11
Ice

> My theory of glacier motion then is this – A glacier is an imperfect fluid, or a viscous body, which is urged down slopes of a certain inclination by the mutual pressure of its parts.
>
> J. D. Forbes, *Travels Through the Alps of Savoy* (1846), p. 365

Scottish physicist J. D. Forbes (1809–1868) was an inveterate mountaineer, observer, and scientific quarreler. His direct observations of glacier flow led him to describe glaciers as highly viscous fluids, a conclusion that could not be reconciled with the mechanical understanding of his day. He spent many years ferociously defending his view against the opposing opinions of another mountaineering physicist, John Tyndall (1820–1893), an Irishman who argued that glaciers deform by regelation, a process in which ice melts and refreezes under pressure.

Both Forbes and Tyndall had become fascinated by the mechanics of glacial phenomena at a time when Swiss paleontologist Louis Agassiz (1807–1873) was announcing his deduction that Europe had once been smothered under immense glaciers during a former age of ice. Although Forbes' views are now accepted as being closer to the truth, a full understanding of how crystalline ice can flow like a viscous fluid was not achieved until the 1950s. Tyndall's regelation idea still finds a place in glacier mechanics, although it is now relegated to sliding at the base of warm-based glaciers.

Ice on the surface is mirrored by ice in the ground. The surfaces of cold regions are often underlain by frozen ground that, because of seasonal thermal cycling, is unusually active and highly productive of distinctive organized patterns.

The slow flow of ice, whether frozen water or other substance, as an "imperfect fluid or a viscous body" creates landforms that are different from those produced by wind or water. Ice occurs either disseminated through the regolith, in small bodies properly called glaciers, and in large ice sheets that can cover large areas.

11.1 Ice on planetary surfaces

Glaciers are the counterparts of rivers in a cold hydrologic cycle. Ice accumulates as snow at high elevations, slowly flows down to lower elevations, and there melts, eroding and transporting the adjacent rock as it moves. Glaciers on Earth currently underlie about 10% of its surface and incorporate about 3% of its water, a number that is rapidly declining in the modern anthropogenic world. During the Earth's recent Ice Age glaciers and ice sheets occupied 30% of the surface and locked up 8% of the water. Water-ice glaciers are apparently active on Mars at the present time and, on Mars as on the Earth, seem to have been much more extensive in the past, although the former Martian Ice Age occurred much farther back in time than the Earth's. Mars may also host glaciers composed of solid CO_2, which require temperatures lower than any achieved on Earth. So far, no glaciers have been found elsewhere in the Solar System: Titan's surface is just a bit too warm for methane to freeze into ice. Ammonia glaciers are theoretically possible but no examples have yet been discovered. Although some researchers have speculated about solid nitrogen glaciers on Triton, no glacial features have yet been identified there. Earth, however, hosts a very unusual type of "glacier" composed of salt (halite), NaCl, that flows by virtue of interaction with small amounts of water.

11.1.1 Ice within the hydrologic cycle

The fluvial hydrologic cycle begins with water falling on the surface as rain, running downhill and picking up sediment, then flowing to an accumulation point where the sediment is dumped into a delta or moved further by longshore and marine currents. In an analogous manner, snow that falls on mountaintops or cold regions, where snowfall exceeds melting, accumulates into a permanent snowfield. If this accumulation were to continue without limit, the Earth's water would soon end up frozen into huge ice mountains. However, snow metamorphoses into ice, which can flow off the land surface as a thick, viscous liquid and, thus, returns the water to lower elevations, where it melts back into water to close the cycle.

Different portions of a glacier or ice sheet can, thus, be distinguished by their function in the overall mass balance as either accumulation areas or ablation areas, which are connected by flowing ice (Figure 11.1). These regions are readily recognized in a terrestrial glacier by color and texture (or thermal inertia) at the end of the summer season: Accumulation areas are bright white, underlain by coarse snow (firn), which is undergoing the transformation into ice, while ablation areas appear as dense blue ice that is in the process of melting. Typical glacier flow velocities range from 0.1 to 2 m/day, exceptionally up to 6 m/day. Large outlet glaciers from ice sheets may flow up to 30 m/day. Some glaciers are observed to suddenly speed up to rates of 70 m/day in brief episodes known as surges. The mechanical basis of surges was once mysterious, but we now know that surges are due to changes in the subglacial plumbing of warm-based glaciers, discussed below.

The heads of glaciers and ice sheets are in areas where more snow accumulates each year than melts, and they flow down to elevations where melting exceeds snowfall, often to altitudes well below the snowline at which accumulation and melting are in balance. A glacier is, thus, in a state of dynamic balance between accumulation and melting.

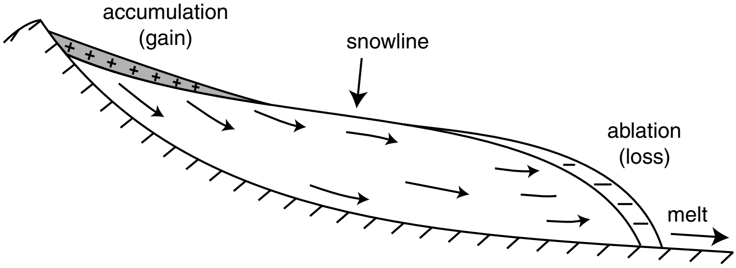

Figure 11.1 Schematic plan of a valley glacier, illustrating the accumulation zone at high altitude, flow of glacier ice downslope, and the melting of the ice in the low-altitude ablation zone. Depending on its temperature, the ice at the base of the glacier may or may not be able to slide over the bed. Inspired by Figure 1 of Sharp (1960).

Accumulation may occur by means other than snowfall directly onto a glacier or ice sheet. Avalanches from adjacent highlands or wind-blown snow (especially important on icecaps where precipitation is low) may deposit snow or even ice directly onto the accumulation area. Valley glaciers, like rivers, may be fed by tributaries that themselves originate in merging ice streams, although the number of links in such networks is usually small.

Once snow falls on a glacier, it undergoes a regular series of changes as it metamorphoses from new snow, to old snow, to firn (density about 550 kg/m^3), then finally into glacier ice (density 820 to 840 kg/m^3). These processes, discussed in Section 7.2.2, involve the interaction with seasonal liquid meltwater and vapor-phase transport within the snowpack. Atmospheric gas bubbles, presently of great importance for measuring the composition of Earth's pre-industrial atmosphere, may be trapped in the process and preserved for many thousands of years.

Ablation, or mass wasting, of the ice is usually by melting. Evaporation is generally unimportant except for tropical glaciers and in the dry valleys of Antarctica. Tidewater glaciers and continental ice sheets, however, may lose most of their mass by calving of icebergs into the sea.

11.1.2 Glacier classification

A common classification of glaciers is based on their morphology. There are three general types: *Icecaps* or *ice sheets* are continuous sheets of ice. Their flow is centripetal, from a high-standing center toward their edges. Terrestrial examples are the Greenland and Antarctic ice sheets or the former Pleistocene ice sheets. A former Martian ice sheet may have covered large areas in its Southern Highlands. *Valley glaciers* are ice streams that have heads in mountainous terrain. They are common in the Earth's high mountains, such as the Alps of Europe, Himalayas of Asia, or the northwest coast of North America. The North American glaciers are noteworthy as most of them head at low elevations – often

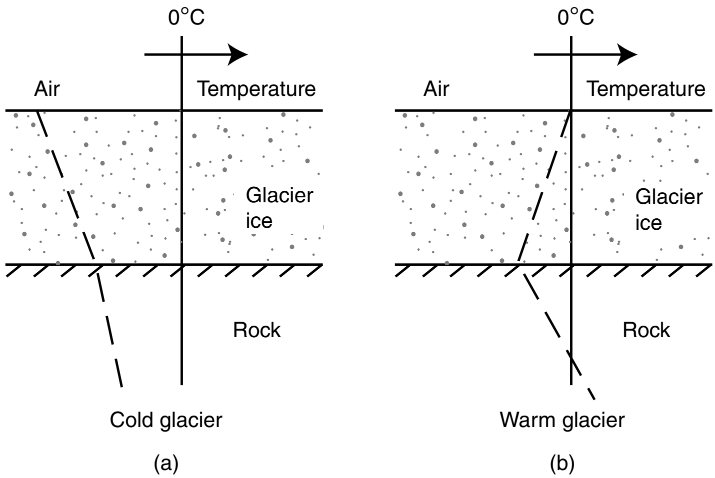

Figure 11.2 Thermal regimes of glacier ice. (a) Illustrates a cold-based (or polar) glacier, in which the temperature is everywhere below the freezing point of water and geothermal heat is conducted to the surface. (b) Illustrates a warm-based (or temperate) glacier in which the temperature is at the pressure-controlled melting point of ice. Because of the inverse dependence of the ice's melt temperature on pressure, the temperature is slightly colder at the base than at the surface, which implies that heat is conducted to the glacier bed. After Figure 5 of Sharp (1960).

only 2000 m – where they are nourished by the heavy winter snowfall from the adjacent Pacific Ocean. *Piedmont glaciers* are ice sheets formed by the coalescence of several valley glaciers on flat terrain at the base of mountains. An example is the Malaspina Glacier in Alaska.

A second classification is based on temperature. Cold glaciers (alternatively called polar glaciers) are easiest to understand. The temperature throughout the ice body of the glacier is below the pressure melting point of ice (Figure 11.2a). The normal gradient of the internal temperature of the Earth is continued through the ice, from warmer below to colder near the surface, although the slope is different because the thermal conductivity of ice is somewhat different from that of rock (Table 4.2). Cold glaciers are frozen to their beds. Their motion occurs by internal deformation of the ice itself through solid-state creep (Section 3.4.3). They are among the most slow-moving of terrestrial glaciers and they are generally ineffective in eroding their beds. Often, where a polar glacier has melted away, there is little evidence of its former existence.

Warm glaciers (alternatively called temperate glaciers) are at the pressure melting point of ice throughout their mass. Because of the inverse slope of the melting curve of water ice, $dT_m/dz = -0.65$ K/km of ice, the temperature actually *decreases* with increasing depth in the glacier (Figure 11.2b). This inverse gradient means that thermal conduction moves heat from the surface *downward*, toward the glacier bed. At the same time, the normal geothermal gradient in the rock below the glacier moves heat upward toward the glacier bed with a slope of about 30 K/km. This creates a thermal crisis for the glacier, which responds by

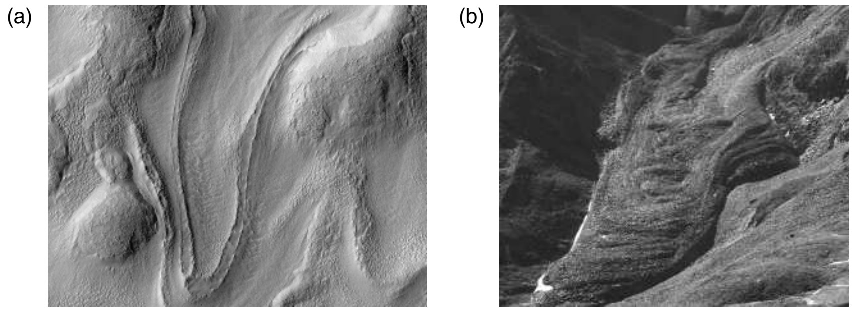

Figure 11.3 (a) A tongue-shaped flow on Mars located on the eastern wall of Hellas Planitia. This flow is about 5 km long and 1 km wide. It is likely to be a Martian analogue of terrestrial rock glaciers. Image PIA09594_fig 1, portion of HiRISE image PSP_002320_1415. NASA/JPL/University of Arizona. (b) Jungtal rock glacier in the Swiss Alps (image courtesy of Dr. Jan-Christophotto, 2011).

melting at its base, converting about 1 cm of ice into water each year. Warm-based glaciers are saturated with liquid water, which is in equilibrium with ice throughout the body of the glacier. Temperatures remain everywhere at the pressure melting point, but a great deal of heat is, nevertheless, transferred in such glaciers by the latent heat from the conversion of solid ice to water, which can flow readily from place to place transferring its latent heat as it moves and then freezes.

Warm-based glaciers can slide over their beds and, with the aid of rocks and debris frozen into the ice, are very effective at abrading and quarrying out the underlying bedrock. These glaciers also deform internally: Typically about half their surface velocity is due to internal deformation and half is due to basal sliding.

Temperature is not, however, always a good classification for the entire glacier, because the thermal regime can change with position in the glacier. Thus, a glacier's upper reaches could be "cold" while its lower parts are "warm." Moreover, parts of the Antarctic ice sheet have meltwater near their beds, indicating a warm-based regime, while their surfaces are cold, well below the pressure melting temperature.

11.1.3 Rock glaciers

Although sometimes not considered "proper" glaciers, rock glaciers are dense mixtures of rock and ice that, despite being mostly composed of rock, nevertheless show clear evidence of flow, albeit moving much more slowly than the mostly pure ice glaciers familiar to glacial geologists. Rock glaciers creep along at rates of centimeters to meters per year, but exhibit the lobate margins, drapery-like ridges, and lateral moraines typical of valley glaciers. Their margins are typically steep, at or close to the angle of repose. They are included here because recently discovered Martian glaciers may be rock glaciers, not solid ice (Figure 11.3).

Mixtures of ice and rock in glaciers form a continuum, running from nearly pure ice, through ice carrying small quantities of rock and debris, to rock glaciers, which are mainly composed of rock debris. Some rock glaciers have ice-rich cores mantled with ice-free rock debris, while in others the ice merely fills the interstices between boulders. Rock glaciers have not received much study, partly because of their rarity and partly because of the difficultly of probing into their interiors: Unlike glaciers, one cannot simply melt boreholes through them with electric heaters or hot steam.

The detailed mechanism by which rock glaciers deform internally is not well understood, in spite of finite element modeling of their flow (see the review by Whalley and Azizi, 2003). The slow creep of the rock/ice mixtures must be due to the included ice, but models of how the heavy burdens of rock debris affect the flow rate and its dependence on factors such as shear stress are not well developed.

11.2 Flow of glaciers

Many theories of how glaciers flow have been proposed since Agassiz brought the importance of glaciers to the attention of geologists. These historical theories include crevasse filling and refreezing, regelation within the mass of ice, and many others. The reason for so many theories was the apparent paradox of a crystalline solid (you can easily see the crystals in partially melted specimens of glacier ice) that, nevertheless, flows like a fluid.

We now understand that "solid" is a poor description of a crystalline material near its melting point, because any material can flow, although the motion is perceptible only very near the melting point. This was first demonstrated for ice by glaciologist J. W. Glen in 1955, although this sort of "creep" had been observed in metals and some rock materials long before. Individual crystals flow by the generation and movement of a peculiar sort of crystal defect known as a dislocation. Dislocations and their dynamics were some of the most important discoveries of twentieth-century materials science.

Figure 11.4 shows how a dislocation (an "edge" dislocation in this figure) can move across a crystal with minimal distortion of the crystal lattice and yet accommodate a net shear displacement. This mode of deformation is common to all crystalline substances, so all can deform plastically and, thus, creep. There is only a minimal threshold stress, so no material has finite strength at high temperatures or over long time periods. Figure 11.4 shows how a dislocation can "glide" through a crystal lattice. In any real material many dislocations are present at the same time and, by gliding across one another, they create kinks in one another that effectively pin the dislocations at the crossing locations. When this occurs, glide ceases after a few percent of strain. A new step is required to free the dislocations from their pinning points. That step requires the bulk motion of atoms through the lattice – diffusion. The process of dislocation glide coupled with diffusive untangling of dislocations is known as dislocation climb. Because diffusion is a thermally activated process, so is the rate of creep. The flow of crystalline solids is, thus, strongly temperature-dependent, with an activation energy similar to that of the bulk diffusion of the solid. This explains why creep is rapid only at high temperatures.

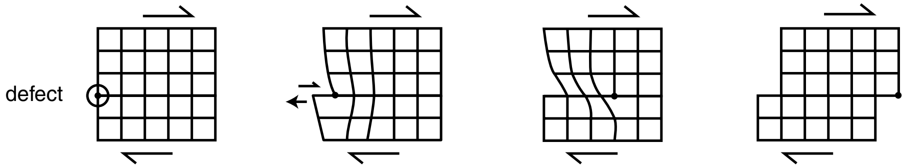

Figure 11.4 Motion of a dislocation through a crystal lattice leads to shear deformation of the crystal. Starting at the left, a line defect is created when the atomic bond of one row of atoms in the upper half of the crystal is shifted one lattice spacing to the right, creating a line of local disorganization known as a dislocation. The atoms in the crystal shift partners as the dislocation moves to the right. After the dislocation finally emerges from the crystal on the right, the upper half of the crystal is shifted one lattice spacing to the right, accommodating an increment of shear strain. This is the mechanism by which ice crystals deform in a creeping glacier.

11.2.1 Glen's law

J. W. Glen (1955) was the first to make careful measurements of the relation between stress and strain in polycrystalline ice and to apply it to the flow of glaciers. Unlike the flow of viscous fluids, he found that the strain rate is not a linear function of stress, but depends upon stress raised to a power n greater than 1. As described in Section 3.4.3, this is the general behavior of creeping hot solids, materials whose flow is dominated by dislocation motion. Glen expressed his rheological law in terms of the strain rate:

$$\dot{\varepsilon} = A\sigma^n = B\,\mathrm{e}^{-Q/RT}\,\sigma^n \qquad (11.1)$$

where Q is the creep activation energy, R is the gas constant, T the temperature in K and σ is the applied shear stress. Glen found that the power n ranged between 2 and 4, with a preferred value of 3.2 (the modern value of n is 4). His estimate of the activation energy was $Q = 134$ kJ/mol, less than the modern value of 181 kJ/mol listed in Table 3.3. His experiments gave the constant $B = 3.5 \times 10^{20}$ MPa$^{-3.2}$s^{-1}. Nevertheless, Glen clarified the important differences between the flow of Newtonian viscous liquids and glacier ice (Forbes was not *completely* right) and showed that very cold ice should flow less readily than ice near its melting point. He also realized that the creep rate changes as the ice recrystallizes during flow, a factor that is still not fully incorporated into modern creep laws.

The implication of a power-law dependence of strain rate on stress is that as stress increases the strain rate increases faster than a direct proportion. Thus, doubling the stress for $n = 4$ means that the strain rate increases by a factor of 16. Most of the strain, thus, becomes concentrated in high-stress areas; at the bottom of an ice sheet on an inclined surface, for example. Non-linear rheological laws such as Equation (11.1) generally resist easy analytic solutions and require numerical methods to get quantitative results. Thus, the creep law for an infinite sheet of power-law fluid flowing down a surface of constant slope can be readily obtained, but it is quite remarkable that there is also an analytic solution for the much less straightforward case of a power-law fluid flowing down a constant slope in a semicircular channel of constant width (Nye, 1952). We give this more interesting result here, leaving the much simpler case of an infinite plane sheet as an exercise for the interested reader.

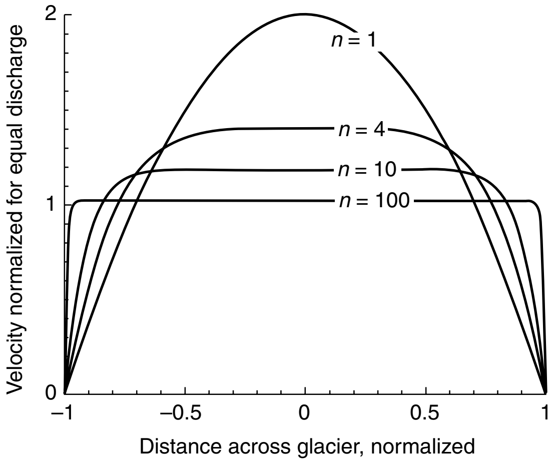

Figure 11.5 Flow profiles across different power-law fluids flowing down a trough-shaped channel with a semicircular cross section. Newtonian fluids with $n = 1$ attain a parabolic profile, while fluids with increasingly high n attain more plug-like profiles, with uniform velocity in the center and steep gradients near their walls.

Let r be the radial distance from the centerline of a power-law fluid flowing through a trough-like semicircular channel of radius R that slopes downhill at angle α. For this case, the downstream velocity $u_z(r)$ is given by:

$$u_z(r) = \frac{2A}{n+1}(\rho g \sin \alpha)^n \left[\left(\frac{R}{2} \right)^{n+1} - \left(\frac{r}{2} \right)^{n+1} \right]. \tag{11.2}$$

For flow in a parabolic or elliptical channel the coefficient of the right-hand side changes slightly, and the scales for vertical and horizontal velocity gradients are different, but the overall behavior is similar. Naturally, this equation also applies to Newtonian flow when $n = 1$.

Plots of this equation in Figure 11.5 show that as n increases the flow becomes more plug-like. Comparison of this equation to the profiles of actual glacier velocity profiles in a straight reach of Saskatchewan Glacier shows good agreement with $n = 3$ and are inconsistent with a Newtonian, $n = 1$ curve (Meier, 1960). Deformation of vertical boreholes through a glacier can also be compared against theoretical velocity profiles such as Equation (11.2). Measurements of this kind show good agreement between the non-Newtonian flow theory and observation (Paterson, 1999).

For more complicated channel geometries with varying cross sections, bends and obstructions, numerical solutions to the flow equations must be constructed. A great deal of progress has been made in the application of finite element methods to prediction of glacier velocity patterns.

The actual rheology of ice is more complicated than Glen's law alone would suggest. Recent investigations indicate that a variety of mechanisms in addition to dislocation climb are important for the deformation of ice, especially at low stress. Such processes as intra-grain diffusion, diffusion through grain boundaries, and grain boundary sliding all contribute to the deformation of ice under various conditions of stress and temperature. The interested reader will find an excellent summary of the complex rheology of ice with planetary applications in the review by Durham and Stern (2001).

11.2.2 The plastic-flow approximation

Figure 11.5 shows that as the exponent n increases, the corresponding flows become more plug-like. For very large n the result would approximate the flow of a perfectly plastic material, one that remains rigid until a yield stress Y is exceeded, after which it flows to an extent determined only by external constraints on the displacement. This observation led to the idea that a power-law fluid with large n can be approximately represented as a perfectly plastic material. Application of this idea to ice suggests adoption of an effective yield stress of about 0.1 MPa for the "strength" of ice.

The plastic model of ice rheology gives a fairly good prediction for the height profile of an ice sheet as well as the cross section of some glacier tongues, as discussed in Section 5.3.2 for lava flows and below for ice sheets. However, there is a serious internal problem with this model and it should be used only with some caution and understanding of this issue. This problem can be seen from dimensional considerations. The fundamental rheological law (11.1) for ice relates a strain rate to a stress, so the coefficient relating the two has the dimensions of inverse stress (to the nth power) and inverse time. However, the plastic law contains no dimension of time, only stress. These two relations, thus, do not transition smoothly over into one another as n becomes large unless some quantity with dimension of time is present. That is, one must quote a timescale τ in addition to a yield stress for this relation to be meaningful. To see this explicitly, rewrite (11.1) as:

$$\dot{\varepsilon} = \frac{1}{\tau}\left(\frac{\sigma}{Y}\right)^n . \tag{11.3}$$

This rewrite is always possible once Y is given. In that case τ is defined implicitly by (11.1). In the limit $n \to \infty$ this equation approaches the plastic yield condition:

$$\begin{aligned} &\dot{\varepsilon} \to 0 \quad \sigma < Y \\ &\dot{\varepsilon} \to \infty \quad \sigma > Y. \end{aligned} \tag{11.4}$$

Equation (11.4) is commonly cited in connection with the plastic approximation (e.g. Paterson, 1999), but without a timescale, τ, Equation (11.3) is not dimensionally correct. The creep rate according to (11.1) is never really zero at low stress – it is only "almost as good as zero" over some timescale that the user considers important.

The widespread neglect of this timescale in the glaciological literature is probably why there is quite a bit of variation in estimates quoted for the yield stress, although it is usually given as "about" 0.1 MPa in round numbers. One can make this approximation precise for Glen's flow law if $\tau = 1.4$ yr is the relevant timescale over which deformation is considered important. This is also the time required for a stress equal to the yield stress to produce a strain of 100%.

11.2.3 Other ices, other rheologies

Deformation by dislocation climb and other mechanisms is not confined to water ice. Ices of other substances, such as solid CO_2, ammonia, and nitrogen, are all of interest for the formation of glaciers on other Solar System bodies. Unfortunately, the rheology of such ices is not as well explored as that of water ice. Table 3.5 lists recent data on the flow law of solid CO_2, but the flow rates of other ices can only be conjectured at the present time. It is, thus, important that theoretical models of steady-state creep exist and, in many cases, are successful in predicting the rheological behavior of complex substances. The detailed examination of creep mechanisms is a large area of research that is beyond the scope of this book. For more information, the reader is referred to the recent monograph by Karato (2008). However, a few simple scaling arguments can be stated that allow crude estimates of the relative creep rates of different substances.

Nearly all rheological creep laws depend upon thermally activated diffusion to permit slow deformation of crystalline material. This explains the observed exponential temperature dependence of relations like Glen's law (11.1). The actual temperature dependence is a function of the binding energy of atoms in the material, the species diffusing, and whether the diffusion occurs along grain boundaries or through the body of the crystal. The dependence of strain rate on stress is a function of whether the deformation is by dislocation climb or by pure diffusion. Dependence of the creep rate on grain size is also a function of the mechanism: Dislocation climb rates do not depend on grain size, whereas diffusion creep processes are functions of grain size because it determines the length of the diffusion path.

The individual equations for different creep mechanisms are, thus, functions of many different variables, not all of which may be known for a new substance. Nevertheless, the overwhelming influence of diffusion makes it possible to create order of magnitude estimates of creep rates. We can crudely write the strain rate at a given stress as:

$$\dot{\varepsilon} = (\text{a bunch of complicated stuff})\, D_0 e^{-Q/RT}\, \sigma^n. \tag{11.5}$$

The complicated term in the parenthesis is a function of the many variables we have mentioned (and others that we have not). The second and third terms, combined, constitute the temperature-dependent diffusion coefficient, while the last term is the shear stress, raised to some exponent ranging between 1 (for pure diffusion creep) and about 5 (for generic dislocation climb).

We can now form the ratio between the creep rate of an unknown substance and that of a known material (water ice, for example) and compare the two. It should be clear that the relative creep rate depends mainly on the relative diffusion coefficients:

$$\frac{\dot{\varepsilon}^{\text{unknown}}}{\dot{\varepsilon}^{\text{ice}}} = (\text{number of order } 1)\frac{D_0^{\text{unknown}}}{D_0^{\text{ice}}} e^{(Q_{\text{ice}} - Q_{\text{unknown}})/RT}. \tag{11.6}$$

This allows us to evaluate the relative effectiveness of different substances as candidate glacier materials if we can make some guess about their diffusion coefficients relative to ice. We may, if we are desperate and in complete ignorance, take one more step into the world of wild and wooly approximations (a world in which planetary scientists must dwell all too frequently) and invoke *Shewmon's rule of thumb*, which states that the diffusion coefficient of *any* substance at its melting point is 10^{-12} m^2/s (Shewmon, 1963). This leads one to expect that, at their melting points, all substances should flow roughly like glacier ice. At lower temperatures, flow rates are slower and depend upon the activation energy for whatever form of diffusion is most effective in permitting the crystal to deform.

Interactions of ices with other substances, especially interstitial fluid, may greatly enhance the rate of creep. Box 11.1 describes the strange case of salt glaciers on Earth, in which the presence of small amounts of intergranular water greatly enhances the creep rate and permits salt to form kilometer-long flows that look superficially like glaciers. Similar enhancements might occur in water ice at low temperatures when small quantities of ammonia are present, as occurs in the outer Solar System.

11.2.4 Basal sliding

The sliding of a glacier over its bed is a quintessentially water-ice process. Warm-based glaciers with melt at their beds are a consequence of water's nearly unique negative-slope melting curve. The process of regelation that makes glacial erosion so effective in warm-based glaciers also depends on this peculiarity of water's melting curve.

When it can occur, basal sliding is an important contributor to the overall motion of a glacier or ice sheet. Many estimates suggest that approximately half of the surface velocity of a warm-based glacier is due to basal sliding. The warm-based Antarctic ice streams similarly depend on basal slip for their high velocities. In these streams the basal ice interacts with water-saturated deformable sediments, not rock. The mechanics of this soft-sediment interaction is complex and not presently well understood. The mechanical behavior of water-saturated sediments themselves is complicated and their confinement beneath a moving ice sheet introduces complex feedbacks that are the subject of current research.

Regelation. Basal sliding over a rigid bed is better understood. When moving ice at the base of a glacier encounters an obstacle on the bed, such as a rock protuberance or wedged boulder, the pressure on the upstream side of the obstacle increases, while that on the downstream side decreases. Because of the negative slope of water ice's melting curve, this lowers the melting point on the upstream side. A small amount of ice melts and the local temperature declines slightly to the pressure melting point. Upstream ice thus melts, but not instantaneously: The rate of melting is regulated by the rate at which the latent heat of melting, 334 kJ/kg, is supplied to the compressed ice by conduction from the adjacent ice and rock. However, as heat is conducted to the upstream ice, the meltwater flows around the obstacle and freezes behind it at a slightly higher than ambient

Box 11.1 **Salt glaciers and solution creep**

Salt on the surface of the Earth would seem to be one of the least likely materials to flow as a glacier. Measurements of the creep of pure halite (NaCl) show that, although it does creep more readily than most rocks, it still requires temperatures in the vicinity of 550 °C for it to creep at a rate comparable to that of glacier ice.

Unlikely as it might seem, glaciers of salt several kilometers long were described from the dry Zagros Mountains of Iran in 1929. The discoverers did not believe that salt at normal surface temperatures could flow at rates comparable to glacier ice and supposed instead that the salt had erupted hot, at temperatures near 300 °C, and that the glaciers are not moving at present.

However, salt is highly soluble in water and a small amount of rain does fall in this region. The theoretical possibility that small amounts of water could enhance the flow rate of salt by the mechanism of pressure solution creep was investigated by Wenkert (1979). Pressure solution creep occurs when a crystal subject to differential stress preferentially dissolves on faces under compression and is deposited on faces under extensional stresses. The dissolved crystal material diffuses much more readily through the solvent than through the body of the crystal. The shear strain rate is given by:

$$\dot{\varepsilon} = 21\frac{V_0 C_L D_L f}{kTd^2}\sigma \tag{B11.1.1}$$

where V_0 is the volume of the diffusing species, C_L its molar solubility in the solvent, D_L is its diffusivity in the solvent, k is the Boltzmann constant, T the absolute temperature, d the grain size in the solid, and f is the fraction of liquid wetting the solid.

Applied to salt glaciers, this equation predicts a strain rate 10^8 faster than that expected for pure halite. This prediction was verified by both direct observations of the creep of salt glaciers following rare rainfall events and by laboratory measurements of damp halite (Urai *et al.*, 1986).

Study of the creep of salt has attracted a large amount of attention because of its importance for proposed nuclear waste storage in salt deposits. Aside from this practical application, solution creep is expected to greatly enhance the flow of limestone in the Earth. It has also been proposed as an agent in enhancing the creep of cold water ice in the outer Solar System through the solution of ice in interstitial ammonia.

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temperature, there releasing its latent heat, which is now available to be conducted to the upstream face.

This process of melting under compression, followed by meltwater flow and freezing in the adjacent low-pressure zone, all regulated by the conduction of heat from the freezing water to the melting zone, is called regelation. Regelation is easily demonstrated in the kitchen (or classroom) by hanging a wire loop weighted at both ends over an ice cube supported at its ends like a beam. The wire very quickly slices through the ice cube and emerges on the other side, leaving the ice cube apparently intact (actually, it is not quite intact – examination under polarized light shows that the ice along the path of the wire has recrystallized).

Regelation at the base of a glacier is very efficient for small obstacles, through which heat is rapidly conducted, but inefficient for large obstacles. On the other hand, the glacier can easily deform around long-wavelength obstructions, but deformation is difficult for small wavelengths, requiring high strain rates. There is, thus, some intermediate wavelength that is maximally obstructive – the expectation is that this wavelength accounts for most of the resistance to basal sliding. Estimates of the size of this most obstructive obstacle indicate that it is about 10 cm.

Glacier Surges. Most glaciers move down their valleys at a sedate speed of a few meters per day. However, a few glaciers are observed to suddenly accelerate to many tens of meters per day in rapid advances known as glacier surges. A surging glacier rapidly lengthens and thins, overrunning forests and roads in its path. Its surface breaks up into a wilderness of crevasses separating large blocks that topple as the glacier moves, making it nearly impossible to cross or even remain safely in one spot on the ice for more than short periods of time. Because they are so difficult and dangerous to study, little was known about the mechanics of surging glaciers until a heroic effort with massive helicopter support was mounted during the 1982–1983 surge of the Variegated Glacier in Alaska (Kamb *et al.*, 1985).

It was discovered that the immediate cause of the Variegated Glacier's acceleration was a large increase in the water pressure at its base, which occurred in conjunction with a rearrangement of the system of subglacial cavities and tunnels that drain the glacier. This increase of water pressure lifted much of the glacier's weight off its bed, decreasing basal friction and greatly enhancing the rate of basal sliding.

Surges are evidently restricted to warm-based glaciers and it may be that, given enough time, all warm-based glaciers will exhibit surge activity. An interesting question is whether the warm-based Antarctic ice sheet is also subject to surges and, if it is, what conditions must be met to cause a surge.

11.3 Glacier morphology

Glaciers are tongue-like masses of ice that flow down valleys, whereas ice sheets are broad plains of ice that spread centripetally from their high centers. Valley glaciers typically carry masses of rock debris along their margins, material that has avalanched from the valley sides onto their surfaces. Where ice streams meet, these lateral moraines merge into long trains of debris within the body of the compound glacier and are known as medial moraines.

The terminus of a glacier or ice sheet may remain at the same location for a long period of time, but this does not mean that the ice is not moving. Instead, the ice is continually pushing forward while it melts back at the same rate, making the terminus a dynamic location that is constantly subject to small oscillations as the balance between flow and melting shifts slightly. Because new ice is constantly arriving at the terminus, debris frozen into the ice melts out and gradually builds up into what may become a large heap – the terminal moraine. Even when the ice front retreats because melting predominates over flow, the internal movement of the ice is still downward: The flow velocity never reverses.

11.3.1 Flow velocities in glaciers and ice sheets

A widespread misconception supposes that the ice in a glacier is squeezed out by the weight of the overlying ice, somewhat like toothpaste from a tube that has been accidentally stepped upon. Called "extrusion flow," this idea is imbedded in many older texts on glacier flow. Unfortunately, it is not supported by observation: Intensive studies of glacier deformation in boreholes, starting during the International Geophysical Year in 1957–1958, have uniformly shown that the maximum velocity in a glacier occurs at its surface. Because of friction on the walls and bed of a glacier, the velocities near the contact between ice and rock are lower than elsewhere. Velocity contours on a transverse section across a glacier are concentric arcs around the maximum, which is on the surface. In bends of the ice stream the position of the maximum shifts from the centerline toward the outside of the bend.

The former beds of vanished glaciers sometimes slope uphill in what was obviously the downglacier direction and observers wonder how a glacier could have been flowing uphill. However, careful consideration of the equilibrium of a block of ice with surface slope α_s and basal slope α_b show that the shear stress on the base of a glacier of thickness H is given by $\rho g H \sin \alpha_s$. That is, the basal shear stress depends only on the surface slope of the glacier ice and is independent of the basal slope – even of its sign. This concept is not as paradoxical as it might seem: The beds of many rivers also slope uphill in reaches between deep pools and shallow riffles, but as long as the surface slope is downhill, the water flows in the expected direction. Thus, so long as the surface of a glacier slopes downhill, the shear forces driving it along continue to urge it downhill, even though its bed might have the opposite slope.

The vertical component of the ice velocity varies systematically along the course of a glacier. In the accumulation area the vertical velocity is downward. A marker placed on the surface of the ice is gradually buried by snow, sinking into the glacier as the snow metamorphoses into new ice and more snow accumulates on top of it. Once our marker (whether it be a meteorite that has fallen onto the snow or the body of some unfortunate early mountaineer) moves into the ablation region, melting ice gives the vertical velocity an upward component and markers once frozen into the ice emerge onto the surface.

The longitudinal velocity of an ice stream does not follow any simple rule, responding instead to variations in the underlying topography and the local thickness of the ice. Where topographic steps occur the ice may flow particularly fast and the surface becomes very steep in reaches known as icefalls. Extensive crevasse systems as well as locally high velocities characterize icefalls.

The downstream velocity of a wide ice sheet of thickness H and surface slope α_s is given by the expression (derived from Glen's law):

$$u_z(y) = u_{\mathrm{bs}} + \frac{2A(\rho g \sin \alpha_s)^n}{n+1}\left[H^{n+1} - (H-y)^{n+1}\right] \quad (11.7)$$

where y is the height above the glacier bed, ρ the density of the ice and u_{bs} is the velocity of basal sliding.

As shown in Figure 11.5, Equation (11.7) predicts that as the creep power law n becomes larger, the flow is more concentrated toward the base of the glacier where the shear stress is higher. In the limit of very large n the flow approximates that of a perfectly plastic material and all of the deformation occurs at the bed.

11.3.2 Longitudinal flow regime and crevasses

As the longitudinal velocity of an ice stream varies along the glacier due to variations in ice thickness and bed slope, the overall strain rate at the surface of the glacier alternates from compressional to extensional. These strain-rate changes are accompanied by corresponding changes in the longitudinal stress. Stress variations might remain unknown to a visual observer, except for the fact that extensional stresses open crevasses that are readily seen in images. The crevasse pattern on the surface of a glacier thus contains clues about the stress state and flow regime of the ice.

Crevasses are gaping open fissures that cut the brittle upper surface of an ice sheet. Their depths are limited (unless they become filled with water) because increasing overburden pressure eventually overcomes the tensile stress and squeezes the crack closed. A crude means of estimating the maximum depth of crevasses derives from the plastic approximation to power-law flow. If the maximum stress in a glacier is limited by a plastic yield limit of 0.1 MPa, then the maximum depth of a crack is reached when the plastic yield stress Y equals the overburden pressure, divided by 2 (this factor of 2 comes from the relation between shear stress and the unidirectional stress exerted by the overburden). Thus, the maximum crevasse depth is roughly:

$$h_{\mathrm{crevasse}} \approx \frac{2Y}{\rho g}. \quad (11.8)$$

Substituting numerical values for ice on Earth, this comes out to about 20 m. Measured crevasse depths in glaciers seldom exceed 25 to 30 m, so this result is the right order of magnitude. A more sophisticated approach incorporating the full rheologic equation is given in Paterson (1999).

Glacier reaches where the flow rate accelerates are in extension, while those in which the flow decelerates are under compression. Crevasses occur on the surface where the stress is tensional, that is, where the flow is accelerating. They are rare where the flow is decelerating. Near the snout of a glacier the flow is typically decelerating as the ice is thinned by

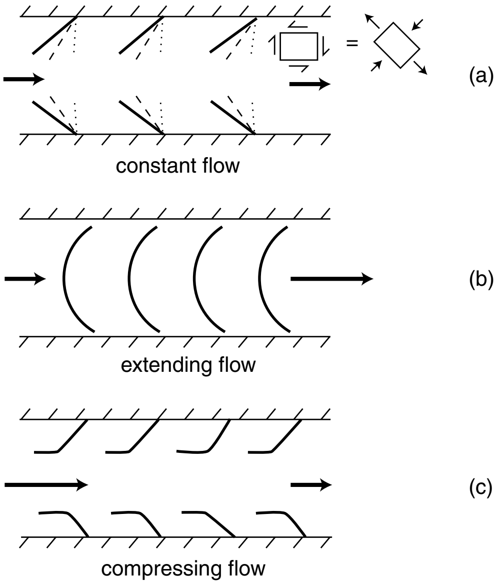

Figure 11.6 The pattern of crevasses on the margin of a glacier indicates the state of longitudinal strain. (a) A glacier flowing in a uniform channel with uniform longitudinal velocity. Friction against the side walls creates shear stress that tends to open crevasses at 45° to the flow in the direction of greatest extension (the inset shows how shear stress can be resolved into extensional and compressional principal stresses). With time, these crevasses rotate down the glacier (dashed lines). (b) Where the flow velocity increases downglacier (extending flow) the crevasses may extend all the way across the glacier nearly perpendicular to its flow direction. (c) Where the longitudinal velocity decreases downglacier (compressing flow) crevasses are suppressed in the center of the glacier stream and curve upglacier. After Figure 9.8 of Paterson (1999).

ablation, the glacier surface is under compression and the enterprising mountaineer may ascend the *Gesundheitstrasse* (German for "healthy route") onto the glacier surface.

Diagonal crevasses often form along the margins of glaciers, where friction against the wall creates shear stresses. Resolving the shear into its diagonal components of compression and extension, one expects the crevasses to form at a 45° angle to the wall of the glacier, with the acute angle facing upstream. However, with time these 45°crevasses rotate to become more transverse to the trend of the glacier because of the differential flow of the ice. Figure 11.6 illustrates the typical crevasse patterns expected for glacier reaches where the flow is (a) neither compressing nor extending, but wall friction is important, (b) extending, and (c) compressing. Moreover, if the glacier spreads out over a broad region, as in a piedmont glacier, crevasses often form perpendicular to the direction of lateral expansion.

11.3.3 Ice-sheet elevation profile

The elevation profile of ice sheets is often well approximated by a parabola, or at least a simple curve resembling a parabola. The parabolic form is a direct consequence of the

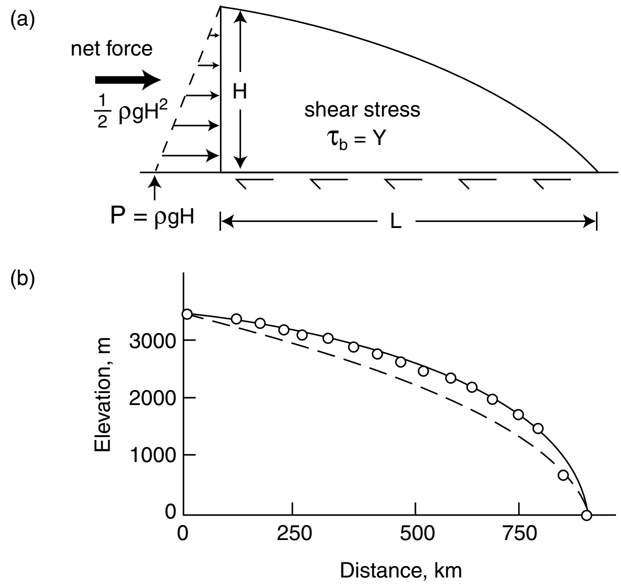

Figure 11.7 The topographic profile of ice sheets can be approximately computed from a plastic-flow model. (a) Illustrates the balance of forces on a mass of ice in which the pressure of the ice mass to the left is balanced against the shear stress at its base, resulting in a parabolic relation between the ice thickness H and the distance to the ice margin L. (b) Comparison between the plastic-flow model (dashed line) and the profile of the East Antarctic Ice Sheet between Vostok and Mirny (circles). A better fit, indicated by the solid line, incorporates a uniform accumulation of ice. After Figure 11.4 of Paterson (1999).

plastic approximation to power-law flow. This same model was used in Section 5.3.2 to argue that the profile of a lava flow consisting of a Bingham fluid should be close to a parabola. As shown in Figure 11.7a, the argument proceeds by balancing the total force from the base of the sheet, $Y L$, against the pressure driving it outwards, $\frac{1}{2} \rho g H^2$. The resulting equation for the thickness of the sheet H as a function of distance from the edge L along a line running radially outward from the center of the sheet is:

$$H = \sqrt{\frac{2YL}{\rho g}}. \tag{11.9}$$

For example, for a point 750 km from the edge of the ice sheet and a yield stress of 0.1 MPa on the Earth, Equation (11.9) predicts an ice elevation of about 3900 m, not far from the observed elevations of the East Antarctic Ice Sheet.

The plastic-flow model is only an approximation and better fits can be attained using the full flow law, coupled with the recognition that temperatures in the upper part of the Antarctic ice sheet are below 0°C and so the ice there is less fluid. Figure 11.7b shows the elevations along a profile from Vostok to Mirny Stations along with a parabola (dashed) and a more exact model (solid line). The parabolic model gives a good first approximation, but the more accurate treatment improves the fit considerably. The simple model also does not take changes in mass balance into account, which is important for most ice sheets.

Because of concerns over the effects of global climate change, modeling of the Antarctic and Greenland icecaps is reaching a high degree of sophistication and incorporates details well beyond the scope of this book. The interested reader is referred to the recent monograph by Greve and Blatter (2009).

11.4 Glacial landforms

Glaciated landscapes betray themselves to the knowledgeable viewer by a variety of characteristic features. Glaciers and ice sheets are effective agents of both erosion and deposition. The present landscapes in high-latitude regions of the Earth bear many scars of the recent series of ice ages. Earth has experienced other episodes of glaciation even farther back in time: during the Permian Era and in the Neoproterozoic. We are aware of these ancient episodes by the changes they produced in rocks exposed at that time, although these are certainly not as apparent as the changes dating from 12 000 yr ago.

Features that seem to indicate glacial erosion also occur on Mars. These appear to be much more ancient than glacial features on the Earth, but thanks to the very slow rate of surface modification on Mars they remain to betray their origin. Acceptance of widespread ice sheets on Mars has been slow, but very recent discoveries of relict ice masses near the equator of Mars point clearly to a former era of extensive ice. In addition, eskers and many other landforms are consistent with a previous age of ice on Mars.

11.4.1 Glacial erosion

Valley glaciers modify the stream valleys they initially followed, grinding the original V-shaped cross profiles into a U-shaped trough. Glacial erosion truncates spurs, creates bowl-like cirques at the head of canyons, and leaves tributary valleys hanging high above their normal level of junction. The longitudinal form of glacial valleys is converted into a giant staircase of treads and risers. The treads frequently slope against the general trend of the valley and, after the glaciers have melted away, trap small lakes called tarn lakes.

The process of glacial erosion proceeds largely by the removal of large blocks from the bed of the glacier, a process called quarrying or plucking. As the glacier slides over irregularities in its bed, it may move fast enough that the ice cannot close in behind the obstacles, leaving open cavities in the lee that fill with meltwater. However, the pressure in the meltwater is far below the overburden weight of the glacier, except perhaps during glacier surges. The pressure gradient between the upstream side of the block on which the ice is impinging and the downstream water pocket is often enough to slide the block out of the bed and incorporate it into the ice. The blocks may be directly fractured from intact bedrock by this pressure differential or may be pre-existing joint blocks. In either case, once the ice mobilizes these blocks they are incorporated into the basal ice and dragged downglacier along the bed, contributing to further erosion by abrading the bedrock still in place.

Abrasion occurs between rock debris already incorporated into the ice and the rock bed of the glacier. Its importance can be judged from recently deglaciated surfaces, which are typically striated, smoothed and, in places, even polished by the action of debris moving along with the ice. Abrasion of solids is a reasonably well-understood process, at least in its dependence upon the force and velocity of the grinding surfaces, so it was a surprise when Geoffrey S. Boulton in 1974 discovered an unanticipated aspect of glacial abrasion that goes far toward explaining some of the details of glacial bedrock erosion.

Boulton (1979) studied glacial erosion by inserting plates of rock directly on the beds of several glaciers beneath which tunnels had been excavated to collect glacial meltwater for water supply. He found, as one might expect, that the thicker the glacier, the harder any rocks frozen into the glacier ice bore down on the bed and the more material was abraded from the bed. However, when the overburden pressure of the ice exceeded about 2 MPa, the increase in the normal pressure of rock on rock ceased, because the embedded rocks simply punched back into the ice instead of transmitting more pressure. When the ice overburden exceeded 3 MPa, the rocks in the glacier simply stalled against the bed and the ice flowed around them, bringing erosion to a halt. At greater pressures the basal debris was deposited beneath the glacier as till.

There is, thus, a limit to how much pressure ice-entrained debris can exert: Abrasion is possible beneath ice about 300 m thick or less. This limit depends somewhat on glacier speed, with the pressure at the peak of abrasion ranging from about 1 MPa at speeds of 5 m/yr up to about 3 MPa at speeds of 100 m/yr. Nevertheless, the qualitative limit to erosion goes far toward explaining the U-shape of glacier valleys.

Consider a glacier initially flowing in a V-shaped fluvial valley. It is deepest at its center, but if its overall thickness approaches the limit of abrasion, it is relatively ineffective at eroding its deepest portion, while removing more material from the walls higher up. The shape then gradually changes from a V to a U as the rate of erosion is equalized across the valley and the glacier continues to grind deeper into the bedrock (Harbor *et al.*, 1988).

The centers of continental ice sheets easily exceed Boulton's abrasion limit, so that most of their work in plucking and grinding their beds is done within a few hundred kilometers of their margins, where the thickness of the ice is relatively low. This prediction accords well with the observation that the continental ice sheets eroded most deeply near their edges, for example excavating the Great Lakes and Finger Lakes beneath the edges of the Laurentide Ice Sheet of North America. Streamlined, ice-shaped hills, such as *roche moutonnée* are best developed near the former margins of the great ice sheets.

11.4.2 Glacial deposition

Moraines. Glaciers are unselective agents of transport. They can and do carry everything from multi-meter blocks of rock to the finest silt. Valley glaciers transport any debris that happens to fall on their surfaces, an occurrence that is common because of rock avalanches from their over-steepened walls. Moving ice of all types picks up material from its bed and carries it along as it moves. When this material in transit reaches the terminus where melting exceeds the rate of ice motion, this debris is dumped in an unsorted heap. This material is called glacial till and the landform it creates is called a moraine.

Besides containing a miscellaneous collection of boulders, pebbles, and silt, glacial till is also compounded of rock flour, an unusual type of sediment unlike that produced by other processes. Rock flour is finely pulverized but otherwise fresh bedrock. Chemical weathering of rock flour is minimal because it is produced by grinding of rock upon rock at low temperatures beneath a glacier. It is composed of grains mostly less than 100 μm in diameter that are easily suspended in meltwater streams. The bluish, milky color of glacial streams and lakes is due to heavy loads of this material. When deposited in front of a glacier it is easily picked up by the wind and blown in dense clouds that make the terminus of a glacier a dirty, gritty place to work. During the ice ages the entire atmosphere of the Earth was laden with dust from rock flour. It was laid down in thick deposits known as loess in extensive plains in China and the midwestern United States that are today valued for their agricultural potential.

Moraines also contain large amounts of sand-sized material that may be mobilized by the high winds that often accompany glacial climates. During the Earth's recent ice age a great sand sheet formed the Sand Hills of Nebraska, created by sand washed out of glacial meltwater streams. The high winds, lack of vegetation, and abundant sand-sized sediment in the Polar Regions led to the surprising development of dune fields in this environment. Given the evidence for former ice sheets on Mars it seems possible that some of the sand-sized material there has a glacial origin.

Glaciologists distinguish several types of moraine, depending on where they form. Terminal moraines pile up at the ends of glaciers, becoming large during times when the ice margin remains at a nearly fixed location. Lateral moraines form at the edge of ice streams. Ground moraines form when the ice retreats rapidly, leaving a thin, loose deposit on the surface. Lodgement moraines form beneath the glacier and their till is often strongly compacted by the weight of the glacier.

While *roche moutonnée* are streamlined hills eroded from underlying bedrock that range in size from kilometers to a few meters, drumlins are streamlined depositional forms molded out of till, whose size range is similar. From morphology alone it is difficult to separate these two features: Indeed, some *roche moutonnée* have downglacier tails of streamlined till and so are hybrid forms. Flowing ice may also produce very elongated hills that grade into fluted surfaces with alternating hills and troughs aligned in the direction of ice flow.

Most glacially deposited material is reworked to some extent by water, for melting and runoff are ubiquitous in the vicinity of glacier ice. A large variety of names have been applied to these deposits depending on the special circumstances of their formation: The interested reader is referred to the more specialized discussions cited at the end of this chapter. In this chapter we refer to only some of the most important features for planetary observations.

Kettles or kettle holes are small, sometimes circular, pits that form in the wake of retreating ice. They have occasionally been mistaken for impact craters, although they almost always lack rims. They form around blocks of ice left stranded by the retreating ice front. These ice blocks are then partially or completely buried by outwash. After the ice melts, a pit remains.

Water at the base of a glacier or ice sheet flows as films along the rock–ice interface, fills pockets and cavities bridged by the moving ice, and eventually collects into streams that form a subglacial drainage network. Subglacial streams are far more dynamic than those flowing over a landscape because the ice flows to fill cavities where the pressure is low. Any tunnel drilled into the ice closes in rapidly until it meets resistance, so the water within a glacier travels in tunnels that tailor themselves to fit the volume of the flow. Increased water pressure opens the passage until it is in balance with the ice pressure, while a decreased head causes the conduit to shrink, always keeping the water pressure in balance with the pressure of the encasing ice.

Eskers. As water moves beneath the glacier it picks up silt and debris and carries it along, depositing it when the current slackens. The deposits of such englacial or subglacial streams are known as eskers. After the ice melts away, eskers stand as branching, sinuous ridges on the land surface. The material that composes eskers is clearly water-laid, with the graded bedding, bedforms, and sorting typical of fluvial deposits. The bedding planes in eskers tend to be anticlinal in cross section, rather than horizontal, as a result of collapse along their margins as the confining ice melted away. An apparently enigmatic feature of esker deposits is that these river-like ridges can travel up and over hills, apparently paying little attention to the slope of the land surface. This is partly because some eskers were draped over the topography after the ice melted away, but a more important difference between esker networks and those of open streams is that they flowed in pressurized conduits for which local slopes are less important than regional pressure (head) variations.

Possible eskers were first recognized on Viking images of Mars and with increasing image resolution in subsequent missions the esker interpretation has become increasingly secure (Figure 11.8). They have been found at both low and high latitudes in the southern hemisphere, but those in the high-elevation Southern Uplands are the most prominent (Banks *et al.*, 2009).

Eskers lead out from underneath the ice sheet and sometimes can be seen to connect with water-laid deposits that form an outwash plain in front of the ice margin. Outwash plains are effectively low deltas or alluvial cones that represent the material deposited by streams or rivers draining from the ice. They are complex deposits in their own right, with many characteristics different from those typical of long-term fluvial deposition

11.5 Ice in the ground

Water in the ground must be frozen wherever the mean surface temperature is below freezing. On the present Earth, about 35% of the land surface satisfies this condition. The mean temperature over the entire surface of Mars is below the freezing point of water. If the

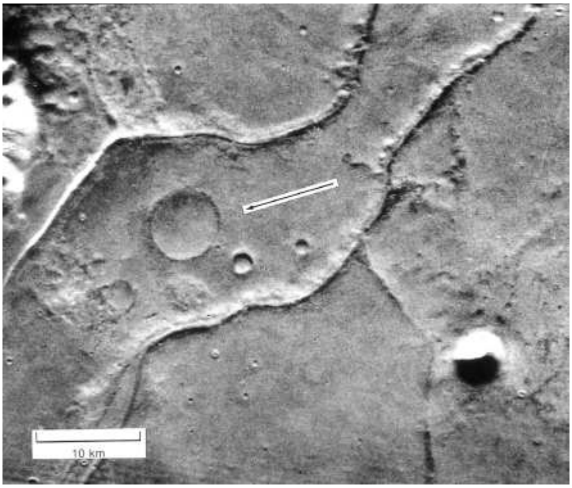

Figure 11.8 Eskers on Mars. It is suggested that these 1 km wide sinuous ridges on the floor of the Argyre Basin at 55.5° S and 40.2° W are eskers formed by subglacial streams that were deposited as the ice sheet melted away. The degraded crater in the center is 7 km in diameter. Arrow points north. Frame is 50 km wide. Viking Orbiter frame 567B33.

temperature remained permanently below freezing, water in the soil would simply stay frozen and not much of interest would occur. Temperature fluctuations, particularly those that cycle about the freezing point, are what produce most landforms and lend geomorphic interest to frozen landscapes.

11.5.1 Permafrost

The strict, but rather pedantic, definition of permafrost is that permafrost is ground that is below 0°C, *whether or not water is present*, independent of rock type. This is pedantic because if no water is present then absolutely nothing of interest happens and there are no special landforms to talk about. All of the following discussion focuses on what happens when water is present and on the effects of freezing and thawing water in the soil. Most of this discussion holds equally well for water and any other substance that undergoes a liquid–solid transformation during temperature excursions that occur on a planetary surface. Methane on Titan freezes at 91 K, just a few degrees below its average surface temperature of 94 K. However, Titan's massive atmosphere prevents temperature variations of even a few degrees from this average, so, at the moment, we do not believe that periglacial processes are relevant to Titan (unless some other common substance on its surface undergoes freeze/thaw cycling).

The nature and behavior of permafrost were not well known in the United States until the later years of World War II. At that time, a Russian-speaking geologist named Siemon W. Muller was employed by the US Army to read and translate the extensive Russian literature

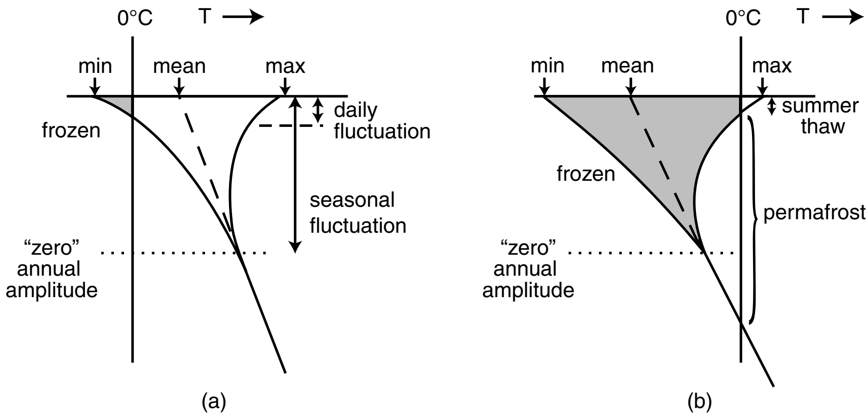

Figure 11.9 Annual temperature variations below the ground surface. (a) Indicates a temperate climate in which freezing takes place only in winter and mean temperatures are above freezing. (b) Illustrates a cold climate in which the mean temperature is below freezing. Thawing takes place only in summer. Ice is stable below the depth of the summer thaw, down to a maximum depth determined by warming due to geothermal heat flow. Permafrost encompasses the entire range over which temperatures remain below 0°C.

on the subject. After the end of the war he published a summary of his gleanings as a book (Muller, 1947) that formed the basis for our modern understanding of permafrost. Although very dated, this book can still be read with profit. Muller coined the word "permafrost" during his research.

Thermal Regime. Figure 11.9a illustrates the subsurface temperature at a location where seasonal cycles allow some freezing temperatures, but the mean temperature is above freezing, and Figure 11.9b illustrates a location where the mean annual temperature is below freezing and a permafrost zone is present. Although the top of the frozen zone is subject to seasonal temperature variations, these become negligible below the level of "zero" annual amplitude (temperature variations are never actually zero, but at this depth they are so small that they can be neglected). Below this, the bottom of the permafrost zone is determined by the planetary geothermal gradient. On Earth the geothermal gradient is about 30 K/km and the thickness of the permafrost in Siberia ranges from 200 to 400 m at 70° N down to a few tens of meters (where it is discontinuous) at 50° N. Unfrozen patches within the permafrost are known as *talik*. Talik occurs for many reasons even deep within permafrost zones; under deep lakes and rivers, for example.

Active Zone. The zone from the surface down to the level where the soil thaws annually is known as the "active zone" for reasons that will shortly be apparent. Seasonal temperatures still vary noticeably below this zone and, because not all the soil water is frozen at 0°C (or even at −30°C; see Section 7.3.2), there is still some movement of liquid water even below the permafrost table. The depth of freezing and amplitude of thermal fluctuations depend sensitively upon the nature of the surface. Surface covers of insulating material such as grass, peat, or snow have a large effect on the thermal regime of the ground underneath.

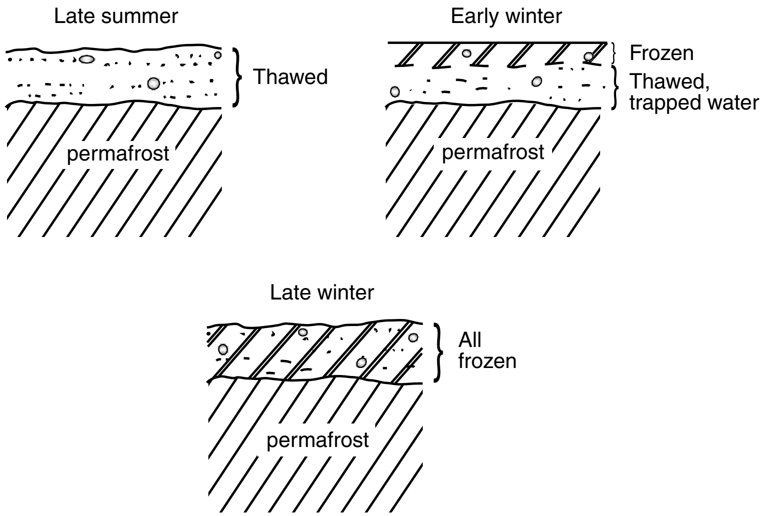

Figure 11.10 Seasonal evolution of the *active layer* overlying a permafrost terrain. In late summer this layer is completely thawed, although the permafrost below creates an impermeable layer that prevents water from draining into the subsurface and so this layer is usually saturated with water. In early winter the top of the active layer freezes, trapping water in the lower part of the active layer between the frozen water above and the permafrost below, a circumstance that promotes many kinds of instability as water pressures rise. By late winter the active layer is entirely frozen. When spring arrives the layer thaws from the top down.

Because such covers depend upon the details of surface topography and exposure, large lateral variations in thermal regime are common in permafrost areas.

Where summer temperatures rise above the freezing point of water, a seasonal cycle of freezing and thawing develops that ranks permafrost terrains among the most unstable on Earth. Permanently frozen ground is highly impermeable to liquid water: Any water that reaches the permafrost table quickly freezes and seals any cracks through which it may have originally entered. The soil overlying permafrost is, thus, commonly saturated with water when temperatures are above freezing, leading to the concept of an "active layer" (Figure 11.10). In late summer a warm thermal wave has propagated to its maximum annual depth. The soil overlying the permafrost table is thawed and often saturated with water. It forms a sea of mud that ranges from tens of centimeters deep in the far north to meters deep farther south. As winter arrives, the upper part of the active layer freezes over, trapping water in the lower portion of the layer between the impermeable permafrost table and the similarly impermeable frozen upper soil. On level ground the trapped water is stable, unless the heavy tread of a caribou or human breaks through to the mud below. However, on sloping ground the trapped water migrates laterally and high pressures can build up only a few tens of centimeters below the surface. Should the surface layer rupture for any reason, near-freezing water flows out in large volumes and quickly freezes in a low mound or sheet on the surface. Such surface layers of ice are known as "icings." Muller, in his book, delighted in showing pictures of cabins built directly on the ground in permafrost regions. Heating the cabin destroys the upper layer of ice, so in the early winter such cabins filled suddenly with icy water that spilled out the windows before freezing solid. Eventually, by late winter, all the water in the active layer freezes and this layer becomes quiescent until melting begins again in the spring.

Solifluction. Soil creep is rapid in the active layer and freeze/thaw processes lower slope angles quickly. High pore pressure in the active layer during early winter greatly enhances the probability that thin landslides develop. Creep is caused by the alternate growth and melting of ice crystals under the surface. Lenses of ice forming above the permafrost table cause intense frost heaving. When large (tens of meters broad and meters high), these ice lenses are called frost blisters: They may tilt overlying trees in the boreal forests and produce what the Russians fondly call "drunken forests." All of these processes mobilize the soil, which flows downhill in a process called solifluction or sometimes gelifluction. This soil motion often organizes into lobe-shaped steps in the surface that range from tens of centimeters to meters in height.

Pingos. Pingos are small, dome-like hills cored with ice. Internally they possess lenses of more or less pure ice beneath a layer of soil. Their mechanics of formation resembles that of igneous laccoliths, and they are sometimes called "hydrolaccoliths." They may reach heights of a few tens of meters (rarely a hundred meters) and diameters of nearly a kilometer. They often exhibit gaping radial dilation cracks at their crests from the uplift and stretching of the overlying sediment as they grew. Pingos are classified as either open-system types, in which the growing ice lens is fed by water flowing from beneath the permafrost layer, or closed-system types that develop where a former lake has frozen and fed water into the near-surface ice lens. Pingos in which the ice lens has melted resemble small volcanoes, with a central collapsed "caldera" surrounded by uplifted sediments. Pingos have been reported on Mars, but it is extremely difficult to differentiate pingos from small volcanic cones ("rootless cones" or hornitos) on morphologic characteristics alone and so these identifications are presently somewhat dubious.

Permafrost on Mars. After many years of conjecture, the presence of permafrost (in the extended sense of including frozen water) has been confirmed on Mars. Permafrost should be stable down to depths of several kilometers in Mars' polar regions, although it is not expected to be stable over the long term near the equator. Mars Global Surveyor studies of thermalized neutrons from primary cosmic rays striking the surface revealed water (more strictly, hydrogen atoms, irrespective of chemical bonding) within a few tens of centimeters of the surface in 2001. This near-surface ice extends poleward from about 50° latitude. The thrusters of the Phoenix spacecraft, which landed at 68° N, directly excavated an ice table about 5 cm below the surface. Fortuitously, five small clusters of impacts imaged by the HiRISE camera aboard the Mars Reconnaissance Orbiter also revealed rather pure ice close to the surface at five locations north of about 45° N, which includes the Viking 2 landing site. Evidently, if the Viking lander had dug just 10 cm deeper it might have uncovered water ice during its operational period from 1976–1979.

11.5.2 Patterned ground

The repeated thermal cycling of the active layer in permafrost terrain affords surface features many opportunities for self-organization. Freezing and thawing in the active layer leads to a poorly understood kind of slow convective motion that sorts fine-grained silts from rocks and organizes them into repeating patterns. Early explorers of permafrost terrains on Earth were astonished at the regular patterns of polygonal troughs, sorted stone circles, stripes, and other forms that develop with such regularity they often appear to be artificial. The size scale of these features is of the order of a few times the depth of the active layer, a few to perhaps ten meters. Larger-scale features are exceptional and require special explanations.

Ice-wedge polygons. The best understood of these features are ice-wedge polygons, thanks to the efforts of Arthur Lachenbruch, whose study of these ice wedges has become a classic of geomechanics (Lachenbruch, 1962). The implications of his study extend far beyond that of ice-wedge polygons themselves. His report should be read by anyone interested in the application of mechanical thinking to geology.

Ice-wedge polygons form networks that cover vast areas of permafrost terrain. The individual polygons may have either high centers bordered by troughs, or low centers surrounded by ridges that are split by troughs. In either case the distance across polygons is typically a few to tens of meters. The intersections of the troughs may be either close to 90°, in which cases the polygons are rectangular, or close to 120°, in which case they approach a hexagonal shape, very much like mudcracks caused by desiccation. A wedge of ice lies beneath the polygonal troughs in active terrain. These ice wedges may extend 30 m below the surface and are a few tens of centimeters to a meter wide at their tops.

Lachenbruch's explanation for how ice wedges form is illustrated in Figure 11.11. During the coldest part of the winter the active layer is frozen. Cold snaps of many days' duration often occur, at which time the frozen ground contracts and strong tensional stresses develop in a layer between the surface and the depth to which the thermal wave from the cold snap extends. If the tensile stress becomes great enough a crack opens. This crack may propagate several times deeper than the depth of the cooling from the cold snap itself, penetrating into the permafrost below the depth where seasonal temperature variations are negligible. The opening of a crack relieves tensional stresses in its vicinity out to a horizontal distance comparable to its depth. A second crack is, thus, unlikely to form close to the first: Cracks tend to be evenly spaced at distances comparable to their depths.

When the cold snap ends, the surface ice expands again, but the crack never quite closes: Dirt and small stones prop it slightly open for the rest of the winter. Upon arrival of the spring thaw, water from the active layer trickles down into the crack and freezes there, forming a thin sheet of ice along the crack surface. During the next winter the cycle repeats, but the crack is now a weak zone: Pure ice is weaker than frozen soil, so cracking during subsequent cold snaps occurs preferentially along the first-formed crack.

The next summer a second layer of water flows into the crack and freezes. After hundreds or thousands of seasonal cycles the thin crack grows into a massive wedge of ice. The

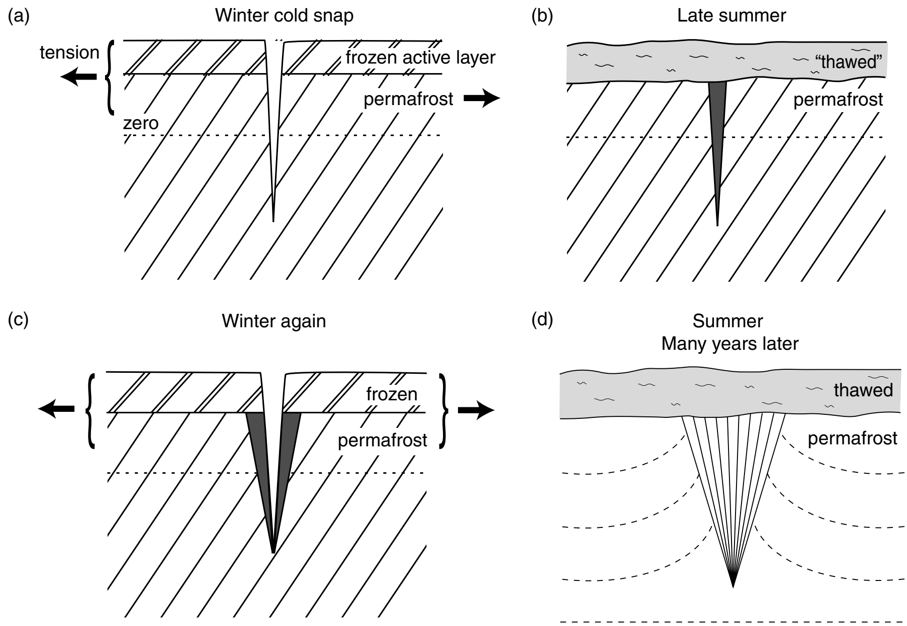

Figure 11.11 Formation of ice wedges, according to the theory of Art Lachenbruch (1962). (a) During late winter a cold snap causes the ground to contract, creating enough tensional stress to open a vertical crack that propagates some distance into the permafrost. The dashed "zero" line indicates the depth below which annual temperature variations are negligible. (b) During the late summer thaw, water percolates into the open crack and fills it, freezing at the subzero temperatures in the permafrost. (c) The next winter another cold snap re-opens the same crack because water ice is weaker than the surrounding permafrost. (d) After many such cycles of crack-opening and water-filling a broad wedge of ice has grown in the original crack, slowly enlarging by forcing adjacent sedimentary layers to deform as it grows.

soil stretches and thins over the opening wedge, which also presents a mechanical and thermal contrast to the rest of the permafrost, being composed of nearly pure ice. Soil adjacent to the growing wedge is slowly pushed aside by the wedge and heaped up into a flanking ridge or thrust farther into the center of the polygon. Such soil deformation is frequently noted in exposed sections of ice wedges.

If the climate changes and the permafrost warms and melts, melting the ice wedges with it, relicts of the ice wedges still remain. As the ice wedges melt away, soil from the active layer flows into the vacated wedge and may be recognized long after by the interruption of the original stratigraphy of the permafrost, distorted layers, and textural differences in the ice-wedge filling. Such fossil ice wedges are frequently discovered in former permafrost terrains and serve to indicate the extent of cold conditions during the Earth's recent ice age era.

The size scale of ice-wedge polygons reflects the depth of penetration of the thermal wave from cold snaps (times a factor of a few to account for the deeper penetration

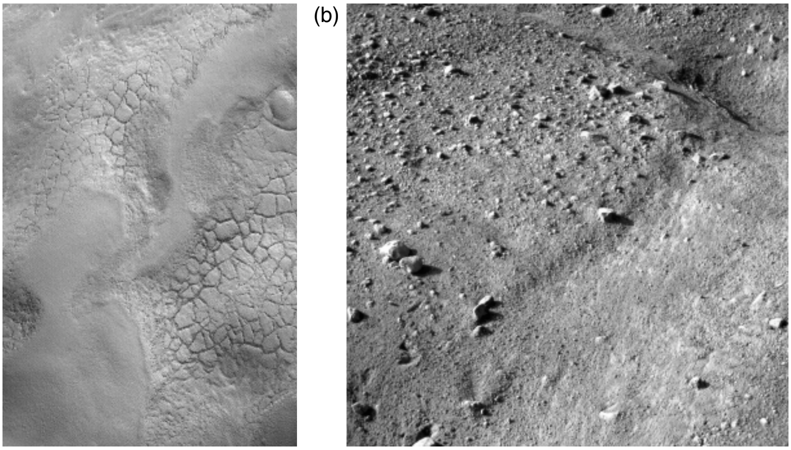

Figure 11.12 Ice-wedge polygons on Mars. (a) Patterned ground seen on the ejecta from Lyot Crater at 54.6° N and 326.6° W. In this case the polygon margins are ridges on which lie large boulders. This image is 3 km wide and is illuminated from the lower left. Image MOC2–564. NASA/JPL. (b) Troughs are spaced 1.5 to 2.5 m apart near the Phoenix landing site at 68°N and 26° W. They are believed to represent ice-wedge activity. On Earth, ice wedges may also be manifested by either ridges or troughs. NASA/JPL/University of Arizona. See also color plate section.

of the cracks beyond the depth of actual tension). It is, thus, a combined function of the duration and intensity of a cold snap and the thermal conductivity of the soil. One might then wonder if more long-continued eras of cold create larger polygons. This is precisely what many geologists thought when 5–10 km scale polygonal troughs were discovered on Viking images of the northern plains of Mars. However, the theory of ice-wedge formation, coupled with the rheology of ice (Glen's law) show that this cannot be the case.

Tensile stresses in ice due to contraction endure only so long as creep does not relax them. Slow freezing that extends to great depths requires a long period of time, during which the ice has time to flow under the applied stress and zero the stress. If the cold snap is not quick, it cannot generate tensile stresses and deep cracks cannot develop. This argument was applied to the polygonal terrain on Mars to show conclusively that such large polygons cannot have been created by thermal contraction (Pechmann, 1980). They may instead be due to draping of compacting sediments over a pre-existing cratered terrain (McGill, 1992). Only recently, with the advent of the very high-resolution imaging possible with the HiRISE camera system and the Phoenix lander have true ice-wedge polygons been observed on Mars (Figure 11.12). These polygons have the expected dimensions of 2–3 m across and closely resemble terrestrial ice-wedge polygons.

The observation of apparently fresh ice-wedge polygons on Mars suggests that liquid water on the surface is not, in fact, necessary for their formation. Perhaps the crack fillings on Mars are dust that has drifted into gaping cracks, not frozen water, and we are really looking at dust-wedge polygons in the Martian permafrost.

Block fields. Block fields are enigmatic periglacial features that have been found in many areas formerly occupied by permafrost. They appear as gently sloping, nearly planar surfaces up to many kilometers in extent, which are covered with boulders of roughly the same size. Block fields lack any obvious matrix material, although when active they may have contained interstitial ice. They may be related to frost heaving in some way as they show little sign of lateral movement: The blocks in these fields seem to have formed in place from jointed bedrock.

11.5.3 Thermokarst

Permafrost betrays its presence most clearly when it is about to disappear. The most dramatic landforms created by permafrost are formed as it is melting during a period of climatic warming. Permafrost does not melt uniformly: Small variations in surface thermal conductivity and exposure become amplified by positive feedback and are expressed as topographic features. The most obvious result of melting permafrost is a volume change. Thawed permafrost expels water and contracts, sagging downward into small ponds that collect more water and enhance melting. Such thaw lakes are common, creating landscapes packed with kilometer-diameter circular to elliptical ponds that are often aligned with the prevailing wind. Such lakes constitute the infamous Carolina Bays, which impact-crater enthusiasts persistently claim to be of impact origin in spite of the complete lack of evidence for impacts. Extensive fields of active thaw lakes occur in lowland areas of northern Alaska, Yukon Territory, and northern Russia. Depressions believed to be thaw lakes have been recognized in the catastrophic outflow channels on Mars, suggesting at least one era of warming on that currently chilly planet.

Once depressions are created by melting permafrost, the scarps that form at the interface between the thawed and still-frozen ground are subject to rapid denudation that removes the insulating surface layer and accelerates the disintegration of the permafrost. These small scarps retreat rapidly, forming shallow cirques that cut into the frozen ground.

Asymmetric valleys are common in permafrost terrains and may be accentuated by its decay. North–south asymmetry develops predominantly because of differences of exposure to solar radiation. East–west asymmetry may also develop because of differing exposure to the prevailing wind and the resulting differences in the depth of wind-drifted snow and chilling by the wind.

The Southern Polar Cap of Mars is subject to another kind of weathering akin to thermokarst disintegration, but not involving liquid water. Thin layers (~10 m) of residual CO_2 ice overlying water ice sublime away to create coalescing circular pits informally dubbed "Swiss-cheese terrain." These pits range from a few hundred meters up to a kilometer across and are approximately 10 m deep.

Further reading

The classic American study of the geological effects of the Pleistocene continental ice sheets is the fat book of Flint (1971). The basic physics of glacier flow, mass balance, and ice sheet formation is Paterson (1999), a book that has gone through many editions (a fourth has just appeared), but remains the most lucid of several such books. Readers seeking to visually feast on glacial features and phenomena should peruse the picture book by Post and LaChapelle (2000), which also contains much wisdom in addition to its magnificent photographs. The ability of glaciers and ice sheets to create landscapes is treated in Sugden and John (1976), a book that has unfortunately not been updated in recent years. The mechanics of cold soil and its implications for landform evolution is the topic of Williams and Smith (1991), while the more observational aspects of periglacial environments is well covered by Washburn (1980). The geomorphology of both glaciated regions and periglacial environments is the subject of a pair of books by Embelton and King (1975a, b).

Exercises

11.1 Rheology in space

The Maxwell relaxation time τ_M for ice at 273K (0°C) is 100 minutes. Use the approximately universal relation:

$$\dot{\varepsilon} = A(\sigma)e^{-gT_m/T}$$

where g ~ 26 for non-metals. Compute τ_M for the following substances on the indicated planetary body (suppose σ is the same as for ice at the τ_M given, and that all materials have nearly the same shear modulus):

Methane, ammonia on Triton, surface temperature 45K
Ammonia, CO_2 on Ganymede, surface temperature 145K
CO_2, water on Mars polar caps, mean temperature ca. 170K
Salt (NaCl), olivine (forsterite) on Venus, surface temperature 750K.

Melting points are:

Methane	$T_m = 91K$
Ammonia	$T_m = 196K$
CO_2	$T_m = 216K$
Salt (NaCl)	$T_m = 1100K$
Olivine (Fo)	$T_m = 2200K$.

Which of the two materials is more likely to show evidence of flow in surface deposits ("glaciers")? Compare your computed flow rates to those of a terrestrial glacier.

11.2 The inner heat

Compute the rate of basal melting of a warm-based glacier on the Earth and on Mars (use some plausible means of estimating the heat flow of Mars, perhaps by assuming that the rate of heat generation per unit volume on Mars is the same as on the Earth). The average heat flow on the Earth is 80 mW/m^2. Does this suggest a way of estimating the minimum rate of snow precipitation necessary to support warm-based glaciers? If so, what is this minimum for Mars?

11.3 Infinite flowing ice

Derive the velocity profile for an infinitely wide sheet of power-law material (Glen's law, Equation (11.1)) of uniform thickness H creeping down a surface with a constant slope α. That is, show how Equation (11.7) comes about in the case of a zero basal sliding velocity. Using the data on Glen's law given in Section 11.2.1, estimate the surface velocity of an ice sheet 3 km thick with a surface slope of 0.01 (about half a degree) at a temperature of 0°C and at −40°C.

11.4 Only the surface matters!

Demonstrate the assertion in Section 11.3.1 that, in a straight reach of a glacier, the basal shear stress depends only upon the surface slope α_s, not the basal slope α_b. Hint: Consider the forces acting on the left and right sides of a block of downglacier length Δx, then get the basal shear stress from the force per unit area (if you get really stuck on this, see Patterson, 1999, p. 241).

11.5 Permafrosty Mars

Estimate the thickness of the permafrost layer on Mars by computing the depth of the 273 K isotherm below the surface at the equator and at 50° N. Assume a global average heat flow of 30 mW/m^2 and a thermal conductivity of the Martian surface layers of about 1.5 W/m-K. The average surface temperature of Mars at the equator is 240 K, but falls to a chilly 200 K at 50° N. Compute the depth to the bottom of the permafrost at these two latitudes. How does this depth compare to the depth of the (Martian) seasonal temperature fluctuation?