Glacial sediments

11

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11.1 Introduction

The concept of extensive ice sheets and glaciers existing during periods of colder climate, and transporting and depositing sediment, originated in continental Europe, primarily during the 1830s. This 'glacial theory' explained the origin of superficial deposits that contain large erratic boulders and which cover broad areas of Europe and North America. By the latter part of the 19th century, the glacial theory had replaced the earlier notion of a universal flood, the Biblically inspired 'diluvial' theory, and its offshoot, the 'drift' theory which invoked extensive iceberg rafting to distribute foreign boulders and other rock debris derived from the far north (Flint, 1971, pp. 11–15). Recognition of pre-Quaternary ice ages also dates back to the mid- to late 19th century with the interpretation of Permo-Carboniferous rocks in Gondwana as ancient glacial deposits (John, 1979). Most poorly sorted, till-like sediments (boulder clays) were interpreted as the deposits of glaciers or icebergs, and therefore as indicators of ancient ice ages. Consequently, ice ages were identified in almost all geological periods (Coleman, 1926).

In the 1950s, recognition of the role of turbidity currents and of slumping and mass movement in deep oceans provided an alternative mechanism for the deposition of till-like sediments, such as pebbly mudstones (Crowell, 1957). The realization that sediments produced by non-glacial processes can be so similar to those deposited by glaciers led to controversy over the accurate interpretation of many of these unsorted deposits (Dott, 1961). Criteria characteristic of glacial conditions, in addition to diamictite, were highlighted by Harland, Herod & Krinsley (1966) and Flint (1975) (see Sect. 11.5.1). However, heated discussions continued as to whether the widely recognized diamictite-bearing units truly record cold climates or whether many, particularly those of Late Proterozoic age, reflect deposition in regions of active tectonics (Schermerhorn, 1974).

Rigorous sedimentological study and re-examination of many ancient inferred glacial deposits, and more thorough investigation of the sedimentary processes and products of many modern glacial and glaciomarine environments, have taken place in the last few decades in response to this phase of scepticism. Through improved communication between those working

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on modern and Pleistocene sediments and those studying ancient glaciogenic rocks, our knowledge and understanding of glaciogenic sedimentary processes and products is now a great deal more sophisticated.

Non-genetic terms ***diamicton*** (unlithified) and ***diamictite*** (lithified) were introduced by Flint, Sanders and Rodgers (1960a,b) to describe sediment 'consisting of sand and/or larger particles dispersed through a muddy matrix' (p. 508, Flint, Sanders & Rodgers, 1960a). In contrast, the genetic terms ***till*** (unlithified) and ***tillite*** (lithified) are used for sediment that has been transported and deposited primarily by glacier ice with subordinate modification by other processes (Dreimanis & Lundqvist, 1984; Dreimanis, 1989). Note that the original definition of diamicton (diamictite) states that the sediment (rock) is matrix-supported, in contrast to some recent uses of the term that include clast-supported sediment and rock (Frakes, 1978; N. Eyles, Eyles & Miall, 1983). Diamicton(ite) should be restricted to matrix-supported sediment (rock) and the terms are used in that sense here. ***Diamict*** is a general term for both consolidated and unconsolidated deposits (Harland, Herod & Krinsley, 1966).

Accurate recognition and interpretation of glacial sediments is extremely important because ice sheets, which are continental-sized glaciers, influence sedimentation over the whole globe by causing changes in climate, sea level, and oceanic circulation patterns. Glacial sediments are very varied, however, and the extent of glacial influence is typically difficult to define. This is because glacial sedimentary processes are commonly superimposed upon processes characteristic of other environments and are themselves modified by other processes, for example fluvial or marine reworking, downslope sediment failure. In these respects, this chapter interconnects with much information provided elsewhere in this book.

11.2 Characteristics of glaciers

A glacier is a mass of ice which deforms and moves due to the force of its own weight. Mass – that is, ice and/or water – and heat are constantly exchanged between the glacier and both the atmosphere above and the bed, or waterbody, below. This continual transfer of mass and heat, which ultimately responds to variations in climate, controls the balance between erosion and deposition and the nature of many sedimentary processes in glacial environments.

Glacial ice may extend from the land into the sea, over areas of low and high relief, at low and high altitudes, and may be subject to small or enormous seasonal fluctuations in temperature. Thus glacial deposits can be preserved in both terrestrial and marine settings, and under a variety of tectonic and climatic conditions. About 10% of the Earth's surface is covered by glacial ice today. During the Quaternary glaciation, maximum coverage was about 30% (Flint, 1971, p. 80), and the resulting sediments were deposited over a large portion of the Earth's surface. In addition to till, glaciofluvial, glaciolacustrine, glacioaeolian and glaciomarine sediments cover substantial areas marginal to glaciated regions.

11.2.1 Mass balance

In general, the formation of glaciers requires low temperatures and high precipitation. Both high latitudes and high altitudes are conducive to glacial growth. Glaciers are nourished in the ***accumulation zone*** (Fig. 11.1), where snow is buried and compacted by subsequent snowfalls and locally by the refreezing of percolating water after a summer thaw. The density increases as air is gradually squeezed out. The resulting glacial ice consists of interlocking ice crystals with isolated air bubbles and is effectively impermeable to both air and water (Paterson, 1969). Material is removed from the glacier in the ***ablation zone*** (Fig. 11.1) by surface melting, basal ice melting, meltwater runoff, sublimation, and iceberg calving.

The algebraic difference between the amounts of accumulation and ablation over a given time is the ***net mass balance***. A glacier in which accumulation is equal to ablation (net mass balance = 0) will have a constant mass, and a corresponding constant thickness and area. A change in mass balance changes the dimensions of the glacier, determining whether the margin is advancing, retreating, or stationary.

11.2.2 Thermal regime

Thermal regime refers to whether most of the ice is above or below the ***pressure melting point*** (the temperature at which ice

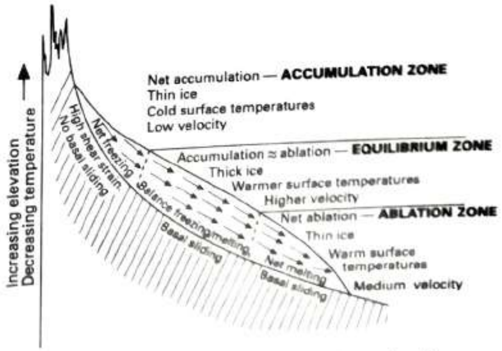


Figure 11.1 Composite model of the thermal regimes of a valley glacier showing velocity vectors, the variety of possible conditions, and controlling parameters (modified after Paterson, 1969; Drewry, 1986).

in a glacier melts, dependent upon the overburden pressure) and is largely determined by climate, mass balance, and ice thickness and velocity (Boulton, 1972b). Thermal regime defines boundary conditions at the glacier sole which have an important effect upon glacial and sedimentological processes. Basal ice temperature is the most important parameter; it is controlled by surface temperature, geothermal heat flux, and heat generated by friction within and at the base of the ice. Three basal thermal regimes can be defined. If the basal ice temperature is above the pressure melting point, there is net melting at the base and the ice is *warm-based* or *wet-based*. The ice slides over the substrate and may be separated from it by a thin layer of water. If the basal temperature is below the pressure melting point, the ice will be frozen to its bed and is *cold-based* or *dry-based*. The adhesive strength of the frozen glacier–bed contact is greater than the shear strength of the ice. An intermediate regime exists in which there is a balance between melting and freezing; the glacier slides but no excess meltwater is produced (Boulton, 1972b).

A cold climate, such as in Greenland and Antarctica, surrounds a glacier with cold air and consequently the glacier is likely to be cold-based. In contrast, a warm climate, such as in southern Alaska and western Norway, leads to summer thaws during which water percolates into the glacier or through it via tunnels, and the glacier is likely to be warm-based. As a result, warm-based glaciers are also called ***temperate*** glaciers, cold-based ones *polar*, and intermediate ones ***subpolar***, but note that these terms are not necessarily indicative of latitude and so can be misleading. Note also that different thermal regimes may exist in different parts of the same glacier. A glacier may be cold-based at high elevations near its source and warm-based near its snout (Fig. 11.1), or it may be warm-based up-glacier where the ice is thicker and cold-based near its terminus.

11.2.3 Types of glaciers

Modern glaciers fall into two main types.

1 *Ice sheets* and ice caps, or continental glaciers, cover large areas and are unaffected by topography (Fig. 11.2). The ice sheets of Antarctica and Greenland today extend over 12.5×10^6 km^2 and 1.7×10^6 km^2, respectively (Fig. 11.3). Ice caps cover less than 50 000 km^2, for example Vatnajökull in Iceland extends over 12 000 km^2 (Flint, 1971). *Outlet glaciers* form where the ice moves through mountains. A ***marine ice sheet*** is an ice sheet that is ***grounded*** below sea level, that is its base is in contact with the sea bed and submerged. Large floating ***ice shelves*** exist locally in polar latitudes where an ice sheet extends into the sea. Ice shelves 200–1300 m thick border parts of the Antarctic coastline today (Fig. 11.3). The effects of erosion and deposition by an ice sheet are geograpically widespread; deposits may be laterally continuous for hundreds of miles.

2 *Valley*, mountain or alpine, *glaciers* are constrained by topography. They originate in highland icefields or cirques and typically form a dendritic pattern similar to that of river systems. Because more surface area of ice is in contact with bedrock and because debris falls on to the glacier surface from adjacent slopes, valley glaciers carry proportionately more debris than ice sheets. Deposits of valley glaciers are geographically restricted and tend to show more rapid facies changes than those of ice sheets.

All types of glaciers comprise three zones. The *basal* or ***subglacial zone*** is the lower part of the glacier which is influenced by contact with the bed (Fig. 11.2). Both erosion and deposition can take place within this zone. The ***englacial zone***, or interior of the glacier, is primarily a region of passive transport of entrained sediment. The ***supraglacial zone*** includes the seasonally influenced upper surface of the glacier as well as detached masses of stagnant glacial ice.

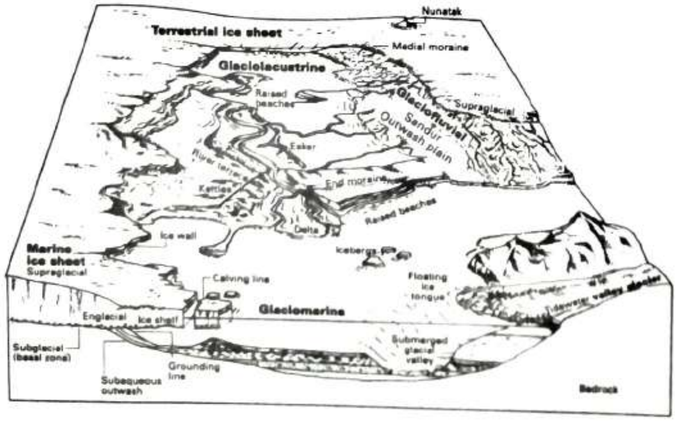


Figure 11.2* Types of glaciers, glacial environments and glacial landforms. *See footnote on p. 454.

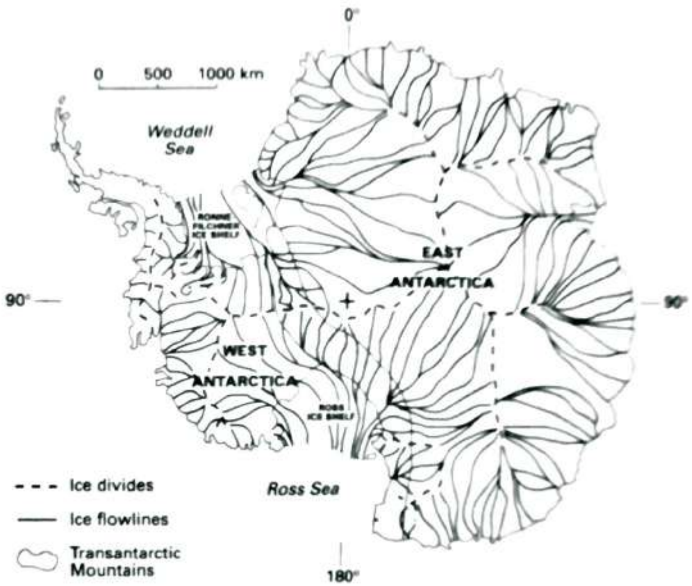


Figure 11.3 Geographic map of Antarctica showing East and West Antarctic ice sheets separated by Transantarctic Mountains, and ice drainage for the continent (modified after Drewry, 1983).

11.3 Processes

Processes associated with ***moving ice*** are unique to glacial environments. ***Moving water, wind*** and ***resedimentation*** also play important roles. Moreover, ***low temperatures*** have an overriding, if sometimes subtle, effect upon many sedimentary processes. For example, cold water is more viscous, sediment may be frozen, biological activity is changed, and frost action is important (Sect. 11.4.7).

11.3.1 Mechanics of ice flow

Ice moves by two main processes: ***internal deformation*** or ***creep***, and ***basal sliding***. These processes affect the ways that debris is eroded, transported and deposited by glaciers. Creep occurs at all levels within a glacier and consists of slippage within and between ice crystals, due to stress caused by the weight of the ice. Creep velocity depends primarily upon ice thickness and slope angle. Basal sliding is concentrated in the basal layers where three main processes occur: (i) ***enhanced basal creep*** is caused by local stress concentrations within the ice due to bed irregularities, (ii) ***regelation*** (pressure melting and refreezing) occurs as ice melts on the upstream side and refreezes on the downstream side of an obstacle, enhancing movement of the ice mass, and (iii) ***slippage over a water layer*** increases ice velocity because of the reduction in friction caused by a layer of water between rock and ice. Another important mechanism of glacier flow is movement within soft, deformable substrate sediment; this can contribute 90% of the total basal movement of the glacier (Boulton, 1979).

Total surface ice velocity is the sum of velocity due to creep and basal sliding. In warm-based glaciers, basal slip is generally dominant. In cold-based glaciers, creep and substrate deformation usually dominate. Longitudinal flow within an ideal glacier follows vectors which are inclined to the glacier surface. In general, there is a downward velocity component within the accumulation zone, and the flow is described as extending (Fig. 11.1). In the ablation zone, there is an upward component of movement and the flow is compressive. Local variations caused by topography are superimposed upon this pattern. The direction of ice flow in an ice sheet depends upon the size and shape of the ice mass (Fig. 11.3). Ice streams are narrow zones within an ice sheet along which flow is much faster than in adjacent broader areas. The distribution of flow lines (slip lines) within ice affects dispersal and transport of englacial debris.

Normal velocities for valley glaciers vary between 1 metre year^{-1} and 7 kilometres year^{-1} (Boulton, 1974; Drewry, 1986). Temporary, catastrophic increases in basal sliding velocity result in glacial surges when velocities may increase 10-, 100-, or even 1000-fold.

11.3.2 Glacial erosion

Ice can be an extremely powerful agent of erosion. Even the lowest measured rates of abrasion alone are twice the world average for erosion of lowland river basins (Boulton, 1974). Thus both large erosional landforms and huge volumes of sediment are produced during glaciations.

Ice erodes primarily by plucking and abrasion. Cold-based ice can only pluck. Large slabs can be quarried, plucked and carried forward by the glacier if weaknesses exist within the substrate and/or the cohesion of the subglacial material is overcome (Boulton, 1972a).

In the presence of meltwater, plucking or debris entrainment may occur either by plastic flow of ice around a block until the block is encased and carried away within the ice, or in association with regelation: debris is frozen on to the glacier base in the lee of bedrock protuberances. Ice within the basal debris-rich zone below warm-based glaciers tends to be melted at the next bump of comparable size. Consequently, basal debris layers of warm glaciers are thinner than those of cold glaciers.

Below warm-based ice, debris carried in the basal layers abrades the bedrock producing *rock flour*, composed of fresh mineral fragments generally smaller than 100 μm (Sugden & John, 1988). Abrasion produces polished surfaces, striations and gouges (Fig. 11.4). Crescentic cracks and gouges form by crushing or fracturing of bedrock below debris-laden ice and can indicate ice flow direction (Fig. 11.4).

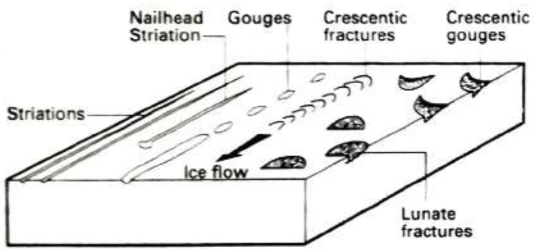


Figure 11.4 Forms of glacial abrasion (modified after Embleton & King, 1975; Shaw, 1985).

11.3.3 Glacial sediment transport

Debris may be carried within any part of a glacier, but the largest volume is generally within the basal zone. The basal zone is usually less than 1 m thick (Boulton, 1972a; Sugden & John, 1988), although it may be as thick as 15 m (Matanuska Glacier, Alaska; Lawson, 1979). Within it, sediment concentrations are variable, averaging 25% but ranging up to 90% by volume for warm glaciers (Drewry, 1986). Since much shearing and abrasion takes place here, the sediment may be layered but it is not sorted. Rock flour is abundant. Clasts are commonly striated and faceted but, due to abrasion, have higher roundness values than supraglacial debris (Fig. 11.5; Boulton, 1978). The development of striations depends primarily upon clast lithology (Kuhn, Melles *et al.*, 1993). Estimates of striated stone abundance range from 0.1 to 28% (Drewry, 1986). Clasts within the basal zone generally show a preferred orientation with their long axes parallel to flow and a small up-glacier dip (Fig. 11.6a;

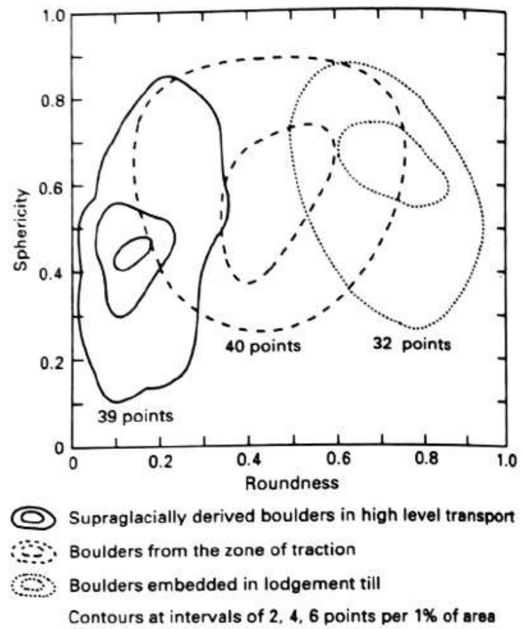


Figure 11.5 Positions of clasts in high level transport, from the zone of traction, and deeply embedded in lodgement till, in Krumbein's sphericity/roundness matrix. Clasts are taken from Breidamerkurjokull in southeastern Iceland (Boulton, 1978).

Lawson, 1979). Under highly compressive flow, long axes may be transverse (Boulton, 1971).

Englacial debris is, for the most part, highly dispersed. It is derived from both the supraglacial and subglacial zones and, since it is not usually modifed during transport, shows textures characteristic of those zones. Supraglacial debris is derived either directly from nearby slopes, in which case the clasts are angular (Fig. 11.5), or from en- or subglacial positions and brought to the ice surface along upward-tilted flowlines. As the debris moves passively on the ice, clast shape may be modified slightly by weathering or abrasion by supraglacial meltwater.

11.3.4 Glacial deposition

Sediment is deposited directly from either moving (active) or stagnant (passive) ice by very different processes. Below moving ice lodgement processes dominate – that is, plastering of basal debris on to the substrate. Above or below stagnant ice deposition occurs during melting.

Lodgement till forms: (i) when the frictional drag on clasts in the basal zone is equal to or less than the tractional force exerted upon them by the moving ice; and (ii) when pressure melting below moving ice allows small particles to be freed and lodged

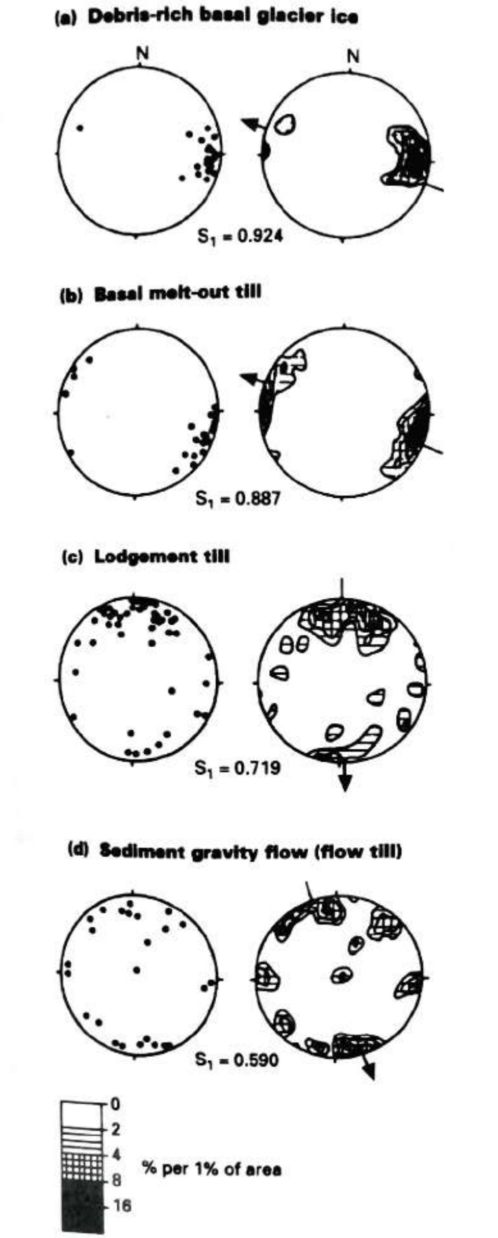


Figure 11.6 Schmidt equal-area projections of clast orientations in glacier ice, tills, and a sediment gravity flow deposit. Arrows indicate direction of ice flow (a,b,c) or of sediment flow prior to deposition (d). Data sources: (a), (b), (d) from Matanuska Glacier, Alaska (Lawson, 1979); (c) from Skalafellsjökull, Iceland (Sharp, 1982). S_1 eigen value for each sample summarizes fabric strength (from Dowdeswell, Hambrey & Wu, 1985).

or plastered on to the glacier bed (Boulton, 1975; Sugden & John, 1988). The latter process is most common beneath warm-based ice. Both processes are affected by irregularities in the subglacial bed, leading to variations in the distribution and thickness of lodgement till.

Melt-out till forms subglacially or supraglacially as stagnant ice melts. Since it is largely a passive process, fabrics from the debris-laden ice are preserved, with slight modification because of volume loss and meltwater movement. In very cold, dry regions, sublimation till rarely forms from debris-laden ice when the ice slowly changes directly into water vapour (Shaw, 1985).

11.3.5 Glaciotectonism

Ice movement and static ice loading can deform consolidated or unconsolidated substrate (Aber, Croot & Fenton, 1989). Deformation occurs where stresses transferred from the glacier are greater than the strength of the stressed material. Ice exerts both drag or shear stress, due to its movement over the bed, and vertical stress, due to its weight. Both ductile and brittle deformation occur and resultant structures are very varied. Much deformation is caused by a lateral pressure gradient. For example, folded and faulted substrate with minor displacement is caused by ice-push at the glacier terminus or by shearing along the glacier sole. Large blocks (up to kilometres across) can be transported long distances from their source and deformed, if, after subglacial plucking, they are carried up into the ice by compressive flow and then moved with it (Moran, 1971).

11.3.6 Related processes: water, resedimentation, wind

Meltwater is a powerful agent of erosion, transport and deposition of glacial sediment, particularly in the region of the glacier terminus and around warm-based ice. Water is present in sub-, en-, supra- and proglacial environments. Because of high sediment load, seasonal fluctuations, and sudden drops in velocity (e.g. where confined flow from a subglacial tunnel enters standing water), sedimentation and aggradation rates can be very high.

Subglacial water flows either in channels or in sheets. Channels are incised either downward into the substrate ('Nye channels') or upward into the ice ('Röthlisberger channels'; Shaw, 1985; Drewry, 1986). Nye channels are stable under moving ice, whereas Röthlisberger channels are only stable under stagnant ice. When subglacial channels cannot accommodate the discharge (Shaw, 1985), or when the basal water pressure is equal to or greater than the ice pressure (Drewry, 1986), subglacial sheet flow occurs. The importance of this sheet flow is controversial. Walder (1982) maintained that sheet flow is probably only quasi-stable at thicknesses <4 mm, whereas Shaw (1985) proposed that large erosional forms can be cut into both bedrock and overlying ice by high magnitude subglacial sheet floods. Ponding of water in subglacial depressions can produce subglacial lakes (Shreve, 1972; Drewry, 1986).

In proglacial environments water exists in rivers, lakes and oceans. One unique feature is catastrophic floods, called *jokulhlaups* (Icelandic). These are caused by sudden discharge of waterbodies that were ponded by ice and can move enormous volumes of sediment (see Sect. 11.4.4).

Sediment gravity flows of all types (see Sect. 10.2.3) are important in most glacial environments. They vary primarily in their water content and exhibit a continuum from slope failure through plastic to fluid behaviour (Lawson, 1979; Fig. 10.3). They are particularly common at and near the glacier terminus in both subaerial and subaqueous settings. In the terminal region of the Matanuska Glacier, Alaska, Lawson (1979) estimated that 95% of the sediments are the result of resedimentation.

Strong *winds* are common in glaciated regions because large ice caps influence atmospheric circulation. Aeolian activity is enhanced by sparse vegetation and the availability of easily eroded sediment. Winds erode and deposit sediment, particularly within the zone fringing the glacier terminus, although *loess* (wind-blown silt) may be distributed over a belt hundreds of kilometres wide and approximately parallel to the ice margin (Brodzikowski & van Loon, 1991).

11.4 Modern glacial environments and facies

Glacial and related environments include a variety of subenvironments each characterized by a distinctive set of processes and sedimentary facies. The *glacial environment* proper embraces all areas which are in direct contact with glacial ice, and includes the *basal* or *subglacial*, *englacial* and *supraglacial* zones (Sect. 11.2.3; Fig. 11.2). The *proglacial environment* occurs around the margin of the glacier and includes: (i) the *ice-contact zone* immediately adjacent to the glacier, where buried stagnant ice (ice no longer flowing with the glacier) is commonly preserved; (ii) *glaciofluvial*; (iii) *glaciolacustrine environments*, which receive glacial meltwater; and (iv) the *glaciomarine environment*, in which glacier ice floats over, or is adjacent to, the sea and/or glacial meltwater influences the marine water structure. Overlapping with the proglacial environment is the *periglacial environment*, which is not directly affected by ice but is influenced by distinctive climatic zones adjacent to an ice sheet.

In the following discussion, descriptions of each glacial environment and facies are based as far as possible upon studies of modern examples, which provide information about processes and spatial configuration of deposits. Sedimentary facies details also come from descriptions of selected Pleistocene deposits. Obviously, some inferences are made here: once the ice has melted we cannot be certain that we know the conditions under which the deposit formed. However, inferences based upon facies analysis of Pleistocene sediments and pre-Pleistocene glacial rocks give important information about stratigraphy and facies preservation and thus provide a valuable contribution to the reconstruction of modern environments and processes, particularly those that are difficult to reach today.

11.4.1 Subglacial zone

The relative importance of ice and water as agents of erosion, transport and deposition within this zone depends upon the thermal regime of the ice and specific position below the glacier. Some direct observations have been made below glaciers, despite obvious logistic difficulties (Kamb & La Chapelle, 1964; Boulton & Vivian, 1973).

Erosional features formed below glaciers depend upon the nature of the substrate. On bedrock surfaces, striae, grooves, crescentic marks (Fig. 11.4) and *roches moutonnées* form beneath warm-based glaciers. Where the substrate is unlithified, deformation of unconsolidated subglacial sediment can contribute significantly to glacier flow, and flutes may form at the ice–substrate interface (Boulton, 1979). This subglacial deformation is characterized by longitudinal extension superimposed on simple shear, resulting in folding, flattening, attenuation and boudinage of original structures. As deformation continues, the till or substrate sediment becomes horizontally laminated and eventually homogenized to produce massive till. Deformation intensity decreases with increasing depth (distance below the glacier base), but may be concentrated along décollement surfaces (Boulton, 1987a). Thus *deformation till* describes weak rock or unconsolidated sediment detached from its source, its primary structures deformed or destroyed, and to which foreign material has been added (Elson, 1989).

Lodgement and melt-out till are important subglacial deposits. Both occur as moraine – that is, ridges or mounds of debris deposited or pushed up by a glacier. Moraine formed subglacially may be fluted, shaped into drumlins (streamlined hillocks generally elongated parallel to ice flow), or form transverse ridges. Lodgement and melt-out till are both composed predominantly of diamicton, although lodgement till may be finer grained due to comminution of debris in the basal zone of the ice. *Lodgement till* is commonly structureless, thrusting and shearing within and below the basal ice causing it to be very compact and sometimes fractured along shear planes and joints that dip up-glacier (Fig. 11.7). Clasts within the till, and deformed sedimentary inclusions if present, are generally orientated so that their long axes are parallel to ice movement, with a gentle up-glacier dip (Fig. 11.6c; Dowdeswell, Hambrey & Wu, 1985; Dowdeswell & Sharp, 1986). Striated and faceted clasts are common as well as bullet-shaped ('flat-iron' or 'elongate pentagonal') clasts positioned with their blunt end down-glacier. Lodgement till usually has a sharp erosional basal contact and forms widespread units up to only a few metres thick (Dreimanis, 1989). However, individual units,

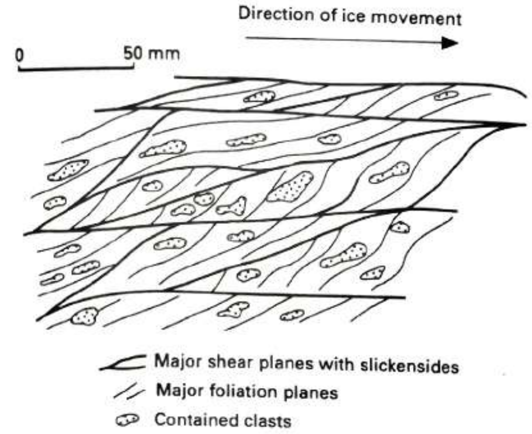


Figure 11.7 Schematic diagram showing the shear structures in lodgement till which may result from stress beneath moving ice (Sugden & John, 1988, modified after Boulton, 1970).

sometimes petrographically distinct, may be superimposed upon one another during a single glacial cycle (N. Eyles, Sladen & Gilroy, 1982). Internal erosion surfaces, sometimes delineated by stone (boulder) pavements, within lodgement till probably form by lodgement, deformation or erosional processes (Hicock, 1991).

Melt-out till is also commonly structureless, but it can contain subhorizontal laminae with gradational contacts derived from debris stratification in the ice, and/or lenses, layers or pods of sorted sediment deposited by escaping meltwater (Lawson, 1981; Shaw, 1982, 1985). The sorted layers may drape over larger clasts due to volume loss as the ice melts. Clasts are orientated parallel to ice movement but usually show a lower dip than in lodgement till due to settling during melt-out (Fig. 11.6b). Because melt-out till is deposited from stagnant ice, it may contain undeformed clasts of unlithified sediment (Shaw, 1982, 1985). Melt-out till units are usually up to a few metres thick, but may be stacked (Dreimanis, 1989). However, Paul and Eyles (1990) emphasize that melt-out till is only of local significance, and is commonly deformed by shear or water escape after deposition.

Subglacial meltwater deposits fill channels aligned subparallel to ice movement which may be tens of kilometres long and are either cut into substrate as *tunnel valleys* or cut upward into the ice to form *eskers*. The former are associated with active, moving ice and the latter with active or passive ice. Tunnel valleys commonly have a trough-like form, an undulating longitudinal profile, and can be a few hundred metres deep and up to 3 km wide. They are filled with sand, gravel, silt and clay and may terminate in a sand–gravel fan at the ice margin (Woodland, 1970; Ehlers, 1981; Patterson, 1994). Eskers are continuous or segmented ridges which can be tens of metres high and hundreds of metres wide (Warren & Ashley, 1994). They are typically composed of massive or stratified boulder gravel to sand with sedimentary structures similar to fluvial deposits (see Sect. 3.2.2) except that, because flow was constrained within a tunnel, antidunes are absent. Scour and fill structures are common. Esker sediments may show fining-upward cycles up to a few metres thick; they may also fine outwards, and show faulting and rotation of beds in the marginal zones (Banerjee & McDonald, 1975; McDonald & Shilts, 1975; Shaw, 1985; Brennand, 1994).

Catastrophic subglacial meltwater sheet floods have recently been recognized as important agents of erosion and deposition. Their effects include areas of giant flutings, tunnel valleys, scoured bedrock tracts, and possibly the deposition and shaping of drumlins (Shaw, Kvill & Rains, 1989; Rains, Shaw *et al.*, 1993).

11.4.2 Supraglacial zone

Many varied subenvironments can exist on top of glaciers, including thick, living vegetation. Once ice is covered by a layer of debris it is protected from solar radiation and so the rate of ablation is reduced. Deposition can occur from melting ice, running or standing water, mass movement or aeolian activity. However, all these sediments have a low preservation potential.

Supraglacial facies formed by passive melting and by mass movement are the most important, amidst a variety of facies types (Fig. 11.8; Brodzikowski & van Loon, 1991). *Supraglacial melt-out* (ablation) *till* is most abundant in areas of mountain glaciation because of the proximity of debris derived from valley walls. It is commonly intercalated with proximal outwash deposited by meltwater (N. Eyles, 1979). Rarely, *sublimation till* forms and delicate englacial structures are preserved distorted only by loss of ice volume (Shaw, 1985, 1989). Various types of *sediment gravity flow deposits*, similar to those described in the proglacial zone, are abundant. This resedimentation is enhanced by the presence of meltwater and an irregular supraglacial topography. Deformation structures, for example collapse features, normal faults and slumps, are ubiquitous in supraglacial facies because these sediments overlie ice that subsequently melts.

11.4.3 Ice-contact proglacial zone

This zone is characterized by very irregular and hummocky topography. End or push moraines parallel the glacier front, whereas differential melting of buried ice blocks and slumping of wet till produce irregular highs and lows, referred to as ice-disintegration topography (Fig. 11.8). Important processes include glaciotectonism adjacent to the glacier snout, resedimentation by sediment gravity flow and mass movement, melting of stagnant ice, and erosion and deposition by running water.

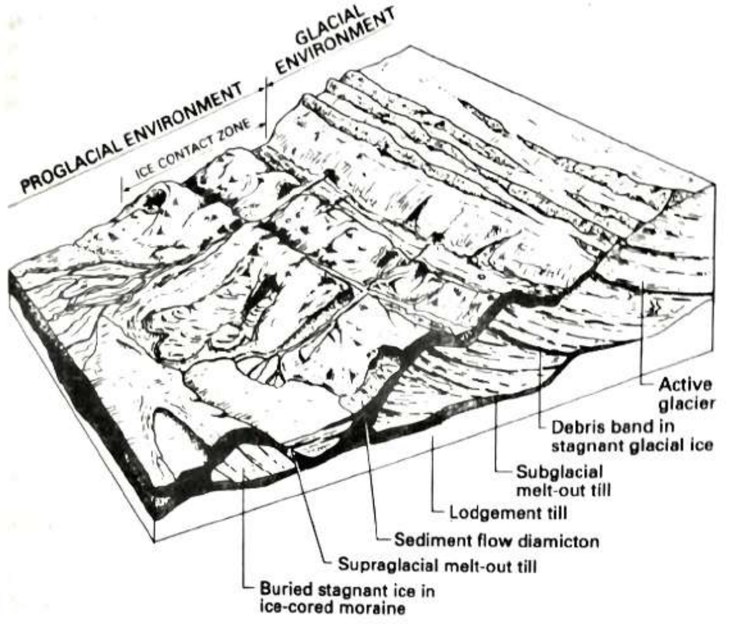


Figure 11.8* Sedimentation in the supraglacial and ice-contact proglacial zones of a slowly retreating, subpolar glacier. Subglacial material is brought into the glacier by basal freezing and thrusting. This debris is released at the surface as the enclosing glacial ice gradually melts. The till is rapidly reworked by flowing meltwater, and may slump and flow downslope to form flow diamictons. Diamicton beds can be intercalated with proglacial stream or lake deposits, and may be extensively reworked (modified from Boulton, 1972a). *See footnote on p. 454.

Proglacial deformation is characterized by compressive structures, for example folding and thrust faulting at various scales. Till or proglacial outwash sediments may be deformed. Glaciotectonic structures are probably most common in glaciers frozen to their bed in the terminal zone (Sugden & John, 1988), but a glacier in the Swiss Alps formed a 5–10 m high push moraine complex composed of deformed older glacial sediments when it readvanced about 100 m between 1971 and 1986 (Fig. 11.9; Eybergen, 1987).

Resedimentation processes are facilitated by local slopes, copious meltwater, and abundant rapidly deposited often clay-rich sediment. Most flows are initiated by backwasting of slopes composed of sediment and stagnant ice. Four types of sediment gravity flows have been recognized at the terminus of the Matanuska Glacier, and the deposits range from massive diamicton to massive to graded sand to sandy silt (Lawson, 1979). Diamictons formed in this way have been called *flow till* (Hartshorn, 1958; Boulton, 1972a), although this term is somewhat misleading since deposition is primarily by resedimentation, which strictly is a non-glacial process (Lawson, 1982). Clast fabrics from sediment gravity flow deposits differ from lodgement and melt-out diamictons in showing either no preferred orientation, a multimodal pattern, or near horizontal long axes aligned parallel to flow (Fig. 11.6d). Other forms of mass movement include talus cones, which accumulate at the base of ablating ice-cored slopes, and slope failure by slumping or spalling (Lawson, 1979; Shaw, 1985).

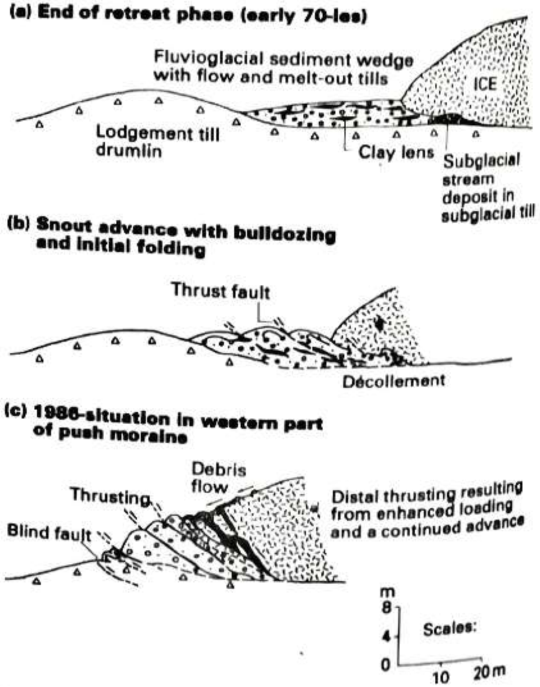


Figure 11.9 Selected stages of a schematic model of push moraine formation at the snout of the Turtmannglacier, Switzerland (modified from Eybergen, 1987).

Melt-out till (Sect. 11.4.1) is deposited in the ice-contact zone, and meltwater flows generally deposit well-sorted silt to silty sand which may be massive, normally graded, parallel stratified, or cross-stratified if deposited over a change of slope. Sorted meltwater deposits are often interbedded with sediment gravity flow deposits (Lawson, 1979).

11.4.4 Glaciofluvial environments

An outwash plain, or *sandur* (Icelandic, pl. *sandar*), generally forms in front of the glacier terminus, beyond the ice-contact zone (Fig. 11.2). Glaciofluvial environments and facies are almost identical to those of non-glacial braided fluvial systems (see Sect. 3.3.2). Glaciofluvial systems differ in that they are affected by fluctuating positions of the ice margin and by buried and transported blocks of ice; they show strong seasonal and weather-dependent discharge variations; and they carry a high sediment load, typically lack vegetation, and contain rare till clasts (N.D. Smith, 1985). Polar sandar differ from temperate sandar in that they grow more slowly due to lower energy and lower magnitude meltwater discharges (Rains, Selby & Smith, 1980).

Sandar may form a wide plain or outwash fan (Fig. 11.2) or may be restricted by topography to form a valley fill (valley train). Outwash fans are typically subdivided into a proximal zone, where flow is confined to a few main channels that are relatively deep and narrow, an intermediate zone, with a complex network of wide, shallow, distinctly braided, shifting channels, and a distal zone, in which flow in shallower, ill-defined channels may merge to form a single sheet during periods of high discharge (see Fig. 3.31). In the proximal zone the sandur may bury patches of stagnant ice, sometimes forming obstacle marks in surrounding sediment (Russell, 1993). When the ice melts, kettle-holes or pits form that may be up to 40 m wide and 6.6 m deep (Price, 1971), and faults may form in the sedimentary strata (McDonald & Shilts, 1975). During periods of low discharge, large inactive areas on the sandar may become vegetated or subject to strong aeolian influence. Although braided streams predominate on outwash fans, meandering and anastomosed channel patterns are also present, particularly in distal areas where silt and clay are abundant (Boothroyd & Ashley, 1975; Boothroyd & Nummedal, 1978; N.D. Smith, 1985).

Facies in the proximal zone are dominated by massive to crudely horizontally bedded gravels, which are locally cross-bedded, with thin and sparse fine-grained units. Downstream, these give way to sandier deposits with abundant tabular cross-beds as well as ripple-drift cross-lamination, trough cross-beds and horizontal stratification (N.D. Smith, 1985). Clast size and gradient decrease downstream (Boothroyd & Ashley, 1975). However, downstream-fining is most pronounced when aggradation rate is high (N.D. Smith, 1985).

Evidence for highly fluctuating discharges shows through frequent vertical changes in grain size and sedimentary structures, abundant scour surfaces, reactivation surfaces, and common fine suspension deposits (clay, silt, fine sand) that also occur as intraclasts (N.D. Smith, 1985). Jökulhlaups transport huge volumes of sediment, and produce distinctive deposits including sand and gravel cross-beds up to 10 m high deposited in channels, graded silt and sand deposited by turbidity currents (Baker, 1973; N.D. Smith, 1985), and sequences of massive homogeneous gravel deposited by hyperconcentrated flows overlain by erosional surfaces, boulder lags and cross-bedded and horizontally bedded sand and gravel (Maizels, 1989).

Vertical facies profiles of glaciofluvial deposits have been characterized by three different types (Miall, 1977, 1983): Scott (proximal), Donjek (medial) and Platte (distal). Miall (1983) emphasized the cyclic nature of Donjek type deposits, whereas N.D. Smith (1985) questioned whether cyclicity is typical of vertical profiles of glaciofluvial outwash. Cyclicity on a scale of several metres in sandar may be rare because of continuously fluctuating discharge, despite Miall's claims. However, some cyclicity would be expected on a larger scale as the outwash facies respond to fluctuations in the position of the ice margin. A change from dominantly braided, steep, low-sinuousity, high width/depth ratio bedload channels to deeper, more sinuous, low-gradient, single thread channels occurs in response to long-term deglaciation (Maizels, 1983). Coarsening-upward sequences 10 m or more thick may form as a result of glacial advance, while fining-upward sequences would result from glacial retreat (Miall, 1983; Ashley, 1988).

11.4.5 Glaciolacustrine environments

Lakes are very common in glaciated landscapes. They form as a result of damming of river courses by ice, formation of irregular topography by glacially deposited or eroded landforms, or on a larger scale by the reversal of regional slope due to isostatic depression caused by glacial build-up. Unique to glacial lakes is the fact that they receive a substantial proportion of their annual water and sediment budgets from meltwater (Ashley, 1989), and it is these *'glacier-fed' lakes* that are discussed here. Glacier-fed lakes are subdivided into *ice-contact* lakes, in which some portion of the lake is in direct contact with glacial ice, and *non-contact* or *distal* lakes, which are located some distance from the ice and are fed by outwash streams.

The main factors that affect physical processes and sedimentation in glacier-fed lakes are proximity to the ice, thermal stratification of the lake water (see Sect. 4.3; Fig. 4.2), and seasonality of inflowing meltwater and ice cover (Ashley, 1989). Thermal stratification is best developed in non-contact lakes where it is most pronounced and stable in mid-summer (see Fig. 4.3). Mixing of the water layers, or *'overturning'*, can occur during the autumn (see Sect. 4.3). In ice-contact lakes, thermal stratification is inhibited by the continuous supply of meltwater at 0°C. Glacier-fed lakes may also show sediment, or density stratification due to a gradual increase in suspended sediment content with depth (Gustavson, 1975).

In ice-contact lakes, meltwater streams may enter near the surface of the lake or at or near the lake bottom from englacial or subglacial channels. In non-contact lakes, all meltwater enters at the lake surface. As they move into the lake, these cold influent streams form overflow, interflow, or underflow plumes depending upon the relative density of the influent versus the standing lake water (see Fig. 4.5; Sect. 4.4.2). Meltwater input into glacial lakes shows marked fluctuations over hours and days or weeks and months. This causes pulses or surges of inflow which overprint the quasi-continuous flows into the lake (N.D. Smith & Ashley, 1985).

Glaciolacustrine environments are subdivided into marginal (proximal) and basinal (distal) ones. Deposition in marginal subenvironments is dominantly by mass movement and underflows. Deltas form at the margins of both ice-contact and non-contact lakes (Fig. 11.10). Gilbert-type deltas (see Fig. 6.22) form where coarse-grained debris is supplied under relatively high-energy conditions to a deep lake margin (see Sects 4.6.2 & 6.5). Where a lower energy, finer-grained system enters a shallow lake basin, a delta with gently dipping foresets (<20°) will form. Density underflows typically build lobes on the delta front, and the deposits show evidence of rapid sedimentation and dewatering (Ashley, 1988). Sequences of ripple drift and draped lamination characterize regions between the lobes (Gustavson, Ashley & Boothroyd, 1975).

Subaqueous fans are built where meltwater enters a lake at or near the base of the ice. Coarse gravel typically forms the

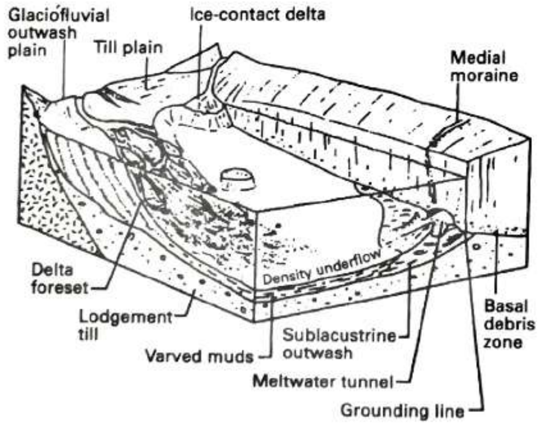


Figure 11.10* Sedimentation in a glacial lake. The three sources of sediment shown here are a sublacustrine outwash fan, an ice-contact delta, and a delta supplied by a glacial meltwater stream. Most of the fine-grained sediment is carried in suspension in a density underflow, and is deposited on the lake bottom. *See footnote on p. 454.

proximal core of the fan which is surrounded by sand. Some sands are massive or show chaotic bedding and/or dewatering structures, and may contain scattered large clasts. They were deposited by sediment gravity flows, filling channels near the apex or forming sheets farther out on the fan. Other sands are cross-bedded, ripple cross-laminated or parallel laminated, deposited by traction currents (Rust & Romanelli, 1975; Rust, 1977; Ashley, 1988).

Ice cliffs or ice ramps (where lake water covers stagnating ice) exist along ice-contact lake margins (Fig. 11.10; Holdsworth, 1973; Barnett & Holdsworth, 1974). In both cases debris accumulates and locally forms a sublacustrine moraine as the ice calves or melts; massive to bedded diamicton is interbedded with massive or cross-bedded sand and mud. In large Late Pleistocene glacial lakes which occupied moderate- to high-relief basins, debris flows deposited thick (<10 m) and extensive (<3 km) diamicton units (N. Eyles, 1987). Debris was contributed by slope failure along lake margins as well as by glaciers. Near the margins of large non-contact lakes, wave and storm reworking may deposit hummocky cross-stratified sand (N. Eyles & Clark, 1988). Shoreline glaciolacustrine sediments can also be deformed by pressure from lake ice (Brodzikowski & van Loon, 1991).

Basinal subenvironments are similar in all glacier-fed lakes and sedimentation processes are more regular because meltwater discharge fluctuations are dampened by the lake body (Ashley, 1988). The effects of seasonal fluctuations are most pronounced here and rhythmites and varves form (see discussion in Sect. 4.8; Figs 4.14 & 4.15). Sediment is deposited from suspension and by density currents, turbidity currents and mass flow (see Sect. 4.6.3). A typical resultant succession contains rhythmically laminated or varved fine sand to clay interrupted by turbidites with complete or incomplete Bouma sequences (see Sect. 10.2.3), current rippled sand deposited by bottom-hugging flows, and massive or graded diamicton beds deposited by mass flows. Soft-sediment deformation, for example loading and slumping, is quite common (Ashley, 1975; Sturm & Matter, 1978; Smith & Ashley, 1985).

A distinguishing feature of ice-contact lake deposits is the presence of ice-rafted debris which interrupts regularly bedded or laminated sediment (Fig. 11.11). Dropstones bend, penetrate, ruck and rupture laminae (see Table 11.2). Dump structures, conical mounds of sediment, form when large quantities of debris are released by the break-up or overturning of icebergs (Thomas & Connell, 1985). Clots of frozen, poorly sorted sediment dropped from floating ice and embedded in laminated sediment are called till pellets, clots or clasts (see Table 11.2; Ovenshine, 1970).

Lastly, biogenic activity may modify glaciolacustrine sediment, for example through burrowing (Ashley, 1975; Gibbard & Dreimanis, 1978). It may also contribute to it: pelletization enhances the rate of suspension settling in some glacial lakes (N.D. Smith & Syvitski, 1982).

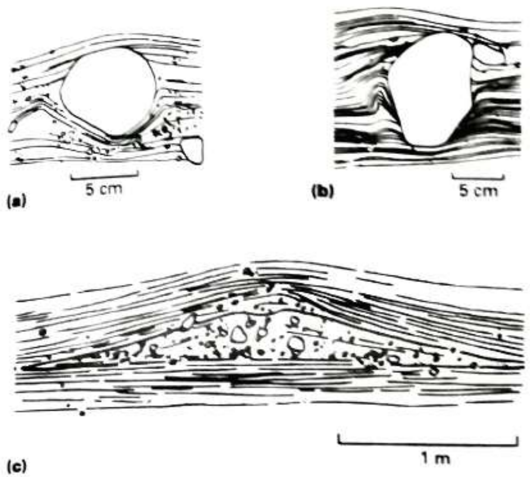


Figure 11.11 Dropstone structures (a, b) and dump structure (c) from Pleistocene glacial lake deposits (Thomas & Connell, 1985).

11.4.6 Glaciomarine environments

In marine settings glacial processes are superimposed upon normal marine processes, their effectiveness diminishing away from the ice terminus. Dropstones are the most obvious indicator that ice was present, as well as iceberg keel scour marks (see below; Andrews & Matsch, 1983; Drewry, 1986). Glacial thermal regime affects sedimentation, as discussed below. Ice sheets enter the sea as ice cliffs (walls, faces), floating ice shelves or ice tongues. Valley glaciers enter with tidewater fronts or floating ice tongues (see Fig. 11.2; Drewry, 1986). Ice cliffs and tidewater fronts form where ice terminates at the *grounding line* or *zone* at which ice entering a waterbody comes afloat.

Glaciomarine environments can be subdivided into: (i) subglacial zone; (ii) grounding line zone; (iii) ice shelf and ice tongues; (iv) fjords; and (v) open ocean, distal, or iceberg zone. The first four zones can only exist as far seaward as the continental shelf break, but the iceberg zone can extend over the continental slope and deep ocean floor.

The *subglacial zone* lies below sea level and landward of the grounding line. Processes of sediment transport and deposition are identical to those in the terrestrial subglacial zone described above (Kellogg & Kellogg, 1988).

Sedimentation in the *grounding line zone* varies with glacial thermal regime. Under polar conditions sediment supply is low because glacial erosion is less effective and meltwater is scarce or absent; grounding line sediment is derived only from the base of the ice. Where an ice sheet terminates in ice cliffs, waves erode the cliffs and bottom currents redistribute the finer sediment leaving only coarse ice-rafted debris and bio-clastic material (J.B. Anderson, Brake *et al.*, 1983; Domack, 1988; J.B. Anderson, 1989). Where ice shelves and tongues exist, most sediment is deposited near the grounding line. At the termini of ice streams, below which deformation till has been recognized, *diamict aprons* form by deposition of subglacial debris and downslope slumping of this sediment. These aprons (also called 'till deltas') are tens of metres thick and tens of kilometres long and include topset, foreset and bottomset beds (Alley, Blankenship *et al.*, 1989; J.B. Anderson & Bartek, 1992; Hambrey, Barrett *et al.*, 1992).

The grounding line zone of temperate glaciers is the one where the largest volume of sediment of all glaciomarine environments is deposited. As the ice terminus advances and retreats, grounding line sediments can be spread over a large area. At a tidewater front or ice cliff, sub-, en- and supraglacial debris melts out or falls during calving or continued melting. A subaqueous moraine forms, composed of a complex mixture of diamicton, gravel, sand and mud (Fig. 11.12a) that is commonly redistributed by mass movement or deformed glaciotectonically during ice advances. Submarine end moraines can be hundreds of metres high, kilometres wide and hundreds of kilometres long (King, Rokoengen *et al.*, 1991). Wedge-shaped deposits, termed *till tongues*, often occur on the distal side of continental shelf moraines and are interbedded with stratified glaciomarine sediment (King, Rokoengen *et al.*, 1991; Anderson & Bartek, 1992; King, 1993). Till tongues are commonly 25–50 m thick in their root area, may extend laterally for tens to hundreds of kilometres, and are characterized by acoustically incoherent seismic reflections (King, Rokoengen *et al.*, 1991). Those formed near the terminus of tidewater glaciers may be cut by tunnel valley deposits (King, 1993).

At the exit point of sub- or englacial streams, grounding line (subaqueous outwash) fans form when the ice terminus is stable (Fig. 11.12b; Cheel & Rust, 1982; Powell, 1988, 1990). As the incoming meltwater jet suddenly decelerates, a down-current fining of traction current deposits may form, namely imbricate gravels, fine granule gravels, and then sands which may be massive, horizontally or cross-stratified, inversely or normally graded, or ripple-drift cross-laminated. At a certain distance from the influx point, the jet will detach from the sea bottom, dump sediment which builds a gravel–sand bar, and form a buoyant plume. Sediment gravity flows may be generated from oversteepened slopes of the bar. Facies on grounding line fans resemble those of coarse-grained fan deltas (see Sect. 6.5) in that traction current deposits are abundant. However, dropstones may be present, as well as diamicton layers in inter-channel areas and deformation caused by melting of buried ice (Cheel & Rust, 1982). Should the ice terminus remain stable for a long period of time, a grounding line fan may aggrade to sea level and form an ice-contact delta with wave- or tide-dominated topset beds and slides and slumps on the delta front. Alternatively, if the grounding line migrates rapidly, sheet-like deposits form. In areas adjacent to the ice terminus

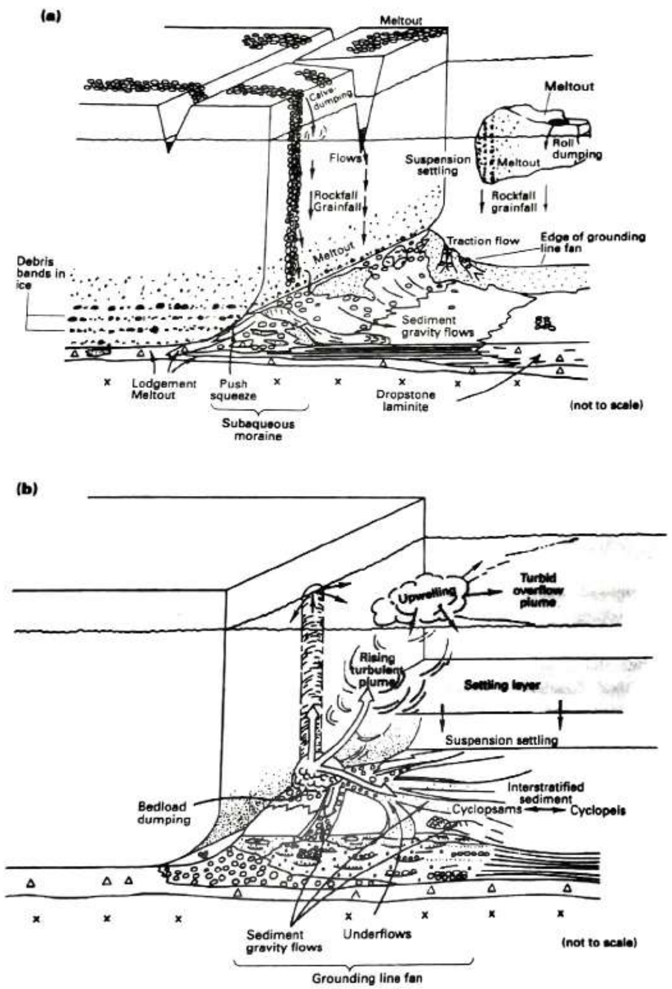


Figure 11.12 Processes acting, and sediment deposited, at and in close proximity to the grounding line zone of a wet-based tidewater glacier (a) where a subaqueous moraine is forming (the margin of a neighbouring grounding line fan interfingers with the moraine), and (b) with submarine discharge of a subglacial meltwater stream forming a grounding line fan (after Powell, 1988).

but distant from active calving or meltwater effluxes, low-energy conditions dominate and mud and ice-rafted debris accumulate. Laminae may result from seasonal fluctuations in meltwater and sediment supply (Powell, 1988, 1990).

Ice shelves and ice tongues are both composed of floating ice attached to and partly fed by land-based ice, but ice tongues are smaller and are long and narrow. The Ross Ice Shelf (see Fig. 11.3) is 420 m thick where drilled (Webb, Ronan *et al.*, 1979) and extends more than 400 km from the grounding line. Ice shelves dampen water movement and currents although thermohaline circulation occurs below them, particularly near the grounding line (Robin, 1979; Jacobs, Gordon & Ardai, 1979). Most sediment is deposited at or near the grounding line because significant basal melting occurs there (see above). Seaward of the grounding line, bottom currents may sort and winnow the sediment (Fig. 11.13).

Beneath floating ice shelves very little sediment is deposited because little debris exists within this ice and sedimentation may be inhibited by basal freeze-on. Massive diamicton with subtle stratification and/or a benthic foraminiferal assemblage

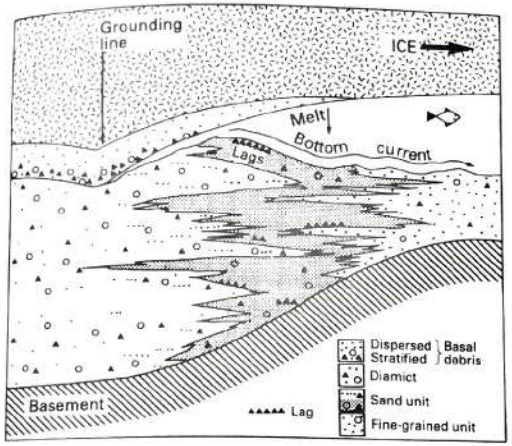


Figure 11.13 Sedimentary facies generated at the inner grounding zone of a floating ice shelf or glacier tongue undergoing grounding-line oscillations (due to sea-level or upstream mass balance changes) but without experiencing major erosion of sea-floor deposits (after Drewry, 1986). Note that below large polar ice shelves, current reworking is minimal and the proportion of massive diamict would be higher than shown here.

is thought to accumulate (Fig. 11.14; J.B. Anderson, Brake *et al.*, 1983; J.B. Anderson, Kennedy *et al.*, 1991). Sub-ice shelf sediments can be thicker if the ice shelf is fed by nearby mountains. Ice shelves only exist in high latitudes, and if ice shelf retreat is rapid, sub-ice shelf deposits may be virtually absent. Consequently, the vertical facies association of massive diamicton (subglacial or sub-ice shelf deposit) directly overlain by open ocean (iceberg) zone glaciomarine sediment with a significant biogenic component (see below) and without intermediate meltwater deposits, provides strong evidence for a polar or subpolar climate (J.B. Anderson, Kennedy *et al.*, 1991).

Fjord sedimentation occurs primarily during glacial retreat and differs from other glaciomarine settings only because of the shape of the basin. In temperate regions, submarine moraines and outwash form in the grounding line zone (Fig. 11.12); homogeneous muds containing ice-rafted debris commonly dominate elsewhere, with sandy gravelly mud on slopes and the outer shallow sill (Elverhøi, Lønne & Seland, 1983). Peak meltwater discharges and sediment gravity flows, commonly generated by slope failures, deposit coarse laminae. If the glacier terminates on land, sand and gravel outwash delta deposits prograde into the fjord and sand intertongues distally with mud, which may be laminated and rarely contains ice-rafted debris (Powell, 1981). Biogenic sedimentation in temperate fjords is minor because of high terrigenous sedimentation rates (Powell & Molnia, 1989). In polar and subpolar fjords, biogenic facies are more common and may dominate over terrigenous sediment with increasing distance from the fjord head (Syvitski, LeBlanc & Cranston, 1990; Domack & Ishman, 1993).

Laminated sediment, including *cyclopsams* (sand–mud laminae) and *cyclopels* (silt–mud laminae), is deposited from over- and interflow plumes 0.5 to several kilometres from the grounding line of tidewater glaciers in Alaska (Mackiewicz, Powell *et al.*, 1984). The couplets have sharp basal contacts and are normally graded (Powell, 1988). Two couplets form per day, due to the interaction of the suspended sediment plume with tidal currents (Cowan & Powell, 1990). Sedimentation rates can be extremely high, for example up to 15.4 cm dry sediment (cyclopsams) deposited in 19 h (Powell & Molnia, 1989). These marine couplets look similar to glaciolacustrine varves, but the gradational upper contact of the marine coarse laminae should distinguish them from freshwater varves in which the coarse–fine contact is commonly sharp because the two varve layers had different modes of deposition (see Sect. 4.8).

The *open ocean* or *iceberg zone* is affected by glaciers in several ways. Normal marine sedimentation is modified by deposition from icebergs and a varied supply of terrigenous sediment. Moreover, the continental shelf is typically depressed by the weight of the ice sheet and has a proglacial depression (Fig. 11.14) and an irregular bathymetry (Sect. 11.5.4).

Polar glaciomarine shelf environments are characterized by accumulation of biogenic material, favoured by low terrigenous

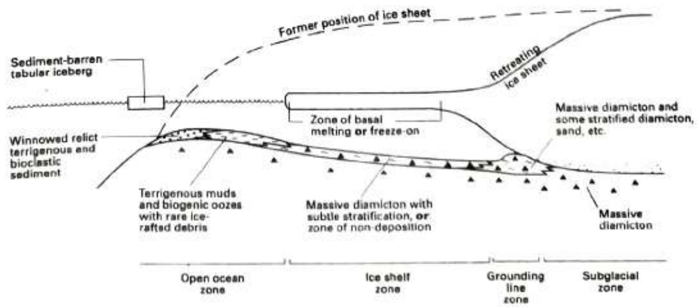


Figure 11.14 Schematic facies model for sedimentation at and around ice shelves associated with polar marine ice sheets during ice-sheet retreat. Note that continental shelf deepens towards continent due to weight of ice sheet. Not to scale, but with vertical exaggeration (adapted from J.B. Anderson, Kennedy *et al.*, 1991).

influx (Orheim & Elverhøi, 1981; Domack, 1988). The Antarctic shelf averages about 500 m deep (Dunbar, J.B. Anderson *et al.*, 1985). Below about 300 m are terrigenous muds and siliceous muds and oozes (biogenic silica concentrations of up to 30–40%), with minor ice-rafted debris (J.B. Anderson, Brake *et al.*, 1983). Carbonate bioclastic material or winnowed relict terrigenous sediment exist at shallower depths (Domack, 1988; J.B. Anderson, 1989).

In temperate regions, such as the Gulf of Alaska and the Beaufort and southern Barents seas, terrigenous sediment dominates (Barnes & Reimnitz, 1974; Vorren, Hald & Thomsen, 1984; Powell & Molnia, 1989). Bioclastic carbonate material is restricted to lag deposits on banks. Although these regions reflect sedimentation during an interglacial, the higher siliciclastic to biogenic ratio is probably indicative of temperate versus polar glaciated shelves.

Iceberg-rafting is common in shelf environments bordered by glaciated coasts. Isolated clasts and dump structures are similar to those described from glacial lakes (Fig. 11.11) (Gilbert, 1990). Icebergs have recently been recognized as important agents in scouring and reworking shelf sediment to depths as great as 500 m around Antarctica (Barnes & Lien, 1988). Surface features include furrows, typically a few metres deep and tens of metres wide, cut by iceberg keels (Fig. 11.15), and subcircular depressions (30–150 m across) formed by grounding of iceberg keels. Ice keels plough and rework shelf sediment to produce structureless diamicton which is termed 'ice-keel turbate' (Vorren, Hald *et al.*, 1983; Barnes & Lien, 1988).

During glacial maxima, large ice sheets transport voluminous unsorted sediment to the shelf break. This sediment moves by sediment gravity flows down the continental slope. On Antarctic continental slopes, slide and slump deposits near the top are replaced by non-stratified, non-sorted debris flows downslope and then by turbidites near the base (J.B. Anderson, Kurtz & Weaver, 1979; R. Wright & Anderson, 1982; R. Wright, Anderson & Frisco, 1983). On the northeast Newfoundland slope, thin but extensive shingled Quaternary debris flow lenses, derived from a line source at the shelf edge, are locally interbedded with stratified hemipelagic sediments (Aksu & Hiscott, 1992).

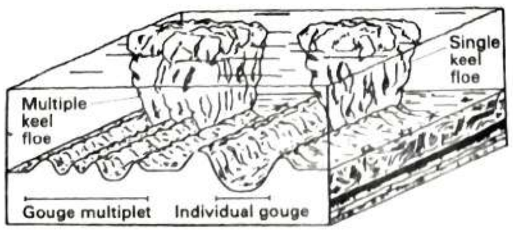


Figure 11.15 Scouring by single and multiple keel icebergs (floes) to produce an idealized ice gouge and gouge multiplet (from Barnes, Rearic & Reimnitz, 1984).

On the deep ocean floor, biogenic sedimentation with an ice-rafted component is typical during glaciation. Cores from the North Atlantic contain specific, 'Heinrich' layers, which have unusually high ratios of ice-rafted debris to Foraminifera shells and record massive influxes of icebergs and/or surges of the Laurentide ice sheet (Bond, Heinrich *et al.*, 1992). In the central Arctic Ocean, uniformly alternating layers of silty and sandy lutites accumulated over the last 5 My (Clark, Whitman *et al.*, 1980). A large submarine fan exists on the abyssal ocean floor of the eastern Weddell Sea (see Fig. 11.3). This fan probably formed during a prior, more temperate glacial setting of Antarctica, when terrigenous sediment supply was higher and before the shelf was lowered by glacial erosion and isostasy (J.B. Anderson, Wright & Andrews, 1986).

11.4.7 Periglacial zone

The periglacial zone is not directly affected by ice but is characterized by intense frost action and at least seasonally snow-free ground. It is frequently underlain by permanently frozen ground, *permafrost*. Aeolian, fluvial and lacustrine processes are also important in this zone.

Frost action leads to formation of many periglacial features, some of which may be preserved in the rock record (Washburn, 1980). Ice wedges and sand wedges, associated with polygons on the ground surface, form by frost cracking in a permafrost environment (Black, 1976). Ice wedges grow laterally and expand upwards, usually causing bedding in adjacent sediments to be bent upwards (Fig. 11.16a). After the ice melts, the wedge is filled with collapsed sediment. Although ice wedges grow preferentially in fine-grained material, they are best preserved in gravels. Sand wedges only form under arid conditions, and the frost cracks are filled repeatedly and incrementally by loose sand. Sand wedges can be distinguished from ice-wedge casts by the different orientation of the infilling laminae (Fig. 11.16b).

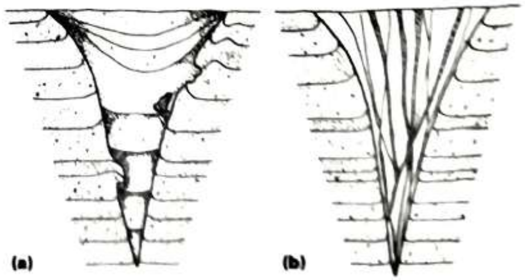


Figure 11.16 Diagrams of (a) ice-wedge-cast cross-section showing slump structures in cast and upturned beds in enclosing material; (b) sand-wedge cross-section showing vertical fabric in wedge and upturned beds in enclosing material (Black, 1976).

Irregular contortions, including deformation, folding and interpenetration of pre-existing strata and termed periglacial *involutions* (cryoturbations), are produced by frost action. Other rarely preserved periglacial features include fossilized pingos (Beuf, Biju-Duval *et al.*, 1971).

Soils of cold climates can show a variety of features caused by ice in the soil including brecciated zones where frozen soil has fallen into position after tabular bodies of ice melted (Retallack, 1990). Buried weathering profiles on Pleistocene tills show vertical grain size variations due to solution of carbonates, movement and alteration of clay minerals, and disaggregation and decomposition of less stable silicate minerals (Willman, Glass & Frye, 1966). Original depositional features may be fundamentally changed through drainage and consolidation, and fines are transported down through the profile by percolating water or lost by wind action (Boulton & Dent, 1974). Palaeosols on tills can be extremely useful for distinguishing different till sheets and interglacial periods (White, Totten & Gross, 1969).

Sand dunes, sand sheets and loess are the main periglacial aeolian deposits. Most dune-forming mechanisms are similar to those in non-glaciated regions (see Chapter 5). However, periglacial sand is usually more heterogeneous in texture and composition than in typical aeolianites, and the dunes migrate more slowly because of moisture freezing in the sand and burial of the dunes by snow (Ahlbrandt & Andrews, 1978). Sand sheets are composed of subhorizontal to low-angle cross-stratified sand and form in areas where limited availability of loose, dry sand inhibits dune formation (Lea, 1990). The inclusion of snow or ice layers and ice-cementation leads to distinctive deformation structures when the ice melts (Calkin & Rutford, 1974; Ahlbrandt & Andrews, 1978). Where fluvial and aeolian processes interact, sediments deposited as channel fills and aeolian sandsheets are truncated by laterally persistent, planar deflation surfaces where permafrost restricts the depth of aeolian reworking (Good & Bryant, 1985).

Loess is composed dominantly of silt, most of which is produced by glacial grinding (glacial flour). Both grain size and deposit thickness decrease with increasing distance from the source (Brodzikowski & van Loon, 1991), whereas the roundness of quartz silt grains increases downwind (Mazzullo, Alexander *et al.*, 1992). Loess deposits may be massive or show faint parallel or undulating laminae. Palaeowind directions may be deduced from the preferred orientation and imbrication of silt grains in loess (Matalucci, Shelton & Abdel-Hady, 1969).

Some periglacial lakes, for example in the 'dry valleys' region of Antarctica, exhibit a characteristic association of sulphate and carbonate evaporites, stomatolitic sediments and sands (Walter & Bauld, 1983). In the ice-covered lakes, small quantities of sediment percolate slowly through the porous ice cover and water-filled, vertical gas channels, and algae living on the lake bottom are not overwhelmed by sedimentation (Nedell, Andersen *et al.*, 1987). The resulting facies association may be an important modern analogue for certain problematic ancient associations of carbonates with glacial facies (Walter & Bauld, 1983) (See Sect. 11.5.5).

11.5 Ancient glacial facies

11.5.1 Characteristics and recognition

Lithofacies deposited in glacial environments are very diverse; all terrigenous sedimentary rock types can be present, and carbonate rocks exist locally (Table 11.1). The diversity results from many glacial facies being hybrids of glacial and other sedimentary processes, for example glaciofluvial, glaciolacustrine deposits. *High sediment supply* is typical in all proximal, and some distal, settings. Whether the sediment is dumped by the glacier or redistributed by meltwater, the rate of sedimentation can be extremely high and the volume of sedimentary deposits large. In addition, *marked seasonal fluctuations* in meltwater and sediment supply characterize glaciation in temperate regions and give a distinctive signature to some glacial facies (Sect. 11.4).

To infer accurately a glacial depositional environment, all sedimentary rocks, structures and contacts as well as the overall facies context must be carefully examined. No individual feature is uniquely diagnostic of glacial processes. None the less, certain specific criteria are useful and are discussed briefly in Table 11.2. Comprehensive discussions of sedimentary features characteristic of glacial deposits are given by Harland, Herod and Krinsley (1966), Schermerhorn (1974), Hambrey and Harland (1981), and Hambrey (1994).

Many authors have used a facies code, particularly one adapted by N. Eyles, Eyles and Miall (1983) from one originally proposed for fluvial sediments (Miall, 1977; see Sect. 3.5), as a shorthand for describing glacial facies. Lithofacies codes should be purely descriptive, and one shortcoming of N. Eyles, Eyles and Miall's (1983) scheme is that it incorporates genetic interpretation of some diamictites facies. Moreover, their distinction between matrix-supported and clast-supported diamictite is inappropriate because diamictites was originally defined as matrix-supported (see Sect. 11.1). The variation in the clast concentration of diamictites can be included through use of a lithofacies code that incorporates 'apparent clast packing density' (Visser, 1986) or a revised classification for poorly-sorted sedimentary rocks that quantifies clast-rich versus clast-poor diamictites (Moncrieff, 1989). Pitfalls of using any lithofacies codes are that they encourage simplistic pigeon-holing such that important variations and details may be overlooked, as may be the nature of contacts between facies. Any scheme should be individually tailored to suite the specific needs and goals of each project (see Sect. 2.2.1).

11.5.2 Glacio-eustasy and glacio-isostasy

The waxing and waning of ice sheets causes important eustatic till (MacAyeal, 1993); and (ii) catastrophic release of meltwater reservoirs (Blanchon & Shaw, 1995).

11.5.3 Glacial facies zones*

Patterns of glacial facies distribution can be divided into two: those that form under terrestrial conditions (Fig. 11.17) and those that form under marine conditions (Fig. 11.18), although thermal regime and relief are also important influencing factors. Since erosion generally dominates in subglacial areas during glacial advance, most deposition occurs during glacial retreat.

TERRESTRIAL GLACIAL FACIES ZONES

During terrestrial glaciation, the glaciated area and adjacent proglacial zone are above sea level. Pleistocene deposits of North America and northern Europe provide well-documented examples of these facies in a continental, low-relief setting and the gross, regional facies distribution can be delineated from many publications and large-scale glacial geological maps (Flint, 1945, 1959; Prest, Grant & Rampton, 1968; Woldstedt, 1970, 1971).

Three main facies zones are differentiated (Fig. 11.17; Sugden & John, 1988). Surrounding an inner erosional zone with thin, sporadic till deposits is the *subglacial facies zone*, which appears on Pleistocene geological maps as a fluted or drumlinized till plain, and is composed predominantly of lodgement till. Till may be the only deposit across much of this zone, though subglacial meltwater facies can be present. Locally, the subglacial facies may be overlain by stratified glacial retreat facies, including widespread varved mud. The upper surface of these deposits is subject to frost action and/or soil development (Sect. 11.4.7). However, with repeated glacial advance and retreat,

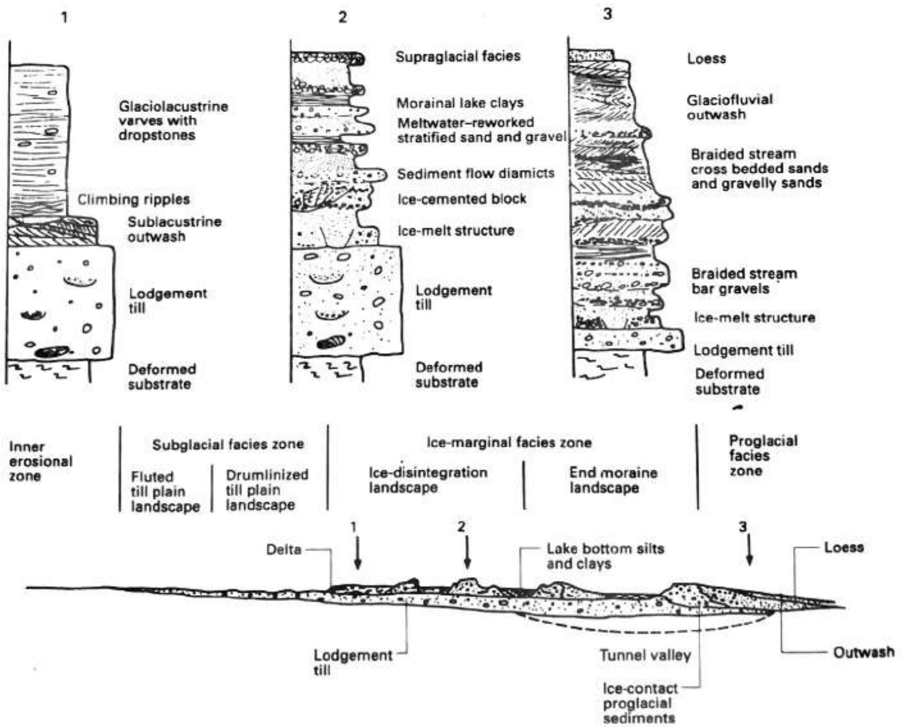


Figure 11.17* Schematic representation of the zonation of terrestrial glacial landforms (from Sugden & John, 1988) and facies deposited by a wet-based ice sheet in a low-relief setting, and characteristic vertical profiles that would be deposited in each zone. The radius of the ice sheet could have been up to 2000 km. The resulting deposits are typically 5–50 m thick. *See footnote on p. 454.

* See footnote on p. 454.

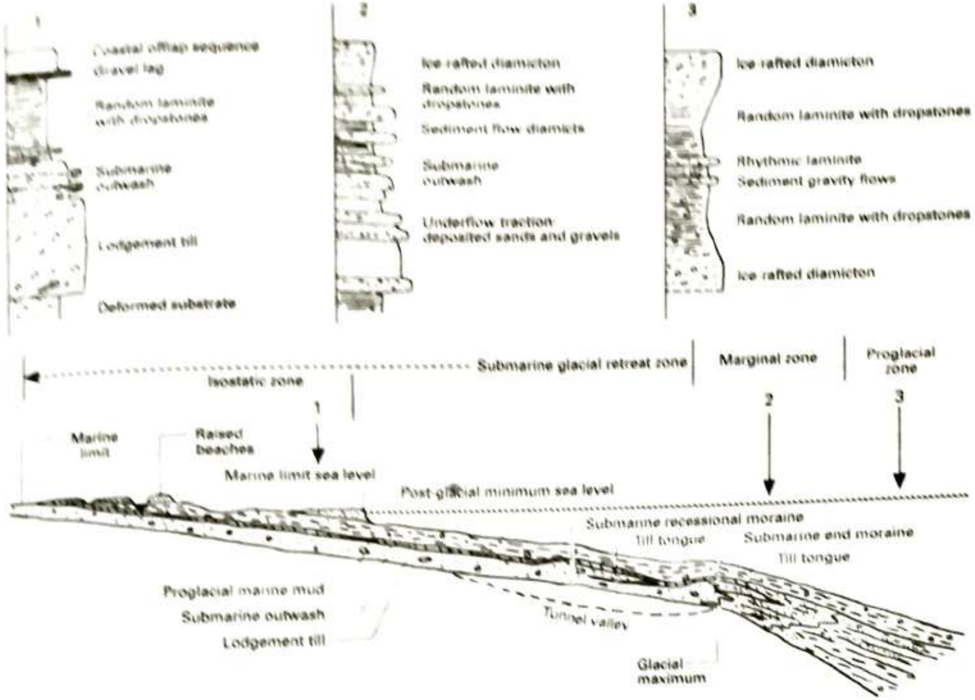


Figure 11.18* Schematic representation of the zonation of marine glacial facies deposited by a wet-based marine ice sheet in a low-relief setting, and characteristic facies sequences that would form in each zone. * See footnote on p. 454.

deposits formed during one retreat may be stripped away during an ensuing advance, so that ultimately a sequence of lodgement tills with thin or no intervening facies is built up.

The *ice-marginal facies zone* occurs in the outer parts of glaciated regions, with two distinct landscapes: end moraines and ice-disintegration topography. Three facies associations occur in this zone: (i) lodgement till and other subglacial facies are deposited throughout, resting upon a regional erosion surface. The underlying deposits may show glaciotectonic deformation; (ii) above are supraglacial and ice-contact proglacial deposits, occurring as end moraines in the outer part of the zone, and as ice-disintegration topography in the inner part; and (iii) widespread varved lacustrine muds may rest on any of these facies. A comparable facies association can develop where the glacier terminus is grounded in a large lake (Landmesser, Johnson & Wold, 1982). In the course of several glacial phases, a complex succession of lodgement tills with thin interstratified outwash sediments can accumulate (White, Totten & Gross, 1969). The region towards the edge of the glaciated area is particularly sensitive to ice-margin fluctuations which may generate a complex stratigraphic record, in some cases including glacial advance as well as retreat successions, and tunnel valley tills. In contrast, more proximal areas remain ice-covered and a simple succession results.

The *proglacial facies zone* includes some ice-contact deposits and all proglacial deposits such as glaciofluvial sands and gravels, lacustrine muds, and windblown sand and silt. The area is largely beyond the zone of till deposition. Apart from the rare occurrence of striated clasts, it may be impossible to deduce glaciation from these deposits alone. Outside of major drainage channels, where valley trains accumulate, these deposits thin rapidly away from the end moraine complex.

MARINE GLACIAL FACIES ZONES

Glaciomarine facies are deposited below sea level and at elevations up to the *marine limit*, the maximum extent of the sea reached during postglacial eustatic sea-level rise. The generalized model presented in Fig. 11.18 is a composite compiled from several Pleistocene examples. Because the bulk of Pleistocene glaciomarine sediments lie below modern sea level and are studied principally by seismic profiling and shallow coring, they are less well known than terrestrial deposits. Moreover, glaciomarine facies associations may be more complex than their terrestrial counterparts, especially where deposition was influenced by isostatic and eustatic effects and the resulting deposit shows the effects of terrestrial, marine and glacial agents.

The critical boundaries that control facies distribution and contacts at any particular time are: (i) glacial maximum; (ii) marine limit; and (iii) postglacial minimum sea level (Fig. 11.18). Because of the number of parameters involved, many scenarios can be reconstructed for glacial retreat sedimentation near sea level. Some of these involve a combination of marine and terrestrial processes because the glacial maximum may rise above sea level during continued glacial recession. To simplify the discussion here three assumptions are made: (i) the glacial maximum was below contemporaneous sea level; (ii) isostatic depression exceeded eustatic fall; and (iii) eustatic rise preceded isostatic rebound and relative fall in sea level. Four glaciomarine facies zones can then be defined, moving from distal to proximal locations: proglacial, marginal, submarine retreat, and isostatic (Fig. 11.18).

The *proglacial marine facies zone* occurs seaward of the glacial maximum. Facies of the open ocean and iceberg zone (Sect. 11.4.6), for example dropstone laminites, sediment gravity flow deposits, ice-rafted and ice-keel reworked diamictons, are gradationally to sharply interbedded with non-glacial facies. Typically, sediment gravity flow deposits are more abundant closer to the ice terminus, and ice-rafted diamictons are more important distally (Boulton & Deynoux, 1981). Therefore, during an advance–retreat cycle a vertical profile such as (3) in Fig. 11.18 may result.

The *marginal glaciomarine facies zone* is formed close to the glacial maximum, generally during major glacial stillstands. It consists of grounding-line-zone sediment, including submarine end moraines, grounding-line fans, and accumulations of diamicton that may form diamicton aprons and/or till tongues (Sect. 11.4.6; Figs 11.12 & 11.13). These facies may rest on thin deposits of basal till and be overlain and interfinger distally with facies of the proglacial zone.

Facies associations in both the marginal and proglacial zones vary depending upon whether the ice is wet- or dry-based. Under wet-based conditions, supply of terrigenous sediment is likely to be high; if the ice is dry-based, terrigenous sediment is probably sparse and biogenic facies may dominate (Sect. 11.4.6).

The marginal facies zone may be areally widespread if the grounding line migrates over time, for example in response to changes in mass balance. Also, submarine outwash may occur locally as recessional moraines, above lodgement till, inside the glacial maximum up to where the ice terminus becomes subaerial during glacial retreat. However, marginal glaciomarine facies can only accumulate as far oceanward as the continental shelf break, the limit for ice sheet grounding, whereas proglacial marine facies can accumulate on the shelf, slope or nearby deep ocean floor.

The *submarine glacial retreat facies zone* occurs between the marine limit and the glacial maximum. It includes the isostatic facies zone, discussed below, and so may be eroded away above the postglacial minimum sea level. The sequence begins with a lodgement till which rests on a subglacial erosion surface. Pleistocene till, deposited in this zone during lowered sea level, is widely recognized on modern high-latitude continental shelves from drilling and geotechnical and seismic properties (Cooper, Barrett *et al.*, 1991; Hambrey, Barrett *et al.*, 1992). Till is overlain by glaciomarine sediments, in many cases dropstone laminite. Under wet-based conditions, submarine outwash will likely be abundant between the till and laminite facies. However, near a retreating dry-based ice sheet basal till may be abruptly overlain by marine proglacial or non-glacial deposits. Following deglaciation of adjacent land areas, normal marine sediments should cap both successions.

Rate of glacial retreat affects facies types and distribution in the submarine retreat zone and the relative abundance of retreat zone versus marginal zone facies. Rate of retreat, or changing position of the ice sheet grounding line, depends upon mass balance, sea level and slope of the substrate near the ice margin (Thomas, 1979; J.B. Anderson & Thomas, 1991). If either mass balance becomes negative or sea level rises and the bed slopes away from the ice centre, then the grounding line will slowly rise and retreat and a gradual submarine retreat sequence will form (Fig. 11.19). Alternatively, if the bed slopes toward the ice centre, the grounded outer part of the ice sheet will become unstable. The ice sheet will rapidly rise to an ice shelf configuration and extensive calving will occur; the grounding line will quickly retreat to a point at which the substrate is shallow enough for the ice margin to stabilize (Fig. 11.19). A large, though short-lived, area of sub-ice shelf sedimentation (Sect. 11.4.6) may be created which is included within the submarine glacial retreat facies zone. Such rapid retreat will result in a catastrophic submarine retreat sequence. Homogeneous diamicts, deposited by rain-out during rapid ice shelf disintegration, may overlie basal tills, with no record of the usually intervening marginal facies zone, and be abruptly overlain by proglacial or non-glacial facies. This ice-sheet decoupling and rapid retreat should be common during large-scale marine deglaciation because most glaciated continental shelves have an irregular bathymetry and proglacial isostatic depression (Fig. 11.14). On a smaller scale, rapid retreat can occur in a fjord from a grounding line situated at a sill (a submarine rise), as demonstrated recently by the Columbia Glacier in Alaska (Hambrey & Alean, 1992).

The *isostatic glaciomarine facies zone* lies between the marine limit and the postglacial minimum sea level, below which

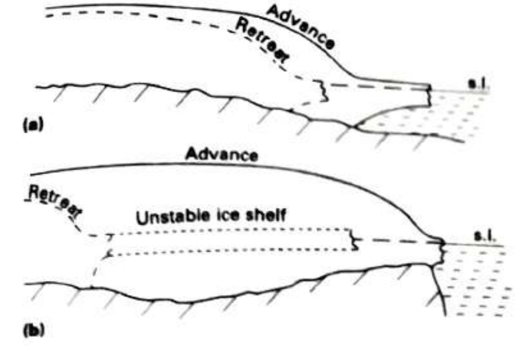


Figure 11.19 Simplified models to show effect of bed slope on rate of marine ice-sheet retreat, holding sea level constant. (a) Bed slopes seaward, away from ice centre: negative mass balance causes gradual retreat and a limited shift in grounding-line position. (b) Bed slopes landward, towards ice centre: negative mass balance causes catastrophic retreat and a large shift in grounding line position; dotted line shows intermediate, unstable ice shelf configuration (modified from Thomas, 1979).

marine sediments are not subject to subaerial erosion. Within this zone submarine glacial retreat zone facies are deposited first and may be capped by muds formed during eustatic sea-level rise, which generally occurs before major isostatic rebound (Boulton, Baldwin *et al*., 1982). Later, when relative sea level falls due to isostatic uplift overtaking eustasy, progradational shoreface sequences form, either as a blanket or locally, as well as thin lags or deep scours formed by subaerial to coastal erosion of the exposed sea-floor deposits. At any one place the resultant succession will depend upon the balance between sediment supply, eustasy and isostasy: it may range from predominantly glacial products (Domack, 1983; McCabe, Bowen & Penney, 1993), through a balance (Miller, 1982), to predominantly coastal marine products (Nelson, 1981). Raised beaches and glaciomarine deltas may record old sea levels, and sediments may be modified by soil-forming and periglacial processes.

11.5.4 Stratigraphic architecture

The stratigraphic architecture, or large-scale three-dimensional stratal geometry, of continental margins bordering large ice sheets differs from that typical of other continental margins for several reasons. The continental shelf around an ice sheet is deeper than most shelves and has a reversed bathymetric profile (deepens from the shelf edge towards the coastline) due to isostatic loading (Fig. 11.14). Glacio-eustatic sea level fluctuations, which strongly affect the architecture of low latitude continental margins, will not therefore subaerially expose this overdeepened shelf, but will affect the distribution of grounded ice and thus of erosion and sedimentation on the shelf. Shelves bordering ice sheets also commonly show rugged relief, including troughs and mounds formed by glacial erosion and deposition when grounded ice covered the shelf (e.g. Boulton, Thors & Jarvis, 1988). Moreover, progradation of the shelf is typically along a broad front, owing to the ice sheet terminus lying close to the shelf edge, as opposed to the point sources of most margins (Larter & Barker, 1989; Cooper, Barrett *et al*., 1991; Hambrey, Barrett *et al*., 1992; Larter & Cunningham, 1993).

Glaciomarine facies architecture, as affected by isostatic and eustatic relative sea-level changes, has been modelled by Boulton (1990) and compared with actual patterns of shelf and slope sedimentation formed during the last glaciation. Boulton proposes that glacier expansion to the shelf edge produces upper slope progradation, because of the large sediment supply to and over the shelf edge (Fig. 11.20). This creates a type of complex sigmoid–oblique clinoform (see Fig. 2.4). During glacial retreat the shelf edge is relatively starved of sediment and reworked by coastal currents. Sediment will collapse and slump from the upper slope to accumulate lower on the slope. Reworked and slumped units onlap the underlying strata (Fig. 11.20). In a comparable model, Larter and Barker (1989) place sequence boundaries at the floating of a previously grounded ice sheet as sea level is rising during a glacial recession.

Seismic stratigraphic studies coupled with drilling on the Antarctic and northern Norwegian continental margins support these generalizations. Prograding sequences that build the continental shelf outward are common; they have complex sigmoidal geometries and progradation is typically from a line source. Troughs and mounds lie parallel and perpendicular to the shelf edge. Aggradational sequences underlie and overlie

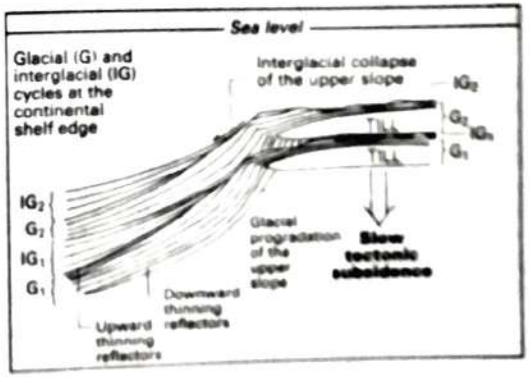


Figure 11.20 Idealized structure of a shelf edge and upper slope in which glacier expansion to the shelf edge produces upper slope progradation, and phases of glacier retreat permit collapse of the upper slope and build up of sediments lower on the slope. Tectonic subsidence permits a vertical sequence to build up which would otherwise be removed by energetic shelf processes. Units labelled 'till' could represent till tongues (Boulton, 1990).

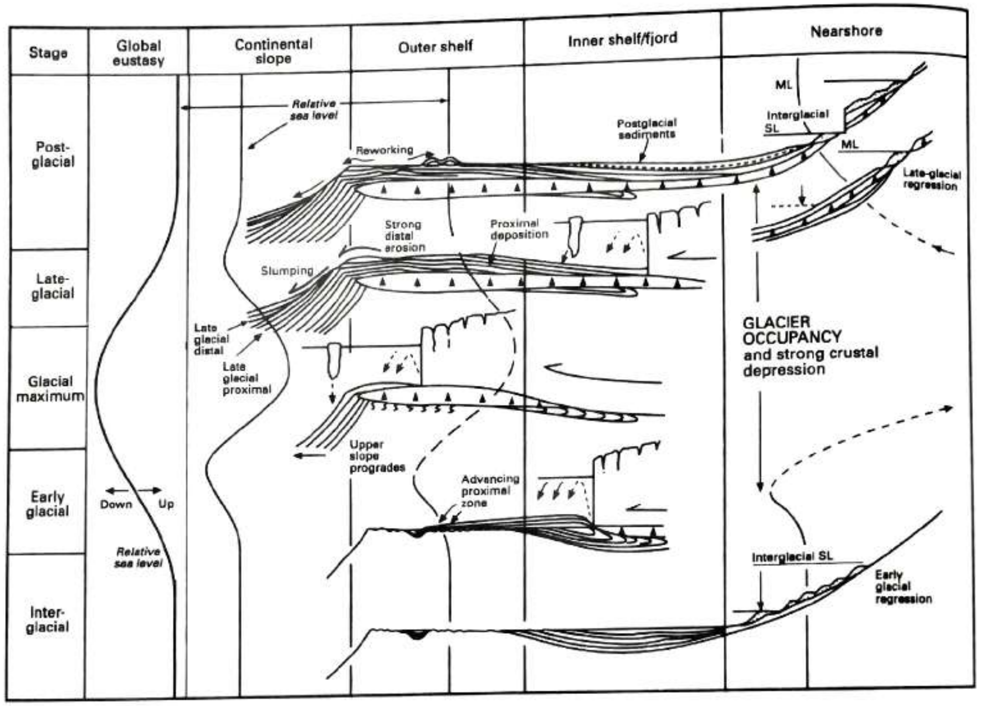


Figure 11.21 Model of glaciomarine architecture in space and time showing facies in different continental margin locations which accumulate through a whole glacial cycle. Relative sea-level changes appropriate to each zone are shown (Boulton, 1990).

the progradational ones. These may have a simple geometry and gently dipping reflections, or may be flat-lying and composed of overcompacted diamict deposited from grounded ice (Vorren, Lebesbye *et al.*, 1989; Cooper, Barrett *et al.*, 1991; Alonso, Anderson *et al.*, 1992; Hambrey, Barrett *et al.*, 1992). Note, however, that stratigraphic architecture is also affected by tectonic setting, particularly rate of subsidence, and differs between passive and active glaciated continental margins (Gipp, 1994).

Few attempts have been made to model, or reconstruct, the effects of glacio-eustasy and glacio-isostasy upon continental margin architecture (from continent across the shelf to the slope) throughout a glaciation, that is from preglacial through glacial to the postglacial stage. Figure 11.21 shows a simple model of this type. Reading and Walker (1966) attempted a large-scale reconstruction for the Late Proterozoic glacials of north Norway, but much has been learnt about glacial sedimentology and stratigraphic architecture since then. Research and modelling of this type are needed, perhaps for the Pleistocene Laurentide ice sheet and/or for older strata where a broad view of the basin through space and time is available.

11.5.5 Ancient glacial facies associations

Glacial facies associations in the rock record are affected by tectonic setting, local relief, and glacial thermal regime as well as their environment of deposition. Cyclicity may also be visible due to repetition of advance or retreat successions. The influences of these parameters, which are commonly interrelated, are illustrated in selected examples discussed briefly below.

Distinguishing between deposition from ice sheets and valley glaciers is extremely difficult because most glacial processes are common to both. The best evidence for large ice sheets are distinctive far-travelled clasts and geographically widespread deposits.

TERRESTRIAL ASSOCIATIONS

Spectacular exposures of terrestrial glacial rocks deposited in a *low relief setting* on a *stable craton* exist within the Upper Ordovician of northwest Africa. Their documented areal extent is approximately 6–8 million km², or about half the area covered by the Quaternary Laurentide ice sheet (Biju-Duval, Deynoux & Rognon, 1981; Deynoux & Trompette, 1981b). Topographical expression of facies in plan view is excellent, allowing three-dimensional reconstructions.

Important regional facies variations are evident which broadly agree with the zonation of Pleistocene continental glacial deposits (Biju-Duval, Deynoux & Rognon, 1981).

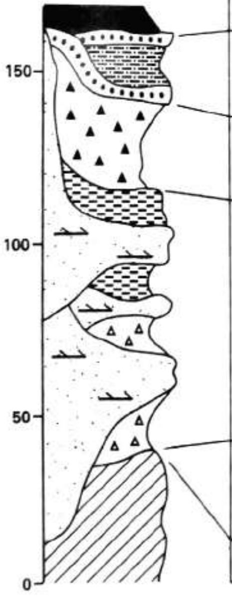
1 In the south, massive diamictites, interpreted as tillites, are abundant, sequences are comparatively thin, and internal erosional unconformities are numerous, have low relief, and attest to successive glacial phases related to ice sheet fluctuations.

2 Farther north, there is greater relief and deposits show rapid lateral thickness changes, reaching about 200 m where they fill large palaeovalleys. Outwash sandstones and other marginal facies are abundant, and massive diamictite is relatively scarce (Fig. 11.22).

3 North of the outcrop area, boreholes indicate a gradual increase in marine strata, with thicknesses up to several hundred metres and fewer signs of glacial erosion.

Facies cyclicity demonstrates repeated episodes of glacial advance and retreat in places within the central zone. A typical succession, comprising an unconformity overlain by massive diamictite (interpreted as till), channelled sandstone (glaciofluvial), and argillaceous sandstone (glaciolacustrine or glaciomarine), is repeated three or four times to fill palaeovalleys (Fig. 11.22; Deynoux & Trompette, 1981b; Deynoux, 1983). Above these valley-fills is massive diamictite with sparse clasts (waterlain diamictite) conformably overlain by shales containing dropstones together recording the transition from glacial to postglacial marine conditions (Fig. 11.22; Deynoux, 1985). Well-sorted aeolian sandstones are present elsewhere, interbedded locally with glaciofluvial deposits, as well as excellently preserved periglacial features, including sandstone wedges and fossilized pingos (Beuf, Biju-Duval *et al.*, 1971; Biju-Duval, Deynoux & Rognon, 1981).

Another well-preserved *low relief*, terrestrial glacial succession exists within the Upper Proterozoic of western Mauritania. Lateral facies relationships are varied and quite complex (Fig. 11.23). Again the sequence is thin, reaching 50 m where it fills a wide, shallow depression (Fig. 11.23; Deynoux, 1983, 1985). Polygonal structures and sand wedges cap the younger glacial deposits (Deynoux, 1982). Overlying them is 3–5 m of baryte-bearing calcareous dolomite with sandy or shaly intercalations, interpreted to mark the beginning of the postglacial



Lithology and structures	Interpretation
Shale with graptolites and brachiopods	Marine sedimentation — post-glacial transgression
Well-bedded **shale** with graptolites and dropstones, interbedded with cross-bedded argillaceous sandstone	Ice-rafting during marine sedimentation
Massive **diamictite** with sparse clasts, including some boulders	Ice-rafting of debris during lacustrine or marine sedimentation
Alternating **diamictite**, overlain by **channelled sandstone** and **argillaceous sandstone**: Argillaceous sandstone — planar or cross-bedded, ripple marks; Channelled sandstone — massive or cross-bedded; Diamictite — foliated, not bedded, locally contains inclusion of underlying units	Shallow marine or lacustrine sedimentation; Proglacial braided stream sedimentation; Terrestrial tillite
Unconformity — striated pavement, roche moutonnées, some substrate deformation	Glacial erosion with some glaciotectonic deformation
	Pre-glacial strata

Figure 11.22 Section showing facies cyclicity and abundance of sandstones in terrestrial glacial deposits in a low-relief setting, Upper Ordovician, southeast Mauritania (modified from Deynoux, 1985).

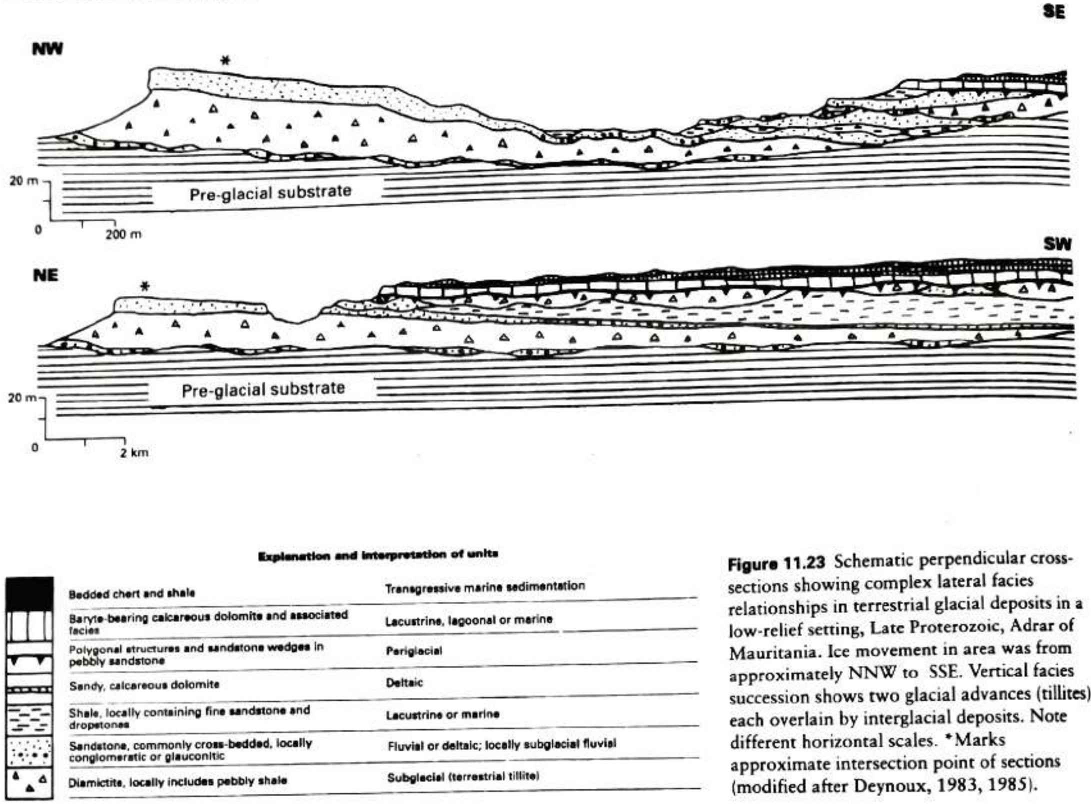


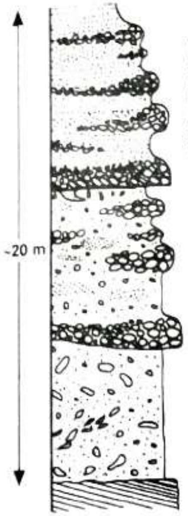
Figure 11.23 Schematic perpendicular cross-sections showing complex lateral facies relationships in terrestrial glacial deposits in a low-relief setting, Late Proterozoic, Adrar of Mauritania. Ice movement in area was from approximately NNW to SSE. Vertical facies succession shows two glacial advances (tillites) each overlain by interglacial deposits. Note different horizontal scales. *Marks approximate intersection point of sections (modified after Deynoux, 1983, 1985).

transgression (Deynoux & Trompette, 1981a). Such carbonate caps are a common and somewhat problematic feature of many Late Proterozoic glacial successions (see below). Spectacular periglacial aeolian deposits representing ergs peripheral to outwash plains exist farther south in western Mali (Deynoux, Kocurek & Proust, 1989).

Ancient examples of terrestrial glacial deposition in *high-relief* settings are rare. Where preserved, the glacial record may be patchy, incomplete and largely resedimented (Collinson, Bevins & Clemmensen, 1989).

Wet-based glacial retreat sedimentation in a terrestrial setting is found within the Upper Proterozoic Smalfjord Formation in Finnmark, north Norway (Fig. 11.24; Edwards, 1975, 1984). At the base of a low-relief palaeovalley, massive diamictite with intraformational clasts (lodgement or melt-out till) is overlain by interbedded diamictite, sandstone and conglomerate (supraglacial and/or proglacial sediment) followed by interbedded conglomerate and sandstone (glaciofluvial deposits) indicating progressive glacial retreat (Fig. 11.24). The upper 75 m of the palaeovalley fill comprises largely massive sandstones with some conglomerate, diamictite and shale, deposited by sediment gravity flows into a lake or fjord after ice left the area (Edwards, 1984).

Repeated glacial advance and retreat near the non-marine terminus of a *wet-based glacier* has caused marked *facies cyclicity* in Permo-Carboniferous strata in the Transantarctic Mountains (Miller, 1989). Sequences differ, depending upon whether advance was subaerial or into standing water (Fig. 11.25). Subaerial ice advance is evidenced by massive diamictite overlying a sharp basal contact with erosional relief and/or glacial grooves and striations. Subaerial ice retreat is recorded by: (i) massive or sheared diamictite which may contain tilted sandstone lenses (lodgement till); overlain by (ii) diamictite containing conformable, discontinuous stringers of sorted sandstone and conglomerate (melt-out till); followed by (iii) massive diamictite interbedded with planar cross-bedded or massive sandstone, with usually planar but occasionally erosive bed contacts (proglacial resedimented till interbedded with glaciofluvial outwash). Glacial advance under subaqueous conditions places diamictite (commonly clast-poor) gradationally above massive or laminated shale which may contain dropstones. Glacial retreat leads to a gradational contact between usually massive



Observations		Interpretation		
Lithology and structure	*Clast composition*	*Relative distance of transport*	*Environment of deposition*	*Means of transport*
Alternating massive and lenticular ***conglomerate*** with erosive bases and pebble imbrication, ***and*** massive, parallel-laminated or low-angle cross-stratified ***sandstone***, broadly fining up	Dolomite and chert, crystallines, sandstone and shale	Distant	Fluvial (glacial) outwash	Fluvial
Massive ***diamictite with*** lenticular bodies of stratified diamictite, parallel- or cross-stratified or massive ***sandstone*** and massive or cross-stratified ***conglomerate***	Crystallines, sandstone and shale	Intermediate	Supraglacial, proglacial till and drift	Upper englacial, supraglacial
Massive ***diamictite***	Sandstone and shale	Local	Subglacial till	Lower englacial

Figure 11.24 Terrestrial wet-based glacial retreat sequence from the late Proterozoic Smalfjord Formation, north Norway (after Edwards, 1975).

diamictite, which may contain sandstone lenses, and overlying shale with sparse clasts and some thin rippled sandstone lenses and beds (Fig. 11.25). In much of this area the succession is remarkably complete and three to six advance–retreat cycles exist in sections 100–200 m thick (Fig. 11.25; Miller, 1989).

In the subsurface in Oman, Upper Palaeozoic *glacio-lacustrine* shales with varve-like laminations and dropstones are associated with diamictites, sandstones and conglomerates, in an oil-bearing succession which has been thoroughly examined using both cores and wireline logs (Levell, Braakman & Rutten, 1988). Varvites are also documented within the Upper Palaeozoic Itararé Subgroup of the Paraná Basin, Brazil (Rocha-Campos, Ernesto & Sundaram, 1981). Rhythmite sequences 30 m or

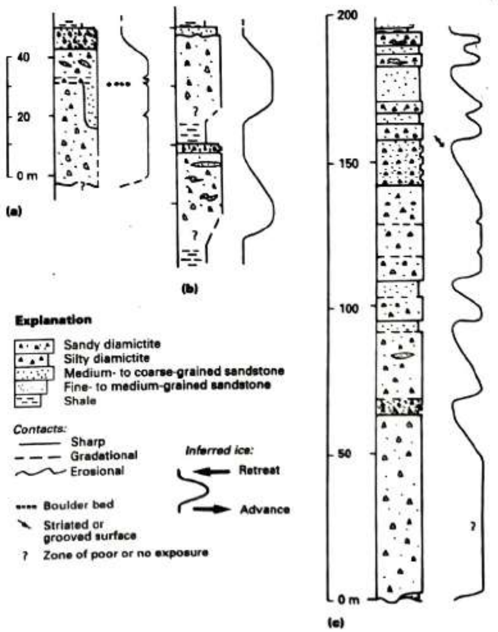


Figure 11.25 (a,b) Parts of measured sections through the Permo-Carboniferous Pagoda Formation, central Transantarctic Mountains, showing glacial advance followed by retreat under (a) subaerial (grounded ice) and (b) subaqueous conditions. (c) One complete section showing approximately six advance–retreat cycles, deposited in this case under dominantly subaerial conditions. Small kinks on advance–retreat interpretative curve signify minor fluctuations of ice margin and/or changes in basal ice dynamics or pauses in sedimentation (Miller, 1989).

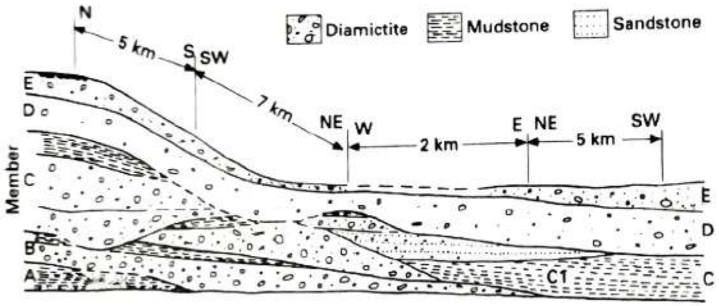


Figure 11.26 Schematic cross-section showing facies relationships during several ice sheet advances and retreats in a marginal low-relief, stable continental shelf setting, late Proterozoic Smalfjord Formation, north Norway (after Edwards, 1984).

more thick are associated with diamictite and sandstone. Annual (seasonal) control upon rhythmite deposition is indicated by concentration of pollen and spores in the coarse layers with the fine layers practically barren, and spectral analysis of thickness variation and palaeomagnetic data from three rhythmite sequences.

Deposition in a *marginal, low-relief, stable continental shelf* area where terrestrial and marine glacial conditions alternated during several advances and retreats of an ice sheet is demonstrated by the upper Smalfjord Formation, Upper Proterozoic, north Norway. Five lithologically distinct diamictite units can be mapped over areas of 10s–100s km^2, and occur within a repetitive vertical sequence of: (i) erosion surface; (ii) generally massive diamictite (lodgement till) which may show banding formed by glacial shearing near its lower contact (Edwards, 1984, 1986); (iii) rhythmic to randomly laminated mudstone, with rare to abundant dropstones and rare intervening sandstone (shallow glaciomarine deposits; Fig. 11.26; Edwards, 1984). Deposition took place near the margin of grounded ice sheets in an area where sea water could encroach (Edwards & Føyn, 1981).

However, at one place in the upper Smalfjord Formation faintly stratified siltstone, with a high silt and low clay content, overlies diamictite (basal till), and is interpreted as indurated wind-deposited silt or '*loessite*' (Edwards, 1979).

MARINE ASSOCIATIONS

Glaciomarine deposits are typically thicker than their terrestrial counterparts and have a higher potential for preservation (N. Eyles, 1993). They are widespread in the rock record, both in space and time (J.B. Anderson, 1983; Andrews & Matsch, 1983). A marine depositional environment is established by the presence of marine fossils; in the absence of fossils it has to be inferred from the facies and facies associations. As with terrestrial deposits, tectonic setting, local relief, glacial thermal regime and climatic cyclicity influence facies associations, as well as the specific depositional environment, but in marine environments glacio-eustatic and glacio-isostatic effects may also be evident.

The effects of *relief* and *thermal regime* upon glaciomarine deposition are both clearly demonstrated in the Late Palaeozoic Dwyka Formation of the Karoo Basin, southern Africa (Visser, 1991). The tectonic setting was a foreland basin near the margin of Gondwana.

Contrasts between southern and northern facies associations are due to differences in *local relief* (Fig. 11.27). The 'shelf facies association', exposed in the south, is widespread, lithologically fairly homogeneous, and up to about 800 m thick. Massive clast-rich diamictite is commoner at the base and massive clast-poor diamictite near the top of the sections (Fig. 11.28). Stratified diamictites increase in abundance northward – that is, nearer the unstable ice front (Fig. 11.28, Elandsvlei section). Mudrock facies can be traced for distances of 400 km and have sharp contacts with adjacent diamictite units. The vertical and lateral homogeneity of the massive diamictites suggests deposition below the inner parts of disintegrating ice shelves. Stratified diamictites were probably deposited near the grounding line. Widespread mudrock units indicate interglacial periods. Most sedimentation occurred during short periods of time when rapidly rising sea level caused the ice sheet and shelf to decouple and become self-destructive (Visser, 1991).

In contrast, the 'valley facies association', exposed in the north and restricted to palaeovalleys (fjords), is up to 200 m thick and characterized by rapid facies and thickness changes. It consists predominantly of massive and stratified diamictite, rhythmite, argillite with isolated clasts ('lonestones') and mudrock, with minor sandstone and conglomerate (Fig. 11.28). Deposition was by tidewater glaciers while mudrocks were being deposited on the shelf (Fig. 11.27; Visser & Kingsley, 1982; Visser, 1991).

The influence of *thermal regime* is shown in the Dwyka Formation through an upward transition from polar to temperate climatic conditions (Visser, 1991). Eroded and glaciotectonized bedrock underlies the formation in most locations, suggesting that initially dry-based grounded ice existed and left little or no sedimentary record. Alternating diamictite and mudrock facies in the shelf association suggest fluctuations between polar and subpolar conditions. A polar regime is required by the inferred existence of ice shelves. Abundant meltwater deposits of the valley facies association and mudrocks at the top of the shelf association indicate deposition under temperate climatic conditions.

Tectonic setting played an important part in glaciomarine sedimentation of the late Miocene to Recent Yakataga Formation in the Gulf of Alaska. The formation is 5 km thick and was deposited in the collision zone between oceanic and continental crust (C.H. Eyles, Eyles & Lagoe, 1991). ***Thermal regime*** and *local relief* were also important: very high sedimentation rates (as much as 10 m ky^{-1}) were due to abundant meltwater from coastal temperate tidewater glaciers carrying sediment derived from rapidly uplifting coastal mountains.

The lower and middle parts of the formation comprise laminated mudstone, with clast-rich bands and sandstone

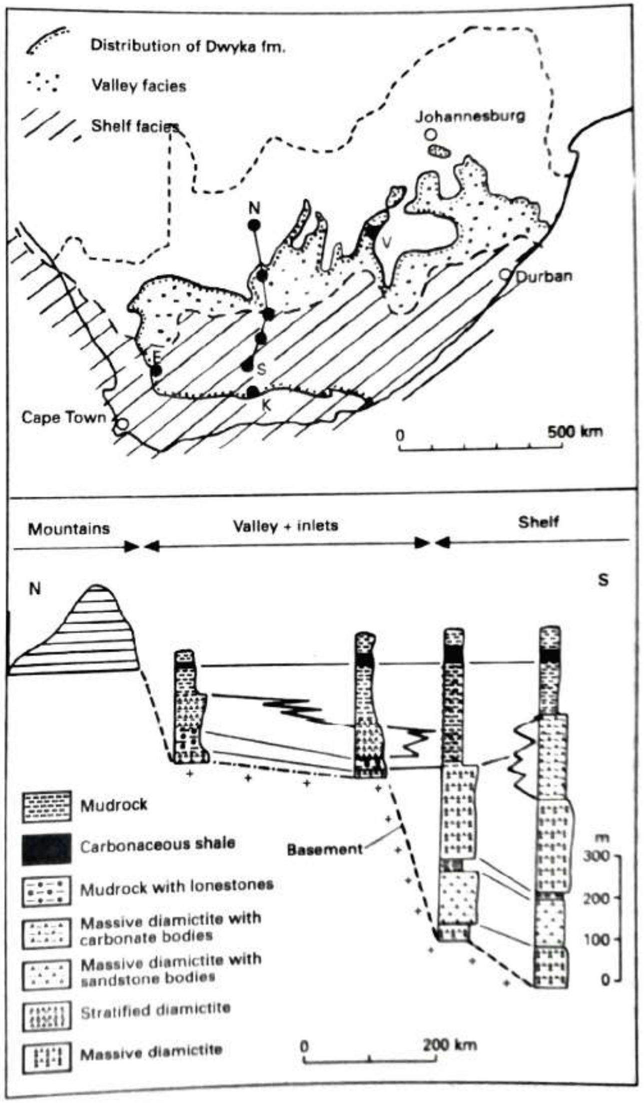


Figure 11.27 Map and section illustrating distribution of and relations between the valley and shelf facies in the Late Palaeozoic Dwyka Formation, Karoo Basin, Republic of South Africa. Note pinch-out of the lowermost shelf units against the palaeoescarpment. N–S, location of section; E, Elandsvlei; K, Klaarstroom; V, Virginia sections shown in Fig. 11.28 (modified after Visser, 1991).

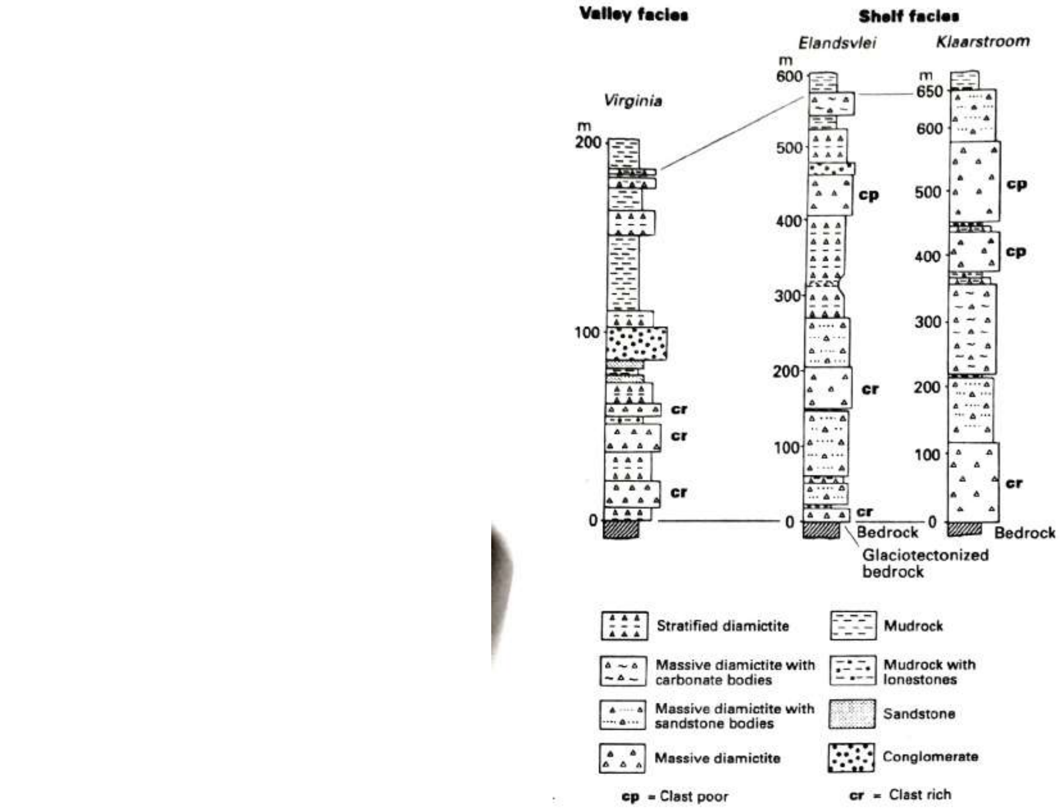


Figure 11.28 Stratigraphic sections representing the valley and shelf diamictite facies of the Dwyka Formation. See Fig. 11.27 for locations of sections. Note different vertical scale for each section (Visser, 1991).

and diamictite interbeds, overlain by sandstone turbidites, massive mudstones, swaley cross-stratified sandstones, and a thick section of broadly lenticular diamictites. Deposition was by turbidity currents and debris flows in upper slope environments followed by downslope collapse of inner neritic glaciomarine and marine facies into deep water. This succession shows how difficult it can be to recognize a glacial imprint upon sedimentation in open ocean environments, particularly in a region of active tectonism. The glacial influence here is inferred from an increase in arctic benthic Foraminifera and the arrival of large volumes of heterogeneous sediment due to initiation of glaciation around the Gulf (C.H. Eyles, Eyles & Lagoe, 1991).

The upper part of the Yakataga Formation is a 1.25-km-thick section exposed on Middleton Island in the Gulf of Alaska (Fig. 11.29). Gravels near the base were deposited by channelized sediment gravity flows which cut into the outer continental shelf and upper slope. Some striated, faceted, and occasionally outsize clasts indicate a glacial sediment source and background ice-rafting (C.H. Eyles, 1987; C.H. Eyles & Lagoe, 1990). In the remainder of the section (Fig. 11.29), massive locally fossiliferous diamictons in extensive, sheet-like beds up to 100 m thick are consistent with deposition by suspension settling and ice-rafting on a low-relief outer shelf. Some stratified diamictons were deposited by sediment gravity flows, possibly triggered by earthquake shocks due to the active tectonic setting. Interbedded striated boulder pavements formed when a grounded marine ice sheet abraded a submarine boulder lag surface during episodic ice advances to the shelf edge (C.H. Eyles, 1988b; C.H. Eyles & Lagoe, 1990). Coquinas (shell beds) record sediment-starved, ice-free conditions during low relative sea level. Bioturbated muds record quiet-water deposition during interglacial conditions with higher relative sea level (C.H. Eyles, C.H. Eyles & Lagoe, 1991). The section therefore records temperate, glacially influenced sedimentation on a low-relief outer shelf with fluctuations of sea level and the ice margin.

Many Upper Proterozoic glaciomarine successions were deposited in active rifts and thus demonstrate the effect of a different *tectonic setting*. Sections commonly show abrupt lateral thickness changes due to contemporaneous faulting, sometimes coupled with glacial erosion. For example, glaciogenic strata in parts of the Adelaide 'geosyncline', South Australia show a change in thickness from less than 200 to 5000 m over about 16 km (Young & Gostin, 1989). Many of the diamictites are crudely stratified and show evidence for resedimentation, another feature typical of glaciomarine deposits in active rifts which complicates recognition of a glacial influence. This succession contains two diamictite-dominated to laminated mudstone-dominated cycles interpreted as deposited during two glacial advance–retreat cycles. A temperate climatic regime is inferred because meltwater deposits are abundant (Young & Gostin, 1991).

Cyclicity on different scales is recognized using a sequence stratigraphic approach in Upper Proterozoic marine to nonmarine intracratonic glacially related strata of West Africa (Proust & Deynoux, 1994). Facies deposited in a range of continental to marine environments form progradational, continental and transgressive wedges and are grouped into depositional genetic units bounded by regionally correlative, maximum flooding or erosional bounding surfaces. Cyclicity of depositional genetic units on the smallest scale is related to glacial advance and retreat rhythms with a duration of about 0.1 My. Stacked depositional genetic units comprise cycles which may represent glacial to interglacial stages, and larger-scale cycles reflect either glacial epochs or tectonic processes.

Glacio-eustatic and glacio-isostatic influences have been emphasized in an Upper Proterozoic glaciogenic succession in Central Australia, where the facies associations and sequence

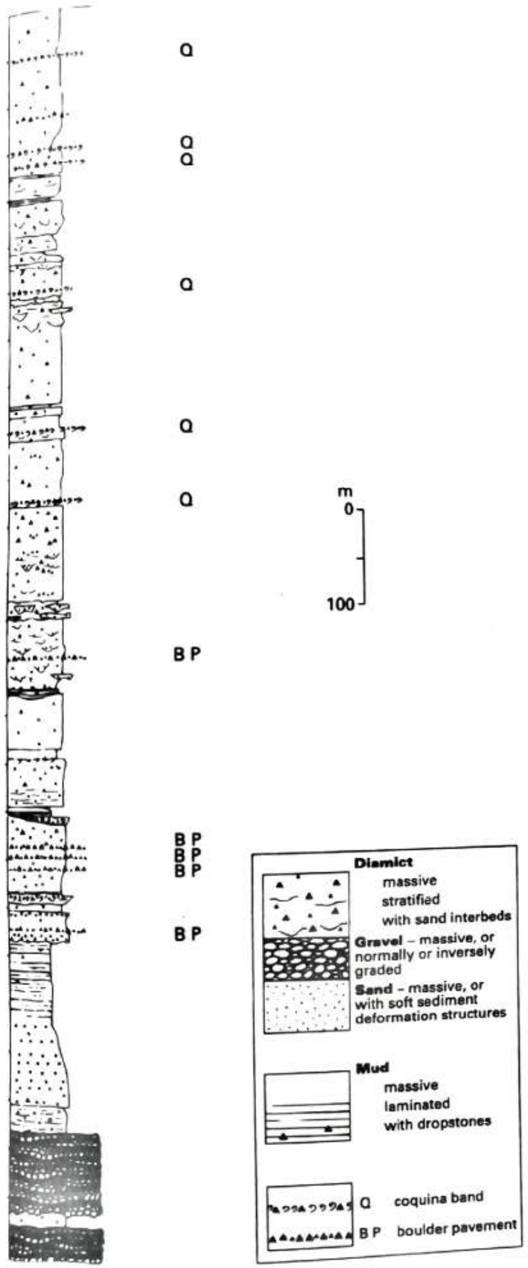


Figure 11.29 Sedimentological log through facies of temperate, glacially influenced sedimentation on a low-relief outer shelf, Yakataga Formation, Middleton Island, Gulf of Alaska (after C.H. Eyles, 1987).

boundaries follow a pattern determined by basin growth and initial fall and subsequent rise of sea level as an ice sheet grew and declined (Lindsay, 1989). Erosion followed by subglacial deposition occurred during lowered sea level. Ice proximal then ice distal and marine sedimentation occurred during glacio-eustatic sea-level rise. Erosion followed due to postglacial isostatic rebound. Diamictites in a younger formation are probably reworked sediment from the underlying formation and belong to a lowstand wedge or fan in the overlying depositional sequence, as opposed to recording a second glaciation.

In many Upper Proterozoic, mostly glaciomarine, successions a *carbonate cap* abruptly overlies glacial diamictite. The limestone or dolomite may be massive or laminated, and may show algal or stromatolitic laminae (Williams, 1979; Fairchild & Hambrey, 1984). If the carbonates record warm conditions, then abrupt climatic changes are implied (Williams, 1979; Fairchild, 1993). However, cold-water carbonates are known to be associated with some glaciomarine facies (Rao, 1981; Sect. 11.4.6). Alternative explanations for this carbonate–diamictite juxtaposition include: (i) carbonate deposition occurred during periods of sea-level rise and deglaciation when terrigenous input was reduced (Tucker, 1986); or (ii) associated carbonate facies may form where detrital carbonate debris exists within the associated glacial deposits (Fairchild, 1993).

11.6 Ice ages in Earth history

Glacial facies are concentrated in certain geological periods, reflecting episodes of cold climates or ice ages during Earth history (Frakes, 1979; Frakes, Francis & Syktus, 1992). The oldest probable glacial rocks occur within the approximately 2.9 Ga Witwatersrand Supergroup of South Africa (Von Brunn & Gold, 1993). Early Proterozoic glacial deposits occur in North America (including the Huronian Supergroup, about 2.3 Ga), southern Africa, Australia, India, and Finland. Late Proterozoic glacial facies are geographically widespread and have been found on all continents, possibly including Antarctica. Glaciation may have peaked during four glacial periods between 900 and 600 Ma (Harland, 1983). Late Ordovician to Early Silurian glacial facies exist in North Africa and South America, and Late Devonian glacials in South America (Caputo & Crowell, 1985), whereas Permo-Carboniferous glacial deposits are well represented on all Gondwanan continents. Cenozoic glacial facies are widespread on all northern continents, on and around Antarctica, and in the Andes (Hambrey & Harland, 1981).

Causes of ice ages include both terrestrial and extraterrestrial factors (Young, 1991). The position of continents and oceans with respect to air–ocean circulation is commonly important (Crowell, 1978). Variations in solar insolation, caused by Earth's orbital variations (the Milankovitch effect; see Sect. 2.1.5), are the most likely driving force for short-term climatic change (i.e. <100 000 years), and for glacial–interglacial cycles (Boulton, 1987b). Particular problems surround the Late Proterozoic glaciation because the deposits are so widespread, some appear to have been deposited at low palaeolatitudes, and a number are closely associated with carbonate facies (Sect. 11.5.5).

Glacial sedimentology has become more sophisticated and accurate in recent years. Frontier areas include further application of sequence stratigraphic concepts and improved recognition and understanding of the significance of discontinuities within glacial successions. Accurate recognition, understanding and interpretation of glacial facies are paramount to elucidating Earth's climatic history.

Further reading

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