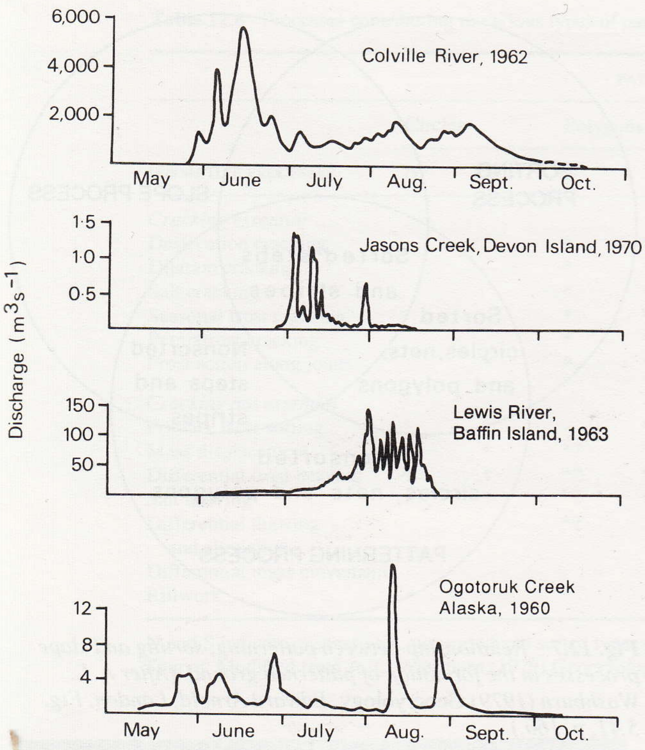

Fig. 12.6 Runoff regimes of selected rivers in the periglacial zone of North America. (Modified from H. M. French (1976) The Periglacial Environment. *Longman, London, Fig. 8.1, p. 169, based on M. Church (1974)* Permafrost Hydrology; Proceedings of Workshop Seminar. *Environment Canada, Ottawa 7–20.)*

current rates of weathering and mass movement supplying more debris than can be transported by existing rivers. Some of this material may, however, be a relic of higher rates of rock breakdown and sediment supply during the Pleistocene.

12.2.6 Aeolian activity

Present-day periglacial areas characterized by extreme aridity are favourable environments for aeolian activity, and there is abundant evidence for much more widespread and intense wind action in areas marginal to the great northern hemisphere ice sheets during the Pleistocene. In addition to strong winds and the freeze-drying of sediments, low precipitation and low temperatures are associated with minimal vegetation cover. Aeolian erosion is evident in faceted and grooved bedrock surfaces, deflation forms in unconsolidated fluvial sediments, and in the formation of ventifacts (see Section 10.2.2.1). Wind is also important in the movement and deposition of snow which, together with topographic features, determines areas of snow accumulation and removal and thereby the distribution of an insulating snow cover.

Of great geomorphic significance are various aeolian deposits, especially the loess which blankets much of those areas in North America and Eurasia which were located on the southern margin of the Pleistocene ice sheets. Loess is composed of angular to sub-angular particles and cohesion between grains allows the development of steep, high cliffs where thick blankets have been dissected by fluvial action. Although loess can form in warm deserts, most is believed to have originated in past or present periglacial environments. Glacial grinding, frost cracking and salt weathering appear capable of producing sufficiently fine particles which can be picked up from sites such as glacial outwash plains and carried considerable distances by the wind before being deposited as extensive loess blankets, exceptionally up to 100 m thick and covering tens of thousands of square kilometres (see Section 10.3.7).

During the glacial phases of the Pleistocene, the steep pressure gradient between the high pressure cells over the Laurentide and Fenno-Scandian ice sheets and low pressure to the south would have produced high mean wind velocities and conditions conducive to significant aeolian transport and deposition. Dating of discrete vertically stacked layers of loess interbedded with soils indicative of more temperate and humid conditions in China and Europe has indeed indicated a close correlation between loess deposition and phases of ice sheet extension (see Section 14.3.2).

12.3. Periglacial landforms

12.3.1 Patterned ground

First described more than a century ago, patterned ground is one of the most conspicuous features of periglacial environments. It can vary widely in scale from patterns composed of elements only a few centimetres across to those with dimensions of 100 m or more. Although some forms of patterned ground are found in other environments, especially deserts, the extent of its development is much greater in the periglacial zone than elsewhere since it is most characteristic of surfaces subject to intense frost action.

The most widely used descriptive classification of patterned ground has been developed by A. L. Washburn. This emphasizes both shape and the degree of sorting of the constituent materials. Five basic patterns are recognized – circles, nets, polygons, steps and stripes – although one form may grade into another. Each of these types is subdivided on the basis of whether the pattern results from the sorting of surface materials into fractions of large and small size or whether it has developed without sorting. Slope angle is an additional factor. Circles, nets and polygons are most common on flat surfaces, whereas steps and stripes tend to form on gradients of between 5 and 30°, with transitional forms developing on gentler slopes of between 2 and 5°. Mass movement is generally too active on slopes steeper than 30° to allow patterns to form. A summary of the characteristics and environmental associations of the various types of patterned ground is presented in Table 12.5.

Table 12.5 Characteristics of different forms of patterned ground

Circles	Occur singly or in groups. Typical dimensions 0.5–3 m. Nonsorted type characteristically rimmed by vegetation, sorted type bordered by stones which tend to increase in size with size of circle. Found in both polar and alpine environments but are not restricted to areas of permafrost. Unsorted circles also recorded from non-periglacial environments
Polygons	Occur in groups. Nonsorted polygons range from small features (<1 m across) to much larger forms up to 100 m or more in diameter. Sorted polygons attain maximum dimensions of only 10 m. Stones delimit polygon border and surround finer material. Nonsorted forms are delineated by furrows or cracks. Some types of polygon occur in hot desert environments but most are best developed in areas subject to frost. Ice-wedge polygons only form in the presence of permafrost
Nets	A transitional form between circles and polygons. Usually fairly small (<2 m across). Earth hummocks comprising a core of mineral soil surmounted by vegetation are a common form of unsorted net.
Steps	Found on relatively steep slopes. They develop either parallel to slope contours or become elongated downslope into lobate forms. In unsorted forms the rise of the step is well vegetated and the tread is bare. In the unsorted type the step is bordered by larger stones. Lobate forms are known as stone garlands. Neither type confined to permafrost environments
Stripes	Tend to form on steeper slopes than steps. Sorted stripes composed of alternating stripes of coarse and fine material elongated downslope. Nonsorted variety delineated by vegetation in slight troughs. Not confined to periglacial environments

Source: Based on discussion in A. L. Washburn (1979) *Geocryology*. Edward Arnold, London, pp. 122–56.

Numerous factors contribute to the form of the great variety of patterned ground including the interactions between sorting, patterning and slope processes (Fig. 12.7). An important distinction is that between those forms of patterned ground in which initial cracking of the surface is essential and those in which it is not (Table 12.6). Causes of cracking include contraction on drying (desiccation cracking), stretching of the surface as a result of heaving (dilation cracking), and thermal contraction of seasonally or perennially frozen ground (frost cracking). Only the latter is restricted solely to periglacial environments. Desiccation and dilation cracking are important in the formation of small non-sorted polygons, whereas frost cracking is instrumental in the development of larger forms. The most important patterning processes which do not depend on initial cracking include differential frost heaving and mass displacement.

Frost heaving is apparently a primary cause of a number of sorted forms of patterned ground; it helps to segregate larger stones, which tend to move upward and outward, from the finer-grained matrix which forms a central core. Continued development at a number of discrete sites eventually leads to coalescence and the formation of polygonal rims formed primarily of stones. Sedimentation and mass displacement contribute to the development of non-sorted circles while surface wash is probably a factor in the formation of sorted stripes.

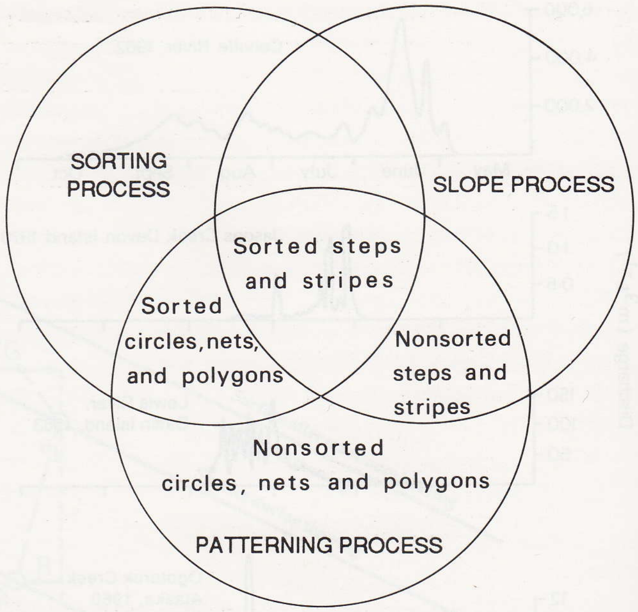

Fig. 12.7 Relationship between patterning, sorting and slope processes in the formation of patterned ground. (After A. L. Washburn (1979) Geocryology. *Edward Arnold, London, Fig. 5.41, p. 160.)*

The regularity of many forms of patterned ground has suggested to some researchers that convection occurring when frozen ground thaws may be an important mechanism in its development. The idea is that water-saturated soil close to the surface warms up while the temperature of that adjacent to the still frozen subsoil remains close to freezing point. Convection in this situation is possible because water is densest at 4 °C, so water at the thawing front is less dense than the slightly warmer water close to the surface. When the active layer thaws the cooler water at the thawing front rises while descending plumes of warmer water lead to localized melting of the underlying ice. This produces a regular undulating interface between frozen and unfrozen soil which, it is thought, can be reflected in the surface topography. The exact mechanism by which this occurs is not clear, but it probably involves a range of processes including frost heave. Striped forms of patterned ground according to this model are explained by the downslope elongation of convection cells within the soil.

12.3.2 Ground-ice phenomena

12.3.2.1 Ice wedges

Ice wedges are downward-tapering bodies of ice (Fig. 12.8). They can be over 1 m across at the top and reach a depth of 10 m or more. They have a polygonal form in plan,

Table 12.6 Processes contributing to various types of patterned ground

	PATTERNED GROUND MORPHOLOGY									
	Circles		Polygons		Nets		Steps		Stripes	
FORMATIVE PROCESSES	N	S	N	S	N	S	N	S	N	S
Cracking essential										
Desiccation cracking			*	*	*	*			*	*
Dilation cracking			*	*	*	*			*?	*
Salt cracking			*	*						
Seasonal frost cracking			*	*	*	*			*?	*?
Permafrost cracking			*	*	*	*?			*	*
Frost action along joints		*	*	*					*?	*
Cracking not essential										
Primary frost sorting		*		*?		*		*?		*?
Mass displacement	*	*	*?	*?	*	*	*	*	*	*
Differential frost heaving	*	*	*?	*?	*	*	*?	*	*	*
Salt heaving	*	*	*?	*?	*	*	*?	*	*	*
Differential thawing and eluviation			*?			*		*?		*?
Differential mass movement							*	*	*?	*
Rillwork									*?	*

N and S indicate, respectively, nonsorted and sorted types of patterned ground.
Source: Modified from A. L. Washburn (1979) *Geocryology*. Edward Arnold, London, Table 5.1, p. 158.

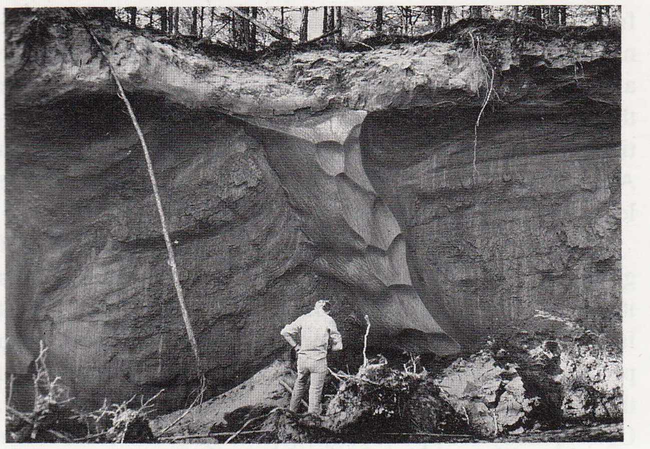

Fig. 12.8 *Ice wedge in the right bank of the Aldan River, 170 km upstream of its junction with the Lena River, Yakutia, USSR. (Photo courtesy A. L. Washburn.)*

and **ice-wedge polygons** are associated with the most impressive form of patterned ground in periglacial environments. Frost cracking appears to be the key mechanism necessary for the development of ice wedges. Extreme cooling of ice-rich permafrost in winter leads to thermal contraction and cracking of the surface (Fig. 12.9). As the wedge grows it compresses the surrounding sediments which tend to bulge up around the perimeter of each polygon forming a low ridge. In other cases thawing and erosion of the ice wedge produces a perimeter trough. In dry permafrost, where meltwater is scarce, the cracks may be filled with loess or other coarser sediments to form **sand wedges**.

Recent field studies have revealed that some ice-wedge cracks originate near the top of the permafrost and then propagate both upwards to the surface and downwards into the ice wedge. Only a proportion of wedges tend to crack in any one year, and the frequency of cracking seems to vary inversely with the depth of snow cover. The average amount of ice accreted each year varies enormously so the total volume of ice present is not necessarily a good indicator of how long an ice wedge has been growing. Observations made of the development of ice wedges after the artificial draining of a lake in northern Canada have demonstrated that under ideal conditions they can grow rapidly and reach a significant size after just a few years.

Ice-wedge casts form when an ice wedge is replaced by sediments as a result of the long-term thawing of the permafrost. The form of the ice wedge is preserved by sediment and may survive long after the permafrost in which the original ice wedge developed has completely thawed. Ice-wedge casts are consequently an important indicator of the previous existence of permafrost. Their palaeoenvironmental significance is also enhanced by the fact that ice wedges only seem to form in present-day permafrost environments where the mean annual temperature is below –6 to –8 °C.

12.3.2.2 Pingos

Pingos, named after the Eskimo word for a hill, are large perennial ice-cored mounds. They are roughly circular to elliptical in plan and range in size from 3 to 70 m high and from 30 to 600 m in diameter (Fig. 12.10). Large dilation cracks generated by the progressive growth of the ice core

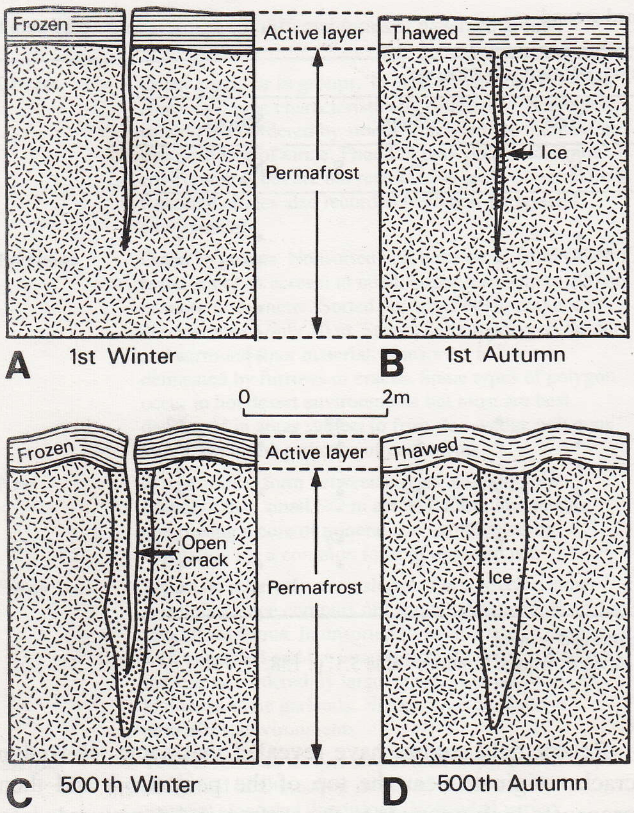

Fig. 12.9 Hypothesized development of an ice wedge initiated by frost cracking due to thermal contraction. The initial crack, which may be only a few millimetres wide, collects snow and meltwater from thawing of the active layer (A). This water freezes below the permafrost table to form a thin ice vein (B). The initial fracture represents a zone of weakness so will be a favoured site for frost cracking in subsequent years. However, cracking does not necessarily occur each year and expansion during summer heating will tend to close cracks developed during the preceding winter. Further ingress of snow and meltwater adds new ice until an ice wedge is eventually formed (C and D). (After A. H. Lachenbruch (1962) Geological Society of America Special Paper *70, Fig. 1, p. 5.)*

commonly mark the summit. These may widen sufficiently to expose the ice core and initiate thawing, thereby giving rise to a **collapsed pingo**. Various materials can envelop the ice core, including clays, silts, gravels and even bedrock. Drilling has shown that in some cases the core of clear ice extends to depths greater that the pingo height. Vertical growth rates of pingos may reach 0.2 m a^{-1} but they tend to decline rapidly as the pingo grows. Large pingos certainly take thousands of years to develop, as indicated by the radiocarbon dating of two pingos in the Canadian Arctic which gave approximate ages of 4500 and 7000 a.

The ice core forming a pingo is thought to grow through the freezing of water forced upwards at pressure. Two mechanisms have been proposed to explain the generation of this pressure. In **closed system pingos** cryostatic pressure is involved. This may be brought about in two related ways. If

Fig. 12.10 Pingo, Wollaston Peninsula, Northwest Territories, Canada. (Photo courtesy A. L. Washburn.)

a lake in a permafrost zone is infilled by sediment and vegetation, insulation of the underlying ground will be reduced (Fig. 12.11) and freezing will encroach from the base, sides and top. This will trap a body of water which on freezing expands and domes the overlying sediments. Alternatively, diversion of a river or draining of a lake may have a similar effect by reducing ground insulation. Support for this model of cryostatic pressure comes from the distribution of pingos in the Mackenzie Delta area of the Canadian Arctic where 98 per cent of the 1380 pingos recorded are located in, or closely adjacent to, lake basins.

The **open system pingo** model involves the freezing of ground water flowing under hydrostatic pressure as it forces its way towards the surface from below a thin permafrost layer (Fig. 12.12). It has been calculated that very high pressures are required to dome pingos and that these exceed those commonly attained within unconfined ground water. Consequently the open system type may in fact form under temporary closed-system conditions as open taliks are frozen in winter. Open system pingos extend into the zone of discontinuous permafrost where mean annual temperature may be only a little below freezing since here water circulation is freer. They tend to occur in clusters in favourable areas of drainage where the hydrostatic pressure is more constant. In contrast, the closed system type usually occur as isolated features on flat surfaces related to lakes and are confined to the zone of continuous permafrost where minimum mean annual ground temperatures are around –5 °C.

12.3.2.3 Palsas

Palsas are mounds or more elongated forms occurring in bogs and containing perennial ice lenses. They differ from pingos in containing peat as a major constituent and a core composed of discrete superimposed ice lenses rather than a

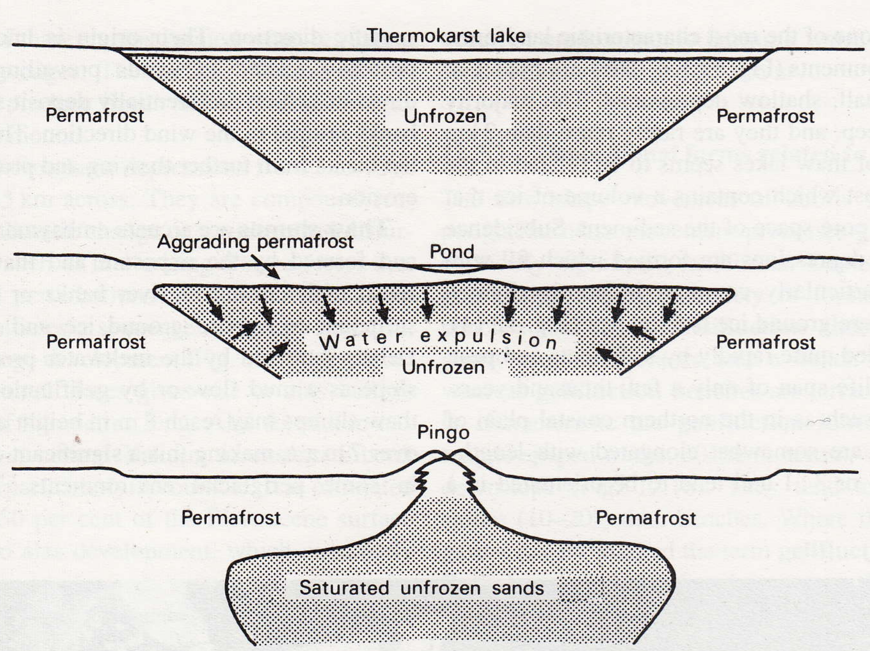

Fig. 12.11 Schematic representation of the development of a closed system pingo following the infilling of a lake. (Modified from A. L. Washburn, (1979) Geocryology. *Edward Arnold, London, Fig. 5.49, p. 183.)*

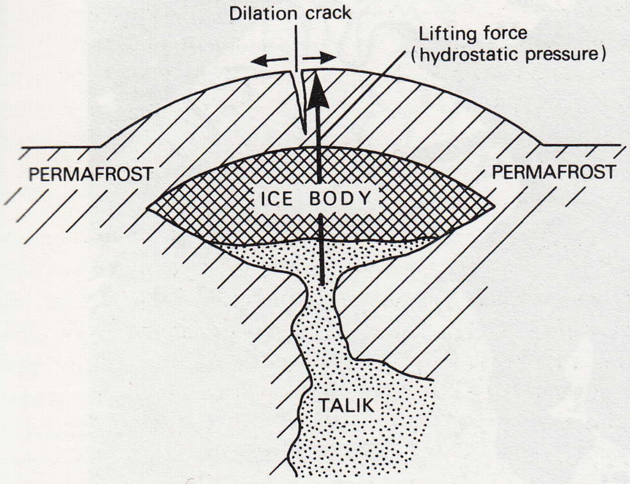

Fig. 12.12 Schematic representation of the development of an open system pingo. (Modified from F. Muller (1968) in: R. W. Fairbridge (ed.) Encyclopedia of Geomorphology. *Reinhold, New York, Fig. 2, p. 846.)*

single ice mass. The mode of ice formation is uncertain but probably originates from differential frost heaving with peat playing a crucial insulating role. Palsas are commonly found in areas of discontinuous permafrost.

12.3.2.4 Thermokarst

Thermokarst is a term used to encompass a variety of topographic depressions formed by the thawing of ground ice. It is so called because of the superficial resemblance of the landforms produced to those characteristic of true karst terrain. In addition to the collapse of material into the space previously occupied by ice, thermokarst features may be significantly affected by flowing water released by thawing. High, short-term rates of erosion can occur, especially where ice masses are exposed along cliffs or stream banks. Although some thermokarst features can be attributed to climatic change, many others form as a result of the 'normal' variability of the periglacial environment. Thermokarst most commonly originates from a modification of surface conditions and this can be brought about in a number of possible ways including disturbance of vegetation, cliff retreat and changes in stream courses.

Thaw lakes are perhaps the most ubiquitous thermokarst

Fig. 12.13 Small thaw lakes, Bathurst Island, Northwest Territories, Canada. (Photo courtesy Geological Survey of Canada.)

form, and arguably one of the most characteristic landforms of periglacial environments (Fig. 12.13). They are bodies of water which fill small, shallow depressions. The majority are less than 5 m deep, and they are rarely more than 2 km across. The origin of thaw lakes seems to be related to the melting of permafrost which contains a volume of ice that exceeds the normal pore space of the sediment. Subsidence occurs and shallow depressions are formed which fill with water. They are particularly common in poorly drained lowland regions where ground ice is abundant (Fig. 12.14). They are usually filled quite rapidly by sediments and peat, and have a limited life span of only a few thousand years. In some localities, such as in the northern coastal plain of Alaska, thaw lakes are somewhat elongated with length–width ratios of 2 : 1 or 3 : 1 and tend to be orientated in a specific direction. Their origin is uncertain, but they are probably related to winds prevailing from a particular direction which preferentially deposit sediments along lake banks normal to the wind direction. These deposits insulate the banks from further thawing and protect them from wave erosion.

Thaw slumps are arcuate embayments facing downslope and formed by the exposure and thawing of ground ice. Basal undercutting of river banks or mass movement are sufficient to expose ground ice and initiate the process. Debris saturated by the meltwater produced moves downslope as a mud flow or by gelifluction. The headwalls of thaw slumps may reach 8 m in height and retreat at rates of over 7 m a^{-1}, making this a significant denudational process in some periglacial environments. **Thermocirques** are

Fig. 12.14 Thaw lakes in the Tuktoyaktuk Peninsula which lies within the continuous permafrost zone of the Northwest Territories, Canada. The area shown is about 180 km across. (Landsat image courtesy R. S. Williams Jr)

large-scale variants of thaw slumps formed when retreating slopes intersect ice wedges. The surface thawing of ice-wedge polygons can produce linear and polygonal troughs surrounding a central mound.

Alases are major thermokarst depressions from 3 to 40 m deep and 100 m to 15 km across. They are compound features resulting from climatic change or a widespread disturbance of the surface such as that arising from a forest fire. These environmental perturbations can lead to permafrost degradation which initiates a sequence of surface collapse, lake formation and pingo development (Fig. 12.15). Eventual coalescence of individual alases gives rise to **alas valleys** which may be tens of kilometres long. Alas formation has affected large areas in central Yakutia in eastern Siberia where conditions are particularly favourable. It is estimated that in this area 40–50 per cent of the Pleistocene surface has been subjected to alas development, which apparently occurred predominantly during a warm phase of the Holocene between 9000 and 2500 a BP.

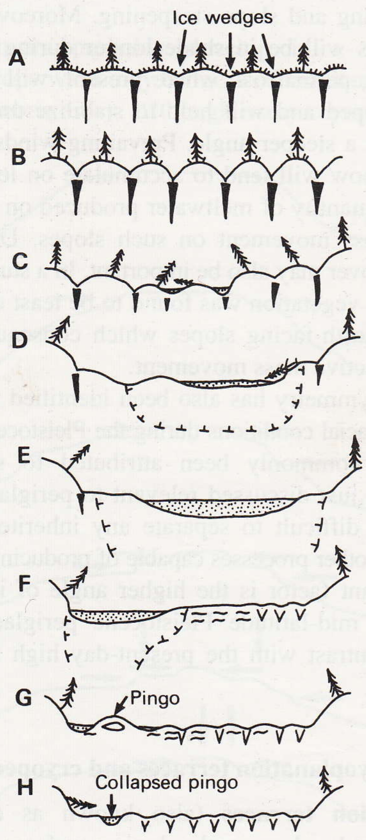

Fig. 12.15 Schematic representation of the sequence of development and decay of alases: (A) original lowland surface with ice wedges; (B) effect of increase in depth of active layer promoted by climatic warming or ground disturbance; (C) initial thermokarst stage; (D) young alas stage; (E) mature alas stage; (F) old alas stage; (G) phase of possible pingo formation; (H) development of collapsed pingos. (After P. J. Williams (1979) Pipelines and Permafrost. *Longman, London, Fig. 2.2, p. 21, modified from T. Czudek and J. Demek (1970)* Quaternary Research *1, Fig. 9, p.111.)*

12.3.3 Depositional forms related to mass movement

The downslope movement of debris in association with periglacial mass movement processes gives rise to a range of locally significant landforms. Deposits associated with gelifluction assume a variety of forms, including sheets, benches and lobes. **Gelifluction sheets** have a smooth, gently sloping surfaces and a bench-like lower margin, whereas **gelifluction benches** are terrace forms. Particularly characteristic are **gelifluction lobes** which consist of tongue-shaped forms, 30–50 m across, which are elongated downslope (Fig. 12.16). They tend to occur on steeper slopes (10–20°) than benches. Where the downslope elongation is very marked the term **gelifluction stream** is used. Both benches and lobes have steep fronts, 1–6 m high, which may be subject to erosion. They appear to form as stones, which have been pushed downslope, accumulate and dam the movement of material above.

Some gelifluction deposits are crudely stratified and most exhibit a preferred downslope orientation of the long axis of their angular constituent particles, which may range in size from boulders down to silt-sized material. Relict gelifluction deposits are difficult to distinguish from other solifluction debris, although they should be less weathered than similar deposits from non-periglacial environments.

Block slopes are slopes mantled with angular boulders, mostly between 1 and 3 m across, covering 50 per cent or more of the ground surface. In the case of **block streams** the boulders are concentrated in valley bottoms or occur as

Fig. 12.16 Gelifluction lobe, Hesteskoen, Mesters Vig, north-east Greenland. (Photo courtesy A. L. Washburn.)

narrow downslope linear deposits on hillsides. Downslope movement is indicated for both forms by the orientation of the long axes of the boulders which tends to be roughly normal to the slope contours. Gelifluction and frost creep are the favoured transporting mechanisms for most occurrences. **Block fields**, consisting of relatively flat boulder-covered surfaces, are also common. They probably form largely by *in situ* frost shattering and the heaving of jointed bedrock and involve no significant downslope movement.

Rock glaciers are tongue-shaped masses of angular boulders resembling in form a small glacier. They have a steep front, exceptionally exceeding 100 m in height, which stands at the angle of repose of the constituent material. Rock glaciers usually descend from cirques or cliff-faces and may reach a length of 1 km or more. Large boulders and smaller stones typically form the surface which commonly exhibits arcuate ridges and lobes. Active rock glaciers contain ice at depth, either filling voids (ice-cemented type) or forming a core (ice-cored type). Some of the ice-cored variety are probably derived from debris-covered glaciers, whereas others may be transitional to ice-cored moraines. In localities where rock glaciers and ice glaciers are both currently active the distinction between them is essentially one of the quantity of debris carried. Where debris-rich glacier ice becomes stagnant sufficient ice may remain mixed with fine sediment at depth to sustain slow movement, largely by internal deformation and basal shear. Frost creep and gelifluction may also contribute to downslope movement. Irrespective of the actual mechanism of movement, rock glaciers are important transportational agents in highland periglacial environments.

12.3.4 Asymmetric valleys

Asymmetric valleys are valleys with one slope significantly steeper than the other. Virtually all valleys are asymmetric to a certain extent, but some display a degree of asymmetry sufficiently marked to warrant some special explanation. Asymmetric valleys are known from a wide range of environments and can clearly be caused by a number of factors, but geomorphologists working in the periglacial zone have long recognized the prevalence of asymmetric valleys there. Collection of field data has indicated no uniformity in the preferred orientation of the steeper slope in periglacial asymmetric valleys (Table 12.7), although it has been suggested that steeper north-facing slopes predominate in the northern hemisphere below a latitude of about 70 °N.

Various conditions existing in periglacial environments may contribute to valley asymmetry. In the northern hemisphere longer exposure to the Sun on south-facing slopes will result in a more rapid and prolonged thawing and a greater abundance of meltwater, promoting gelifluction and other mass movement processes. Consequently, south-facing slopes will experience a more active reduction of slope angle, while their colluvium will tend to divert streams towards opposing north-facing slopes which will be subject to undercutting and slope steepening. Moreover, as north-facing slopes will be in shade longer during the summer thaw season, permafrost, where present, will be preferentially developed and will help to stabilize unconsolidated sediments at a steeper angle. Prevailing winds may play a role since snow will tend to accumulate on lee slopes and the greater quantity of meltwater produced on thawing will augment mass movement on such slopes. Differences in vegetation cover may also be important. In a study in Alaska, for instance, vegetation was found to be least developed on the colder north-facing slopes which consequently experienced less active mass movement.

Table 12.7 Nature of slope asymmetry in the periglacial zone of the northern hemisphere

LOCALITY	ORIENTATION (ASPECT) OF STEEPER SLOPE	VALLEY ALIGNMENT
Central Alaska	N	E–W
North-west Alaska	N	E–W
Southampton Island, Canada	N	E–W
Banks Island, Canada	SW	NW–SE
Caribou Hills, Canada	N, S	E–W
Yakutia, Siberia (3 studies)	N	E–W
Andreeland, west Spitsbergen	S	E–W
Conwayland, west Spitsbergen	W	N–S
Kaffioya-Ebene, west Spitsbergen	S	E–W
Wollaston-Vorland, east Greenland	N	E–W

Source: Modified from H. M. French (1976), *The Periglacial Environment*, Longman, London, Table 8.3 p. 179.

Valley asymmetry has also been identified in areas subject to periglacial conditions during the Pleistocene. Although these have commonly been attributed to some of the mechanisms just discussed relevant to periglacial environments, it is difficult to separate any inherited periglacial effects from other processes capable of producing asymmetry. One important factor is the higher angle of inclination of the Sun in mid-latitude Pleistocene periglacial environments in contrast with the present-day high latitude periglacial zone.

12.3.5 Cryoplanation terraces and cryopediments

Cryoplanation terraces (also known as **altiplanation terraces**) are level or gently sloping surfaces found in the periglacial zone which are cut into bedrock on hill summits or upper hillslopes. **Cryopediments** are a similar form developed at the foot of valley sides or on marginal slopes. Unless they transect structure it may be difficult to distinguish them from structurally controlled benches, and in any case lithological and structural factors are important in their formation. Cryoplanation terraces range from 10 m to 2 km across and up to 10 km in length. The risers between terraces may reach a height in excess of 70 m and stand at an angle of 30° or more where debris-covered, and up to 90° where bedrock is exposed. The terraces are mantled by gelifluction debris and may contain patterned ground.

Cryoplanation terraces are thought to be formed by several processes working in conjunction. In essence they appear to form by the combined effects of the break-up of bedrock by frost action and scarp recession. Various stages are involved (Fig. 12.17). According to one model, development begins with the formation of a nivation hollow or bench associated with snow patches. Nivation then proceeds to erode a cliff which recedes as debris is transported away from the cliff base, largely by gelifluction and slope wash. Continued cliff recession on either side of an interfluve finally forms a summit terrace with residual rock masses. Some researchers, however, believe that most descriptions of cryoplanation terraces are of relict forms, and that in many cases the features described may be no more than benches related to the differential erosion of contrasting lithologies.

Cryopediments probably develop in a similar way except that slope wash may be more active than gelifluction in carrying debris away. Although many cryoplanation terraces and cryopediments are associated with present-day permafrost, frozen ground does not seem to be a prerequisite for their formation.

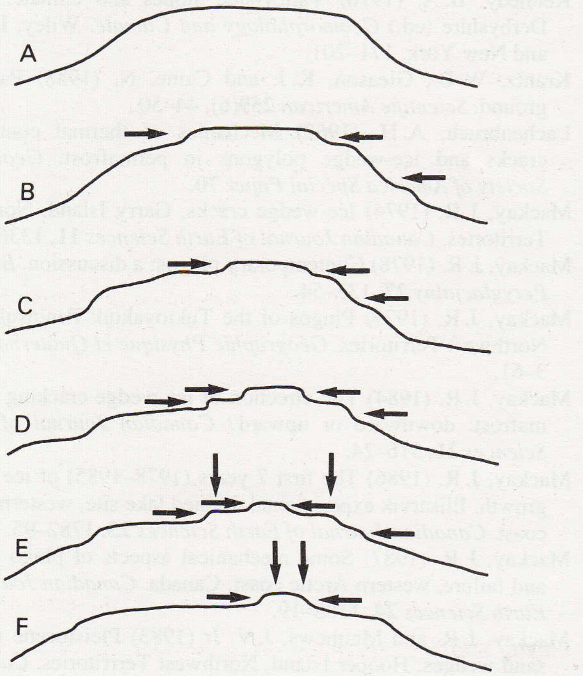

Fig. 12.17 Stages in the development of cryoplanation terraces in resistant rock: (A) original surface; (B) formation of nivation hollows; (C) initial development of cryoplanation terrace; (D) mature stage of cryoplanation terrace development; (E) initial formation of cryoplanation surface; (F) development of cryoplanation summit surfaces. Arrows indicate the direction of surface modification. (Based on model of J. Demek (1969) Biuletyn Peryglacjalny ***18****, 115–25, after H. M. French (1976)* The Periglacial Environment. *Longman, London, Fig. 7.11, p. 160.)*

Further reading

Students of periglacial geomorphology have no shortage of excellent texts which they can consult. Washburn (1979) provides a fully illustrated and referenced survey of processes and landforms; a considerable strength of this text is the extensive reference made to the work of Russian and East European scientists who have made very important contributions to periglacial studies. Other general sources include Embleton and King (1975), French (1976) and Williams and Smith (1989). The volumes edited by Church and Slaymaker (1985) and Clark (1988) provide detailed up-to-date treatments of many aspects of periglacial geomorphology, while a number of useful papers are contained in *Zeitschrift für Geomorphologie Supplementband*, 71. Various regional studies are included in Boardman (1987), although here the emphasis is on relict periglacial forms in the British Isles.

The problems involved in the human utilization of high latitude environments have received much attention. Although these are not specifically considered here, discussions of various aspects of applied periglacial geomorphology are to be found in Brown (1970), Harris (1986), Sugden (1982) and Williams (1979). Journals containing articles on periglacial geomorphology include *Biuletyn Peryglacjalny* (a Polish publication but with many articles in English), *Canadian Journal of Earth Sciences, Quaternary Research* and *Arctic and Alpine Research*, in addition to other journals which publish papers on geomorphology generally. Useful annual reviews of periglacial geomorphology are to be found in *Progress in Physical Geography*.

Harris (1986) provides a detailed discussion of the nature and formation of permafrost, while the pioneer work by Taber (1929, 1930, 1943) on frost action is still worth consulting. Thorn (1979) presents some intriguing data on the occurrence of freeze–thaw cycles and the efficacy of frost shattering in the alpine environment of the Colorado Rockies, and White (1976) compares the role of frost shattering and hydration shattering. Both field and laboratory investigations of frost shattering are reviewed by McGreevy (1981), and Pavlik (1980) provides a brief discussion of the factors controlling ground ice formation. Salt weathering in relation to tor formation is considered by Selby (1972).

Mass movement processes are covered by Harris (1987) and Rapp (1986), and also in the more general texts, especially Washburn (1979). Benedict (1976) gives a useful review of frost creep and gelifluction and also provides a detailed discussion on mass movement mechanisms in alpine environments (Benedict, 1970). Fluvial processes in specific periglacial environments are discussed in Church (1972)