Pergamon

Journal of Structural Geology, Vol. 16, No. 10, pp. 1447 to 1462, 1994

0191-8141/94 $7.00+0.00

0191-8141(94)E0008-M

The structural geology of a surge-type glacier

WENDY J. LAWSON

Department of Geography, University of Auckland, Private Bag 92019, Auckland, New Zealand

MARTIN J. SHARP*

Department of Geography, University of Cambridge, Cambridge CB2 3EN, U.K.

and

MICHAEL J. HAMBREY

School of Science and Technology, John Moores University, Liverpool L3 3AF, U.K.

(*Received* 8 *June* 1992; *accepted in revised form* 10 *January* 1994)

Abstract—The range of deformation structures found in surge-type Variegated Glacier, Alaska, is similar to that found in many thin-skinned thrust belts, and includes foliations, folds, dip-slip and strike-slip faults, thrust faults and fractures. The development of structural relationships and the overall structural assemblage can be related to the deformation histories of quiescent and surge phases of motion. The surge phase dominates the formation of brittle structures because of the large stresses and strain rates associated with it. The relationship between downglacier-verging overturned folds with hinge lines transverse to flow, and a family of transverse arcuate thrust faults is characteristic of structural development in the lower part of the glacier during the surge. Longitudinal foliation develops mainly from sedimentary stratification in the upper part of the glacier in a marginal transpressive deformation regime during quiescent phase flow from broad accumulation basins. The structural relationships in this surge-type glacier may aid in the interpretation of structures in other gravity-driven tectonic settings.

INTRODUCTION: RATIONALE AND AIMS

GLACIERS serve as excellent natural laboratories in which to study rock deformation because analogous structures develop at rates which allow direct observations to be made concerning their formation (Hambrey & Milnes 1977). In addition, strain rates can be measured and used as a basis for quantitative structural interpretation. Temperate glacier flow (see Paterson 1981) provides a particularly good mechanical analogue for the motion of a gravity-driven thrust sheet on a décollement horizon above which displacement occurs by a combination of lubricated frictional sliding (cf. Hubbert & Rubey 1959) and distributed quasiviscous creep-like flow. Surge-type glaciers, of which Variegated Glacier in south-east Alaska is one of the most intensively studied examples (Bindschadler *et al.* 1977, Kamb *et al.* 1985, Raymond & Harrison 1988), have a pulsating flow regime in which sustained periods of relative inactivity (quiescence) alternate with brief, regularly-spaced periods of intense activity (surges). During surges, velocities and strain rates reach 10–100 times quiescent phase values (Meier & Post 1969). Surges begin at a nucleus, and expand by downglacier propagation of a wave-like surge front at speeds two–three times the ice velocity. This pulsating behaviour contrasts markedly with that of non-surging glaciers in which the average velocity field is essentially invariant over time and fixed in space. The motion and deformation of surge-type glaciers are conceptually similar to those of thrust sheets which are emplaced intermittently (Gretener 1981) and in which displacement is initiated at a point on the basal detachment and propagates away as a dislocation (Price 1988).

This paper has three specific aims:

(i) to outline the structural geology of surge-type Variegated Glacier, emphasizing those structures which have analogues in thrust sheets and which occur systematically over a large area (Gretener's 1972 'fundamentals');

(ii) to propose mechanisms for the formation of these fundamental structures on the basis of the chronology of deformation as suggested by their interrelationships, and by comparing this chronology with what is known of the mechanics of surge-type glaciers;

(iii) to evaluate the implications of these results for the interpretation of structures in rock tectonic settings.

*Present address: Department of Geography, University of Alberta, Edmonton, Alberta, Canada.

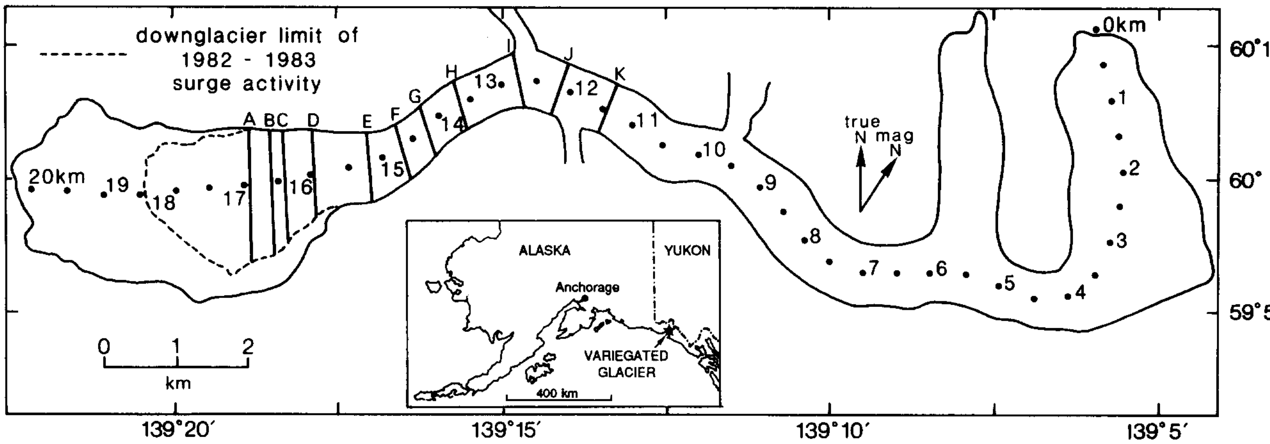

Fig. 1. Location and map of Variegated Glacier, showing distances from the head of the glacier along the centreline. The structural survey profiles, lettered A–K, are also shown.

VARIEGATED GLACIER AND ITS LAST SURGE CYCLE

Variegated Glacier is a 20 km-long temperate surge-type valley glacier that flows from east to west on the southern side of the St. Elias Mountains, southeast Alaska (Fig. 1). It drains from wide accumulation basins into a narrow, steep-sided, meandering valley (the 'confined glacier'). From the confined glacier it debouches westwards and southwards onto the coastal plain, ending in a broad terminal lobe that is partially in contact with the Hubbard Glacier on its otherwise steep-sided north side. Variegated Glacier has surged five times this century at intervals of about 18 years, in 1905–1906, sometime between 1930 and 1933, in 1947–1948, 1964–1965 and most recently in 1982–1983 (Bindschadler *et al.* 1977, Kamb *et al.* 1985).

Detailed descriptions of the motion and deformation history of the glacier since its 1964–1965 surge are presented elsewhere (Raymond & Harrison 1988, Sharp *et al.* 1988, Lawson unpublished dissertation 1989, Lawson & Sharp in preparation). Here we briefly summarize the findings of this work to provide a kinematic framework to assist with the interpretation of structural assemblages.

During the last surge cycle (between the end of the penultimate surge in 1965 to the end of the last surge in 1983) ice between 4 and 16 km from the head of the glacier (see Fig. 1) was displaced a mean distance of 3.1 km. On average, 60% of this displacement occurred during the 1982–1983 surge. During quiescence, motion occurred primarily by internal deformation of glacier ice, although basal sliding did also occur, particularly during the summer months, in the upper reaches of the glacier and towards the end of the period (Raymond & Harrison 1988). During the surge, intense ice deformation occurred in the vicinity of the propagating surge front (Raymond *et al.* 1987), but in the fully surging part of the glacier 95% of motion was by basal sliding with subglacial water pressures within 2–3 bars of overburden (Kamb *et al.* 1985).

During the quiescent phase of the cycle, the ice experienced weak longitudinally compressive strain, except in the region between 8 and 10 km from the valley head, where a localized reversal of the longitudinal velocity gradient resulted in weak longitudinal elongation. During this period, the spatial pattern of the longitudinal strain rate field remained unchanged, although the magnitude of the strain rates generally increased over time as the upper glacier was reactivated prior to the 1982–1983 surge. Deformation was essentially coaxial near the glacier centreline, but non-coaxial within well-developed marginal shear zones.

The 1982–1983 surge occurred in two phases (January–June 1982 and October 1982–July 1983) and affected ice from 2 to 18.5 km from the head of the glacier (Fig. 1). The surge originated at a nucleus around 4.8 km from the glacier head, and propagated up and down-glacier from there. Downglacier propagation involved the formation of a velocity peak which increased in magnitude to a maximum around 70 m day^{-1} with increasing distance downglacier. Downglacier from this velocity peak was 1–1.5 km wide zone of longitudinally compressive flow in which the glacier experienced rapid thickening and the development of a prominent topographic ramp known as the surge front (Raymond *et al.* 1987). Upstream of the velocity peak, ice experienced weaker longitudinal elongation. Since the surge front and associated velocity peak propagated downglacier at rates two–three times the ice velocity, ice within the glacier experienced deformation histories which were strongly dependent upon location relative to the surge nucleus and the final position reached by the velocity peak (Fig. 2) (Sharp *et al.* 1988). Ice downstream of the final position of the velocity peak experienced continuous and large cumulative shortening, while ice upstream of the nucleus experienced continuous and cumulative elongation. Ice in the intermediate region experienced shortening followed by elongation as the velocity peak passed by, and low cumulative strains. After being reactivated, ice within the surging part of the glacier moved as a plug defined by marked lateral wrench faults (Kamb *et al.* 1985). Within this plug, deformation was essentially coaxial across the width of the glacier.

Ice exposed at the surface of the ablation area has taken 10–100 years since deposition to reach its present

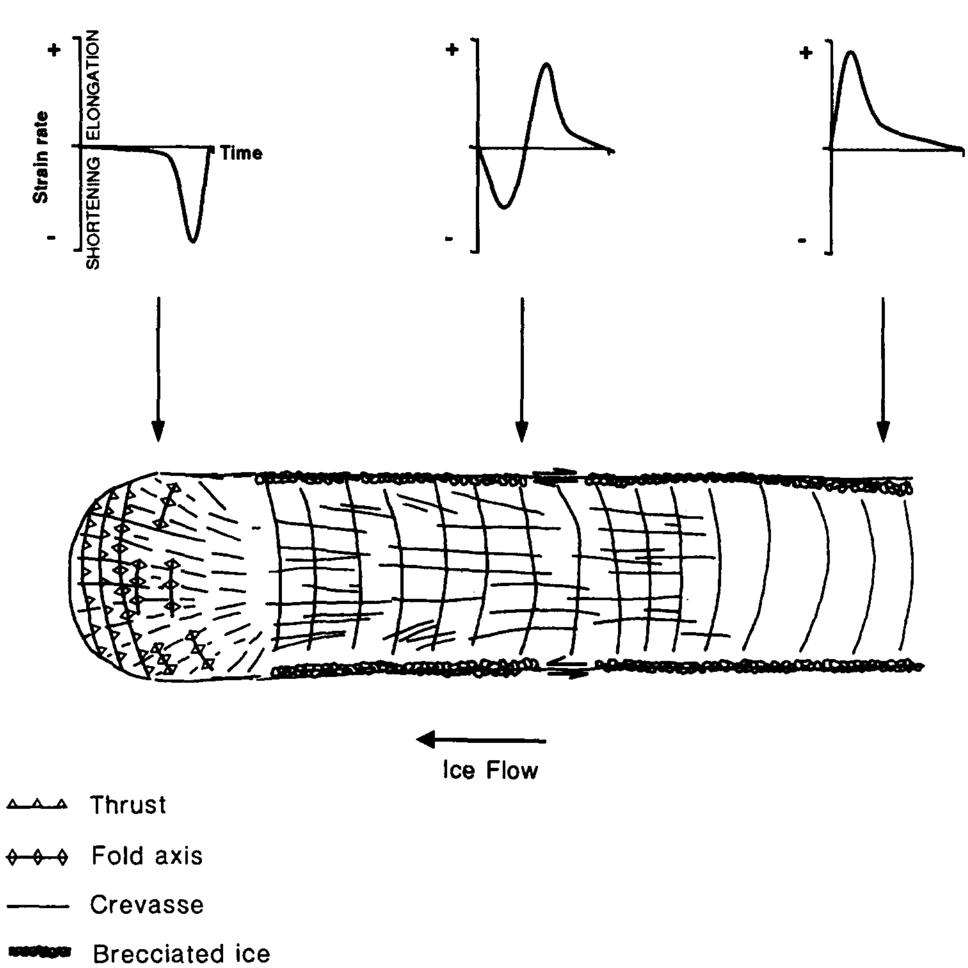

Fig. 2. Schematic diagram showing structural patterns at the end of the 1982–1983 surge, and the deformation histories experienced by ice in different parts of the glacier that were affected by the surge. Note the three-fold structural zonation, comprising an extensional regime towards the hinterland, and compressional regime towards the foreland, and a superimposed compressional–extensional regime between the two.

position, depending on its location. The most distal ice in the terminal lobe has probably taken about 100 years to travel downglacier since being deposited high in the accumulation area and has experienced a total deformation history that includes approximately six surges.

Structural information presented in this paper is derived largely from field mapping conducted in 1986 and 1987. Additional information about structural patterns at other times has been obtained from aerial photographs. In the field the glacier was mapped as far downglacier as the limit of the 1982–1983 surge, and as far upglacier as the position of the late summer snowline allowed (Fig. 3). Measurements of the orientation of the principal structures were made primarily along a series of transects across the glacier (Fig. 1).

THE STRUCTURAL GEOLOGY OF VARIEGATED GLACIER

The pattern of outcrop of the fundamental structures of Variegated Glacier in 1986–1987 is illustrated on Fig. 3. The nature and sequence of development of the fundamental ductile structures are summarized in Table 1. Brittle structures are not included in this table for reasons which are outlined in the following discussion.

In this paper we use the terms *ductile* and *brittle* to describe the way in which a structure has accommodated deformation at the outcrop scale, and do not intend to imply that brittle and ductile (or plastic) deformation mechanisms have operated at the crystal scale (cf. Rutter 1986).

Compositional layering

Compositional layering is observed throughout the study area and consists of irregular alternating layers of coarse-grained clear blue ice and coarse-grained bubbly white ice tens of centimetres thick with small amounts of fine-grained ice. This layering is similar to primary sedimentary stratification found in many non-surging glaciers. Its attitude is typically low-lying and variable (a few orientations are plotted on the map in Fig. 3). Rock debris of variable lithology, clast size, roundness and degree of sorting that is found within the ice is disposed parallel to and between some compositional layers. The lithologies correspond to those found in the headwalls of the glacier above the accumulation basins (unpublished work by B. Kamb), suggesting that the debris was deposited on the glacier surface by various forms of mass movement and subsequently buried by snowfall. This implies that the compositional layering is primary sedi-

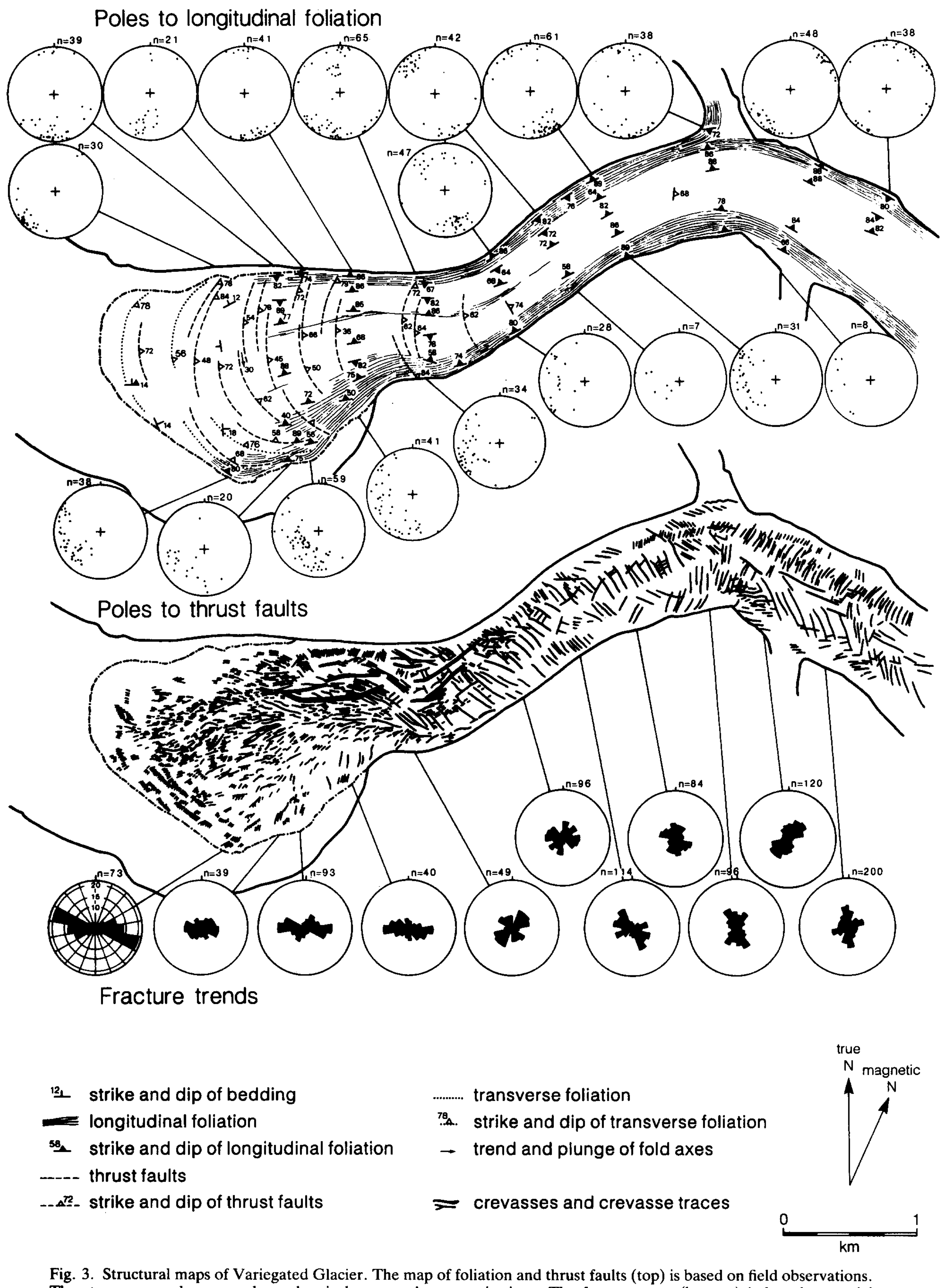

Fig. 3. Structural maps of Variegated Glacier. The map of foliation and thrust faults (top) is based on field observations. The stereograms shown are lower-hemisphere equal-area projections. The fracture map (bottom) is based on aerial photographs taken on 28 August 1983, 6 weeks after the end of the 1982–1983 surge. Rose diagrams of fracture trends are shown. (Figure 3 is continued on the following page.)

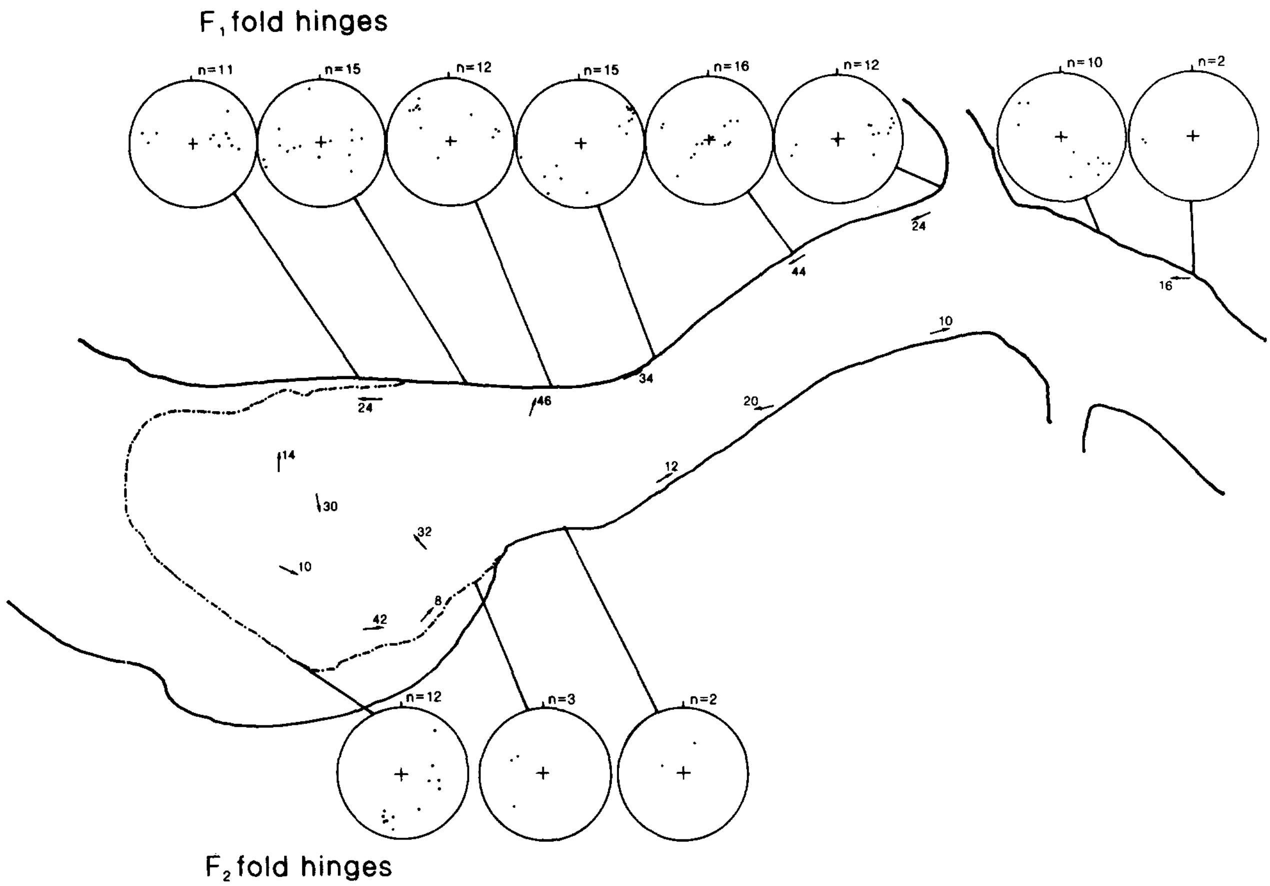

Fig. 3. (Continued.) Structural map of Variegated Glacier showing fold hinges, based on field structural observations.

Table 1. Summary of characteristics of fundamental ductile structures at Variegated Glacier

Element	Characteristics
S_0	Pervasive compositional layering
S_1	Longitudinal foliation
F_1	Upright, tight to isoclinal folds with axial planes parallel to flow and gently plunging hinge lines
S_2	Transverse foliation
F_2	Overturned folds, verging downglacier with axial planes transverse to flow and gently plunging hinge lines
F_3	Large-scale upright, tight to isoclinal folds with axial planes transverse to flow and subvertical hinge lines

mentary stratification, and it is thus denoted S_0 (Table 1).

Foliations

Two penetrative foliations which occur in different areas of the glacier are defined by layers of ice of varying bubble content, and differentiated on the basis of their orientation. The older foliation (S_1) develops further upglacier and has suffered more subsequent deformation than the younger (S_2). S_1 is steep and parallel to flow within the confined glacier (Fig. 3). In the terminal lobe, divergence of flow lines and flow-parallel shortening have resulted in large-scale folding of the S_1 foliation into isoclinal folds with vertical hinges. As a result, S_1 lies parallel to the margin where it is exposed at the south edge of the terminal lobe. S_1 is fairly consistent in attitude within a small area, with variations in orientation being limited to 25°, and variations in dip to 20°.

Intensely developed S_1 consists either of discontinuous layers and lenses of bubbles in coarse clear ice, or of steeply-dipping, attenuated compositional layering. The latter type of S_1 is not seen at locations where flat-lying sedimentary stratification is readily observable. On a horizontal ice surface, weathering of strong S_1 produces a distinctive longitudinal trend in the microtopography (Fig. 4a). Individual foliae vary from millimetres to centimetres in width and from centimetres to metres in length. Associated bubbles are generally $\geqslant$ 1 mm in diameter.

In the lower part of the confined glacier, the form of S_1 changes between the margins and the glacier centreline. At the margins, S_1 is strongly developed and consists of steeply-dipping compositional layering, whereas towards the centre of the glacier it consists of a weak cleavage. The compositional nature of S_1 near the margins suggests that it is formed by the transposition of sedimentary layering. In places, offset of crevasse traces indicates strike-slip between adjacent steep layers of compositional S_1. In the centre of the glacier, compositional S_1 is absent, except beneath and parallel to medial moraines in ice which was at the margins of tributaries prior to their confluence with the main glacier. There is also an upglacier transition in the form of S_1 found near the glacier margins. In the terminal lobe and lower part of the confined glacier, S_1 consists of steep compositional layering, but in the upglacier part of the mapped area, it consists of a well developed cleavage.

S_2 foliation is distinguished from S_1 by its transverse-to-flow orientation and arcuate pattern of outcrop. It dips upglacier in the centre of the glacier, and inwards at the margins, thus having a 'nested spoons' geometry similar to that of transverse foliations in non-surging glaciers (Fig. 3) (Allen *et al.* 1960, Ragan 1969, Hambrey *et al.* 1980). The folia of S_2 are generally finer than those of S_1, and the associated bubbles are smaller (≤ 1 mm in diameter). As with S_1, the amount of clear ice and the orientation of the folia are variable but, unlike S_1, the character of S_2 does not vary systematically over the mapped area. A transposed type of S_2 is, however, locally developed around F_2 folds (see following section) where the steep overturned limbs of sedimentary stratification become parallel to S_2 bubble foliation (Fig. 4c). S_2 occurs only in the lower part of the terminal lobe and is best developed towards the centreline. At the south margin, S_1 predominates to the east, while S_2 is prevalent to the west. Although it frequently occurs directly beneath moraines in the terminal lobe, S_2 has no debris directly associated with it.

Because of the topographic controls of the morphology of the terminal lobe as outlined in the introductory sections of the paper, structures related to the divergence of flow-lines and the corresponding flow-parallel shortening, such as the isoclinal folding of S_1 described above, are better developed on the south side of the terminal lobe than on the north.

Folds

There are three fold systems in the ice. F_1 and F_2 are mesoscale (observable at outcrop scale), while F_3 is large scale and identifiable from moraine patterns on aerial photographs. Both mesoscale fold systems have surface-parallel hingelines. F_1 folds are upright with hinge lines parallel to flow, tight to isoclinal, and symmetrical and similar in style (Figs. 3 and 4b). They affect only S_0 compositional layering. Although their hinge lines are typically horizontal or subparallel to the ice surface, they display a wide range of plunges consistent with the effects of subsequent superimposed deformation. First-order folds of this system have amplitudes and wavelengths of centimetres to metres. These folds are well developed near the margins of the confined glacier where they are isoclinal; away from the margins the folds become progressively more open and have a smaller amplitude-to-wavelength ratio. F_1 folds are also well developed towards the south margin of the terminal lobe, where their hinges lie parallel to the strike of S_1 and the margin. The relationship between F_1 folds and the longitudinal S_1 foliation is discussed in more detail below.

F_2 folds have shallow dipping hinge lines perpendicular to the flow direction (Fig. 3). They are asymmetrical to overturned, verge downglacier and are open to tight (Fig. 4c). They can have either similar or parallel morphology, although the 'parallel' folds are not strictly parallel in the sense of Ramsay class 1B folds, but lie somewhere between similar 1C and parallel 1B folds (Ramsay 1967). The 'parallel' F_2 folds are less common. Although it is primarily the compositional layering which is folded by this system, foliations and thrust faults are also affected. F_2 folds have amplitudes which vary from a few centimetres to metres, and half wavelengths of the same order of magnitude. They are best developed and most common in the terminal lobe, where they occur across the width of the glacier. Towards the south margin of the terminal lobe, where flow lines diverge as ice flows out of the confined glacier, the fold hinges lie oblique to the margin (Fig. 3). Some larger F_1 and F_2 folds have smaller parasitic second-order folds on their limbs. In F_2 folds, parasitic folds typically occur on the gently dipping upglacier limb.

F_3 folds occur only in the terminal lobe and are defined by folds in medial moraines (Fig. 5). They are tight to isoclinal and their axial planes are steep and lie transverse to flow. Their pattern of outcrop suggests that they have hinges that dip steeply upglacier. An analysis of changing moraine patterns indicates that these folds form by the amplification of bulb-shaped irregularities in the medial moraines during passive folding (cf. Hudleston 1976). These bulb-shaped irregularities themselves are a well-known morphological feature of surge-type glaciers which develop as a result of the asynchronous surging of tributaries (Post 1972, Driscoll 1980). A detailed survey of the structures in ice lying within one of these bulb-like hoops indicates that ice within the loops becomes structurally incorporated into the main glacier and does not remain isolated. These loops therefore do not comprise autonomous structural units although their morphology is marked at the surface of the glacier by the pattern of the medial moraines. The F_3 folding results in the S_1 foliation becoming aligned transverse to the flow direction at the south side of the terminal lobe (Fig. 3).

F_2 and F_3 folds are therefore both related to longitudinal shortening and are both best developed in the terminal lobe. The difference between them is largely a function of the nature of the layer being deformed and their distribution. Although the F_2–F_3 designation may therefore be considered somewhat arbitrary, F_2 is inferred to pre-date F_3 because F_2 folds are found further upglacier. These two structures serve to illustrate clearly that very different types and scale of structures can form in the same place as the surge passes through.

Fractures

Fractures at Variegated Glacier vary from huge surge-phase chasms through currently active marginal crevasses to short closed fractures. At the time of structural mapping the majority of fractures were open and inactive but had clearly been active a few years previously in the 1982–1983 surge. The only active crevasses at the time of mapping lay in the confined glacier.

The active crevasses were of two types. Small longi-

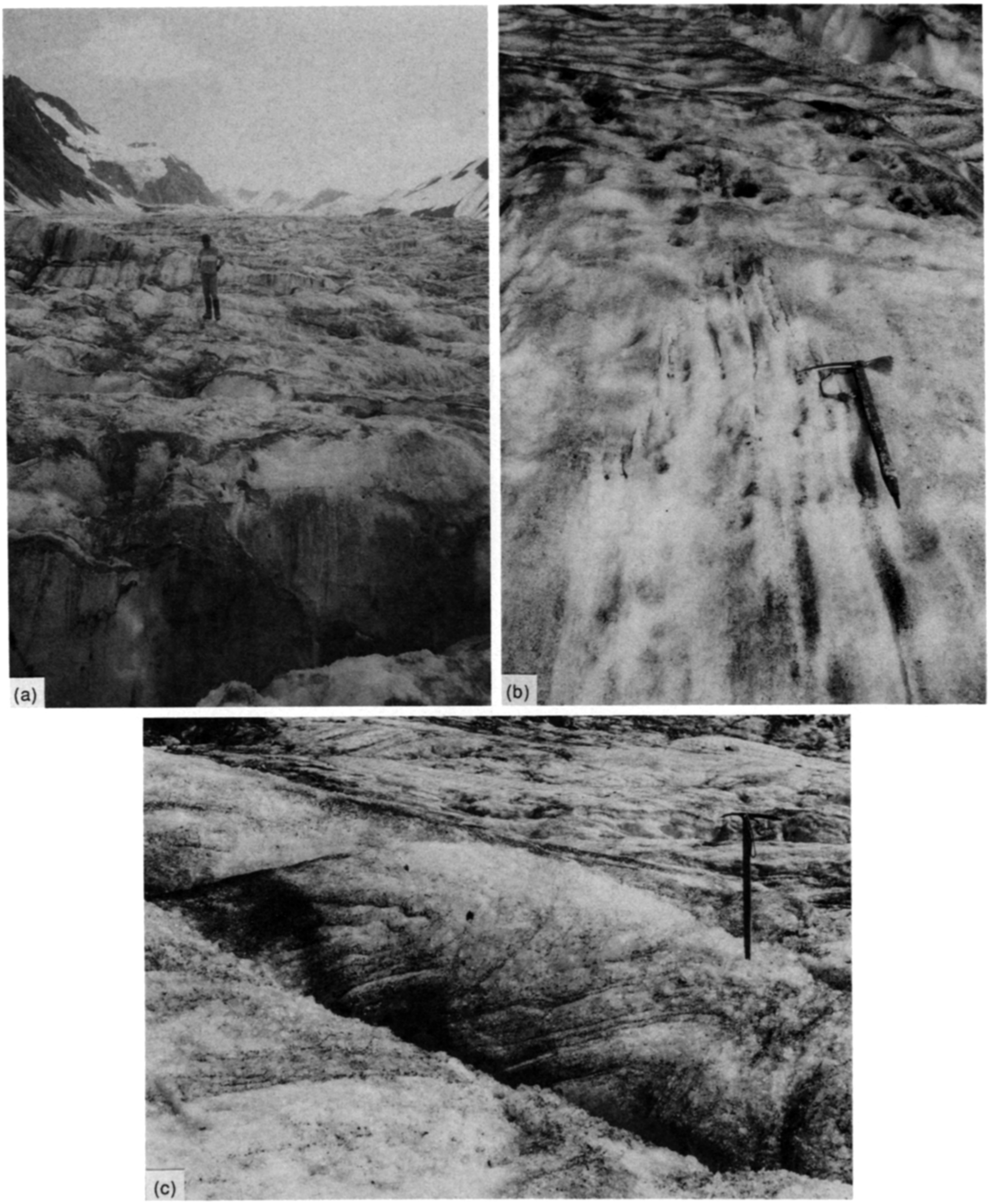

Fig. 4. (a) Intensely developed S_1 longitudinal foliation at the north side of the glacier in the vicinity of Profile J (see Fig. 1). Foliation at the margins of the confined glacier is transposed from sedimentary stratification (Fig. 7, i). (b) An upright isoclinal F_1 fold. Flow is parallel to the penetrative axial planar S_1 longitudinal foliation, and out of the photograph. (c) An overturned F_2 fold in sedimentary stratification, exposed in a longitudinal crevasse in the centre of the terminal lobe. Flow is left to right. Note the two S_2's in the bottom right of the photograph, which are axial planar to the folds, but which in this case have no displacement across them.

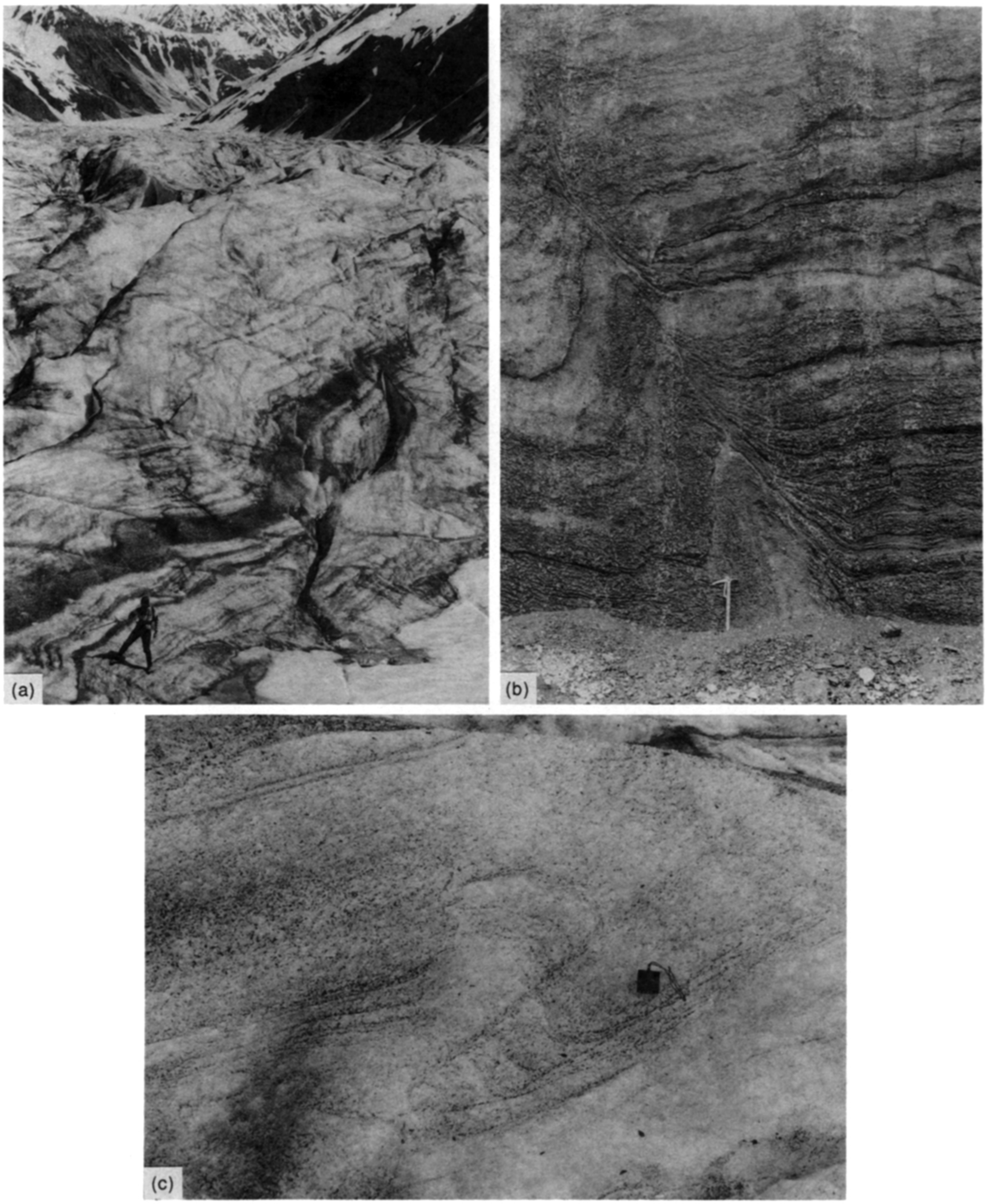

Fig. 6. (a) Folding and thrusting of low-lying sedimentary stratification in the terminal lobe. Flow is from top right to bottom left, and out of the page. (b) Thrust fault axial planar to an F_2 fold in sedimentary layering, seen in a vertical flow-parallel section at the south margin of the terminal lobe. (c) Pop-up between two faults with a dihedral angle of about 40° in a flow-parallel vertical section near the centreline of the glacier near Profile F.

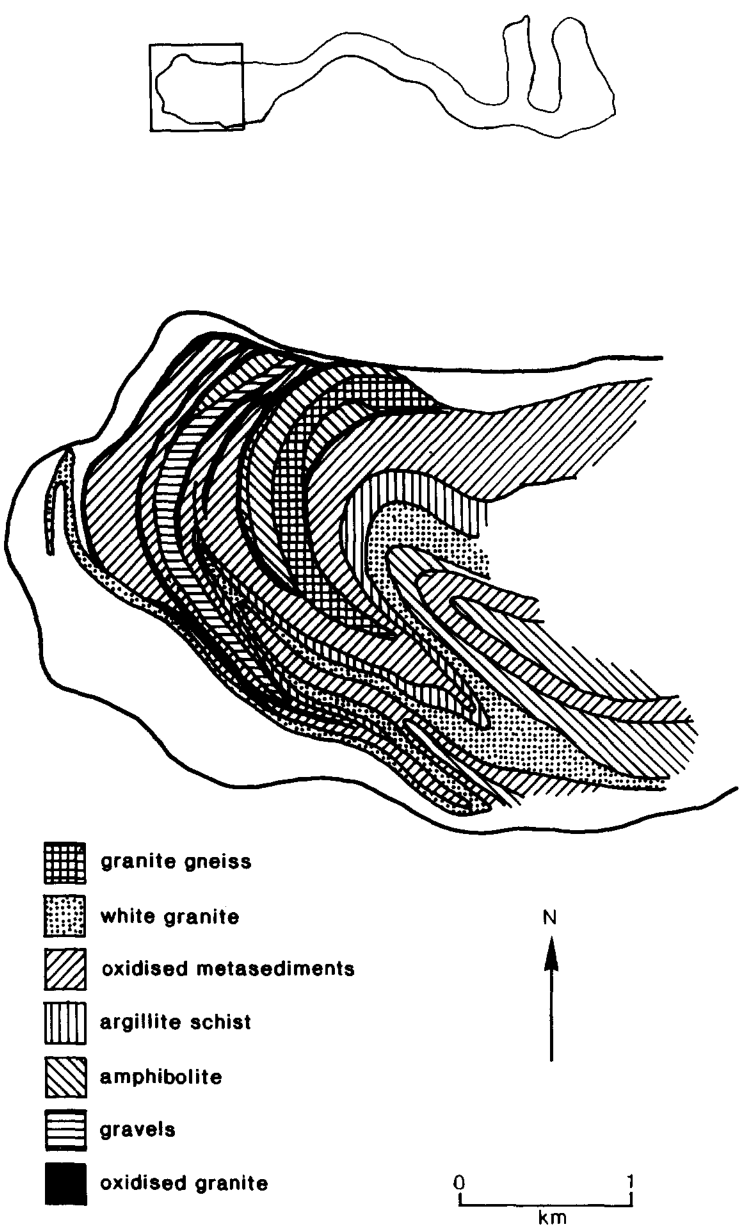

Fig. 5. Map of the lithologies of medial moraines in the terminal lobe showing the form of the large-scale F_3 folds which develop from the passive folding of bulb-like loops in medial moraines.

tudinal crevasses occurred in the centre of the glacier in the vicinity of Profile E, and large-scale en échelon arrays occurred at the margins at various locations. The en échelon arrays lay parallel to the margins, and crevasses within them made an angle of about 45° with the valley walls upstream. A detailed analysis of aerial photographs indicates that these marginal arrays exist at all stages of quiescence, especially in the vicinity of tributary confluences. Crevassing also occurs throughout quiescence in the upper reaches of the glacier where transverse crevasses develop in the accumulation basins. Significant transverse crevasses develop from mid-quiescence in the region of 10 km from the head of the glacier and longitudinal splaying crevasses at about 7 km during late quiescence.

Crevasse traces (any fracture with no opening; cf. Hambrey & Milnes 1977, p. 672) were the most pervasive of all observed structures. They have a variety of forms which reflect their mode of formation (Hambrey & Müller 1978). Typically they consist of straight veins of coarse-grained clear ice a few centimetres to a few millimetres wide, sometimes with a central fracture surface or layer of bubbles. Such crevasse traces have matching walls and offset earlier structures, and are therefore dilatational in origin (Ramsay 1980). Some have a comb structure formed by the syntaxial growth of elongate crystals perpendicular to the vein walls (cf. Durney & Ramsay 1973) and are inferred to be syntectonic veins which formed as basal or lateral extensions of active crevasses.

A second type of crevasse trace consists of layers of ice formed by refreezing of standing water. These have a characteristic form comprising large clear crystals growing radially inwards from the vein walls and converging at a central plane of suture (cf. Stenborg 1968). They often consist of several discontinuous layers of crystals suggesting intermittent formation, perhaps due to repeated crack–seal (Ramsay 1980) or interrupted freezing. A third type consists of closed fracture surfaces with no associated vein filling. Straight planes of bubbles in clear ice mark their location in the terminal lobe. Open inactive crevasses also occur throughout the mapped area. These range in size from large chasms to small cracks.

The pattern of measured fracture orientations varies between the terminal lobe and the confined glacier (Fig. 3). In the terminal lobe flow-parallel fractures are most common, although crevasses occur with all orientations. Conjugate sets of crevasse traces occurred, with the bisector of their acute dihedral angle having the same orientation as flow-parallel crevasses. Whilst the flow-parallel crevasses are aligned parallel to the most compressive principal stress axis as it existed during the 1982–1983 surge (Sharp *et al.* 1988), the conjugate fractures intersect in a line parallel to the intermediate principal stress (cf. Hobbs *et al.* 1976, fig. 7.31). These conjugate fractures are therefore inferred to be conjugate shears rather than tensile fractures. Although these conjugate shears cannot be considered to be one of the fundamental structures of the glacier, they are interesting because they demonstrate that shear as well as tensile fracture of ice occurs.

In the confined glacier, flow-parallel and transversely aligned fractures occur with roughly equal frequency. Although definitive cross-cutting relationships are often hard to identify, it seems that the transversely orientated crevasses most commonly post-date the flow-parallel ones. Here the ice is so pervasively fractured that individual sérac blocks may be only tens of centimetres across. The orientations of the most common fractures in the confined glacier are therefore consistent with those of tension fractures formed in longitudinal compression (flow-parallel crevasses) and longitudinal extension (transversely oriented crevasses).

Fractures are widely associated with dip-slip and strike-slip displacements. Strike-slip displacements accommodate transverse stresses caused by velocity gradients or adjustments of flow to valley geometry and occur on crevasses with any orientation. Particular parts of the glacier tend to be associated with characteristic senses of strike-slip displacement. For example, south of the centreline in the terminal lobe the sense of strike-slip motion is sinistral, reflecting shear during flow line divergence. At the north margin of the confined glacier below the bend sinistral strike-slip displacements reflect rotation as the glacier flowed around the bend. Dip-slip displacements are most common on transverse fractures and accommodate longitudinal stresses which result in both shortening and elongation.

Although there are typically many sets of fractures characterized by particular orientations at a single locality, the complexity of the crevasse patterns means that it is generally impossible to identify a complete sequence for their development. However, there are two particular temporal relationships that occur consistently. In the confined glacier, as indicated above, recent activity on transverse crevasses post-dates recent activity on longitudinal crevasses. In the terminal lobe longitudinal crevasses are the most recently active. In terms of the relationships between fracture and the ductile structures, cross-cutting relationships indicate that the most recent activity on the majority of fractures post-dates the formation of both S_1 and S_2 foliations. The widespread existence of crevasse traces throughout the mapped area points to the possibility that crevasses may form, close up and the later be reactivated at a location further downglacier.

Thrust faults

Many transverse fractures in the terminal lobe display thrust displacements (Figs. 6a & b). These fractures typically dip 40–80° upglacier (Fig. 3), although a few have dips as low as 10° and some dip downglacier and show evidence of backthrusting. The thrust 'planes' are often actually layers (rather than planes) comprised of coarse-grained clear ice up to a few decimetres wide and containing some kind of internal planar feature. This internal feature can be a layer of fine-grained ice, a fracture plane or a plane defined by aligned crystal boundaries. Other types of thrust fault include isolated layers of fine-grained ice millimetres to centimetres wide probably originating in dynamic recrystallization, single fracture planes and layers of ice with characteristics typical of formation by refreezing of meltwater, as mentioned above. Composite features were also observed. Some of the thrust faults had debris entrained along them, and turbid water was seen gushing from some of the fault surfaces. Syntectonic observations of thrusting were made during the surge, when turbid water was also seen to gush from actively thrusting faults (Sharp *et al.* 1988).

In some cases, thrust faults with moderate upglacier dips have antithetic backthrusts branching off them. These dip downglacier and define a pop-up between the two faults. Pop-ups also form between synthetic faults with opposite senses of displacement (Fig. 6c), and between faults which branch along strike, rather than along dip, displacing ice sideways in what might by analogy be termed a pop-out.

Thrust faults have an arcuate pattern of outcrop and listric shape, similar to that of S_2 foliation, and show no systematic longitudinal variation in attitude over the mapped area. At a single locality, thrust faults display a variety of dips, and faults with different attitudes cross-cut and displace each other by tens of centimetres. Some thrust faults are folded by F_2 folds whereas some are axial planar to F_2 folds and displace the limbs relative to each other (Fig. 6b). The spacing of faults varies from decimetres to, more typically, several metres, and in some places they crop out in groups. Individual thrust faults are typically tens of metres long in outcrop. Thrust faults occur mainly in the terminal lobe, especially towards the centreline. The terminal lobe therefore consists of an imbricate fan thrust complex (Boyer & Elliot 1976). In the confined glacier, thrust faults are uncommon, and none were seen on Profiles I and K.

Cross-cutting relationships indicate that recent activity on the thrust faults pre-dates that on the recently active longitudinal crevasses. The ambiguous relationship with F_2 folds suggest that thrust fault activity is roughly synchronous with the formation of F_2, but is more recent that the formation of F_1.

The sequence of structural development

The three morphologically-distinct types of fold structure, together with the presence of sedimentary stratification, define a sequence of ductile structural development at Variegated Glacier which is summarized in Table 1. Geometric relationships between foliations corroborate this structural history. However, establishing a unique history for brittle structural development is impossible because of the complexity of geometric relationships between contiguous sets. The most recent activity on most—but not all—crevasses post-dates the formation of the S_2 foliation. Crevasses were also actively forming at the time of the fieldwork. Thrust activity occurred roughly synchronously with the formation of F_2.

CHARACTERISTIC STRUCTURAL INTERRELATIONSHIPS AND THEIR INTERPRETATION

In the previous section we have outlined the types and characteristics of structures found at Variegated Glacier at one particular stage in its surge cycle, and assigned the ductile structures relative ages suggested by their geometric relationships. In this section we examine the nature of geometric, morphological and temporal relationships between different structures in order to elucidate the processes of their formation.

Deformation histories in glaciers

Although throughout this paper we have used the usual subscripted historical notation denoting structural type and relative age, there are particular limitations in its application to glacier tectonic settings. In non-surging glaciers deformation phases are typically defined spatially by the geometry of the valley, rather than temporally by distinct deformation events. As a result, structures associated with all deformation phases may be forming simultaneously at different locations within a glacier. The situation is more complicated in surge-type glaciers where, during surges, a propagating deformation front moves downglacier through the ice at speeds greater than the glacier velocity. The passage of such a wave-like deformation front produces deformation events which are both spatially and temporally defined. Whilst both longitudinal compression and longitudinal extension occur simultaneously at different locations in the glacier, individual parcels of ice may be subjected to compression followed by extension as the deformation front passes. Whilst we relate the deformation episodes recognizable from structural interrelationships to a single ideal surge cycle, we stress that ice experiencing D_n in one surge cycle may already have experienced D_{n+1} and D_{n+2} in a previous cycle.

Sedimentary stratification

The extent to which sedimentary stratification survives into the ablation areas of glaciers depends upon channel geometry and ice flow characteristics. In some glaciers (for instance where channel geometry produces strong transverse compression and cumulative strains are large) stratification may be overprinted by or transposed to foliation at an early stage (e.g. White Glacier, Axel Heiberg Island; Hambrey & Müller 1978), whilst in others (particularly small glaciers of relatively simple geometry) it may be preserved as far as the snout (observations on Swedish glaciers, Hudleston personal communication). At Variegated Glacier, sedimentary stratification extends well into the terminal lobe. This persistence is probably due to the high proportion of the downglacier displacement of ice that occurs by basal sliding under conditions of plug flow. The brittle deformation that occurs under such conditions dissects the stratification, but does not alter its fundamental nature. Hence S_0 is well preserved in the central areas of the glacier.

F_1 *folds and longitudinal* S_1 *foliation*

The compositional nature of S_1 foliation in marginal areas of the glacier and its axial planar relationship to F_1 folds suggest that this foliation developed by transposition of S_0 sedimentary stratification during the deformation event that produced the F_1 folds. This event is denoted D_1. The style of F_1 folds is consistent with formation in simple shear with a superimposed element of pure shear induced by transverse shortening (cf. Gilbert & Merle 1987). The spatial distribution of these structures suggests that D_1 deformation is concentrated close to the glacier margins and occurs in the upper reaches of the glacier. Since comparable folds and foliations have frequently been observed in non-surging glaciers (e.g. Blue Glacier, Washington, Allen *et al.* 1960; Saskatchewan Glacier, Alberta Rockies, Meier 1960; White Glacier, Axel Heiberg Island, Hambrey & Müller 1978) it seems likely they form during the quiescent phase when the processes of flow are most similar to those that operate in non-surging glaciers (Bindschadler *et al*, 1977). The ductile nature of the structures also argues against a surge-phase origin, because marginal deformation during surges is concentrated along well-defined lateral wrench faults (Kamb *et al.* 1985) rather than distributed through wider shear zones.

During quiescence, strong transverse compression must occur as ice flows from the 3 km wide accumulation basins into the 1 km wide confined glacier (Fig. 7,i), and it seems likely that this is where D_1 deformation occurs. If this transverse shortening (of > 0.6 of original width) were experienced by a static body of ice in pure shear, the deformation would be approximately homogeneous

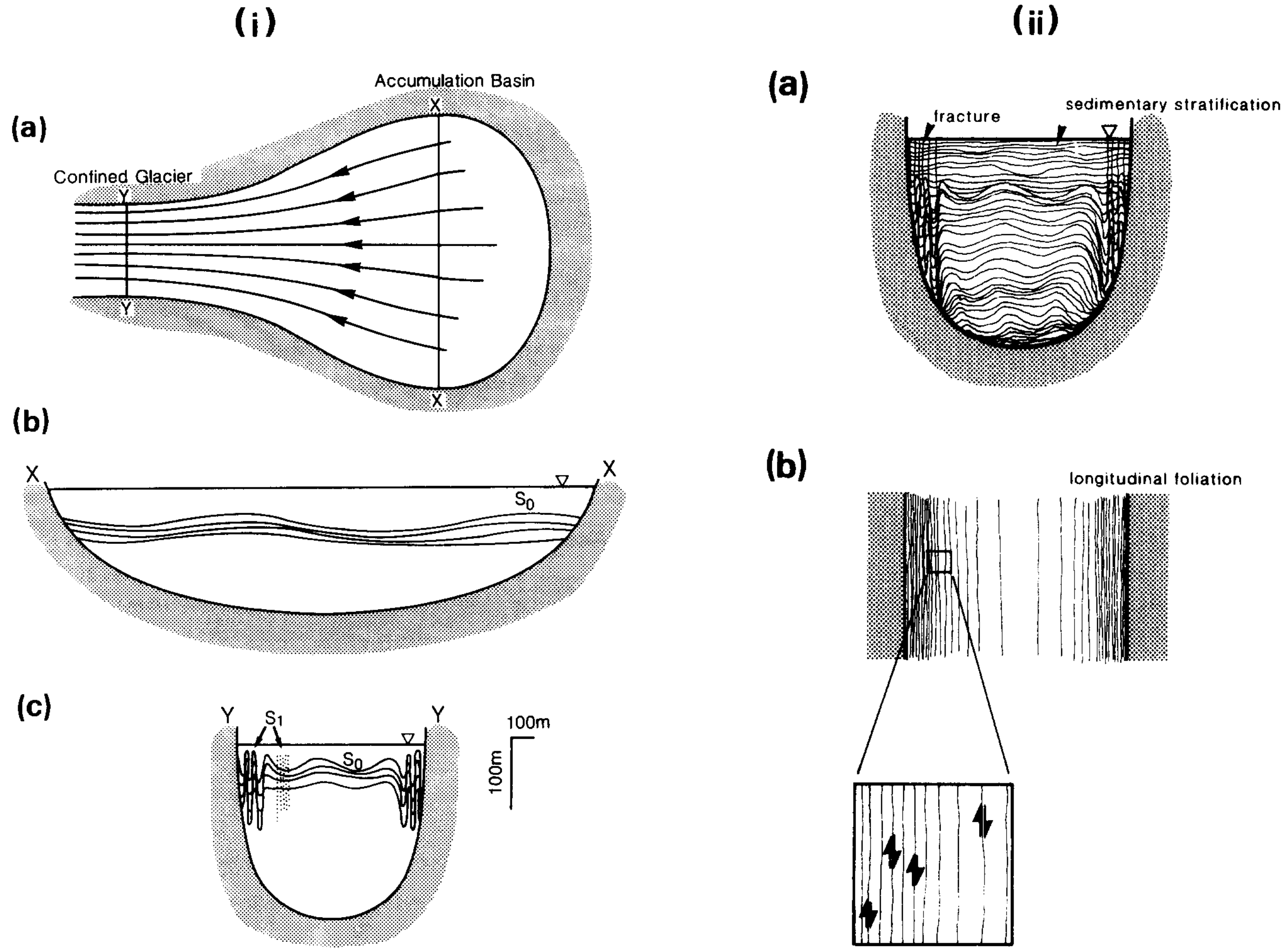

Fig. 7. Models for the development of longitudinal foliation at Variegated Glacier. (i) The development of longitudinal foliation in ice deposited in the upper accumulation basins. (a) Map view showing a downglacier reduction in width (*XX* to *YY*) comparable to the amount of transverse shortening experienced by ice moving from the wide accumulation basins to the narrow confined glacier. (b) & (c) Cross-sections showing the downglacier evolution of upright F_1 folds and transposed S_1 near the margins, and axial planar cleavage away from the margins. (ii) The development of longitudinal foliation in ice which is deposited in the narrow part of the accumulation area and which does not therefore undergo marginal transpression. (a) Cross-section showing isoclinal flow-parallel folds transposing sedimentary layering near the margins at depth, and longitudinal fracture planes nearer the surface. (b) Plan view, showing how the fracture cleavage that comprises the longitudinal foliation accommodates transverse shear by strike-slip.

throughout the cross-section. However, during normal glacier flow transverse shear stress is largest towards the glacier margins and consequently transverse shear strain is concentrated in marginal ductile shear zones. The viscosity of ice is inversely related to its deformation rate (Glen 1955), and hence ice towards the margins has a lower effective viscosity than ice in the centre of the glacier. Ice at the margins thus undergoes a kind of strain softening, which may involve rotation and recrystallization of grains into an orientation suitable for intracrystalline glide (Hudleston 1980), and which allows a disproportionate amount of the transverse shortening to be taken up there (Fig. 7,i). A component of pure shear is therefore added perpendicular to the simple shear.

The result of the concentration of deformation at the margins is that the intensity of folding is greatest there and decreases away from the margins. In this model for the formation of longitudinal foliation, the deformation experienced by the ice at the margins can be characterized as a regime of simple transpression, in the specific sense of oblique convergence (Harland 1971). If ice in the marginal shear zones undergoes strain softening, it is in effect experiencing transpression between two relatively rigid blocks, the margin on one side and the unsoftened ice on the other. At the ice–rock contact, there is a deformation discontinuity, whereas at the inward edge of the shear zone there is a locally steep deformation gradient rather than a discontinuity. Volume conservation during transpression of the marginal zone is achieved by longitudinal extension (but see Sanderson & Marchini 1984) which also reaches a peak in the marginal shear zone.

This model for the development of longitudinal foliation in marginal transpression explains the relationship between longitudinal foliation and upright flow-parallel folds, but it cannot account for the formation of the well developed cleavage type of foliation found in ice which is deposited in the narrow part of the accumulation area between 6 and 9 km from the head of the glacier. Another mechanism for the development of longitudinal foliation must therefore operate in this ice.

It is suggested that the formation of longitudinal foliation in ice deposited in the confined glacier, which does not experience marginal transpression, is closely linked to the presence of pre-existing longitudinal foliation at depth (Fig. 7,ii). In the confined glacier, marginal simple shear which occurs subsequent to the formation of the transposed foliation is partly accommodated by strike-slip displacement along the subvertical layers of the transposed sedimentary stratification. It is postulated that the well developed flow-parallel cleavage which crops out near the margins of the upglacier part of the mapped area is a shear cleavage that develops when this slip at depth is transmitted upwards into younger, unfolded ice. This type of foliation is therefore unusual in that its characteristics are not directly derived from any pre-existing inhomogeneity (cf. Hambrey 1975, Hambrey & Milnes 1977, Hooke & Hudleston 1978).

F_2 *folds,* S_2 *foliation and thrust faults*

The early D_2 phase of deformation produced overturned F_2 folds, and S_2 foliation and thrust faults which lie parallel to their axial planes (Fig. 6b). Thrust faults frequently lie along the axial planes of F_2 folds and displace their limbs, indicating that folding was associated with, but preceded thrusting (as described by Fischer & Coward 1982 for the Heilam thrust sheet, northwest Scotland). As in the Heilam thrust sheet, folding was accompanied by flexural slip between adjacent layers of S_0 stratification, as indicated by displacement of thrust faults along layer boundaries and ramping between adjacent layer boundaries. During the 1982–1983 surge thrusts developed in response to the intense shortening which occurred in association with the downglacier propagation of the surge front (Raymond *et al.* 1987, Sharp *et al.* 1988).

Although the occurrence of thrust faults is largely restricted to the terminal lobe of the glacier, the thrust faults vary considerably in their local dip (Fig. 3). This suggests that they may not be primary fractures, but rather are developed along pre-existing fracture surfaces. Their arcuate listric form suggests that the most likely parent structures are transverse crevasses and crevasse traces which have been sheared into a 'nested spoons' geometry during quiescent phase flow. The cross-cutting behaviour of thrust planes suggests that several generations of transverse fracture may be affected.

Progressive deformation of initially vertical transverse fractures in a regime of simple shear results in a progressive reduction in the dip of fractures, such that at a given location younger transverse crevasses are less steep than older ones. This trend is reinforced by the fact that surface ablation results in exposure of progressively deeper parts of the structures with increasing distance downglacier. As fractures are advected downglacier, they are progressively rotated into an orientation which approaches that of the expected direction of shear failure in a regime of flow-parallel compression. Furthermore, as ice thins downglacier the magnitude of the confining stress, which inhibits brittle failure, is reduced. Thus thrust faults are found mainly in the terminal lobe, where pre-existing fractures are orientated favourably to accommodate thrust displacements and brittle failure is possible. In slightly thicker ice, blind thrusts may occur at depth along the less steeply inclined sections of crevasse traces.

The pervasive nature of transverse crevassing during surges, and the fact that ice in the terminal lobe has been affected by several surges during its passage downglacier mean that it is impossible to trace the origin of the crevasse traces to specific parts of the glacier.

Crevassing during the 1982–1983 *surge*

Flow-parallel crevasses are also related to the phase of longitudinal shortening associated with the propagating surge front. The longitudinal crevasses formed after the thrust faults, and are therefore considered to be features characteristic of later in D_2. In the terminal lobe, crevasse formation lagged about one day behind thrust-faulting during the 1982–1983 surge (Sharp *et al.* 1988), but further upglacier flow-parallel crevasses were the principal structure formed during the episode of compression. Both primary fracture and reactivation of older crevasse traces appear to have occurred, the latter being particularly common in the terminal lobe, where ice has the longest and most complex deformation history and hence the highest density of pre-existing fractures.

Aerial photographs taken during the 1982–1983 surge show that the huge longitudinal chasms in the terminal lobe were formed by the linkage of small longitudinal crevasses within en échelon arrays (cf. Nicholson & Pollard 1985). These arrays developed on the south side of the glacier where flow diverges from the confined valley. They make an acute angle with the flow downstream and trend roughly WNW–ESE. Their association with diverging flow and transverse dilation implies that the initial cracks were tensile fractures. This inference is consistent with measurements of the relationship between the orientations of cracks and principal surface strain rates at the time of their formation (Sharp *et al.* 1988). The initially straight cracks within the array dilate in the diverging flow and are sheared into sigmoidal shapes, and the bridges between them bend and eventually fail when cross fractures form and adjacent en échelon cracks connect. In Beach's (1975) terms, these are Type 2 arrays which form by the nucleation of tensile cracks in a zone which may later undergo shearing deformation due to its lowered competence, rather than Type 1 arrays which form by the opening of shear fractures within a zone which is actively experiencing shear.

Other longitudinal chasms developed downglacier of Profile F by the collapse of highly fractured intercrevasse blocks (Raymond *et al.* 1987). Aerial photographs suggest that longitudinal collapse occurred particularly along the suture between tributary ice in the loop on the north side of the terminal lobe and ice in the main body of the glacier, suggesting that the suture is a zone of weakness.

The wave of longitudinal extension which moved downglacier behind the surge front caused transverse crevassing in the confined glacier. 1982–1983 surge phase transverse crevasses are only found in the confined glacier, reflecting the fact that ice further downglacier experienced only longitudinal shortening ahead of the surge front. Transverse crevasse traces in the terminal lobe (Fig. 3) were not active as extensional structures in the 1982–1983 surge, and are therefore products of earlier deformations.

Ice that experienced longitudinal shortening followed by longitudinal extension displays superimposed compressional and extensional structures and, in particular, both longitudinal and transverse fractures. The lag between the formation of these two orthogonal fracture sets was not directly observed during the surge, but centreline velocity profiles (Raymond 1984) suggest a typical lag of a few days between shortening and extension for ice in the lower part of the glacier. This sequence is not reflected in cross-cutting relationships, which are often inconsistent along strike. The longitudinal fractures did not therefore become completely inactive during the subsequent longitudinal extension, and continued to take up displacements during the formation of transverse crevasses.

Crevassing during quiescence

Between the end of the surge in August 1983 and the start of fieldwork in 1986, longitudinal crevasses formed at about 15.5 km from the head of the glacier, in ice which had been relatively inactive since the end of the surge. These crevasses opened along pre-existing fracture surfaces, which probably lowered the regional strain rate required for their formation.

The geometric relationships in marginal en échelon crevasse arrays which have formed since the end of the surge are typical of en échelon fractures within a ductile shear zone (Ramsay & Huber 1983, fig. 2.11), and reflect the sense of shear across the zones (dextral at the north margin, sinistral at the south). In some cases in marginal snow or firn, smaller scale en échelon arrays occur parallel to the cracks within the larger scale array. This arrangement, of en échelon arrays within en échelon arrays, indicates that, like the conjugate shears found in the terminal lobe, the crevasses in the larger scale array form as shear fractures rather than as tensile fractures.

Summary: structural development and longitudinal position

Ice flowing from the wide accumulation basins into the narrow confined glacier during quiescence becomes tightly folded towards the margins. As a result of this folding, sedimentary layering is transposed near the margins to a flow-parallel foliation (Fig. 7,i), but elsewhere is unusually well preserved into the terminal lobe. Ice deposited within the confined glacier acquires a shear cleavage parallel to the underlying foliation. A second, transverse listric foliation associated with overturned, downglacier-verging folds develops as ice approaches the terminal lobe as a result of surge-phase longitudinal shortening. As ice moves out into the broad terminal lobe and further shortening occurs, large-scale passive folds with steeply-dipping hinges develop.

Ice is fractured early in its history either as a result of quiescent phase crevassing high in the accumulation area or as a result of the first surge by which it is affected. Longitudinally and transversely orientated crevasses, once formed, are probably reactivated during later surge events. Transverse crevasse traces are reactivated as thrust faults in the terminal lobe.

IMPLICATIONS FOR GEOLOGICAL INTERPRETATION: STRUCTURE AND MECHANICS

In the light of the mechanical similarity outlined in the Introduction, it is not surprising that the range and assemblage of structures found at Variegated Glacier are comparable to those found in many thin-skinned fold-and-thrust belts. In particular, the patterns and types of structures are reminiscent of those found in the discrete, rapidly moving parts of thrust belts called surge zones (Coward 1982). These surge zones, which were identified as zones of enhanced deformation within the Heilam thrust sheet in northwest Scotland (Coward 1982, Fischer & Coward 1982), comprise a body of material similar in overall morphology to an idealized valley glacier, being bounded towards the foreland by an arcuate listric contractional fault and towards the hinterland by an arcuate listric extensional fault. In the foreland parts of these surge zones, as at Variegated Glacier during its surge, folds form due to layer parallel shortening. These folds, like the analogous folds in the terminal lobe of Variegated Glacier, have an arcuate pattern of outcrop and verge downstream. The main contrast between the overall configuration of structures in the surge zones and those at Variegated Glacier is that in the surge zones, the listric extensional faults can be traced into the strike slip faults and then into the contractional faults, which together then form a continuous detachment surface (Coward 1982, fig. 9). At Variegated Glacier the three sets of detachments are separate, if not mechanically independent.

One of the most exciting possibilities suggested by the work presented in this paper is that by analyzing and understanding the general characteristics of the relationships between flow mechanics and structural assemblages in this well understood tectonic setting, structural information can be used to infer flow mechanics—and hence mechanisms of emplacement—in other tectonic settings. This possibility is elucidated below, by first characterizing the structural patterns produced in each of the quiescent and surge phases of flow in the context of their mechanics, and then by outlining patterns that are characteristic of the alternation of the two.

In order to be able to make the link between flow type and structural pattern, the differences in structural pattern between quiescent and surge phases of flow that can be attributed to change in flow *type* must first be clearly identified and separated from those attributable to changes in flow *rate*. In the case of Variegated Glacier, the alternation between creep-dominated and sliding-dominated regimes coincides with an alternation between low and high strain rate regimes, respectively. This association arises from the fact that the strain rates occurring during the surge are sufficiently large to cause marginal and basal detachment, and thus to allow motion by sliding. The net effect is that brittle structural development is much more common in the surge phase than in the quiescent phase, and that ductile structural development is a more significant part of overall structural development in the quiescent phase.

Structural pattern and flow type

The creep-dominated quiescent phase of motion produces deformation which is generally non-coaxial. As a result, fractures that form during quiescence are curved at the time of their formation, such that longitudinally-trending crevasses splay downglacier and transverse crevasses are concave downglacier. Also, planar transverse features such as crevasses or crevasse traces become deformed during quiescence and obtain a three-dimensional concave upglacier geometry. This process is progressive, with the result that the oldest planar transverse features at a given location are those with the shallowest dips. Features that are diagnostic of a creep-dominated regime therefore comprise: (1) fractures that are curved at the time of their formation; (2) an upstream curvature of transversely-orientated planar features; (3) an inverse relationship between the age of a planar transverse structure and its dip at a given location; and (4) the presence of simple shear features towards the margins.

The sliding-dominated surge phase of motion produces deformation which is coaxial across the width of the glacier. As a result, both brittle and ductile structures are straight at the time of their formation, and aligned either parallel or orthogonal to the flow direction across the width of the glacier. Flow-parallel and flow-orthogonal crevasses are widespread, and overturned, downglacier-verging folds form towards the foreland with hingelines perpendicular to the flow direction. As a result of the large proportion of the total displacement of the ice that occurs as a result of sliding, pre-existing layering is well preserved. Features that are diagnostic of a sliding-dominated regime are therefore: (1) planar fractures; (2) flow-alignment (either parallel or perpendicular) of various structures at the time of their formation across the width of the deforming body, including fractures and fold hinges; and (3) persistence of pre-existing layering, despite very large strain rates and displacements. Pre-existing layering may persist in creep-dominated regimes, but generally where the total displacement is relatively small.

A particular characteristic of the sliding motion at Variegates Glacier is that it propagates through the glacier as a wave. The wave-like propagation of the sliding dislocation is evidenced by the zonation of extensional and compressional structural regimes.

Cumulative emplacement at Variegated Glacier occurs by a combination of creep and sliding, in which a periodic variation of sliding rate is superimposed on a fairly constant creep rate. At other glaciers and in other tectonic settings, sliding and creep may both occur with roughly constant magnitudes through time. In either case, a combination of the two flow processes is reflected in the presence of: (1) the coexistence of structures that were flow-aligned at the time of their formation with those that were flow-oblique at the time of their formation towards the margins; (2) transverse planar features with a range of curvatures, whereby the youngest of them are the least curved; and (3) the extended persistence of pre-existing layering.

The coexistence of flow-aligned and flow-oblique structures, especially towards the margins, is therefore characteristic of emplacement resulting from a combination of creep and sliding. Creep-dominated regimes produce a similar pattern as a result of the operation of cumulative simple shear on structures that are initially flow-oblique, but do not produce this range of orientations at the time of formation.

Structural pattern and flow rate

Ductile structural development reflects relatively slow strain rates, and brittle structural development reflects relatively high strain rates. In terms of the rates of emplacement, then, the coexistence of ductile structural development with very intense brittle structural development suggests that strain rate has been highly variable through time.

Structural pattern and overall flow regime

Together then, the combination of structures that are flow-aligned on formation and brittle in nature (that is, sliding–surge related) with structures that are flow-oblique on formation and ductile in nature (creep–quiescence related), as found at Variegated Glacier, indicates that the flow mechanism has varied through time, and that high strain rates occurred during sliding and relatively low strain rates during creep.

The variation of dominant flow mechanism and the corresponding variation of flow rate can therefore be fairly confidently inferred from overall structural pattern. However, it is considerably more difficult to infer a periodic oscillation of flow mechanisms/rates, rather than a simple linear variation in time. Perhaps the most significant indicators that together strongly suggest this period oscillation are: (1) a zonation of structural regimes, indicating the wave-like propagation of a sliding dislocation; and (2) the extended and unusual persistence of pre-existing layering resulting from a large proportion of the total displacement having been achieved by sliding.

Acknowledgements—This work was supported by The Royal Society, The Nuffield Foundation, The Royal Geographical Society, and Fitzwilliam and St. Edmund's Colleges, Cambridge. W. J. Lawson was supported by a NERC Research Studentship. Thanks are due to C. F. Raymond for the opportunity to work at Variegated Glacier and for access to unpublished data, and to Mike Ivers of Gulf Air Taxi, Yakutat for flying us there. The U.S. National Park Service granted permission for work in the Wrangell-St.Elias National Park, and the U.S. Forest Service for work in the Tongass National Forest. Clarence Summers of the Park Service provided invaluable support in Yakutat and beyond. The clarity of the paper has been substantially improved as a result of comments from Peter Hudleston.

REFERENCES

Allen, C. R., Kamb, W. B., Meier, M. F. & Sharp, R. P. 1960. Structure of the lower Blue Glacier, Washington. *J. Geol.* **68**, 601–625.

Beach, A. 1975. The geometry of en-echelon vein arrays. *Tectonophysics* **28**, 245–263.

Bindschadler, R., Harrison, W. D., Raymond, C. F. & Crosson, R. 1977. Geometry and dynamics of a surge-type glacier. *J. Glaciol.* **18**, 181–194.

Boyer, S. M. & Elliot, D. 1976. Thrust systems. *Bull. Am. Ass. Petrol. Geol.* **66**, 1196–1230.

Coward, M. P. 1981. Surge zones in the Moine thrust zone of NW Scotland. *J. Struct. Geol.* **4**, 247–256.

Driscoll, F. G. 1980. Formation of the Neoglacial Surge Moraines of the Klutlan Glacier, Yukon Territory, Canada. *Quat. Res.* **14**, 19–30.

Durney, D. W. & Ramsay, J. G. 1973. Incremental strains measured by syntectonic crystal growth. In: *Gravity and Tectonics* (edited by DeJong, K. A. & Scholten, R.). John Wiley, New York.

Fischer, M. W. & Coward, M. P. 1982. Strains and folds within thrust sheets: an analysis of the Heilam Sheet, northwest Scotland. *Tectonophysics* **88**, 291–312.

Gilbert, E. & Merle, O. 1987. Extrusion and spreading beyond a closing channel. *J. Struct. Geol.* **9**. 481–490.

Glen, J. W. 1955. The creep of polycrystalline ice. *Proc. R. Soc. Lond.* **A228**, 519–538.

Gretener, P. E. 1972. Thoughts on overthrust faulting in a layered sequence. *Bull. Can. Petrol. Geol.* **20**, 583–607.

Gretener, P. E. 1981. Pore pressure, discontinuities, isostasy and overthrusts. In: *Thrust and Nappe Tectonics* (edited by McClay, K. R. & Price, N. J.). *Spec. Publs geol. Soc. Lond.* **9**, 33–39.

Hambrey, M. J. 1975. The origin of foliation in glaciers: evidence from some Norwegian examples. *J. Glaciol.* **14**,(70), 181–185.

Hambrey, M. J. & Milnes, A. G. 1977. Structural geology of an Alpine glacier (Griesgletscher, Valais, Switzerland). *Eclog. geol. Helv.* **70**, 667–684.

Hambrey, M. J., Milnes, A. G. & Siegenthaler, H. 1980. Dynamics and structure of Griesgletscher, Switzerland. *J. Glaciol.* **25**,(92), 215–228.

Hambrey, M. J. & Müller, F. 1978. Structures and ice deformation in the White Glacier, Axel Heiberg Island, Northwest Territories, Canada. *J. Glaciol.* **20**,(82), 41–66.

Harland, W. B. 1971. Tectonic transpression in Caledonian Spitsbergen. *Geol. Mag.* **108**, 27–42.

Hobbs, B. E., Means, W. D. & Williams, P. F. 1976. *An Outline of Structural Geology*. John Wiley, Chichester.

Hooke, R. Le B. & Hudleston, P. 1978. Origin of foliation in glaciers. *J. Glaciol.* **20**(83), 285–299.

Hubbert, M. K. & Rubey, W. W. 1959. Role of fluid pressure in the mechanics of overthrust faulting. *Bull. geol. Soc. Am.* **70**, 115–166.

Hudleston, P. J. 1976. Recumbent folding in the base of the Barnes Ice Cap, Baffin Island, Northwest Territories, Canada. *Bull. geol. Soc. Am.* **87**, 1684–1692.

Hudleston, P. J. 1980. The progressive development of inhomogenous shear and crystallographic fabric in glacial ice. *J. Struct. Geol.* **2**, 189–196.

Kamb, B., Raymond, C. F., Harrison, W. D., Engelhardt, H. F., Echelmeyer, K. A., Humphrey, N., Brugman, M. & Pfeffer, T. 1985. Glacier surge mechanism: 1982–1983 surge of Variegated Glacier. *Science* **227**, 469–479.

Lawson, W. J. 1990. The structural evolution of Variegated Glacier, Alaska. Unpublished Ph.D. thesis, University of Cambridge, Cambridge.

Meier, M. F. 1960. Mode of flow of Saskatchewan Glacier, Alberta, Canada. *Prof. Pap. U.S. geol. Surv.* **351.**

Meier, M. F. & Post, A. 1969. What are glacier surges? *Can. J. Earth Sci.* **6**, 807–817.

Nicholson, R. & Pollard, D. D. 1985. Dilation and linkage of echelon cracks. *J. Struct. Geol.* **7**, 583–590.

Nye, J. F. 1952. The mechanics of glacier flow. *J. Glaciol.* **2**,(12), 82–93.

Paterson, W. S. B. 1981. *The Physics of Glaciers* (2nd edn). Pergamon Press, Oxford.

Post, A. 1972. Periodic surge origin of folded medial moraines on Bering piedmont glacier, Alaska. *J. Glaciol.* **11**, 219–226.

Price, R. A. 1988. The mechanical paradox of large overthrusts. *Bull. geol. Soc. Am.* **100**, 1898–1908.

Ragan, D. M. 1969. Structures at the base of an icefall. *J. Geol.* **77**, 647–667.

Ramsay, J. G. 1967. *Folding and Fracturing of Rocks*. McGraw-Hill, New York.

Ramsay, J. G. 1980. The crack–seal mechanism of rock deformation. *Nature* **284**, 135–139.

Ramsay, J. G. & Huber, M. I. 1983. *The Techniques of Modern Structural Geology, Volume* 1: *Strain Analysis*. Academic Press, London.

Raymond, C. F. 1984. Interim report on surges of Variegated Glacier. Fourth year report on NSF grant number EAR-79-19530. Unpublished Report.

Raymond, C. F. & Harrison, W. D. 1988. Evolution of Variegated Glacier, Alaska, U.S.A., prior to its surge. *J. Glaciol.* **34**, 154–169.

Raymond, C. F., Johannesson, T., Pfeffer, T. & Sharp, M. 1987. Propagation of a glacier surge into stagnant ice. *J. geophys. Res.* **92**, 9037–9049.

Röthlisberger, H. & Iken, A. 1981. Plucking as an effect of water-pressure variations at the glacier bed. *Ann. Glaciol.* **2**, 57–62.

Rutter, E. H. 1986. On the nomenclature of mode of failure transitions in rocks. *Tectonophysics* **122**, 381–387.

Sanderson, D. J. & Marchini, W. R. D. 1984. Transpression. *J. Struct. Geol.* **6**, 449–458.

Sharp, M. J., Lawson, W. & Anderson, R. S. 1988. Tectonic processes in a surge-type glacier. *J. Struct. Geol.* **10**, 499–515.

Stenborg, T. 1968. Glacier drainage connected with ice structures. *Geog. Ann.* **50A**, 25–53.