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Glacio-tectonic structures: a mesoscale model of thin-skinned thrust sheets?

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Abstract—A series of seven groups of push-ridges ranging from 7 to 40 m in height, 50 to 280 m in length, and occupying a total width of more than 2 km, mark the marginal zone of the A.D. 1890 maximum of Eyjabakkajökull, an outlet glacier of the Vatnajökull ice cap, Iceland. The internal structure of one ridge complex comprises two distinct elements: a proglacial part which has been subject to compressional stresses, resulting in the development of imbricate thrust sheets; and a subglacial part which comprises low-angle normal fault structures. The two sub-systems appear to be linked via a floor thrust and to have evolved together as the glacier reached the limit of its rapid advance in A.D. 1890.

INTRODUCTION

THIS PAPER arises from a piece of fieldwork primarily concerned with glacial geology and landform genesis. The results of this field investigation were so clear and precise, however, that comparisons with features of interest to structural geologists were inevitably drawn. The aim of this paper is to draw attention to the fact that the products of macroscale fold and thrust belts are being replicated at a more easily studied scale, by processes of glacio-tectonic deformation. Similar analogies between glaciotectonic structures of Pleistocene age and hard-rock tectonic structures in fold and thrust belts have been drawn by Berthelsen (1979) and Pedersen (1986).

The cross-application of knowledge between glaciologists and structural geologists is not new (e.g. Berthelsen 1979, p. 260). Furthermore, a number of recent papers have 'borrowed' theories and observations created and made by glaciologists and applied them to structural geology (Coward & Kim 1981, Ramberg 1981). In the present study comparisons are made between a glacier, advancing into a sheet of unconsolidated proglacial sediments, and thin-skinned thrust tectonics. Mechanistically, the explanation put forward at the end of the paper is most similar to a 'push-from-the-rear' model (Hubbert & Rubey 1959, Elliott 1976, Chapple 1978, Price & McClay 1981), rather than a gravity spreading or gliding model (Berthelsen 1979, Pedersen 1986), since the relatively rapid advance of the glacier in question appears to have provided a mechanical 'jolt' or push to the rear end of the static sedimentary pile.

Landforms very similar to those described in this paper have been described as occurring at the snouts of a number of glaciers in the Arctic (Gripp 1929, Kälin 1971). Indeed it was the early work by Gripp which drew attention to the Jura-like appearance of glacier-pushed-ridges ("stauchendmorane" in Gripp's paper). However, no data have yet been published on the internal structure of such contemporary features, although a considerable volume of literature has been published on the internal structure of Pleistocene glacio-tectonic features in Northern Europe (see Houmark-Nielsen & Berthelsen 1981, and Ehlers 1983, for extensive bibliographies). Consequently comparisons with large-scale geologic structures have remained at the qualitative level. The term 'push-moraine' has been used to describe ridges developed proglacially, predominantly by a pushing process. The term includes not only such features as those described as Jura-like ridges or ridge complexes by Gripp and Kälin, but also much more simple, smaller ridge forms such as that found at the snout of the Frias Glacier, Argentina (Rabassa *et al.* 1979). The genetic classification and subclassification of glacio-tectonic forms is currently being reviewed by a Working Group of the International Quaternary Association Commission on the Genesis and Lithology of Glacial Deposits. For the present, the term 'push-moraine' is an adequate genetic definition of the features considered in this paper, but the term 'push-moraine' and 'push-ridge complex' are also used synonymously.

The following sections deal with the field description of the site and the morphometry of the push-ridge complexes which occur at the glacier limit achieved in A.D. 1890. The results of the detailed structural mapping of three such push-ridge complexes is followed by the description of the internal structure of one ridge complex. An attempt is then made to restore the proglacial section using line balancing to produce data on shortening, extension and bulk strain. The field and derived data are then used to compare the characteristics of these mesoscale structures with those produced in large-scale orogenic belts and those of ideal models suggested by Elliott (1976) and Chapple (1978) and reviewed by Siddans (1984).

CHARACTERISTICS OF THE GLACIER AND FORELAND

The ridge complexes described and analysed below have been formed by the advance of Eyjabakkajökull, a north-eastern outlet glacier of the Vatnajökull ice cap, Iceland, into pre-existing valley floor sediments (Fig. 1).

Tephrochronological evidence demonstrates that the valley was deglaciated in the area of the ridge complexes until the 19th century (Thorarinsson personal communication 1979). Since that time rapid advances (surges) of the glacier have occurred at regular intervals [1890, 1931, 1938 (partial), 1972] to a variety of positions in the valley floor (Thorarinsson 1938, 1964, 1969, Williams 1976, Croot 1978, Sharp 1982). The surges have normally lasted several months and involved advances of the ice margin by as much as 3 km. The extent of the 1890 advance was not exceeded by the 1931, 1938 or 1972 surges, and consequently the features described below have remained intact since 1890.

The current glacier foreland is divisible into two major morphological units separated by the push-ridges under discussion (Fig. 1). The recently deglaciated area between the push-ridges and current glacier margin comprises a suite of landforms and sedimentary units that are products of the stagnation of the ice following the 1890, 1931, 1938 and 1972 surge advances (Croot 1978, Sharp 1982). Outside the limits of the 1890 advance, the valley is generally broad and flat, comprising a range of fluvioglacial stratigraphic units with an extensive, well-developed herbaceous cover (Fig. 2). Standing above this vegetated sandur are low (10 m or less) isolated basaltic hills with little or no vegetation. The meltwater streams from Eyjabakkajökull and isolated perennial snow patches meander and braid across the sandur and eventually coverge into a single channel at the north-eastern neck of the valley before plunging down a gorge into the deeply entrenched valley beyond.

PUSH-RIDGE MORPHOLOGY

Morphological mapping (at a scale of 1:2500) and topographic surveys of the push-ridges were carried out in the field using enlargements of excellent air photographs (Landmaelingar Islands 1967). The results of these surveys are produced in Figs. 3 and 4.

Although describing a broad arc across the valley floor representing the ice limit achieved in 1980, the push-ridges can be subdivided into seven groups or complexes, each one representing a discrete unit of deformation (Figs. 3 and 4). These complexes are in places separated by meltwater channels that have exploited the topographical low areas at the junctions of adjacent groups of ridges. The western part of the valley floor is devoid of ridges, being an area of very active meltwater erosion and accretion. There may have been push-ridges in this area immediately after the 1890 surge, but any evidence of such occurrence is now missing. Elsewhere, the forms of the ridges do not appear to have changed at all since their evolution in 1890. Consequently morphological and structural details are excellently preserved. The single exception is ridge complex (a) (Figs. 1 and 4), which is adjacent to the active sandur described above. These ridges have been and are being eroded by meltwater activity and their original morphology and dimensions are impossible to ascertain. Despite this erosion, however, they remain the largest of all the ridge complexes.

Each ridge complex is lobate or crescentic in plan form, and slopes from the proglacial margin (usually ice-contact) to the distal limit (Fig. 4). However, the size and shape of each group vary across the area [compare for example ridge groups (a) (profiles 1 and 2), (e) (profiles 10 and 11) and (f) (profile 12)]. In all cases the former position of the ice margin is very clearly marked on the air photographs (change of tone) and on the ground (by the presence of a thin layer of meltout till). The glacier-proximal area of all the ridge groups is a topographic low, either occupied by a shallow lake [ridge groups (c), (d) and (e)] or filled with post-deformation outwash gravel [ridge groups (a), (f) and (g)].

Although the dimensions of ridge groups (b)–(g) vary (Fig. 5 and Table 1), each group is similar in that it comprises individual ridge forms aligned parallel (or nearly so) to the former ice margin. These individual ridges can often be traced across the lobate complex, although some rise and are pinched out again within the lateral margin. Invariably, the ridge nearest the former ice margin is the highest, and crest heights fall successively towards the glacier-distal limit. The larger ridges [particularly of complexes (c), (d) and (e)] nearer the former ice margin are asymmetrical with steep slopes facing down-valley and very gentle slopes facing in towards the former glacier margin. The lower, distal ridges of these groups are symmetrical anticlinal forms. Ridges comprising groups (f) and (g) are quite different in form. These ridges are spatially more extensive, but of low amplitude above the sandur level (Fig. 4, profile 12-12′).

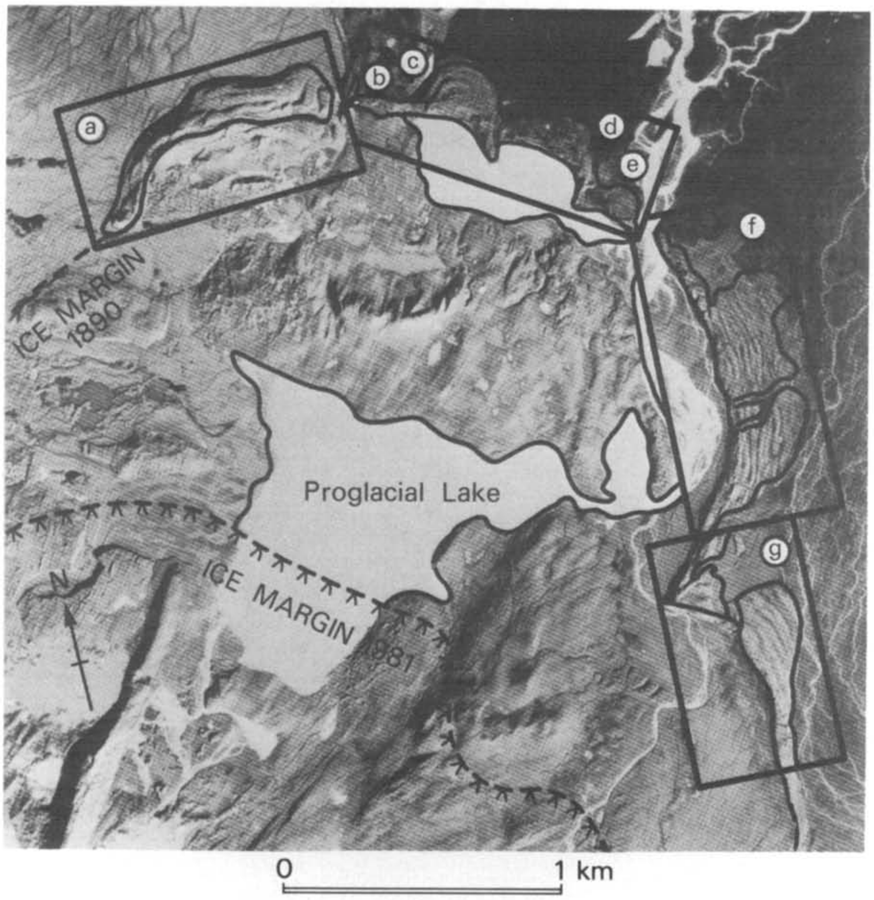
Other landforms are found within the push-ridge complexes. These include deeply incised blind-head gullies or canyons, orthogonal to the trend of the ridge crests, and less frequently topographic flats. Interlobate areas are commonly masked by thin gravel and sand deltas.

It is possible, on the basis of morphology alone, to classify the groups of ridges into: (a) incomplete, but

Table 1. Dimensions of the ridge complexes (see Fig. 5 for definitions)

Ridge complex	H (m)	L (m)	W (m)	θ (°)	Aspect ratio*	Volume† ($\times 10^5$ m^3)	Mass‡ ($\times 10^8$ kg)
a	40	220	360	7	5.5	15.8	26.9–37.9
b	7	50	111	9	7.1	00.19	0.3–0.4
c	15	230	330	3.5	15.3	5.69	9.7–13.6
d	8	200	305	4	25	2.44	4.15–5.9
e	9§	110	200	8	12.2	0.99	1.7–2.4
f	5§	280	527	2	56	3.68	6.3–8.8
g	5§	270	270	2	54	1.82	3.1–4.4

* Aspect ratio = L/H.
† Volume estimated to be $V = ((L/2)\ wh)$ m^3.
‡ Density values probably range from 1700 (silt) to 2400 kg m^{-3} (gravel); min.–max. range of mass values is therefore given.
§ These values are more representative of height than values read from profiles.



Photograph reproduced with permission of Landmaelingar Islands

Fig. 1. Air photograph of part of the proglacial area of Eyjabakkajökull and location map of Iceland. Areas of push-ridges formed adjacent to the 1890 ice margin are labelled (a)–(g) (see text for description). Photograph reproduced with permission of Landmaelingar Islands.

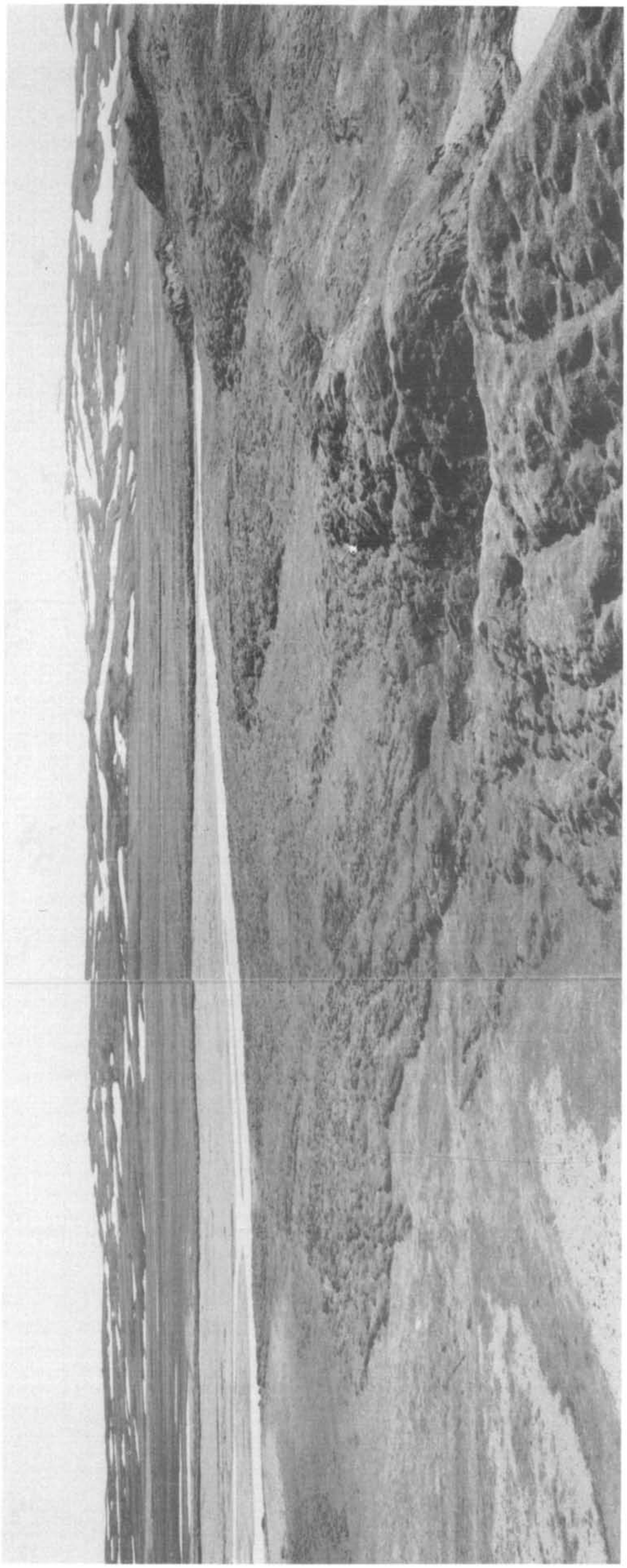


Fig. 2. Panorama of part of push-ridge complexes (c), (d) and (e). Glacier advanced from right to left, terminating part-way up the glacier-proximal slope of the push-ridges. Note the continuity of vegetation cover (pre-1890) from deformed to undeformed sandur.

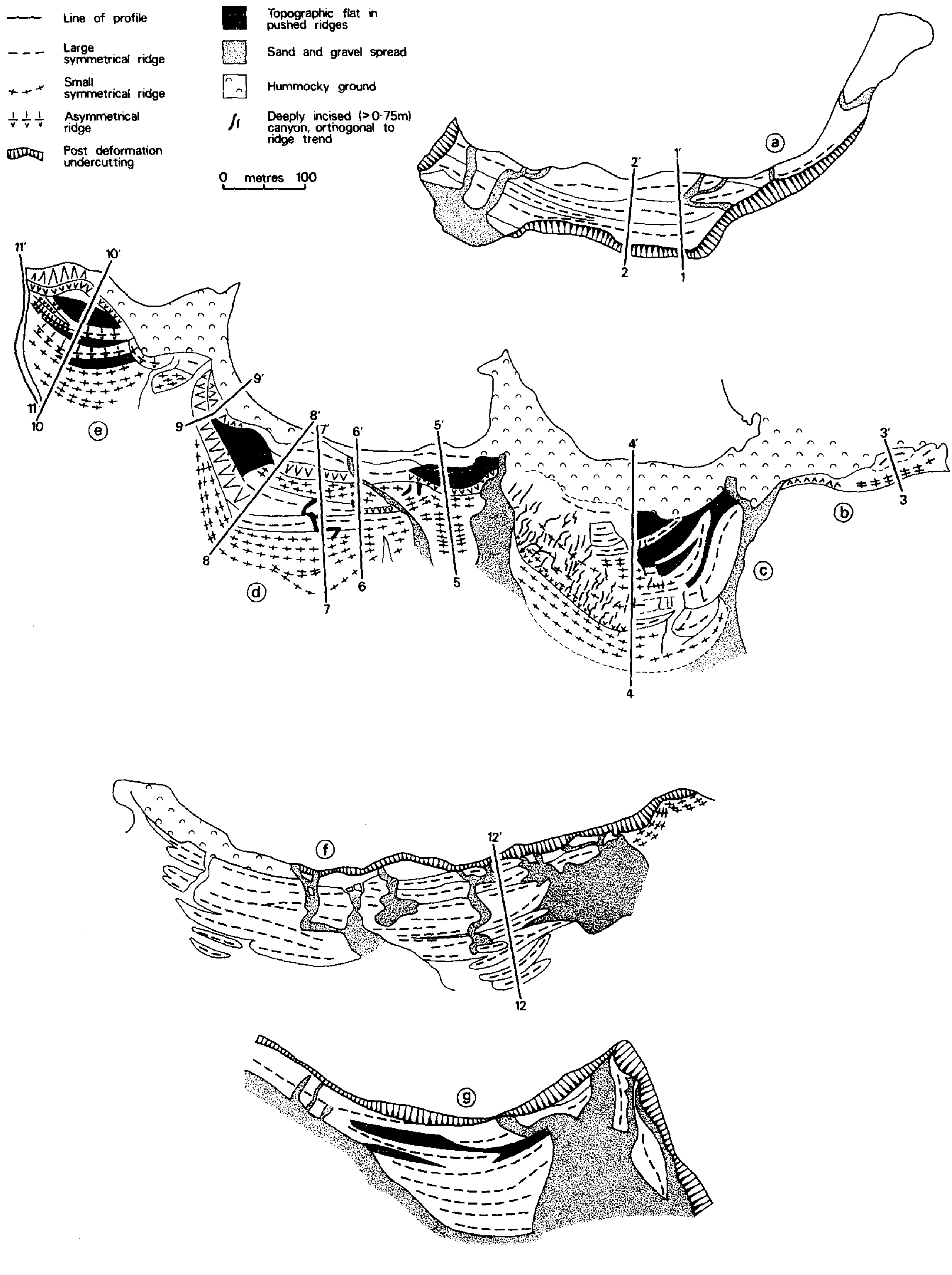


Fig. 3. Morphology of push-ridge complexes (a)–(g) (see text and Fig. 1 for location).

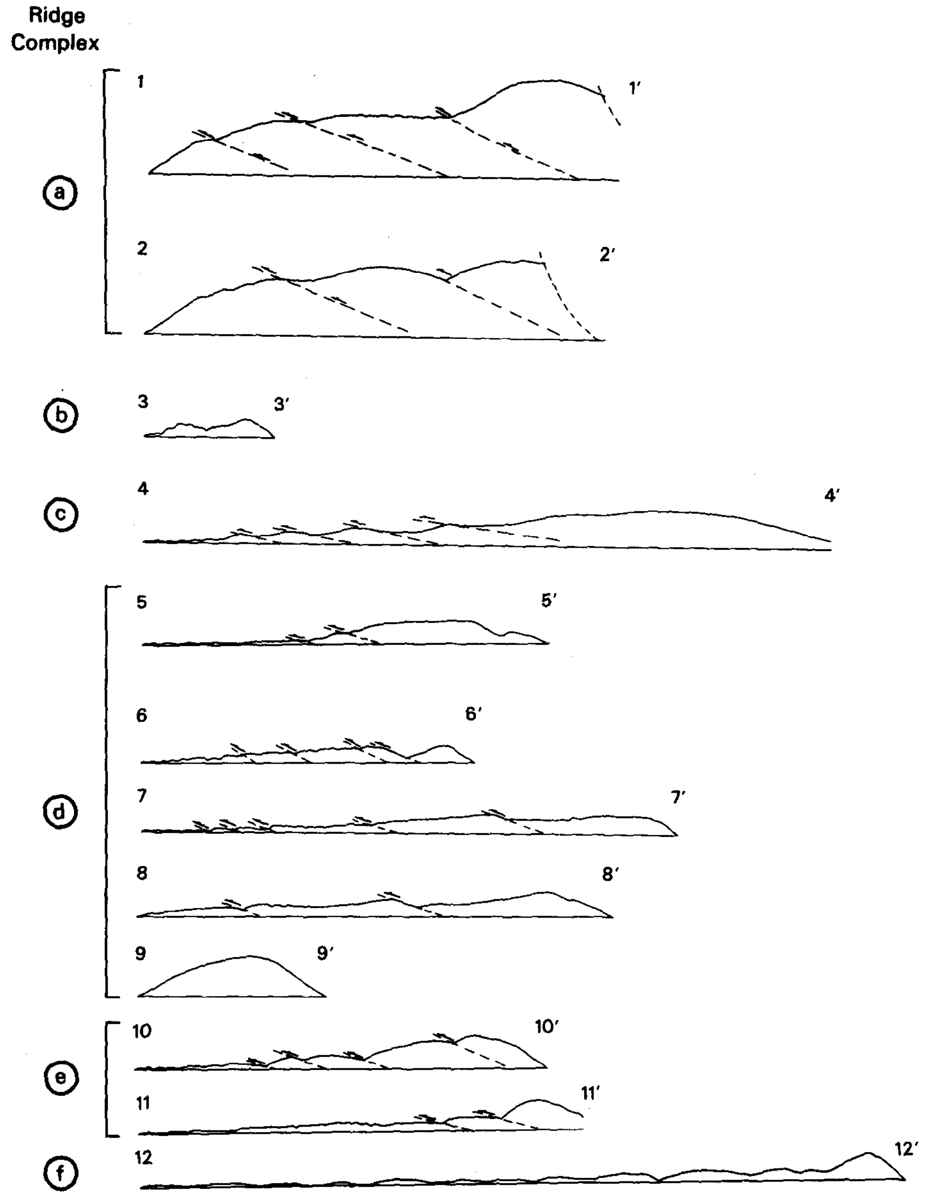


Fig. 4. Profiles 1–1′ to 12–12′, surveyed across ridge complexes (a)–(f) (see Fig. 3 for locations).

bulky features; (c), (d), (e) extensive, complete complex-morphology features; and (f) and (g) (extensive, simple symmetrical anticlinal structures of low amplitude. More detailed analysis of ridge groups (c), (d) and (e) was carried out since a section through ridge complex (e) afforded a great deal of vital data regarding the internal structure of this type of ridge complex. This included stratigraphic analysis and detailed plotting of the large section exposed at the eastern end of ridge group (e), along the line of profile 11–11′ (Fig. 3).

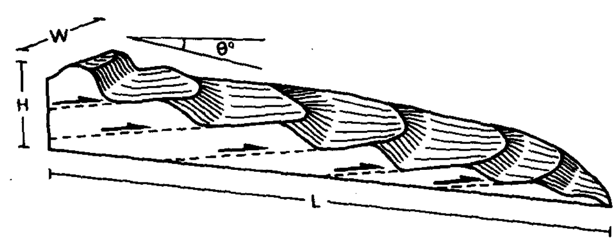


Fig. 5. Idealized model of push-ridge complex, showing definitions of dimensions given in Table 1.

STRATIGRAPHIC ANALYSIS

Riverbank sections along profile 11–11′ demonstrate the stratigraphic continuity from the undisturbed sandur into the ridges. This is supported by the continuity of the turf and near-surface stratigraphy elsewhere in the ridges and sandur, proven by shallow pits dug to a depth of 50 cm.

Sandur sedimentation produces lateral variability in sediment texture over short distances, and this could have given rise to problems in the interpretation of structures. Fortunately, however, the area has received uniformly thick deposits of tephra from volcanic eruptions during the period of sandur accumulation. These tephra layers provide excellent marker horizons and occur 60–65, 68–70 and 102–105 cm below the topographic surface in an undisturbed section. The remainder of the strata comprises sands, silts and gravels. Some of these sediments are loose-textured, whilst others are compact and impermeable. Compact blue silts have permeability values (k) of 0.0047 cm^{-1}, whilst the open-work gravels have permeability values (k) of 0.172 cm^{-1}. Prior to the 1890 advance, the area had developed an extensive vegetation cover, with a substantial depth of turf, organic soil and peat in some places. This has been incorporated in the push-ridges, and comprises the bulk of the sedimentary mélange which forms the core of the largest ridge at the former ice margin (see Fig. 9).

STRUCTURES

Structural mapping

Well-defined thrusts and thrust sheets in the riverbank section along profile 11–11′ are expressed at the surface. Further away from the former ice margin the smaller symmetrical ridges which are similarly truncated by the meltwater channel, exposing the major section along profile 11–11′, are seen to be simple anticlines or overturned anticlines. In this way a straightforward relationship between topography and underlying structures could be established, and the structural mapping of the remainder of ridge complexes (c), (d) and (e) completed (Fig. 6). The bulk of these ridge complexes comprise imbricately stacked thrust sheets, the structure of which determines the surface topography.

Description of main section in ridge complex (e)

Fluvioglacial incision through the proglacial ridges has revealed the internal structure of some of them. One such section exposed in the western end of ridge complex (e) was cleaned by spade and trowel. The technique adopted for recording the structural data was similar to that used by Cooper *et al.* (1983). The section was photographed using a Polaroid camera, from distances of 1–3 m. The individual slips from the Polaroid camera were then used to construct a large-scale photo-mosaic onto which the details of section were recorded.

These details were subsequently transferred from the less accurate Polaroid prints to a reference section compiled from surveying and high-resolution photographs taken using a tripod-mounted camera 60 m from the face. The end-product of this process is Figs. 7, 8 and 9.

For ease of description the overall section is divided into two elements characterized by contrasting styles of deformation: firstly, the section beneath the former (1890) ice margin ('subglacial', Fig. 8), and secondly that portion which lay down-valley from the former 1890 ice margin ('proglacial', Fig. 9). Only the major marker horizons and tectonic features have been illustrated in the figures for the sake of clarity. The terminology used to describe the section is broadly in line with that defined by Butler (1982).

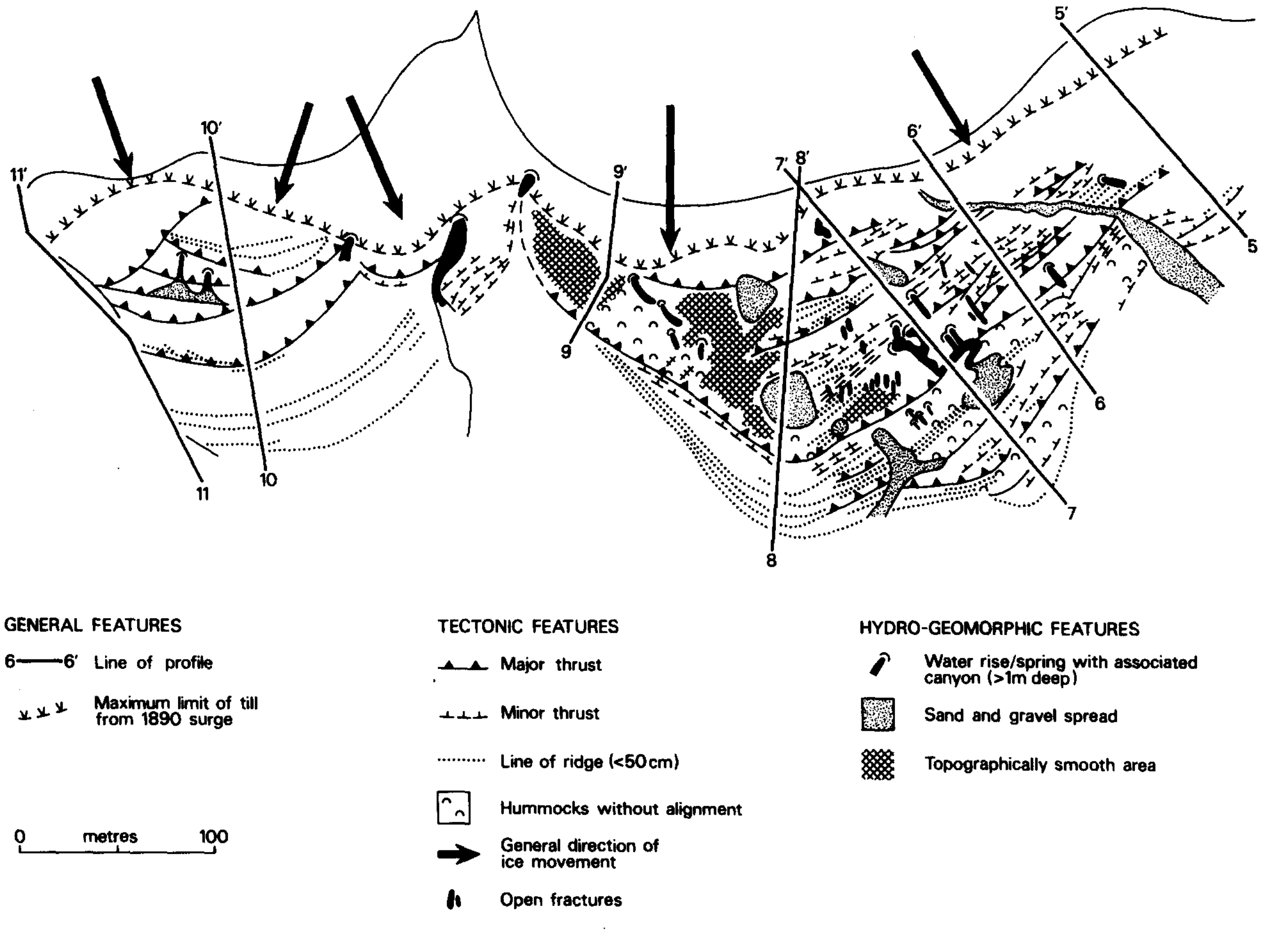


Fig. 6. Tectonic and hydro-geomorphic features of ridge-complexes (d) and (e). Lines of profiles 5–5′ to 11–11′. Detailed structural analysis was made along the line of profile 11–11′.

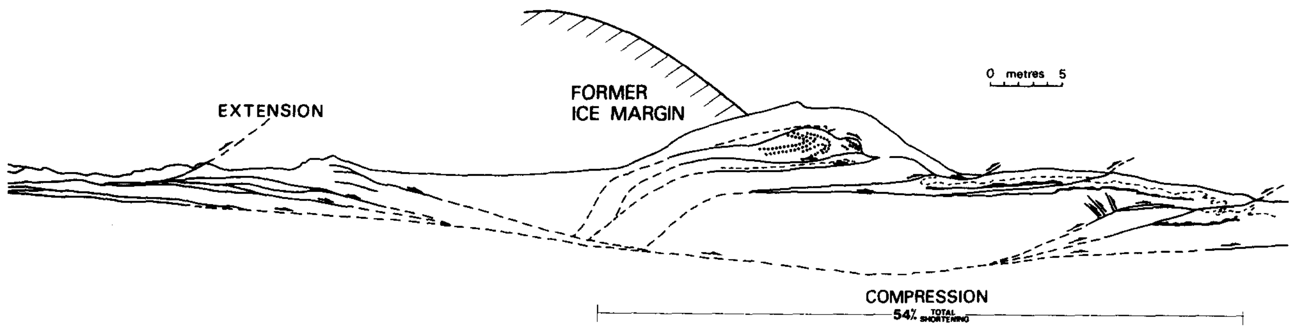


Fig. 7. Principal components of the tectonic system comprising ridge complex (e), divisible into two main zones; proglacial compression and subglacial 'extension', separated by a zone of transition.

The main part of the 'subglacial' section is characterized by a number of low-angle faults dipping in the direction of ice movement (called transport direction in Fig. 8). A marker horizon of tephra within the sand and the gravel that comprise most of this section enabled the amount of displacement along each fault to be measured. The down-valley dip of the faults increases in the direction of transport from 4 to 17°. The direction of slip is invariably normal (i.e. upper elements have moved to the right of lower ones in the diagram). The amount of slip varies from 0.5 m to just less than 6 m with no apparent systematic variation. The arrangement of the fault set implies a convergence of faults at a depth of 4–5 m some 10 m within the former ice margin. Although the lower-most part of the section is obscured, the gravel below the lowest visible fault appears to be compacted, but undisturbed stratigraphically. It is suggested therefore that the lowermost, visible fault is the basal décollement.

In addition to this set of faults dipping down-valley, there is an upper horizontal roof thrust that has carried horizontally bedded gravel and a mixture of glacial till and gravel down-valley. This overthrust unit is itself cut by an upward curving thrust, which converges with the roof thrust of the extension fault system at a shallow angle.

Between these well-defined 'subglacial' structures and those which developed beyond the former ice margin is a 'transition zone' of substantial deformation, that forms the largest ridge of the complex. This ridge apparently formed partly beneath the ice margin and partly in front of the glacier, but structurally the style of deformation is most similar to the down-valley section (i.e. compression) and is included in Fig. 9. The ridge comprises at least three sheets of highly disturbed brown sandy silt, compact blue silt and soil horizons, bounded top and bottom by well-defined thrusts. The original stratigraphy, as indicated by tephra layers, appears to have been identical to that found immediately down-valley, but the deformation within each thrust sheet has been so extreme that it is impossible to demonstrate it diagrammatically.

Most of the 'proglacial' section, illustrated in Fig. 9, displays quite different characteristics from the 'subglacial' section described above. The whole section is divisible into a number of imbricate thrust sheets, dipping opposite to the direction of transport (up-valley, up-glacier). The majority of thrusts are concave-upward in their upper parts, and convex-upward further down, a characteristic common to imbricate thrust sheets in major orogenic thrust belts (Boyer & Elliott 1982). The thrusts are all of low dip (between 3 and 20°), the majority cutting out at the surface.

The amount of displacement along three of the major proglacial thrusts (1, 2, 3, Fig. 9) can be assessed using the well-defined tephra horizons. Displacement along

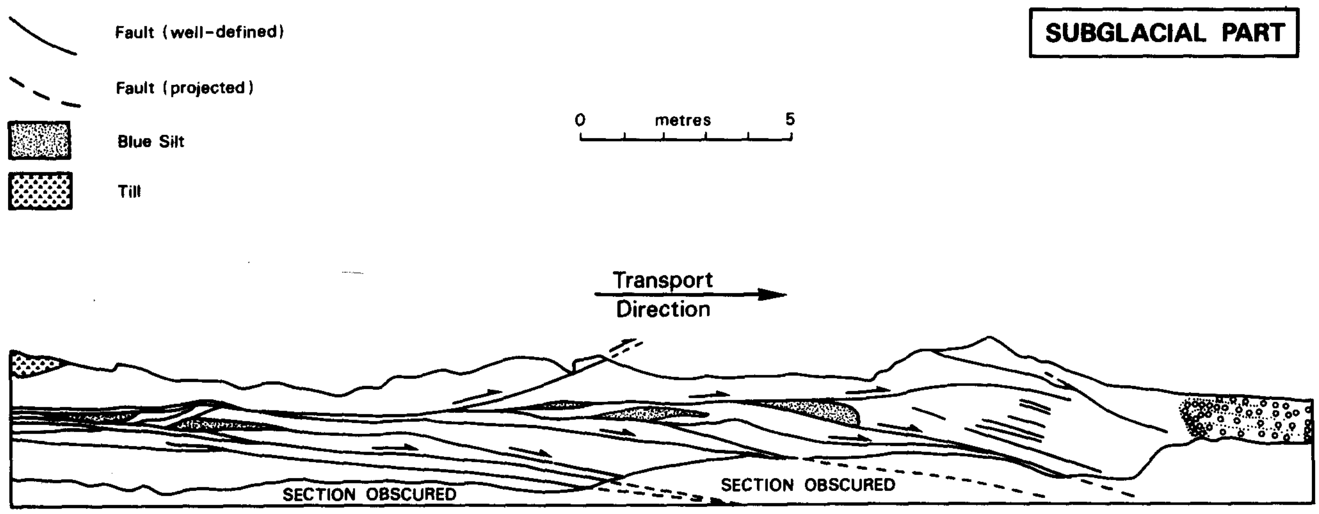


Fig. 8. Structure of part of push-ridge (c) overridden by ice in 1890 (subglacial).

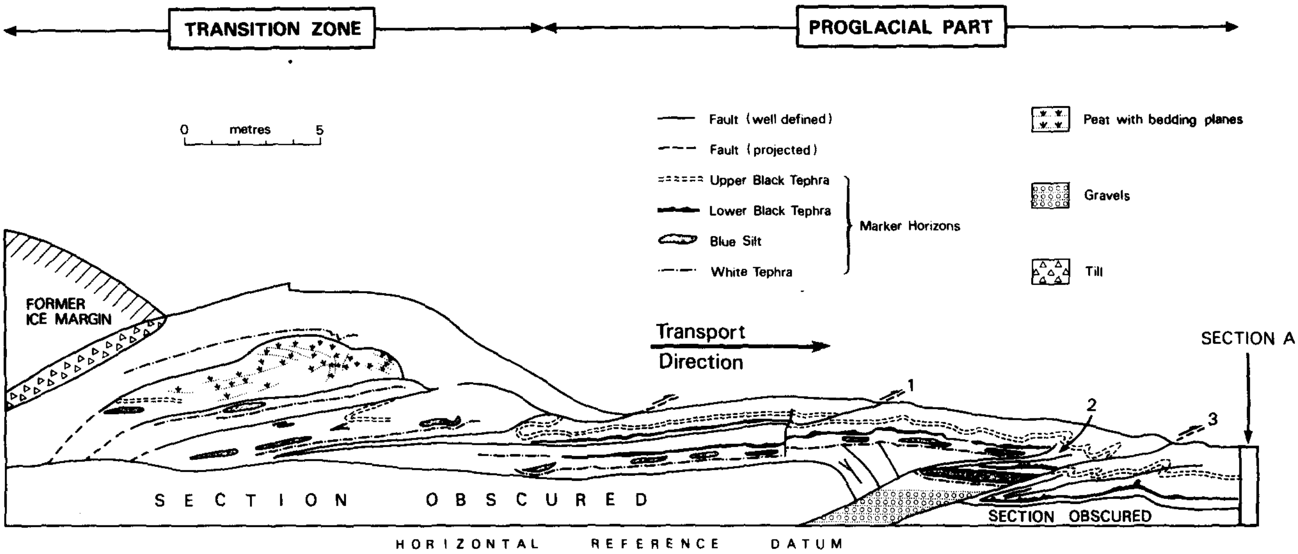


Fig. 9. Stratigraphy and structure of part of push-ridge (c) which remained proglacial during the rapid advance (surge) of 1890.

these individual thrusts which cut the surface decreases down-section from 8.4 to 2.8 m and finally to 2.3 m. Some thrusts die out upwards into fold structures. This is particularly evident furthest from the former ice-margin position, adjacent to the location of section A in Fig. 9. Folding is a feature mainly associated with the upper horizons, especially in the area of blind thrusts, in the toe of the movement. Some normal faults are developed in the strata, displacing the major thrusts by a few centimetres (between thrusts 1 and 2, Fig. 9).

RESTORATION

The detailed measurements of the direction and amount of faulting, thrusting and folding using two identifiable marker horizons facilitated the construction of a palinspastic section for the proglacial element (Fig. 10), using line balancing (Dahlstrom 1969, Hossack 1979). There is no apparent thickening or thinning of beds within each proglacial horse, and consequently bulk strain is solely attributable to shortening by thrusting and folding, as found by Cooper *et al.* (1983).

Restoration of both the 'transition zone' and 'subglacial part' of the overall section is hindered by the lack of more than one marker horizon and the degree of deformation within each horse, particularly in the transition zone (Fig. 9).

However, several features of the transition zone and subglacial part are significant to the restoration procedure: the orientation and dip of thrust planes within the

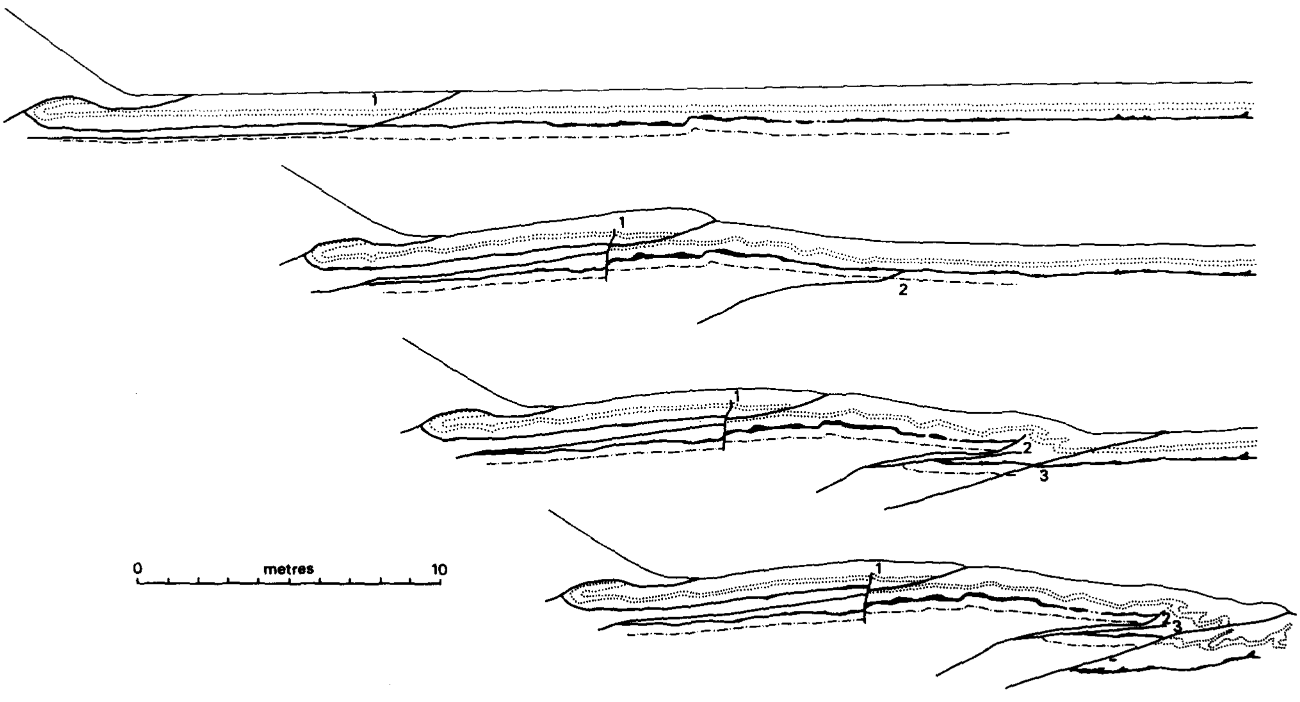


Fig. 10. Development of the proglacial part of the push ridge-complex (c), along profile 11–11′, according to the piggy-back model.

transition zone; the contrasting structural style of the proglacial and subglacial parts and the apparent convergence of faults and thrusts beneath the transition zone. These features do facilitate restoration, albeit at a less sophisticated and accurate level than is ideal.

Restoration of the proglacial part is achieved by assuming that the movement occurred as a piggy-back thrust system in which older movement planes occur higher in the sequence and have been carried forward by subsequent displacement along lower, more recent, thrusts (Dahlstrom 1970, Elliot & Johnson 1980, Butler 1982, Van der Wateren 1985). The total restoration of the proglacial part is achieved by successive restoration of thrusts numbered 1–3 in Fig. 9, and this is illustrated in Fig. 10.

It is evident that the materials in each of the horses comprising the transition zone are uniquely identical to the upper layers of undisturbed stratigraphy at the site of section A (Fig. 9) even though the detailed stratigraphy has been highly disturbed. Similarly each horse in the transition zone is similar in thickness to those in the proglacial part of the section, and all horses in both the proglacial part and the transition zone have developed in thrust planes approximately 30 cm below the white tephra layer. Any restoration must therefore logically place these horses from the transition zone into a stratigraphic position which replicates the undisturbed section as closely as possible. Using the piggy-back model employed above, the three major horses in the transition zone can be restored above the gravels in the subglacial part.

The same model of piggy-back thrusting cannot be used to restore the subglacial part. Indeed, the field evidence seems to conflict with established models of stress–strain relationships in marginal parts of modern glaciers (Paterson 1983). These models suggest that movements of subglacial material should reflect compression, not extension, although horizontal shearing along fault planes would be compatible. In most cases, however, downward-normal faulting could not be accommodated because the subglacial materials would be confined vertically and horizontally. In the case presented here, however, there is clear evidence to suggest that the normal faults in the subglacial part and thrusts in the proglacial part converge at the same point. The two movements (subglacial extension and proglacial compression) appear therefore to be structurally and mechanically linked via a basal thrust, which is concave upwards (Fig. 7). Such a model would allow normal faulting in the subglacial part of the movement to take place because any tendency to increase the volume of material in the transition zone would be compensated by movement along thrust planes in the proglacial part. Kälin (1971) carried out detailed analysis of the push-moraine developing in front of the Thompson Glacier, Axel Heiberg Island in the Canadian Arctic. Although he does not present structural data from sections, his model of development of the Thompson glacier push-moraine would enable subglacial normal faulting to develop. Provided therefore that extension can be compensated by observed compression of the same order of magnitude, there is no reason to reject subglacial extension.

EVOLUTION

Prior to deformation the stratigraphy throughout the proglacial area appears to have been very similar to that in the undisturbed sandur, and restoration extends this stratigraphy as a planar surface, some 50–70 m further towards the glacier (Fig. 11).

As Eyjabakkajökull surged to within 50–70 m of the maximum limit achieved in 1890 (in the vicinity of the lake-filled depressions up-valley from the ridge complexes) it appears to have begun to detach the overlying beds of turf, silt and compact silt in sheets from the gravels beneath. Such detachment would have been facilitated by high pore pressures generated within the gravels, providing a natural plane of décollement either within the gravel or at the base of the overlying beds. The measured rate of permeability for undisturbed silts and gravels ($k = 0.172$ and 0.0047 cm s^{-1}) demonstrate the contrasting ability of these materials to transmit water. Recent measurements of meltwater production during a glacier surge show that it is normally generated in large volumes at the glacier–substratum interface (Kamb *et al.* 1985). Provided that this meltwater permeates the glacier substratum and becomes linked with the groundwater in the proglacial area (which is the generally accepted model), then any meltwater generated at a rate in excess of the permeability of the gravels will induce significant pore water pressure within the gravels, since they are confined by relatively impermeable compact silts above. The overburden stress imposed by the rapidly advancing ice front would have added significantly to the rate of loading (Van der Wateren 1985).

As the sheets of silt and turf became imbricated and were pushed along the surface in front of the advancing ice margin, excess pore pressures continued to be generated within the gravel beds under the 'transition zone' and proglacial part of the formation. A synchronous movement then appears to have taken place, involving both the subglacial faulting and proglacial thrust movements. The two systems appear to be physically linked by a floor thrust, which is concave-upward, and along which development seems to have propagated. Although only small elements of this floor thrust appear in the section, the projection of all thrust planes in both the subglacial and proglacial sections indicates its presence. The concept of such a floor thrust or plane of décollement is well-established in regard to large-scale tectonic features (Price & McClay 1981) and implicit in most models of glacio-tectonics, although it is commonly assumed to be the depth to the permafrost table.

The inter-relationships of thrusts in the proglacial section suggest that they developed in the sequence shown in Fig. 11. The upper beds, particularly the compact blue silt, appear to have behaved as competent

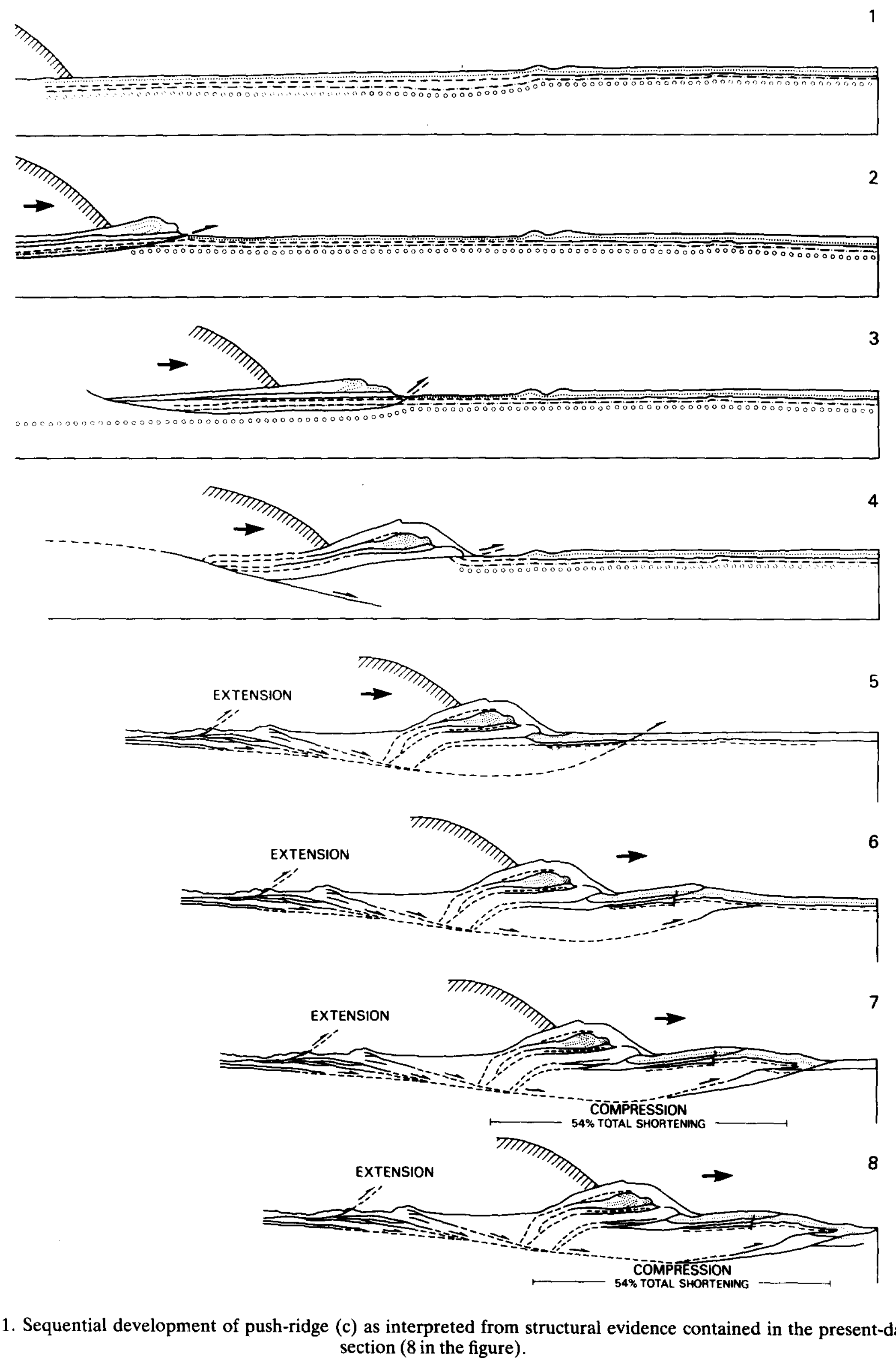


Fig. 11. Sequential development of push-ridge (c) as interpreted from structural evidence contained in the present-day section (8 in the figure).

strata, whereas the thrusts propagated through the gravel and cut up-section through the rest of the more competent units, a process typical of large-scale tectonic movements. As stresses increased, strain was initially accompanied by folding, followed by the initial thrust propagating close to the advancing glacier snout.

The numerous channels which are superimposed across the trend of the proglacial ridge complexes originate at the outcrop thrust planes. This suggests that the thrusts were water-lubricated during movement, and provided natural routes for the release of encapsulated meltwater or groundwater under high pressure. As the stresses in each thrust were dissipated, such outlets at those sites became unnecessary. As each thrust movement waned, the next thrust took up the movement, initially by ductile folding, but subsequently by failure along the gravel, cutting up-section through the silts and turf. The final movements involved sequentially lower levels of stress, and led to blind thrusts and eventually to simple fold structures at the toe of the system.

DISCUSSION

Standard methods of structural analysis can be successfully used to describe geological features of very recent age. The results of this investigation suggest that landforms which have properties similar to those of orogenic belts are currently being produced at the margins of some Arctic glaciers. The development of these mesoscale structures takes place within a time period which would allow *in situ* measurements to be made of the structures as they develop. Such a study would fill a wide gap between those of laboratory-scale models and analyses of orogenic belts. It is hoped that the publication of this paper will stimulate a cross-disciplinary exchange of ideas and comments to the benefit of those interested in glacio-tectonics and structural geology.

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