The sedimentary and structural evolution of a recent push moraine complex: Holmstrømbreen, Spitsbergen

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Abstract

The glacier Holmstrømbreen, in Spitsbergen, surged into the ice contact scarp of a proglacial outwash sequence at some time during its Neoglacial maximum. The outwash sediments were pushed along a decollement to produce a moraine in which deformation extended for 1.5 km beyond the furthest extent of the glacier front. The style of folding and faulting and the nature of the pre-, syn- and post-tectonic sedimentary sequence across the whole push moraine is described from a continuously exposed section of the push moraine which extends from its proximal to distal extremities. The precise extent of incremental compressive shortening of the pushed sediments, of some 900 m, is established. The depth to the underlying decollement is inferred to be an average of about 30 m, indicating that stresses and movement were transmitted through a thin nappe with an aspect ratio of about 1 in 30. It is suggested that this nappe was frozen and that an artesian water pressure head of 60 m immediately beneath it reduced friction along its base to a very low value. It is calculated that a glacially generated force of about $1.5 \times 10^7$ kN was responsible for pushing the sediment nappe. The nature of the glacially controlled groundwater flow system rather than the magnitude of longitudinal forces generated by the glacier is the principle determinant of large-scale push moraine characteristics.

The changing ice topography produced by the pushing event during the surge and in the post-surge decay had a major influence on the evolution of the meltwater drainage system and the style of fluvial sedimentation. The structure, sedimentary architecture and evolution of the whole glacial-tectonic/fluvial complex can best be understood by considering the impact of the surge on a complete outwash system.

It is suggested that the setting and processes which produced the Holmstrømbreen push moraine could account for many broad, multi-ridge and fold push moraines which formed proglacially, and that the thickness of original sediments above the decollement can be a guide to the thickness of contemporary permafrost. © 1999 Elsevier Science Ltd. All rights reserved.

1. Introduction

Major moraines with large-scale internal tectonic structures are widespread within the areas occupied by large ice sheets in the mid-latitudes of the Northern Hemisphere during the late Cenozoic. They are presumed to have formed when the ice sheet margin advanced, pushing up proglacial sediments to form a moraine. In the areas occupied by ice sheets during the last glacial period in Europe and North America, push moraines are concentrated near to the maximum extents of the ice sheets. In North America, this reflects a long period during which the ice sheet margin lay at or near to its maximum extent, whilst fluctuating strongly, possibly in a surging mode (Wright, 1973). In Europe, large push moraines appear to have formed during an early stage of slow net retreat from the maximum extent, during which small readvances took place.

The conditions under which push moraines form are relatively poorly understood and there are several key questions which need to be addressed:

- Does pushing take place most readily when sediments have accumulated against the glacier margin, so that
the advancing glacier can readily transmit a large force through a large sediment mass (Boulton, 1986)?

- How can the frictional resistance to movement of a large, thick sediment mass be overcome when the shear stress which an ice sheet can apply is limited by its own low shear strength (about 100 kPa)?

- How can stresses be transmitted horizontally through proglacial sediment plates of low aspect ratio which can be greater than kilometres in extent from their proximal ice contact front to their distal leading edge?

- How is the pattern of strain in the pushed mass related to glacial and environmental processes?

Although extensive sections in push moraines produced by the mid-latitude ice sheets can help to reconstruct patterns of strain, the most useful studies must be those on modern or recently produced push moraines, where sedimentary and tectonic style can be related to observable glacial regimes and conditions. Unfortunately, most large modern glaciers have undergone substantial retreat through most of the 20th century, apart from short periods of surging. However, the last century was a period of widespread glacier advance which, in many parts of the world produced large push moraines. There are many such glaciers in Spitsbergen which produced large push moraines marking their late Holocene maximum extent during the late 19th century. Many of them were investigated by Gripp (1929) in 1927, and found to have great mutual similarities. We chose to study the push moraine of Holmstrømbreen in central Spitsbergen (Fig. 1) in order to address the questions posed above, and because
the internal structure of the push moraine was particularly well exposed in two major stream gorges, which permitted us to relate the internal structure of the push moraine to the glacial and environmental regime.

In analysing the origin and setting of the Holmstrøm-breen push moraine we distinguish four principal zones (Figs. 2 and 3):

- The exposed glacier.
- The ice-cored moraine zone.
- The glacially pushed sediment zone.
- The proglacial outwash zone.

2. The exposed glacier

Looped patterns of isoclinically folded medial and lateral moraines on the surface strongly suggest that the glacier’s last advance to its recent terminus was achieved by a surge (Fig. 4). The glacier surface in the terminal zone lies well below the level of the ice-cored moraine (Fig. 2) and the downwasting of the glacier surface indicated by this reflects relative inactivity since that surge. We have no direct evidence of the date of the surge, but it clearly pre-dates Gripp’s observations.

We estimate the glacier surface in 1927 to have lain some 40 m below the crest of the ice-cored moraine, compared with at least 80 m at present. Comparison of our photographs with those of Gripp suggests that the crest of the ice-cored moraine was not significantly lower.
in 1984 than in 1927. We thus conclude a net downwasting of the exposed ice of some 40 m in 57 years. Gripp’s photographs clearly indicate however that the slope in the terminal zone has reduced since 1927 and that the glacier is probably still decaying. Gripp’s map is clearly not sufficiently accurate to be used to estimate whether medial moraines have been significantly displaced by glacier flow since 1927, although it does show moraine patterns very similar to those of the present day.

The major medial moraine loops on the glacier surface are overlain by between 1 and 2 m of bouldery diamicton which can be shown to have melted out from the underlying debris-rich ice. The englacial debris is largely composed of supraglacially-derived material incorporated into lateral debris septa after falling onto the glacier from valley walls in the accumulation area. Melting out of these septa in the ablation area has generated a supraglacial diamicton, added to by material which fell from flanking hillsides in the ablation area, and which never became englacial unless it fell down crevasses and moulins. (The term till is applied to this diamicton. It is widespread on the glacier surface, is relatively uniform in composition and cannot be confused with a sediment from any other source. In several places the till has flowed, either over ice or over other sediment, but is still regarded as till, or flow till, for the sake of simplicity.)

The sources of individual medial moraines in the terminal area can be estimated with reasonable confidence. The two large detached loops near the southern margin of the glacier (Fig. 4) were probably derived from tributary glaciers on the southern side of Holmstrømbreen, whose confluence with Holmstrømbreen is marked by ridges of supraglacial till (e.g. 14°70’E, 78°50’N) reflecting underlying debris septa. We presume that the Holmstrømbreen surge plucked away the terminal zones of these glaciers and their terminal debris septa, and incorporated them as detached lenses into the surged mass.

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Fig. 3. Map of the terminal zone of Holmstrømbreen made from the 1966 vertical aerial photograph. It distinguishes four major zones: the exposed glacier ice surface; the sediment covered glacier zone; the zone of pushed pre- and syn-tectonic fans; and the post-tectonic outwash fans. Note that the sand and gravel units in the pushed zone represent the locations of the major pre- and syn-tectonic fans. The lineations on these fans represent channel and bar features.

Fig. 4. Holmstrømbreen and its tributary glaciers drawn from the 1996 aerial photographs and showing the pattern of folding of lineations and medial moraines in the terminal zone of Holmstrømbreen. It is suggested that the major, detached moraine loops (eg. a, b/c, u, v, etc.) represent the terminal zones of tributary glaciers which were pulled into the flow of Holmstrømbreen at the time of the surge. The letters on these loops relate them to the tributaries from which they are suggested to have been derived.
Fig. 5. Meandering stream cut into the glacier surface showing deeply incised meander bends.

The medial moraines and the outermost ice-cored zone form topographic ridges and the depressions between them contain lakes. These are fed by typically complex drainage channels which are frequently incised deeply and intimately associated with medial moraines (Fig. 5). We suggest that this association, typical of many glaciers, results from the formation of frequent crevasses in the zones of relatively high shear strain rates which tend to coincide with medial moraines. These are then exploited by surface drainage to produce complex channel systems similar to those in limestone areas. Many large surface streams disappear into englacial tunnels.

The paths of many former streams can be reconstructed with accuracy. Where stream capture causes the abandonment of a channel, a frequent occurrence, melting down of the glacier surface may reveal the old channel and its sediment infill. The latter tends to inhibit ablation in the underlying ice, whilst flanking ice melts relatively rapidly. The end result is an esker, a ridge of stream sediment standing on an ice core high above the flanking ice. Many active streams lie immediately to one side of eskers, or even series of eskers which represent successive generations of the same drainage line (Fig. 6).

3. The ice-cored moraine zone

Ice cored moraine forms a ridge of complex and relatively high relief up to 2 km wide, about 105 m above sea level and 80 m above the outer part of the exposed glacier zone (Fig. 7). A seismic and resistivity survey, to be reported in detail in a later paper, shows the ridge to be underlain by a mass of dead ice about 200 m thick, resting on a subglacial surface which lies at about 100 m below sea level. It is overlain by a till stratum, generally about 1–2 m thick, which has melted out from the underlying ice (Fig. 8). Supraglacial fluvial and lacustrine sediments are also ubiquitous in this zone (cf. Boulton, 1972).

The distinction of an ice-cored moraine zone from the medial moraines of the exposed glacier is arbitrary, as it is clear that the ice-cored zone is primarily composed of medial and lateral debris septa, and intervening clean ice zones, which have been strongly compressed longitudinally (Fig. 4). Compression is reflected by lineations of non-fluvial origin in the ice-cored zone, which reflect local thickening of the till cover produced by melting out of zones of relatively high debris-content.

The supraglacial till is inherently unstable and undergoes a typical cycle of failure (often as arcuate slump scars – Fig. 8a; Boulton, 1972), remoulding and re-stabilisation followed by a further similar cycle. During this process the originally massive till may acquire more or less well defined sorted and bedded horizons (cf. Boulton, 1972; Boulton and Paul, 1976; Lawson, 1979; van der Meer et al., 1992).

The crest of the ice-cored zone lies near to the outer margin of the zone, giving it a long glacier-proximal slope and a short distal slope. Thus, much of the modern drainage on its surface is inwards, towards the exposed glacier and the lakes which lie on it (Figs. 2 and 3). As late as 1966, the date of the aerial photographs from which Fig. 3 was constructed, these lakes drained englacially for 1.5 km into the head of the ‘Red River’ (Gripp’s term). Since that time, surface melting, and possibly the collapse of the tunnel roof has exposed the Red River so that it drains the exposed glacier basin area via an entirely subaerial channel.

Within the dead ice zone there are complex series of long, well-defined narrow ridges lying parallel to the strike of foliation in the underlying ice and composed of sorted and bedded materials ranging from coarse gravel to fine sands (Fig. 9), many of which can be seen to rest on ice cores. We suggest that they are eskers reflecting the former positions of supraglacial and sometimes englacial streams. Although the trend of these eskers reflects the dominant modern pattern of drainage (towards the
Fig. 6. (a) The location of abandoned (1-4) and active (5) stream channels on part of the exposed glacier surface along a major drainage routeway, in which stream locations have shifted through time. (b) Cross section through the zone, showing how abandoned channel sediments have inhibited ablation of underlying ice so as to form elevated, ice-cored esker ridges. Note how in channel 5, stream gravels undercut the eastern ice wall of the channel. This reflects undercutting at meander bends such as those shown in Fig. 5.

Fig. 7. Profile across the outer part of the ice cored moraine zone and the pushed zone. The location of the interface between dead ice and pushed sediments is based upon seismic evidence.
Fig. 8a–b. Panorama across a) the outer part of the ice cored moraine zone and b) the inner part of the pushed zone (lower picture), looking east across the Red River. Note the extensive slump scars in the ice cored zone, with major sediment flows emanating from them. Much of the sediment is till, but some fluviatile and lacustrine sediments are also involved, producing flow mixing and lamination of sediments. Note that ice is exposed at several points at the incised margin of the Red River.
Fig. 9. Explanation of the origin of the ridges shown in cross section in e). (a) Plan of a supraglacial meandering stream such as shown in Fig. 5. (b) Long profile along the stream showing coarse sediments in shallow sections and fine sediments in meander bends. (c) Section across the stream whilst it remains active. (d) Accumulation of flows above fine-grained sediments in an abandoned meander pool. Prior to abandonment, the flows are disaggregated by stream flow. (e) Resultant meander bend forms on the glacier surface after abandonment and reversal of topography. The fine sediments below the flow till capping are illustrated in figure 9a.

trending towards several clearly defined points at the distal extremity of the ice-cored moraine zone (Fig. 3). We suggest that these esker systems reflect supraglacial (and possibly englacial) drainage patterns during the final phases of the Holmström breen surge and immediately after its completion. The drainage channels were located in the troughs between the highly compressed sequences of medial moraine ridges, and were only able to break out beyond these ridges at several well-defined localities. After collapse of the surge, the glacier, unprotected from ablation by a till cover, melted down more than the ice-cored zone, thus producing a counter slope and reversing the predominant direction of drainage in the ice-cored zone. The glacier’s main drainage network was no longer routed at surface through the ice-cored zone but flowed directly into the low terminal area of the exposed glacier, finding its way through the ice-cored zone via an englacial tunnel. Surface drainage in the ice-cored zone, became merely local. Thus, the pattern of drainage in the dead ice zone changed and the size of streams and the resultant fluviatile sediment bodies was reduced.

The early post-surge topography of the dead ice zone, of till-covered dead ice ridges with intervening fluviatile channels, changed after the diversion of drainage. The thicker, more stable, fluviatile deposits provided a better protection against ablation than the thinner less stable intervening till areas, and thus the topography was reversed, to the present topography in which fluviatile channels form ridges and till areas the intervening hollows (Figs. 6b and 9).

On the eastern side of the ice-cored moraine zone, the esker ridges do not trend parallel to glacier structures and to the general trend of the ice-cored ridge, but cross this trend (Fig. 3). The eskers do not discharge from the ice-cored zone at well defined points, but across a broad front. We speculate that the reason for this was the relatively unfurrowed surface of the ice-cored zone in this area at the end of the surge because the surge was directed along the axis of the valley towards the south and southeast. In this area, many medial debris septa have been highly compressed, whereas the eastern and northeastern margin, occupying a lateral position in relation to the surge axis, had a relatively small number of compressed debris septa (Fig. 4). Water was thus relatively free to drain downslope, away from the glacier, without being deflected laterally by moraine ridges lying above debris septa.

At many points along the lines of eskers in the ice-cored moraine zone there occur steeply-dipping masses of well-bedded, generally fine-grained sand with occasional outsize clasts. They are typically of the form and structure shown in Fig. 9e. They are common on dead ice surfaces in many geographic areas, but as yet there have been no published descriptions of them. They are often arcuate in plan with the sediment beds dipping towards...
the concave side of the arcs. Flow tills sometimes occur on their surface. We suggest that these may be analogous to features which are currently forming on the glacier surface. Streams flowing over homogeneous materials such as ice tend to meander. One such large meandering stream, cut into unstable, till-covered dead ice is shown in Fig. 5. Such well-developed arcuate meanders, which cut deeply into the ice walls on the outside of bends, are common features of well-established streams on ice surfaces, which, because of the high thermal capacity of meltwater tend to incise deeply by melting. These deep channels can sustain large meltwater discharges at relatively low velocities, and thus the sediment in them is frequently relatively fine grained, apart from the outsize clasts which fall into them from the till on the ice surface or from melting-out of debris from their ice walls. The channels are largest at the bends, with the ice surface or from melting-out of debris from their outsize clasts which fall into them from the till on the ice surface. Streams flowing over homogeneous materials such as ice tend to meander. One such large meandering stream, cut into unstable, till-covered dead ice is shown in Fig. 5. Such well-developed arcuate meanders, which cut deeply into the ice walls on the outside of bends, are common features of well-established streams on ice surfaces, which, because of the high thermal capacity of meltwater tend to incise deeply by melting. These deep channels can sustain large meltwater discharges at relatively low velocities, and thus the sediment in them is frequently relatively fine grained, apart from the outsize clasts which fall into them from the till on the ice surface or from melting-out of debris from their ice walls. The channels are largest at the bends, with narrow inter-bend straights where higher velocities are attained. Thus, we expect the largest finest-grained sediment masses to accumulate on meander bends. Fig. 9a–e shows the sediment character in such a channel and how it might develop after it is abandoned by an active stream. As stream downcutting occurs elsewhere and the site dries out, ice underlying the thin sediment on the inside of the bend will tend to melt more rapidly than that beneath sediments at the outside of the bend, thus tilting the whole mass towards the concave sides of the form. Small scale faulting is common in such sediment masses.

Where relatively fine grained fluviatile and lacustrine sediment masses have accumulated on the glacier surface, they are prone to the same instability and flow as supraglacial till. The flanks of many of the supraglacial meander deposits discussed above give rise to major sediment flows. In many cases substantial masses of supraglacial fluviatile and lacustrine sediment have completely lost their original form and structure and given rise to sediment flows which, by successive episodes of failure and flow, have been dispersed over relatively wide areas and mixed with flow till and till-derived flows (Fig. 8a). In these areas, inter-laminated diamiction and silt and sand flows are common.

About a kilometre from the southeastern margin of the ice-cored zone, the floor of a large trough lies between two medial moraine ridges and is occupied by a thick (>10 m) mass of fine sand and silt (Fig. 3). A very distinct upper margin to these deposits is marked at several points by small deltaic masses. They clearly represent the deposits of a former supraglacial lake. The lake overflowed via a point marked CH in Fig. 3 through a currently dry channel which cuts through the crest of the highest ice-cored ridge to the southeast and is confluent with the Red River gorge. Drainage from this channel ran into the Red River at the level of a terrace at CH1, some metres above the present river level.

4. The glacially pushed sediment zone

4.1. General

There is evidence that a belt of ridged terrain some 1.5 km wide from proximal to distal extremities (Figs. 2 and 3), which stretches as a broad arc across almost the whole width of the valley of about 5 km, has been pushed forward as a mass by the surging glacier, but not overridden by it. We refer to it as a push moraine, rather than, as has been suggested by Huddart and Hambrey (1996) for similar features, a thrust moraine. As we shall show, it has been emplaced by pushing rather than, for instance, subglacial shearing, although deformation during pushing has involved thrusting and folding.

On its surface are a series of ridges, elongated transverse to the valley, which reflect the overall arcuate form of the pushed masses. The ridges are metres in width (occasionally several tens of metres). The majority of major ridges can be clearly traced for 100–200 m in the transverse direction and occasional very large ridges can be traced for up to 1 km (Figs. 2 and 10). Most of the ridges reflect an underlying anticlinal structure, and their limited lateral extent suggests that they reflect fold culminations which die out laterally. This can be confirmed in most of the smaller folds (Fig. 11).

The pushed zone is cut by three major radial channels (Fig. 3), of which only the most easterly (the ‘Red River’) currently carries a major stream, and several minor channels with complex structurally-controlled courses. The walls of these channels expose the stratigraphy and structure of the pushed belt.

An almost complete, well-exposed section through the pushed sediment zone, roughly perpendicular to tectonic strike, was found on the eastern wall of the Western Gorge and in a tributary channel which joins it on its left bank 400 m from the proximal extremity of the pushed mass.

A detailed survey was made by taking a continuous series of polaroid photographs of this section. These were then overlain, in the field, by transparent sheets on which the details of the section were drawn. By mapping the location from which each photograph was taken and the distance to the base of the cliff, it was possible to correct the photographic distortion and to project each sketch section onto any given plane. The details are illustrated in Boulton, van der Meer et al. (1989). Fig. 12 shows the resultant geological section.

4.2. Sediments of the pushed belt

We distinguish six principal lithofacies in the pushed belt, exposed both at surface and in section. They are:

(a) a diamicton;
(b) a rhythmically bedded sequence of sand, silt and clay;
Fig. 10 (a) Plan showing the distribution of ridges on the surface of the pushed zone. (b) Strike lines of beds and fault planes intersecting the surface in the pushed zone. Several clear fold closures and fault displacements can be seen.
Fig. 11a–b. Fold structures coinciding with ridges in the pushed zone. Tectonic transport (direction of pushing was from left to right). (a) A box fold in the external zone (Fig. 12). The fold dies out laterally. (b) A tight fold coinciding with a ridge in the intermediate zone of asymmetric folds.

(c) a massive, homogeneous blue-black mud;
(d) a chocolate brown, fine sandy, silty mud; sands;
(e) sands;
(f) sands and gravels.

(a), (b) and (c) have a restricted occurrence, (a) and (b) at the proximal extremity of the pushed belt, and (c) at a single outcrop near the southwestern side of the pushed belt. In the Western Gorge, (d) and (e) are the main units. Near the ice-cored zone, (e) passes gradually into (f).

Unit (a): The diamicton is a subglacially deposited till. Where exposed it contains abundant striated bullet-shaped clasts and is dense and clearly overconsolidated. A seismic survey carried out in 1985 shows it not to be ice-cored at the proximal end of the Western Gorge.

Unit (b): In an outcrop at the proximal extremity of the folded sequence, a rhythmically bedded sequence of fine sand, silt and clay with a thickness of at least 5 m was found (Fig. 13). The stratigraphic position of this unit with regard to the other units of the pushed belt is unknown. The unit consists of a regular alternation of parallel bedded laminae of fine sand, silt and clay, varying in thickness from a few mm to a few cm. Sand and silt predominate over clay. Sand laminae have a sharp lower and gradational upper boundary. Apart from a small frost crack, no other macroscopic sedimentary structures occur. Based on the limited amount of data, we suggest that the unit has been deposited below wave base in a standing body of water, with the sand and silt deposited by turbidity currents, and the clay by settling from suspension. The unit is incompatible with the depositional environment of a proglacial outwash plain. It is a typical varve sequence and appears to reflect the former presence of a lake.

Unit (c): This is a massive, well-consolidated blue-black mud, with shells and occasional clasts, which has only been found in a sub-surface exposure at a single locality near the southwestern extremity of the pushed mass (see Boulton, van der Meer et al., 1989). It contains numerous sub-horizontal slicksided joint planes which cut the sediment into well-defined lenticular shear lenses some centimetres or decimetres in thickness. It weathers to a reddish-brown colour. According to Spaink (1986), two groups of molluscs are found in this sediment. One lives in seawater depths of 10–30 m. The other lives intertidally. The former reflects cold climatic conditions and may date from the late-glacial. The latter reflects a distinctly warmer climate and may have lived during the postglacial. The sediment is glaciomarine.

Unit (d): This unit, with its characteristic chocolate brown colour, dominates the pushed belt. We suggest that it was deposited in a subaerial environment. It underlies unit e, the outwash plain sands, with a sharp intervening contact. Although fine-grained sands are present in the unit, in particular in its upper part, silt and clay are the dominant constituents. Two sections were measured, one reaching 2.5 m (A) below the sand/mud boundary, the other (B) about 5 m below it (Fig. 14). Section A, between 45 and 65, represents entirely the supratidal part of the mudflats; section B, between 205 and 225, the supratidal and intratidal part. The common presence of driftwood on top of the pushed belt, obviously weathered from the exposed muds, suggests that intratidal and subtidal deposits must form an important constituent of the folded muds.

Both sections show a distinct coarsening-upward trend with more sand in the upper part and more clay below. Alternation of parallel bedded, thin laminae of sand, silt and clay dominates and point to deposition from sheet flows. Parallel bedding is interrupted by trains of ripples of fine-grained sand, which increase in number and thickness upwards. These could represent small gulleys as shown in Fig. 23. Loading and strong syn-sedimentary deformation of these fine sandy and silty beds is common, indicating quicksand formation associated with mud volcanoes and collapse structures. In section A, rooted plant remains occur near its base, indicating that
Fig. 12. A composite section in the Western Gorge. Note the decreasing intensity of deformation from the proximal to distal extremity of the pushed belt: an internal zone of thrust and gliding nappes; an intermediate zone of asymmetric folds and overthrusts; and an outer zone of parallel folds, box folds and small scale thrusts.
in a 30 cm thick top layer of imbricated gravel up to 5 cm thick. Most current-generated structures indicate a palaeo flow direction towards the south; from 5.60 to 4.75 m northerly directions dominate. At about 590 m, a pit was dug on top of a gravel-capped ridge. Permafrost at 1.3 m prevented further digging. Most sections examined in lithologies (e) and (f) were similar. Locally, climbing ripple cosets or muddy fine sands and silts may be dominant. The distribution of these sediments at the surface is shown in Fig. 3.

If the effect of folding and faulting is removed, the stratigraphy exposed in the Western Gorge is relatively simple and comprises a predominantly muddy unit of facies (d), overlain by sandy and gravelly facies (e) and (f). At the proximal extremity of the pushed mass, beyond 820 m, muddy and sandy facies are mixed in a more complex fashion than elsewhere. This may be largely an original sedimentary feature or it could be a product of repetition due to folding, an ambiguity difficult to resolve because of the strong folding and faulting of this zone. At 920 m rhythms (facies (b)), underlie sands (facies (e)), and at 940 m till (facies (a)) occurs, although we were unable to establish its relationship to facies (b).

4.3. Evolution of drainage in the pushed belt

Comparisons have been drawn between the lithofacies represented in the modern Holmström outwash system and those which occur in the push moraine; in particular the similarity between the gravel and sand units in the push moraine and the proximal and distal fan deposits of the ‘Red River’ (Fig. 3), and between the muddy facies (d) in the ridges and the modern pro-fan mudflat deposits. The simplest explanation of the sequence exposed in the Western Gorge is that it represents progressive progradation of an outwash fan over pro-fan mudflats. In such a system we would expect the fan deposits to be generally relatively thin, as they are in most of the Western Gorge section. Thickening of the sandy facies between 340 and 410 m and between 190 and 260 m suggests either that major channels developed in these areas, or that tectonic thickening of the sands has occurred.

Sections elsewhere in the numerous channels which cut the pushed mass, and in the walls of the Red River, show a stratigraphy similar to that exposed in the Western Gorge, although in many places the sand and gravel capping is lacking.

It proved possible to map the extent of this capping, both in the field and from aerial photographs. Fig. 3 shows the distribution of sandy and gravelly units exposed on the surface of the pushed complex. Two principal areas of former fan sediment occur, one flanking the Red River and a second flanking the Western River. They are located beyond the points where supraglacial esker systems discharged from the ice-cored zone. At several
Fig. 14. Sections in the intratidal and sub-tidal sediments of unit (d), which form the dominant lithologies of the push belt, and the lower part of the overlying glacial outwash fan sediments of unit (e). Note the soft sediment deformation and water escape structures in unit (d).

points on the folded surfaces of the former gravel fans, lineations occur which we interpret as the margins of former channels and channel bars (Fig. 3). These lineations appear to emanate from the distal extremities of eskers in the ice-cored zone. We therefore interpret the superficial distribution of sands and gravels as a reflection of the location of two former outwash fans, which were ancestors of the present Western Gorge and the Red River, and were deformed by the glacier-pushing episode.

It is clear however, that not merely did fan deposits pre-date the main phase of folding, as most of those within the pushed sequence clearly did (Fig. 12), but that some fan sediments were contemporary with the folding episode. In some places, a channel, in one part of its course, may be deflected around several folded ridges, suggesting that it post-dates them, but in a more distal position the channel gravels, still showing clear channel margins and channel bars, pass over a ridge (Fig. 15), suggesting that here it pre-dates ridge formation. Such syn-tectonic outwash features are common in the vicinity of the two former fans, and generally suggest that tectonic activity migrated outwards in the push moraine.

Whilst the major masses of pre-tectonic fan sediment retain the form of fans spreading widely over the surface, it can be shown in several places that much syn-tectonic outwash deposition occurred in channels in the folded belt.

We can thus distinguish four principal phases of outwash activity in the glacially pushed zone.

Phase 1: A pre-tectonic, mudflat phase, when the area was analogous to the present pro-fan mudflat belt. This
Phase 1: A phase of pre-tectonic fan development when two broad fans prograded into the area, fed by major rivers located roughly in the position of the present Red River and Western Gorge, and a minor southern margin stream. Lithofacies (d) represent this phase.

Phase 2: A phase of syn-tectonic drainage represented by the major esker systems in the ice-cored moraine zone and by the syn-tectonic fan gravels in the pushed zone. Channel fills which cross-cut planar units of lithofacies (e) and (f) (e.g. 220 and 535 m, Fig. 12) presumably reflect incision related to the rejuvenation of streams as their glacial source approached and tectonic uplift in the pushed belt increased stream gradients. These incising syn-tectonic streams probably fed fans located in the same positions as the modern fans, and were fed by the eskers of the ice-cored moraine zone.

Phase 3: A phase of syn-tectonic drainage represented by the major esker systems in the ice-cored moraine zone and by the syn-tectonic fan gravels in the pushed zone. Channel fills which cross-cut planar units of lithofacies (e) and (f) (e.g. 220 and 535 m, Fig. 12) presumably reflect incision related to the rejuvenation of streams as their glacial source approached and tectonic uplift in the pushed belt increased stream gradients. These incising syn-tectonic streams probably fed fans located in the same positions as the modern fans, and were fed by the eskers of the ice-cored moraine zone.

Phase 4: A post-tectonic phase in which downcutting continued, producing highly incised streams and feeding fans beyond the pushed belt. Downcutting during this phase probably encouraged the development of the Red River, which captured water from the western channel and channel CH-CH1 which had hitherto been fed by subaerial systems.

The evolution of drainage during the later part of the syn-tectonic phase and the post-tectonic phases can be reconstructed with some confidence (Fig. 16). During an early post-tectonic phase prior to 1927, both the Red River and the Western Gorge were major drainage outlets of water from the dead ice zone (Fig. 16A). When Gripp visited the site in 1927, the Western Gorge had already ceased to discharge significant quantities of meltwater. The predominant discharge was by the Red River, which flowed for some distance in an englacial tunnel (Fig. 16B). The oblique aerial photographs of 1936 show this situation (Boulton et al., 1989). Vertical aerial photographs from 1966 and 1970 suggest two modes of flow of the Red River at this time. In the 1966 photographs, the Red River disappears down a large moulin (Fig. 16D), to reappear lower downstream. Along the upper course of the river, extensive lake sediments occur, with a well-marked lake shoreline and lake-margin deltas. In the 1970 photographs, the whole of this area is occupied by a lake up to the level of the shoreline in the earlier photographs, and draining through a surface channel at CH (Fig. 3), from which water flowed into the Red River. It is probable therefore, that during this phase, drainage alternated between the Red River’s englacial tunnel (mode 1, Fig. 16D) and the channel at CH when the Red River was blocked and a supraglacial lake formed (mode 2, Fig. 16D). This may have been a seasonal event, associated with cessation of drainage in winter, and freezing of the Red River’s englacial channel. By 1984 the englacial tunnel of the Red River had collapsed, producing an entirely sub-aerial channel which drained a very large lake between two major moraine loops (Fig. 16E).

4.4. Structures in the glacially pushed sediment zone

The structures exposed in the Western Gorge and its tributary permit a structural geological analysis of the whole fold belt. The analysis has four goals:

- to define structural styles in different zones;

![Fig. 15. Panorama showing syn-tectonic gravel sediments in the pushed zone. The well-defined surface lineations, particularly prominent in the central part of the panorama (marked T), are channels and bars formed on the surface of an outwash fan. The surface has been tilted and the lineations can be seen folded with the push ridges to the right (marked F). In a more proximal position, the channel pre-dates local folds. These features reflect outward migration of tectonic activity, with proximal elements of drainage systems being folded and uplifted whilst fluvial activity continued in their distal parts. The features are located in the zone of lineations immediately to the west of the Red River shown in Fig. 3.](image-url)
- to establish finite strain within folds and other structures, and to give an accurate estimate of tectonic shortening;
- to determine the bulk tectonic shortening of the fold belt;
- to reconstruct the depth and geometry of the decollement surface of the fold belt.
- to understand the processes which have led to push moraine formation.

We divide the section of the fold belt exposed in the Western Gorge into three zones on a structural basis (Fig. 12):

- **An external zone**, from the foreland thrust (4 m) up to 255 m, dominated by parallel folds and small-scale thrusts.
- **An intermediate zone**, from 255 to 690 m, characterised by strongly asymmetric folds and overthrusts.
- **An internal zone**, from 690 m to 950 m, dominated by thrust and gliding nappes. Between 780 and 850 m, structural and lithostratigraphic relationships are complicated and differ from the other structures in this zone.

### 4.4.1. Structural style in the external zone—4–255 m (Fig. 12)

Folds in the external zone are mainly open folds with axial surfaces inclined up-glacier at about 45° (Fig. 17). Synclines are usually tighter than anticlines (e.g. 73–80 and 110 m, Fig. 12). Thrusts frequently occur in the lower limb of distally asymmetric folds, dipping steeply towards the hinterland with movements up to several metres along the fault planes. A few small backthrusts are

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**Fig. 16.** Evolution of the Holmstrømbreen drainage from the pushing event until 1985.
found near the front of the fold belt, some of which are refolded (e.g. 25 m). More proximally, most folds are asymmetric boxfolds with steep or even overturned southern limbs (73–80, 110 m, Figs. 12 and 18). Both wavelength and amplitude of the folds decrease towards the foreland from 30 to 9 m and from 15 to 3 m respectively.

Minor thrusting and drag folding has occurred on the limbs of the box folds (40, 90, 170 and 180 m) as a normal consequence of parallel folding. Post-tectonic erosion has removed only a little material from the tops of structures. We have not found a well-developed gravel lag on top of the folds in this part of the push-moraine, suggesting that wind erosion has left the structures here reasonably intact.

Where possible (mainly in the external zone) we have classified folds according to Ramsay’s (1967) classification. There is a marked contrast in fold style between the sandy beds and the silty and clayey beds. Within the same structure coarse-grained beds tend towards class 1B to class 2 folds, with a maximum in class 1C, whereas the muds score highest in class 1C to class 3, with a maximum in class 3. Sand beds near the sand-silt interface show a slight preference for class 1B (parallel) fold geometries. The overturned folds in particular show a development towards similar fold style (class 2) in either direction away from the sand/silt interface. This may in part be attributed to flexural slip and to volume problems near fold hinges.

These observations let us conclude that the folds in the lowest sand beds are the nearest approximations to ideal parallel folds. We thus use the form of the sand/silt interface to estimate bulk tectonic shortening in the fold belt. This general conclusion does not however appear to apply to the section between 130 and 140 m, where the sand/silt interface appears to have acted as a glide plane along which the sand sequence has been strongly compressed into a sequence of tight folds and faults not reflected in the underlying silts. This zone also coincides with an abrupt thickening of the sand sequence. From 130 m to the foreland it is 1–2 m in thickness, from 130 to 430 m it is 4–10 m in thickness. A sudden sedimentary thinning of the sand overburden would reduce the strength of the sand/silt interface and the resistance to
buckling of the sand stratum so that the edge of the thicker sand unit would act as a ram, able to buckle the thinning sand and cause it to slide over the silt surface. We thus regard this disharmony between folding in the sand and the form of the sand/silt interface as a local reflection of a sedimentary change. We suggest in general that the variations in thickness of the sandy unit result from syn-tectonic erosion and sedimentation.

At the distal extremity of the cross-section, folds appear to die out and merge imperceptibly with an untectonised foreland. At the most southerly extent of the fold belt, between the Red River and the Western Gorge, and to the west of the latter, pre-tectonic sediments (lithofacies (e) and (f)), continuous with those of the fold belt, underlie a very thin horizon (less than 10 m) of recent pro-fan sediments. In many places these sediments lie in the troughs between the outer, low-amplitude folds belt. As the pre- and syn-tectonic sediments which form these folds pass beneath the recent, post-tectonic sediments, the folds die out and no major tectonic discontinuity has been found. We therefore conclude that the distal extremity of the fold belt was fixed during folding.

4.4.2. Structural style in the intermediate zone, 255–690 m (Fig. 12)

Thrusting and folding are much more intense in the intermediate zone than in the external zone (e.g. Figs. 19 and 20). At least three major thrusts have been observed in this zone, defined by mud-on-sand-sequences:

(a) between 260 and 280 m upward facing sands are overlain by downward facing silt and clay beds (transport >12 m);
(b) between 433 and 440 m sandy beds form a footwall ramp beneath intensely folded silt and clay beds (transport >8 m);
(c) between 485 and 517 m hanging wall deformation is even more intense (transport >20 m).

These three are fold nappes, showing different degrees of internal deformation as a consequence of their different transport distances. Fig. 21 shows examples of hanging wall deformation at the base of nappes (b) and (c). Where the front of a nappe has climbed a footwall ramp, hanging wall folds are tighter, as at 520–540 m.

The anticline-syncline pair near the nappe front at 260–270 m is oriented obliquely to the main structural trend, probably caused by an oblique footwall ramp below the anticline. Fold axes plunge at 20 and 10°, respectively towards 305°, while neighbouring structures trend east-west (see map, Fig. 10).

Anticlines such as those at 570–575 and 585–590 m (Fig. 12), 610–620 and at 640–660 m still preserve some of the characteristics of box folds, but the syncline-anticline pair at 590–600 m is completely different. It has a fold style we more frequently find further to the north (e.g. at 630 and 660 m), resembling an intermediate stage of an evolution towards the glide nappes at 755–790 m.

4.4.3. Structural style in the internal zone, 690–950 m (Fig. 12)

Strong overthrusting along low-angle planes is developed in this zone, demonstrating an overall increase in the shear component of deformation compared with the buckled zone beyond. Between 740 and 790 m, a series of flat-lying shallow rooted nappes occurs, some of which even dip towards the foreland. We suggest that these may have initially developed as high angle structures which reached their extreme development as gravitational flows when the muddy units had broken through their competent sandy capping, then advancing as gravity-driven flows bearing, in some cases, their sandy overburden with them.

Between 825 and 855 m, a series of backthrusts and hinterland directed nappes occur. It has proved very difficult to unravel the intricate structural relationships here, as the lithostratigraphy seems to differ from that in the rest of the fold belt. This may lead to a gross underestimate of the tectonic shortening of this part of the push moraine. On their proximal side (850 m) these structures appear to overlie the nose of a recumbent anticline which is much less deformed. We correlate the recumbent anticline with a structure at 920 m, on the north side of a broad gulley, which also shows relatively minor small scale deformation compared with the structures between 825 and 855 m.

These structures do not lie far from the buried stagnant ice zone; the remains of the glacier which produced the folding. We suggest that the relatively less deformed sediment masses between 850 and 920 m behaved as a passive wedge in front of the advancing glacier. At 850 m it drove into the sediments beyond it, which mounted the wedge along backthrusts and backwardly directed nappes. As the deformation continued, folding and thrusting developed in the direction of the foreland. Thus, the passive wedge mechanically coupled the glacier and the fold belt.
We conclude that each zone is characterised by a limited number of structures, belonging to different structural ‘families’ (Dahlstrom 1969). The external and intermediate zones may be compared with Dahlstrom’s ‘foothill structural family’, which consists of concentric folds, detachments (decollements), low-angle thrusts, tear faults (strike-slip faults across the main structural trend) and late normal faults. The external zone is dominated by open box folds with minor high-angle thrusting. Low-angle thrusting is more important
in the intermediate zone and the similar (classes 2 and 3) folds tend to be tighter. Tear faults may be recognised from the map of the push moraine. They accommodate differential tectonic transport due to the curvature of the fold belt.

Structures in the internal zone are predominantly attenuated recumbent folds and rootless gliding nappes, of which the most proximal ones are hinterland directed. The internal zone has some remarkable parallels with the gliding nappes of the northern Apennines (van Wamel et al., 1985).

A single ridge or several associated ridges of till lie between the proximal extremity of the folded sediments and the dead ice zone. We suggest that these are pushed ridges which mark the advancing tip of the glacier as it rode over a ramp above the folded sediment mass.

The extreme proximal zone between 910 and 940 m is particularly interesting. At 920 m, a closely spaced series of overthrusts have carried upwards a sequence of clay, rhythmically bedded sand-silt/clay couplets, and sands. The clays and rhythmites (Fig. 13) are lithofacies not seen elsewhere in the fold belt. The rhythmites are similar in appearance to lacustrine varves.

The sands beneath the thrust contain very distinctive structures consisting of separated bulbous basin structures between which tightly folded antiforms occur. Although the axis of symmetry of the lower parts of most basins is vertical, their upper parts are often overturned in the direction of overthrusting. They tend to occur in sets beneath obvious planes of shearing, suggesting that they are genetically related to the thrusting process. They are both cut by and deform faults, demonstrating that they are penecontemporaneous with deformation. We suggest that the structures may be water-expulsion structures reflecting drainage and consolidation of the mass during deformation. In such a case, a thrust plane and the sediments immediately above it would be a zone towards which water would naturally migrate, as movement along the thrust would produce drainage by advection in the sediments immediately above it. It would not therefore be surprising to find such structures immediately beneath a thrust plane.

4.5. Strain in the pushed belt

The net longitudinal shortening apparent in the surficial horizons of the fold belt can be assessed with relatively little ambiguity if we assume that in the sandy units at least, most of the strain shows itself in faulting and folding. Even in the sandier parts of the mud unit, thin sections support this view (Fig. 20). Thus, the shortening per unit length (Fig. 22a) and the cumulative shortening (Fig. 22b) of the interface between sandy and muddy units can be determined. Shortening increases towards the proximal extremity of the fold belt, an index of increasing intensity of folding and thrusting. Between 320 and 400 m (Fig. 12), we are not sure whether a thrust plane has been refolded to trap a muddy unit between sandy beds, or whether the thickening of the sandy beds with an intervening mud unit reflects change in the location of distributary channels on the fan. If a folded thrust is the correct explanation of this feature, the aggregate shortening will increase from this point by about 100 m at all points. However, as such a structure would appear to be out of place in this zone, we prefer a sedimentary explanation of the sequence.

We estimate the overall shortening of the proximal extremity to be 670 m, or 710 m if we accept the folded thrust hypothesis. This gives a net longitudinal compressive strain in the fold belt of 0.58–0.46.

There are, however, some identifiable errors in this simple estimation of strain. The mud-on-sand-interface near the front of fold nappe (a) is almost strain free. The muds rest upside down on the sandy beds, and we may conclude that a large recumbent fold of silt and clay has less or less rolled over the sands without apparent shear at the interface. The front of the fold is strongly brecciated and shows the crushed remains of the steep limb.
that have probably been dumped into a snow-filled depression. We have found another breccia related to such a ‘roll-over-fold’ at 793 m. This mechanism implies a large amount of simple shear and pure shear (flattening) within the muds, so that the original interface length is smaller than its present length. We can envisage the structure as a conveyor belt carrying a mass of sand over a sandy substrate. The overlap of the sand units can then be used to approximate tectonic shortening here.

This flattening is small in the external zone, but increases substantially towards the internal zone. It is possible to estimate the amount of flattening strain (pure shear) with the aid of the orthographic method (De Paor, 1986) to correct the calculated shortening. We do this for the sand beds immediately above the sand/silt interface (which we use for our estimate of bulk shortening), assuming plane strain (flattening and extension in the plane of the cross section only). A first approximation of the strain ellipses’ axial ratios gives values in the order of 4.8 (578 m), 3.9 (530 m), 2.0 (410 m) and 1.4 (220 m). We obtain the following values for the extra horizontal shortening due to flattening: 0.54, 0.49, 0.29 and 0.15, respectively. The extreme values of horizontal pure shear are invariably found near the leading edges of nappes, i.e. places where we may expect a high concentration of strain. For example, near the front of nappe (c), between 500 and 580 m, we find folds which have apparently undergone a very large amount of flattening superposed upon the buckling strain. This flattening drops to very low levels immediately behind the zone of high strain concentration, as can be deduced from the fold geometries between 600 and 690 m. The sand beds in the footwall of this thrust are class 1A/1B and seem to have undergone little flattening.

We conclude that in the hanging walls and footwalls of large overthrusts, a substantial amount of horizontal pure shear is added to the horizontal shortening produced by buckling. We calculate that shortening values between 500 and 580 m should be multiplied by 0.90 and those between 380 and 430 m by 0.95. This correction is locally significant but not generally important.

Fig. 22a and b shows that there is a general increase of compressive strain from distal to proximal extremity of the fold belt, although a great deal of the strain is concentrated in well-defined anticlinal structures (e.g. 250 m, 500 m). The lower limbs of these anticlines are often cut by thrust planes (e.g. 210 m) which must have acted as ramps along which the proximal sediment masses have overridden distal masses. These important structures, and the inferred ramps which they reflect, coincide with locations where a sand and gravel capping is lacking. It is unlikely that wind deflation has eroded the crests of these anticlines to achieve this, as an armoured lag surface would have built up, and nearby higher sandy crests have survived (Riezebos et al., 1986). We suggest therefore that major anticlines and thrusts which represent concentrations of compressive strain, have developed where the sandy upper unit was thin or absent. The underlying finer-grained beds were thus permitted to expand upwards and strain to be taken up where there was no stiffer overlying sand unit (unit (e)).

We conclude that the original distribution of sandy sediment at the surface strongly affected the rheological response of the fold belt to the glacier’s driving force. The sands were relatively stiff compared with the underlying mud sequence and must thus have played an important role in transmitting glacially-generated forces from the proximal to the distal extremity of the fold belt. Where it was thin or absent at surface, a zone of weakness occurred in the fold belt which led to a concentration of strain in these zones and the development of local ramps. The concentrations of strain in the muds in these zones is reflected by small scale tight disharmonic folds which contrasts with the relative absence of such folds where the muds lie beneath a thick sand unit (e.g. Fig. 12, 585, 460 m).

4.6. Evolution of the folded mass

The structural attributes of the folded mass and the distribution of pre-, syn- and post-tectonic fluviatile sediments in the mass lead us to conclude that the spatial sequence from undisturbed foreland through the distal to proximal extremity of the fold belt also reflects a temporal sequence, in which the inner zone of nappes has developed through an early phase of box-folding through a phase of overthrust faulting to a final phase of nappe development. If this is so, it implies that the advancing glacier initially pushed up a narrow belt of low amplitude box folds and that as it advanced, these evolved into overturned folds with thrust lower limbs and a further belt of low amplitude box folds developed beyond them. Thus, a wave of progressively greater deformation spread out beyond the advancing glacier. As the proximal extremity of the moraine was pushed forward by the advancing front of the glacier and was subject to progressive compressive strain, so did the leading, distal margin of the moraine advance by the development of new folds and fractures in hitherto undeformed fan sediments (Fig. 27).

We presume that the distal edge of the moraine has sometimes been marked merely by buckling of the foreland (as at the southern end of the Western Gorge), and sometimes separated from it by a basal thrust, as might have happened at the southern end of the Western Gorge if further pushing of the moraine had occurred. The Jura-type style of folding in the outer zone suggests strongly the existence of an underlying decollement. Even though at the latest stage, displacement along the decollement, roughly reflected by the shortening shown in Fig. 24, fell to zero near to the toe of the fold belt.
At each phase of distal extension of the push moraine, the basal decollement must come to surface and strain would have been taken up by uplift of the surface. At some point, further strain would have been more readily relieved by development of a new decollement in front of the contemporary push moraine and outward movement of the moraine mass along it. Some of the thrusts exposed in the section may have performed this role before further more distal decollements became active, although some of the thrust planes will reflect local accommodations to stress within the deforming mass.

As the intensity of deformation and degree of tectonic uplift increased in the internal zone we would expect the locus of fan sedimentation to be displaced distally. The area where the sand unit is thickest, between 130 and 430 m, is also one where several possible syn-tectonic sand masses occur. It is probable that sand masses at 405, 360 and 230 m which lie in synclinal folds and show shallower dips or dips discordant with the underlying beds, are also syn-tectonic features and represent channel fillings in the troughs between pre-existing fold ridges which themselves underwent deformation as the folds developed further. There is some evidence that post-tectonic sand beds which overlie folds in the internal zone can be traced into folded syn-tectonic sands in the intermediate zone.

4.7. Position of the decollement and its relation to folding

Sedimentological evidence suggests that the topmost sedimentary unit (unit (e)) in the pushed belt was deposited on an alluvial outwash fan similar to the fans which emanate from the mouths of the Red River and the Western Gorge (Fig. 25). Both these fans have similar slopes. If we assume that a pre-tectonic fan of similar surface profile to the modern fan formed by the Red River (Fig. 26) was deformed to create the Holmstrømbreen push moraine, whose profile is known; and using our conclusion that pure shear has been the principal agent of geometric distortion, it is possible to infer the depth of the decollement surface along which movement has occurred from the evidence of shortening.
Fig. 24. Highly schematised view of the inferred geometric evolution of the Holmstrombreen push moraine. The figure shows the position of the inferred ice contact face into which the surging glacier drove, the suggested slope of the pre-tectonic fan which emanated from the ice contact slope, the form and structure of the moraine at the end of the pushing episode, the depth of the decollement beneath the push moraine and the point to which the earlier ice contact face had been driven by the surge. Faults and folds observed in the Western Gorge section have been projected downwards to the decollement surface. The decollement plane is drawn as a smooth surface. In reality, it may have contained small-scale ramps and irregularities.

(Fig. 22). The incremental shortening ($S$) of a mass of average original thickness ($t$) produces uplift of a mass of longitudinal section area ($A$) such that $St = A$. The reconstructed decollement depth is shown in Fig. 24. The consequences of the structural ambiguity at 320 m (Fig. 22a and b) shown in the reconstruction of shortening are clearly seen. If the shortening at 320 m is small, the decollement plane will be depressed in this zone by about 20 m in comparison with the decollement reconstructed from the case where there is large local shortening at 320 m. In the absence of an obvious reason for the depression on the decollement surface, we prefer the interpretation consistent with major shortening at 320 m and no deepening of the decollement. The thickness of the thrust nappe is therefore inferred to be about 30 m. Even if we presume that there has been horizontal stretching under shear of individual units in the pushed mass, the calculated decollement depth represents a minimum.

A seismic and resistivity survey of the dead ice and push moraine zones were conducted during 1985, and will be reported in a later paper. The survey of the proximal extremity of the pushed mass suggests that glacier ice in the dead ice zone extends to a depth of about 40–50 m below surface (Fig. 26). This is consistent with a pushing event in which the inferred decollement surface is a direct continuation of the ice/bed interface at the point of contact of the pushed mass.

The back-folding between 800 and 900 m (Fig. 12) can then be explained as sediment which glides up the advancing face of the glacier, which acts as a ram as it moves forward during the pushing event. The folds and faults exposed in the western gorge have been extended downwards in Fig. 24 to indicate their possible relationships to the underlying decollement surface.

5. The proglacial outwash zone

The push moraine is cut by the two major post-tectonic channels of the Western Gorge and the Red River, of which only the latter contains a major river. Relatively minor streams also drain the glacier at the eastern and western margins of the valley; of even less importance are the small streams emanating from the pushed belt. The Red River and other streams feed a broad outwash plain which drains into a fjord-head delta at the head of Ekmanfjord (Figs. 2 and 23).

The outwash plain is 1–3 km wide, has a general slope of less than 1:200, is characterised by braided stream systems, and dominated by the coarse gravel fans of the Red River and the Western Gorge. The fjord-head delta is about 5 km from its proximal to distal extremities, including its intertidal part and excluding the delta foreslope. It has a general slope of more than 1:1000, is characterised by a few sinuous channels and dominated by mud deposition.

Outwash plain and delta have been fed by channel systems which have changed with time (Fig. 25). A large part of the gravel fans were fed by streams draining the glacier at an earlier stage. This is evident in the fan at the mouth of the Western Gorge, which must have been fed by a major river in the early post-tectonic stage, but which ceased to be active prior to 1936. Oblique aerial photographs from that year show the Western Gorge to be as dry as today. Comparing aerial photographs from 1936 and 1966 with the situation in 1984, it is evident that the active fan apex of the Red River has shifted southward for a distance of several hundred metres because of incision by the stream flowing out of the gorge section into the earlier fan (Fig. 25). The disused fan, with its apex
at the mouth of the gorge section, was still in use in 1936, and probably mainly dates from the time that the eastern section of the glacier was drained through channel CH-CH1. Capture of this river, and the other streams to the west by the Red River caused this to cut down through the earlier fan, and to displace its active apex to the south. Displacement of the fan apex is even more conspicuous in the interval ‘66–’84, and can only be explained by continued downcutting of the Red River in its higher reaches as part of the continuing readjustment of the fluvial system to the glacier pushing event.

Sediments of the Red River fan are mainly coarse-grained channel and bar gravels, grading into sandy gravels and sand where the braided streams reunite to form the well-defined channels of the mud flats proper. As the marginal streams and those discharging through the pushed belt mainly carry mud, the fan of the Western Gorge is largely covered by the fine grained fan deposits of these streams. The non-active lobes of the Red River fan are partly covered by fine fan deposits. Because of the slope of this part of the outwash plain, these fans consist of a braided system of small shallow gulleys carrying fine grained sand separated by sheet flow sequences of silt and clay. These fine-grained sediments have a high water content, probably because of inhibition of drainage by underlying permafrost. Small mud volcanoes commonly occur, probably due to overpressuring produced by seasonal freezing.

In the outer mudflat zone, the braided patterns of the fans reassemble into a few sinuous channels. The main discharge is through low sinuosity tidal channels along the eastern side of the fjord. In addition, a number of meandering ‘overflow’ channels discharge the Red River fan. Comparing the suspension plumes at the delta front, more than 80% of the discharge is by way of the main channel which has a depth of over 5 m at the delta front. It carries sand up to its mouth. All channels have pronounced levees, indicating that deposition on the mudflats mainly occurs during peak discharge or as a result of high water level in Ekmanfjorden during southerly storms. The flats between the channels consist of silt and clay, probably largely deposited from sheetflows. Crevasse channels occur in the transition area of fans to channelled mudflats. On the latter, crevassing is uncommon, probably because of the cohesive nature of the sediments.

The tidal range at the delta front is small (1–1.5 m), and the influence of the tides on deposition on the mudflats is probably not important. During incoming tides, some suspended sediment from the plumes in front of the delta may drift into the intertidal bays and settle. These intertidal bays or flats between the channel levees have a poorly developed drainage pattern which is largely independent of the main delta channels.

Despite the relatively high sinuosity of parts of the overflow channels on the mudflats, lateral migration of the channels is small. Channel positions and shapes of the levees on the oblique aerial photographs of 1936 are very similar to those of the 1966 and 1984 photographs, except for the outlet of the main channel, which was bifurcated in 1936. Before 1966 the western branch of the forked outlet silted up, but on the 1966 photographs, its former presence can still be recognised from the shape of the levees. Stability of the channels is probably largely due to the erosion-resistant, cohesive nature of the muds. Cohesiveness may be re-enforced by the presence of segregation ice. The nature of discharge may be another
Fig. 26. Schematic section from the head of Holmstrømbreen to the foot of the delta foreslope.

factor. As drainage of the glacier occurs by way of englacial lakes, the latter might buffer the system so that peak discharges causing erosion of the channel banks and flooding of the levees are rare. If this is correct, most deposition on the outer mudflats has occurred prior to 1936. Since then almost all suspended load has been transported directly to the delta front. As the photographs of 1936, 1966 and 1984 were taken at different tidal levels, it was not possible to estimate progradation of the delta front since 1936, but in any case, it must be small.

Bedforms related to depositional processes are scarce on the mudflats. Most conspicuous are polygonal patterns of ridges on the channel levees. The ridges are about 10–20 cm high and occur as quite regular polygons with a diameter of 50–200 cm. In cross-section the ridges appear to be formed by the expulsion of sediment and water. Similar polygonal patterns have been described from mudflats of sub-arctic Canada (Dionne, 1987) and are attributed to the presence of polygonal lenses of segregation ice in the sediments, or to loading by ice fœs and the intrusion of sediment in a polygonal pattern of cracks in the icefoot. The ridges enclose flat areas with occasional small mud volcanoes and collapse structures. Linear series of mud volcanoes are believed to be related to cracks in underlying segregation ice.

The foreslope of the Holmstrømbreen delta and the fjord bottom beyond it were mapped morphologically by a series of echo-sounder profiles from a small boat (Fig. 23). Channel levees extend onto the delta foreslope. The upper foreslope has an angle of slope of $1/12^\circ$; typical of muddy deltas. An echo-sounder profile aligned at right angles to the delta front in its central part showed a large mound some 4 m high, which breaks the smoothness of the foreset/bottomset intersection about 800 m from the delta front. We suggest that this may be the toe of a slump derived from failure of the upper part of the delta foreslope.

6. Origin of the Holmstrømbreen tectonic-sedimentary system and the significance of push moraines

6.1. General geometry of the system

Fig. 26 shows a schematic longitudinal section through the whole of the Holmstrømbreen system, from the ice divide at the top of the glacier catchment (Fig. 4) to the foot of the delta foreslope. The section through the glacier from the origin to 23 km is based on a radio-echo traverse kindly supplied by Dr D. Drewry (personal communication 1988) and the interface between dead glacier ice and the pushed-sediment zone at 29 km was established by the seismic and resistivity survey carried out in Spring 1985. The radio-echo data suggests that the subglacial bed of Holmstrømbreen descends below sea level some 7 km from the glacier terminus. This taken together with the seismic data which shows the glacier/sediment interface dipping steeply below sea level, to at least $-70$ m, suggests that a general sub-sea level...
depression of the glacier bed occurs in the terminal 7 km of the glacier.

The material in the pushed zone above the extrapolated level of the outwash plain, or even above sea level, is at least ten times smaller than the volume of the depression below sea level which lies up-glacier of it. This basin may have been produced during the surge, and could be analogous to other rapidly produced glacier basins (Boulton, et al., 1983), or it may have been a pre-existing basin. If it were the latter, the surge of Holmstrømbreen may have involved pushing against an ice contact scarp hitherto separated from the glacier by a basin. This interpretation is supported by the existence of lacustrine sediments at the proximal extremity of the pushed sediment mass (Fig. 13). We thus suggest there existed a pre-surge state when ancestors of the present delta and outwash complex were separated from the whole or part of the glacier front by a lake basin. If this was a large basin it would have acted as a sediment trap, and the outwash complex beyond would have been starved of sediment. The recent readvance would have brought a major rejuvenation of the sedimentary system.

The whole push moraine/outwash complex which lies between Holmstrømbreen and the sea must be of relatively recent date. In early Holocene times, sea level in the central Isfjorden region of Spitsbergen lay about 80 m above sealevel in earliest Holocene times and has only approached modern levels in the last 1–2000 years (Salvigsen, 1984). Prior to this recent period, we expect much of the modern outwash surface to have been open fjord. We expect the black fjord muds (lithofacies c) exposed in the western side of the pushed zone beneath fan sediments, to reflect this stage. Radiocarbon dates from the pushed belt (Boulton et al., 1989) suggest that marine conditions were succeeded by terrestrial conditions between 1900 and 1800 BP.

The production of the major supratidal system in recent times, reflecting an important sediment source, is a direct result of the presence of Holmstrømbreen, and as is often found in large ice-contact sediment masses, the outwash sediments are intimately associated with glacier pushing.

If the glacier were to be removed (Fig. 26), a most unusual sediment mass would be left behind, a major coarse sediment accumulation truncated at its proximal extremity, and lacking any obvious source.

What are the conditions under which the glacier would override the ice contact front rather than push it forward? If the resistance of the mass to glacier pushing is large, overriding will occur. The non-linear form of the flow law for ice (Paterson, 1994) and the behaviour of natural ice masses show ice to flow very readily under stresses in excess of 100 kPa, which can be regarded as a yield stress for ice. We assume therefore that the maximum resistance \( R \) at the ice contact face at the initiation of pushing would be 100 kPa. The force \( F \) against unit width of the proximal side of the pushed mass (height \( h \)) will be \( F = hR \). For \( R = 100 \) kPa; \( h = 33 \) m; \( F = 3.3 \times 10^3 \) kN. The total force against the ice contact slope of about 4.5 km width will be about \( 1.5 \times 10^7 \) kN. This was transmitted to the ice contact face as a longitudinal force which was clearly sufficient to push the proglacial sediment mass.

We have demonstrated that as the proximal, ice contact extremity of the moraine was pushed forward and tectonically shortened, the distal extremity also advanced through new extensions of the basal decollement and associated folds. Fig. 27 shows an interpretation of the evolution of the push moraine, in which the distal extremity advances at a greater rate than the proximal extremity, evidenced by the contrast in fold styles. We conclude therefore that the present width of the push moraine is a maximum.

The pushed moraine’s resistance to movement is provided by friction \( (\sigma_o) \) across the basal decollement, which can be estimated using a simple Coulomb law. It is

\[
\sigma = C + \text{N} \tan \phi
\]

\( \sigma \) is cohesion, \( \text{N} \) is the effective normal pressure and \( \phi \) is the angle of internal friction). The effective normal pressure \( (\text{N}) \) at the level of the decollement is given by

\[
\text{N} = df[1 - n (\rho_s - \rho_w) g]
\]

(\( d \) is the depth to the decollement, \( n \) is void ratio, \( \rho_s \) is the density of sediment grains, \( \rho_w \) is the density of water and \( g \) is the gravity). For the silty sand which makes up much of the Holmstrømbreen push moraine, and assuming a water table to be at surface, the increment of pressure per unit depth is about 10 kPa m\(^{-1}\), producing a mean effective pressure along the decollement of 400 kPa. Assuming that sediment cohesion can be taken as zero during sliding and assuming a residual angle of friction of \( \phi = 22^\circ \), the frictional resistance along the basal decollement would have been about 160 kPa. If the water table lay below the surface, or if sediments were frozen, effective pressures and frictional resistance would have been greater.

The average shear stress at the level of the decollement \( \tau_s \) is made up of the stress \( \tau_q \) produced by the longitudinal force generated by glacier pushing and by the
gravitational stress $\tau_h$ from the sloping surface of the push moraine:

$$\tau_h = \frac{F}{L} + (\rho_s - \rho_a) gd \sin \alpha$$

(where $L$ is the length of the decollement, and $\alpha$ is the average slope of the push moraine surface). During the last phase of pushing, when the push moraine had its present form, imposed by pushing at the level of the decollement, for a total glacier generated force of $3.3 \times 10^3$ kN at the ice contact slope, when $L = 900$ m and $\sin \alpha = 0.033$, then

$$\tau_g = 3.7 \text{ kPa}, \quad \tau_s = 1.3 \text{ kPa}$$

giving a total mean shear stress at the decollement of 5 kPa, compared with a mean frictional resistance of about 160 kPa, a safety factor of 32. Even if the decollement had lain immediately beneath the floor of the Western Gorge, the safety factor would still have been 18.4, although we have argued above that the calculated decollement depth is likely to be a minimum.

It is clear that the only circumstances in which large-scale movement of a pushed mass of this magnitude can occur is one in which excess pore water pressures are generated. The average effective normal pressure on the decollement would need to be less than 9.5 kPa to permit failure. As the average bulk density of the sediments in the pushed mass is about $2 \text{ Mg m}^{-3}$, giving a load of about 800 kPa at the base of the pushed mass of average thickness 40 m, the water pressure would need to be 790.5 kPa, implying a water head of 79 m above the base of the pushed mass and an artesian head of about 40 m in order to maintain movement of the pushed mass along the decollement. An initial decollement depth of about 25 m (Fig. 28) would have required an artesian head of about 25 m. As the thickness of the pushed mass above the decollement has increased during the pushing episode, and supposing that there has not been a dramatic increase in the longitudinal stress imposed by the glacier, there must have been a progressive increase in water overpressures. An estimation of this increase is shown in Fig. 28, assuming that the conditions for failure were achieved at all points along the decollement. Progressive deformation was associated with an increase of water pressure head and head gradient as the deforming mass shortened and thickened. It is clear that the production of large-scale push moraines is as much dependent on glacial enhancement of water pressures as in provision of a tractive force.

It is not difficult to understand how water pressures are glacially enhanced by undrained loading. As the glacier advances over a sediment, the additional ice load, or the increase in horizontal stress, is initially born entirely by interstitial water, so that water pressures equal ice pressures (Boulton and Dobbie, 1993). This sets up a pressure gradient between the loaded and unloaded proglacial zones and water drains from the system, tending to lower water pressures, at a rate governed by the transmissivity of the bed. Rates of drainage can however be slow compared with rates of loading (particularly so in a surge) such that the increase in water pressures inferred in Fig. 28 is much as would be expected.

We have discussed above the static balance of forces for the specific geometry of the Holmstrømbreen push moraine, but how is the geometry produced in the first place? The difficulty lies in understanding how a force applied at the proximal extremity of the pushed mass could be transmitted for about 1 km through a thin plate of soft sediment between 20 and 40 m in thickness (aspect ration 1/40–1/20) without being entirely absorbed by buckling in the sediments in a relatively narrow zone near to the ice margin. The geometry of the system is analogous to that of the retaining wall problem for which Rankine obtained the analytic solution that the plane of shear failure at the base of a wall retaining a horizontal cohesionless soil mass would have an angle of $45^\circ - \phi/2$, an angle of $34^\circ$ for $\phi = 22^\circ$. This agrees with typical angles of shear planes in the outer parts of the pushed belt where they are unlikely to have been rotated since formation. Thus, an ice contact slope extending 20 m below the sediment surface would push along a sliding plane which would come to surface no more than 30 m from the advancing ice front. This assumes that the angle of friction between the ice and sediment is less than $\phi/3$. Pushing of such a mass would naturally increase the slope of the proglacial surface, such that further sliding
planes would adopt a lower inclination. However, the slope of the push moraine is far too small to be able to account for the creation of a primary decollement near to the horizontal.

A very obvious solution, suggested by Rutten (1960) and Mackay and Mathews (1964), is that the upper part of the pushed sediment mass is cemented by permafrost, which gives the sediment short term stiffness, thereby enhancing its capacity to transmit stress and minimise buckling, concentrates stress at its lower surface and also acts as an aquiclue or aquitard so as to permit high glacially driven pore fluid pressures to build up beneath it. In this model, the push moraine acts as a relatively rigid mass, Coulomb failure, facilitated by high pore water pressures, occurs at the base of the permafrost and determines the location of the decollement. The key conclusion however, is that, given the small size of the disturbing forces in a large push moraine, it is the glacial control of pore water pressure which is the key attribute of the system.

We expect Holmstrømmbreen to be similar in its basal thermal regime to other glaciers of Spitsbergen in being of a type in which basal melting occurs (Schytt, 1969; Drewry and Liestøl, 1985; Hagen and Saetran, 1991; Hagen et al., 1993). This meltwater is able to recharge subglacial aquifers (Boulton et al., 1995). Meltwater reaching the bed of the glacier will be driven outwards because of the glacial pressure gradient, and as the Carboniferous and Devonian limestones, evaporites and sandstones over which Holmstrømmbreen flows have a relatively high intrinsic permeability, significant fluxes of water will be discharged by groundwater flow. We have used the model described by Boulton et al. (1995) to simulate the head distribution produced by a water flux from beneath a glacier into a proglacial area with permafrost and without permafrost (Fig. 29). This shows clearly that for a hydrogeological setting such as that at Holmstrømmbreen, the impact of permafrost is to extend the zone of pore fluid overpressures much further beyond the glacier than in the non-permafrost case. An initial time dependent simulation associated with prescribed forward movement of the glacier, which computes the effect of undrained loading, shows a broader zone of excess water pressures in the

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**Fig. 28.** Schematic longitudinal section showing the position of the glacier forefront (black) immediately prior to the pushing event (A-A) and immediately after (B-B) and, at the same periods, the form of the water pressure head (blue) at the same periods (A_w and B_w), the surface of the pre-tectonic (red) and post-tectonic (green) fans (A_f and B_f) and the calculated form of the decollement surface along which the pushed mass slid. The head increase from A_w to B_w is a product of undrained loading of sediments during the rapid advance of the glacier front.
Fig. 29. Model of two-dimensional water flow beyond a glacier in which (a) there is proglacial permafrost and (b) no proglacial permafrost. Subglacial discharge is prescribed. There is no recharge at the proglacial surface, and any surface discharge is assumed to be removed by surface drainage. Flow occurs in an unfrozen aquifer above a planar impermeable surface. Water overpressures occur in a narrow proglacial zone in the absence of permafrost and a broad zone in the presence of proglacial permafrost.

non-permafrost case, but still far smaller than required to account for glacial failure on a kilometre scale.

The head distribution will also have an important influence on the outer extent of the push moraine. It is most likely to have been determined by the outward extent of diffusion of high excess water pressures at the time the surge ended. However, Boulton and Caban (1995) have suggested that the generation of high water overpressures beneath proglacial permafrost will tend to create tensile forces in the permafrost and thereby produce hydrofracturing. Such hydrofracturing during push moraine development could determine the extent of the push moraine by drawing down water pressures and inhibiting failure in sediments. The existence of patches of surface ice beyond the point at which the Red River emerges from its gorge in the push moraine could be a modern zone of groundwater discharge at which supercooling of rising groundwater (Lawson et al., 1995) produces icing of the surface.

Van der Wateren (1995) has used data from Boulton, van der Meer et al. (1989) to analyse the formation of the Holmstrøm-breen push moraine. He appears to assume that the push moraine sediments behave as a Newtonian viscous fluid (although the viscosities he assumes are clearly much too large) which creates a necessary relationship between ice surface profile and push moraine form, that push moraine behaviour is comparable to the assumed behaviour of accretionary wedges (Davis et al., 1983) being in a critical state of shear failure throughout, and thus that gravitational body forces in the moraine play a key role in its deformation and emplacement.

Experiments on the rheological behaviour of soils and sediments show that a Coulomb frictional threshold must first be overcome before viscous behaviour develops, and thus that a Bingham fluid rheology is most appropriate when considering the initiation of movement, even though a viscous approximation may be appropriate for post-failure modelling purposes. As the disturbing forces are so small in an extensive push moraine, it is necessary to develop very high pore fluid pressures, and as we do not see how such pressures can develop in Van der Wateren’s model, we do not understand how the threshold of pervasive Coulomb failure can be achieved. Nor do we believe that the structures in the push moraine reflect pervasive deformation in the critical state. As we show above, given the small size of the disturbing force, the glacially-powered water pressure regime is the key control on push moraine development. Any push moraine theory must therefore show how an appropriate pressure regime can develop.

Our conclusions lead us to suggest a general hydraulic setting for the Holmstrøm-breen push moraine. It has long been accepted (e.g. Weertman, 1969) that detachment of a glacier from its bed due to increasing water pressures is the cause of surging behaviour in glaciers. Surging is common amongst Spitsbergen glaciers, which has attributed to their sub-polar thermal regime, characterised by an internal zone of melting and an outer zone of freezing, conducive to the build up of large subglacial water pressures. The outward gradient of ice pressure will tend to drive subglacial meltwater towards the glacier margin, and some of this can be expected to flow as groundwater, with a head gradient towards the terminus (Boulton et al., 1995). A permafrost horizon beyond the glacier will confine groundwater flow, but permafrost will terminate near to the coastline, because of the large thermal capacity of large water bodies. The groundwater head in the coastal zone will therefore be atmospheric, but there will be a head gradient beneath proglacial permafrost because of the groundwater which must be discharged beneath it. As a consequence, the groundwater head beneath proglacial permafrost will be atmospheric, effective pressures will be low, and the resistance to glacier pushing beneath a frozen sediments mass will be minimised.
6.3. Implications for large modern and Pleistocene push moraines

6.3.1. Modern glaciers
Modern glaciers produce several types of push moraine in which freezing does not appear necessarily to be involved:

- single-ridge push moraines a metre or so in width and a metre or so in height as a result of winter readvances (e.g. Boulton, 1986)
- single-ridge push moraines up to several metres in width and height as a result of more general, climate-driven, though small, readvances (e.g. Krüger, 1985)
- multi-ridged and multi-fold push moraines in which significant deformation has been transmitted horizontally for the order of 50 m. beyond the glacier and through a thickness of about 10 m of sediment (aspect ratio 1:5) (e.g. Croot, 1988).

In all these cases the distance from the glacier front across which failure occurs is consistent with a simple Rankine-model of pushing without need for special hydraulic circumstances.

In Spitsbergen and elsewhere, push moraines analogous in form to those of Holmstrømmbreen are common. They differ from those listed above in their tightly sub-parallel fold-ridges, their great width, of several hundred metres to over a kilometre, and the very small aspect ratios of thickness to lengths, often between 1/20 and 1/100. Their distribution has been described by Gripp (1929) and Croot (1988). Most are located on coastal plains. In these areas, proglacial permafrost terminates at the coastline, and permafrost is likely to be thin because of the history of strong recent land uplift in Spitsbergen, where most coastal plains have been recently sub-marine and permafrost has only begun to develop since emergence. We suggest therefore that the ideal circumstances which favour the production of large and broad push moraines are

- the glacier margin is initially stationary so as to produce a well-defined ice contact face against a thick sediment sequence so that in a subsequent, possibly surging, advance it is able to transmit a large stress to proglacial sediments;
- proglacial sediments are frozen so that a stress can be transmitted through a considerable distance by relatively stiff sediments;
- the glacier is melting basally so that groundwater, forced to flow beneath impermeable permafrost, is able to develop large porewater pressures;
- the permafrost is thin so that effective pressures and the frictional resistance at the base of the permafrost are small;
- the permafrost is broken distally so that its distal extremity can advance relatively freely.

Hambrey and Huddart (1995) and Huddart and Hambrey (1996) have described ‘thrust moraines’ produced at or near the margins of two adjacent valley glaciers (Uversbreen and Comfortlessbreen) in Spitsbergen. The structural style of these moraines appears to be quite different from that of the Holmstrømmbreen push moraine, being dominated by a series of low-angle parallel thrusts dipping up-glacier with little sign of large-scale folding. The overall structure of the moraine is less clear than at Holmstrømmbreen and in the case of Uversbreen, the evidence presented of thrusting is equivocal.

The relative absence of buckling in thrust nappes is most likely to derive from stiffness within the nappes and almost friction-free gliding beneath them. A result most likely to be achieved where thrusting exploits very soft sediment horizons, such as the widespread glaciomarine sediments which underlie many coastal plains in Spitsbergen (Sollid et al., 1994; Huddart and Hambrey, 1996) which may have played this role. We would conclude therefore that emplacement of very wide push moraines must be facilitated by high interstitial fluid pressures and suggest a continuum classification of push moraines:

- thrust dominated push moraines, which reflect very low or friction-free sliding and may show great proximal-distal width;
- fold-thrust-dominated push moraines, which reflect greater friction along decollement planes but which may also show considerable width;
- fold-dominated push moraines, which reflect relatively high internal and basal friction, in which the folds may be rooted reflecting very small displacements and with very small proximal-distal widths, or lie above a basal decollement, with greater displacement and greater widths.

6.3.2. Pleistocene push moraines
There are many Pleistocene push moraines which are similar in overall structure to that of Holmstrømmbreen, consisting of a series of tightly parallel fold/fault structures transverse to glacier movement, and occurring as thin but longitudinally extensive nappes overlying a well-defined decollement. Kupsch (1962) described such systems from western Canada which were up to 150 m in thickness and suggested, following Rutten (1960), that they could only have formed under permafrost conditions for similar reasons to those suggested here. Eismann (1987) reviewed similar structures from Germany, describing Saalian push moraines from the Elbe and Spree valleys which have basal decollement planes lying some 50–75 m below the original surface at the time of initiation of pushing. Van der Wateren (1987) described thin plates of ice pushed sediment with aspect ratios of length to thickness of 20–40/1 from the Saalian Dammer Berge push moraines in western Germany.
We suggest that these push moraines have developed by pushing of frozen sediment nappes along a basal decollement coincident with the depth of permafrost and that frictional resistance was low along the decollement because of high pore fluid pressures developed under an hydraulic regime similar to that shown in Fig. 29. In places such as the Dammer Berge, soft clay units may deflect the decollement away from the permafrost boundary, although van der Wateren (1992) believed that permafrost was not necessary in explaining the Dammer Berge push moraines.

If this conclusion is correct, it implies that the depth of the decollement surface at the beginning of the pushing episode is an important guide to the depth of contemporary permafrost. It then becomes particularly important to establish the time of pushing. If a pushing episode occurs during glacier advance or at the glacial maximum, it gives evidence of permafrost during build up phases. If pushing takes place during readvances in periods of overall decay, permafrost depth will depend upon whether freezing occurred beneath the ice sheet, the period of proglacial exposure of a surface prior to readvance and would therefore be the result of a complex local temperature history rather than the reflection of a regional pattern of permafrost thickness.

Acknowledgements

We gratefully acknowledge the assistance and companionship in the field of Ilona Castel, Irene Quinn, Peter Riezebos, Mary Thornton and Dick van der Wateren. Dick van der Wateren is thanked for the use of Figure 21. The project was supported by ZWO (The Netherlands Organisation for Pure Scientific Research), the Natural Environment Research Council (UK), the University of Amsterdam and The Geological Survey of the Netherlands. Mike Hambrey made valuable comments as a referee.

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