Chapter 16

Glaciers as Landforms: Glaciology

Of all the water near the surface of the earth, only about 2 percent is on the land, in the solid state as glacier ice (Figure 2-5). Even this small fraction is sufficient to cover entirely one continent (Antarctica) and most of the largest island (Greenland) with ice to an average thickness of 2.2 km over Antarctica and 1.5 km over Greenland. At the present time, about 10 percent of the earth's land area is ice covered. An additional 20 percent has been ice covered repeatedly during the glaciations of the Pleistocene Epoch, much of it as recently as 15,000 to 20,000 years ago. Glaciation has been the dominant factor in shaping the present landscape of North America northward of the Ohio and Missouri rivers and of Eurasia northward of a line from Dublin eastward through Berlin to Moscow and beyond the Urals (Figure 16-1). In addition, mountains and plateaus in all latitudes have been glaciated to an altitude 1000 to 1500 m lower than their present snowlines.

GEOMORPHOLOGY OF GLACIER SURFACES

The simplest morphologic subdivision of glaciers is to distinguish those that flow between confining rock walls, or **valley glaciers,** and those that bury the rocky landscape and flow unconfined by virtue of their great thickness, the **ice caps** and **ice sheets** (Sharp, 1988). Although the morphology of all glaciers is controlled by the rheidity of ice (p. 13), important differences are to be noted whether flow is confined or unconfined.

Glaciers originate in regions where snow accumulation exceeds loss, and they flow outward or downward to regions where losses exceed accumulation (Figure 16-2). The *terminus,* or downglacier extremity, represents the line where losses by all causes (melting, sublimation, erosion, and calving into water: collectively termed **ablation**) equal the rate at which ice can be supplied by accumulation and forward motion. Some valley glaciers have steep ice faces at their termini, but others end deeply buried in transported rock debris, so that the exact location of the terminus is uncertain. In the *zone of accumulation,* vectors of particle motion are downward into the ice mass as each year's snowfall adds a new surface layer to the glacier. Erosion at the glacier bed is intensified. In the *zone of ablation,* vectors of particle motions point toward the ice surface, because a surface layer is annually removed, exposing progressively deeper ice (Figure 16-2). Net sediment deposition is probable. This is the predominant method of bringing sediment to the surface of a glacier, more important than thrust faulting (p. 366). At the cross section of the glacier where upglacier net accumulation and downglacier net ablation are in exact balance, internal flow vectors are parallel to the glacier surface. This equilibrium cross section of a glacier is located beneath the **equilibrium line,** the downglacier edge of net annual accumulation.

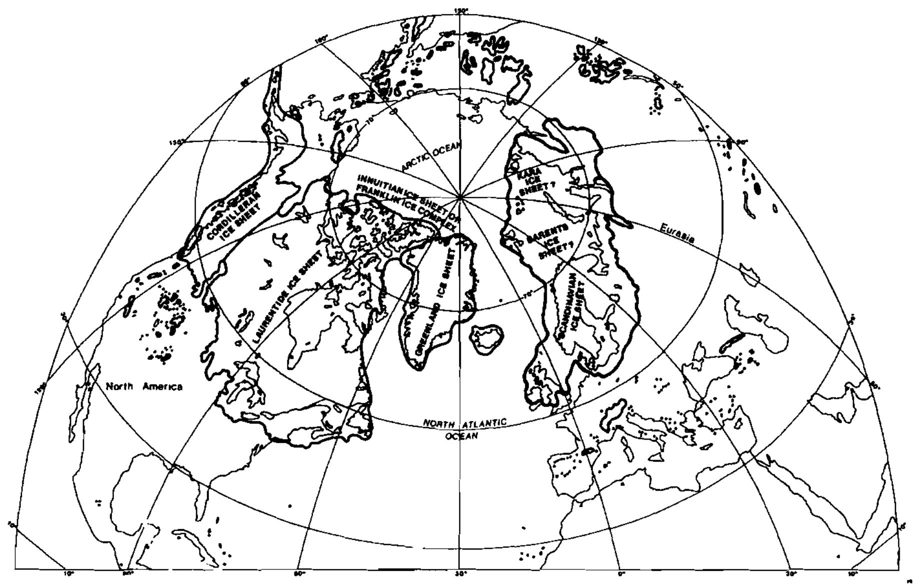


FIGURE 16-1. Northern hemisphere ice sheets and other glaciers during the last ice age. Possible floating ice shelves not shown (Denton and Hughes, 1981, p. viii, reprinted by permission. Copyright © 1981, John Wiley & Sons, Inc.).

Valley Glaciers

In a mountainous terrain, the colder temperatures and usually higher precipitation at higher altitudes produce ice fields at the heads of glaciers. This is the zone of accumulation (Figure 16-2), and as a rule of thumb, about 65 percent of the surface area of a valley glacier is in this zone. The annual discharge of ice through the cross section beneath the equilibrium line is an important measure of the glacier's regimen. The equilibrium line is approximated by the annual **snow line** on the glacier and adjacent mountain slopes.

Downglacier from the equilibrium line, a glacier has net annual loss of ice. Melting is the major factor in valley glaciers, but sublimation, wind ablation, and iceberg calving into lakes or the ocean also contribute. A glacier is in equilibrium only if annual discharge through the equilibrium-line cross section is equal to the net annual accumulation upglacier and the net annual ablation downglacier. Otherwise, it will grow or shrink.

Many large valley glaciers are fed by one or more *tributary glaciers* (Figure 16-3). Although tributary glaciers do not erode their floors as deeply as does the trunk

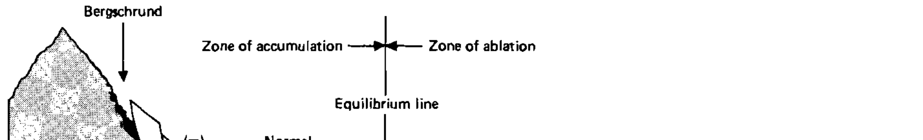


FIGURE 16-2. Trajectories of particle motion in a valley glacier. The large area of the zone of accumulation is not well shown in this longitudinal cross section. Sections labeled (+) and (−) are local areas of compressing and extending flow.

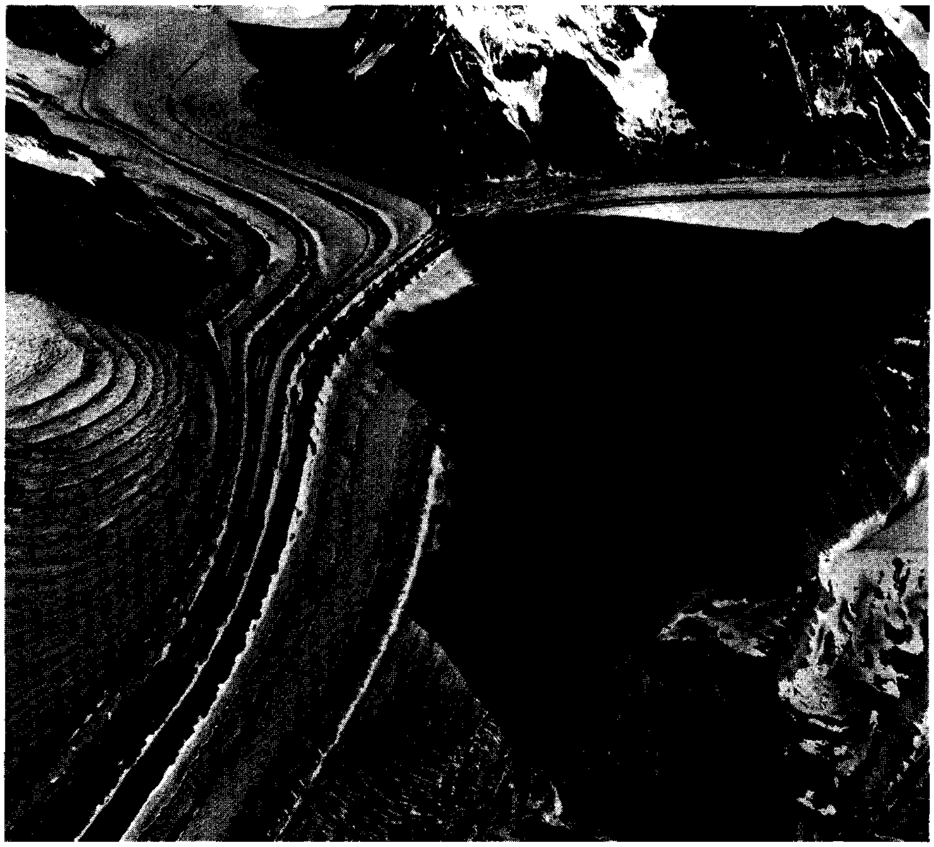


FIGURE 16-3. Yentna Glacier, Alaska Range. Late summer view; equilibrium line is visible toward the upper left. Rhythmic ridges on the leftmost tributary are *ogives,* formed by seasonal melting as the glacier descends a buried ridge. Lateral moraines of tributary glaciers become medial moraines in foreground, but maintain their individual identity (photo: Austin Post, U.S. Geological Survey).

glacier, the surfaces of merging glaciers join at a common level because ice is easily deformed and responds to the lateral stresses imposed by an entering tributary. Tributary glaciers do not turbulently mix with the trunk glacier, as rivers do. Because of the nature of flow in ice, the tributary glacier retains its identity as a separate stream or surface ribbon of ice (Figure 16-3) although it may become progressively attenuated or enfolded into the trunk glacier.

Among the oldest observations of glacier flow were reports that lines of stakes emplaced straight across valley glaciers would bulge progressively downvalley year after year, proving that the center part of the ice surface moved faster than the margins. Many subsequent experiments have confirmed the fact (Figure 16-4). Because most of the differential shear is very near the base and lateral margins of a glacier, the mean velocity of the surface ice is within a few percent of the mean velocity of the entire glacier cross section (Paterson, 1994, p. 270). As a useful result of this approximation, glaciologists can estimate the annual discharge of ice through the equilibrium-line cross section of a glacier by measuring surface velocities only. It is not necessary to drill boreholes and measure velocities within the ice. In this respect especially, the laminar flow of a glacier is much simpler than the turbulent flow of a river, for which an entire velocity cross section must be measured to determine mean discharge (p. 203).

Valley glaciers have much higher surface gradients than rivers, commonly averaging 10 percent (6°). On ice cliffs or ice falls steeper than 45°, ice avalanches are nearly continuous. Ledges buried beneath the ice are reflected by *crevasse fields* and ice pinnacles (*séracs*) on the ice surface (Figures 16-3 and 17-4). The long profile of a glacier surface is likely to be marked by level areas and steep drops, much like the surface gradient of a river through rapids and across pools.

Within the zone of accumulation of a valley glacier, velocity generally increases downglacier. This is called *extending flow* because an imaginary mass within the glacier becomes longer and thinner as it moves downvalley (assuming that the valley sides are parallel). Within the zone of ablation, longitudinal velocity generally decreases, perhaps to zero at the toe of the glacier. This is called *compressing flow* because an imaginary mass within the glacier becomes shorter and thicker as it flows (again assuming no change of the valley width). The equilibrium cross section is therefore the place where the flow rate is at the maximum.

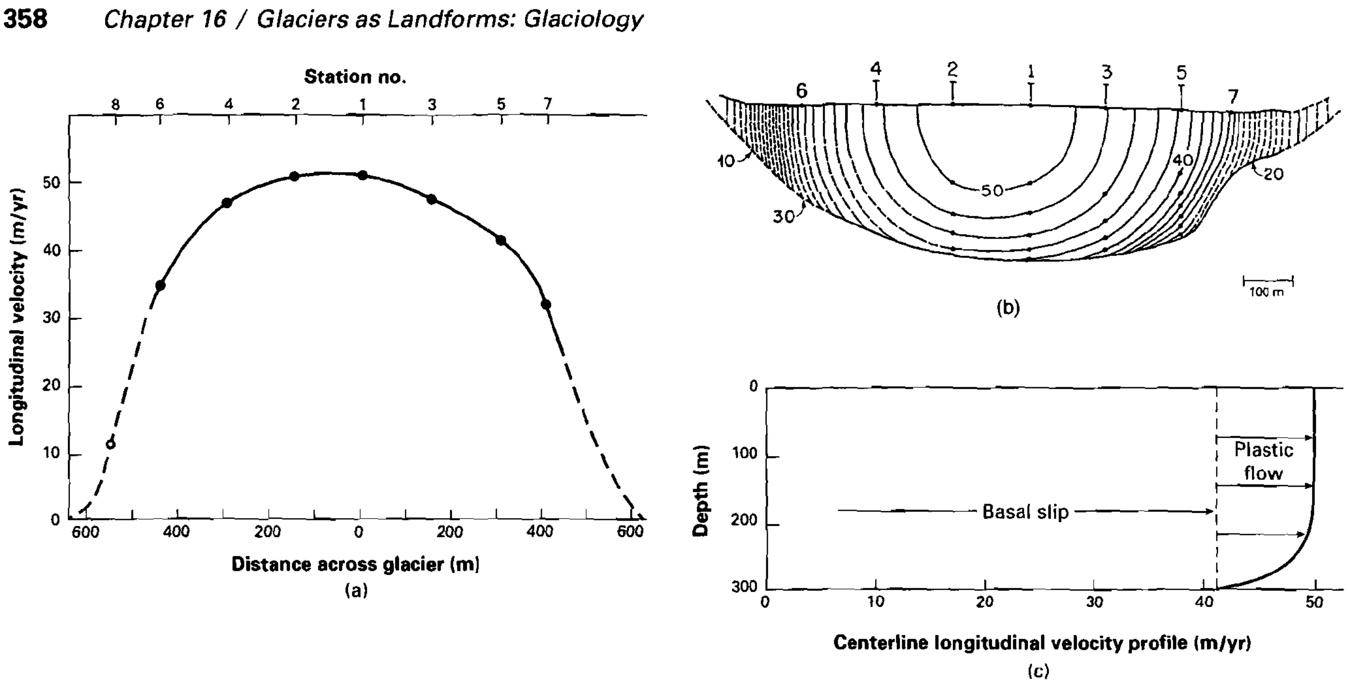


FIGURE 16-4. Flow in the Athabasca Glacier, Alberta, Canada. (a) Lateral variation in surface velocity. Solid circles, surveyed markers; open circle, estimated from adjacent measured velocities. (b) Distribution of longitudinal velocity in cross section from boreholes at stations 1-5, with extrapolated marginal velocities. (c) Inferred longitudinal velocity profile at the centerline of flow half way between boreholes 1 and 2. Basal slip accounts for 81 percent of the surface velocity in this profile, and more than 70 percent of the surface velocity across half the glacier width (data from cross section A, Raymond, 1971).

Locally within a glacier, other areas of extending and compressing flow also can be identified (Figure 16-2). For example, flow may be compressive upstream from a bed obstruction, with the ice thickening and the glacier surface bowing upward as a reflection of the buried obstacle. Extending flow is common where ice passes over a buried sill and drops rapidly. At such places, the ice may fracture into a crevasse field or simply thin rapidly by extending flow. In plan view, compressing and extending flow can be expected, respectively, on the inside and outside of curves, where shear rates are high. Similar surface fracture patterns on floating ice shelves show regions of extending and compressing flow (Figure 16-8). Beneath the ice, regions of extending and compressing flow can control basal slip (p. 366), and determine whether the glacier erodes its bed or deposits sediment (Chapter 17).

On the surface of most valley glaciers in the ablation zone, a variety of ephemeral landforms develop due to melting, collapse of ice tunnels, and erosion by streams flowing on, in, or along the glacier. Many features are analogous with karst features and may be collectively called **glacier karst.** *Thaw lakes* and swallow holes (*moulins*) mark the ice surface. Segregations of debris-rich ice stand as ice-cored mounds or moraine ridges above the debris-covered surface (Figure 16-3). Lakes may drain and then abruptly form again when a crevasse or moulin opens at depth, and then deformation or clay-rich sediments seal it closed again. Ice tunnels within or at the base of the glacier discharge sediment-laden water from the terminus.

Ice Caps and Ice Sheets

Unconfined glaciers may be so large that they form their own orographic precipitation pattern. The accumulation zone is in the central region of the dome, and the ablation zone is on the periphery, below the equilibrium line. The regimen of such an ice cap or ice sheet is very sensitive to slight climatic changes. Ice caps and ice sheets, unlike valley glaciers, are examples of *positive feedback systems.* For example, any increase in accumulation will increase the height and therefore the area of the accumulation zone relative to the ablation zone, resulting in further accelerated growth. Conversely, a decrease in accumulation will lower the central dome, further decrease the relative area of accumulation, and accelerate shrinkage. Because the climate has warmed since they formed, numerous small or moderate-size ice caps on Iceland, Scandinavia, and the Canadian Arctic islands could not re-form today if their summits ever lowered to near the equilibrium line; the present sustaining orographic precipitation is a consequence of the ice caps themselves, not of any mountain mass beneath them.

The radial profiles of continent-size ice sheets and the somewhat smaller, but otherwise similar, ice caps differ very little from each other (Figure 16-5). Theories of radial flow predict parabolic or elliptical radial profiles very similar to those actually observed (Figures 16-5 and 16-6). The ice must spread radially in plan view from the area of accumulation, as well as outward along any cross section. Except for the complexity of outlet glaciers and ice streams (p. 360–361), the general flow lines (Figure 16-7) are probably much as inferred by early workers: downward beneath the zone of accumulation near the center and then outward to the zone of ablation. Obstructions such as enclosing mountain ranges obviously complicate the flow patterns. Note the considerable vertical exaggeration on Figures 16-5, 16-6, and 16-7. At true scale, the thickness of the ice could barely be shown by a slight thickening of the horizontal base lines. Published maps and radar profiles (Walker et al., 1968; Drewry, 1983; Bogorodsky et al., 1985) give an excellent idea of the surface relief, bottom relief, and thickness of large ice sheets.

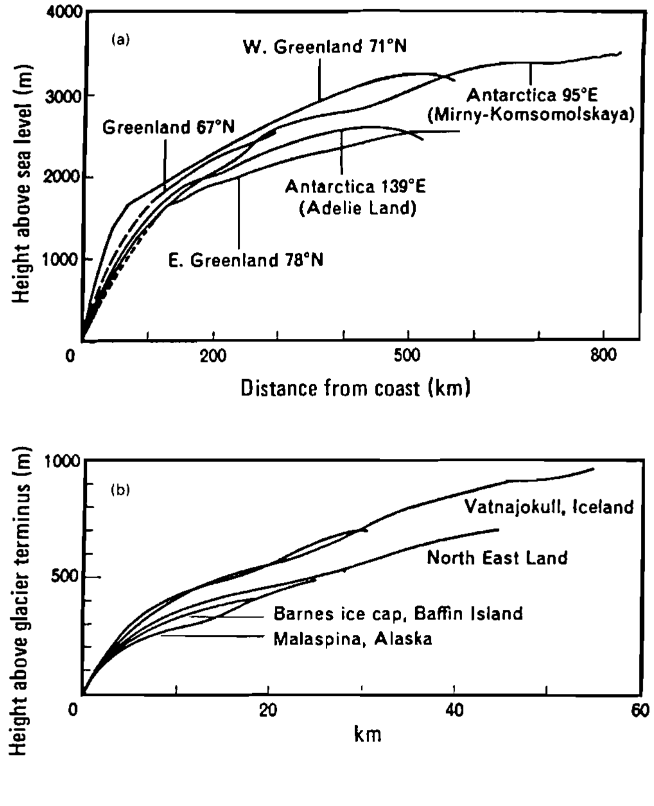


FIGURE 16-5. Surface profiles of (a) the ice sheets of Greenland and Antarctica, and (b) smaller ice sheets and ice caps. In both figures, bedrock relief is slight relative to ice thickness (Robin, 1964, Figure 5).

Most of the Greenland ice sheet perimeter is a series of broad, gently sloping ramps. Along some parts of the perimeter, where the terminus is on land, the glacier ends as a vertical ice cliff as much as 100 m in height. The cliffs may be due to a combination of factors, such as low temperatures, compass orientation relative to a low sun angle that causes sublimation directly from the cliff face, and cold, brittle ice almost free of enclosed sediment except at the base. The Greenland ice sheet (known as the *Inland Ice*) rests on a basin-shaped interior plateau surrounded by mountain ranges (Figure 16-6) (Reeh, 1989). Around much of its margin, the ice moves between exposed or slightly buried mountain peaks as **outlet glaciers.** These have many of the geomorphic properties of valley glaciers. At the head of each outlet glacier is an area of shallow concavity sloping toward the outlet, with many wide, deep crevasses. This is the "drawdown" area, or "iceshed" of that particular outlet glacier.

At least two discrete outlet glaciers drain significant portions of the Greenland ice sheet at rates much more rapid than the average radial spreading rates. Jakobshavns Isbrae reaches the west coast of Greenland as a high gradient, thick (2800 m) outlet glacier that may be the world's fastest moving nonsurging glacier (Echelmeyer et al., 1991, 1992). The continuous forward motion is 7 km/yr near the floating terminus, and it produces many of the icebergs that float south through Davis Strait. This single outlet glacier drains 6.5 percent of the total area of the Greenland ice sheet although it is only 6 km wide and 85 to 90 km in length. Spectacular surface zones of marginal shearing separate the ice stream from adjacent slow-moving ice. A similar tongue of fast-moving ice in northeastern Greenland was identified from repeated satellite radar images (Fahnestock et al., 1993). It drains at least twice the area of Jakobshavns Isbrae. The factors that control such rapidly moving portions of an ice sheet have yet to be determined, but they violate generalizations about uniform trajectories of flow radially from the center of an ice sheet as implied by Figure 16-7.

Much of the perimeter of the Antarctic ice sheet is also a great cliff, but that cliff ends at a *grounding line* in the ocean (Figure 16-7). A few **nunataks,** peaks surrounded by ice, rise above the surface (Figure 16-8). The West Antarctic sector, in the longitudinal zone that faces northward toward the Pacific Ocean between New Zealand and South America, is grounded 1 to 2 km below sea level in two large tectonic basins from which ice flows into the floating *ice shelves* of the Ross Sea and Weddell Sea and into numerous other smaller ice shelves (Figure 16-8). Episodically, slabs of ice with

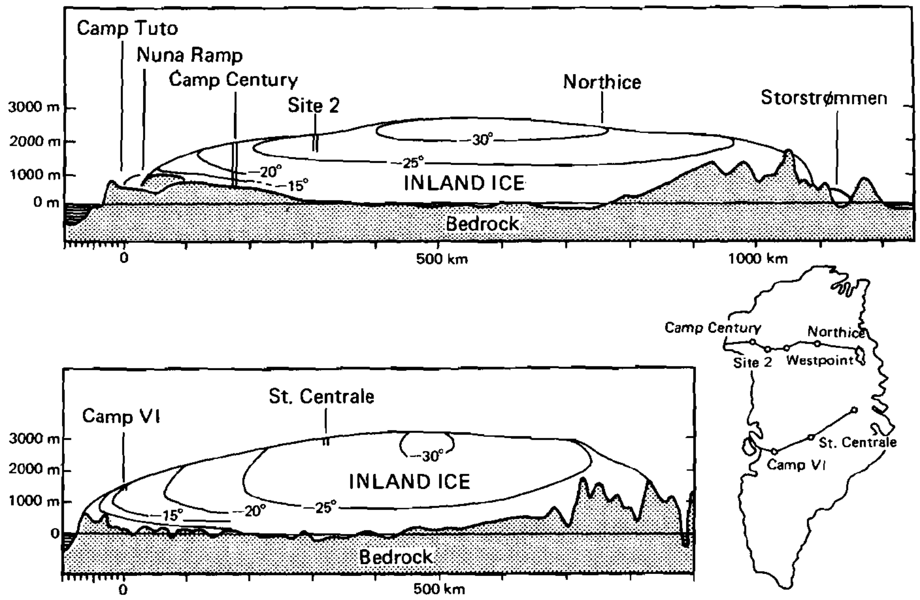


FIGURE 16-6. Profiles and estimated temperatures of the Greenland ice sheet (the Inland Ice). Vertical exaggeration 50:1 (Weidick, 1975, Figure 4).

areas measured in thousands of square kilometers separate from the ice shelves and drift into the southern ocean (Swithinbank, 1988; Rott et al., 1996). Large outlet glaciers are held back by *ice rises*, buried hills over which the ice rides in the transitional zone where outlet glaciers become floating ice shelves (Figure 16-8). Separated by the ice rises are several rapidly flowing *ice streams*, moving over water-saturated sediments at rates of up to several meters per day (Engelhardt et al., 1990; Alley and Whillans, 1991).

The surface of the East Antarctic ice sheet is a monotonous dome with only minor areas of exposed bedrock, mostly near the coast (Drewry, 1983; Radok, 1985). Flow lines define ice divides and drainage basins, but surface gradients are very low, and little relief is apparent. Much of the East Antarctic ice sheet

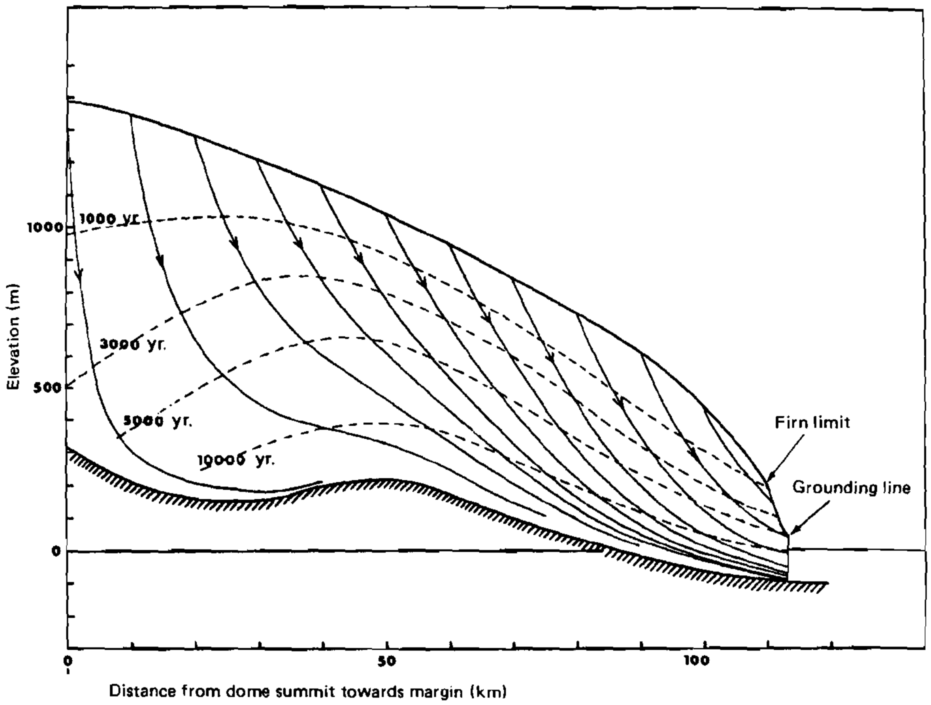


FIGURE 16-7. Particle paths and calculated ages of ice in a cross section of Law Dome, a small independent ice dome on the edge of the Antarctic ice sheet in Wilkes Land (111°E longitude) (from Budd and Morgan, 1973, Figure 4).

has precipitation of less than 50 mm/yr of water equivalent, making it one of the driest regions on earth. Surface temperatures are the coldest known on earth.

GLACIOLOGY

Glaciology is the scientific study of all ice, not just of glaciers. It includes the study of ice crystals in high clouds, hail, and snow; frozen lake, river, and ocean water; and glacier ice. The Voyager and Galileo spacecraft exploration of Jupiter, Saturn, and Uranus and their moons has broadened the interest of glaciology to include the ices of ammonia (NH_3) and methane (CH_4), for these along with H_2O ice are probably the materials of which many of the moons of the outer planets are composed (Klinger et al., 1985; Jankowski and Squyres, 1988). Glaciologists can be metallurgists, meteorologists, physicists, geographers, or geologists. What we know about glacier ice has been learned from a wide field of experiment, observation, and theory (Lliboutry, 1965; Hutter, 1983; Weertman, 1983; Paterson, 1994; Hooke, 1998).

Formation of Glacier Ice

The conversion of snow to glacier ice follows a regular sequence. Fresh snow is full of entrapped air and may have a bulk specific gravity even lower than 0.1. Snowflakes readily *sublime* (vaporize directly from the solid state without melting), so aging snowflakes lose their frilly margins and become more globular. Meltwater often freezes onto snowflake nuclei, so aging also tends to increase the size of the ice granules.

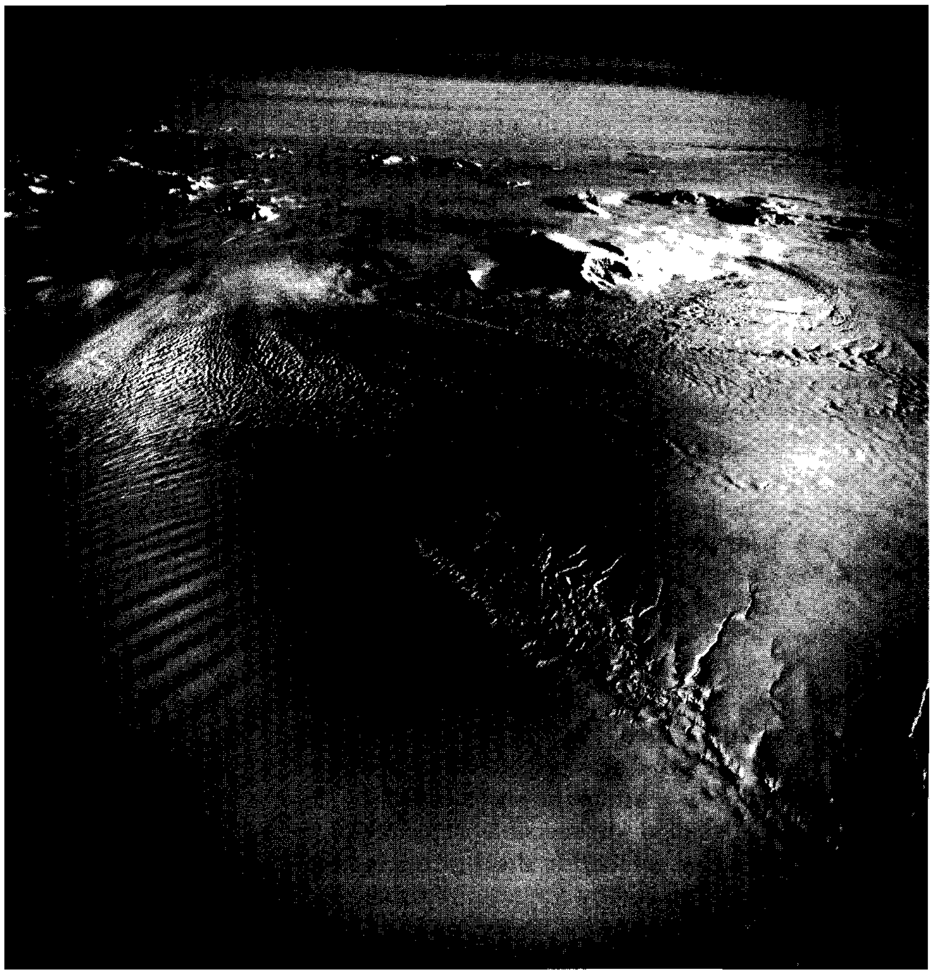


FIGURE 16-8. Oblique aerial photograph up the Fleming outlet glacier, Graham Land, Antarctica. The grounding line is somewhere in midphoto. Buffer Ice Rise (foreground; about 3 km wide) deflects and fractures the floating Wordie Ice Shelf (photo: U.S. Navy for U.S. Geological Survey, TMA 1835F33, courtesy R. S. Williams, Jr.).

Old snow, such as remains on sheltered mountain slopes in early summer, has the texture of very coarse sand. This is the late-season *corn snow* or *buckshot snow*, familiar to avid skiers. The loose granular mass is about one-half ice and one-half entrapped air and has a bulk specific gravity of about 0.5.

If snow has survived a summer melting season, it is called **firn** in German or **névé** in French. Both terms are widely used in English. Firn, or névé, is an intermediate step in the conversion of snow to glacier ice, with a bulk specific gravity between 0.5 and 0.8. It is granular and loose unless it has formed a crust. It represents the net positive balance between winter accumulation and summer loss.

As successive annual layers accumulate, the deep firn is compacted. Individual ice grains freeze together. By definition, when the grains are frozen together so that the mass is impermeable to air, firn becomes glacier ice. The bulk specific gravity is usually about 0.8 by this stage of consolidation. Glacier ice is a polycrystalline mass that includes a variable amount of entrapped air, as well as dust and rock fragments that have fallen, washed, or been blown onto the ice surface, and rock that has been eroded from beneath the glacier. The bulk specific gravity of glacier ice ranges from about 0.8 to 0.9, close to the specific gravity of pure, gas-free ice (0.916).

Ice Deformation

The only form of ice that occurs naturally on planet Earth crystallizes in the hexagonal crystal system, and is characterized by three intermolecular glide axes 120° apart on a plane perpendicular to the *c* optical axis (Figure 16-9). The glide axes define a *basal glide plane* on which single ice crystals easily deform under very low shear stress. The basal plane is the orientation of the incomplete hexagonal skeletal ice crystals we call snowflakes. Under a low sustained shear (directional) stress of 0.3 MPa (3 bar), an ice crystal can be drawn out into a thin ribbon in only a few days (Figure 16-9) (Glen, 1952).

When an ice specimen is subjected to a slight shear stress (pressure or tension) and then released, it elastically recovers its original shape. When the shear stress is increased to above a certain threshold called the *elastic limit* or *yield strength*, irreversible deformation or *strain* results (Figure 16-10a). If shear stress is maintained, the sample deforms by continuous plastic strain or *creep*.

Polycrystalline ice under shear stress deforms very much as iron deforms when it is heated to a bright red color. Either in single crystals or in polycrystalline masses, at only slightly increased stresses above the yield stress the strain rate of ice becomes very rapid. By the simplifying assumption that ice is a perfect plastic, which does not deform at all at stresses less than 0.1 MPa (1 bar) but deforms infinitely rapidly at a slightly larger stress, theoretical models of ice flow have been developed that closely approximate the flow of glaciers. The approximation is sometimes referred to as the "one-bar flow law." It is plotted on Figure 16-10b as a horizontal line at a shear stress of 0.1 MPa (1 bar). Several lines of evidence suggest that shear stresses at the base of glaciers are seldom far from the range 0.05 to 0.15 MPa (0.5 to 1.5 bar), so the simplifying assumption of the one-bar flow law has proved useful (Paterson, 1994, p. 240).

More precisely, the plastic deformation of ice compiled from laboratory experiments and measurements in ice tunnels and boreholes (Figure 16-11) fits a power-function equation of the form

$$\dot{\varepsilon} = A\tau^n$$

where $\dot{\varepsilon}$ is the creep rate, or rate of permanent deformation measured in percent change of length per time (which simplifies to reciprocal times; see p. 12), τ is the shear stress, and A and n are constants. This equation is called "Glen's flow law" in recognition of the original experimenter (Glen, 1952; Paterson, 1994, pp. 85; Menzies, 1995, pp. 139, 150; Hooke, 1998, pp. 41ff). Experimental values of n for both single crystals and polycrystalline ice generally have a mean of about 3 (Figure 16-10b). When plotted on a log-log graph, the value of n of about 3 is illustrated by a thousandfold increase in creep rate for each tenfold unit increase in shear stress (Figure 16-11).

If an ice grain is constrained from easy deformation on its basal glide plane, as for example, if it is completely surrounded by other grains of ice and held in an unfavorable orientation relative to the applied shear stress, it will still deform, but much more slowly. The creep rate is 500 times more rapid for a given shear stress in the "easy glide" basal plane of a single crystal than in various "hard glide" crystal orientations or in polycrystalline aggregates (Figure 16-11) (Weertman, 1983, p. 224; Hooke, 1998, p. 34). The cluster of experimental and field-determined strain-stress relationships in the lower central part of Figure 16-11 are regarded as the best ones to use in calculations or modeling of glacier flow. They demonstrate that, in general, a mass of polycrystalline ice subjected to a shear stress of 0.3 MPa (3 bar) will deform by an amount equal to its original length in about 1 year.

Although absolute creep rates vary greatly between single crystals in the easy glide orientation and polycrystalline glacier ice with varying amounts of

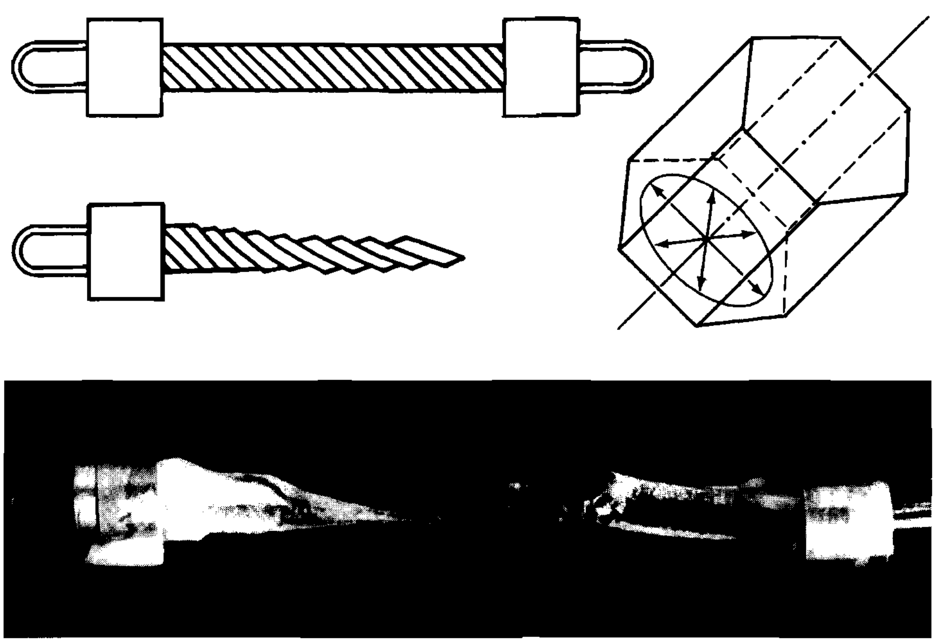


FIGURE 16-9. Single crystal of ice, 13 mm^2 cross section, *c* optical axis 45° to specimen axis, deformed by sustained tension of 0.4 kg (= 0.3 MPa) (photo: J. W. Glen, reproduced from the Journal of Glaciology by permission of the International Glaciological Society).

impurities, the power function (n) that describes the increase in creep rate with increased shear stress is similar in most experiments. The constant A, but not n, is strongly temperature-dependent, so that for a given shear stress, the strain rate is only one-fifth as great in ice at −25°C as in ice at −10°C (Paterson, 1994, p. 86).

It is important to note that confining, or cryostatic, pressure does not directly affect the strain rate of ice. Only shear, or directional, stress causes plastic deformation. As with rock and other solids, confining pressure only prevents brittle fracture and thereby permits creep to be measured. Confining pressure does not make the ice at the bottom of a glacier any more "plastic" than the ice at the surface. Many serious errors in interpreting glacier flow have been made because of this misconception.

Thermal Classification of Glaciers

Confining, or cryostatic, pressure does affect glacier deformation indirectly. Pressure depresses the freezing temperature of water almost linearly from 0°C at 0.1 MPa (1 bar) to -22°C at a pressure of 207 MPa (2070 bar), beyond which other forms of ice crystallize. Given

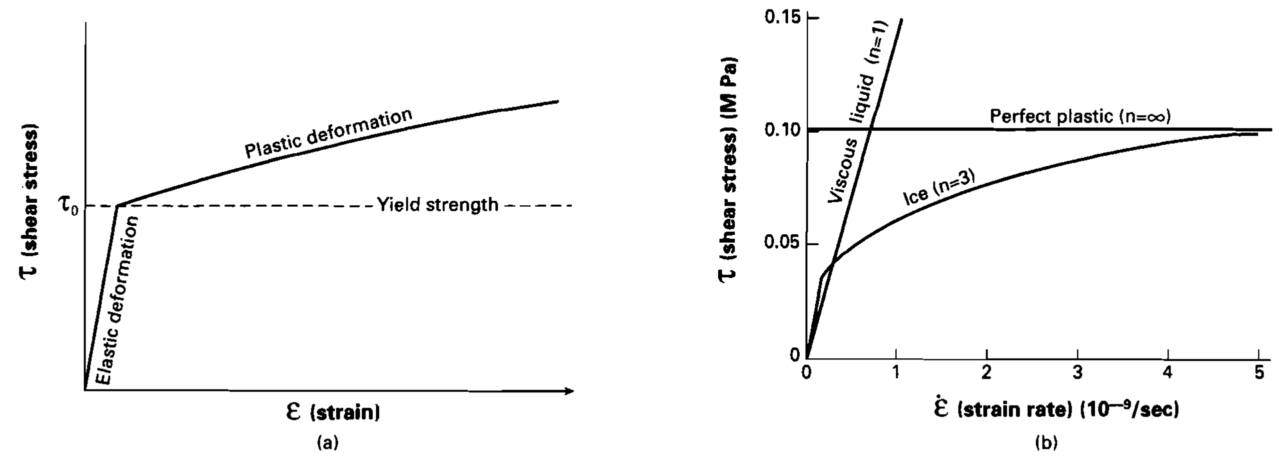


FIGURE 16-10. (a) A schematic typical stress-strain curve. τ_0 is the yield strength, marking the onset of plastic creep, ε is the strain, or deformation. (b) Shear stress τ versus strain rate $\dot{\varepsilon}$ for ice, with n = 3 (p. 362). For comparison, the strain rates of a "perfect" plastic (n = ∞) and a viscous liquid (n = 1) are shown.

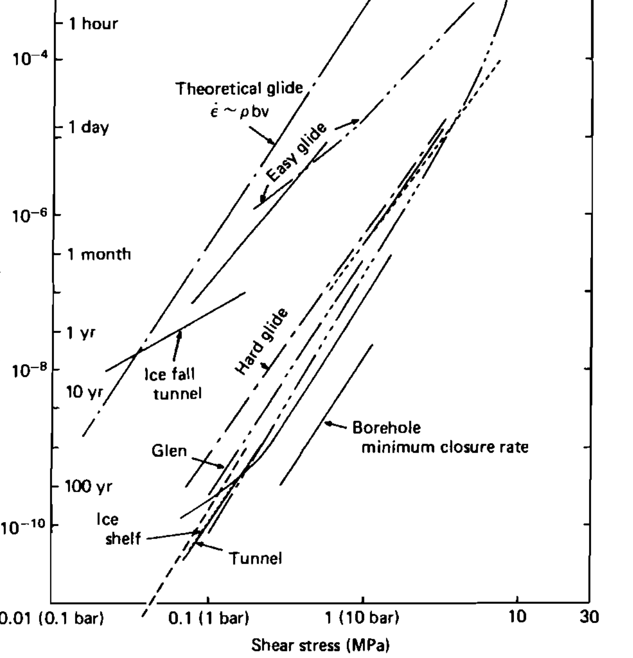


FIGURE 16-11. Log-log plot of steady-state or the minimum creep rate versus stress. Creep rate is normalized to −10°C. Creep rate and stress are for, or have been changed to, the equivalent of a uniaxial tension or compression test. Axes are reversed from Figure 16-10b. Data from numerous sources listed in Weertman (1983) (reproduced with permission from the Annual Review of Earth and Planetary Sciences, v. 11, © 1983 by Annual Reviews Inc.).

the nearly uniform specific gravity of glacier ice, the pressure-melting temperature of ice (the temperature at which ice and water phases coexist) decreases at a rate of about 0.7°C/1000 m of ice thickness. Thus, if water is encountered at any depth within a glacier, the temperature at that depth can be precisely calculated (Figure 16-12). From the foregoing discussion, two categories of glaciers can be visualized. One category is of glaciers that are cold from top to bottom, with the entire thickness of ice well below the pressure-melting temperature. The other category is of glaciers that are at or very close to the pressure-melting temperature so that small pressure variations can produce melting. Actually, this thermal classification is better applied to parts of glaciers rather than to entire valley glaciers or ice sheets because the thermal status of one part of a glacier may be different from that of another part. A discussion of the thermal status of glaciers and parts of glaciers uses terminology first proposed more than 60 years ago (Lagally, 1932; Ahlmann, 1933, 1935), but the examples and conclusions are much more recent. The thermal condition of glaciers has a very important bearing on both their form and their ability to shape landscapes.

Cold, Polar, or Dry-base Glaciers. A 1387-m borehole through the Greenland ice sheet at Camp Century, about 100 km from the west edge of the ice sheet, was completed in 1966. At the time, it was the deepest borehole made in ice and the first to penetrate entirely through an ice sheet. The temperature profile of the Camp Century borehole (Figure 16-12) illustrates well the characteristics of a **cold,** or **polar,** glacier. The ice temperature at a depth of 10 m was −24°C, very close to the mean annual air temperature at Camp Century. A minimum temperature of −24.6°C was reached at a depth of 154 m; then the temperature rose steadily to −13.0°C at the base of the ice. The entire thickness of ice at the Camp Century borehole is far below the pressure-melting temperature. No water phase can exist, and the rather slow surface motion of only 3.3 m/yr must be due entirely to plastic creep and minor brittle fracturing near the surface.

Cold, or polar, glaciers are also called **dry-base** glaciers, because the basal ice must be frozen firmly to the rock beneath. Experiments have demonstrated that the surface bond between rock-forming minerals and ice is considerably stronger than the yield strength of ice, so that any motion of a cold glacier must be by internal plastic deformation. Just one consequence of this fact is that cold glaciers cannot be expected to erode by sliding over bedrock. Most of the motion is concentrated in the basal, relatively "warm" ice, however, and differential shear in the basal zone might break off rock projections or fracture larger rock fragments during transport.

Temperate Glaciers. As the confining pressure due to the weight of the overlying ice increases downward in a glacier, the melting temperature is depressed. If the mean annual surface temperature of the accumulating snow and ice is close to 0°C and melting occurs, then with progressive burial, pressure melts additional ice, and the ice and water phases coexist at progressively lower temperatures with depth. This is the definition of a **temperate** glacier.

An important characteristic of a temperate glacier derives from the negative slope of the pressure-melting curve (Figure 16-12). Because the temperature at the base is less than at the surface, no geothermal heat can flow upward as it does in a polar glacier. The geothermal heat that reaches the base of a temperate glacier cannot be conducted upward but must melt basal ice. This geothermal heat, plus an increment of heat generated by sliding friction that is also trapped at the base of the ice, assures that a temperate glacier must move on a wet base.

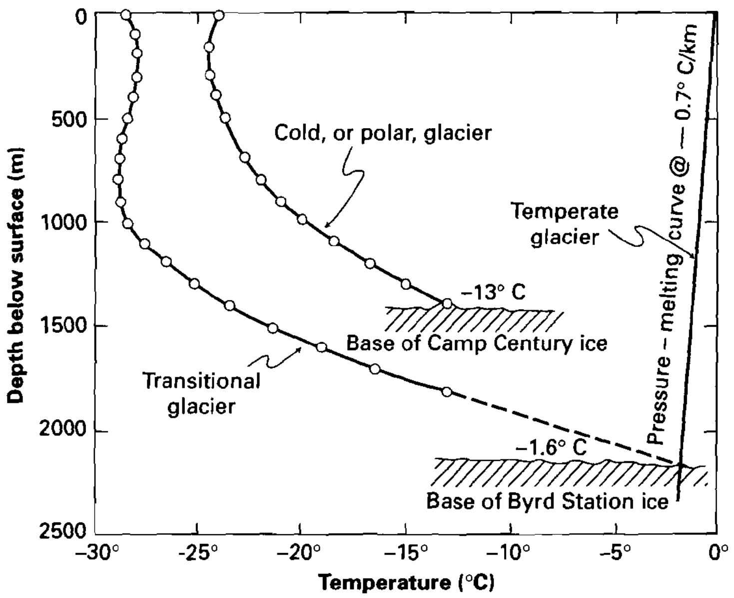


FIGURE 16-12. Temperature profiles of boreholes at Camp Century, Greenland (Weertman, 1968), Byrd Station, Antarctica (Gow, et al., 1968), and a hypothetical temperate glacier.

Flow rates in temperate glaciers are likely to be at least twice as fast as in polar glaciers. Striated, grooved basal rock surfaces are possible, as are numerous other significant forms of glacier transport and erosion (Chapter 17) because the ice slides over wet rock or saturated sediment. In addition, the subice water adds its erosional, transportational, and depositional work to that being done by the ice (Chapter 17).

Intermediate Categories. Ahlmann (1933, 1935) subdivided polar glaciers (*arctic* glaciers in his original terminology) into two classes, *sub-polar* and *high-polar.* Although both are cold at depth, summer melting on the sub-polar glacier might produce a surface layer 10 to 20 m deep, which is at the melting temperature. The various zones of dry snow, wet snow, and water percolating into surface ice and refreezing are well shown by satellite radar images (Fahnestock et al., 1993). Ahlmann's high-polar glacier would be typified by the Greenland ice sheet at Camp Century (Figures 16-6 and 16-12) in which not even a surficial layer of snow melts in the summer.

Lagally (1932), in a classification otherwise almost identical to Ahlmann's polar and temperate glaciers, proposed a *transitional* type that would be cold or polar at the top but be at the pressure-melting temperature near the base. The transitional category of Lagally was accepted as an interesting theoretical possibility but was generally overlooked until 1968, when a borehole was drilled at Byrd Station in Antarctica (Figure 16-12) (Gow et al., 1968). At a depth of 2164 m, the bottom of the ice was reached, but at the basal contact was a film of water judged to be at least 1 mm in thickness, under enough pressure to rise 50 m up the drill hole and displace the antifreeze solution that had been used to keep the hole open. At the base of the ice sheet, the calculated melting temperature is −1.6°C. The surface temperature is −28.8°C. The steeper-than-average thermal gradient in the lower half of the Byrd Station borehole suggests a geothermal heat flow about 50 percent greater than the world average. Because West Antarctica is an ice-covered volcanic archipelago, the higher heat flow is reasonable. The effectiveness of the wet-base condition is demonstrated by the fact that the Byrd Station drill bit was unable to obtain a sample of the subice material because the ice moved laterally several centimeters per day and constantly pinched the drilling tool in the borehole. The substrate may be either rock or saturated, clay-rich basal sediment.

Subsequent radar depth soundings of the entire Antarctic ice sheet have indicated that large areas of the basal ice are wet and have a basal shear stress of only about 20 kPA (0.2 bar) (Oswald and Robin, 1973; Drewry, 1983; Radok, 1985). Surprisingly then, much of the Antarctic ice sheet is wet based in spite of the extremely cold surface temperatures. Even more surprising is the identification by radar and seismic exploration of a large lake, 14,000 km^2 in area and as much as 500 m in depth, 4 km below the central Antarctic ice (Kapitsa, et al., 1996; Ellis-Evans and Wynn-Williams, 1996). Since liquids have no yield strength (Figure 16-10), this and other large areas at the base of the Antarctic ice sheet are floating with zero basal shear stress. The implications for ice-sheet stability are the subject of intense debate.

GLACIER MOVEMENT

Plastic Flow

A slab of ice frozen to an inclined rock surface will deform downslope as a solid, without any melting and refreezing. The rate of movement at the ice surface is in nonlinear proportion to the thickness and surface slope of the ice (Paterson, 1994, p. 240). Theory and observation of actual glaciers agree that the motion is most rapid at the surface and progressively less rapid with depth (Figures 16-4 and 16-13). Theory and observation also agree that the direction of the flow is controlled by the surface slope of the glacier, not by the slope of the rock floor beneath it, as long as the ice thickness is significantly greater than the bedrock relief. It is quite possible for basal ice to move into and out of an ice-filled basin or over an obstruction, controlled only by the slope on the surface of the ice above the basin or obstruction. Thus ice can flow "uphill" over and around obstacles on its rock floor.

Theoretical models that assume ice to be a plastic solid with rheidity defined by Glen's flow law give excellent predictions of the shape, thickness, and rate of flow both of glaciers and of floating ice shelves, but the results are not perfect. An obvious reason is that real glaciers are complex masses of ice, with internal temperature gradients and impurities, that move over rock surfaces along discontinuities that cannot be modeled theoretically. Also, during movement, deforming ice crystals in a polycrystalline mass grow and reorient themselves in the stress field by a process of solid-state recrystallization. Thus the grain size and orientation, which greatly affect the rate of deformation, change during deformation (Hooke, 1998, pp. 36ff). Ice under high shear stress may fracture or form small, interlocking grains, but if the stress is low and uniform, these small grains "anneal" to form oriented grains as wide as 10 cm. Ice recovered from deep boreholes in glaciers usually has crystals several cm across, oriented with their *c* optical axes parallel and vertical; in this configuration, horizontal "easy glide" flow is facilitated, and the deformation rate increases by a factor of 4 over the rate for randomly oriented grains (Figure 16-11) (Weertman, 1983, p. 229).

Brittle Fracture: Faults and Crevasses

In their thin downstream or marginal regions, glaciers also "flow" by movement along discrete shear planes such as thrust faults. Pressure of moving ice against marginal ice that is either frozen fast to rock, or too full of rock debris to move, or blocked by a large obstruction, causes a thrust fault to form, dipping upglacier. Actively moving ice shears forward and over the stagnant basal portion. Thrust faults of this sort may reach entirely through the glaciers and are one means by which rock debris from the base of the ice is brought upward to the surface.

Brittle failure also contributes to surface topography. Wherever a glacier moves over an obstruction at shallow depths or drops down a buried cliff, the surface ice is fractured by a maze of crevasses, mostly vertical and extending tens of meters in depth. These are tension cracks that heal at the depth where the tensional stresses can be compensated by plastic deformation. The deepest known crevasses, in the cold brittle ice of Antarctica, are about 50 m deep. Crevasses often become bridged by drifting snow and are one of the major dangers for explorers. A prominent arcuate crevasse, called the **bergschrund,** is at the extreme head of most valley glaciers (Figure 17-3). It is caused by snow compaction and ice motion away from the rock face. Deep crevasses in valley glaciers provide a route for surface rock debris to be incorporated into deeper ice.

Basal Slip: Enhanced Plastic Flow and Regelation Slip

The strain rate of glacier ice is typically proportional to the third power of the shear stress (Figure 16-11). Near

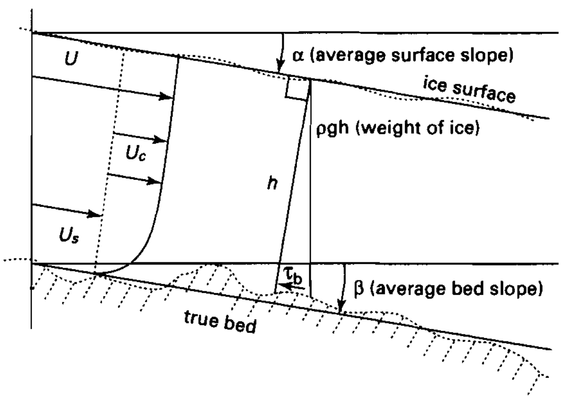


FIGURE 16-13. Simplified cross section through a glacier moving at a constant velocity on a sloping bed. Surface velocity U is the sum of plastic creep (U_c) and basal slip (U_s). Forces at the bed show that the resisting component τ_b (basal shear stress) balances the *driving stress* parallel to the bed, which is the weight of the overlying ice (ρgh) multiplied by the sine of the surface slope: $\tau_b = \rho gh \sin \alpha$. For typical gradients < 10°, bed slope (β) is irrelevant.

the base of a glacier, where bedrock irregularities protrude into the ice, local shear stresses increase, and local ice deformation is correspondingly very greatly increased. This **enhanced plastic flow** is proportional to the size of the obstruction, so the larger the obstruction, the more important is the enhanced plastic flow. Although strictly speaking the process is one of plastic flow, it is restricted to a zone near the ice-rock interface, and is usually regarded as a form of basal slip.

If ice is at or very close to its melting temperature, an additional form of basal movement can occur, called **regelation slip.** This is not a solid-state phenomenon but involves pressure-dependent melting and freezing. If the ice is at the melting temperature for the appropriate confining pressure, on the upglacier side of an obstruction the ice presses against rock, locally increasing pressure, depressing the melting temperature, and producing water *at the same ambient temperature as the surrounding ice.* The water migrates along grain boundaries within the ice or over the rock surface to regions of less pressure, especially on the downglacier side of the obstruction where ice is being drawn away from the rock. Here the water refreezes as bubble-free clear blue *regelation ice* (Figure 16-14).

Regelation is essentially an isothermal process. The only thermal gradients are those produced by the pressure gradients around the obstruction. The heat released by refreezing on the downstream side of the obstruction migrates upstream, primarily through the obstruction, to be absorbed by the melting at the upstream side. The heat transfer is faster through small obstacles, so the smaller the obstruction the more important is regelation slip.

Mathematical models of basal slip are imperfect because it is essentially the result of boundary conditions between unlike materials and phases. The controlling factor, assuming that the temperature is at or very near the melting point, is the bed roughness, or size of the limiting obstruction (Drewry, 1986, p. 10; Paterson, 1994, p. 139). Large obstructions are more efficiently passed by enhanced plastic flow, but smaller obstructions are more efficiently passed by regelation slip. As shown by borehole measurements (Figure 16-4), basal movement (including regelation slip) can account for 50 percent or more of the total observed surface motion of valley glaciers. Theory suggests that the controlling wavelength of bedrock obstruction, which would permit the optimal contributions by both plastic deformation and regelation slip, is about 0.5 m (Paterson, 1994, p. 142). However, obstructions of this size are not common on glaciated rock surfaces. Streamlined glaciated landforms about ten times as large are common and are thought to be stable forms under moving ice, somewhat like ripple marks and dunes under moving water and air. The lack of agreement between theory and observation illustrates the imperfections of present understanding of basal sliding.

Water at the Base of Glaciers: Surges

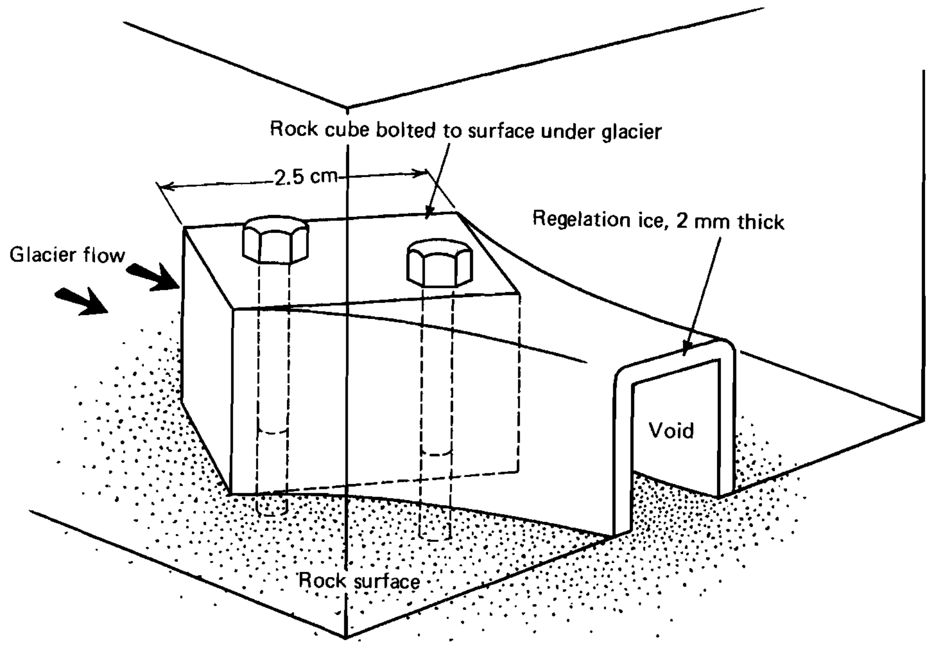
Glacier Hydrology. Water occurs on (*supraglacial*), within (*englacial*), and beneath (*subglacial*) temperate glaciers, at the base of transitional glaciers, and seasonally on and in some cold glaciers as well. It flows as rivers or collects in meltwater lakes on the surface of glaciers, flows along pressure gradients through karst-like enlarged fractures within the ice (Figure 16-15), and flows along the base of the ice over impermeable beds or permeates into suitable substrates. Many analogies with karst hydrology (Chapter 8) can be found in glaciers, although the opening of conduits in glaciers is by melting rather than solution, and the flow of ice may close established conduits and open new ones (Menzies, 1996; Hooke, 1998, Chapter 8).

Moulins serve as conduits for surface meltwater to flow down into the glacier, eventually joining subglacial channel systems. They are inclined as much as 45° toward the glacier margin because the water tries to flow along the internal water pressure gradients, perpendicular to the equipotential lines (Figure 16-15) in the saturated zone of the glacier (Holmlund, 1988).

Enhanced plastic flow and regelation slip are typically associated with a film of water at the ice-rock interface, but the most important effect of the water layer is to reduce the resistance to basal sliding. Bed roughness is an important factor in controlling glacier flow (Figure 16-13). Roughness typically includes a wide range of dimensions, as shown by fractal analysis (Figure 1-2). If the water film covers and negates the smaller part of the roughness spectrum, the flowing ice "sees" only the fewer, larger irregularities, and will slide much faster (Weertman and Birchfield, 1983, p. 22). By carefully mapping the glacial erosional features on a recently deglaciated limestone surface, Walder and Hallet (1979) concluded that at least 20 percent of the surface area had been separated from the base of the glacier by water-filled channels, with the remainder covered by a thin water film.

Cavities with dimensions of several meters may be opened on the downglacier side of rock knobs that protrude into temperate glaciers. These become water filled, especially during the summer melt seasons, and may coalesce into a system of water-filled channels with water pressure almost equal to the cryostatic pressure. Channels may be closed by ice flow during the winter season, and reopen or form in new areas during subsequent meltwater seasons.

FIGURE 16-14.
Experimental demonstration of the formation of regelation ice beneath a temperate glacier (from Peterson, 1970).



Complex sets of subice tunnels have been observed under an outlet glacier from a small Norwegian ice cap that is used for hydroelectric power generation (Hooke et al, 1985), as well as under other temperate glaciers (Hock and Hooke, 1993).

Deformable, or Soft, Beds. Where wet-base glaciers move over a layer of saturated sediment rather than hard rock, an additional important mode of glacier movement has been demonstrated. The sediment layer becomes a **deformable bed**, dragged in the direction of glacier movement by the transfer of the glacier's basal shear stress downward into the sediment layer (Boulton, 1996). As shown by the Coulomb equation that governs mass movement (Figure 9-1), pore pressure within the sediment layer reduces the overburden pressure, and the *effective* shear stress on a glacier bed of saturated sediment can be reduced to near zero (Iverson et al., 1995). If the sediment is fine grained, especially silt-size, drainage is inhibited and pore pressure is maximized. Sand and gravel beds may be well drained and have much lower pore pressure. Clay-rich beds inhibit infiltration from the overlying ice, and by decreasing the effective shear stress at the base of the overlying ice, can produce rapid basal slip without much deformation of the underlying sediment.

Surging Glaciers. In a small percentage of glaciers, the ice has been observed to move suddenly forward, advancing at a rate of kilometers per year instead of the average rate of a few hundred meters per year (Paterson, 1994, Chapter 14). These **surging glaciers** typically flow normally for 10 to 100 years, then suddenly surge forward. The zone of rapid motion begins high in the glacier, and advances wavelike to the terminus. The surging part of a glacier is dominated by extending flow, but the lower end is in compressing flow until the surge reaches it, when it rapidly thins and surges forward (McMeeking and Johnson, 1986).

Surges are unknown in many glaciated regions, yet over 6 percent of the 2356 glaciers in the St. Elias Mountains on the Alaska-Canada border have surged, and other concentrations are in specific regions in Asia, the Andes, Iceland, eastern Greenland, and Svalbard

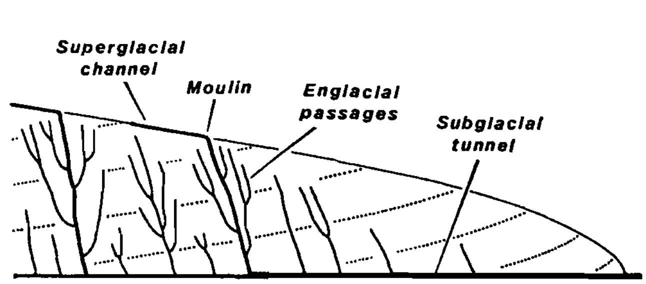


FIGURE 16-15. Equipotential lines (dotted) and internal drainage in an ideal cross section of a temperate glacier on a horizontal, impermeable bed. Flow is perpendicular to the equipotential lines and toward the outlet of the subglacial tunnel (Shreve, 1985, Figure 4).

(Clarke et al., 1986). Surges rarely last for more than a few months to a year, but they create a chaotic pattern of crevasses on the glaciers (Figure 16-16) (Kamb et al., 1985). They were first reported on a group of Alaskan glaciers early in this century (Tarr, 1909) and were attributed to an excessive accumulation of snow from avalanches produced by an earthquake in 1899. This idea has not been supported by subsequent observations, but after the major Alaskan earthquake of 1964, a close watch was kept on the glaciers of the Alaska Range in hopes of observing the initial phases of surging.

Events leading up to the well-documented 1982-1983 surge episode of Variegated Glacier in Alaska had been monitored for a decade (Kamb et al., 1985; Raymond and Harrison, 1988). The ice in the middle and upper parts of the glacier had been thickening since 1973, and the velocity of flow had been increasing, even though the terminus remained stagnant. Until 1978, most of the increased rate of flow was during the summer, but subsequently the ice began to move more rapidly in winter as well. This indicated that basal sliding was accelerating even in winter. In 1982, the surface of the glacier in the accumulation zone began to move forward at rates of 2 to 15 m/day. The surge then propagated rapidly downglacier, increasing from a presurge flow rate of 0.2 m/day or less to 40 to 60 m/day. The surface of the glacier was intensely crevassed by the rapid motion. A borehole to the base of the glacier confirmed that 95 percent of the surging motion was by basal sliding. The zone of water saturation in the ice had risen to within 90 percent of the ice thickness. A basal layer of impermeable, shearing sediment was detected by seismic reflection techniques beneath Variegated Glacier in Alaska prior to the 1982-1983 surge. It may have helped confine water near the bed and build the water pressure to a level at which the glacier was nearly floating during the surge (Richards, 1988). When the water in the ice migrated down to the terminus and drained away in a spectacular flood, the surge stopped. The excess water that accumulated within the glacier was only indirectly correlated with air temperature and precipitation. Apparently, it is a characteristic of Variegated Glacier to accumulate water over an interval of about 20 years, then rapidly drain during a surge (Kamb et al., 1985).

Flow in Ice Sheets

For various reasons excluding hydrostatic pressure, shear strain (the vertical gradient of velocity) in an ice sheet, as in a valley glacier (Figure 16-13), should be most rapid near its base. Ice at higher temperatures deforms more rapidly, and the highest temperatures are near the base of an ice sheet because of the poor thermal conductivity of ice and the geothermal heat flow (Figure 16-12). The increased grain size of the ice crystals and their progressively better orientation with each other in the horizontal plane also facilitate plastic flow. A contribution of frictional (deformational) heat must also be considered. Further, the greatest shear stress is at the base of the ice although this factor does not vary greatly. If any regelation slip occurs, it too will be at the base. Therefore, reasonable models of flow in ice caps are obtained by assuming that all the shear takes place in a basal layer only a few meters to tens of meters in thickness, and the overlying ice is passively transported. Under such assumptions, flow rate is proportional only to the surface slope of the ice, which develops an elliptical or parabolic profile (Figures 16-5 and 16-6). An excellent approximation of the thickness of many ice sheets and ice caps within a few hundred kilometers of their margins is given by the equation

$$h = 3.4\sqrt{d}$$

where h is the thickness of the ice (in meters) on a horizontal base at distance d in from the margin of the ice sheet (in meters). This equation is derived from Glen's flow law (p. 362) with an assumed basal shear stress of 50 kPa (0.5 bar) (Paterson, 1994, p. 242). A few minutes with a calculator will demonstrate how well this equation approximates actual surface profiles of ice sheets and ice caps (Figure 16-5). However, many of the western and southern lobes of the Laurentide ice sheet during the last ice age (Figure 16-1) had surface slopes considerably flatter than would be predicted by the equation, implying basal shear stresses as low as 1 to 30 kPa (0.01 to 0.3 bars) for ice that flowed northwest across the Canadian arctic near the Alaskan border (Beget, 1987), and only 0.7 to 22 kPa (0.007 to 0.22 bars) for ice lobes in the Great Plains and Great Lakes lowlands (Clark, 1992, 1994). Such shear stresses are much too low to be the result of ice deformation. Rather, they have been attributed to shearing in a deformable bed of water-saturated sediment at the base of the ice.

The pattern of flowlines in the former North American Laurentide ice sheet (Chapter 17) suggests that where the ice was moving primarily over resistant rock such as the metamorphic and igneous terranes of the Canadian Shield it moved primarily by plastic deformation, enhanced basal plastic flow, and regelation slip. However, when it expanded onto the more easily erodible sedimentary-rock terranes of the northern Great Plains and the Interior Lowland around the Great Lakes (Figure 18-7), a deforming bed of fine-grain sediment greatly accelerated movement (Clark, 1994). The implications of such a deforming bed, with resulting

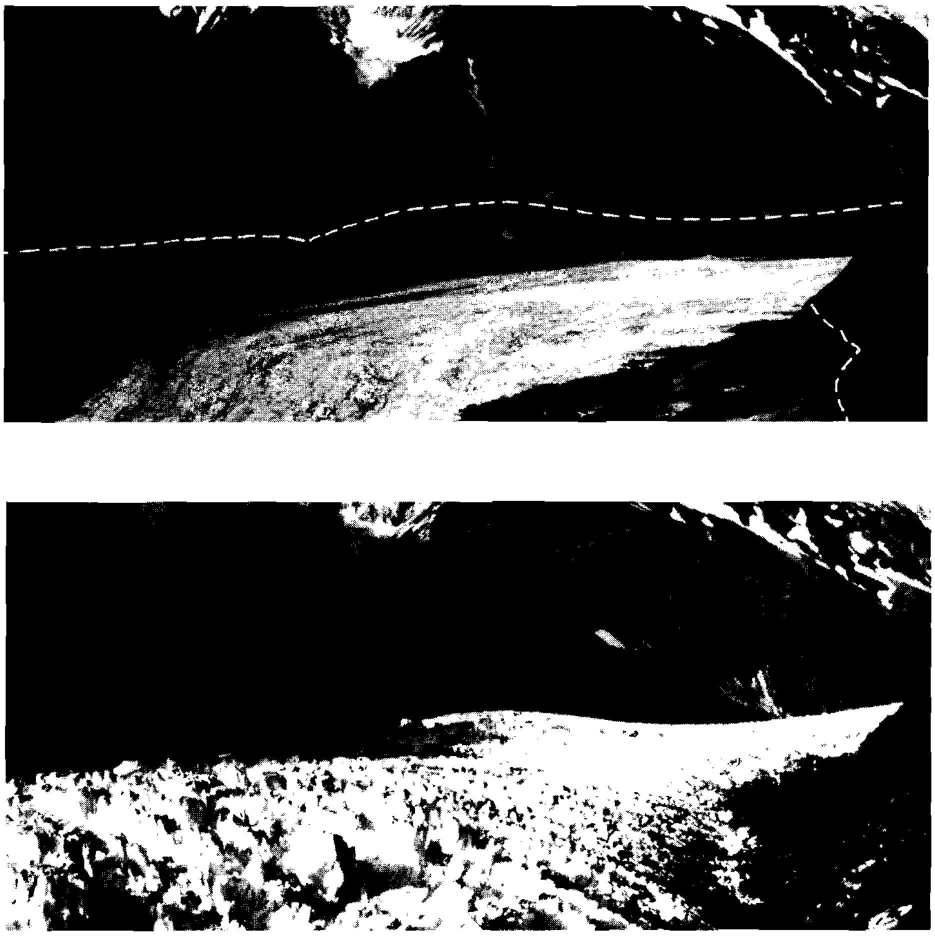


FIGURE 16-16. Views upglacier of the Variegated Glacier, Alaska (a) before a surge, July 1982, and (b) during a surge, 4 July 1983 (Kamb et al., 1985, photos courtesy B. Kamb).

(a)

(b)

very flat surface gradients on the ice, much thinner ice than previously estimated, highly lobate and rapidly fluctuating ice margins, and less than expected postglacial isostatic recovery, are considered further in Chapters 17 and 18.

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