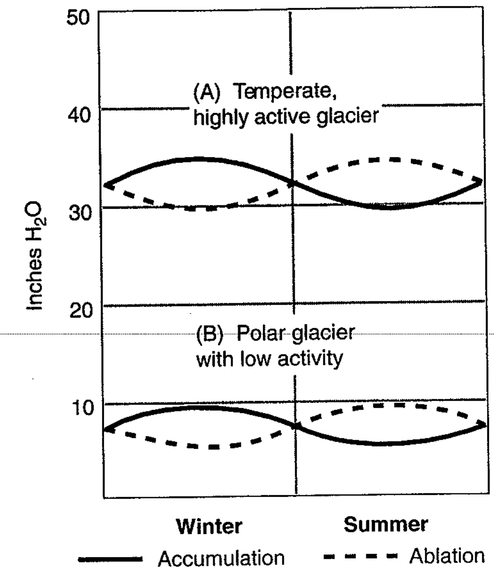

Figure 9.6
Annual budgets on two glaciers in which the net mass budget of each equals zero. Active, temperate glacier (A) has greater total amounts of ablation and accumulation than the low-activity polar glacier (B). Even though both are in equilibrium and their fronts are stationary, (A) does more work because it must transfer more ice during the year from the accumulation zone to the snout.

and, as such, represents a net accumulation of mass. This *superimposed ice* is found in enough cases to warrant the distinction between the firn line and the equilibrium line (fig. 9.5), and it sometimes creates a rather complex transitional zone between the accumulation area and the ablation area. Nevertheless, equilibrium lines and firn lines are usually close enough to approximate one another.

What does all this have to do with glacial mechanics and the resulting processes of erosion and deposition? If glaciers are viewed as open systems, both the mass balance and the absolute total amounts of accumulation and ablation are important factors in the character of glacial movement. When the net budgets are perfectly balanced (total mass balance equals zero), no expansion or shrinkage of the glacier occurs and the glacial extremities remain stationary. This equilibrium condition, however, is seldom maintained for a long period of time, and so the glacial front and sides usually fluctuate constantly. Glaciers with a positive mass balance actively advance and characteristically maintain a relatively steep or vertical front. In contrast, an overall negative budget induces a recession of the glacial front, and the snout will be gently sloping and often partially buried by debris released from the ice. The mass balance, therefore, is closely related to the position and type of morainal system constructed by the ice.

The absolute amount of snow and ice relative to the area of a glacier (gross accumulation and ablation) directly affects its internal activity. Large gross accumulation values on small glaciers promote very rapid flow from the accumulation area to the ablation area; in most cases temperate glaciers are likely to have large accumulation and ablation values and, consequently, high flow velocities. In contrast, polar glaciers, which tend to be rather passive, usually have small values of accumulation and ablation and low internal flow velocities. It is very important to recognize that glaciers with different gross values can have the same mass balance, and if they are in an equilibrium condition, the glacier front does not advance or recede even though the internal transfer of ice may be enormous. Figure 9.6 is a hypothetical diagram of annual budgets for a polar glacier with low activity and a highly active temperate glacier, both of which have a mass balance equal to zero. The snouts and lateral edges are stationary in both cases, but the temperate glacier, having a large value for accumulation and ablation, is moving at a higher rate internally. This dynamic activity causes pronounced erosion and rapid transportation of debris through the system. With the proper budgets, large moraines may result at the terminal and lateral boundaries. The polar glacier depicted in figure 9.6 will have little if any internal motion of its ice and, as a result, will probably not form any significant depositional features. The depositional and erosional character of every glacier, therefore, is fundamentally determined by the characteristics of the snow and ice added to or lost from its surface.

The Movement of Glaciers

Several hundred years ago, through direct observations, Alpine residents realized that glaciers move. Measurements of the rate of glacier flow were made as far back as the early eighteenth century. It is now generally accepted that glaciers move by two mutually independent processes: internal deformation of the ice, called *creep*, and *sliding* of the glacier along its base and sides. In addition, where glaciers override a layer of unconsolidated debris, the ice mass may move along with the underlying sediment as it deforms or flows beneath the glacier.

Internal Motion

Basic Mechanics The first real attempts to explain the physical dynamics of englacial ice motion are found in the work of J.D. Forbes, who in 1843 suggested that glaciers respond to stress much like a substance undergoing plastic deformation. By the beginning of the twentieth century, numerous models based on a plastic flow mechanism had been developed, although other ideas were also being proposed. They included flow caused by differential shearing along numerous, closely spaced planes (Phillip 1920; Chamberlain 1928) and the widely accepted notion that ice behaves as a viscous liquid that deforms in linear proportion to stress (Lagally 1934). Each of these concepts suggests that glacier flow manifests a relationship between ice deformation (strain) and the stress (force/unit area) that produces it. Stress generated at any point can be separated into two parts, hydrostatic pressure and shear stress. *Hydrostatic pressure,* which is related to the weight of the overlying ice, is exerted equally in all directions. *Shear stress,* which causes glacier motion, is a function of the overlying weight and the slope of the glacier surface. For the purposes of demonstration, consider the two simplest possibilities: (1) that ice behaves as a viscous fluid and (2) that ice deforms as a perfectly plastic substance.

If ice behaves as a Newtonian viscous material, and we assume a constant viscosity, the application of stress should result in a linear relationship between the stress value and the strain rate (fig. 9.7). In addition, deformation will begin in the ice as soon as stress is applied and will maintain the linear proportionality regardless of the changes occurring in the stress. In contrast, a plastic substance shows no immediate response to stress but is capable of supporting a certain amount of stress without sustaining any deformation (fig. 9.7). Thus, at low stresses the strain of a plastic will be zero. As stress increases, however, it eventually attains a value, called the *yield stress,* where the ice will experience limitless and permanent deformation. An entire glacier would behave plastically only if the shear stress along its base were equal to the yield stress.

Neither of these simple cases prevails in the mechanics of ice motion. However, in many instances glacial properties predicted according to pure plasticity closely approximate the observed characteristics (Nye 1952b), and thus glacial movement is often referred to as pseudoplastic (Meier 1960; Johnson 1970). Laboratory studies since the late 1940s have shown that single ice crystals under stress deform as soon as stress is applied. Under any given stress the strain will increase

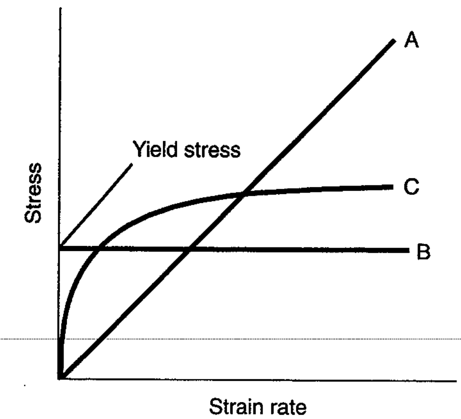

Figure 9.7
The relationship between stress and deformation.
(A) Newtonian viscous fluid. Relationship is normally plotted as a function of stress and the rate of strain. (B) Perfect plastic deformation. The material experiences no deformation until stress is increased to the value of the yield stress. The material will then continue to deform as long as the stress is applied. Normally plotted as a function of stress and strain. (C) Ice creep according to Glen's flow law.

rapidly at first (Glen 1952, 1955), but within a short time (tens of hours) it approaches an almost steady value that plots as a nearly straight line on a graph (fig. 9.7). This continuous deformation with no increased stress is the creep process that allows glacial ice to flow steadily under its own weight.

The rate of strain during the creep process is related to varying stress values by the following equation derived by Glen (1952, 1955) and now commonly referred to as the *power flow law:*

$$\dot{\epsilon} = k\tau^n$$

where $\dot{\epsilon}$ is the strain rate ($d\epsilon/dt$), τ is stress, and k and n are constants. The values of n, determined by a number of investigators, seem to vary from approximately 2 to 4 for individual crystals. In polycrystalline ice they range from 1.9 to 4.5, with the mean value being close to 3. In any case, the n values associated with the power flow law are significantly greater than 1, the value required for linear viscous flow. The power flow law indicates that minor changes in stress can produce a major response in the strain rate. For example, if $n = 4$, doubling of τ increases the strain rate 16 times.

Glacier ice is always polycrystalline. The flow law, although based on the deformation of a single crystal, still seems to predict the response of glaciers (Thomas et al. 1980), although minor modifications of the equation may be needed (Meier 1960; Colbeck and Evans 1973). The value of k in glacier flow might be reduced by interference of adjacent grains and recrystallization. In addition, as shown in figure 9.8, cold ice deforms less readily than temperate ice, mainly because the constant k is

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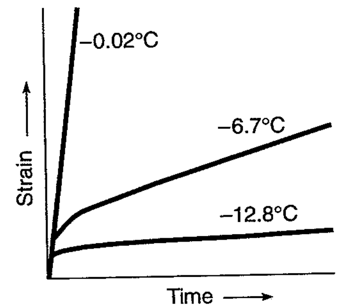

Figure 9.8
The effect of temperature on the creep deformation curve. All curves represent deformation under a stress of 6 bars.
(After Glen 1955)

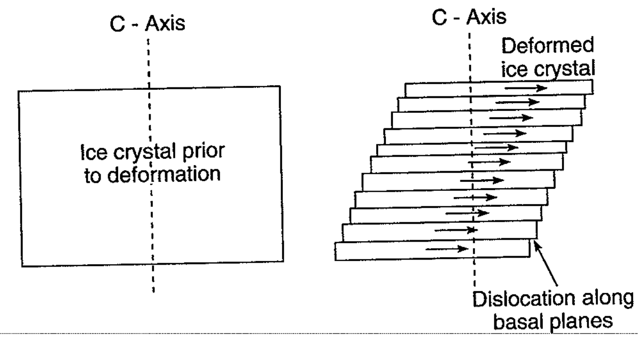

Figure 9.9
Ice deformation involving the dislocation of atoms along the basal planes of an ice crystal. As crystals deform, recrystallization occurs, reestablishing the equidimensional shape of the grains.

also dependent on temperature. In Glen's experiments the value of k decreased by two orders of magnitude (from 0.17 to 0.0017) when the temperature was lowered from 0° C to −13° C. In fact, Paterson (1969) suggests that the strain produced in ice at −22° C by any stress is only 10 percent of its value when the ice is at 0° C. Impurities contained within the ice, including gas bubbles, dissolved constituents (ions), and rock debris, are also known to influence strain rates (Benn and Evans 1998). The effects of rock debris on ice deformation have received the most attention because ice at the base of many glacier contains considerable amounts of sediment. The influence of solid particles on deformation is still unclear, however, as the observed effects on strain rates differ between individual studies (see, for example, Nickling and Bennett 1984; Hubbard and Sharp 1989; Echelmeyer and Wang 1987). In some cases, solid particles appear to harden the ice, in others they soften it. Benn and Evans (1998) point out that these variations are probably due to the numerous factors that are involved in controlling the deformation process including the grain size of the sediment, the orientation between the debris and the ice crystals, and the temperature-related melting of ice around individual particles.

Given the above, it should be clear that ice does not behave like a viscous fluid, although it may approximate a viscous response under low stress when the strain rate is still in its transient phase. At stress <1 bar in temperate ice, for example, n values as low as 1.3 have been measured (Colbeck and Evans 1973). However, when nearly steady strain is attained or the ice is under high stress, the deformation becomes more plastic. Although this condition approximates plasticity, no distinct yield stress is associated with the creep process, indicating that ice is not a perfectly plastic material.

Much research has been done to determine why ice under high stress takes on a nearly plastic behavior. The answer probably rests in the mechanics of strain. Ice deforms by a process involving the dislocation of atoms along planes of weakness inside individual crystals (see Glen 1958, 1987; Weertman and Weertman 1964). The dislocation process works most efficiently through gliding on the basal planes of the ice crystal (fig. 9.9), the movement somewhat resembling the slip observed in a stack of playing cards (Sharp 1988). If deformation were allowed to continue, it would produce sheetlike ice grains. As the crystals deform, however, recrystallization occurs, reestablishing the equidimensional shape of the grains (fig. 9.9). It was suggested years ago that plastic flow was closely associated with ice crystals lying in a preferred orientation (Perutz 1940). It is now clear that a consistent orientation of grains in a polycrystalline ice mass (ice fabric) produces strain exerted nearly parallel to the basal planes (Russell-Head and Budd 1979; Duval 1981).

In sum, the power flow law derived in Glen's laboratory seems to fit the actual observed motions of glaciers. We cannot expect that polycrystalline ice will respond in precise agreement with the flow law, however, and modifications of the equation are generally required based on the individual situation. In addition, where stress systems are complex because of irregular valley floors, cross-sectional shapes, or nonisothermal ice, the flow law will have to be generalized or modified (Nye 1957, 1965). Attempts to go beyond the general approach and replace the simple power law with more sophisticated mathematical models become extremely difficult (Hutter 1982). Thus, in spite of inherent difficulties, Glen's power flow law provides an excellent approximation of glacial motion, between theoretical predictions and observed flow data, and with proper modification probably best describes the internal mechanics of glaciers.

A Simple Model of Internal Flow We can now develop a simple model of internal ice flow and determine whether glacial velocities agree at all with the basic mechanics of ice motion. In developing this theoretical model, we will assume that an active glacier remains un-

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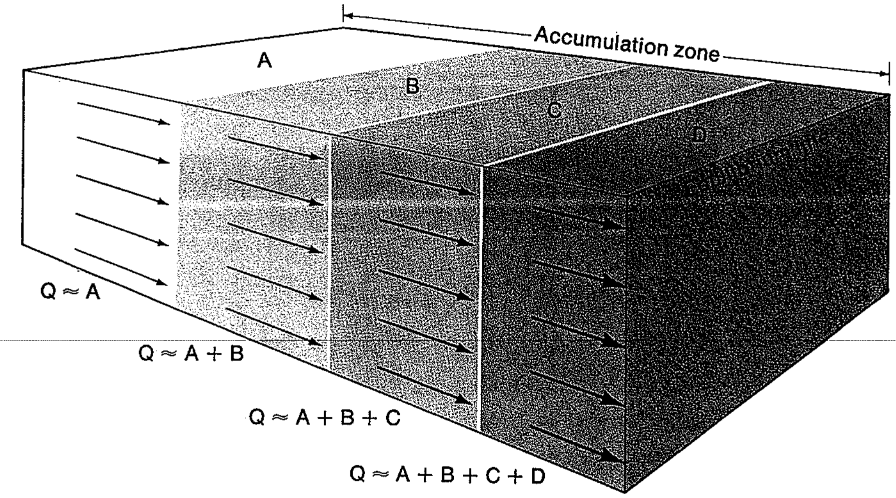

Figure 9.10
Hypothetical diagram showing why the velocity of glacier flow reaches a maximum at the equilibrium line. If no change in cross-section area (width × thickness) is allowed, increasing discharge down-ice requires the increase in velocity shown by arrows.

changed in its total and local dimensions over an entire budget year; that is, the total accumulation equals the total ablation, and all cross-sections of the glacier are equal in area and remain constant in area during the year. To maintain its area, each successive downglacier cross-section must transport from its lower boundary exactly the amount of ice and snow delivered to its upstream boundary. As figure 9.10 shows, in the accumulation zone the cross-section at the highest elevation has only a small area of accumulation above it and so must discharge only a small volume of ice, equivalent to the snow accumulated in that restricted surface area. Each section farther down-ice, however, must transfer a progressively larger volume of ice, since it moves not only the ice-equivalent of accumulation on its surface but also the cumulative volumes of all the higher sections. Without a change in cross-sectional area, the velocity of flow must increase to a maximum at the equilibrium line. This follows because discharge is increasing to that level and because glacier discharge is equal to the cross-sectional area times the velocity. In the ablation area we can similarly expect a gradual decrease in velocity from the equilibrium line to the terminus.

The model assumes that glaciers strive for and maintain some type of equilibrium and that the ice movement obeys the power flow law in some form. Mathematical treatment of even this simple glacial model is quite complicated. Therefore, we will make some further assumptions, as Nye did (1952a), that simplify the quantification of ice flow; specifically, we assume that ice motion is two-dimensional, plastic, and laminar such that the lines of flow are parallel to the bed and surface at all places (fig. 9.11). Under these conditions the shear stress at the base of a glacier, measured along the central longitudinal axis and perpendicular to the surface, is given as

$$\tau_b = \rho g h \sin \alpha$$

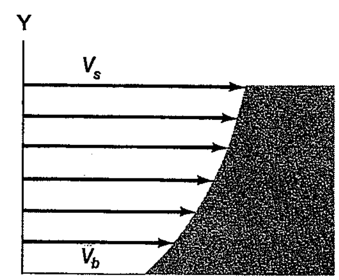

Figure 9.11
Velocity distribution in longitudinal section of glacier. Surface velocity (V_s) is sum total of flow rates at every level within the ice. Strain rate due to shear stress is highest at base of ice and is shown as the basal velocity (V_b).

where τ_b is the shear stress at the glacier base, ρ the ice density, g the acceleration of gravity, h the ice thickness, and α the slope of the surface.

By assimilating Glen's flow law into the equation, the velocity at any depth along the central axis (*xy*-axis) can be estimated by assuming that shear stress is proportional to depth and that the strain rate directly relates to that stress. Theoretically, then, the internal velocity profile of a glacier in a longitudinal section should show a decreasing rate of flow from the surface to the bedrock floor. Even though the strain rate is greatest where shear stress is the highest (at maximum thickness), the velocity increases from the base to the surface because each internal layer not only moves in response to the shear stress generated at that level but is also being carried on top of, and at the speed of, the adjacent lower layer. Thus, the surface velocity (V_s) is the sum total of strain rates for all the layers within the ice mass (fig. 9.11). The shear stress equation also shows us that internal velocity is proportional to the product of surface slope and ice thickness. The product seems to be fairly consistent, meaning that where the ice is thin the surface gradient will be steep, and vice versa.

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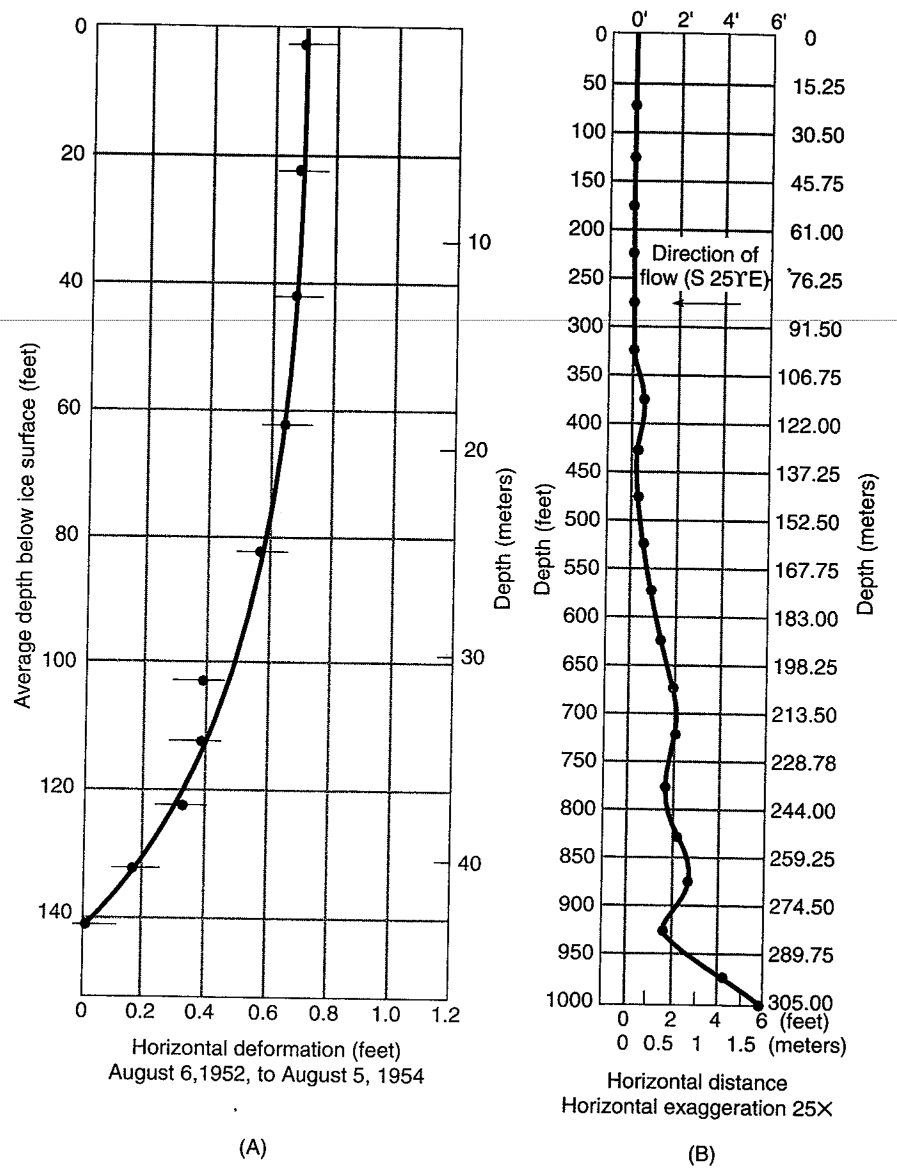

Figure 9.13
Internal velocity of two glaciers. Horizontal displacement of boreholes indicates the velocity.
(A) Saskatchewan Glacier
(B) Malispina Glacier.
(A): (Meier 1960)
(B): (Sharp 1953)

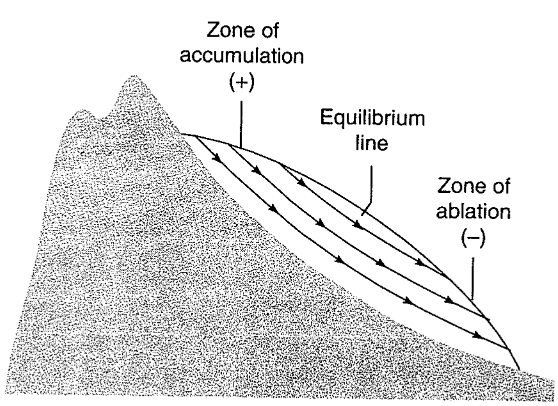

Figure 9.14
Longitudinal flow lines in hypothetical glacier.

with the surface of the ice. The potential slip planes show that in zones of extending flow (accumulation zones), the predominant downglacier slip direction will be downward at the surface, while in compressive zones, such as ablation areas, the low-angle downglacier slip will be upward.

This behavior seems to coincide with the predicted mode of flow in our ideal glacier (see fig. 9.14), but in real glaciers the pattern is much more complicated. An irregular bedrock profile will cause reversals of flow type in the ice (fig. 9.16), and where measurements have been made for an entire glacier, the flow pattern is not the simple one of our ideal case (fig. 9.17). Furthermore, more than one velocity maximum may occur, and the equilibrium line does not necessarily mark a zone of maximum flow velocity. In fact, on the South Cascade Glacier (Washington), the equilibrium line is usually near a zone of lower surface velocity, but the thickness is so great that the total discharge through the section is still very high.

Sliding

Real glaciers seem to possess the general traits that were predicted in idealized internal flow, at least in the distribution of velocity. Nonetheless, we cannot expect glacier characteristics to be predictable in detail. The assumptions made for the balanced glacier are invalid on a local scale, and velocity itself is controlled by several factors that are external to the system defined by the flow laws.

Part of the difficulty in applying the power flow law directly to glaciers is due to the fact that glaciers also move by sliding over the subglacial bed, this motion being in addition to creep within the ice mass. Few direct measurements have been made, but those available show that sliding may account for less than 10 percent of the surface velocity in some glaciers and more than 90 percent in others. In addition, sliding velocities may be extremely variable within different portions of the same glacier. For example, abnormally high velocities are sometimes encountered at the glacier sides (Meier et al. 1974).

Sliding processes operate either at the contact between the underlying glacial bed and the base of the glacier or within the lower ice layers. The processes themselves are poorly understood because direct observation of their action requires tunneling through the ice mass to the bedrock floor, an endeavor that is both costly and extremely difficult. Nonetheless, it is generally accepted that ice temperature, the character of the bedrock-ice interface, and the nature of the subglacial drainage system are the prime factors that determine how sliding actually works.

Perhaps the most intuitive and acceptable explanation of the process of sliding is based on the phenomenon of *glacial slippage,* in which the glacier slips over a water layer that rests between the underlying rock surface and the base of the glacier. This water is derived from precipitation and meltwater on the ice surface that is able to penetrate through openings and fractures to the

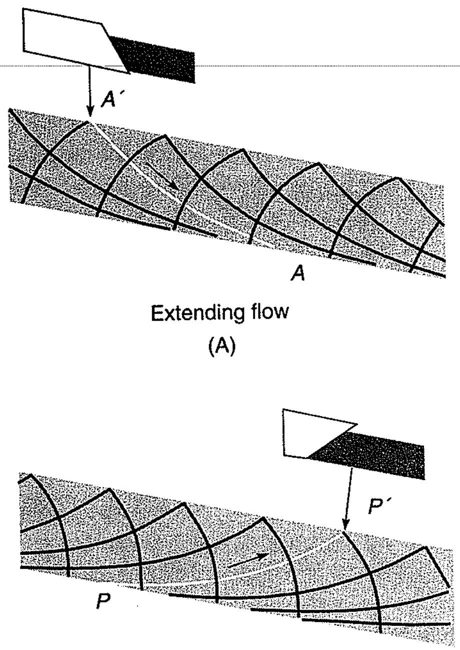

Figure 9.15
Potential slip planes under (A) extending and (B) compressive flow. The preferred downglacier slip paths will follow *A-A′* and *P-P′*.
(Nye 1952a)

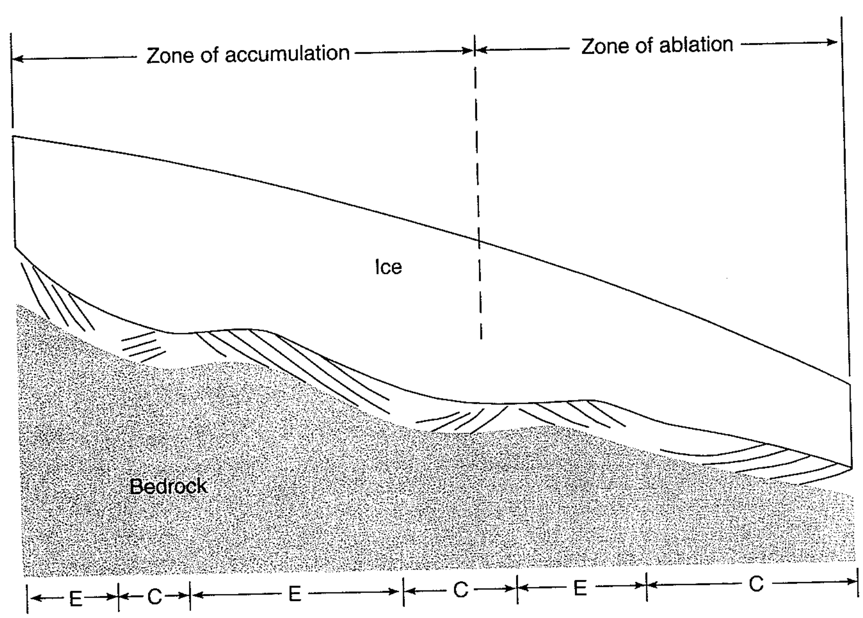

Figure 9.16
Longitudinal section of hypothetical glacier showing irregular bedrock profile and preferred slip planes within the ice. Zones of extending flow and compressive flow indicated by E and C.
(Nye 1952a)

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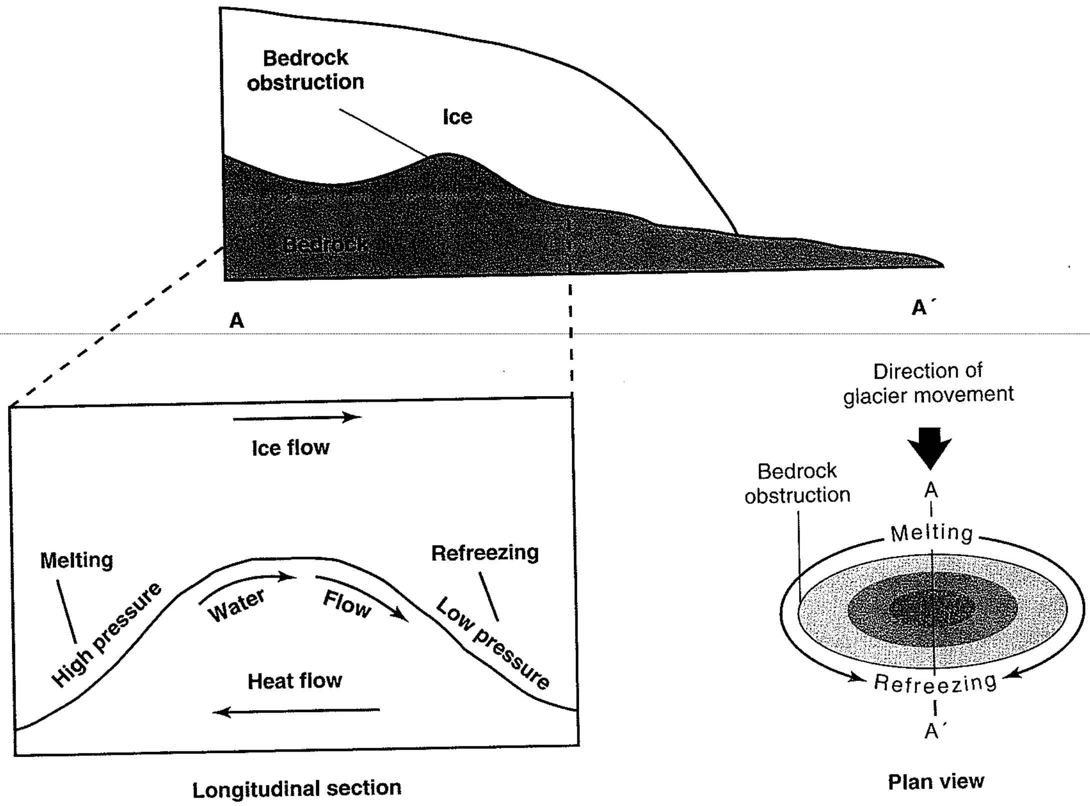

Figure 9.20
The process of regelation as it functions beneath a glacier. Melting occurs on upstream flank of obstruction where pressure is greatest. Refreezing occurs downstream of obstruction where pressure is least, although some cavitation may form.

where pressure is greatest, will have a higher strain rate than that at the downstream edge, and will exhibit *enhanced creep.* Larger obstacles will augment the stress and cause higher flow velocities; that is, the velocity will be directly proportional to the size of the bedrock knob. Because of these complications, Weertman (1957, 1964) suggested that sliding over an irregular bed is produced by components of the two processes: regelation is the predominant control on sliding when the barriers are small, and enhanced creep when the obstacles are large. It follows that some intermediate size of obstacle, called the *controlling obstacle size,* determines which process will function. In the subglacial bedrock topography, ice moving over or around protuberances smaller than the controlling obstacle size will do so mainly by regelation; where surface bulges are larger, creep will probably dominate. Field observations and theoretical studies suggest that the controlling obstacle size is typically on the order of 0.05 to 0.5 m.

Motion with Deforming Subglacial Sediments

The theoretical models of internal flow and basal sliding presented earlier generally assume that the glacier bed is a rigid, impervious bedrock surface. Although such surfaces do exist beneath glaciers, observations of regions recently exposed by glacial retreat illustrate that at least the margins of many glaciers are underlain by unconsolidated debris. In addition, Boulton and Hindmarsh (1987) point out that during the last glacial period, 70 to 80 percent of the continental ice sheets in Europe and North America were underlain by unconsolidated or poorly consolidated sediments of Quaternary or Tertiary age. It is now clear that these unconsolidated, subglacial sediments may deform, allowing the overlying ice to actually move along with the mobilized material. In fact, investigators have recently found that as much as 90 to 95 percent of the forward motion of some temperate glaciers results from the deformation of subglacial sediments (Boulton 1979; Alley et al. 1986; Blankenship et al. 1986; Boulton and Hindmarsh 1987).

The deformation of subglacial sediments is governed in large part by the shear stress applied to the material by the overlying ice and the shear strength of those sediments. Boulton (1979) used a form of the Coulomb equation to express the shear strength of subglacial material in an unconsolidated state:

$$S = C + (P_i - P_w) \tan \phi$$

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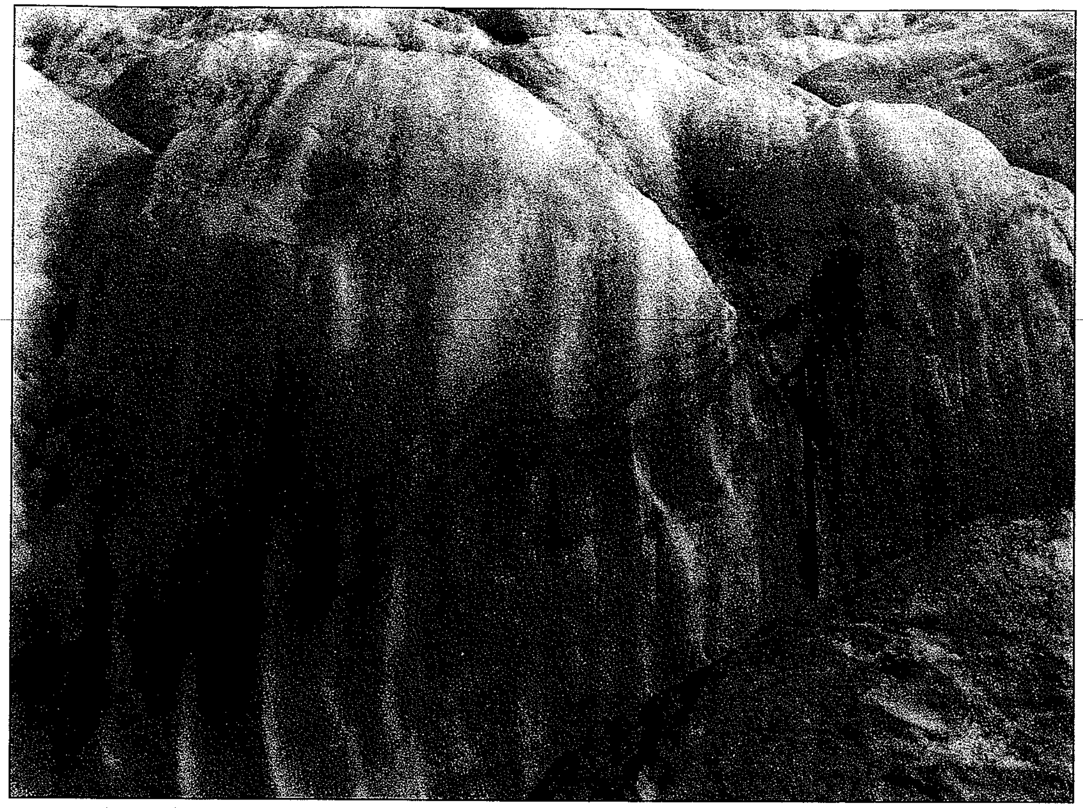

Figure 9.24
Saskatchewan Glacier showing gently dipping stratification wrinkled and intersected by nearly vertical foliation. Exposed on east wall of a crevasse, 4.5 km below firn limit in midglacier. Province of Alberta, Canada.

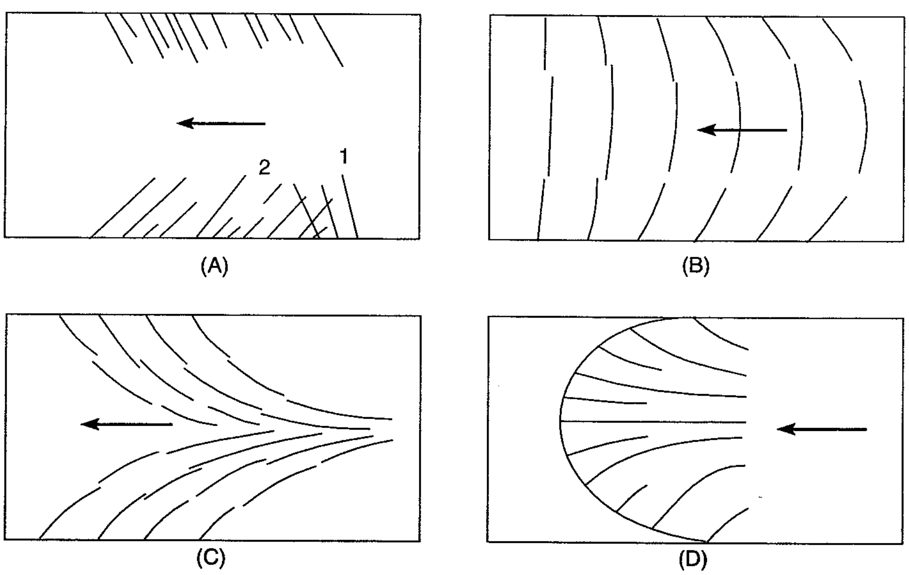

Figure 9.25
Types of crevasses in valley glaciers: (A) marginal or chevron—(1) old rotated crevasses; (2) newly formed crevasses; (B) transverse; (C) splaying; (D) radial splaying. Arrows show flow direction.
(Sharp 1960)

deeper than 30 m. It is generally assumed that crevasses develop perpendicular to the direction of maximum elongation of the ice. On some glaciers, however, the crack directions do not precisely coincide with the measured surface strain rates (Meier 1960). Nonetheless, crevasse types (fig. 9.25) do reflect a local tensional stress environment. *Splay crevasses* or *radial crevasses* form near the flow centerline under compressive flow where spreading exerts a component of lateral extension. In contrast, near the ice margins the shear stress parallels

10

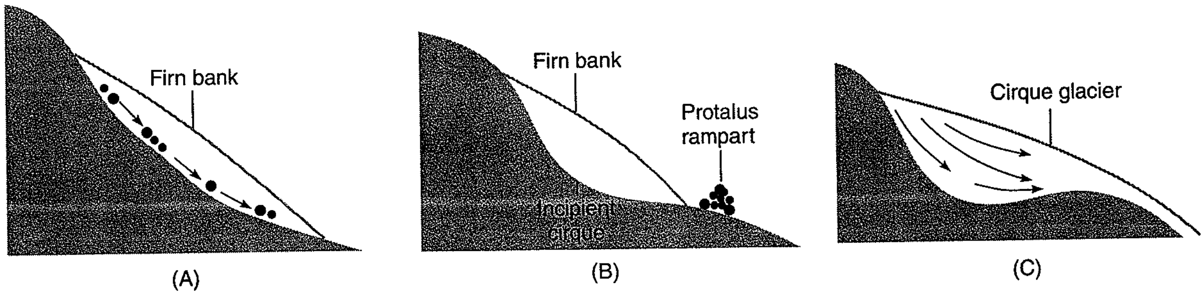

Figure 10.9
Stages of cirque development. (A) Nivation beneath firn bank. (B) Nivation cirque. (C) Cirque with fully developed cirque glacier.

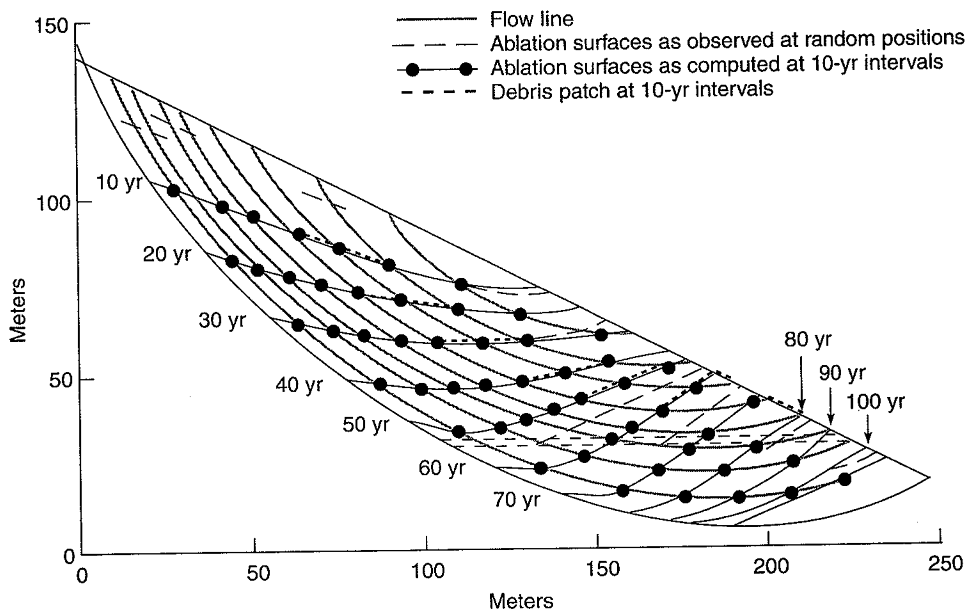

Figure 10.10
Long section through a cirque glacier in Norway showing ablation surfaces and debris patches at 10-year intervals. Note rotation of ablation surfaces in the down-ice direction.

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 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.

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.

Johnson (1904) suggested that surface meltwater gains access to the base of the headwall by percolating down the bergschrund. A **bergschrund** is a crevasselike opening near the headwall that separates actively moving ice of the glacier from nonactive ice frozen to the headwall. Once the water had permeated fractures in the wall rock, it was assumed to produce extensive frost shattering as the water was 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

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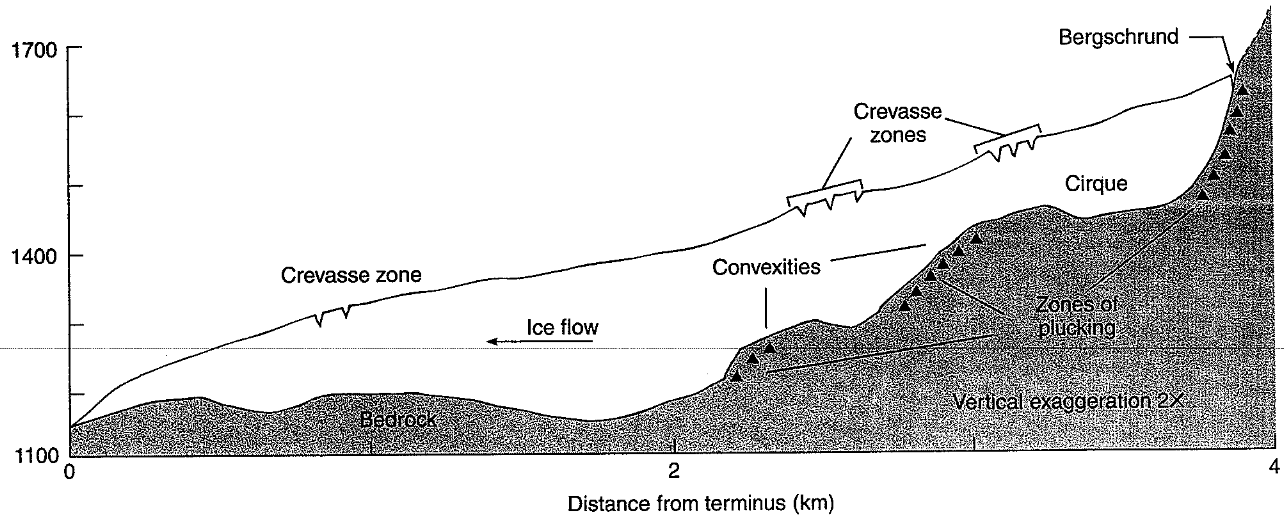

Figure 10.11
Longitudinal profile of Storglaciaren, Sweden, showing the relationship between zones of plucking and points of water input along crevasse zones and the bergschrund. The downward flow of water through these openings may lead to rapid subglacial water-pressure variations and localized plucking.
(Hooke 1991)

detailed analysis of freeze-thaw mechanics, indicates, however, 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. 10.11). Utilizing a model similar to that proposed by Iverson (1991b) and discussed earlier in the chapter, he argues that rock shattering is initiated along the headwall by abrupt decreases in subglacial water pressure, while subsequent water pressure increases lead to plucking of the shattered rock.

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. 10.12). The transformation of a V-shaped river valley into a U-shaped glacial valley has been the topic of considerable analysis and speculation (Johnson 1970; Boulton 1974). Most recently Harbor (1992) addressed the mechanical development of a glaciated valley utilizing a numerical modeling approach. Harbor's analysis assumes that the rate at which any point on the valley wall is eroded is a function of the velocity at which the basal ice moves over the bedrock. Given this constraint, erosion rates tend to increase from the margins of the glacier toward its center (fig. 10.13). A zone of lower flow should exist, however, near the valley axis. Thus, initially, erosion rates reach a maximum part way up the valley sides. As differential erosion causes the sides to bow outward, the central region of reduced erosion is progressively removed because the velocity distribution changes as a more parabolic cross-valley profile is produced. Eventually, erosion systematically increases toward the axis of the valley, and erosion rates tend to maintain the classic U-shaped morphology (fig. 10.13).

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 formation of the U-shaped cross-section occurs by both lateral and vertical erosion of the preexisting valley. Whether the parabolic form is derived predominantly by widening or predominantly by