# 11

# Glacial processes and landforms

## 11.1 Glacier characteristics and dynamics

In some ways it is appropriate to describe the Earth as an ice planet. At present some 10 per cent of the surface of the continents is covered by ice, but as little as 18 000 a BP the figure was nearer 30 per cent. This greater extent of ice sheets has characterized much of the past 2–3 Ma and consequently landforms created by glacial action cover large areas that are now experiencing non-glacial climates. Since the majority of glacial landforms are formed beneath ice sheets and glaciers it is only when these ice bodies retreat that many types of glacial landform are revealed. This situation poses particular problems in understanding the processes whereby glaciers produce landforms since the mechanisms at work can only rarely be observed directly. In this chapter we examine specific landforms created by glaciers, but as many glaciated landscapes are located in regions which are now experiencing quite different temperate conditions we consider the features of glacial landscapes as a whole in Chapter 14.

### 11.1.1 Glacier distribution and classification

Ice covers some 14.9 million km$^2$ at the present day. Most of this is accounted for by the two vast ice sheets of Antarctica and Greenland, the remaining 4 per cent comprising ice caps and glaciers located mainly in high latitudes, and innumerable glaciers found at high elevations in all latitudes (Table 11.1). Glaciers can be classified morphologically on the basis of their relationship to the underlying bedrock topography (Table 11.2; Figs 11.1 and 11.2). It is difficult to generalize about the distribution of the different glacier types since, apart from ice sheets, most occur over a broad range of latitudes.

The occurrence of glaciers is determined not only by climate but also by topography since there must be a suitable surface on which ice can accumulate. Glaciers can only form where snow persists from year to year. Whether this occurs is determined by the rate of snow accumulation, which is largely a function of the amount of precipitation falling as snow, and the rate of melting, which is largely a function of temperature (Fig. 11.3). In the mid-latitudes the seasonal distribution of precipitation is more crucial than the total amount. Extensive glaciers are not found, for instance, in the high mountains along mid-latitude west

**Table 11.1** Present global distribution of glaciers

| REGION | | APPROXIMATE AREA (km$^2$) |
|---|---|---|
| *North polar area* | | |
| Greenland ice sheet | | 1 726 400 |
| Queen Elizabeth Islands | | 106 988 |
| Other Greenland glaciers | | 76 200 |
| Spitsbergen and Nordaustlandet | | 58 016 |
| Other | | 114 012 |
| | Subtotal | 2 081 616 |
| *Continental North America* | | |
| Alaska | | 51 476 |
| Other | | 25 404 |
| | Subtotal | 76 880 |
| South American Cordillera | | 26 500 |
| Europe | | 9 276 |
| Asian continent | | 115 021 |
| African continent | | 12 |
| Australasia | | 1 015 |
| *South polar area* | | |
| Antarctic ice sheet (excluding ice shelves) | | 12 535 000 |
| Other glaciers on Antarctic continent | | 50 000 |
| Sub-Antarctic islands | | 3 000 |
| | Subtotal | 12 588 000 |
| | Total | 14 898 320 |

*Source*: Data from R. F. Flint, (1971) *Glacial and Quaternary Geology*, Wiley, New York, Table 4–B, pp. 76–7, based on various sources.

**Table 11.2** Morphological classification of glaciers

| BASIC TYPE | COMPONENT OR SUBTYPES | CHARACTERISTICS |
|---|---|---|
| **Ice sheet** and **ice cap** (unconstrained by topography)* | **Ice dome** | A dome-like ice mass with a convex cross-profile formed in response to the basic flow characteristics of ice |
| | **Outlet glacier** | Glaciers which radiate out from an ice dome often occupying significant depressions (Fig. 11.1). Within the ice dome they can be distinguished by a zone of rapidly moving ice termed an ice stream |
| **Ice shelf** | | A floating ice cap or part of an ice sheet only partially constrained by the coastal configuration and which deforms under its own weight |
| Glaciers constrained by topography | **Ice field** | A roughly level area of ice distinguished from an ice cap because of the absence of a dome-like form and the control on ice flow exerted by the underlying topography |
| | **Cirque glacier** | A small ice mass usually occupying an armchair-shaped bedrock hollow and characteristically wide in relation to its length |
| | **Valley glacier** | A glacier which occupies a rock valley and is overlooked by rock cliffs (Fig. 11.2). They may originate in an ice field or a cirque glacier into which they may imperceptibly merge at their upper end. Large valley glaciers may be joined by tributary glaciers, forming a dendritic pattern of ice flow |
| | Other small glaciers | Glaciers which occur in a wide variety of topographic positions, but all of which are closely controlled by the underlying topography |

*Ice sheets and ice caps are essentially distinguished on the basis of size, the former exceeding around 50 000 km$^2$ in area.
*Source*: Based on classification and discussion in D. E. Sugden and B. S. John (1976) *Glaciers and Landscape*. Edward Arnold, London.

*Fig. 11.1* *Outlet glaciers near Bartholins Brae, Blosseville Kyst, Greenland. Note the medial and lateral moraine. (Photo courtesy of the Geodetic Institute, Copenhagen, Denmark.)*

*Fig. 11.2* *Harker Glacier, a valley glacier discharging into a fjord, South Georgia. (Photo courtesy D. E. Sugden).*

coasts in spite of heavy precipitation because most falls as rain in summer and the relatively high summer temperatures lead to substantial melting of winter snows. By contrast the largest ice sheets receive very small amounts of precipitation although this nearly all falls as snow. The mean annual precipitation over the northern part of the Greenland ice sheet is only 150 mm, but rates of summer melting are so low that ice accumulation takes place.

Although there are other sources of heat supplied to glaciers, solar radiation is by far the most significant in

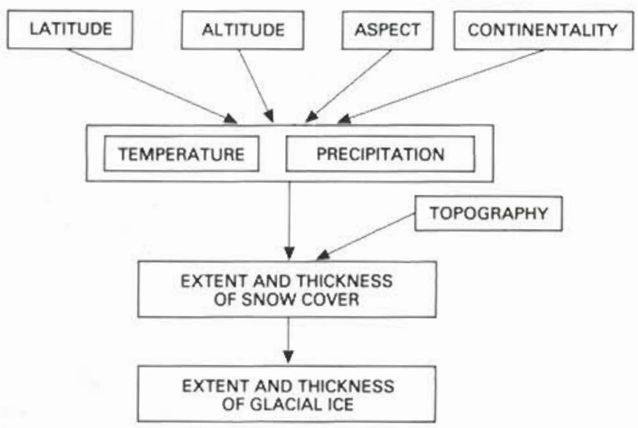


*Fig. 11.3* *Major factors controlling the occurrence of glaciers.*

accounting for their global distribution. Because melting can occur only at temperatures above 0 °C it is the mean summer temperature rather than the mean annual value that influences the amount of snow-melt that will occur. Latitude, elevation, aspect and continentality all exert an indirect control over the distribution of glaciers through their influence on precipitation and temperature (Fig. 11.3). The effect of continentality on precipitation is a particularly significant factor in the development of ice sheets. This is illustrated by the pattern of snow accumulation over the Antarctic ice sheet which is closely related to the distance from the nearest area of ice free ocean in summer.

### 11.1.2 Characteristics of glacier ice

Glaciers are composed not only of ice but also of smaller amounts of air, water and rock debris. A property of ice of considerable geomorphic importance is its ability to deform and flow under its own weight. Although glacier ice can form directly from the freezing of liquid water or water vapour on the glacier surface, snowfall is the most important source. Snow with a density of only 50–70 kg $m^{-3}$ is gradually converted to ice by a complex process of compaction and recrystallization through an intermediate stage known as **firn**. This is composed of a loosely consolidated mass of ice crystals with a bulk density of around 400 kg $m^{-3}$. The transformation into ice with a density of about 800 kg $m^{-3}$ involves an increase in crystal size and the closing of voids, but the final conversion to true glacier ice (density of about 900 kg $m^{-3}$) only occurs when the pressure of overlying ice leads to the elimination of most of the remaining air bubbles.

Temperature is a key characteristic of glacier ice as it exerts a profound influence on glacier behaviour. In addition to solar radiation heat can be supplied to a glacier from the surface by the incorporation of firn which is warmer than the existing glacier ice and by the release of latent heat as water is refrozen. A glacier is also warmed from below by geothermal heat and by frictional heat generated by sliding and the deformation of ice at its base. Variations in the supply of heat from these sources give rise to two types of glacier ice which have fundamentally different geomorphic properties.

**Cold ice** is at a temperature below melting point, but **warm ice** is so close to melting point that it contains liquid water. Strictly we should use the term **pressure melting point** here as the freezing point of water decreases as pressure increases. For example, the melting point at a depth of 2164 m at the base of the Antarctic ice sheet at Byrd Station is –1.6 °C. Cold ice occurs where the glacier surface experiences very low winter temperatures or where low summer temperatures lead to negligible surface melting.

Warm ice occurs where geothermal or frictional heat are sufficient to raise the temperature of at least the basal ice to the pressure melting point. In small glaciers the percolation of surface meltwater may be the most important heat source since the refreezing of 1 g of water releases sufficient heat to raise the temperature of 160 g of ice by 1 °C. The rate of firn accumulation can also be important since a high rate can introduce a significant quantity of cold material into the glacier. This helps to counteract the geothermal heating at the glacier base and so reduces the rate of increase in temperature with depth. Once the base of a glacier reaches pressure melting point all the heat from geothermal and frictional sources can be used in melting rather than simply raising the temperature of the ice.

It has been accepted practice to classify glaciers as temperate or polar depending on whether they are supposed to be composed of warm or cold ice. However, it is now recognized that not only do both types of ice occur in glaciers in both temperate and polar environments but that both types may be present in an individual glacier or ice sheet. The Antarctic ice sheet, for example, is composed predominantly of cold ice but is now known to contain basal layers of warm ice in places. In the following discussion we shall refer to those glaciers or parts of glaciers with a basal layer at pressure melting point as **warm-based** and those with a basal layer below pressure melting point as **cold-based**. A further important distinction is that between **active ice** which is moving downslope and being replenished by fresh accumulations of snow in its source region and **stagnant ice** which is decaying *in situ* and has ceased to experience any significant lateral movement.

### 11.1.3 The glacier mass balance

The gains and losses of ice experienced by a glacier constitute its **mass balance** or **glacial budget**. The ice, firn and snow added to the glacier constitute its **accumulation**. The losses, or **ablation**, arise largely from melting in most glaciers, but evaporation, **sublimation** (the direct conversion of ice to water vapour), wind erosion and **calving** (the breaking away of blocks of ice into standing water) may also be important in certain situations.

A mass balance is normally calculated over the **balance year**. This is the interval between two successive times at which ablation has reached a maximum value. Although this normally occurs at the end of the summer the balance year may not be exactly 365 days. The total masses added to and lost from a glacier during the balance year are called, respectively, the **gross annual accumulation** and **gross annual ablation**. The difference between these two amounts is the **net annual accumulation** or **net annual ablation**, depending on whether there has been a net gain or loss in mass. The **net specific balance** is the net annual accumulation or ablation at a particular point on a glacier, and the integration of a large number of such point measurements

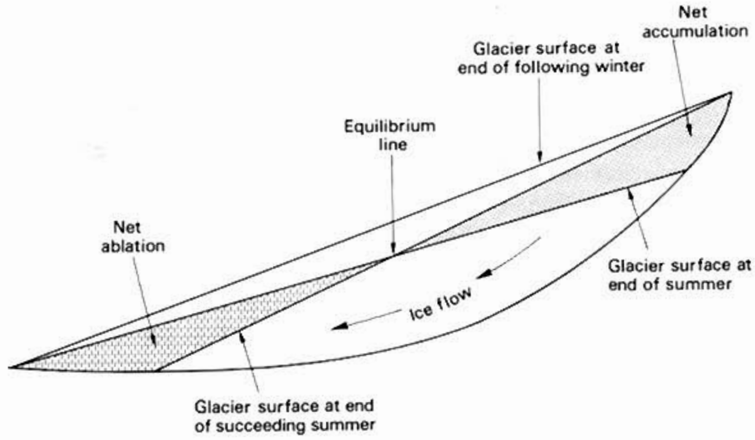


*Fig. 11.4 Variations in accumulation and ablation down an idealized glacier. The distribution of net annual accumulation and ablation over a glacier is of great significance to glacier movement. Net ablation generally increases down-glacier below the equilibrium line while net accumulation tends to increase up-glacier above this point. The rate of increase in the net balance with height up a glacier is important for the rate of ice movement because the higher the rate of increase the faster is the rate of ice movement needed to maintain the same glacier profile.*

can provide an estimate of the total mass balance.

The taking of such measurements is a difficult and time-consuming task and consequently there are few accurate mass balances for real glaciers. The data that do exist clearly show that the annual net balance generally varies over a glacier in a systematic manner with a positive balance (net accumulation) in the upper part of the glacier and a negative balance (net ablation) in the lower part. These two zones meet at the **equilibrium line** where accumulation is exactly compensated by ablation and where the mass balance is therefore zero (Fig. 11.4). It is important to draw a distinction between the situation over an entire balance year and seasonal variations since, depending on the season, much or all of a glacier may be experiencing accumulation or ablation.

### 11.1.4 Glacier motion

#### *11.1.4.1 Mechanisms of ice movement*

The movement of glaciers can in many respects be treated in the same way as other forms of mass movement, although ice masses have special properties which must be taken into consideration. Two, more or less distinct, types of ice movement can be distinguished – **internal deformation** and **basal sliding**. The relative importance of these two mechanisms varies significantly with basal sliding accounting for up to 90 per cent of the movement of warm-based glaciers, but being largely inoperative in cold-based glaciers. Glacier ice deforms internally because it is subject to stress. At any one point in a glacier this stress has two components: hydrostatic pressure, which is exerted in all directions and is related to the mass of overlying ice, and shear stress, which is related not only to the thickness of overlying ice but also to the surface slope of the glacier. High shear stresses are generated towards the base of thick glaciers with steep slopes whereas lower basal shear stresses are produced by thin, gently sloping glaciers. The variations in shear stress at the base of most glaciers are, however, not large, usually lying between 0.05 and 0.15 MPa with a mean of about 0.1 MPa.

The main mode of internal deformation involves slippage within and between ice crystals and is called **creep**. Although the exact mechanism is not fully understood, various attempts have been made to model ice deformation by this process. In one such formulation, known as **Glen's power flow law**, the predicted rate of deformation depends not only on the shear stress but also on the temperature of the ice (Box 11.1). This model of creep appears to accord with several observations about the behaviour of glaciers. First, deformation rates are at a maximum at the base of the glacier, where both stresses and, in the case of cold ice, temperatures are highest. Secondly, it explains the faster rates of movement of warm ice. Thirdly, it accounts for the way glaciers can regulate their discharge through negative feedback since an increase in glacier thickness will increase the basal shear stress which will in turn accelerate the rate of ice flow and thereby reduce ice thickness. Where stresses within the glacier cannot be accommodated sufficiently quickly the ice may move by fracturing. This is most likely to occur on the margins of glaciers where thrusting can give rise to shear fractures and tensional stresses can produce crevasses (see Section 11.1.5.2).

Basal sliding involves four major processes. One mechanism is the slippage of the glacier bed over a thin

### Box 11.1 Glen's power flow law

On the basis of laboratory experiments J. W. Glen established that the strain rate in a block of ice soon attains a steady value on being subject to a constant stress. The relationship determined by Glen, known as the power flow law, has been adapted by J. F. Nye to apply to glaciers. The relationship is

$$e = A\tau^n$$

where $e$ is the strain rate, $\tau$ the effective shear stress, $A$ a constant related to temperature and $n$ an exponent.

The values of $n$ have been determined by several investigators and vary from 1.7 to 4.5, the mean value being around 3. This exponent makes the strain rate highly dependent on shear stress since with a value of $n$ of 3 a doubling of the shear stress produces an eightfold increase in strain rate. In Glen's experiments the value of $A$ was found to be significantly affected by differences in temperature, changing from 0.17 at 0 °C to only 0.0017 at −13 °C, a variation of two orders of magnitude.

layer of water only a few millimetres thick. This can significantly reduce the friction between the glacier and its bed and considerably increase rates of glacier movement. A second mechanism, known as **regelation creep**, involves the movement of a warm-based glacier over minor irregularities in its bed. The higher pressures on the up-glacier side of an obstacle lead to melting and the water produced migrates to the zone of lower pressure on the lee side of the obstruction where it refreezes. The process is most effective where the obstacle is small enough (less than about 1 m across) to enable the latent heat released by the refreezing to be effectively conducted to the up-glacier side where it promotes further melting. Where larger obstacles are present on the glacier bed a third process known as **enhanced basal creep** can become important. The increased stress on the up-glacier side of such an obstruction allows the ice to flow round it. This mechanism promotes the movement of basal ice even where it is below pressure melting-point.

Traditionally, the analysis of basal sliding has been founded on the assumption that glaciers move over a rigid rock bed. That this is frequently not the case is clear from the sediment-covered surfaces found immediately in front of retreating glaciers. Furthermore, some 80 per cent of the area of Europe and North America covered by ice sheets during the glacial advances of the Pleistocene is mantled by unconsolidated sediments rather than bedrock. Many instances of deformation structures attributable to ice sheet movement are known from these sediments, and recent experimental work has demonstrated that even quite coarse-grained sediments can be easily deformed beneath a glacier. All this suggests that a fourth, and possibly widespread, mechanism of glacier motion involves a significant part of ice movement being accomplished through the deformation of subglacial sediments. By facilitating movement such **bed deformation** would increase the rate of ice flow or allow flow to occur at a similar rate over a lower bed gradient or lead to a combination of these effects. In order for bed deformation to occur the subglacial material must be saturated with water and maintain a high pore-water pressure. Such conditions will only be met if the glacier bed is at the pressure melting-point and if basal water accumulates in the sediment rather than being drained away.

#### *11.1.4.2 Glacier flow*

Average rates of movement vary enormously from one glacier to another but typically lie in the range 3–300 m a$^{-1}$. The velocity of ice flow also varies spatially and temporally within an individual glacier. If we look at a vertical profile we find that the flow velocity is at a maximum at the glacier surface and decreases downwards (Fig. 11.5(A), (B)). This at first sight appears paradoxical since, as we have seen, most movement occurs in the basal layers of a

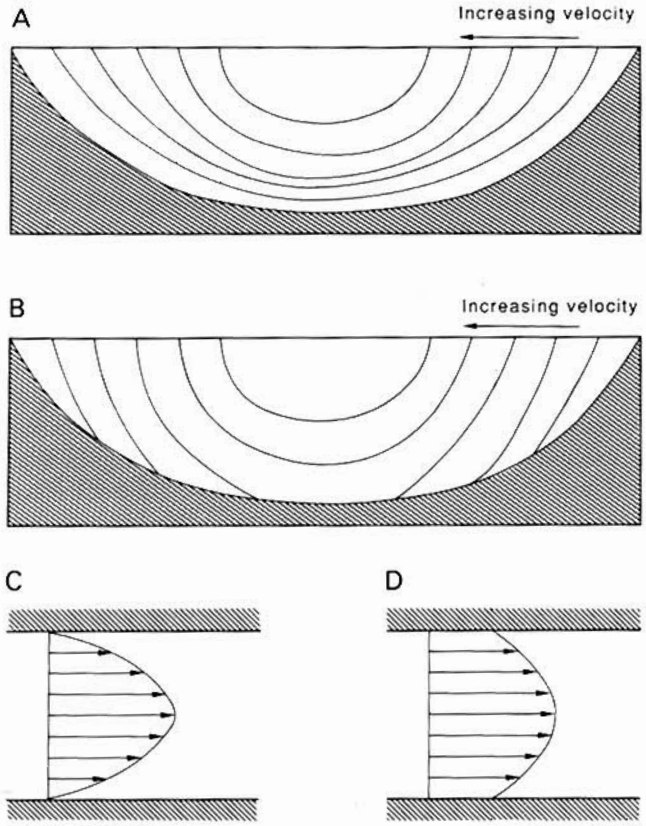


***Fig. 11.5*** *Schematic representation of vertical and horizontal variations in the velocity of valley glacier movement: (A) cross-profile vertical variation in velocity in a glacier with no basal sliding; (B) vertical cross-profile variations in velocity where basal sliding is significant (isolines intersect valley floor); (C) plan view of valley glacier illustrating variations in velocity at surface where there is no sliding along valley walls. (D) plan view of velocity variation at surface where there is significant sliding along valley walls.*

glacier where the shear stress is at a maximum. We have to remember, however, that each imaginary layer of ice is not only moving as a result of the shear stress at that level but is also being carried along by the moving layer below it. The velocities at each level in the glacier are thus cumulative from the base to the surface. There are also significant variations in the velocity across confined glaciers since the increased friction between the ice and rock wall retards ice movement there (Fig. 11.5(C), (D)).

Changes in discharge longitudinally down a glacier are strongly influenced by variations in total accumulation and ablation. Discharge is generally highest around the equilibrium line because the cumulative volume of ice increases from the head of the glacier to reach a maximum at this point. Below the equilibrium line discharge decreases progressively down-glacier with the increasing net loss of ice through ablation (Fig. 11.4).

Because snow is being added from above in the accumulation zone there must be some downward movement of ice from the surface. Similarly, in the ablation zone there must be some upward movement to maintain the surface form by replacing ice lost through ablation. Evidence of these vertical components of movement is provided by the burial of surface debris in the accumulation zone and the emergence at the surface of previously buried debris in the ablation zone. Such longitudinal variations in ice movement are associated with two different types of flow regime within a glacier (Fig. 11.6(A)). Above the equilibrium line **extending flow** predominates since the ice becomes more 'stretched out' down-glacier as the velocity of flow increases with the steeper angle of movement. Conversely, there is a reduction in velocity in the zone of **compressive flow** below the equilibrium line as the upward component of ice movement becomes more marked. In reality the longitudinal zonation of extending and compressive flow is more complex than this since the flow regime is also affected by bedrock topography; extending flow is promoted over convex bedrock surfaces, especially in **ice falls** which occur where a glacier flows down a steep slope, whereas compressive flow prevails over concave surfaces (Fig. 11.6(B)). The upward and downward component of ice movement associated with zones of compressive and extending flow is accommodated along **slip lines** which meet the glacier surface at around 45° and represent trajectories of maximum shear stress. Movement along these slip lines can occur through creep as well as by thrusting along 'faults' within the ice.

In ice sheets and ice caps relatively narrow zones of ice occur in which movement is much more rapid than the adjacent parts of the ice mass. These **ice streams** are responsible for a significant proportion of ice movement in large ice masses and often feed outlet glaciers (Fig. 11.1). Several major ice streams have been identified in the Antarctic ice sheet and seismic measurements indicate that at least one of these in the west Antarctic, known as Ice Stream B, is moving over a bed composed of saturated, deformable sediment.

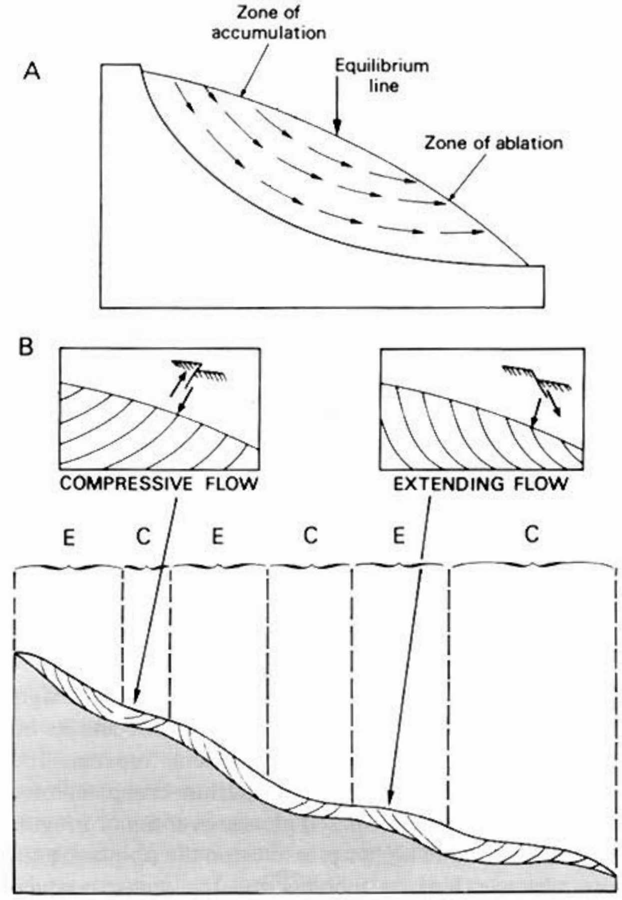


*Fig. 11.6* *Flow patterns within a glacier: (A) basic pattern of downward movement of ice in the accumulation zone and upward movement in the ablation zone; (B) zones of extending and compressive flow related to irregularities in valley floor morphology (E = extending, C = compressive). The lines indicating the direction of ice movement with respect to the glacier surface are slip lines which indicate the direction in which the ice has the maximum tendency to shear. (Part (B) modified from J. F. Nye (1952)* Journal of Glaciology **2**, *Fig. 7, p.88 and Fig. 8, p. 90.)*

Changes in ice movement over time occur in response to alterations in the mass balance arising from particular meteorological conditions. Cooler weather promotes accumulation and the glacier thickens and increases its velocity, whereas in warmer conditions increasing ablation induces thinning and a reduction in velocity, both adjustments progressing until a new equilibrium profile is attained. Changes in mass balance can be transmitted down-glacier by **kinematic waves**. These are complex phenomena, but we can grasp a basic understanding of how they function by imagining an increase in accumulation which causes a local raising of the glacier surface. The thicker ice under the bulge will begin to move more rapidly than the surrounding ice because the basal shear stress will be higher. The zone of thicker ice then moves down-glacier as a wave at between two to five times the velocity of the ice itself. Adjustments to the profile of a glacier arising from changes in mass balance can consequently be transmitted to the glacier snout much more rapidly by kinematic waves than by physical movement of the ice itself. Although kinematic waves have been observed moving down glaciers, in exceptional cases involving a surface rise of some 100 m, they need not necessarily have any surface expression since the increase in discharge represented by the wave can occur through adjustments in the ice flow other than those involving alterations in glacier thickness.

The rapidity with which glaciers respond to changes in mass balance and attain a new equilibrium form varies enormously but it is generally related to the size of the ice mass. Relaxation times vary from 3 to 30 a for valley glaciers to thousands of years for ice sheets. Such long and variable relaxation times make it difficult to relate changes in glacier movement to particular climatic or meteorological conditions.

The most spectacular temporal variations in glacier movement are **glacier surges**. These events, during which ice-flow velocities temporarily reach between 10 and 100 times their normal value, are experienced by some, but not all, glaciers. Surging glaciers are fairly common, some 204 having been identified in North America alone, but they seem to have few characteristics in common. In some, surges occur on a more or less regular cycle with a periodicity ranging from 15 to 100 a or more, but in others surging is unpredictable. They also occur in a wide range of glacier types, both warm and cold-based. Surging is initiated when a threshold of instability is reached and ice in the upper ablation zone begins to move rapidly down-glacier. In some cases this may precipitate a rapid movement of the glacier snout, such as the 45 km advance at a rate of up to 5 m $h^{-1}$ recorded during a surge by Bruárjökull, Iceland. In many North American glaciers, however, the effect of a surge is mainly confined to a thickening of ice towards the snout.

Surging is still one of the least understood aspects of glacier movement; it is necessary to explain both the mechanism that triggers a surge and the high flow velocities subsequently attained. The greater abundance of meltwater commonly associated with surges suggests that the rapid movement may be related to greatly enhanced rates of basal sliding through the effect of basal water reducing friction along the glacier bed. The trigger mechanism is much more difficult to explain. It may be linked to external factors such as earthquakes or increased precipitation, but the regularity of surging in many glaciers and the rather similar quantities of ice moved in each surge suggests that it is more likely an intrinsic element of the behaviour of some glaciers. Such glaciers may experience a gradual build-up of basal water as a result of changes in the subglacial hydrological system which eventually precipitates an increase in basal sliding. This could then promote a cycle of positive feedback in which the initial increase in basal sliding generates more meltwater because of the greater frictional heating associated with the higher rate of basal movement.

The concept of glacier movement assisted by the deformation of subglacial sediments has also led to the idea that periodic changes in pore-water pressure could promote surging behaviour. Data collected from boreholes drilled into Trapridge Glacier, Yukon Territory, Canada, indicate that it rests, at least in part, on a bed of potentially deformable sediment. It has been suggested that during non-surging interludes this substrate is efficiently drained through a system of channels and pipes; consequently, pore-water pressures in the sediment are low and significant deformation does not occur. An increase in basal shear stress, perhaps related to an increase in glacier thickness, could promote an initial increase in subglacial sediment deformation leading to the disruption of the subglacial drainage system. Such a reduction in permeability could then lead to a build-up of water within the sediment voids, and the resulting increase in pore-water pressure could induce a considerable increase in bed deformation and thereby glacier movement.

#### *11.1.4.3 Short-term glacier fluctuations*

Glaciers advance and retreat in response to changing meteorological and climatic conditions over a range of time scales. Here were are concerned only with the shorter-term fluctuations of up to a few hundred years, since longer-term adjustments are caused by major climatic changes and are therefore more appropriately considered in Chapter 14. The relationships between changing meteorological conditions and short-term climatic fluctuations and glacier behaviour are far more complex than was once assumed. Indeed, changes in weather and climate are so rapid in comparison with typical relaxation times of glacial systems that glaciers in true equilibrium with prevailing climatic conditions are rare. Only over the past 40 a or so have detailed records of mass balance been kept and these for only a few glaciers. Over a longer period less detailed information on glacier fluctuations can be gained from maps, photographs, drawings, historical reports and, over longer periods, botanical evidence.

Variations in rates of glacier movement over weeks and months appear to be related to the abundance of meltwater associated with seasonal changes in ablation rates. Year-to-year fluctuations follow, with a lag, variations in winter snowfall and mean annual temperature, while the widespread and significant phases of glacier advance and subsequent retreat documented over the past 300 a or so appear to be related to the cool climatic interval known as the 'Little Ice Age'. Not all glaciers, however, respond in the same way to a particular climatic fluctuation. Glaciers with a low

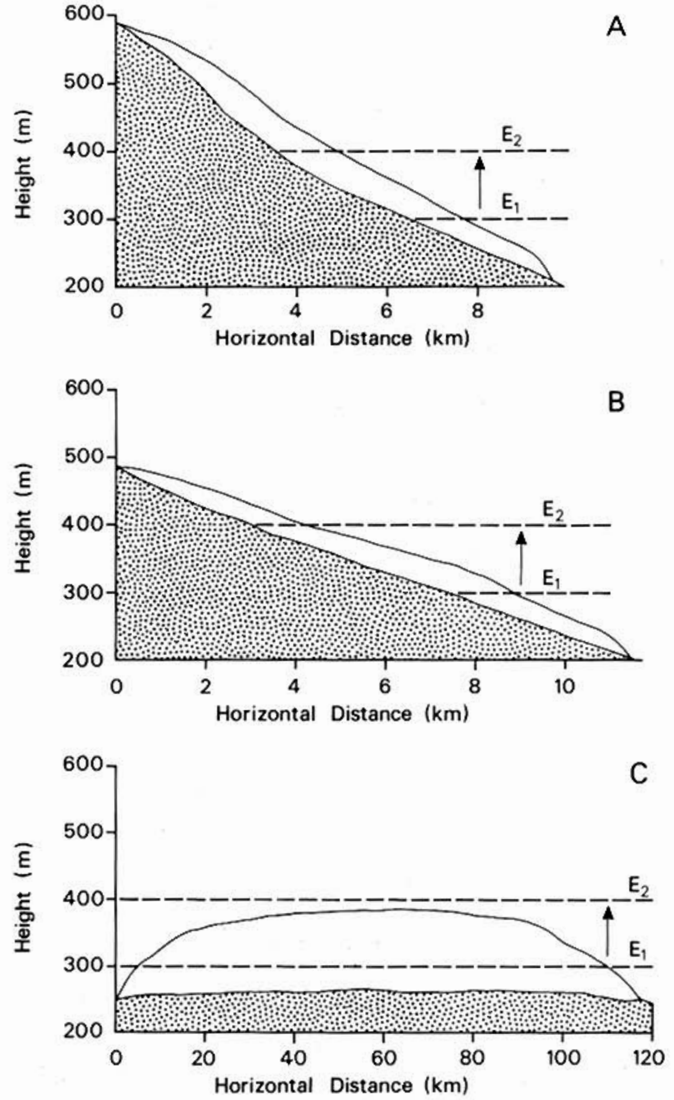


*Fig. 11.7* *Effects of a rise of 100 m in the altitude of the equilibrium line ($E_1$ to $E_2$) on a steeply inclined valley glacier (A), a more gently inclined valley glacier (B) and on an ice cap (C). Note that as the slope of the ice surface decreases from A to C the proportion of the glacier affected by the change in the altitude of the equilibrium line increases.*

surface gradient are more susceptible to such changes than those with a steep slope since a vertical change in the equilibrium line due to, for instance, a change in mean annual temperature, will affect a much larger proportion of the latter and thus produce a much more substantial change in the areas of ablation and accumulation (Fig. 11.7).

### 11.1.5 Glacier morphology

*11.1.5.1 Large-scale forms*

While the overall shape of confined glaciers is of course closely related to topography, that of unconfined ice masses

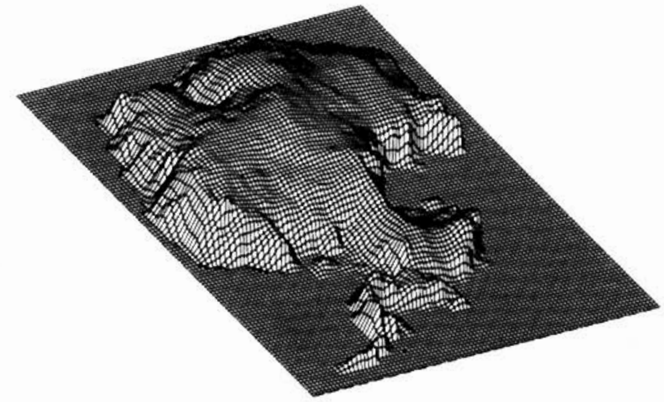

*Fig. 11.8* *Morphology of the Antarctic ice sheet illustrated by a computer-generated three-dimensional reconstruction. (From D. J. Drewry (ed.) (1983)* Antarctica: Glaciological and Geophysical Folio. *(Scott Polar Research Institute, Cambridge Sheet 2, Fig. 2b.)*

is largely controlled by the dynamics of ice flow. This can be seen most clearly in the convex form of ice domes (Fig. 11.8). These have a maximum thickness which may exceed 4000 m in Antarctica. Where there is sufficient accumulation the ice builds up until the level of basal shear stress promotes significant ice deformation. As we have already noted the amount of shear stress is related to glacier thickness and surface slope so where the ice is thinner around the margins of the dome a steeper slope is required to maintain the ice flow, and conversely where the ice is thicker below the crest of the dome a less steep gradient is needed to sustain movement. Theoretical models of ice deformation generally accord quite well with the form of most ice domes (Box 11.2), but in reality a smoothly curved convex shape is prevented by factors such as bedrock irregularities underlying the glacier and spatial variations in ice tempera-

### Box 11.2 Estimating the profile of an ice dome

On the basis of a number of assumptions J. F. Nye proposed a simple formula which enables an estimate of the cross-profile of an ice dome to be made. The most important assumptions are that the ice is actively flowing and the ice dome is in equilibrium, but it is also assumed that the ice flow is not affected by variations in temperature, by localized accumulation or by bed irregularities. The elevation (m) of the ice surface at any point ($h$) is given by

$$h = \sqrt{(2h_0 s)}$$

where $h_0$ is 11 m, and $s$ the horizontal distance from the margin of the ice dome (m).

This formula provides a fairly good approximation of the form of most ice domes, although it tends to overestimate the maximum elevation. Not only are such calculations useful for examining existing ice domes, but they are also a useful way of estimating the form of the large ice sheets of the Pleistocene.

ture and accumulation. Movement of ice in an ice dome occurs normal to its contours and it can operate through basal sliding as well as internal deformation.

*11.1.5.2 Small-scale surface features*

Two sets of processes create forms on glacier surfaces – those related to ice movement and those associated with accumulation and ablation. Perhaps the most evident features are **crevasses** formed through ice fracture. They are vertical to subvertical cracks up to several metres across, but do not generally extend to a depth of more than 30 m since below this depth the ice is less rigid and is likely to accommodate to tensional stress by creep rather than fracture. Three types of crevasse can be identified and related to contrasting stress patterns (Fig. 11.9).

A feature common in the ablation zone of many glaciers and attributed to the metamorphism of ice during flow is a banding of white ice containing abundant air bubbles between denser bluish ice. Another form of patterning is that produced by **ogives**, or **Forbes bands**. These are alternating bands of light and dark ice lying across the glacier surface and are developed in some glaciers below ice falls (Fig. 11.10). Where movement is more rapid in the middle of the ice flow the bands are deflected down-glacier. The total width of each pair of light and dark bands corresponds to the distance moved by the glacier in a year. The dark bands are generated during the passage of ice across the ice fall in the summer when melting occurs. Conversely, light bands are produced in winter when fresh snow is incorporated into the ice as it traverses the ice fall.

On the surface of the large ice sheets of Greenland and Antarctica a variety of aeolian features have been recorded. The best documented are dunes called **sastrugi** constructed of hard-packed snow and aligned with the prevailing wind. They most commonly form in the lee of obstacles but are also known in open situations. Wind is also funnelled around rock outcrops protruding above an ice sheet, and snow is excavated from their lee side. On most glaciers the more significant small-scale surface features associated with ablation are produced by meltwater (see Section 11.4.1).

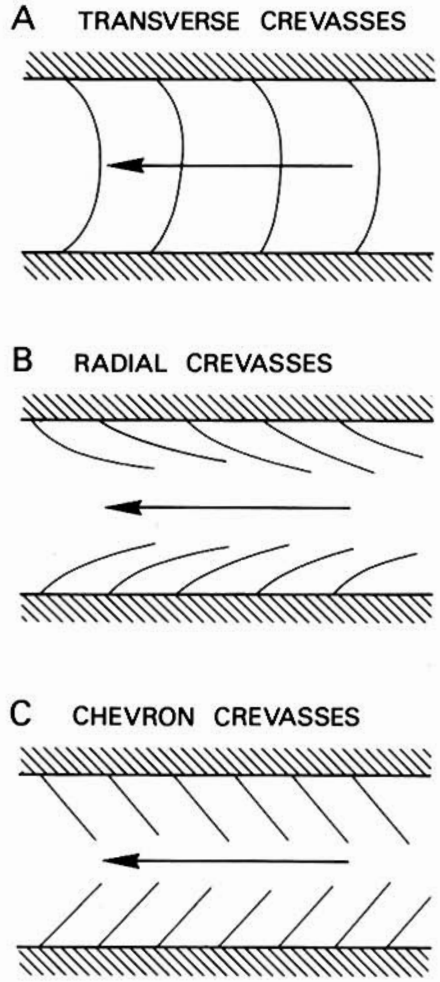


***Fig. 11.9*** *Types of crevasse occurring in valley glaciers: (A)* ***transverse crevasses*** *are associated with extending flow and result from expansion normal to the direction of ice flow; (B)* ***radial*** *(or* ***splay****)* ***crevasses*** *occur in zones of compressive flow, and although they also arise from expansion normal to the direction of ice flow they differ from transverse crevasses in being concave up-valley; (C)* ***chevron*** *(or* ***en echelon****)* ***crevasses*** *associated with the tensional stresses generated by the drag of the valley walls on the margins of a glacier.*

## 11.2 Glacial erosion

### 11.2.1 Mechanisms of erosion

Glacial erosion is accomplished by three major processes – abrasion, crushing and fracturing, and **joint-block removal**. Abrasion involves the scratching, grooving and polishing of bedrock by debris carried in the base of a glacier (Box 11.3). Evidence of its efficacy is provided by smoothed bedrock surfaces, often exhibiting parallel sets of fine grooves known as **striations**, and the production of fine particles (usually < 0.1 mm in diameter) known as **rock flour** and occurring in high concentrations in most glacier meltwater streams. The operation of abrasion has been further confirmed by limited direct observations of glacier beds and synthesized in laboratory studies. Its effectiveness is determined by glacier, bedrock and basal debris characteristics and the role of these factors is summarized in Table 11.3.

Evidence for crushing and fracturing comes from series of crescentic cracks and grooves called **chattermarks** found on some bedrock surfaces and thought to represent the effects of pressure exerted by basal glacier debris, and from the presence in glacial deposits of blocks which have clearly been sheared off from fresh bedrock. The typical basal shear stress of around 0.1 MPa is quite inadequate on its own to shear off bedrock protrusions, but where large boulders are being carried along the glacier bed a much more

***Fig. 11.10*** *Ogives visible on the surface of Austerdalsbreen, a valley glacier in southern Norway. Note the precipitous valley walls.*

## Box 11.3 Frictional shear at the bed of a glacier

Ice thickness determines the normal stress at the interface between debris and substrate at the base of a glacier, but for warm-based glaciers this stress is counteracted to some extent by basal water pressure. The basal frictional shear ($F$) is given by

$$F = A_c\mu(\nu h - p)$$

where $A_c$ is the area of contact, $\mu$ the coefficient of friction, $\nu$ the unit weight of ice, $h$ the ice thickness, and $p$ the basal water pressure.

As the effective basal pressure increases, the rate of abrasion, for any given sliding velocity and combination of basal debris/glacier bed hardness, will rise (Fig. B11.3). Eventually, however, as the basal normal pressure increases, the friction between a particle and the glacier bed will reach a point where the ice begins to flow over the particle and the rate of abrasion begins to fall. Ultimately, with a further increase in effective basal pressure the particle will stop moving and abrasion will cease.

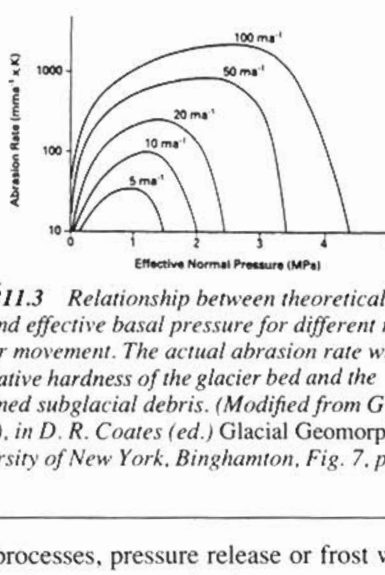


***Fig. B11.3*** *Relationship between theoretical abrasion rate and effective basal pressure for different rates of glacier movement. The actual abrasion rate will depend on the relative hardness of the glacier bed and the entrained subglacial debris. (Modified from G. S. Boulton (1974), in D. R. Coates (ed.)* Glacial Geomorphology. *State University of New York, Binghamton, Fig. 7, p. 52.)*

substantial force can be exerted. The process is probably most effective where the bedrock protrusion is much smaller than the transported boulder, and where the shear strength of the protrusion is low.

Joint-block removal involves the 'plucking' or 'quarrying' of large joint-separated blocks by an overriding glacier. Such bedrock joints may originate prior to glaciation through tectonic processes, pressure release or frost wedging. Pressure release, however, can develop as a consequence of rapid glacial erosion since bedrock is replaced by much lower density ice. Moreover, joint widening by frost wedging may possibly occur under warm-based glaciers with abundant meltwater. Subglacial meltwater is, of course, another potent erosional agent and is considered in Section 11.4.2.

**Table 11.3** Factors affecting the mechanisms of glacial abrasion

| FUNDAMENTAL FACTORS | COMMENTS |
|---|---|
| Presence of debris in basal ice | Clean ice is unable to abrade solid rock. The rate of abrasion will increase with debris concentration up to the point where effective basal sliding is retarded. |
| Sliding of basal ice | Ice frozen to bedrock cannot erode unless it already contains rock debris. The faster the rate of basal sliding the more debris passes a given point per unit time and the faster the rate of abrasion. |
| Movement of debris towards glacier base | Unless particles at the base of a glacier are constantly renewed they become polished and less effective abrasive agents. Thinning of the basal ice by melting or divergent flow around obstacles brings fresh particles down to the rock–ice interface and increases abrasion. |
| *Other factors affecting nature and rate of abrasion* | |
| Ice thickness | The greater the thickness of overlying ice the greater the vertical pressure exerted on particles on the glacier bed and the more effective is abrasion. This is the case up to a depth where friction between particles and the bed becomes so high that movement is significantly retarded and abrasion decreases. |
| Basal water pressure | The presence of water at the glacier base, especially when at high pressure, can reduce the effective pressure on particles on the bed and thus abrasion rates by buoying up the glacier. However, sliding velocities may tend to increase because of the reduced friction. |
| Relative hardness of debris particles and bedrock | The most effective abrasion occurs when hard rock particles in the glacier base pass over a soft bedrock. If the debris particles are soft in comparison with the bedrock the former are abraded and little bedrock erosion is accomplished. |
| Debris particle size and shape | Since particles embedded in ice exert a downward pressure proportional to their weight, large blocks should abrade more effectively than small particles. Moreover, angular debris will be a more efficient agent of abrasion than rounded particles. |
| Efficient removal of fine debris | To sustain high rates of abrasion fine particles need to be removed from the ice–rock interface since they abrade less effectively than larger particles (assuming the latter are continually being supplied from above). Meltwater appears to be the main mechanism for the removal of fine (<0.2 mm) debris. |

*Source*: Based on discussion in D. E. Sugden and B. S. John (1976) *Glaciers and Landscape*. Edward Arnold, London, pp. 153–5.

The thermal regime of a glacier exerts a pervasive influence over processes of glacial erosion. For instance, the potential for abrasion in cold ice is negligible because of the lack of basal slip and the absence of a significant load of basal debris. However, the greater adhesion of cold ice to bedrock would be expected to increase the effectiveness of joint-block removal. Abrasion is probably confined largely to warm-based glaciers since here the presence of basal water both promotes sliding and provides a means for removing fine debris. The fact that a single glacier may have both warm- and cold-based sections and that glaciers may alter from one type to the other in response to climatic change, makes patterns of glacial erosion potentially very complex.

### 11.2.2 Debris entrainment and transport

Effective glacial erosion can only continue if the eroded material is entrained and transported away by the ice flow (Fig. 11.11). The size of glacially transported debris ranges from very small rock fragments up to huge boulders. **Subglacial debris** is transported along the base of a glacier (Fig. 11.12), but glaciers also acquire **supraglacial debris** through material falling on to the ice surface from rock

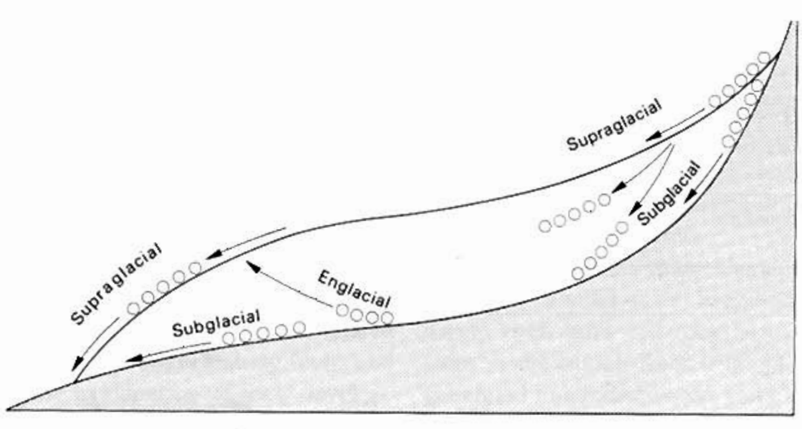


*Fig. 11.11 Schematic illustration of travel paths of debris transported by valley glaciers.*

*Fig. 11.12 Debris exposed at the base of Russell Glacier, Søndre Strømfjord, west Greenland. (Photo courtesy D. E. Sugden.)*

walls or other ice-free areas (Fig. 11.1). Naturally, supraglacial debris is likely to be most abundant in valley and cirque glaciers and absent over large areas of ice sheets. It can continue to be carried on the glacier surface within the ablation zone, but above the equilibrium line it will become progressively buried because of the accumulation of ice from above. Once buried it becomes **englacial debris** and it can travel as such to the glacier snout. Alternatively, it may emerge at the surface in the ablation zone through melting of overlying ice or it can move down to the glacier bed and be trapped by existing subglacial debris. Movement along slip lines in zones of compressive flow tends to transform subglacial into englacial debris, and especially in the lower section of the ablation zone, into supraglacial debris.

In some glaciers englacial debris is distributed fairly evenly with depth, but in others it is concentrated in bands separated by relatively clean ice. Such bands run parallel with flow lines and most are probably formed by the freezing of subglacial debris prior to its upward movement through the glacier. Under warm ice, basal melting produces a high concentration of subglacial debris which can be attached on to the base of the glacier by refreezing or **regelation**; but as the regelation layer is only a few centimetres thick this process can only account for thin bands. Creep provides another mechanism for the incorporation of subglacial debris into the base of a glacier, and in warm-based glaciers small amounts of material can probably be squeezed up into subglacial cavities. Finally, where a glacier is overriding a large obstruction on its bed, thrusts can develop in the ice which provide a further means of emplacing englacial debris.

### 11.2.3 Erosional landforms

Glacial erosion produces landforms at a wide range of scales, but irrespective of scale the overriding control is the attempt by glaciers to modify their beds in such a way as to improve the efficiency of ice flow. The extent of the ice cover, the thermal regime of the glacier, the amount of basal debris, the characteristics of the underlying bedrock, the period of time during which glacial erosion has been active and the form of the pre-existing topography all affect

**Table 11.4** Landforms and landscapes of glacial erosion

| PREDOMINANT PROCESS | ASSOCIATED MORPHOLOGY | LINEAR DIMENSIONS (0.01 m, 0.1 m, 1 m, 10 m, 100 m, 1 km, 10 km, 100 km, 1000 km, 10 000 km) | |
|---|---|---|---|
| Unconfined ice flow | Positive, streamlined | ◄— Whalebacks —► Streamlined spurs —►<br>◄— Rock drumlins —► | Landscape of areal scouring |
| | Positive, partially streamlined | ◄— Roches moutonnées — Flyggbergs —► | |
| | Negative, streamlined | ◄— Striations —► ◄— Grooves —► | |
| | Negative, partially streamlined | ◄— Rock basins —► | |
| Channelled ice flow | Negative, streamlined | ◄— Troughs — | Landscape of linear ice-sheet erosion |
| Interaction of glacial and periglacial processes | Negative | ◄— Cirques —► | Valley glacier landscape |
| | Positive | ◄— Arêtes —►<br>◄— Horns —► | |
| | | ◄— Nunataks —► | Nunatak landscape |

*Source*: Modified from D. E. Sugden and B. S. John (1976) *Glaciers and Landscape*. Edward Arnold, London, Table 9.2, p. 169.

the nature of glacially eroded landscapes. A useful distinction can be made between those forms associated with largely unconfined ice movement, those produced by channelled glacier flow and those created through the interaction of glacial and periglacial processes. This classification forms the basis for our discussion of glacially eroded landforms (Table 11.4).

*11.2.3.1 Forms associated with unconfined ice flow*

Unconfined ice flow is restricted to ice sheets and ice caps. Although the erosional activity below such ice masses is probably for the most part limited to regions of warm basal ice a range of positive landform features have been attributed to ice-sheet erosion. They include small, smoothly eroded forms a few hundred metres in length known as **whalebacks**, larger streamlined hills sometimes called **rock drumlins** and tapered interfluves and spurs up to several kilometres long. Although there are doubts about the interpretation of some of the larger landforms their alignment with known directions of ice movement and their smoothed surfaces support a glacial origin. There is certainly no doubt about the glacial nature of the asymmetric streamlined form known as a **roche moutonnée**, and its larger variant known as a **flyggberg**. Roches moutonnées range in size up to the dimensions of whalebacks. They have a smoothed end facing the direction of ice flow but a craggy, steeper lee side. Although it is apparent that they are formed by abrasion on their up-glacier side and joint-block removal on their lee side, it is not clear why their down-glacier side is not also smoothed by abrasion as is the case in fully streamlined forms. This may be related to the lower ice pressure on the lee side of a bedrock obstruction in an ice flow. The lower pressure would not only reduce abrasion but would also enable meltwater to migrate to the lee side and possibly aid joint enlargement by frost wedging. Low lee-side ice pressures would be favoured by thin or fast moving ice flowing around a prominent obstruction, since in these circumstances the ice would be less able to mould itself closely around the obstacle.

Negative relief forms attributed at least in part to the effects of unconfined ice flow include grooves aligned with a known direction of ice movement, and shallow **rock basins** often filled by a lake. Grooves up to 30 m deep, 100 m across and 12 km long occur in the Mackenzie Valley in northern Canada while rock basins, from several metres to a few hundred kilometres across, are more ubiquitous and are to be found over much of northern Canada. A glacial origin for rock basins is suggested not only by evidence of abrasion but also by their overdeepening, that is, their erosion to levels below regional base levels related to fluvial systems. Some basins are aligned along faults and major joint systems and one explanation of their development involves preferential joint-block removal where bedrock joints are more closely spaced. Abrasion also probably plays a role since it might be more effective where basal shear stresses are highest under thicker ice occurring over topographic depressions; in such a case an initially shallow depression might then be prone to overdeepening.

*11.2.3.2 Forms associated with channelled ice flow*

Ice flow concentrated in channels gives rise to steep-sided, **glacial troughs** (Figs 11.1, 11.2, 11.10). These may be formed by valley or outlet glaciers, or by ice streams occurring within ice sheets and ice caps. The walls of glacial troughs, which may exceed 1000 m in height, frequently truncate the tributary valleys and spurs of the pre-existing fluvially-eroded landscape thereby giving rise to **hanging valleys** (Fig. 11.13). The cross-profile of many glacial troughs approximates to a parabola and this form probably results from a combination of higher basal flow rates, and therefore more effective erosion, beneath the middle of the glacier and fluctuations in the height of the surface over time leading to less direct glacial erosion of the upper parts of the trough.

Three varieties of glacial trough can be distinguished. **Alpine troughs** are eroded by valley glaciers whose accumulation zones lie below a mountain mass. **Icelandic troughs**, by contrast, are formed by glaciers flowing from ice sheets or ice caps over the trough head and some enormous examples have been formed by outlet glaciers flowing from the Antarctic ice sheet. Where coastal Icelandic troughs are partially drowned by rising sea level they form **fjords**. A third type, **open troughs**, are so called because they are open at both ends. They frequently breach watersheds and most seem to have been eroded by ice streams within ice sheets or ice caps.

Alpine troughs characteristically have an overdeepened long profile; near the trough head the floor is steeply inclined, but down-valley there is a much lower or even slightly reversed gradient. Icelandic troughs typically have an excessive steepening towards the trough head and flatter trough floors, whereas open troughs normally have a high point roughly in the middle of their long profile. The location of overdeepening in alpine troughs is probably related to the position of the equilibrium line since here the ice is thickest and we would expect this to coincide with the zone of maximum erosion. The overdeepening of Icelandic troughs so near the trough head is more difficult to explain, but may be associated with a localized zone of warm-based ice which promotes more rapid erosion than the surrounding cold-based parts of the glacier.

Superimposed on these characteristic long profile forms are often smaller-scale irregularities consisting of alternating **rock bars** and rock basins which form a series of steps in the trough floor (Fig. 11.13). These features have generated much discussion and explanations have included more effective erosion where ice flow is constricted through a local narrowing of the trough or where tributary glaciers

*Fig. 11.13 Valley below the present location of the snout of Austerdalsbreen, southern Norway. Note the hanging valleys and truncated spurs, and the rock bar in the centre of the photograph.*

enter the main trough, and differential erosion controlled by variations in the spacing of joints in the underlying bedrock or the degree of pre-glacial weathering. If a trough floor contained small initial irregularities then any depressions would tend to be preferentially eroded, especially under warm ice, because of the higher basal shear stress associated with the greater thickness of overlying ice. Deepening of depressions in this way would cause further ice thickening and so irregularities would tend to be accentuated by this positive feedback mechanism. This provides an interesting contrast with fluvial systems in which adjustments between channel bedform and changes in flow depth and velocity are characterized by negative feedback and therefore promote the reduction of irregularities. In glacier flow the development of a stepped trough-floor profile is further aided by the existence of zones of compressive and extending flow since concavities in the glacier bed tend to be eroded to conform to slip lines.

*11.2.3.3 Forms associated with periglacial action*

**Cirques** (also called **corries** and **cwms**) are the major landforms developed by a combination of periglacial action (see Chapter 12) and glacial erosion and are usually located at the heads of deep valleys. A cirque consists of a bowl-shaped rock basin which extends from a steep **headwall** of shattered rock to a low rim. Fully developed cirques have a fairly consistent ratio of height to length suggesting that they are equilibrium forms which maintain their relative dimensions as they grow in size. They range from modest depressions a few hundred metres across, to massive amphitheatre-like forms several kilometres wide and with headwalls several hundreds of metres in height. Wind-blown snow is important in the mass balance of cirque glaciers, so it is not surprising that they are best developed where prevailing winds can bring in large quantities of snow, or at least where accumulated snow is not readily removed by the wind.

Cirque development starts with the formation of a firn bank in a suitable depression. This is gradually enlarged by a combination of processes known as nivation (see Section 12.2.4) involving active frost weathering and mass movement processes promoted by the presence of meltwater around the firn bank. Deepening proceeds until the firn turns to ice which then begins to flow. Because cirque glaciers have a rather unusual bed configuration their flow characteristics are rather different from other glacier types.

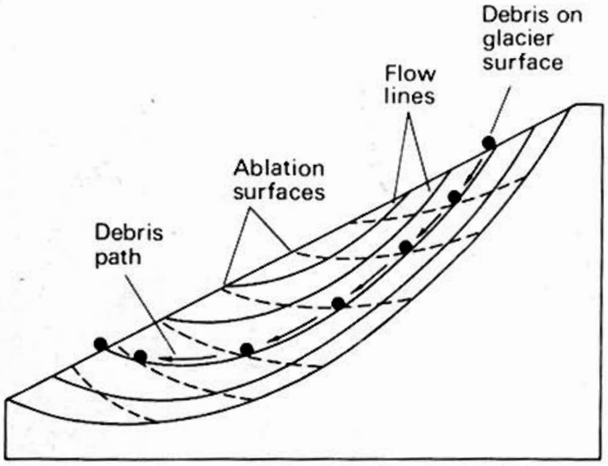


*Fig. 11.14 Schematic longitudinal section through a cirque glacier. Note the form of the flow lines and the path taken by debris which falls on to the glacier surface.*

The regular concave long profile form induces rotational sliding, and the flow lines are inclined away from the glacier surface near the headwall and towards the surface near the terminus. This is illustrated by the movement within the glacier of rock debris which falls on to the surface of the ice at the base of the headwall (Fig. 11.14). As cirque glaciers become larger it is likely that rotational slip becomes less important and internal deformation more so.

In order to maintain the rather constant ratio between height and length, deepening of the cirque floor, largely by glacial abrasion, must be accompanied by retreat of the headwall. This occurs mainly through the removal of large blocks of bedrock, but uncertainty surrounds the mechanism whereby these blocks are produced. It has long been observed that the top layer of actively moving ice on a cirque glacier is usually separated from the headwall by a deep chasm known as a **bergschrund**. The bottom of a bergschrund would be a likely collecting point for meltwater and, being exposed to the atmosphere, it might also experience frequent temperature fluctuations across freezing point and therefore be a site of active enlargement of pressure-release joints by frost weathering. This superficially plausible hypothesis unfortunately lacks supporting evidence since the little data collected do not indicate temperature changes which are either sufficiently frequent or rapid to promote intense frost weathering. A possible alternative mechanism for rock break-up is hydration shattering (see Section 6.3.1) although the effectiveness of this process in a wide range of rock types has yet to be firmly established.

Whatever the process of headwall erosion, cirque growth can consume large areas of upland. The early stages of cirque development focused along plateau margins create deeply notched plateau edges giving rise to so-called **biscuit board topography**. Cirque growth may eventually progress to the point where two encroaching cirque headwalls are only separated by a narrow ridge, known as an **arête**, or where several headwalls converge, by a sharp mountain peak or **horn** (Fig. 11.15). A summary of the sequence of features typically associated with the glaciation of mountain regions and created by the combined action of channelled ice flow and the interaction of glacial and periglacial processes is provided by Figure 11.16.

Periglacial processes also play an important role in the shaping of **nunataks**. These are mountain peaks of limited extent and entirely surrounded by glacial ice. They are most common where an ice sheet or ice cap almost entirely buries a mountain range with only the highest peaks remaining above the ice surface. After wasting of the ice there may be a significant contrast between the irregular, sharp-edged topography characteristic of nunataks and the somewhat smoother and more subdued forms of the surrounding glacially modified terrain.

## 11.3 Glacial deposition

Before we look at the processes of glacial deposition it is useful to draw a distinction between the deposits laid down by ice, known as **till** or **boulder clay**, and the landforms produced by such deposits, which are termed **moraine**. Although tills are highly variable deposits they usually have certain characteristics in common. These are listed in Table 11.5, but it must be emphasized that any one till deposit is unlikely to exhibit all these features; moreover, deposits originally deposited by ice may subsequently be reworked by glacial meltwater (see Section 11.4.1).

**Table 11.5** Characteristics of till

| | |
|---|---|
| 1. | Poor sorting – there is a large range in grain size, with large clasts up to boulder size frequently contained in a finer, sometimes clayey, matrix |
| 2. | Lack of stratification – laminations and graded bedding (progressive change in grain size with depth) are generally absent except in deposits modified by meltwater which may exhibit stratification |
| 3. | Mixture of lithologies – particles may have been derived from widely separated sources, especially in the case of deposits laid down by large ice sheets. |
| 4. | Frequent presence of particles with abraded facets and striations. |
| 5. | Preferred orientation of particles |
| 6. | Compaction associated with pressures developed during deposition |
| 7. | Overlies striated rock or sediment basement |
| 8. | Predominantly subangular particles due to a combination of fracturing and rounding by abrasion |

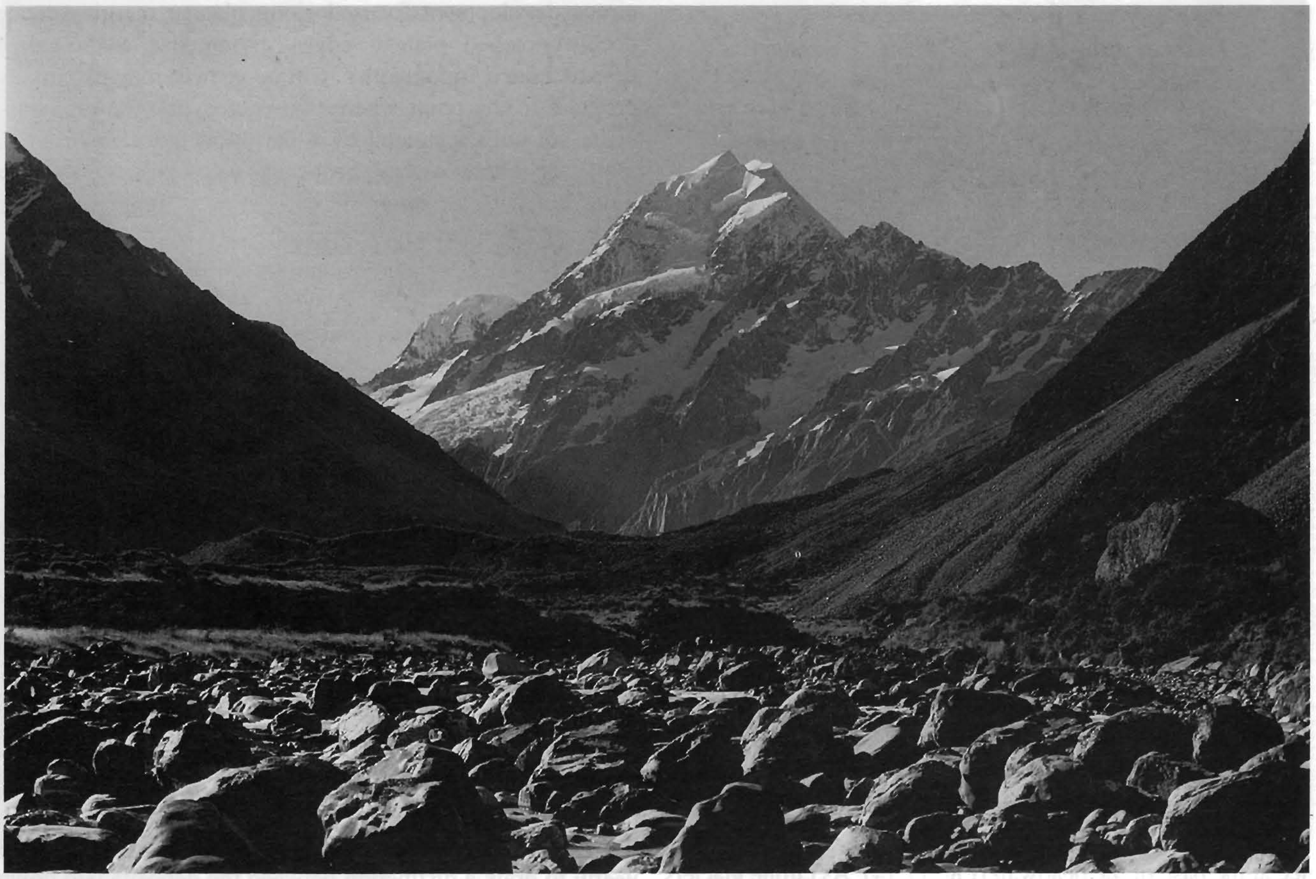

*Fig. 11.15 The horn of Mount Cook, the highest mountain in the Southern Alps, South Island, New Zealand. (Photo courtesy G. M. Robinson).*

### 11.3.1 Mechanisms of deposition

Glacial deposition involves a range of processes and any classification is somewhat arbitrary (Table 11.6). An added complication is that researchers have used different terms to describe the same process. Here we make a broad grouping of processes on the basis of whether they operate at the base, on the surface, or around the margins of a glacier. Nevertheless, material deposited in one situation may be subsequently re-entrained and re-deposited elsewhere (Fig. 11.17).

At least three distinct mechanisms of subglacial deposition can be identified. **Undermelt** involves the deposition of material through melting of the underlying ice. Melting may arise from geothermal heat, through frictional heating or as a result of increased pressures around the up-glacier side of obstructions. A second process, called **basal lodgement**, entails the smearing of predominantly fine material

**Table 11.6** Classification of mechanisms of glacial deposition

| LOCATION RELATIVE TO GLACIER | PROCESS | THERMAL AND DYNAMIC CONDITIONS |
|---|---|---|
| **Subglacial** | Undermelt | Warm-based only, active or stagnant |
| | Basal lodgement | Predominantly warm-based, active only |
| | Basal flowage | Warm-based only, active or stagnant |
| **Supraglacial** | Meltout | Active or stagnant |
| | Flowage | Active or stagnant |
| **Marginal** | Dumping | Active only |
| | Pushing | Warm- or cold-based, active only |

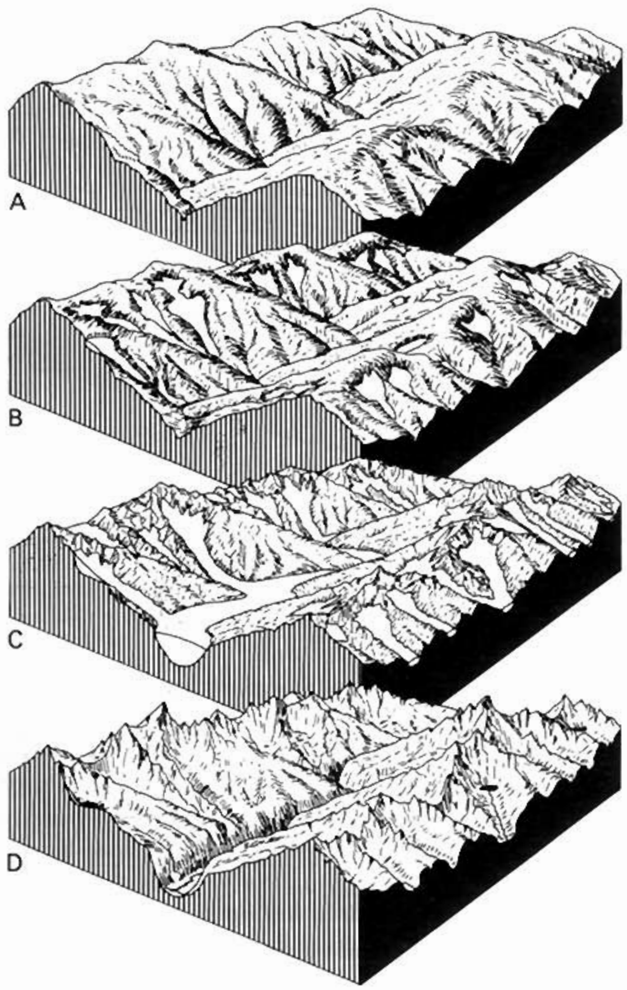


*Fig. 11.16 Development of landforms associated with alpine glaciation: (A) a fluvially eroded mountain landscape prior to glaciation; (B) the initial accumulation of snow and ice; (C) full development of a network of valley glaciers; (D) after deglaciation showing landforms of mountain glaciation including cirques, arêtes, horns, truncated spurs, hanging valleys and a glacial trough. (After R. F. Flint (1971)* Glacial and Quaternary Geology, *Wiley, New York, Fig. 6.3, p. 140.)*

on the glacier bed. Lodgement of subglacial debris occurs where the friction between a particle being transported by the ice and the glacier bed becomes so great that its further movement is retarded. This situation is favoured by thick ice promoting high basal pressures which increases friction on the glacier bed and by low flow velocities which are associated with weak basal shear stresses. **Basal flowage** is a third mechanism of subglacial deposition but it is also partly erosional. It involves both the squeezing of unconsolidated water soaked debris into basal ice concavities and the streamlining of till by overriding ice (see Section 11.3.2.1).

Supraglacial debris deposition can occur by **meltout** or by flowage. Meltout involves deposition of sediment through melting of the glacier surface. It is most active in the snout of warm glaciers where up to 20 m of ablation may occur in a single summer. **Supraglacial flowage**, involving the movement down the ice surface of the debris-rich upper layers of a glacier, has also been observed and is particularly common near the snout where intense supraglacial meltout occurs. The movement ranges from slow creep to a rapid liquid flow.

Deposition around a glacier margin can arise from a variety of mechanisms. Water-soaked till can be squeezed from under the ice, and if the glacier margin remains stable for some time significant accumulations of supraglacial and englacial debris can form through dumping of material by meltout. Existing glacial deposits can also be pushed by an actively advancing glacier and deposited when the advance stops, or be overridden and recycled towards the snout as englacial and supraglacial debris. Material deposited at a glacier margin commonly suffers extreme disruption prior to deposition so that any original orientation of particles in the ice is usually lost, except possibly where a frozen mass of till is bulldozed *en masse*.

### 11.3.2 Depositional forms

Most ice-laid deposits have a definite three-dimensional shape, that is, they form moraines. Few moraines are composed entirely of till – they also contain stratified sediments deposited by meltwater and, in some cases, even a bedrock core. Another characteristic of many moraines is their transience. Those occurring around the margins of a glacier often have an ice core and they collapse into insignificant forms when this melts. Moraines can also be destroyed or drastically modified by meltwater and, in the case of those formed subglacially, can be modified by glacial activity at the ice margin when the glacier retreats.

Perhaps the most helpful way to classify such complex landforms as moraines is to categorize them in terms of their relationship to the direction of ice flow. On this basis there are three major types: those aligned largely parallel to the direction of ice flow, those orientated roughly transverse to the direction of ice flow and those lacking any consistent orientation (Table 11.7).

#### *11.3.2.1 Forms parallel to the direction of ice flow*

Moraines orientated roughly parallel with the direction of ice flow can be formed subglacially, supraglacially, or at the ice margin. Most of the subglacially produced forms seem to result from streamlining related to variations in stress across the glacier bed. Often formless sheets of till, known as **ground moraine**, can be modified subglacially by the ice flow into fluted forms up to 10 m high and 1 km long, or exceptionally into megaflutes up to 25 m high and 20 km long. That they can be either predominantly depositional or erosional features is highlighted by the distinction made between **fluted ground moraine**, in which the fluting is

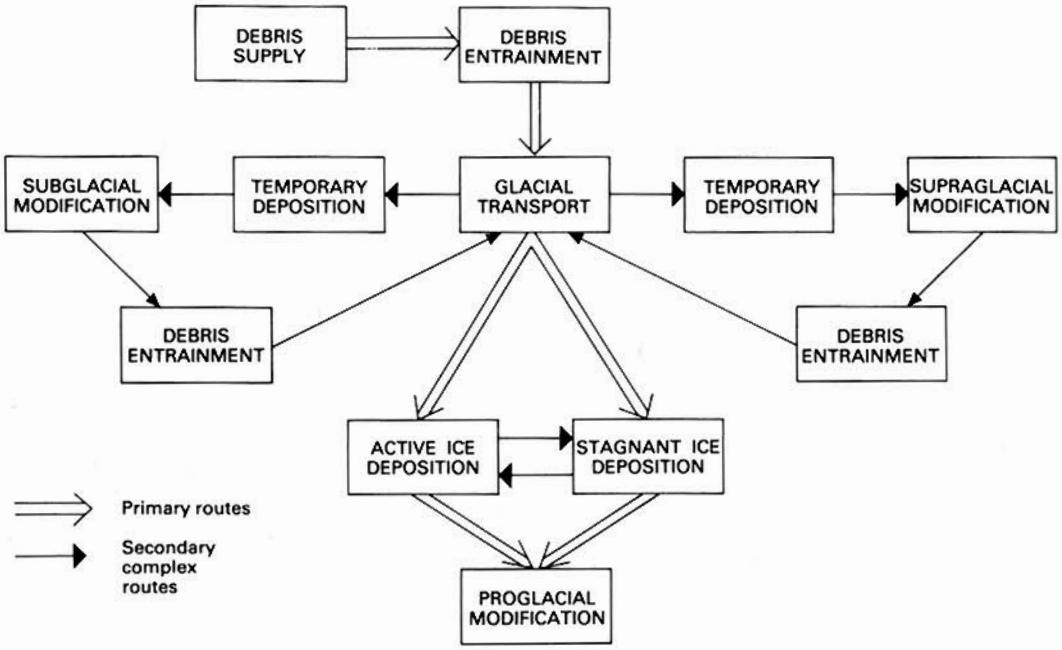


***Fig. 11.17*** *Schematic representation of the possible routes that debris may follow in the glacial and fluvioglacial system prior to final deposition.*

**Table 11.7** Classification of the major types of moraine

| PARALLEL TO ICE FLOW | TRANSVERSE TO ICE FLOW | LACKING CONSISTENT ORIENTATION |
|---|---|---|
| *Subglacial forms with streamlining*<br>Fluted and drumlinized ground moraine | *Subglacial forms*<br>Rogen or ribbed moraine | *Subglacial forms*<br>Low-relief ground moraine |
| Drumlins and drumlinoid ridges | De Geer or washboard moraine | Hummocky ground moraine |
| Crag-and-tail ridges | Subglacial thrust moraine | |
| | Sublacustrine moraine | |
| *Ice-pressed forms*<br>Longitudinal squeezed ridges | *Ice-pressed forms*<br>Minor transverse squeezed ridges | *Ice-pressed forms*<br>Random or rectilinear squeezed ridges |
| *Ice marginal forms*<br>Lateral and medial moraines | *Ice front forms*<br>End moraines | *Ice surface forms*<br>Disintegration moraines |
| Some interlobate and kame moraines | Push moraines | |
| | Ice thrust/shear moraines | |
| | Some kame and delta moraines | |

*Source*: Modified from D. E. Sugden and B. S. John (1976) *Glaciers and Landscape*. Edward Arnold, London, Table 12.1, p. 236, after V. K. Prest (1968) *Geological Survey Papers Canada* 67–57.

due to troughs within the till, and **drumlinized ground moraine** where ridges rise above the general level of the till surface and more clearly represent constructional relief forms. There are numerous theories for the formation of fluted moraine, but several point to the role of large boulders on the glacier bed (Fig. 11.18). Constructional flutings could be formed by unfrozen till being forced into cavities on the up- and down-glacier sides of an obstacle.

The most intensively studied subglacial moraine form is the **drumlin**. Although morphologically somewhat variable, in plan drumlins tend to be bluntly rounded on their up-glacier margin and sharply pointed on their lee side, while in profile they are usually highest towards their up-glacier end. Typical dimensions are 1–2 km in length, 5–50 m in height and around 500 m in width. While some drumlins appear to be composed entirely of clay-rich till, others have a bedrock core. They rarely occur singly, much more commonly forming **drumlin fields** in which individual drumlins are randomly spaced.

Drumlins remain difficult landforms to explain. The

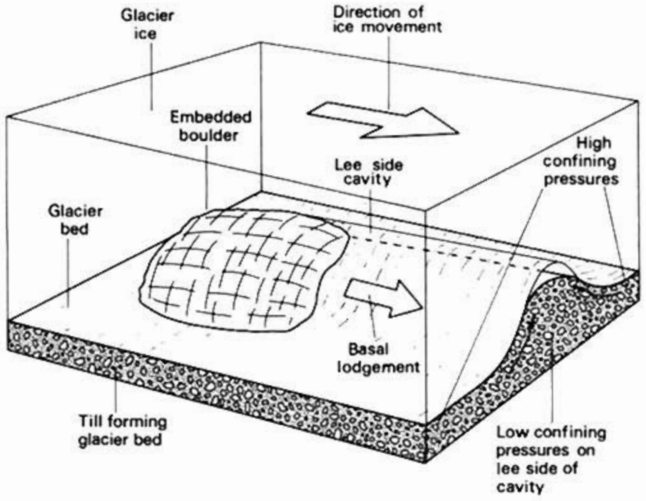


*Fig. 11.18 A probable model of fluted moraine formation which may be applicable to most types of fluted moraine ridges. (Based on D. E. Sugden and B. S. John, (1976)* Glaciers and Landscape. *Edward Arnold, London, Fig. 12.2, p. 239.)*

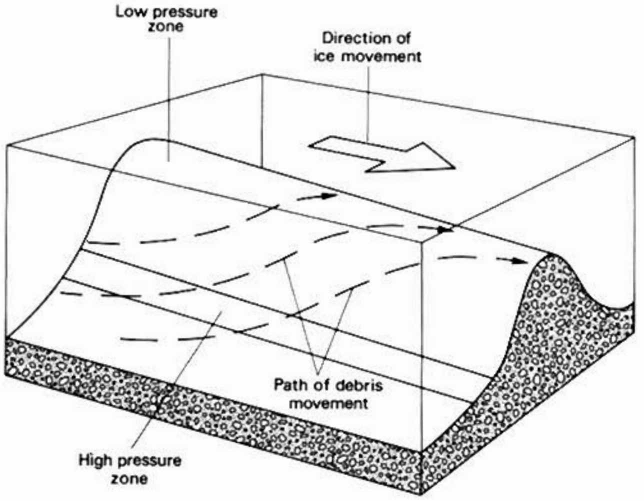


*Fig. 11.19 Schematic representation of migration of debris on the flanks of a growing drumlin. (Based on D. E. Sugden and B. S. John (1976)* Glaciers and Landscape. *(Edward Arnold, London) Fig. 12.4, p.241.)*

existence of a bedrock core in some examples suggests that they may form around obstructions. But this is at best an incomplete explanation since it does not account for those forms lacking a bedrock core. The concentration of drumlins in specific areas suggests that the initial accumulation of material is influenced by local conditions. Probably the most significant are variations in bed roughness with subglacial debris being preferentially deposited where the bed is irregular. Bedrock protrusions or large boulders will encourage debris to collect on their up-glacier side while deposition will also occur on the lee side where shear stresses are less. An initial stage of this kind of deposition is represented by **crag-and-tail** forms which consist of a tail of glacial deposits in the lee of a rock obstruction. The drumlin form appears to develop through variations in basal pressure associated with ice flow over an initial irregularity (Fig. 11.19). Drumlin crests would be zones of low pressure while the troughs between would experience high pressures. Particles being transported over the drumlin by the ice flow would tend to move across the gradient of decreasing basal ice pressure and be deposited on the crest. This kind of mechanism accords with the observation that in many drumlin fields there is only a very sparse till cover between drumlins.

Variations in the mechanical properties of till may also be important in drumlin formation, with subglacial deposition occurring where till is more resistant to deformation. This might take place where subglacial meltwater is efficiently drained from till which overlies a permeable bedrock. Another explanation focuses on the role of dilatancy (see Section 7.1.1). Most tills have the property of dilatancy; that is, in very simplified terms, they expand when subject to stress within certain limits. It is envisaged that debris on a glacier bed will resist movement until the stress builds up to a sufficiently high level to induce the dilatant characteristic. Once this level is reached the till will begin to deform readily and will continue to deform, even when the stress is decreased, until a lower critical threshold is reached and the dilatant property is lost. Imagine debris below thick ice. Here subglacial debris is moving freely in its dilatant state since the basal stress is high. As the ice thins towards the glacier margin the basal stress will gradually decrease until the critical lower limit for dilatancy is reached. The basal debris will now suddenly revert to a compact form and resist further movement by the ice. As it is likely that there will be variations in stress across the glacier bed, zones of compact, stationary till will be surrounded by dilatant material which continues to be transported. The zones of compact till will then start to be moulded by the still-moving, debris-rich basal ice into a streamlined form offering least resistance to ice flow.

Although widely regarded as ice-moulded forms, it has also been suggested that some drumlins, at least, might be formed by subglacial meltwater. One idea is that drumlins can be formed when sedimentation occurs in cavities eroded by meltwater flow directed upwards towards the base of the ice. Another possibility is that drumlins can be remnant erosional ridges created by meltwater flow. These kinds of drumlins are composed of pre-existing bedrock or sediment left protruding as the surrounding material is removed by meltwater erosion.

Moraines deposited on an ice surface or along a glacier margin have a far lower chance of survival than forms of subglacial origin. **Medial moraines** are formed from supraglacial debris concentrated in a thin ribbon in mid-glacier below the confluence of two tributary glaciers (Fig. 11.1).

They can be significant features in the upper part of a glacier but they usually only remain as a thin cover of till near the snout. **Lateral moraines** are formed primarily from frost-shattered debris which has fallen on to the edge of a glacier from the adjacent rock walls and are more likely to survive since they may lie partly on a bedrock substrate. Vertical fluctuations in the level of the glacier can lead to lateral moraines being perched high above the ice surface.

*11.3.2.2 Forms transverse to the direction of ice flow*

Perhaps the most obvious place to expect moraines aligned perpendicular to the direction of ice movement is along the front of a glacier. **End moraines** form in this position and they mark the maximum extent of a glacier margin. Those associated with continental ice sheets may be up to 100 m high and extend for tens of kilometres. A more or less parallel series of end moraines can often be found representing stages of ice retreat. End moraines are usually composed of both supraglacially and englacially transported debris together with fluvioglacial and lacustrine sediments, and are formed by a combination of dumping, meltout and flowage. Debris brought to the ice surface along flow lines and shear planes developed near the glacier snout can be dumped by sliding off ice-cored ridges on the glacier surface or by collapse due to melting of underlying ice. The rate of melting can significantly affect the detailed form of the moraine. Where melting is rapid and meltwater abundant supraglacial till can flow down even shallow slopes and rather featureless accumulations of till are produced. Where melting occurs slowly debris can collect in troughs lying between ice-cored ridges. As melting progresses these ridges collapse and the originally debris-filled troughs become upstanding ridges as the topography is inverted. Small end moraines, less than 10 m high, can also be formed by the squeezing of water-soaked debris from beneath the ice front or into subglacial crevasses. Moraines developed by this processes of basal flowage may be produced seasonally when summer meltwater percolates to the bed of a warm-based glacier and saturates the underlying till.

**Push moraines** are another form constructed along an ice front. They are produced by the bulldozing of glacial, and even non-glacial, deposits by an advancing ice margin and can attain considerable dimensions – up to 100 m high and 30 km long. The pushing action of the ice can create complex thrust and fold structures in the till and typically results in a distal slope which is much steeper than the slope facing the glacier front.

Ice-margin processes are not the only mechanisms capable of generating transverse moraines. Transverse ridges up to a few metres in height, in many cases associated with lakes or former lakes, are known as **washboard moraines**, or **De Geer moraines**. Some examples may be subaqueous push moraines, but others may owe their origin to subglacial thrusting and therefore be allied to **ribbed moraines** or **Rogen moraines**. These consist of fairly regularly spaced ridges, usually arcuate in plan and concave up-glacier, typically 10 m or more in height and up to 1 km long. Many, though not all, have streamlined 'drumlinized' crests. One interpretation is that they form through preferential deposition where shear stresses transverse to the direction of flow are relieved, and experience subsequent streamlining by the ice (Fig. 11.20). Another explanation involves a genetic link between Rogen moraine and fluting, the development of the former being seen as a response to variations in basal stress, or the presence of topographic irregularities, across the glacier bed.

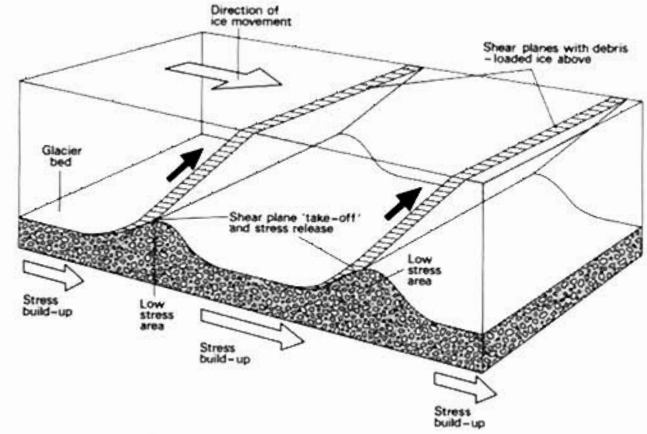


*Fig. 11.20* *A model of the formation of ribbed (Rogen) moraine. Debris accumulates at the base of shear planes in zones of compressive ice flow. In addition to particle movement along pressure gradients, the process of formation may also involve pressure melting and lodgement where the shear planes 'take off' from the glacier bed. (Based on D. E. Sugden and B. S. John (1976)* Glaciers and Landscape. *Edward Arnold, London, Fig. 12.8, p.245.)*

*11.3.2.3 Forms lacking consistent alignment*

Some moraines have no obvious orientation or linear development and therefore fail to fit into either of the morphological categories discussed above. Ground moraine may occur as virtually featureless sheets of till, but it can also form a gently undulating topography or even a chaotic hummocky terrain with a relief in excess of 100 m. This may develop from debris which has accumulated in basal ice concavities or from the meltout of supraglacial debris spread extensively over an area of wasting ice. On the margins of a retreating ice sheet large masses of ice can become stagnant through being isolated from a source of active ice flow. Such stagnant ice will gradually be reduced by ablation to discrete masses of decaying debris-rich ice and eventually a chaotic assemblage of disintegrating moraines will be created forming **dead-ice topography**.

## 11.4 Fluvioglacial erosion and deposition

### 11.4.1 Glacial meltwater

**Meltwater** is an integral part of the glacial system (Fig. 11.21). It both significantly influences glacier behaviour through its effect on rates of basal sliding and has an erosional and depositional role associated with, but distinct from, that of glacier ice. On most glaciers surface melting is by far the most important source of meltwater, the amount released increasing down through the ablation zone. In temperate regions this surface source may be supplemented seasonally by rainfall and, in valley glaciers, by runoff from valley-side slopes. The chief basal and internal sources of meltwater are basal melting through geothermal and frictional heating, although ground water can be a significant additional source in temperate valley glaciers. Geothermal heat is capable of melting about a 6 mm layer, and heat generated by basal sliding and internal deformation a 10–15 mm layer, of warm ice annually. Further basal melting can arise from the heat brought down by surface

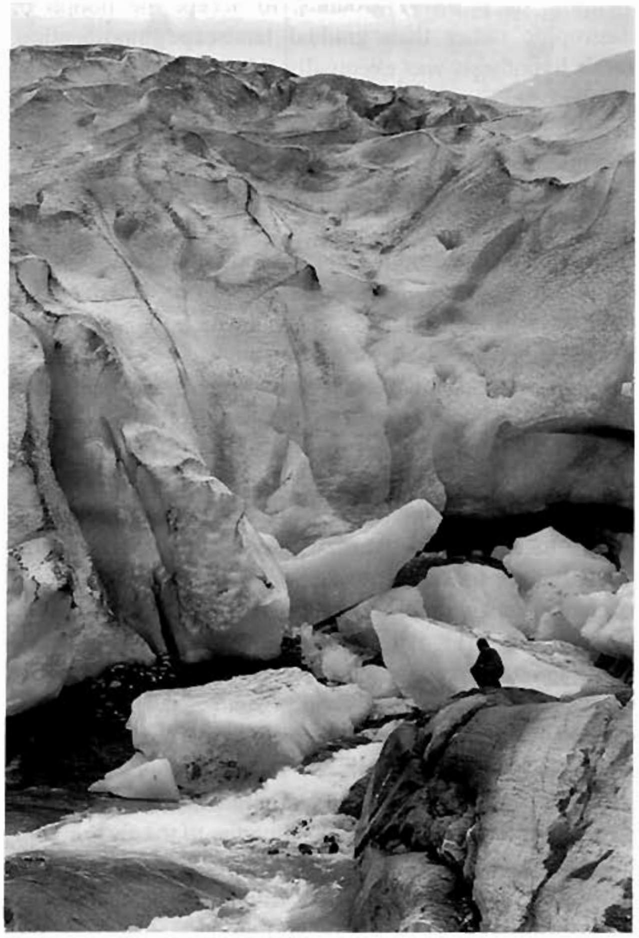

*Fig. 11.21 Meltwater emerging from the snout of Nigardsbreen, a valley glacier in southern Norway.*

water and by the frictional heat generated by the flow of the meltwater itself.

Depending on the thermal regime of the glacier, meltwater may travel over the glacier surface or move englacially and subglacially (Fig. 11.22). Surface runoff may be concentrated into stream channels up to several metres deep, in some cases sunk into 'valleys' on the ice surface. Extensive drainage networks covering hundreds of square kilometres can develop on ice sheets. Stream courses may be influenced by ice surface irregularities, but very regular meandering patterns have also been reported. Internal drainage within a glacier evolves in a manner somewhat analogous to the formation of underground drainage in highly soluble rocks such as limestones. Initially, water permeates warm ice along the boundaries separating individual ice crystals, but eventually a secondary permeability develops with the creation of surface sinkholes, known as **moulins**, and tunnels which tend to form preferentially along cracks and cavities in the ice. Glacier tunnels form dendritic networks both within, and at the base of, the ice, their direction and gradient being controlled both by the general slope of the glacier surface and the subglacial topography. In addition to basal channels, water can collect to form large subglacial lakes. It is suspected that these may be quite extensive below large ice sheets.

Rates of meltwater flow from glaciers vary over a range of time scales. Short-term fluctuations occur in response to diurnal variations in ablation and changing meteorological conditions. Seasonally the lowest flows usually occur in late winter and the highest in early summer – how early depending on the rapidity with which the internal drainage channels are re-established after winter freezing and closure by ice flow. Significant longer-term maxima in meltwater discharge are associated with glacier surges, but the most spectacular peak flows are caused by the catastrophic draining of ice-dammed lakes. Discharges several orders of magnitude greater than normal are attained in a few hours and diminish again even more quickly. Such jökulhlaups are particularly common in Iceland where some are associated with volcanism (see Section 5.2.3.4). One originating from the western part of the Vatnajökull ice cap with a periodicity of about 10 a has reached an estimated peak discharge of 40 000–50 000 $m^3s^{-1}$, while another in Baffin Island, Canada, involved the draining of a $5 \times 10^6 m^3$ subglacial lake in only 30 h. Such massive discharges could arise in various ways, but one possibility is that a gradually extending internal drainage system eventually taps an ice-dammed, subglacial lake.

Even more cataclysmic jökulhlaups were apparently associated with the draining of Late Pleistocene Lake Missoula in the north-west USA (Fig. 11.23). This **proglacial lake**, which extended across a large area of western Montana, was the largest of a number in the region impounded by ice advancing from the north. It is estimated that at its maxi-

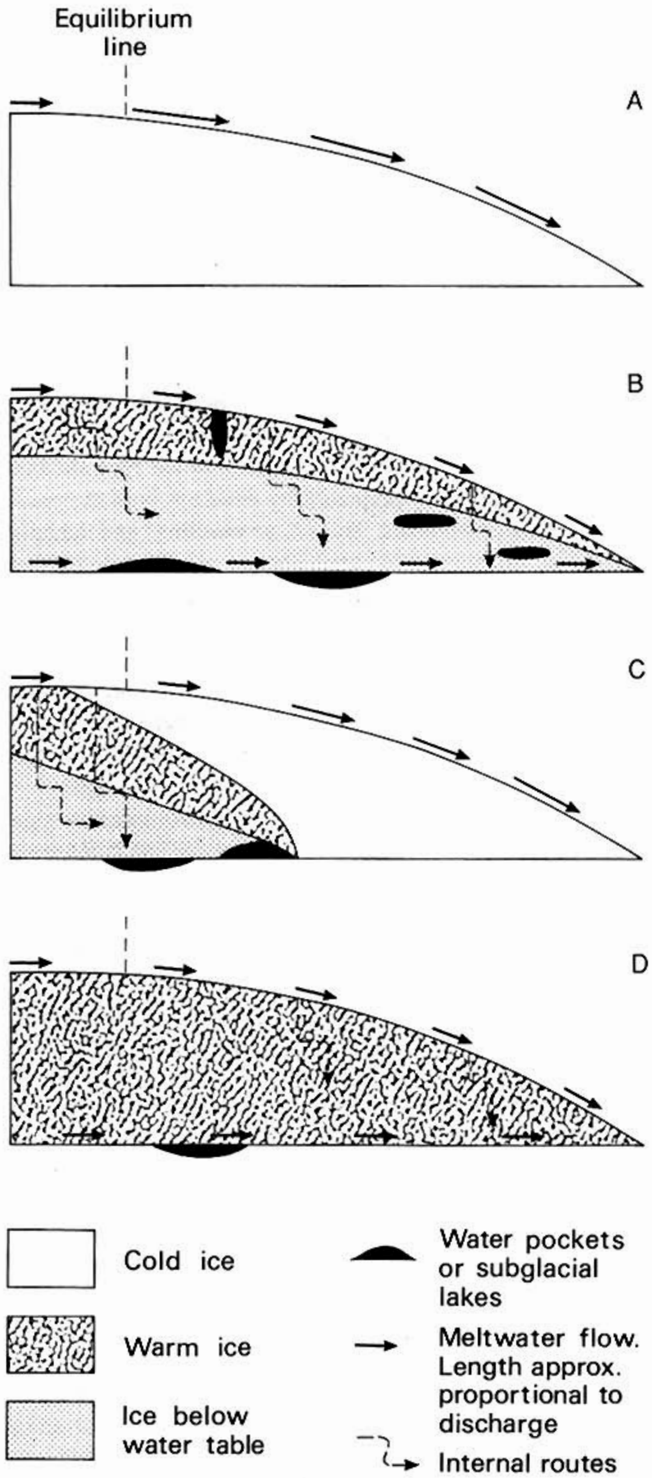


*Fig. 11.22 Possible meltwater routes in different types of glaciers and glacial ice: (A) cold ice; (B) warm ice; (C) cold ice in the ablation zone and warm ice in the accumulation zone; (D) a high altitude glacier in equatorial latitudes. (After D. E. Sugden and B. S. John (1976)* Glaciers and Landscape. *Edward Arnold, London, Fig. 14.18, p. 298.)*

mum Lake Missoula reached a volume of 2000 $km^3$, and much of this water was released during repeated failures of its ice dam. The peak discharge attained during break-out events is thought to have been a staggering $21.3 \times 10^6\ m^3\ s^{-1}$, or approximately 20 times the mean discharge of *all* the world's rivers! Several separate flood events occurred between 16 and 12 ka BP which have been reconstructed from the extraordinary landforms they produced. These include a series of enormous streamless canyons, or **coulees**, which form an anastomosing network and which contain giant 'ripples', plunge basins, deep rock basins and vast potholes (Fig. 11.24).

This whole region of anomalous topography is called the Channeled Scabland and its origin was the subject of intense debate after a catastrophic flooding hypothesis was first put forward by J. Harlen Bretz in a paper published in 1923. This explanation was initially rejected by his contemporaries, both because Bretz had no convincing source for such enormous volumes of water (he was unaware of literature documenting the existence of Lake Missoula published in the 1880s), and because they were unwilling, on *a priori* grounds, to accept the notion of catastrophic rather than gradual landscape modification. Bretz's hypothesis was eventually substantiated by his own persistent field work and by other workers who made the link between catastrophic flooding and Lake Missoula. Subsequent work has begun to relate the specific dimensions of the landforms present to flow conditions. Of particular interest are the giant current ripples found in the channels. These are composed of gravel and are typically around 5 m high and have a wavelength of 100 m or so. From these dimensions it is possible to infer mean flow velocities, bed shear stresses and stream power values. It seems that the floods that formed these ripples could transport boulders as large as 10 m across.

### 11.4.2 Fluvioglacial denudation

The sediment-charged nature of glacial meltwater streams is testament to their transportational role and suggestive of their erosional capabilities. Although data on sediment yields are scarce it is known that the suspended loads of meltwater channels may attain concentrations in excess of 3000 mg $l^{-1}$. Their characteristic greyish-white colour is due to an abundance of the fine sediment generated by glacial abrasion, but up to 25 per cent or more of the total load may be carried as bed load. Peak yields are usually reached in the early part of the summer presumably due to the flushing out of sediment generated during the preceding winter. The limited information available indicates that meltwater streams also probably carry a significant solute load. The high surface area of unweathered rock provided by fine suspended sediment and the highly turbulent flows in which it is transported might be expected to promote active chemical

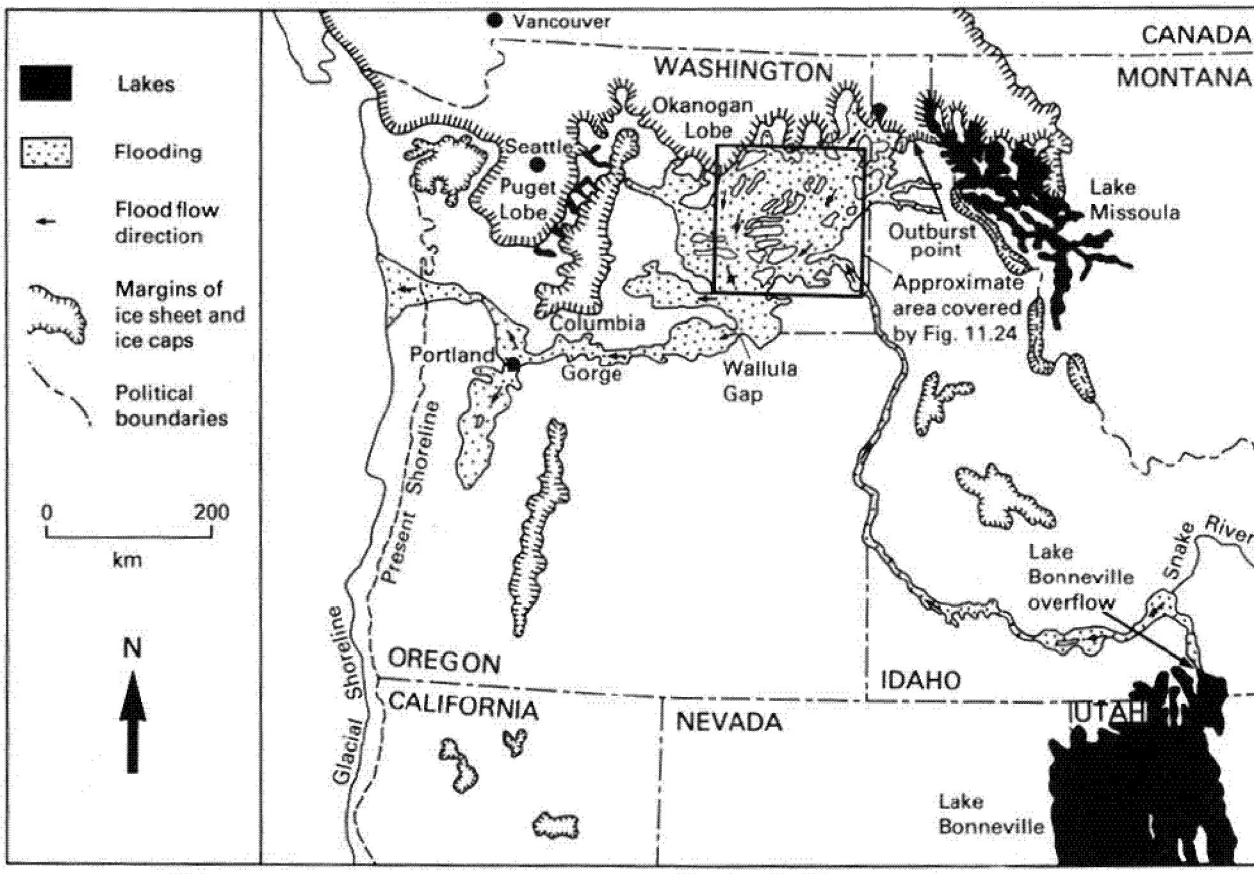


*Fig. 11.23 Area of the north-west USA affected by catastrophic flooding at the end of the last glacial. In addition to catastrophic outbursts from Lake Missoula, flooding also occurred as a result of overspills from Lake Bonneville in Utah and from other ice-dammed lakes in the Columbia Basin. (Modified from V. R. Baker and R. C. Bunker (1985)* Quaternary Science Reviews *4, Fig. 1, p. 2.)*

weathering. Much chemical weathering undoubtedly occurs subglacially and this may indirectly increase erosion rates by weakening the bedrock. During winter low flows the solute load may even exceed the solid load.

High flow velocities and especially high peak discharges suggest that meltwater streams should be potent erosional agents. Although fine rock flour usually represents the major component of the suspended load it probably has a negligible erosional effect. The coarser suspended and saltating load, however, is apparently a highly effective abrasional agent; indeed there is observational evidence of the cutting of meltwater streams into resistant bedrock in just a few years. Since high flow velocities in the range 8–15 m s$^{-1}$ are probably quite common in meltwater streams, and because rough channel beds will be the norm, cavitation may also be an important erosional process especially where stream channels narrow.

Various minor erosional forms consisting of a range of smooth rock depressions have been attributed to the action of meltwater, although for some of these alternative explanations involving glacial erosion have been proposed. Collectively they are known as plastically sculptured forms, or simply as **p-forms**. Elongated varieties exhibiting striations are certainly likely to be due to glacial abrasion but the origin of another type, known as **sichelwannen**, which occur in resistant crystalline rocks, is less certain (Fig. 11.25). They consist of crescentic depressions up to 5 m or more across and a fluvioglacial origin is suggested by the formation by differential fluvial erosion of similar features in less resistant rock.

The major large-scale erosional fluvioglacial landform is the **meltwater channel**. These can be up to 100 m or more in depth and extend for tens of kilometres. They take various forms but can be broadly grouped into subglacial ice-directed forms, where active ice movement has exerted a major control, and marginal and submarginal forms which tend to run parallel to the glacier margin. Ice-directed channels are generally aligned parallel to the direction of ice movement. Some breach pre-glacial drainage divides while others run downslope parallel with divide crests. They occur singly or as bifurcating or anastomosing networks. A subglacial origin for such channels is supported by their up-and-down long profiles, which are difficult to explain by anything other than subglacial water flow under hydrostatic pressure, and

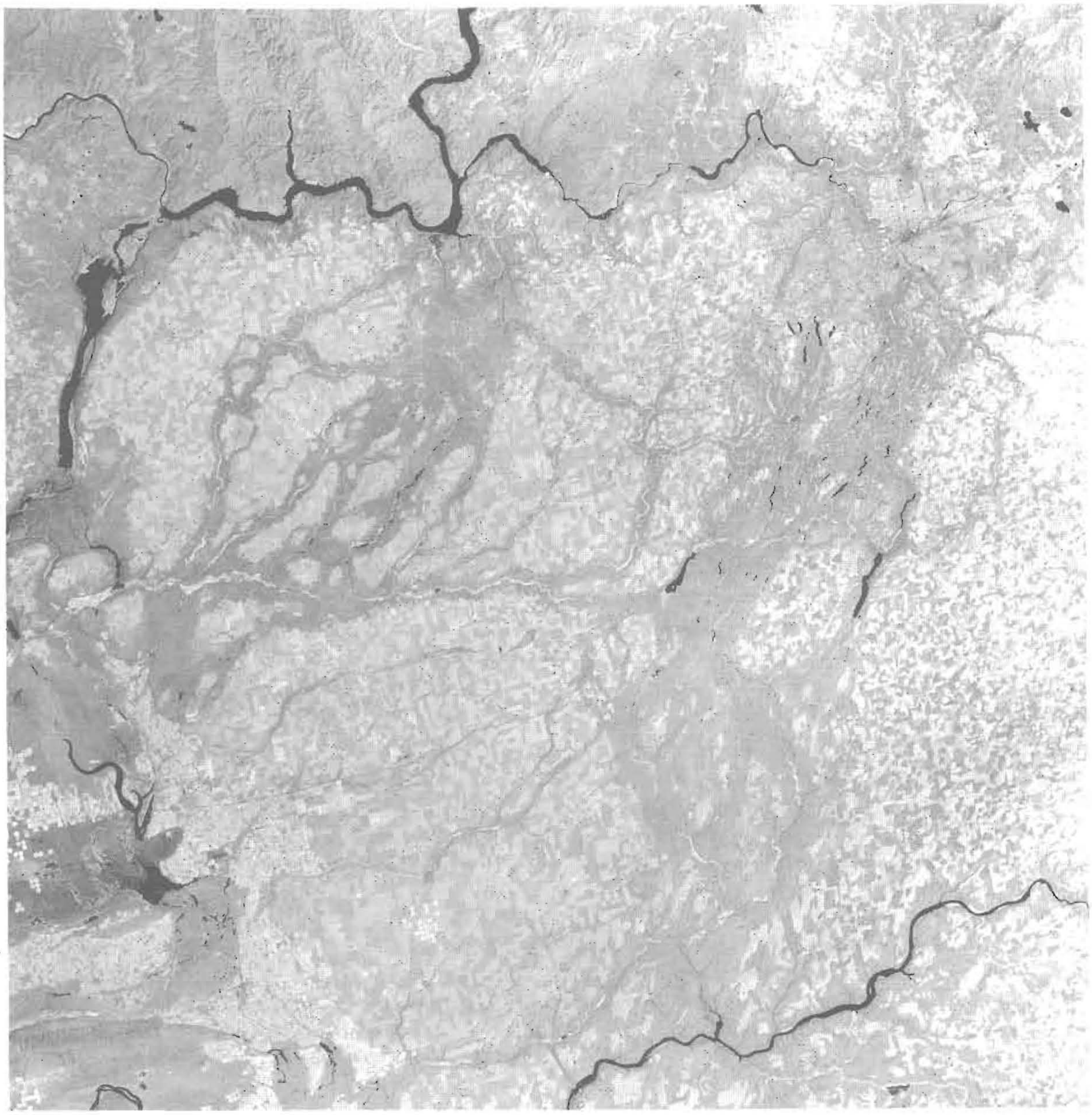

*Fig. 11.24 Landsat image of the Channeled Scabland, Washington, USA. The vast anastomosing channel patterns are clearly visible because they were cut through a cover of loess, thereby exposing the dark-coloured underlying basalt of the Columbia Plateau. The area covered is about 180 km across. (Image courtesy N. M. Short.)*

by the presence of ice moulding on the upper slopes of some examples.

The distinction between subglacial and marginal channels is not always clear since the latter can range from channels cut wholly in lateral moraine or bedrock to submarginal channels located subglacially at the extreme edge of the glacier. The marginal and submarginal environment is certainly a likely focus for meltwater generated in the zone of very active ablation of a valley glacier close to the trough wall, and for non-glacial runoff from adjacent valley slopes. While submarginal channels are more likely to form in association with warm ice, true marginal channels will develop along the margins of cold glaciers since here there is little opportunity for meltwater to percolate along the frozen ice–rock interface.

### 11.4.3 Fluvioglacial deposition

Sediment transported by meltwater can be deposited either in contact with the glacier or beyond the ice front in the

*Fig. 11.25* *Crescentic scour marks, or sichelwannen, Søndre Strømfjord, west Greenland. These features are probably formed as a result of erosion by high-velocity, sediment-charged subglacial meltwater flows. (Photo courtesy D. E. Sugden.)*

proglacial environment. This distinction provides a convenient basis for classifying fluvioglacial landforms (Table 11.8). Wide temporal variations in meltwater discharge and the potential for reincorporation within a glacier of fluvioglacially deposited debris by refreezing means that many ice-contact forms are transient features, at least in active ice. Proglacial features are much more likely to survive especially if the ice front is receding.

The basic mechanisms of deposition from meltwater are the same as those for ordinary fluvial systems, except that deposition within a glacier may occur under high hydrostatic pressures if meltwater completely fills an englacial or subglacial tunnel. There are some important differences, however, in the environment of deposition; for instance, tunnels within ice are not usually free to migrate laterally as fluvial channels may do and the rapid temporal fluctuations in meltwater discharge can produce equally rapid vertical changes in the calibre of sediment deposited which may range from large boulders to sand. The key characteristics which distinguish water-laid fluvioglacial sediments from glacial deposits is the absence of fine particles due to sorting and the presence of stratification. Meltwater deposits may be less easy to differentiate from ordinary fluvial sediments, although the glacial sediment source and the typically short distance of transport before deposition means that the former are usually less rounded.

Meltwater can deposit sediment on, in, under and along the margins of glaciers (Fig. 11.26). Although most of the ice-contact landforms to which this deposition gives rise are formed in channels, some may also be constructed either subglacially or marginally in lakes. The terminology used to describe these features is extremely confused. Here we

**Table 11.8** A classification of fluvioglacial deposits

| DOMINANT SEDIMENT | ENVIRONMENT | GENERAL FORM | RELATIONSHIP TO ICE | GENETIC TERM |
|---|---|---|---|---|
| *Ice-contact deposits* | | | | |
| Sand and gravel | Fluvial | Ridge | Marginal, subglacial, englacial, supraglacial | **Esker** |
| | | Mound | | **Kame** **Kame complex** |
| | | Spread with depressions | Marginal | **Kettled sandur** |
| *Proglacial deposits* | | | | |
| Sand and gravel | Fluvial | Spread | Proglacial | **Sandur** |
| Silt and clay | Lacustrine | | Proglacial/ marginal | **Lake plain** |
| Sand and gravel | | Terraces, ridges | | **Beach** |
| Clay sand and gravel | | Terrace | | **Kame delta** |
| Silt and clay | Marine | Spread | | **Raised mud flat** |
| Sand and gravel | | Terraces, ridges | | **Raised beach** |
| Clay, sand and gravel | | Terrace | | **Raised delta** |

*Source*: After R. J. Price (1973) *Glacial and Fluvioglacial Landforms*. Oliver and Boyd, Edinburgh, Table 3, p. 138.

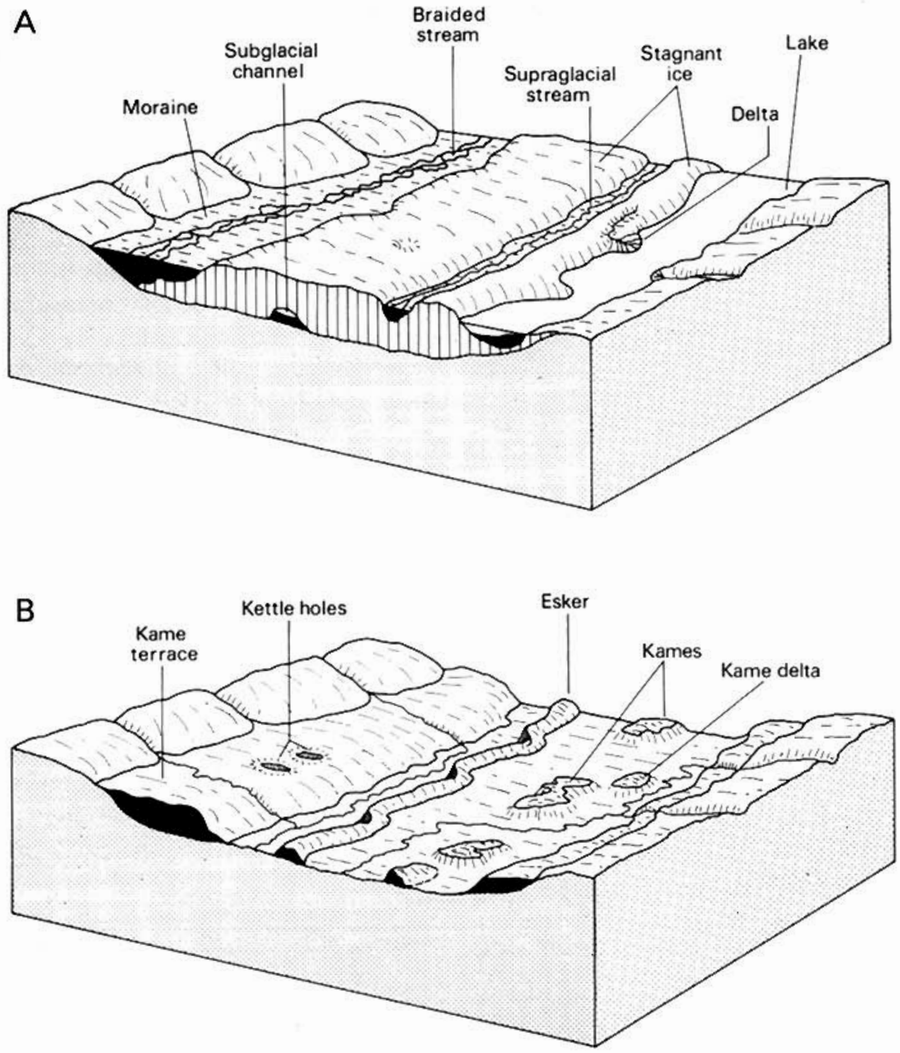


*Fig. 11.26 Development of ice-contact fluvioglacial landforms and deposits: (A) late stage of deglaciation with extensive areas of stagnant ice and abundant meltwater; (B) after deglaciation. (Based on R. F. Flint, (1971)* Glacial and Quaternary Geology. *Wiley, New York, Fig. 8–4, p. 209.)*

will simply distinguish between two main types of feature – **eskers** and **kames**.

Eskers are sinuous, sometimes discontinuous, ridges formed of sand, gravel or boulders and range up to 200 m in height, 3 km in width and 100 km or more in length. Some consist of mounds, lined by narrow ridges and they may form single ridges or interconnected networks. Eskers may initially be deposited in a variety of positions; they can be formed in ice tunnels either subglacially or, much more rarely, englacially, or they may be deposited in channels on the glacier surface. Where initially deposited supraglacially or englacially they can eventually be superimposed on to the subglacial topography as the underlying ice melts. Eskers may also be deposited along the ice margin, either in association with marginal and submarginal channels, or in ice-marginal lakes.

Kame is a broad term describing a mound of sediment formed by the initial deposition of material within a cavity in the ice followed by slumping of this material as the supporting ice walls melt away. Often the term is used as an adjective to describe landforms constructed in particular position within, or along, a glacier. **Kame terraces** for instance, are formed by the lateral or frontal accumulation of fluvioglacial deposits along the ice margin, while a **kame complex** results from the letting down of numerous sediment-filled supraglacial depressions and cavities in stagnant ice on to the subglacial surface. All varieties of kame may be further modified by slumping resulting from the melting of ice cores within the sediments thereby giving rise to **kettle-holes**. Some landscapes described as **kettle-and-kame topography** are dominated by mounds originally representing sites of deposition in depressions and the secondary effects of sediment collapse over melted ice masses.

Fluvioglacial deposition is dominant in the proglacial zone both because of the decreased capacity for transport of meltwater streams once they emerge from the ice front and the common presence of ice-margin lakes in this environment. The rapid dumping of large quantities of coarse debris immediately beyond the glacier margin and the associated shifting of stream channels within this zone produces an extensive depositional plain known as a **sandur** (plural **sandar**). Sandar are broadly analogous to alluvial fans with characteristic entrenchment by the main feeder channels, but they experience much more drastic seasonal fluctuations in discharge. **Valley sandar** form in the laterally confined environment of glacial troughs, whereas **plain sandar** or **outwash plains** can develop along the margins of ice sheets where braided rivers from numerous outlets along the ice front coalesce. Typically the proximal zone of a sandur close to the glacier margin has only a few main meltwater streams discharging from the ice front and a characteristically pitted surface, known as a **kettled-sandur**, produced by the melting of ice masses buried by the fluvioglacially deposited sediments. Channel braiding becomes accentuated further away from the ice front in the intermediate zone and in the distal zone the channels are so shallow that overflow commonly occurs especially during peak discharges. The distal zone commonly merges into an extensive delta system formed along a proglacial lake. The sandur may extend away from the glacier front, gradually burying the delta as it advances.

## Further reading

There are several comprehensive surveys of glacial geomorphology available including Embleton and King (1975), Flint (1971) and Price (1973), while Andrews (1975) provides a brief but stimulating introduction to the subject. Drewry (1986) provides a detailed treatment of glacial processes but these are rather poorly related to landforms. My own favoured treatment is that by Sugden and John (1976). This excellent text combines breadth with depth and has provided much of the framework of discussion on which this chapter is based. Useful supplementary material on a variety of topics is to be found in Coates (1974), while the volume edited by Gurnell and Clark (1987) contains various papers on glacial and fluvioglacial sediment transport with particular reference to Alpine terrains.

Articles on glacial geomorphology are published fairly frequently in the general geomorphology journals, but the other important sources are the *Journal of Glaciology* and *Annals of Glaciology* which emphasize the characteristics of ice and mechanics of glacier flow. Some papers of geomorphic interest are also published in *Arctic and Alpine Research*, *Quaternary Research*, *Quaternary Science Reviews*, *Journal of Quaternary Science* and *Boreas*.

Within the general field of glacier characteristics and dynamics Paterson (1981) is the basic reference, while Armstrong *et al.* (1973) provide illustrated definitions of ice and snow features. Chapter 5 of Sugden and John (1976) contains a discussion of the factors controlling the global distribution of glaciers, and the temperature characteristics of glacier ice are considered in Harrison (1975) and Paterson (1981). The complexity of the mass balances of real glaciers is examined in Mayo *et al.* (1972) and Østrem (1975) looks at the possibilities of the satellite monitoring of glacial budgets.

A general review of glacier movement is provided by Kamb (1964), while a detailed treatment of internal deformation of glacier ice is contained in the classic papers by Glen (1952, 1955) and Nye (1957, 1965a). For a similarly in-depth analysis of the mechanisms of basal sliding see Kamb and La Chapelle (1964), Lliboutry (1968), Morris (1976), Nye (1970) and Weertman (1964). More general reviews can be found in Kamb (1970) and Weertman (1979). The role of bed deformation in glacier movement is assessed by Boulton and Jones (1979) and Boulton and Hindmarsh (1987). Shreve (1984) challenges the conventional distinction between cold and warm-based ice and considers the mechanism of glacial sliding at subfreezing temperatures. Ice flow in the Antarctic ice sheet is considered by Drewry (1983) and Bentley (1987), while Alley *et al.* (1986) and Blankenship *et al.* (1986) discuss the role of substrate deformation in the movement of Antarctic Ice Stream B. It should be pointed out that a full understanding of most of these articles on glacier dynamics requires a fairly sophisticated level of mathematical expertise. The brief but very useful review by Hutter (1982) is, however, readily accessible.

Turning to the movement of glaciers as a whole, extending and compressive flow is treated by Nye (1952a) and applied to a specific glacier by Meier and Tangborn (1965), while the operation of kinematic waves is analysed by Nye (1960, 1965b). Possible mechanisms of surging are considered by Budd (1975), Jarvis and Clarke (1975), Kamb *et al.* (1985) and Robin and Weertman (1973), and Clarke *et al.* (1984) examine the role of bed deformation in the surging behaviour of Trapridge Glacier, Canada. Raymond (1987) and Sharp (1988a) review ideas on the mechanisms of surging and Clarke *et al.* (1986) examine the glacier characteristics common to surging behaviour and Sharp (1988b) looks at its geomorphic consequences. Short-term glacier fluctuations are covered in Chapter 6 of Sugden and John (1976).

The large-scale form of ice masses is discussed by Nye (1952b) and Weertman (1961), and Reeh (1982) presents a model which takes into account the effects of subglacial topography. The properties of the Antarctic ice sheet are documented by Drewry *et al.* (1982) and Drewry (1983). Numerous ice-surface features are illustrated in Post and La Chapelle (1971) while Nye (1952a) considers the generation of crevasses and Hambrey (1976), Kamb (1964) and Lliboutry and Reynaud (1981) discuss the formation of various kinds of ice banding.

The processes of glacial erosion are reviewed by Boulton (1974, 1979) and detailed discussions of particular erosional mechanisms are provided by Kamb and La Chapelle (1964) and McCall (1960) (abrasion), Glen and Lewis (1961) and McCall (1960) (crushing and fracturing) and Lewis (1954), Linton (1963) and Trainer (1973) (joint-block removal). The formation of fluted surfaces at a variety of scales is discussed by Boulton (1974, 1976), Flint (1971), Goldthwait (1979) and Linton (1963). The form of glacial troughs is considered by Graf (1970), Harbor *et al.* (1988), Hirano and Aniya (1988) and Linton (1963), and their stepped long profiles are examined by Bakker (1965) and, in the broader context of contrasts between glacial and fluvial systems, by King (1970). The morphology of cirques and factors controlling their development are treated by Derbyshire and Evans (1976), Haynes (1968) and Olyphant (1981). Overviews of the creation of glacially eroded landscapes are provided by Linton (1963) and Sugden (1974) while Sugden (1978) provides a more detailed treatment of the morphology generated by the Pleistocene Laurentide ice sheet of North America.

There are several good reviews available on glacial deposition and associated landforms, including Goldthwait (1971), Price (1973) and Schluchter (1979). Boulton (1974) and Kamb and La Chapelle (1964) look at the processes of debris entrainment and transport and Boulton (1972, 1975) discusses mechanisms of deposition. Prest (1968) provides a comprehensive classification of depositional landforms. Fluted moraines are discussed by Baranowski (1970), while drumlins are reviewed by Menzies (1979) and considered in detail in the symposium volume edited by Menzies and Rose (1987). Specific processes of formation are proposed by Evenson (1971), Shaw and Sharpe (1987) and Smalley and Unwin (1968). Embleton and King (1975) and Price (1973) contain extended treatments of end moraines, and Cowan (1968) discusses the formation of ribbed (Rogen) moraine.

For excellent discussions of meltwater in glaciers see Shreve (1972), Stenborg (1969) and Weertman (1972). Nye (1976) provides an additional recent analysis with particular reference to the generation of jökulhlaups, and Thorarinsson (1953) contains a vivid description of the 1934 Grímsvötn jökulhlaup. The Late Pleistocene Lake Missoula cataclysmic flooding is discussed by Baker and Bunker (1985) and Waitt (1985), but the classic papers by Bretz (1923, 1969) are well worth consulting. The relationships between sediment yield and discharge in a meltwater stream are analysed in Østrem *et al.* (1967). Plastically sculptured forms (p-forms) are discussed by Dahl (1965), but Boulton (1974) and Gjessing (1965) provide alternative explanations for their development. The origin of meltwater channels is discussed in the classic paper by Mannerfelt (1949) and by Clapperton (1968) and Price (1973). Flint (1971), Parizek (1969) and Price (1973) consider the complex range of depositional fluvioglacial forms and Church (1972) and Krigstrom (1962) provide detailed treatments of the proglacial environment.

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