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The morphology, structural evolution and significance of push moraines

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Abstract

Push moraines (glaciotectonic ice-marginal moraines) have a restricted distribution at modern glacier margins and consequently are of potential value in reconstructing Pleistocene ice sheets, providing data both on former glaciodynamics and on the palaeoenvironment of the glacial foreland. To the wider earth science community, push moraines are of interest as analogues for thin-skin tectonics within orogenic belts. This paper reviews the morphology, structural geology, formation and significance of push moraines. The morphological and structural characteristics of small, seasonal, push moraines through to large, multi-crested, examples produced by sustained glacier advances are reported, before the primary controls on push moraine formation are examined. These controls include the nature of the applied glacial stress field, the presence and properties of décollement horizons, and the shear strength and rheology of the glacial foreland. A conceptual model of push moraine formation is introduced, in which the range of observed morphological and structural forms are viewed within a matrix defined by the main variables which control their formation. The absence of consensus over which of these variables is of greatest importance currently limits the significance of push moraines in palaeoglaciological research. As a consequence, this review emphasises the need for future research in order to realise the true potential of push moraine in the reconstruction of Pleistocene environments. © 2001 Elsevier Science B.V. All rights reserved.

Keywords: push moraines; ice-marginal landforms; glaciotectonics

1. Introduction

1.1. Aims and rationale

The spatial analysis of glacial landforms is an essential tool in reconstructing ancient ice sheets. These landforms are the clues in the landscape from which the dimensions, geometry, dynamics and history of former ice sheets can be reconstructed. Our understanding of the physics and mechanics of glaciers, and the manifestation of these processes in the landform and sedimentary record, has developed rapidly over the last 30 years as accessibility to contemporary glaciers has increased. Morphological and sedimentological data is increasingly available on individual landforms, and landform assemblages, at contemporary glacier margins where the glaciological processes, environmental conditions or events which control their formation can be constrained. As a consequence when used in reverse, to interpret the Pleistocene record, this geomorphological data has

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enhanced the sophistication and finesse of ice sheet reconstructions (e.g. Boulton and Clark, 1990a,b; Kleman and Bergström, 1996; Kleman et al., 1997). There is a growing need for such reconstructions both to link the behaviour of regional Pleistocene ice sheets with the global environmental signature recorded in marine- and ice-cores (e.g. Dansgaard et al., 1993; Keigwin et al., 1994) and to test models of environmental change and ice sheet behaviour (e.g. Boulton, 1996). In order to support this increased demand for high quality ice sheet reconstructions, there has been a recent trend to define landform-sediment assemblage models (e.g. Bennett et al., 1999a, 2000; Evans and Rea, 1999; Hart, 1999; Evans et al., 1999) and to synthesis knowledge on individual landforms (e.g. Brennand, 2000). It is with this background that the current paper reviews the morphological diversity, formation, and significance of push moraines with a view to providing data of value in the interpretation of the landform record of Pleistocene ice sheets.

Push moraines (glaciotectonic ice- or sub-marginal moraines) contain valuable information about the interaction of glaciers with the foreland into which they advance. The distribution of push moraines is restricted; they do not occur at every ice margin and are, therefore, associated with either specific glaciological scenarios, the occurrence of conditions favourable for deformation in the glacial foreland, or more likely some combination of both. As a consequence push moraines have considerable potential in palaeoglaciological investigations, since their occurrence in the landform record may provide data on both glaciodynamics and the palaeoenvironment of the glacial foreland. They are also a landform type which has been the subject of considerable recent research (e.g. Hart, 1990; Hambrey and Huddart, 1995; Bennett et al., 1999a; Boulton et al., 1999), reflecting the emergence of glaciotectonism as a key process in glacial geology (e.g. Hart, 1995a; Benn and Evans, 1996). Push moraines also have a much wider geological context than many glacial landforms since they provide potential analogues for thin-skin tectonics within orogenic belts (e.g. Gripp, 1929; Pedersen, 1987; Aber et al., 1989). It is in this context that there is a need for a review of push moraine formation to service the student in a rapidly advancing field, the broader geological community interested in push moraines as examples of thin-skin tectonics, and those interested in the contribution that push moraines can make in the reconstruction of Pleistocene ice sheets.

1.2. Questions of terminology and scope

A practical and universal taxonomy of ice-marginal moraines does not exist and a plethora of terms and descriptors have been used within the literature. Non-genetic terms such as lateral, terminal, or recessional moraines occur juxtaposed with such genetic, or sub-genetic, descriptors as push moraines, thrust moraines, dump moraines, ablation moraines and hummocky moraine. The landform classification proposed by INQUA illustrates this point, since it uses a mix of genetic and non-genetic terms (Goldthwait, 1989), making it difficult to use. The problem lies in the fact that specific genetic descriptors are frequently vague and have been applied to a range of different phenomena. Added to which moraines are commonly polygenetic and their formation may involve a range of different processes, thereby defying simple classification. In contrast, glaciotectonic landforms have been classified more effectively (Aber et al., 1989) into: (1) hill-hole pairs; (2) composite ridges and thrust-block moraines; (3) cupola hills; and (4) megablocks and rafts. Although this classification has been widely followed (e.g. Hambrey, 1994; Benn and Evans, 1998), it is confusing in the context of ice-marginal moraines, since the relationship between these forms and the ice margin is not always clear.

It is not the intention of this review to try and resolve these wider taxonomic problems, but a definition, with a defined scope, is required to constrain the paper and consequently I propose to use the term ‘push moraines’ to describe all forms of glaciotectonic, ice-marginal or sub-marginal moraine. As a descriptor the term push moraine has been widely used since Chamberlin (1890, p. 28) first defined it, although earlier descriptions of the process exist (e.g. De Charpentier, 1841; Heim, 1885). Some workers have argued that the term ‘push moraine’ is a genetic, and therefore a formal, descriptor which should be restricted to moraines produced by sediment bulldozing and is less appropriate where glaciotectonic thrusting, for example, dominates (e.g. Evans, 1989; Benn and Evans, 1998). In contrast, others have used 'push moraine' in a more informal fashion as an all-inclusive term (e.g. Kalin, 1971; Boulton, 1986; Etzelmüller et al., 1996; Boulton et al., 1999). This argument goes to the heart of the debate about the driving forces responsible for ice-marginal deformation, a choice between those who emphasise glacioisostatic (gravity-spreading) and those who favour glaciodynamics (pushing) forces. In practice it is difficult to separate these forces, while other sources of stress may also be involved, and neither is exclusively associated with one type of deformation structure. As a consequence it is often difficult to infer the origin of the forces involved in push moraine formation simply from the morphology and structural geology observed. Consequently, in taxonomic terms, it makes little sense to use the term 'push moraines' in a formal sense and therefore have to introduce an alternative, particularly when it is commonly used informally and well understood. The glaciotectonic landform classification of Aber et al. (1989) does not offer an alternative, since the component landforms are not necessarily associated with the ice-marginal zone, and the term 'composite-ridge' does not encompass the full range of morphological forms observed as the product of glaciotectonism at modern ice margins.

For the purposes of this review the term 'push moraine' is defined as the product of construction by the deformation of ice, sediment and/or rock to produce a ridge, or ridges, transverse or oblique to the direction of ice flow in front of, at, or beneath an ice margin. This definition embraces such things as thrust-block moraines, thrust moraines, composite ridges and hill-hole pairs where they occur at, or close to, an ice margin and can be clearly linked to it.

Push moraines display a very wide range of different morphologies, at a range of scales from a few metres to features which extend for several kilometres, and are composed of an equally diverse range of sediments and/or rocks. In general, however, one can identify a morphological continuum from small, discrete, ice-marginal ridges formed by seasonal readvances of an ice margin, to multi-crested push moraines, with wide proximal–distal widths, formed by more sustained advances, and as with any continuum identifying distinct classes, is arbitrary and open to debate. It is, however, possible to devise a number of categories on the basis of such variables as: morphology, style of deformation, and the magnitude of proglacial shortening involved. Fig. 1 defines some of the variables with which to describe the anatomy of a push moraine and thereby subdivide the morphological continuum. One of the most important variables is the size of the wedge of glacial foreland that is deformed to give the push moraine. This wedge may be composed of both proglacial and subglacial sediment/rock and in some cases may also incorporate the ice margin itself. Prior to deformation, this wedge can be defined in terms of its thickness and width to give an aspect ratio (thickness:width) and when compared to the aspect ratio of the final push moraine, it gives an indication of the degree of glaciotectonic shortening involved (Fig. 1). The aspect ratio of the foreland wedge which is deformed, and the style of deformation within it, is controlled by such variables as the rheology of the sediment, the level of friction along the basal décollement surface, its ability to transmit stress in a proglacial direction, the compressional stress regime applied, and the geometry of the ice–sediment coupling at, or close to the ice margin (Fig. 1). Despite the range of variables involved, it is possible using a combination of morphology, deformation style, and the scale of shortening to recognise the following broad categories of push moraine associated with increasing compressive stress (Boulton et al., 1999):

(1) Small (≤ 5 m high) push moraines, with a single crest orientated parallel to the ice margin. Deformation occurs close to the ice margin in a narrow zone, often as a consequence of seasonal ice-marginal fluctuations. These push moraines are normally referred to as seasonal, or annual push moraines (Sharp, 1984; Boulton, 1986).

(2) Large (≥ 5 m high) push moraines, with a single crest orientated parallel to the ice margin, which result from a more sustained advance, usually due to a marked change in glacier mass balance.

(3) Narrow, multi-crested, push moraines in which significant deformation has been transmitted horizontally for the order of 50 to 300 m beyond the glacier margin, and through a thickness of perhaps 10 to 20 m, giving an aspect ratio for the undeformed foreland wedge of between 1:5 and 1:20. The style of

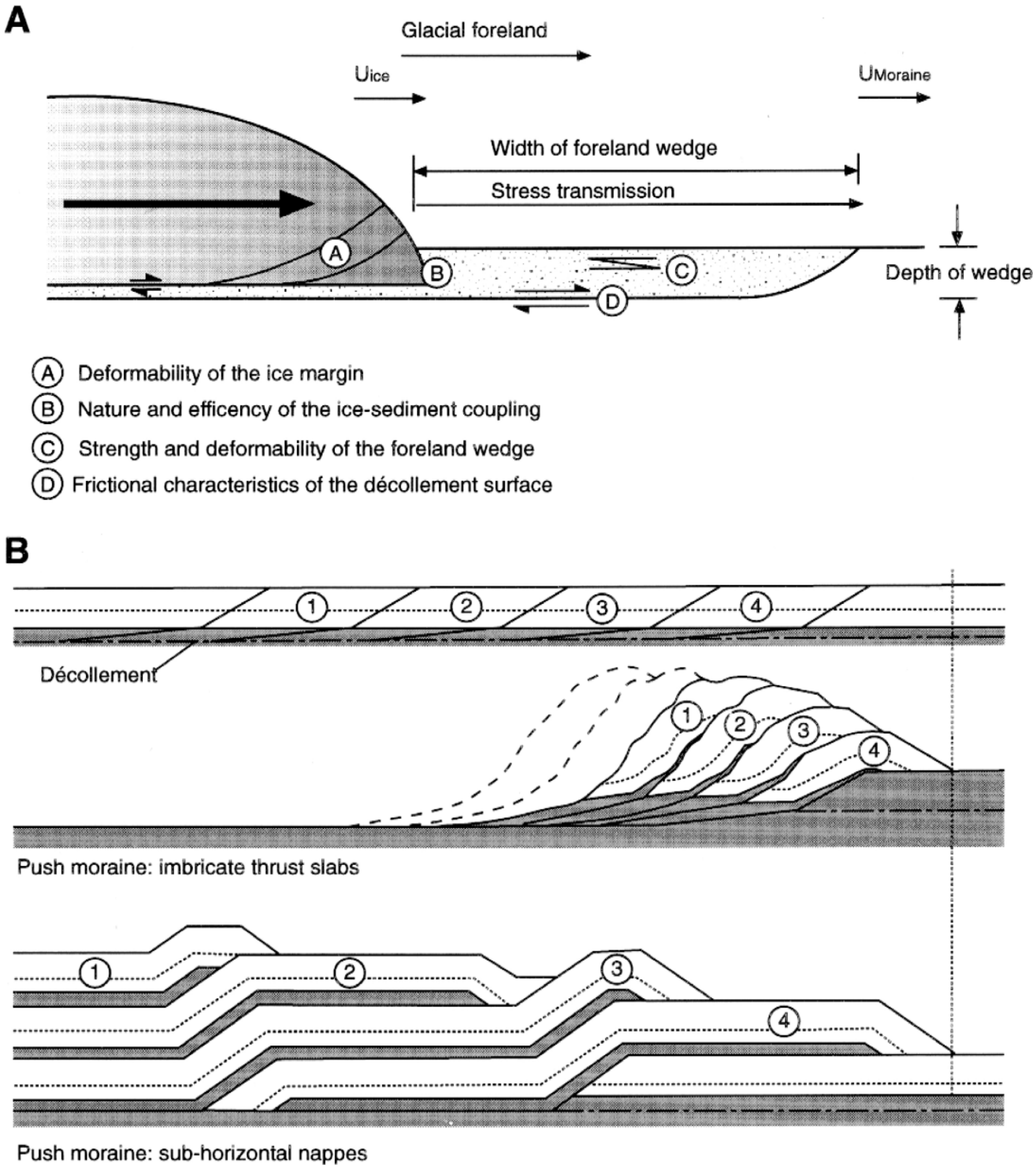


Fig. 1. The anatomy of a push moraine. (A) Some basic definitions and key variables. (B) Geometry of an imbricate push moraine compared to one built of nappes (modified from: Van der Wateren, 1995b).

deformation may involve multiple folds, or fans of listric thrusts.

(4) Wide, multi-crested, push moraines in which deformation has been transmitted in excess of 300 m beyond the glacier giving aspect ratios for the undeformed foreland wedge in excess of 1:20 and typically as much as 1:50. The style of deformation commonly involves either fans of imbricate thrusts, or superimposed sub-horizontal nappes produced by overthrusting.

Although somewhat arbitrary, particularly when distinguishing different examples of multi-crested push moraines, this classification does provide a starting point from which to review the morphology, structure and evolution of these landforms. Each class of push moraine identified above is examined in turn, before the generic mechanisms and controls are considered, and their value to inverse modelling discussed. It is important to note that push moraines occur both in terrestrial and subaqueous settings, and that the mechanisms and principles involved are very similar (Boulton, 1986); however, this review uses for the most part terrestrial examples.

2. Seasonal push moraines

Extensive glacier retreat during the twentieth century has led to the formation of glacier forefields with sequences of recessional push moraines, each formed by a winter readvance (Fig. 2; Hewitt, 1967; Sharp, 1984; Boulton, 1986). Individual ridges vary in height from as little as 0.3 to 5 m, and typically have asymmetrical cross-profiles with gentler ice-proximal and steeper ice-distal slopes and are often associated with flutes (Van der Meer, 1997). In planform, individual moraines frequently show bifurcations and cross-cutting patterns, associated with variation in activity along the length of an ice margin (Fig. 3A; Sharp, 1984; Boulton, 1986; Bennett and Boulton, 1993; Krüger, 1994). Planform geometry is also controlled by the morphology and continuity of the ice margin. For example, the more crevassed and irregular a margin, the more irregular the resultant moraine pattern (Horsefield, 1983). At Bødalsbreen (southern Norway), Matthews et al. (1979) attributed a distinctive 'saw-tooth' planform to a crevassed glacier margin and also noted that the embayments were higher than the re-entrants due to the preferential concentration of debris within them. Moraines have also been observed forming on the distal faces of snow-banks or lake/river ice (e.g. Worsley, 1974; Birnie, 1977). Consequently, moraine formation may occur some distance in front of the active ice margin.

Small push moraines, similar in morphology to those just described, have been widely reported from subaqueous environments (e.g. Andrews, 1963a,b; Aartolahti, 1972; Holdsworth, 1973a,b; Boulton, 1986; Larsen et al., 1991). They are referred to by a variety of names, including De Geer moraines, wash-board moraines or cross-valley moraines. Caution is required, however, as some of these moraines clearly form as push moraines either formed by seasonal readvances (Boulton, 1986; Larsen et al.,

Fig. 2. Seasonal push moraine in front of Brieðamerkurjökull, in Iceland (photograph courtesy of G.S. Boulton).

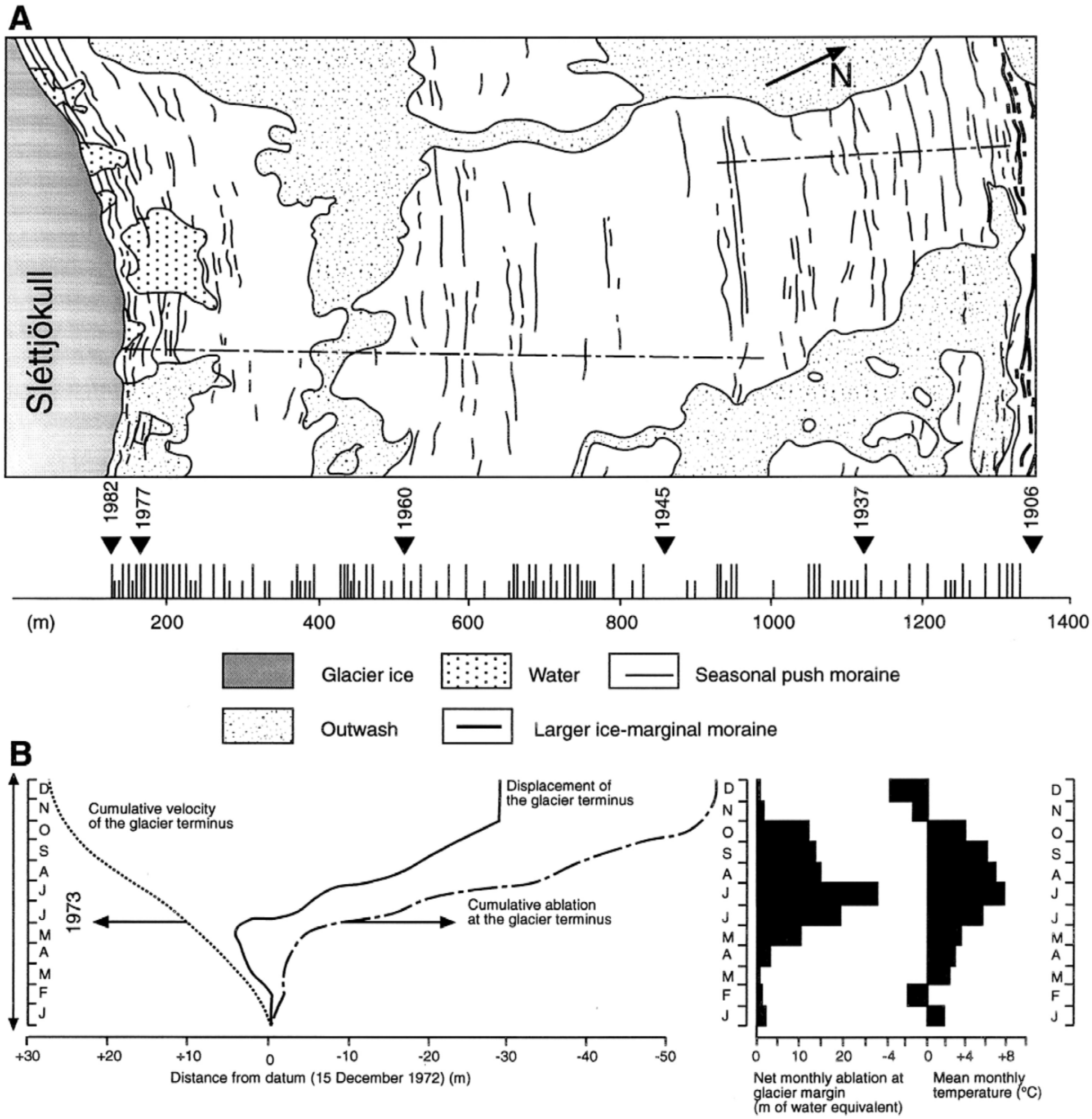


Fig. 3. Seasonal push moraines. (A) Seasonal push moraines in front of Sléttjökull mapped from arial photographs (map) and recorded in the field (long clear, short dash indistinct) along two overlapping transects (bar). The long lines on the bar represent clear examples, while the shorter dashes equal more indistinct moraines. Note the greater number of moraines identified in the field (modified from: Krüger, 1994). (B) Data from Brieðamerkurjökull to illustrate the balance between ablation and ice velocity at the margin that determines the presence or absence of a seasonal readvance. Also note the similarity between monthly ablation and mean monthly temperature (modified from: Boulton, 1986).

1991), or by iceberg calving (Holdsworth, 1973a,b), while other examples form by squeezing of subglacial sediment into basal crevasses (e.g. Elson, 1957; Hoppe, 1957; Stromberg, 1965; Lundqvist, 1989; Zilliacus, 1989), or simply by the coalescence of subaqueous fans along a stable ice margin.

Terrestrial examples tend to consist of a core of deformed sediment, with a re-worked surface and varying amounts of supraglacial debris. The importance of the supraglacial component is a function of the geometry of the ice margin which controls delivery and distribution of debris (Rogerson and Batterson, 1982), while the debris structure, and velocity of the glacier involved controls the rate of supply. The sediment within the deformed core depends on that available within the proglacial zone, but com-

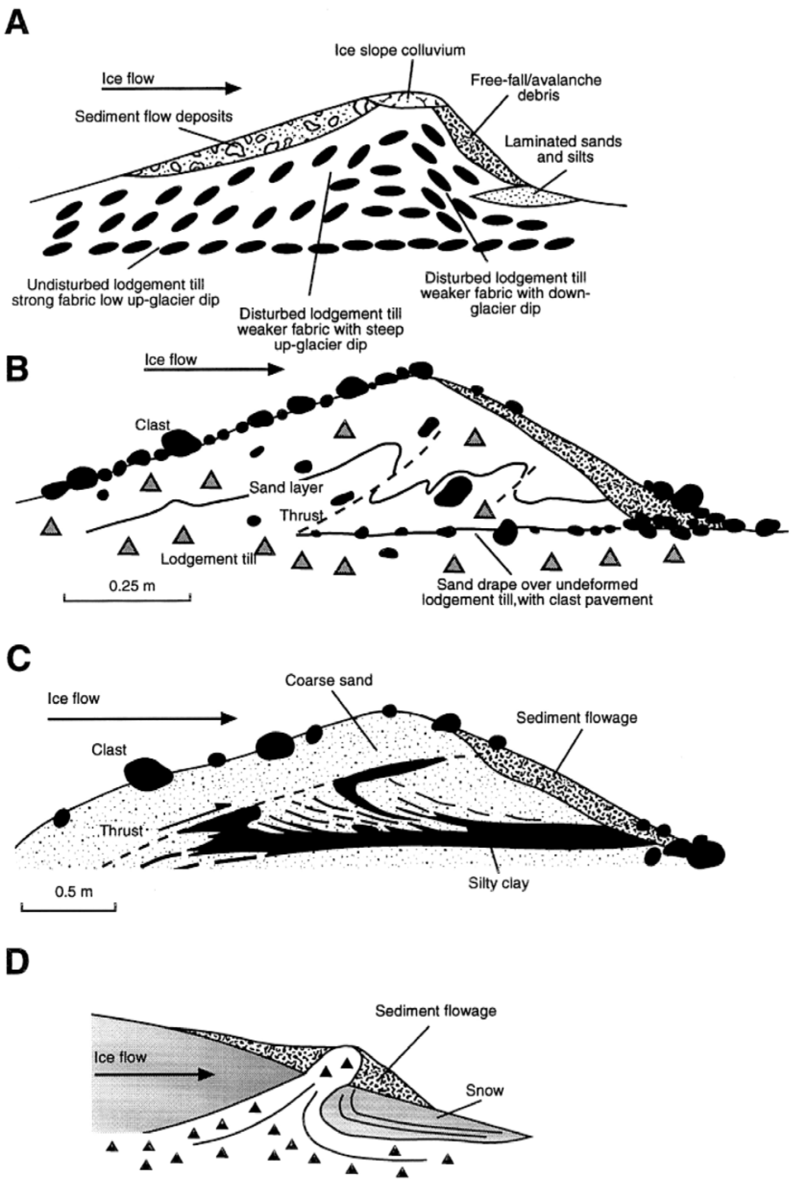


Fig. 4. Internal composition and structure of seasonal push moraines. (A) Idealised model of clast fabric and facies variation within a seasonal push moraine according to Sharp (1984). (B) Section through a ridge formed in the winter 1979/1978 in front of Sléttjökull (modified from: Krüger, 1994). (C) Cross-section through a seasonal push moraine in front of Heinabergsjökull in Iceland. (D) Seasonal push moraine formed by compression of sediment between a seasonal snow bank and an advancing ice margin. Similar structures form as a result of basal 'freezing-on' of sediment at the ice margin of Sléttjökull according to Krüger (1994). Based on data in Birnie (1977) and Krüger (1994).

monly consists of a mix of subglacial diamicton and a surface veneer of supraglacial clasts. A number of investigations have examined clast fabric within this diamicton core in order to infer patterns of strain experienced during deformation. One of the most systematic set of observations are those of Sharp (1984) who's results are summarised in Fig. 4A. These data show a strong up-glacier dipping fabric on the ice-proximal face and a down-glacier dipping fabric on the distal face. Price (1969, 1970) also noted a strong preferred orientation perpendicular to ridge axis (see also Andrews and Smithson, 1966). There are few observations of deformation structures within the core of these ridges. Krüger (1994) describes the presence of several listric thrust faults and small deformation structures in sands within a push moraine in front of Sléttjökull (Iceland; Fig. 4B), while Fig. 4C shows the internal structure within a push moraine in front of Heinabergsjökull (Iceland). Here the moraine is composed of coarse sand and silt units truncated by at least one thrust, which is highlighted by a footwall drag fold in silty clay (Fig. 4C).

Despite numerous investigations, the origin of the moraine core remains a source of debate. Price (1970) advocates a mechanism of moraine squeezing, in which saturated till is pressed from beneath the ice margin to form the ridge (see also Hoppe, 1952; Andrews and Smithson, 1966). Similarly, Sharp (1984) advocates a mechanism of till advection by subglacial deformation to the ice margin. In contrast, Krüger (1994) suggests that 'freezing-on' of a slab of subglacial diamicton to the sole of a thin ice margin, prior to its seasonal readvance, may lead to moraine formation as the sediment slab is released by summer warming. Other models invoke the presence of ice-proximal snow-banks and the deformation of sediment along the suture between the advancing glacier and the frontal snow-bank (Fig. 4D; Birnie, 1977). In these models, the incorporation of winter snow within the ice distal slope causes significant post-depositional re-sedimentation. In practice, a range of different modes of moraine formation may operate to produce the core of these ridges.

The geographical distribution of seasonal push moraines is restricted to warm-based and active glaciers. This reflects the critical interplay between glacier activity and the rate of ice-marginal ablation, which defines the presence or absence of a seasonal readvance. The position of an ice margin through time is determined by the linear component of ablation and glacier velocity at the ice margin. Where horizontal ice velocity exceeds the linear component of ablation, the glacier advances and where ice velocity is less than ablation, the glacier margin recedes. Ice-marginal ablation is controlled by such variables as air temperature, or the number of snow-free days at the ice margin, and consequently varies seasonally. In contrast, ice flows throughout the year and as a consequence may exceed ice-marginal ablation in winter, causing the margin to readvance and thereby producing a push moraine. It is this seasonal readvance which is essential for the formation of seasonal push moraines. The mechanics of such seasonal readvances are well illustrated by data from Breiðamerkurjökull (Fig. 3B). In more continental locations, glacier velocity is too small to initiate seasonal readvances even though winter ablation rates are minimal (Boulton, 1986), which emphasises the climatic control on the spatial distribution of this type of push moraine.

A consequence of this seasonal signature is that moraine spacing should correlate with the rate of glacier retreat and consequently with such variables as air temperature, or in locations inhibited by late lying snow, the number of snow-free days each year (Timmis, 1986). The linkage between moraine spacing, retreat rates, and air temperature has been examined in several investigations and good correlations between these variable have been obtained (e.g. Sharp, 1984; Timmis, 1986; Krüger, 1994). This implies that moraine spacing may be used to obtain data on inter-annual variability in summer ablation, or even as a proxy for such variables as summer air temperature.

3. Large, single ridge, push moraines

The push moraines described in this section are typically greater than 5 m in height, and are the product of sustained advances of the ice margin. The distinction between this scale of push moraine and those produced by seasonal readvances is not clear-cut and the 5-m height chosen here is rather arbitrary. The key distinction, however, is between moraines produced by seasonal ice marginal fluctuations and those associated with more sustained advances.

One of the most comprehensive studies of a single-crested push moraine formed by a sustained glacier advance is provided by a Danish team, working at Höfðabrekkujökull, an outlet glacier of the Myrdalsjökull ice cap in Iceland (Humlum, 1985; Heim, 1983; Krüger, 1985, 1994). Between 1979 and 1982, Höfðabrekkujökull advanced at a rate of 10 m per year (Humlum, 1985) and then at a rate of 32 m per year between 1982 and 1985 (Krüger, 1985, 1994). This produced an ice front 10 to 20 m high, with an angle of between 25° and 56° (Humlum, 1985). A push moraine between 1 and 5 m high developed preferentially along those sections of the ice margin characterised by glaciofluvial sediments (Humlum, 1985), although lodgment till was also deformed in some locations (Krüger, 1985). Transverse sections revealed an internal structure dominated by an imbricate stack of thrust slabs (Fig. 5). Each slab was between 0.5 and 1 m thick, 10 m wide, with an up-glacier dip of between 25° and 30°, and a lateral extent along the ice margin of between 50 and 150 m. Internally, the slabs were either undeformed, contained small-scale kink folds with fold axes parallel to the moraine crest and axial planes dipping up-glacier at between 0° and 15°, or had a recumbent form. The initiation of thrusting may have been facilitated by a groundwater surface at or close (< 1 m deep) to the ground surface (Krüger, 1985).

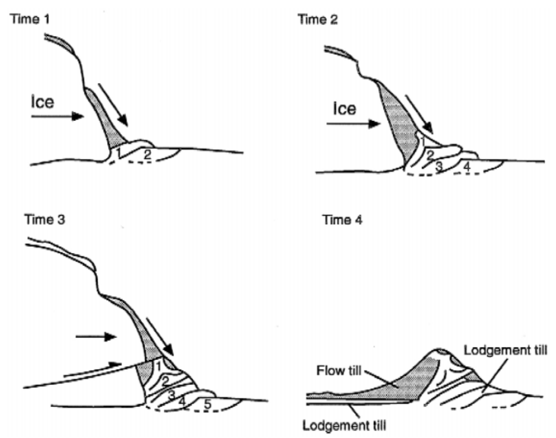


Fig. 5. Morphology and internal structure of the Höfðabrekkujökull push moraine (modifed from: Humlum, 1985; Krüger, 1985).

These push moraines formed as the glacier dislocated slabs of either lodgment till or glaciofluvial sediment in a progressively ice-distal direction to form an imbricate stack of small thrust sheets (Fig. 5). Humlum (1985) identified rapid glaciofluvial sedimentation (3 cm per day) against the ice margin as an important factor in moraine formation since burial of the glacier snout facilitated the coupling of the ice margin to unfrozen proglacial sediment.

Subsequent work by Krüger (1993) at Sléttjökull (Myrdalsjökull, Iceland) has demonstrated that a moraine of similar internal composition can form at a stationary ice margin, without recourse to proglacial tectonics. At Sléttjökull, the ice margin was both thin and stationary, and except for minor seasonal oscillations yet had a large, actively forming push moraine composed of an imbricate stack of lodgment till slabs. Krüger (1993) argued that this moraine formed as a result of seasonality in the basal thermal regime of the ice margin. He suggested that each winter, the thin marginal ice become cold-based, 'freezing-on' a slab of lodgment till to the glacier bed, which was then ripped-up and moved forward as the ice margin advanced at the end of each winter season. These slabs were then deposited in summer as the margin becomes warm-based to form, over a number of yearly cycles, a superimposed stack. Internally, the structure of these moraines is similar to that described from Höfðabrekkujökull formed by a sustained glacier advance (Krüger, 1985), except that the slabs contain no evidence of internal deformation. A similar type of moraine formation has been described by Matthews et al. (1995) from Styggedalsbreen in Norway, although in this case it involves the deposition of a double layer of sediment. Sediment frozen both on top and below a thin ice margin advances with it each winter and then melts to deposit a double layer of sediment. In this way, the moraine grows each year by the addition of each double layer of sediment. This work illustrates how ice-marginal moraines with similar morphological and structural architectures may form as a result of very different processes. This is an observation that emphasises the need for caution when interpreting the geomorphological record of former ice margins.

4. Narrow, multi-crested, push moraines

This category of push moraine involves the displacement of sediment/rock over 50 to 300 m in front of the glacier in response to a sustained advances, or surges, producing multiple ridge crests. In contrast to the previous forms, the ice no longer lies along each of the ridge crests.

Croot (1987) describes the multi-crested, push moraine which resulted from a surge of Eyjabakkajökull (Iceland) in 1890. The moraine complex consists of seven groups of ridges with widths, perpendicular to the former ice margin, of between 50 and 280 m. Individual ridge heights vary from 5 to 40 m. The ridges display a pronounced asymmetry close to the former ice margin, which decreases in a distal direction, along with ridge height. The ridges are composed of units of outwash, peat, tephra, turf and organic soils. Internally, the ridge complex can be divided into an ice-proximal and a proglacial zone. The ice-proximal zone is dominated by a series of large down-glacier dipping faults; the product of extensional deformation. This is balanced in the proglacial zone by compression in the form of a stacked sequence of listric thrusts which dip up-glacier, and separate an imbricate stack of nappes (54% shortening). The two tectonic systems are linked by a sole thrust, or basal décollement (Fig. 6). Croot (1987) argued that the sediments were unfrozen during deformation and that release of subglacial meltwater and pressurised groundwater along thrusts played a major role in their lubrication. As evidence for this he cited the presence of dry drainage channels that radiate from the ice margin and from thrust surfaces. A distinct tectonic gradient was also noted across the push moraine from the ice-proximal zone, where individual ridges were high, asymmetrical and associated with nappes, to the distal zone where ridges were low, symmetrical and formed by individual anticlines, or rooted recumbent folds. The push moraine appears to have evolved in two phases (Croot, 1987): firstly, an imbricate stack of thrust slabs developed in front of the ice margin, and then secondly, this stack was rotated and augmented along a basal décollement which experienced simultaneous extension in the hinterland beneath the ice margin and compression in the glacial foreland (Fig. 6).

A moraine complex of similar dimensions to that at Eyjabakkajökull, but with a different style of tectonic regime, has been documented by Boulton et al. (1976) in front of the Maktak Glacier (Baffin Island). This moraine complex marks the Neoglacial ice maximum and unlike that at Eyjabakkajökull was not formed by a surge. In this moraine, the glaciofluvial sediments of which the moraine is composed yielded predominantly by folding as opposed to fracture and were again unfrozen during deformation (Boulton, 1986).

Eybergen (1986) describes a similar scale of push moraine from Turtmannglacier in Switzerland (Fig. 7). Prior to moraine formation, the glacier forefield consisted of a drumlinised lodgment till surface, upon which an outwash fan had built-up. During a series of readvances since 1971, this unfrozen outwash fan was deformed, along with some of the underlying lodgment till, to form a multi-crested push moraine. The whole moraine complex was 100 m long, and 40 m wide, with a maximum elevation of 11 m. In 1986 the glacier was advancing at between 5.6 and 9.2 cm per day, while the ice-prox-

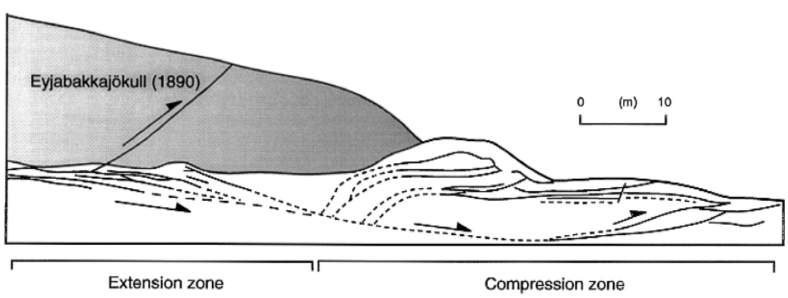


Fig. 6. Morphology and internal structure of the Eyjabakkajökull push moraine (modified from: Croot, 1987).

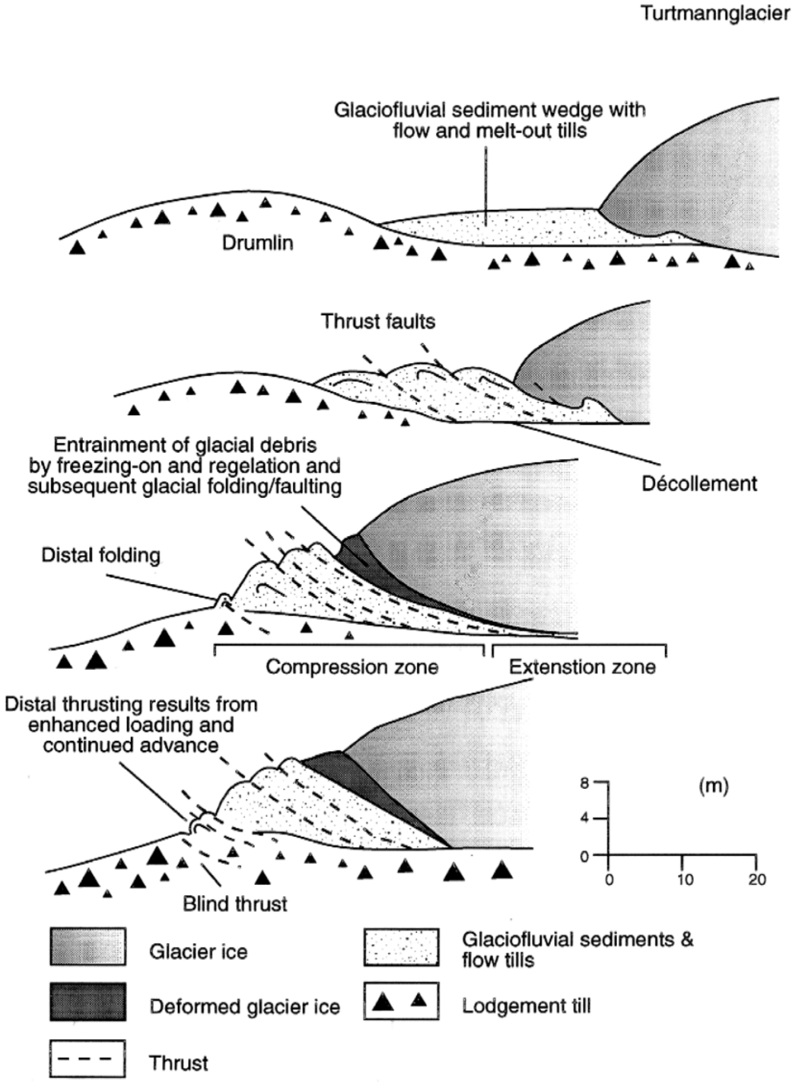


Fig. 7. Morphology, internal structure and evolution of the Turtmannglacier push moraine in Switzerland (modified from: Eybergen, 1986).

imal face of the push moraine was advancing at 4.7 cm per day and the ice-distal face at 1.7 cm per day. Internally, the push moraine consisted of an imbricate stack of thrust slabs, separated by listric thrusts dipping up-glacier at between 20° and 30°. Some folding was evident within the thrust slabs and folding of debris-rich basal ice occurred at the glacier margin as it was compressed against the push moraine. This resulted in the elevation of basal-debris within the ice margin and consequently the development of ice-cored topography immediately behind the push moraine. Deformation appeared to have involved the successive displacement of thrust slabs in an ice-distal direction, a process maintained by a positive mass balance and associated thickening of the ice margin (Eybergen, 1986). The décollement, or sole thrust, from which the listric thrusts rise was located on the upper surface of the lodgment till and was lubricated by the build-up of water pressure along the junction between the overlying porous gravels and the relatively impermeable lodgment till below (Eybergen, 1986).

Bennett et al. (2000) describe a similar scale of push moraine formed by a Neoglacial advance of Hagafellsjökull Eystri (Iceland) against a steep reverse bedrock slope. The significance of this push moraine complex is that it appears to have been completely over-run during the glacial maximum and its morphology is the product of subsequent meltwater erosion during deglaciation. Similarly, Hart (1994) describes a truncated moraine complex at Melabakkar–Ásbakkar (Iceland) where the core of a push moraine has been overrun and truncated by subglacial deformation. Benn and Clapperton (2000) describe a series of drumlins around the central Magellan Straits in Chile, which appear to have originated as ice-thrust, push moraines at a polythermal ice margin. These moraines were overrun by advancing ice and re-shaped to form the core of drumlins. This work simply emphasises that push moraines need not form at ice-margins and that they may be overrun and re-worked to form new landforms.

The Bride Moraine is one of the best exposed moraine complexes in the British Isles and forms a large moraine ridge just south of the Point of Ayre on the Isle of Man (Slater, 1931; Thomas, 1984a). The stratigraphy consists of the Shellag Till, which is overlain in turn by sands and gravels. The core of the moraine ridge consists of a series of diapirs, which are cut by high-angle normal and reverse thrusts (Thomas, 1984a). In front of this in an ice-distal direction are a series of low angle overthrusts. The till shows a well-developed tectonic foliation, not dissimilar to slaty cleavage. Thomas (1984a) argued for the presence of both proglacial permafrost, on the grounds of epigenetic and syngenetic ice wedges, and for elevated water pressures within the Shellag Till, in order to facilitate deformation. He suggested that saturated, low-permeability till, beneath a proglacial lid of permafrost, resulted in elevated water pressures beneath and in front of the former ice margin. As the ice margin advanced, loading and compression of the till occurred resulting in the formation of a series of rooted, diapiric folds at the ice margin. Continued compression led to thrusting within these diapiric folds and to over-thrusting in the foreland (Thomas, 1984a). The significant feature of this push moraine is that the folds remained rooted and that widespread décollement did not occur at the base of the moraine complex. As a consequence, the moraine complex remained relatively narrow and located close to the former ice margin. The absence of widespread décollement also explains the intensity of deformation observed, since the stress was absorbed by internal deformation rather than by propagation of the moraine complex into the foreland.

All the examples described so far have involved the deformation of a significant wedge of sands and gravels, in most cases due to ice advance into some form of outwash fan. A range of deformation styles have been observed, but the presence of an imbricate stack of thrusts linked to a basal décollement is common (Fig. 1B). There is, however, one example of a push moraine complex which fits within this category in terms of the scale of deformation, but is different in form, reflecting a specific process regime. Boulton et al. (1996) described the landform assemblage associated with a soft-bed surge of Sefstrømbreen (Svalbard). The glacier surged in 1882/1886 into Ekmanfjorden and over-ran the shoal/island of Coraholmen. The landform assemblage revealed on deglaciation consists of a network of crevasse-fill ridges behind a low push moraine complex. The push moraine complex is 250 m wide and consists of numerous low ridges up to 2 m high, separated by between 5 and 20 m (Fig. 7). In planform the ridge crest-lines have a conformable and parallel trace, although the ridges become more arcuate in an ice-distal direction. The ice-proximal face of the moraine complex consists of a steep, and deeply embayed, ice-contact slope. Except for the conformable planform pattern of individual ridge crest-lines, the moraine complex resembles a field of seasonal push moraines (Boulton et al., 1996). However, investigation of the internal composition of this moraine complex reveals a series of mushroom type, or diapiric folds, at or close to the former ice-contact slope, beyond which there are structures consistent with the proglacial flow of sediment. Boulton et al. (1996) suggest that the surge was propagated by rapid subglacial deformation over the glaciomarine and marine muds of the fjord floor. Subglacial sediment was advected by deformation towards the ice margin where it began to accumulate. At the end of the surge the glacier settled into this soft-bed intruding sediment into basal crevasses and extruding it proglacially. This fluid sediment flowed proglacially to form the push moraine complex, with each moraine crest essentially representing a pulse of flowing sediment (Boulton et al., 1996).

5. Wide, multi-crested, push moraines

Push moraines with a significant proximal–distal width (0.5 to 1.5 km) have been described from both the high-Arctic and the margins of Pleistocene mid-latitude ice sheets. They involve multiple ridge crests and the deformation of proglacial sediment wedges, with aspect ratios of the order of 1:30 to 1:50. One of the most detailed descriptions of such a large push moraine has been provided by Boulton et al. (1999) at Holmstrømbreen in Svalbard (Fig. 8; see also Van der Wateren, 1995a). Holmstrømbreen surged during its Neoglacial maximum into a large outwash fan, resting on glaciomarine and estuarine sediments. This resulted in a 1.5-km-wide zone of deformation and a moraine complex with a width of the order of 900 m. The moraine complex was superbly exposed in 1984 and recorded in detail by Boulton et al. (1989). An ice-proximal to ice-distal transition was observed with three main structural zones—an external, intermediate and internal zone—each characterised by a different intensity of deformation. The external zone (4 to 255 m) is characterised by open folds, with upright or slightly inclined axial surfaces. Most of these are Jura-style, box and concentric folds. Strong variation in thickness within the sand and gravel units is attributed to syn-tectonic erosion and deposition. This is replaced in the intermediate zone (255 to 690 m) by an increasing intensity of folding and thrusting. Tight over-turned folds and thrusts dominate and appear to have evolved from the open concentric folds typical of the external zone. The internal zone (690 to 950 m) is highly deformed, with strong evidence of overthrusting. Attenuated recumbent folds and a number of sub-horizontal, rootless nappes are present, some of which appear to have evolved through sediment flow on the surface of the moraine. In summary, the push moraine at Holmstrømbreen consists of an imbricate structure of thrust nappes showing a gradual increase in the intensity of deformation towards the former ice-contact face (Boulton et al., 1999). The thrust surfaces are linked via a basal décollement surface located at a depth of 30 m (Fig. 8). Deformation appears to

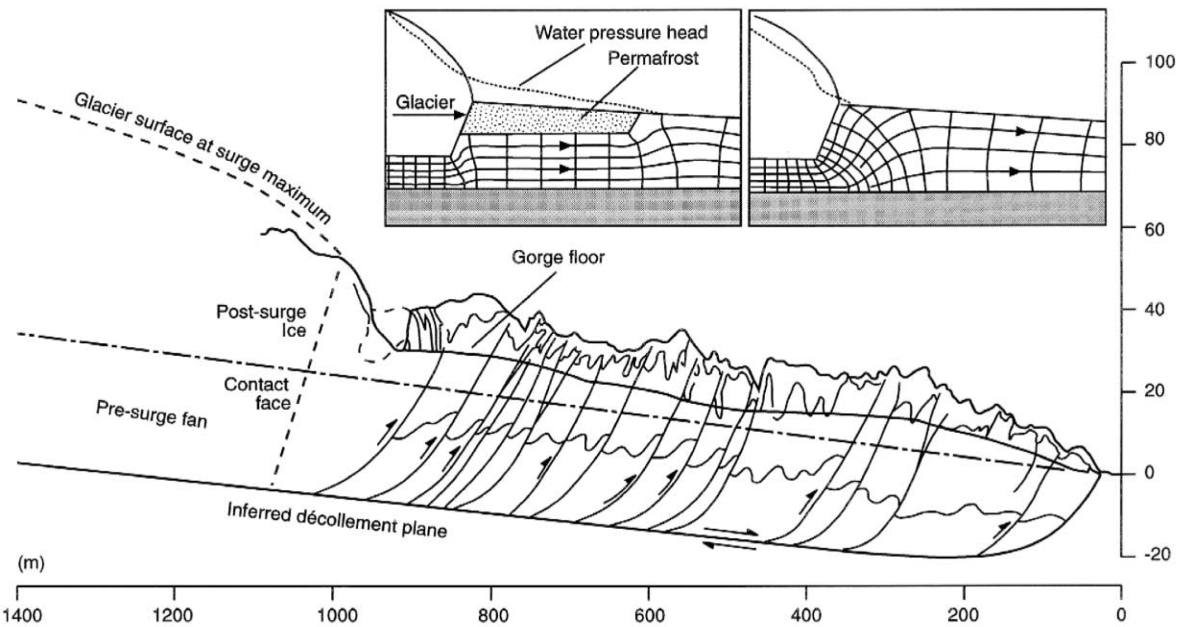


Fig. 8. Holmstrømbreen push moraine in Svalbard. (A) Schematic cross-section (modified from: Boulton et al., 1999). (B) Hydrogeological regime in front of Holmstrømbreen with and without proglacial permafrost (modified from: Boulton et al., 1999).

have started as a narrow zone of concentric folding above a décollement surface bordering the ice front. As the glacier margin continued to advance, these initial folds developed into recumbent tight folds and fold nappes, while a new zone of concentric folding developed in the foreland. Continued deformation lead to the development of rootless nappes and further propagation of the deformation front into the proglacial zone (Boulton et al., 1999).

Boulton et al. (1999) suggested that permafrost played an essential role in the formation of the Holmstrømbreen push moraine on two counts. Firstly, they argued that permafrost was necessary in order to transmit glacial stress over 1 km through a thin layer (20 to 40 m thick) of proglacial sediment. Secondly, permafrost was essential to explain the elevated water pressures necessary to reduce friction along the basal décollement. Boulton et al. envisaged a thin layer of proglacial permafrost with a warm-based and melting glacier behind. Subglacial melting and groundwater flow from beneath the glacier was unable to escape at the margin due to this thin lid of proglacial permafrost thereby promoting elevated pore water pressures immediately below the permafrost layer. As Holmstrømbreen surged, these high pore water pressures reduced friction along the décollement surface, located at the base of the permafrost, facilitating movement along the slip-plane. Boulton et al. (1999) use this model to suggest that permafrost was an essential pre-requisite for large-scale push moraine formation and that consequently, such push moraines in the Pleistocene record may in turn provide evidence of permafrost conditions.

It is frequently assumed that large complex push moraines, such as that at Holmstrømbreen, result solely from glacier surges (see Croot, 1988; Lefauconnier and Hagen, 1991). This has been challenged by Hambrey and Huddart (1995), who describe the formation of a large push moraine at Uvêrsbreen (Svalbard), not by a significant glacier advance, but by flow compression within a polythermal ice margin (Fig. 9). The moraine complex formed during the Neoglacial maximum and structural investigations of the glacier indicated that it has not experienced surge-type behaviour (Hambrey and Huddart, 1995). The moraine complex is over 500 m wide, and has a relative relief of 45 m. It consists of a broad belt of ridges and moraine mounds. Individual ridge crests

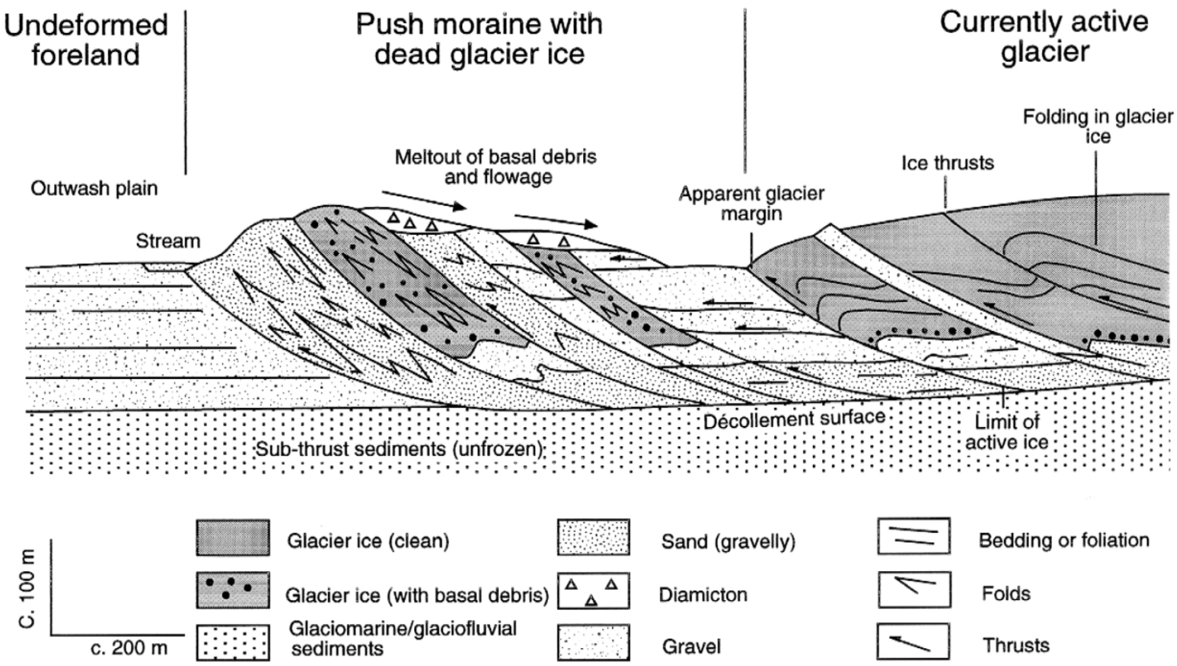
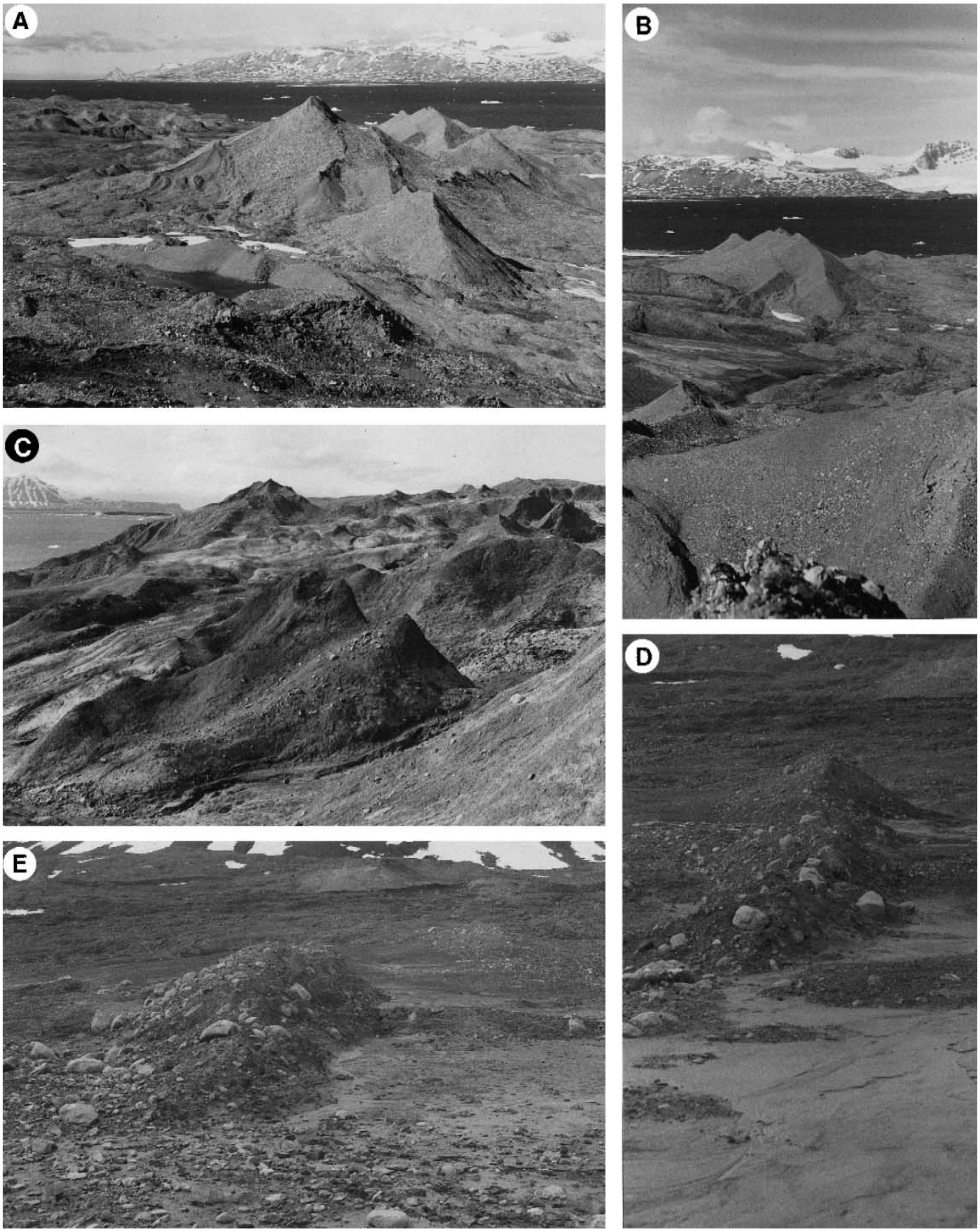


Fig. 9. Schematic cross-section through the Uvêrsbreen push moraine in Svalbard (modified from: Hambrey and Huddart, 1995).

are oriented transverse to the direction of ice flow and have well-developed ice-proximal faces, which dip up-glacier at between 30° and 45°. The complex is composed of gently inclined units of diamicton, sand, gravel and debris-rich ice, separated by high-angled thrust faults that dip up-glacier at between 30° and 40°. Individual thrust blocks frequently form identifiable mounds, or ridges, with the up-glacier rectilinear face of the mound/ridge being formed by the thrust plane. Each of the thrust sediment slabs shows relatively minor internal deformation, with only small scale folding of less competent strata, although the slabs of glacier ice within the moraine complex are highly deformed. Hambrey and Huddart (1995) argued that thrusting in the foreland is an extension of thrusting within the glacier, which is visible on the glacier surface where basal sediment is elevated along thrust planes into the body of the glacier. On ice retreat, this debris and intercalated ice is added to the moraine complex (Fig. 9). The thrusts within the ice are linked to those in the foreland by a sole thrust, located at the junction between glaciofluvial gravels and underlying glaciomarine muds. Hambrey and Huddart (1995) suggested that thrusting was initiated by flow compression within the glacier, associated with the thermal transition from warm-based sliding in the glacier interior, to cold-based non-sliding ice at the glacier margin. It is this compression which generated the push moraine, as opposed to any significant ice advance. Similar push moraines have been subsequently described in neighbouring parts of Svalbard by Huddart and Hambrey (1996) and Bennett et al. (1996a). This type of moraine architecture is consistent with the Ground Penetrating Radar data obtained from the push moraine complex at Scott Turnerbreen (Svalbard) by Lønne and Lauritsen (1996). Ronnert and Lankvik (1993) also stress the importance of thrusts within Svalbard push moraines.

The basic mechanics of the process of moraine formation described by Hambrey and Huddart (1995) were clarified by Bennett et al. (1996b) in an ice cliff orientated parallel to the direction of ice flow at the margin of Kongsvegen (Svalbard; see also Glasser et al., 1998). The ice cliff exposes, in cross-section, several thrusts within the marginal zone of Kongsvegen. Some of these thrusts are debris-rich, while others are debris-poor. The debris-rich thrusts contain rafts of subglacial diamicton varying in thickness from a few centimeters to over 2 m. In several cases these sediment rafts were thickened by folding and in general show increased thicknesses towards the base of the thrust at the glacier sole. The meltout of this debris-fill was observed at the ice margin; sediment in the upper part of the thrust simply melted out and was draped over the glacier forefield, while sediment at the base of the thrust was revealed as a distinct ridge without any associated ice core (Figs. 10 and 11). Variation in the angle and planform geometry of the thrust controls the height and morphology of the resultant ridge. The morphology of the moraines produced by this process is largely a function of thrust spacing. The thrusts observed in the ice cliff at Kongsvegen are widely spaced (<10 m) and consequently produce discrete ridges. However, large push moraine complexes, formed by this process, occur at the Neoglacial maxima of several glaciers, such as at Uvêrsbreen; here a concentration of proglacial and englacial thrusting occurred. Where thrusting is less concentrated a system of moraine mounds may form, similar in morphology to areas of hummocky moraine, and this type of moraine forming process has in fact been used to re-interpret some areas of hummocky moraine in Britain (Hambrey et al., 1997; Bennett et al., 1998a).

The role of rising debris bands or debris-rich thrusts has been emphasised in the past with respect to the formation of moraine mound complexes, or kame and kettle topography, particularly at polythermal ice margins, such as those in Svalbard (Boulton, 1972a; Paul, 1983; Sollid and Sørbel, 1988). In this case, the outcrop of the debris-rich thrusts and rising sediment bands produce an ice-cored topography in which re-sedimentation into ice-cored troughs and associated topographic inversion are the key features (Boulton, 1972a). This model is not applicable to the push moraine or moraine mound complexes described above as the product of thrusting, which leads to the question of why in situ thrust mounds form in some locations and re-sedimented moraine mounds or kames in others. Bennett et al. (1998a) suggest that one key variable may be the intensity with which thrusts crop out on the glacier surface. If the outcrop density is high, a thick supraglacial debris cover may result in which significant quantities of buried ice are incorporated. In this case the resultant landform is a moraine mound complex of kames formed by re-sedimentation and topographic inversion. These may contain cores of thrust material as described by Boulton (1967) at Sørbreen (Svalbard). If the outcrop density is less and the debris thickness greater, a moraine complex of in situ thrust



slabs results. These ideas have yet to be explored in detail, but suggest a continuum may exist between kame and kettle topography associated with re-sedimentation in ice cored topography and the formation of push moraines and moraine mound complexes formed by englacial thrusting.

So far the examples of push moraine, attributed to englacial thrusting, have not involved a glacier surge. However, this process of moraine formation has been used to explain the characteristics of a large push moraine complex at the head of Kongsfjorden in Svalbard formed by a surge (Fig. 12; Bennett et al., 1999a,b; Hambrey et al., 2000). The tidewater glacier complex of Kongsvegen and Kronebreen surged in 1948 (Fig. 12) down Kongsfjorden, resulting in moraine complexes on the south-west and north-east sides of the fjord. At the peak of the surge, the ice limit was located along the crest of a distinct ramp on the south-west side of the fjord and was associated in the north-east by a proglacial complex of thrust blocks. On deglaciation, a large moraine complex was revealed inside this ice limit, formed by thrusting within the body of the glacier (Figs. 10 and 13; Bennett et al., 1999a). On the north-eastern side of the fjord, moraine formation resulted from normal compression as the ice advanced against the fjord-walls, while in the south-west an element of transpression appears to have been involved (Fig. 12). In the north-east, the moraine complex, inside the proglacial thrust blocks, is composed of an imbricate stack of mounds and ridges (Fig. 10), while on the south-western side of the fjord a huge moraine ridge system (Figs. 10, 12 and 13), cored in part by glacier ice and composed of subaqueous outwash derived from the fjord floor, dominates the moraine complex (Bennett et al., 1999a,b). In general, the scale and density of thrusting within the ice body appears to have been much larger than that described from the examples documented in front of polythermal glaciers (Hambrey and Huddart, 1995; Huddart and Hambrey, 1996; Bennett et al., 1996a) and large slabs of fjord bottom sediments, mainly sands and gravels of a former grounding-line fan, have been transported on shore to form the moraine complex. Although the morphology and structure of the moraine complex are clear, the mechanism involved in its formation, and in particular the entrainment of large slabs of fjord bottom sediments into the body of glacier, are both uncertain and challenging. It is unlikely that the glacier was cold-based during the surge and consequently compression was probably due to both the passage of the surge front through the ice tongue and subsequent, rapid, ice-flow against the reverse bedrock slope of the fjord sides. The displacement of large proglacial thrust slabs is also a feature of the fjord bottom (Whittington et al., 1997; Hambrey et al., 2000).

Thrusting within the body of the ice margin appears therefore to be an additional mechanism by which large push moraines may form. Examples have been recorded from polythermal glaciers where compression is due to deceleration across a thermal boundary (Hambrey and Huddart, 1995; Bennett et al., 1996a; Huddart and Hambrey, 1996), at surge-type glaciers where compression results from the passage of the surge-front (Hambrey et al., 1996; Bennett et al., 1999a), and at glaciers advancing against steep reverse slopes (Hambrey et al., 1997; Bennett et al., 1998a,b, 2000). Several points are worth emphasising: (1) push moraines of this sort need not necessarily result from glacier advances; (2) they do not form exclusively in the proglacial zone and therefore need not mark the former ice front; (3) the mechanism appears to operate under a range of thermal regimes; and (4) the moraines produced are similar in morphology and structure irrespective of the cause of the flow compression (e.g. surge or no surge).

Fig. 10. (A and B) Kongsvegen push moraine complex (Svalbard) formed during the 1948 surge on the southwest side of the fjord (Fig. 12). These gravel ridges were formed by thrusting of sediment into the body of the advancing glacier. Fig. 13 gives a oblique view of the same moraine complex. The ridge system is composed of sands and gravels deposited in grounding-line fan on the fjord floor. These gravels were transported on-shore during deformation (see Bennett et al., 1999a). (C) Part of the Kronebreen push moraine complex (Svalbard) formed by normal compression on the north-eastern side of Kongsfjorden during the 1948 surge of Kongsvegen (Fig. 12). The moraine mounds visible are formed from glaciomarine diamctons transported on-shore within englacial thrusts (see Bennett et al., 1999a). (D and E) Small (<3 m high) ice transverse ridges formed at the base of indivual thrusts (Fig. 11). Similar ridges can be seen melting out of the base of thrusts within a vertical section, parallel to the direction of ice flow, within the margin of Kongsvegen (see Bennett et al., 1996b).

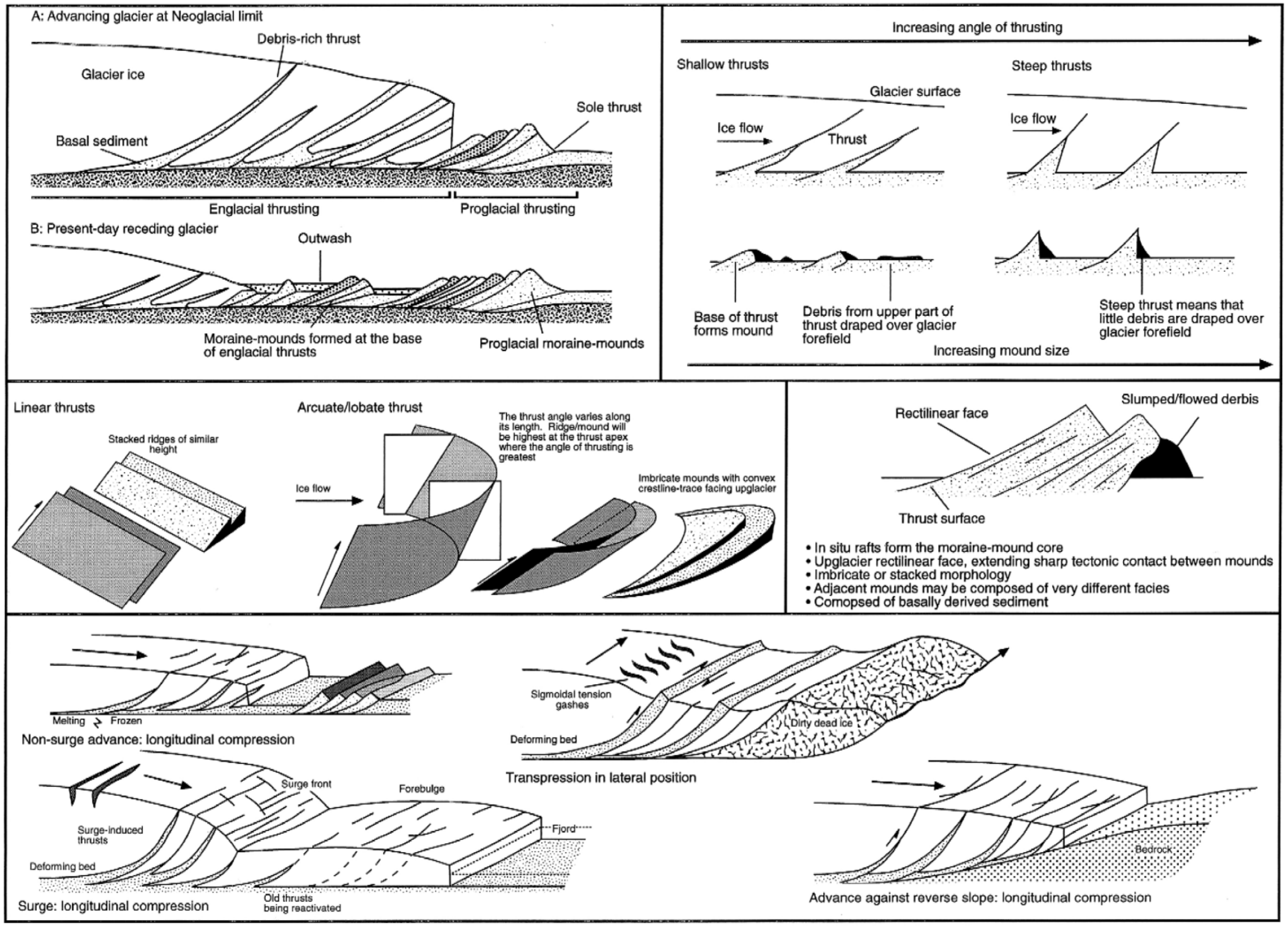


Fig. 11. Model of thrust moraine formation Svalbard (based on: Bennett et al., 1998a; Hambrey et al., 2000).

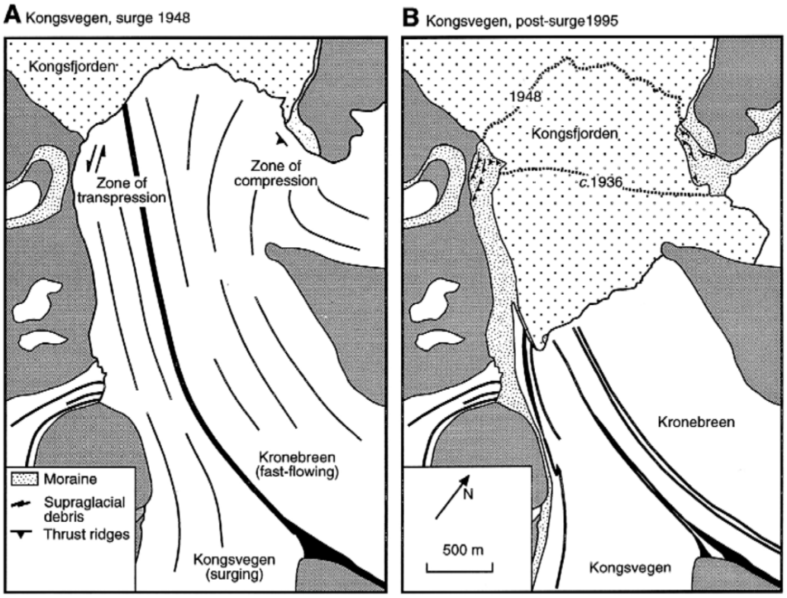


Fig. 12. Kongsfjorden during the 1948 surge of Kongsvegen.

One of the most challenging aspects of this model of moraine formation is understanding the processes by which sediment is entrained along thrusts into the body of the glacier. The traditional explanation for sediment entrainment within englacial thrusts is that the glacier 'freezes-on' basal sediment, due to a change in basal temperature, prior to thrusting (e.g. Weertman, 1961; Boulton, 1972b; Kleman and Hättestrand, 1999). However, this mechanism presents a number of difficulties. Firstly, the 0°C isotherm has to descend into the underlying sediment to a considerable depth in order to account for the thickness of some of the sediment slabs observed, although bed-parallel thrusting and folding may help to thicken the slabs. Secondly, not all the glaciers associated with this type of process are polythermal. A more radical explanation is that the glacier, its subglacial zone and proglacial zone acting as a single tectonic unit which is deformed by compression in a similar way to layers of strata, of varying competence, are commonly cut and shortened by a series of listric thrusts, rising through each layer from a basal décollement. This décollement must therefore run both beneath the glacier and into the proglacial zone and is effectively the surface on which the glacier and associated sediment body is moving en masse. This type of model is viable irrespective of the thermal regime of a glacier, provided that sliding is occurring along a basal décollement within the sediment pile below the glacier bed and that the overlying sediment is sufficiently rigid and strongly coupled to the glacier. In the case of polythermal glaciers this décollement may be associated with the base of the permafrost, or in warm-based examples, such as Kongsvegen, associated with the base of a layer of free-draining, and therefore non-deformable, gravels resting on deformable, mud-rich, diamictons. This is a challenging model, but fits the observed facts and emphasises the idea that the glacier, and sediment beneath and in front of it, may act as a tectonically coupled system.

The importance of proglacial thrusting within the formation of large push moraine complexes has also been well illustrated in the Canadian Arctic archipelago, but here the glacier does not appear to be deformed and incorporated into the push moraine complex, as is the case in the examples described above from Svalbard. The term 'thrust-block moraines' has been used to describe these proglacial push moraine complexes in the Canadian Arctic (Kalin, 1971; Klassen, 1982; Evans, 1989; Evans and England, 1991; Lehmann, 1992). These push moraines typically involve the proglacial dislocation, along listric thrusts, of slabs of permafrozen gravels.

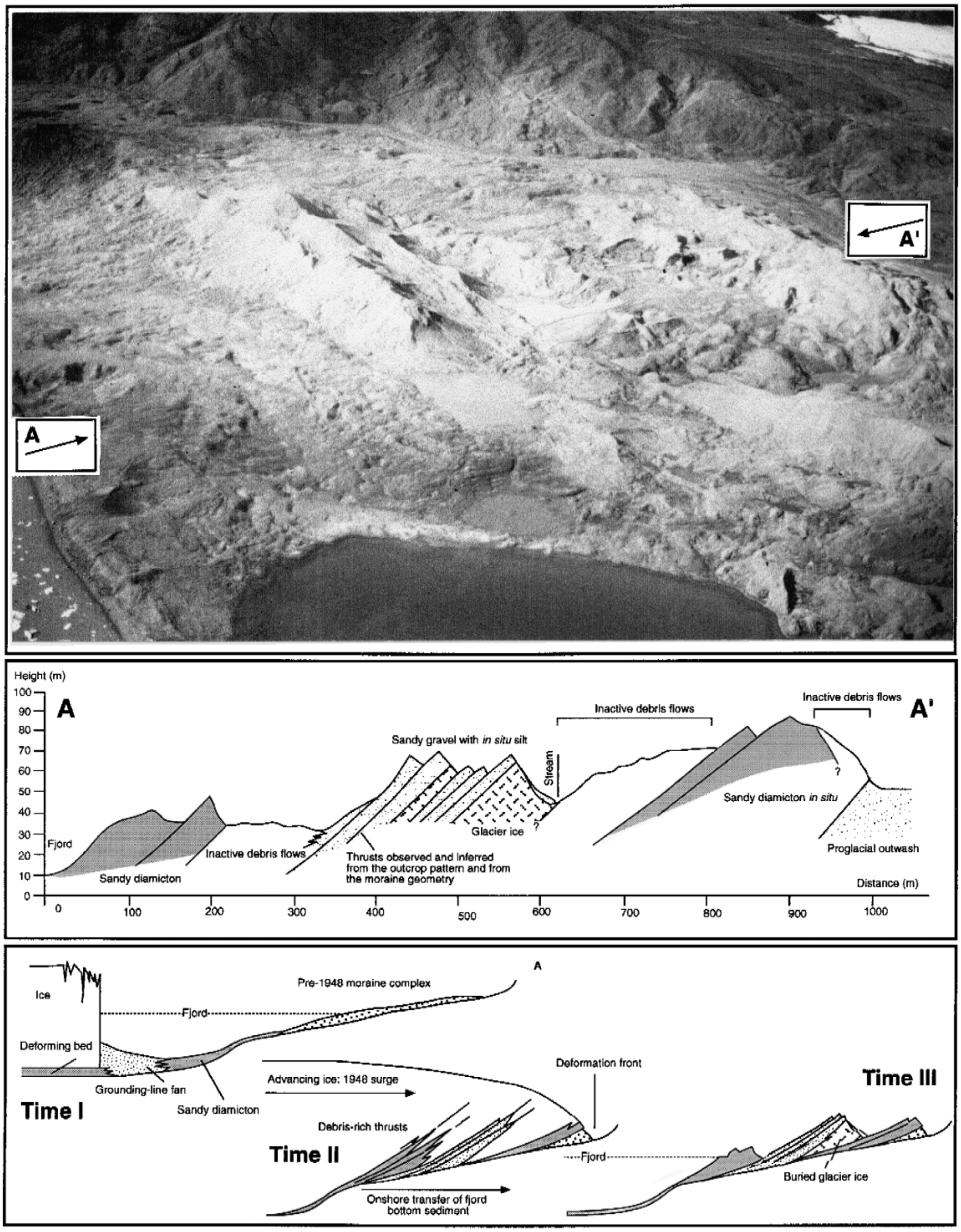


Fig. 13. Push moraine formed by englacial thrusting during the 1948 surge of Kongsvegen on the south-west side of Kongsfjorden in Svalbard. The upper panel shows an oblique areal photograph of the push moraine and the line of section shown in the middle panel (photograph courtesy of M.J. Hambrey). The lower panel is a schematic of the moraine formation (based on data in: Bennett et al., 1999a,b).

The décollement surface is located between the gravels and underlying finer-grained sediments. Perhaps the best and most famous example is the push moraine described by Kalin (1971) from the Thompson Glacier, an outlet of the McGill Ice Cap (Axel Heiberg Island) which underwent sustained advance during the 1960s. Between 1960 and 1967, the Thompson Glacier advanced over 180 m (ca. 7.1 cm per day; Kalin, 1971), to produce a push moraine over 900 m wide (distal–proximal width) along the eastern part of the ice margin (Fig. 14). The moraine consists of back-tilted blocks of outwash that have been displaced along a series of thrusts. The complex is dissected by an intricate drainage system of both active and abandoned channels. Many of these channels are blind and are intersected by thrust ridges, reflecting a complex pattern of syn-tectonic drainage evolution. Kalin (1971) examined rates of movement and deformation within the moraine complex over a 3-year period during the late 1960s (Fig. 14). These observations suggested that the rate of deformation may vary seasonally, with faster rates during the summer. This may reflect either seasonality in the rate of glacier flow and therefore compression, or alternatively, in the rate of deformation within the moraine itself. Given that the ice tongue is assumed

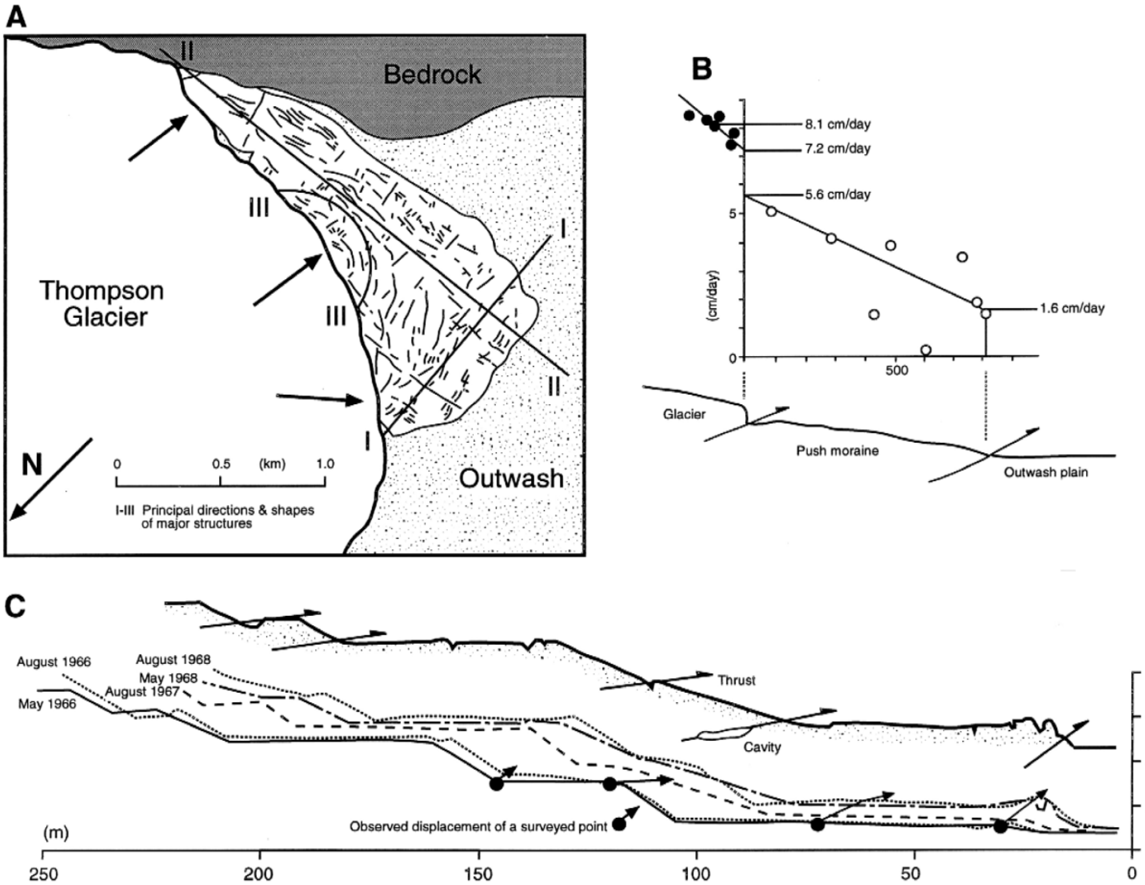


Fig. 14. Thrust-block push moraine in front of the Thompson Glacier, Axel Heiberg Island, in the late 1960s as recorded by Kalin (1971). (A) Map of the main structural elements within the push moraine. Key elements are picked out by the lines and roman numerals. (B) Surface velocity profile across the glacier, push moraine and outwash system during 1967 and 1968. The discontinuity between the various velocities can be accounted for by ablation at the glacier snout and by compression within the push moraine. (C) Serial cross-sections showing the evolution of the push moraine through time. The same cross-section is shown twice, once with the main structural elements and once with the sequential changes from May 1966 to August 1968. All three panels are simplified from Kalin (1971).

to be frozen to its bed at the ice margin (Lehmann, 1992), it seems more likely that seasonal variation in water pressure within the moraine complex itself, and therefore in the friction along thrusts and other structures, causes the observed seasonality in deformation rates. This emphasises the importance of the hydrogeological setting of a push moraine in its formation, evolution and deformation history. Structurally the moraine complexes described by Kalin (1971) and elsewhere in the Canadian Arctic (e.g. Evans and England, 1991; Lehmann, 1992) are similar to the proglacial component of the thrust-dominated push moraines reported from Svalbard (e.g. Hambrey and Huddart, 1995). The snout of the Thompson Glacier shows extensive evidence of thrusting (Hambrey, personal communication), but these structures are largely debris-free and do not contribute sediment to the moraine complex.

Large thrust moraines (hill-hole pairs and glaciotectonic composite ridges) involving both bedrock and glacial sediments have been widely reported from the margins of Pleistocene mid-latitude ice sheets (e.g. Rutten, 1969; Dellwig and Baldwin, 1965; Andrews, 1980; Pedersen, 1987; Pedersen et al., 1988; Aber, 1988; Aber et al., 1989). The interpretation of these glaciotectonic ridges as ice-marginal or sub-marginal moraines is not always clear. Some features are clearly associated with former ice margins, others appear to have formed at ice margins before being over-ridden by advancing ice (e.g. Benn and Clapperton, 2000), whilst in most examples the relationships to a former ice margin are unclear (Aber et al., 1989). Consequently, their significance within a review of push moraines is uncertain. They do, however, show a similar internal structure, consisting for the most part of imbricate stacks of thrust slabs, or nappes composed of rock/sediment, with varying degrees of internal deformation. The ridge complex at Møns Klint in Denmark is perhaps the most famous European example (Slater, 1927; Aber, 1985). This complex consists of an imbricate stack of thrust Cretaceous chalk rafts, characterised by structural relief in excess of 150 m, formed by a succession of ice advances during the Pleistocene (Aber et al., 1989). In Britain a smaller example of similar thrust chalk rafts is found on the Norfolk coast at Sidestrand (Hart, 1990). In North America, the Dirt Hills and Cactus Hills in Saskatchewan, Canada, provide an equally impressive example. Here a ridge network between 3 and 5 km wide and with a surface area in excess of 1000 km^2 has been formed, which consists of imbricate thrust and fold blocks of Cretaceous sandstones and mudstones (Kupsch, 1962; Aber 1988). Like many of the thrust ridge systems in North America, this ridge system appears to define the margins of several ice lobes and has for the most part been over-ridden, although some sections appear to have formed completely within the proglacial environment. Aber (1988) argued that they are the product of rapid and differential loading of saturated mudstones at the base of the succession, which induced a horizontal pressure gradient (gravity-spreading) beneath the advancing ice margin sufficient to dislocate bedrock slabs and stack them in front of the advancing ice. Moran et al. (1980) identified three common associations for this type of thrust ridge complex in North America, in particular the presence of: (1) bedrock escarpments; (2) subsurface aquifers, particularly if they are in some way confined; and (3) identifiable ice-marginal positions. Ridge complexes tend to occur as 2- to 5-km-wide belts immediately inside ice margins, behind which more streamlined forms occur, such as cupola-hills. This reinforces the widely held assumption (Aber et al., 1989) that thrusting occurs either in proglacial permafrost, or at zones of thermal transition within the ice-tongue, while streamlining of earlier thrust blocks occurs in the wet-based interior.

The push moraines reviewed so far in this section are commonly composed of an imbricate stack of steeply inclined thrust slabs, with each thrust rising from a basal décollement (Fig. 1B). However, Van der Wateren (1981, 1987, 1995a) has provided evidence from northern Europe for an alternative structural form; namely the superposition of sub-horizontal nappes (Fig. 1B). One of the most spectacular Pleistocene end-moraine complexes in northern Europe is the Rehburg Line, which stretches over 500 km from the North Sea in the west to Hanover in the east and marks the approximate, if not synchronous, extent of glacier ice during the Rehburg Phase (Drenthe advance) of the Saalian Glaciation (Fig. 15A; Van der Wateren, 1995a). Several push moraines have been documented along the Rehburg Line, for example, Van der Wateren (1981, 1985) has described the morphology of the Utrecht Ridge

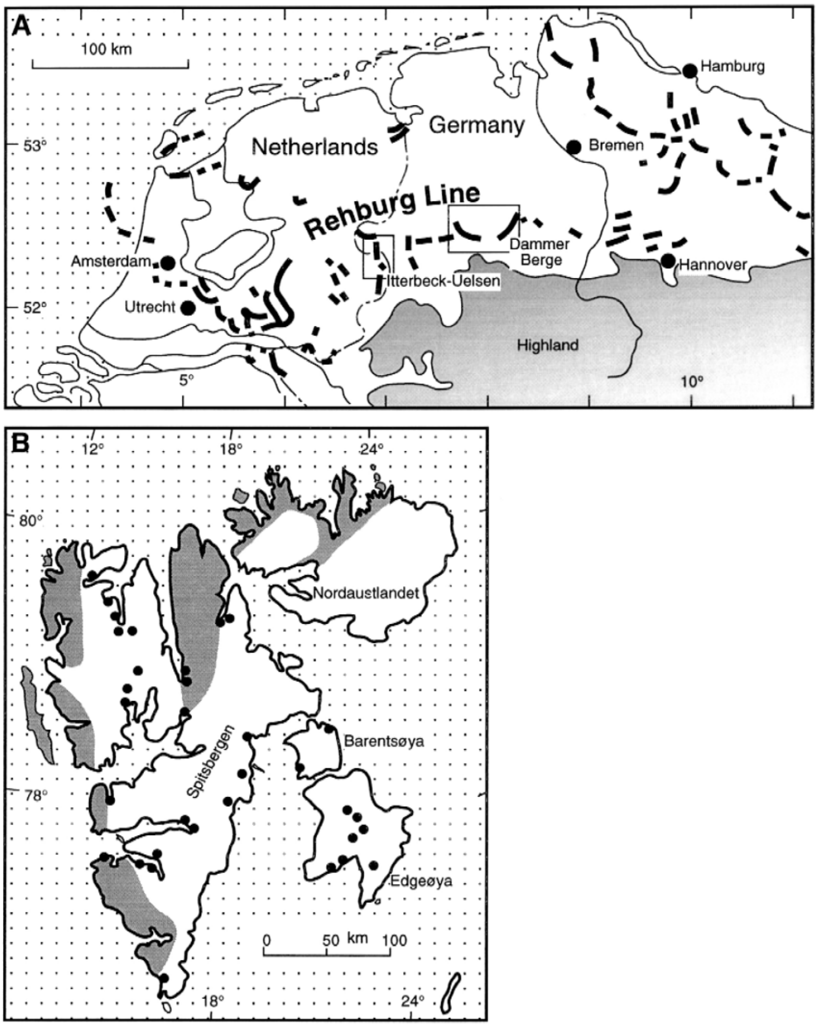


Fig. 15. (A) Map of northern Europe showing the Rehburg Line of push moraines. (B) Map of the distribution of push moraines in Svalbard.

in the Netherlands, and more recently that of the Dammer and Fürsterenauer Berge (Dammer Berge) in northern Germany (Van der Wateren, 1987, 1994, 1995a). In both cases, the push moraine architecture consists of a series of sub-horizontal nappes which have been displaced horizontally, in some cases as much as 6 km, producing forms that are typically 15 to 25 m thick, with aspect ratios of between 1:20 to 1:100. Each nappe is normally composed of coarse-grained glaciofluvial or Tertiary sands and gravels that have been displaced over a substratum of ductile clay or silt. Nappes are bounded above and below by shear zones. Folding and thrust imbrication occurs within some of these nappes, particularly towards their distal extremities, but these structures branch from the floor thrust which bounds the base of each nappe and do not rise from a basal décollement surface (Fig. 1B). Successively younger nappes appear to have formed progressively beneath older ones, carrying them forward into the foreland (Van der Wateren, 1995a). A similar structural architecture was observed by Klint and Pedersen (1995) in the Hanklit Thrust Complex in Denmark (Fig. 16).

From his observations along the Rehburg Line, Van der Wateren (1995a,b) developed the argument that a push moraine architecture of horizontal nappes

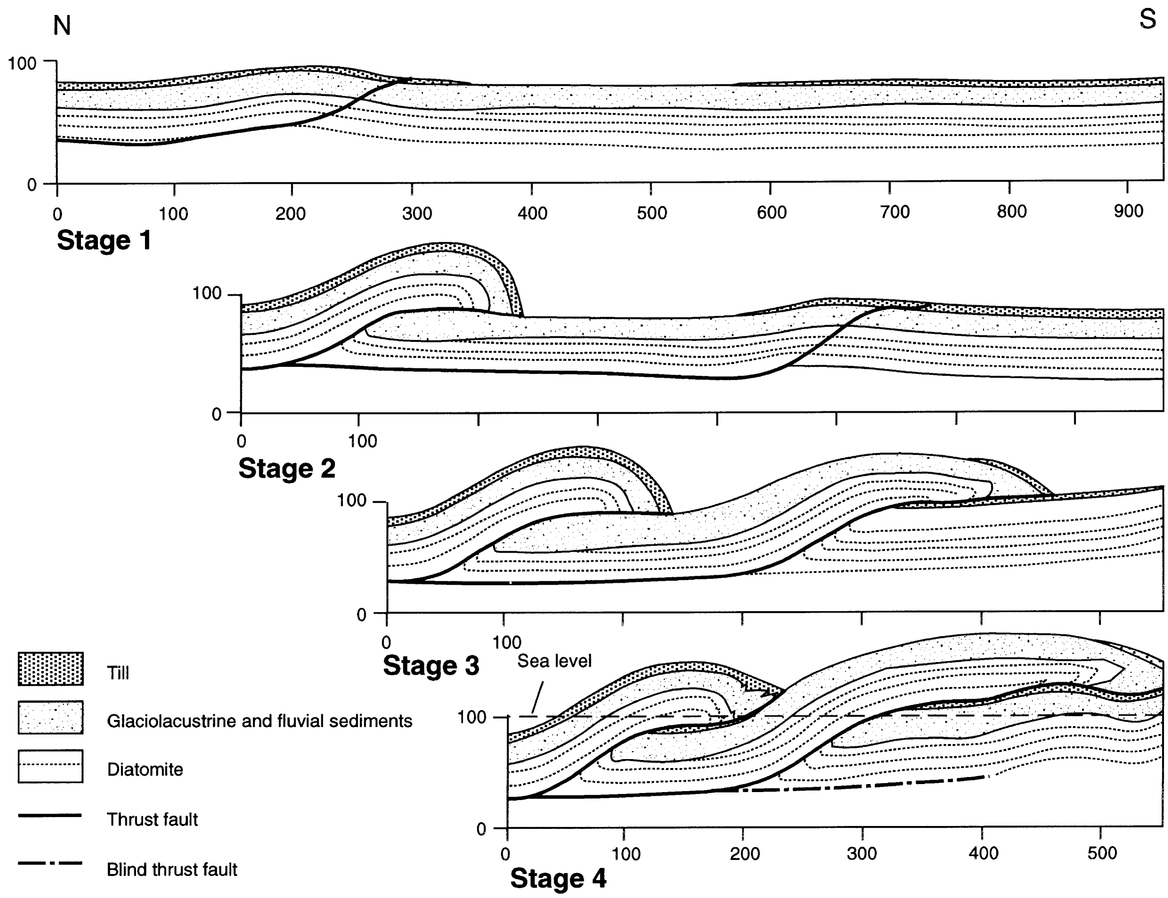


Fig. 16. The development of the Hanklit Thrust Fault Complex in Denmark (modified from Klint and Pedersen, 1995).

(Fig. 1B) is more common than the literature suggests. He argued that researchers have been heavily biased towards an imbricate thrust model following the work of Richter et al. (1951) on the push moraines in the Itterbeck–Uelsen area on the Dutch–German border (Fig. 15A), an influence he traces in both European work (e.g. Gripp, 1954; Woldstedt, 1954; Köster, 1958; Viete, 1960; Domoslawska-Baraniecka, 1961) and that from North American (e.g. Mackay, 1959; Kupsch, 1962; Mackay and Mathews, 1964; Bluemle, 1966; Moran, 1971; Clayton and Moran, 1974). He suggested that there is tendency to adopt an imbricate thrust model too readily and makes the point by re-plotting several structural sections from different push moraines, without vertical exaggeration (e.g. Grahle, 1960; Höfle and Lade, 1983; Wilke and Ehlers, 1983; Van Gijssel, 1987; Kluiving, 1994). In their original form these sections appear to be consistent with a fan of steeply dipping listric thrusts rising from a basal décollement, but in practice, when plotted at a true scale, they are more consistent with a series of low angle or sub-horizontal nappes, which are partially superimposed and onlap (Fig. 17). A push moraine architecture of sub-horizontal nappes may, therefore, be more common than is perhaps reflected in the literature (Van der Wateren, 1995a,b).

This emphasises the difficulty in making structural interpretations from limited borehole, or field exposures. The point is well illustrated in relation to

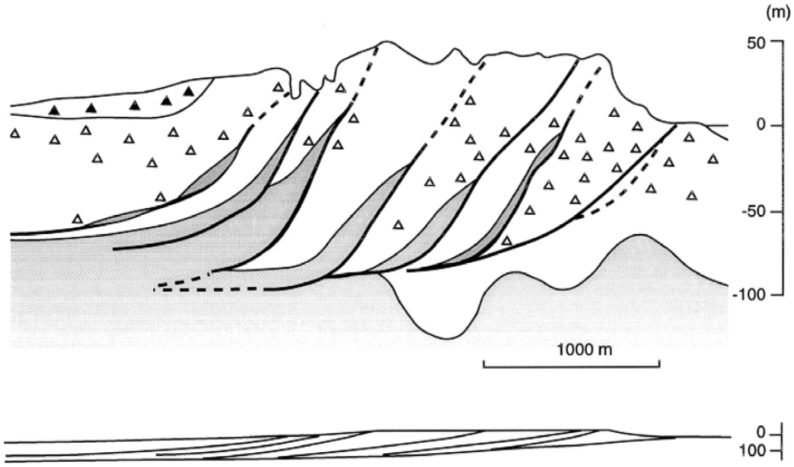


Fig. 17. Replotted cross-section through the Hamburg–Blankenese push moraine after Van der Wateren (1995a,b). The original (upper) cross-section is after Wilke and Ehlers (1983). This illustrates how the presentation of the data may bias its interpretation towards an imbricate model.

Dinas Dinlle in North Wales where a coastal exposure of deformed sediments has been subject to differing field interpretations based on the extrapolation of the visible structures (Hart, 1995b,c), and only with addition of subsurface geophysics was the true tectonic structure revealed and the ridge correctly interpreted as a form of push moraine (Harris et al., 1995, 1997).

6. Models of push moraine formation

It is apparent from the above review that push moraines have a wide range of morphologies and structural architectures and that they have been interpreted via a diverse set of tectonic models. It is possible, however, to recognise some form of continuum, from single ridges composed of rooted folds and small-scale imbricate fans of listric thrusts, to multi-crested complexes composed of de-rooted folds and nappes organised in either horizontal or imbricate stacks. The push moraines described by Krüger (1994) at Höf∂abrekkujökull represent one end of the spectrum, while those of Holmstrømbreen (Boulton et al., 1999), Uvêrsbreen (Hambrey and Huddart, 1995), or those along the Rehburg Line in northern Europe (Van der Wateren, 1995a), represent the other. The key difference is the size, or aspect ratio (thickness to width), of the slab of foreland which is deformed to create the push moraine and the extent to which the structures become de-rooted and travel along the décollement surface beneath the deforming slab. This is a function of several factors, including: (1) the magnitude of the applied glacial stress, the rate at which it is applied, and the efficiency with which it is coupled to the foreland; (2) the occurrence and properties of the décollement horizon; and (3) the strength and rheology of the foreland above the décollement.

6.1. Glacial stress field

The magnitude and origin of the applied glacial stress and the efficiency with which it is coupled to the foreland is important in push moraine evolution. In theory the greater the applied stress the greater the scale of deformation should be; large advances should produce big push moraines. In practice, this is complicated by the properties of the foreland which control the degree to which the glacial stress field is coupled to the ice margin and the susceptibility of the foreland to deformation. The general principles were, however, illustrated by a series of numerical simulations undertaken by Van der Wateren (1985, 1995a). He modelled the evolution of push moraines based on the principle that they are analogous to the behaviour of accretionary prisms at subduction zones. These push moraine models start from the premise that the sediment within them is subject to pervasive shear and consequently sediment viscosity and gravitational body-forces within the growing moraine (i.e. it begins to deform under its own weight) become key variables. One of the conclusions from these formulations is that for materials of the same density and viscosity, the speed of the glacier advance (rate of deformation) determines the thickness and gradient of the push moraine, with rapidly advancing glaciers producing thicker, steeper push moraines. Although these results are of interest the presence of pervasive shear within push moraines has been questioned by Boulton et al. (1999).

For deformation to occur the glacial stress field has to exceed the strength of the sediments or rock within the foreland along the plane of décollement, allowing the foreland as a whole to deform. The following are four main ways in which this stress can be generated (Fig. 18):

(1) Push-from-the-rear. Here, deformation is caused by the forward advance of the glacier into a sediment pile that causes its lateral compression. The critical variables in this model are the transmission

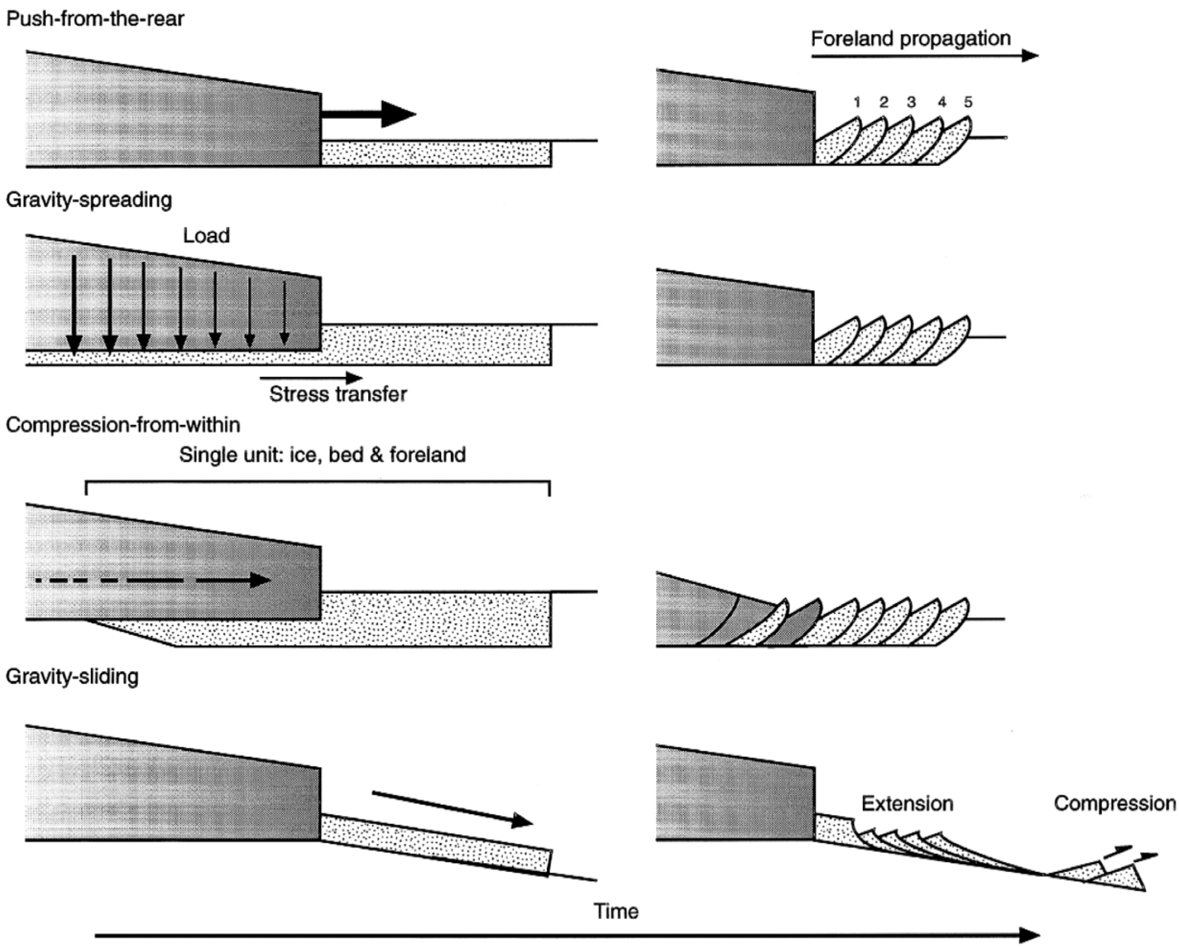


Fig. 18. Models of applied glacial stress.

of stress through the foreland and the coupling of the foreland to the ice margin.

(2) Gravity-spreading. This is a result of gravitational forces, associated with lateral pressure gradients induced by vertical loading (Aber et al., 1989; Van der Wateren, 1985, 1995a). Glaciers impart a vertical load, due to their weight, which decreases as the glacier thins towards an ice margin. A component of this normal load is, therefore, directed laterally from areas of maximum ice thickness towards the margin. If this stress field exceeds the strength of the weakest layer within the foreland, it will deform and collapse to give a décollement surface that rises to the surface to define an initial thrust block or nappe. Displacement of the first block causes it to rise either vertically along a steep listric thrust, or to overthrust the foreland. At some point, the applied stress is more easily accommodated horizontally, by further collapse of the weak layer in the foreland, than by elevating the thrust block. As a consequence, the basal décollement surface is extended horizontally, and rises to the surface to define a new thrust block, this block is displaced beneath the combined load of the ice and the first thrust block. Lateral extension of the push moraine will continue until the combined load of the ice and push moraine no longer exceeds the strength of weakest plane within the foreland. One of the key structural features with which to recognise the products of gravity-spreading are the presence of load structures, such as boudins and, in particular, mud diapirs (e.g. Pedersen, 1987, 1996; Klint and Pedersen, 1995; Sadolin et al., 1997). It is worth noting that many researchers consider gravity-spreading to be the principal glaciotectonic force involved in glaciotectonic deformation (Rotnicki, 1976; Van der Wateren, 1985; Aber et al., 1989; Benn and Evans, 1998).

(3) Push-from-within. In this model, compression occurs within the terminal zone of the glacier due to deceleration of ice flow, which may occur: across a thermal boundary, due to a change in substrate, due to the passage of a surge-front, or simply due to the presence of a reverse bedrock slope. The glacier, the subglacial zone and proglacial zone are strongly coupled and act as a single unit that is deformed by a series of listric thrusts rising from a basal décollement which lies beneath both the glacier and its foreland (e.g. Hambrey and Huddart, 1995). Where the ice margin is cold-based, then the coupling is a thermal one. However, where the glacier bed is unfrozen, then the nature of the coupling is less apparent, but may reflect the presence of a basal décollement surface that is able to accommodate movement more easily than other potential décollements, such as sliding or sediment deformation at the ice-bed interface.

(4) Gravity-sliding or -gliding. Here deformation is caused by the down-slope movement of thin sediment packages under their own weight. Gravity-sliding was once considered to be an important mechanism in some orogenic belts (e.g. Elter and Trevison, 1973), but in the context of push moraines its relevance is uncertain. For this process to operate in push moraine formation, one first requires the creation of some form of proglacial bulge in front of the glacier, providing a surface slope away from the glacier, down which sediment can move. As a primary mechanism for the formation of a push moraine this is unlikely, but it may operate on an established push moraine as it continues to grow. Pedersen (1987) provides criteria to distinguish between the products of gravity-sliding from those of gravity-spreading.

In practice, a combination of all four models may be applicable; a fact which may help to explain the range of forms observed. It is worth drawing attention, in this context, to the push moraine complex on Coraholmen in Svalbard (Boulton et al., 1996), which formed both by gravity-spreading and unusually by gravity-tectonics. Sediment was advected by subglacial deformation to the ice margin during the surge and subsequently extruded (gravity-spreading) from beneath the glacier as it stagnated post-surge. This sediment appears to have subsequently flowed into the foreland under it own weight and provides one of the few good examples of gravity-tectonics in push moraine formation (Boulton et al., 1996).

The rate at which the glacial stress field is applied to the foreland is also critical. It determines the rate at which pore-water pressures are elevated within foreland sediments and consequently the relative importance of ductile to brittle deformation. If the stress field is applied quickly, by a rapid ice advance for example, then brittle deformation is more likely, while ductile deformation is favoured by slower rates of deformation. The rate of ice advance, controlled by ice velocity relative to ice-marginal ablation, becomes critical therefore (Van der Wateren, 1995a).

Not only is the magnitude and rate of change in the applied glacial stress important, but the way in which it is coupled to the glacial foreland is also critical. Reverse bedrock slopes (Bennett et al., 2000), or the accumulation of sediment against a stable margin prior to pushing (e.g. Humlum, 1985; Boulton, 1986), are all considered necessary for the effective transfer of stress from the glacier to the proglacial zone in order to facilitate the formation of a push moraine. Boulton (1986) argued that the formation of an outwash fan is an important pre-requisite for push moraine formation and therefore some form of glacial still-stand is necessary prior to a glacial advance for push moraine formation. It follows from this that push moraines must, therefore, mark a stable and perhaps climatically important ice-marginal position (Van der Wateren, 1995a). The Holmstrømbreen push moraine illustrates this point, with the glacier surging into an outwash fan to create the moraine (Boulton et al., 1999), an association which has been widely documented (e.g. Thomas, 1984a; Humlum, 1985; Eybergen, 1986; Van Gijssel, 1987; Croot, 1987). It is also worth noting that deformation of ice-marginal landforms, such as grounding-line fans (e.g. Ashley et al., 1991) and deltas (Thomas, 1984b), is also well documented in subaqueous environments.

Although Van der Wateren (1994, 1995a) stresses the importance of pre-, syn- and post-tectonic deposition of outwash during push moraine evolution, he argued that outwash fans are neither an integral part or pre-requisite for push moraine formation. Instead he suggested that ideal conditions for the formation of a push moraine are found along advancing ice margins at which ice flow is decelerating either due to a reduction in basal sliding or in the rate of subglacial deformation. Van der Wateren (1995a) noted that ice advanced towards the Rehburg line over fine-grained sediments, which would have facilitated subglacial deformation. However, close to the former ice margin the substrate consisted of coarser-grained fluvial deposits, which would have precluded subglacial deformation, reducing ice velocity, and therefore causing the ice margin to thicken. He suggested that the build-up of potential energy in this way would have been sufficient to couple the glacial stress field to the foreland without recourse to an outwash fan. With this type of coupling push moraines need not represent climatically significant or synchronous ice margins.

6.2. The décollement horizon

The presence and depth of a ductile horizon along which décollement can occur is essential for large-scale foreland deformation and in addition determines the aspect ratio of the slab of foreland which is deformed. The occurrence and distribution of such a ductile horizon, a function of foreland stratigraphy, has the potential to control, therefore, the distribution of push moraines along an ice margin. This is illustrated by Berg and Beets (1987) who found a good correlation between the distribution of push moraines and the presence of a ductile sub-horizon in the glacial foreland stratigraphy in the Netherlands. Similarly, Van der Wateren (1995a) shows how the distribution of push moraines along the Rehburg Line, in Germany (Fig. 15), is also subject to substratum control, with the occurrence of push moraines closely following a zone where Mesozoic and Tertiary clays come close to the surface. Similar conditions are associated with the push moraines along the Missouri Coteau in Canada (e.g. Kupsch, 1962; Aber, 1988). The correlation is not always perfect, for example at some locations on the Rehburg Line push moraines are absent despite favourable geology (Van der Wateren, 1995a). This suggests that the glacial stress field was not always sufficient to cause deformation, perhaps due to the presence of a deforming layer precluding the coupling of the glacial stress field to the foreland (Van der Wateren, 1995a). This type of observation emphasises the importance in the formation of large push moraines of the occurrence of a weak, ductile or brittle, layer within the foreland stratigraphy. The occurrence of fine-grained sediments, with a high water content and low permeability is particularly important. The facies present within the foreland and their spatial variability is therefore an important consideration. The formation of a décollement horizon, particularly in less favourable lithofacies, is strongly influenced by the hydrogeology of the foreland. The presence of pressurised groundwater or the occurrence of confined aquifers is particular critical. For example, a saturated sediment layer confined vertically by two aquicludes would form an ideal horizon for décollement. Rapid loading of such a layer, elevating pore-water pressures, would favour its deformation and collapse to form a décollement plane. Boulton et al.'s (1999) interpretation of the Holmstrømbreen push moraine illustrates the importance of the broader hydrogeological setting in elevating groundwater pressures and thereby facilitating the formation of a décollement surface (see also Boulton and Caban, 1995).

The hydrogeology of a glacial foreland also determines the likely fluid pressures and therefore friction along a basal décollement once it has formed. This is important in determining the mobility of structures along a décollement, with low levels of friction favouring the transport of nappes with little internal deformation, while high level of friction favours more rooted folds. Boulton et al. (1999) have argued that fluid pressure along the basal décollement may help define a continuum of structural forms within large, multi-crested, push moraines. This continuum can be amplified such that it consists of the following:

(1) Thrust-dominated moraines, which result from very low or friction-free sliding along the basal décollement and may show very great proximal–distal width. These moraines may have either a classic imbricate thrust architecture, or an internal structure more consistent with overthrust nappes (e.g. Thompson Glacier, Kalin, 1971; Uvêrsbreen, Hambrey and Huddart, 1995; Dammer Berge, Van der Wateren, 1995a).

(2) Fold-thrust-dominated moraines, which result from greater friction along the décollement plane, but may also show considerable proximal–distal widths. These moraine may display a more imbricate structure (e.g. Eyjabakkajökull, Croot, 1987; Holmstrømbreen, Boulton et al., 1999).

(3) Fold-dominated moraines, which occur when internal and basal friction is high. The width of the deformation is a function of whether or not the folds become attenuated and detached from the root zone. If they do become derooted, along a décollement surface, the proximal–distal widths will be greater than if they do not (e.g. Bride Moraine, Thomas, 1984a; Maktak Glacier, Boulton et al., 1976).

The boundaries between these three categories are somewhat arbitrary and open to interpretation, but this threefold division does serve to emphasise the potential role of fluid pressure along the décollement surface and in general the importance of the hydrogeological setting for push moraine formation. It is worth noting that other variables may also control this continuum, such as the rate at which deformation occurs, with rapid loading and compression favouring more brittle styles of deformation.

The importance of the hydrogeology of a push moraine is a recurrent, if poorly understood, theme in the literature. For example, there are frequent references to the presence of dry valleys intersecting and radiating from thrust surfaces within push moraine complexes, formed by the drainage of water from thrusts during deformation (e.g. Croot, 1987; Kalin, 1971). Equally, Kalin (1971) provides tentative evidence of seasonality in the rate of deformation of the Thompson Glacier push moraine caused by seasonal changes in ablation, and therefore in water supply. As water supply increases, relative to the rate of groundwater discharge through the moraine, water pressures will rise, thereby facilitating increased rates of slip along thrusts. In addition, the main theme which has emerged from the literature on thrust-blocks in North America is their association with confined aquifers (e.g. Bluemle and Clayton, 1984). Groundwater flow beneath, and in front of, an ice sheet has recently been modelled by Boulton et al. (1995). They have suggested that mid-latitude Pleistocene ice sheets re-organised groundwater flow in lowland regions in response to the influx of subglacial meltwater and to variations water potential caused by ice thickness variations. Boulton and Caban (1995) use these groundwater flow models, in conjunction with a model of ice sheet loading, to calculate the principal effective stresses in the subsurface beneath and in front of an ice sheet. This work suggests that overpressured groundwater may occur where permafrost is present in the glacial foreland. If this permafrost is discontinuous, narrow plates of permafrost at the ice margin may be associated with a significant head of groundwater, sufficient to facilitate décollement and movement of plates of permafrozen sediment to form push moraines.

The hydrogeology of the foreland, and in particular of the horizon in which décollement occurs, is critical therefore not only to the initiation of deformation along décollement surfaces, but also to the tectonic style of deformation which occurs. A wide range of site specific variable control this hydrogeology, and therefore push moraine formation and character, some of the key variables are listed below.

• The rate of subglacial melting and water supply to the groundwater system, which is a function of rainfall, ice velocity and basal temperature.

• The glacially induced water-potentials which drive groundwater flow, relative to gravitational potential. The scale and thickness of the glacier relative to the topography of the groundwater basin are the key elements here.

• The distribution and depth of permafrost in the foreland. The presence of thin-slabs of permafrost may confine groundwater at the ice margin and thereby elevate groundwater pressures beneath them.

• The geometry and permeability of the main hydrogeological units present. Not only the hydrogeological characteristics of the lithified-substrate present are relevant here, but also the unlithified sediments of the foreland and their spatial variability.

6.3. Strength and rheology of the foreland

The extent to which stress can be transmitted beyond the glacier margin, into the glacial foreland, is clearly important in determining the scale of deformation. This is controlled by the rigidity of the foreland relative to shear strength of the décollement horizon and the level of friction along it. If the foreland, as a whole, is too rigid then the ice will simply deform against it. However, if the glacial stress field is sufficient to overcome the shear strength of a décollement horizon present within the foreland, then in general, the stronger the foreland slab, above the décollement horizon, the further deformation will extend into the foreland. The width of foreland which is deformed for a given foreland strength is also controlled by the level of friction along the basal décollement. The greater the friction, the greater the strength the foreland must have to extend far beyond the glacier and vice versa. Consequently, two glaciers applying the same compressive force, but to two different forelands may result in push moraines of different sizes. The importance of the décollement horizon in the foreland has been discussed in the previous section, here we focus on the broader properties of the foreland and in particular its strength and rheology above any décollement horizon.

One of the most disputed aspects is the role of permafrost in push moraine formation. Observations of thrust dominated push moraines both in the Arctic and around the margins of the Laurentide and European ice sheets have frequently emphasised the importance of proglacial thrusting of ice-rich permafrost (e.g. Gry, 1940; Richter et al., 1951; de Jong, 1967; Rutten, 1969; Mathews and Mackay, 1960; Kupsch, 1962; Kaye, 1964; Kalin, 1971; Clayton and Moran, 1974; Andrews, 1980; Moran et al., 1980; Kaplyanskaya and Tarnogradskiy, 1986; Astakhov et al., 1996). Permafrost has also been invoked in push moraine formation within Alpine regions (Haeberli, 1979). Permafrost is considered in this work to be crucial in order to impart sufficient strength to the glacial foreland, thereby facilitating the transmission of stress from the glacier, pushing at the rear, through the thin thrust-blocks or nappes of frozen sediment. Some workers have suggested that basal décollement is located at the base of the permafrost and therefore the depth of décollement provides an indication of the permafrost thickness and consequently climatic conditions within the glacial foreland at the time of deformation (see Richter et al., 1951; Andrews, 1980; Boulton et al., 1999). Alternatively, others have suggested that décollement may occur within ice-rich, or fine-grained, horizons in permafrost since both tend to have lower shear strengths than ice-poor and coarse-grained horizons (e.g. Mathews and Mackay, 1960; Astakhov et al., 1996; Etzelmüller et al., 1996).

Etzelmüller et al. (1996) make a useful contribution to this body of literature by emphasising the variability in the mechanical properties of permafrozen sediment. They argue that the formation of large push moraines in Svalbard is essentially controlled by the mechanical properties of the permafrost. If the permafrost is too strong, push moraine formation is precluded. The mechanical properties of permafrost are a function of such variables as ground temperature, grain-size, and groundwater chemistry, which all determine the amount of unfrozen water content of the permafrost (Mathews and Mackay, 1960; Andersland and Alnouri, 1970; Williams and Smith, 1989). Fine-grained sediments, with a saline water chemistry, and temperatures close to freezing give permafrost with a higher water content and are consequently weaker than coarse-grained, cold, permafrost which is likely to retain little water at sub-freezing temperatures. Tsytovich (1975) distinguishes between weak permafrost, which is plastically deformable, and hard permafrost, which is not. Etzelmüller et al. (1996) reviewed the distribution of push moraines in Svalbard (Fig. 15B), and suggested that they are largely confined to locations below the Holocene marine limit and therefore to weaker permafrost due to the presence of significant amounts of unfrozen saline pore-water. Although there are several notable exceptions to this pattern (e.g. Lønne and Lauritsen, 1996), this work does emphasise that permafrost may not only provide foreland strength, but that variation in its mechanical properties may have a significant impact on the rheology of the foreland and therefore influence the style of deformation. It is possible to suggest that where permafrost is very rigid, no push moraine will form, but as the strength falls thrust dominated moraines in which the thrust blocks show little internal deformation may occur, such as those found in Arctic Canada (e.g. Thompson Glacier, Kalin, 1971), and as the strength falls further more fold-thrust-dominated examples may form (e.g. Usherbreen, Hagen, 1987; Etzelmüller et al., 1996).

Not all researchers accept the need for permafrost in push moraine formation (e.g. Aber, 1988; Van der Wateren, 1995a,b; Schlüchter et al., 1999). The importance of permafrost was originally questioned by the observation that many thrust bedrock landforms were associated with the presence of confined aquifers (e.g. Mackay and Mathews, 1964; Moran, 1971; Rotnicki, 1976; Bluemle and Clayton, 1984; Aber, 1985). Rapid loading of these aquifers may decouple sediment and/or rock along extensive décollement surfaces and when combined with compressive ice flow could induce thrusting of thin nappes without need to strengthen them with permafrost (Bluemle and Clayton, 1984). These interpretations tend to stress the importance of gravity-loading and consequently nappes tend to be more confined, both by the ice and the growing moraine, again reducing the requirement for foreland strength. Van der Wateren (1981, 1985, 1994, 1995a,b) has produced strong evidence to support the idea that permafrost is unnecessary for push moraine formation, pointing out some of the push moraines of the Rehburg Line formed by glacial advances into ice-marginal lakes in which permafrost is unlikely. This argument is also supported by occurrence of similar structures in non-glacial environments such as the Mississippi mud lumps where permafrost is definitely not present (Morgan, 1961; Morgan et al., 1968; Aber, 1988). The Mississippi mud lumps are structures similar in scale to some push moraines and are formed by gravity-loading due to delta progradation. In this analogue near frictionless sliding is possible due to the saturated nature of the sediments and they emphasise the importance of the hydrogeological setting.

From the range of observations available, it would appear, therefore, that permafrost is an essential pre-requisite for some types of push moraine, but not necessarily for all. Recognising which moraines are the product of glacier-permafrost interaction is not likely to be easy. In some cases there may be sedimentary evidence in the form of pre- or syn-tectonic ice wedges, such as in the Bride Moraine (Thomas, 1984a), but such evidence is not always present. A more realistic approach is to focus on the hydrogeology of the décollement horizon, and its rheology relative to the overlying sediments. If one cannot generate sufficient fluid water pressures along the décollement for near frictionless sliding, then it is likely that the nappes within large push moraines (aspect ratios $> 1:50$) were frozen.

Irrespective of the presence, or absence, of permafrost the rheology of the sediments and/or rock within the foreland, and its vertical and horizontal variability, will influence the style of deformation. This is a function of grain-size and facies variability within the foreland (Hart and Watts, 1997). Thick competent units tend to give long wavelength folds, while thin incompetent layers give folds of shorter wavelength and amplitude. Van der Wateren (1995a) makes a similar point arguing that the coarse-grained sediments of the Dammer Berge push moraines favour the formation of overthrust nappes, while the finer-grained, more ductile sediments at Holmstrømbreen favour folding and the formation of an imbricate stacks of folded nappes. The fluid pressure within the sediments and the overburden pressure from the evolving push moraine may also influence the style of deformation. It is worth noting that the modelling undertaken by Van der Wateren (1995a) suggests that, for a given rate of advance, thinner and flatter push moraines result in materials of lower viscosity.

Another characteristic of the foreland which is frequently considered in the formation of push moraines is the general availability of rock or sediment suitable for deformation. Implicit in this idea is that the widespread availability of debris or deformable sediment within a glacial foreland will favour push moraine formation. For example, Etzelmüller et al. (1996) note that push moraines in Svalbard are closely correlated with the outcrop of sedimentary rocks (Fig. 15B), which produce abundant debris via frost weathering and glacial erosion. If there is no debris or sediment to be deformed push moraines cannot form unless deformation of in situ bedrock occurs. The importance of the susceptibility of the foreland to deformation is also illustrated by Bennett et al. (1999b) in Reindalen (Svalbard). Here

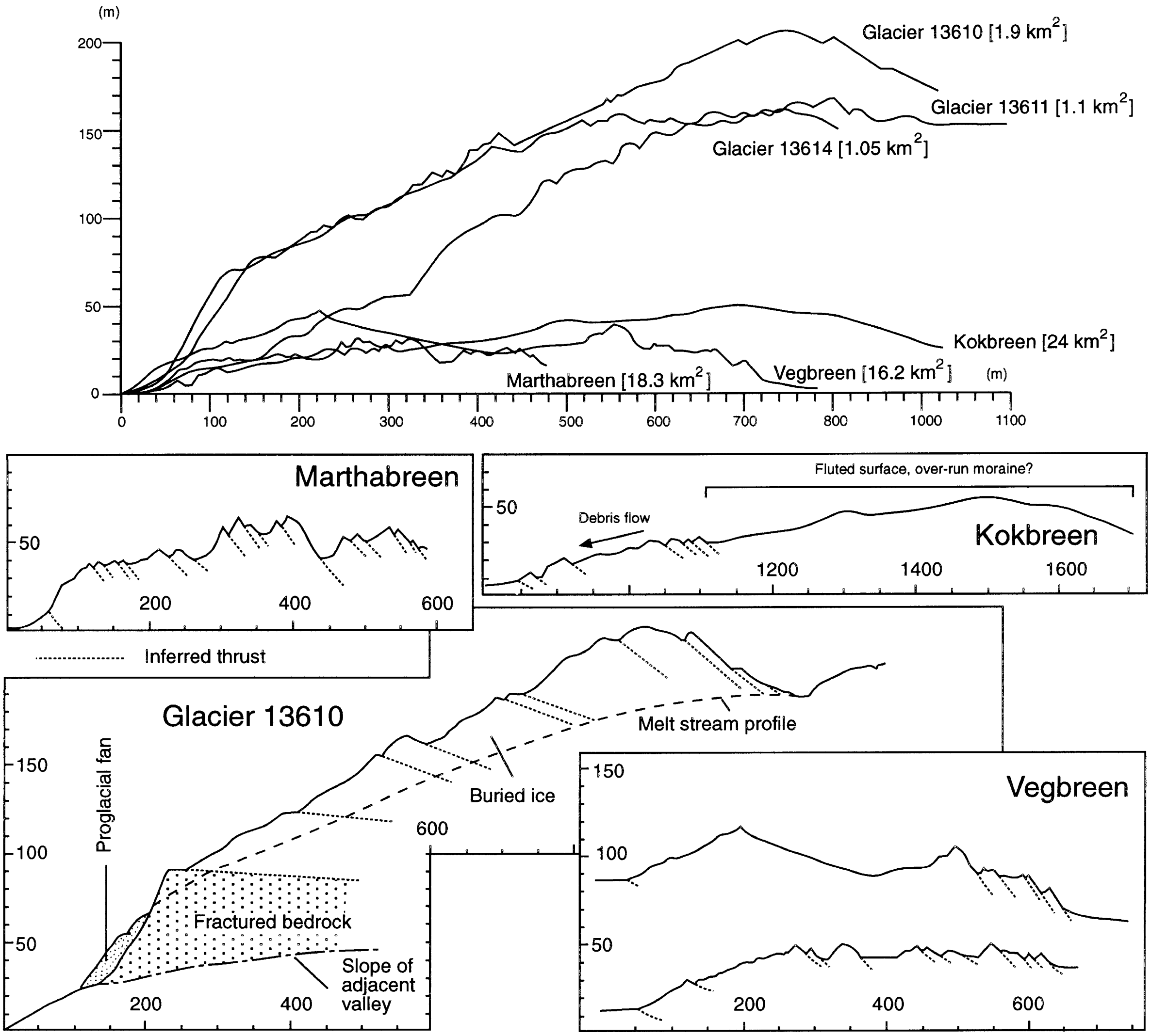


Fig. 19. Cross-sections, along a central flow line, through a selection of push moraines in Reindalen, Svalbard. Note that glacier area and size do not equate to the largest push moraines. In particular three small cirque glaciers are associated with three very large push moraines, which contain fractured bedrock rafts. Based on data in Bennett et al. (1999b).

a series of exceptionally large push moraines form in front of very small cirque glaciers, while the moraines in front of adjacent, but much larger glaciers, are comparatively small (Fig. 19). These exceptionally large cirque moraines contain significant amounts of displaced bedrock and Bennett et al. (1999b) argued that these small cirque glaciers were able to exploit specific conditions in the foreland. In particular, the outcrop of well fractured, sub-horizontal sedimentary strata, with fine-grained lithologies at the base of the rock succession, and an absence of lateral support which allowed the glaciers to push rock units out of the slope. This example serves to illustrate that the properties of the foreland have a critical role in determining the scale of push moraines that results from a given set of glaciological conditions.

7. Conclusions

It is clear from the above review that the size, morphology, structural architecture and evolution of a push moraine, reflect the subtle interplay of both specific glaciological conditions and the occurrence of conditions favourable for deformation in the glacial foreland. A range of morphologies and structural architectures exists from small single single- or multi-crested push moraines composed of imbricate fans of listric thrusts, to multi-crested complexes composed of de-rooted folds and nappes organised in either horizontal or imbricate stacks. Fig. 20 attempts to depict this range of push moraines within a matrix defined by some of the variables that control their formation as discussed in the previous sections. This model scopes the variables that are relevant in understanding the morphology, structural architecture and therefore significance of push moraines. The key variables are the presence of a décollement layer, its rheology, permeability, porosity, and hydrogeology. Further data about these variables are required to help constrain the detailed mechanism involved in the evolution of push moraines and to perhaps quantify the conceptual model presented in Fig. 20. It is the properties of the foreland that seem to dominate in the formation of push moraines rather than those of the glacier. The relative importance of glaciological versus foreland characteristics in push moraine formation is clearly difficult to determine, but critical to the significance of Pleistocene push moraines in palaeoglaciological investigations. One is left, in conclusion, with the question: what palaeoglaciological and environmental inferences can be made from the occurrence of push moraines in the geological record, if any?

In the context of small, seasonal, push moraines the potential is considerable, although limited by the poor preservation potential of these small features. They form in front of a specific type of glacier—warm-based and active—and are consequently indicative of these conditions (Boulton, 1986). More importantly, where an annual sequence of moraines can be established, their spacing may provide a proxy record of summer ablation and therefore of such variables as summer air temperature, or the number of frost-free days (e.g. Timmis, 1986). In addition, small push moraines, of this sort, provide information about former ice-marginal geometry (e.g. Bennett and Boulton, 1993), and crevasse patterns (Matthews et al., 1979; Horsefield, 1983). Again if one assumes, or can establish, an annual spacing, then such moraines provide data on the rate of deglaciation (e.g. Larsen et al., 1991). In practice, all these applications are largely restricted to recent landform assemblages of contemporary glaciers, due to the poor preservation potential of these small push moraines.

The significance of larger push moraines, in palaeoglaciology, is much more difficult to establish. Large, single-crested, push moraines may still provide information about the geometry of former ice margins, but the geometry of multi-crested, push moraines may bear little relationship to former ice margins, particularly where they have been overridden (e.g. Bennett et al., 2000; Benn and Clapperton, 2000). If the development of outwash fans is a pre-requisite for push moraine formation (Boulton, 1986; Boulton et al., 1999), then they are likely to mark a climatically significant and synchronous ice margin. If as argued by Van der Wateren (1995a) they form anywhere where the glacial stress field is effectively coupled to the foreland, by for example a change in subglacial substrate, then they need not reflect a line of still-stand and consequently may be diachronous. One of the implicit assumptions frequently made is that large, multi-crested, push

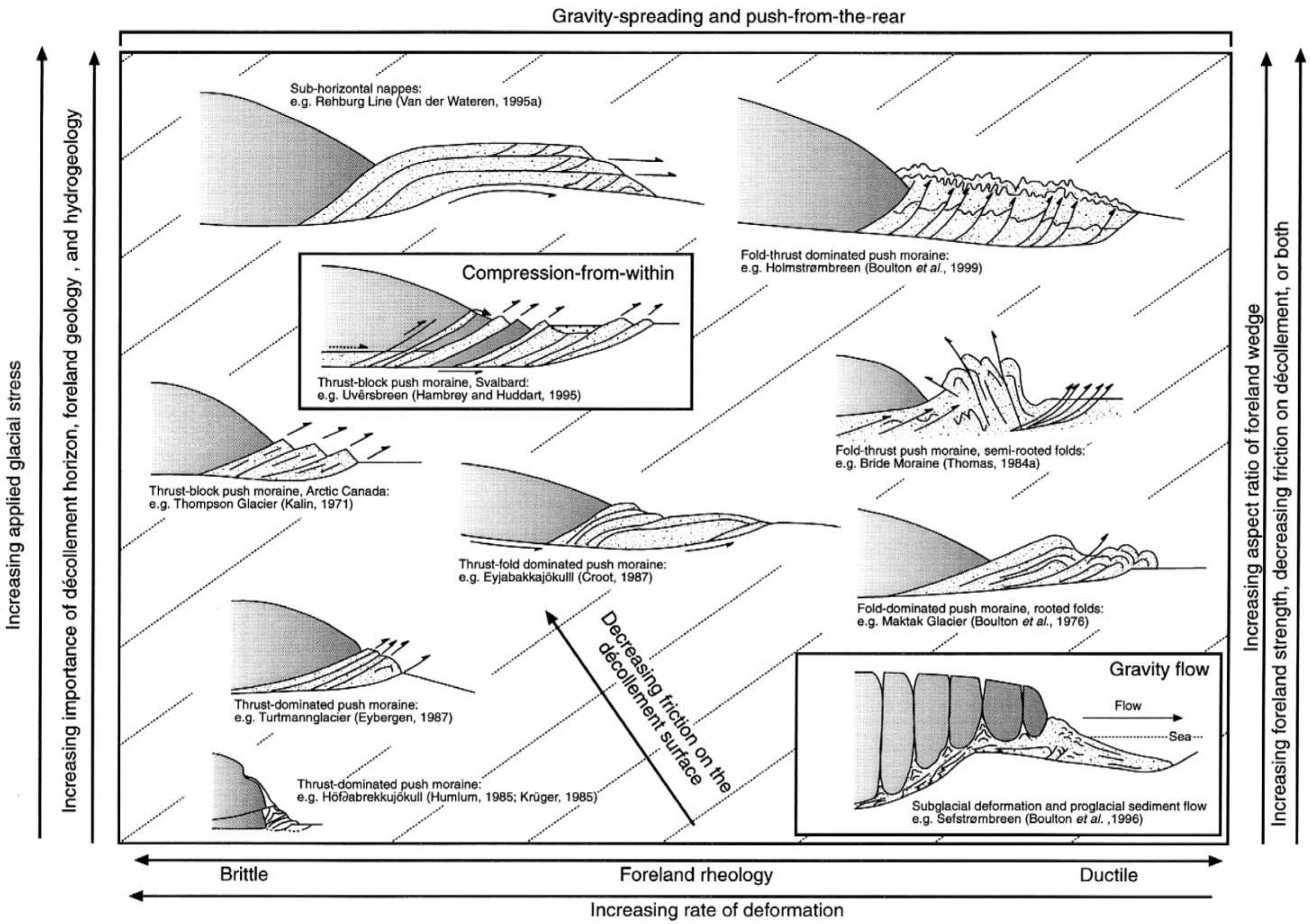


Fig. 20. Schematic model showing how some push moraines may relate to selected variables used to define a broad matrix.

moraines tend to form in front of surge-type glaciers. This idea was advanced by Croot (1988) in relation to large push moraines found in Svalbard, which he used as evidence for the number of surge-type glaciers in Svalbard. On this basis, he suggested that 90% of glaciers were surge-type, a figure reduced to 35% by Hamilton and Dowdeswell (1996) who used a range of other, more reliable, criteria. Hambrey and Huddart (1995) also demonstrate how large multi-crested push moraines need not form in front of surge-type glaciers in Svalbard.

Perhaps the most significant inference from large multi-crested, push moraines is their potential relationship to permafrost, which as reviewed above, is both ambiguous and contentious. The traditional view that permafrost is essential to large-scale proglacial deformation has led to the idea that push moraines may be indicative of a permafrozen foreland and that the depth of décollement may provide palaeoenvironmentally significant data on the depth of permafrost (e.g. Richter et al., 1951; Boulton et al., 1999). The problem is that there is no consensus about this point. What is emerging, however, is an emphasis on the importance of the presence, rheology, and hydrogeology of the décollement layer. In order to explain the deformation of thin slabs of foreland, of considerable areal extent, very low levels of friction along underlying décollements are required. This may simply be a function of the lithofacies present within a foreland, for example the occurrence of thin, saturated, layers of fine-grained sediment, or the presence of a confined aquifer, which may both favour low friction décollement. However, where such conditions do not occur, pressurised groundwater within a décollement layer may be required to explain the scale of deformation observed. One of the most plausible methods of elevating groundwater pressure is through the presence of permafrozen sediment within the foreland (Boulton and Caban, 1995). In addition, as levels of friction along a décollement increase, the need to strengthen the foreland with permafrost, to facilitate stress transmission, increases. It is consequently impossible to make generalisation about the significance of permafrost in push moraine formation. One approach, however, is to consider at each site, how friction along a décollement layer was lowered sufficiently to allow the scale of movement observed, and whether this can be achieved without recourse to permafrost, before its presence is assumed and any inferences made on this basis. Important in this context, therefore, is an understanding of the hydrogeology of the glacial foreland and of the rheology of décollement layer. If the hydrogeology of the glacial foreland is important in push moraine formation, then it follows that relict push moraines may contain hydrogeological information about conditions during their formation. If the glacier is the principal factor controlling the hydrogeology of the foreland, then push moraines may contain valuable data on the groundwater and melt regimes beneath former ice sheets; if, however, the hydrogeology is determined more by the properties of the foreland, then the glaciological significance of push moraine in the geological record may be limited. Further research is required to understand more clearly the mechanism of décollement and in particular the role of pore-water pressure and its link to the broader hydrogeological regime of a glacial foreland. To date, these issues have only been addressed through theoretical modelling (Boulton et al., 1995; Boulton and Caban, 1995). These experiments need to be constrained by field data on the hydrogeology of glacial forelands and in particular within actively forming push moraines.

To answer the question posed at the start of this section, small, seasonal, push moraines are of some value in reconstructing the palaeoglaciology of recently deglaciated areas, but their utility is ultimately limited by their poor preservation potential. Larger push moraines have considerable potential for palaeoglaciology, since their formation is a consequence of either the occurrence of specific glaciological conditions, or the occurrence of particular conditions in the glacial foreland. However, further research is required to clarify the key variables involved and in particular the relative importance of glaciological versus foreland properties in controlling the formation and structural evolution of push moraines before their true potential can be realised.

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