Structures and Fabrics in Glacial Ice: A Review

PETER J. HUDLESTON

Department of Geology and Geophysics, University of Minnesota, Minneapolis, MN

55455, U.S.A.

hudle001@umn.edu

Abstract

Glaciers, ice sheets and ice caps represent tectonic systems driven by gravity. Their movement can be studied in real time and the rheological properties and strength of ice determined from laboratory experiments and field measurements. All glacial ice has primary stratification, exhibited by variations in grain size, bubble content and debris content. As it deforms, with deformation dominated by plastic flow and recrystallization, accompanied locally by fracture under tension, a suite of structures develops that reflects the primary fabric of the ice and the anisotropy that develops as a result of cumulative deformation. Initial variations in solid impurity content and strain dependent anisotropy as a result of a crystallographic fabric give rise to effective viscosity increases or decreases compared to isotropic polycrystalline ice of about a factor of ten. Foliation develops from inherited (mostly stratification) or introduced (mostly ice veins or fracture traces) fabric elements and from dynamic recrystallization. It is largely dependent on the accumulated strain, which is highest at the base and near the margins of glaciers, ice

24 sheets and ice streams. Folds develop largely passively due to initial amplification of
25 irregularities in the primary stratification, to variations in flow with time or to
26 inhomogeneous flow associated with shear zones and ductile accommodation
27 around open fractures. Buckle folds and boudinage, mostly on a small scale, occur
28 where viscosity contrast is large, mostly in basal ice. Thrusting and wrench faulting
29 are documented in surging glaciers but theoretically most unlikely and rare or
30 absent elsewhere. Many structures interpreted as faults are not due to shear failure
31 but rather result from shear displacements during opening and closing of tensile
32 fractures.

33 **1. Introduction**

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35 Glaciers have particular appeal to structural geologists, providing natural 'tectonic'
36 systems in which a variety of brittle and ductile structures develop under known
37 driving forces and at rates that can be directly measured. Analogies between
38 structural development in ice and in rocks have been made by many authors (e.g.
39 Wegmann, 1963; Hambrey and Milnes, 1977; Kamb et al., 1985; Hudleston, 1992;
40 Lawson et al., 1994), and have been drawn from the scale of the crystal – ice in some
41 ways being a good analogue for quartz (e.g. Wilson and Russell Head, 1979; Wilson
42 et al., 2014), to individual structures such as folds (e.g. Hudleston, 1977b) and
43 boudinage (Hambrey and Milnes, 1975), to tectonic systems (e.g. Herbst and
44 Neubauer, 2000).

One benefit of studying structures in ice is that the rheology is relatively well understood, with a well-established flow law for isotropic polycrystalline ice and with recent advances that allow characterization of the rheology at low stresses and with preferred crystallographic orientation (e.g. Pettit et al., 2011; Treverrow et al., 2012; Wilson and Peternell, 2012). Considerable experimental work has also been done on the effect on rheology of solid particles dispersed in ice and behavior of ice-saturated sediment (Moore, 2014). Much recent work has been done on the microstructure of ice and the development of preferred crystallographic orientation with strain (e.g. Duval et al., 2010; Montagnat et al., 2012, 2014), with particular focus on the Greenland and Antarctic ice caps. In those locations, understanding the flow history has important implications for interpreting proxy signals for climate change, since deformation overprints and disrupts stratigraphy deep under ice caps due to the nature of the ice flow.

Geophysical methods are employed to determine ice flow and material properties at depth within and over large areas of ice sheets and glaciers. Radio-echo sounding (e.g. Drews et al., 2009) and ground-penetrating radar (e.g. Vaughan et al., 1999; Holschuh et al., 2014) are used to determine ice thickness and elucidate internal structure. Not only can these techniques pick up internal reflectors representing stratification (isochrones), they may also detect changes in crystallographic fabric (Eisen et al., 2007), as may seismic wave velocity analysis (Diez et al., 2104). Interferometric synthetic aperture radar (InSAR), together with optical-image-tracking methods provided by satellite imagery and GPS, are used to measure

surface velocities over large areas and over time frames of from seconds to years (e.g. Joughin et al., 1999, 2010). This is especially useful for studying ice streams in ice sheets.

Hambrey and Lawson (2000) have provided a comprehensive summary of structural studies on glaciers from the 1950s until the turn of the century. Good examples of detailed studies are those of Allen et al. (1960) on the Lower Blue Glacier, Washington; Meier (1964) on the Athabasca Glacier, Alberta; Hambrey (1976a, 1976b) on Charles Rabots Bre, Norway; Hambrey and Milnes (1977) on Griesgletscher, Switzerland; Hambrey and Müller (1978) on White Glacier, Axel Heiberg Island, Canada; Sharp et al. (1994) and Lawson (1996, 1997) on the Variegated Glacier, Alaska; and Goodsell et al. (2005) on Haut Glacier d'Arolla, Switzerland.

The approach taken in this paper is first to provide brief reviews of ice rheology, the kinematics of flow in glaciers and larger ice bodies (Fig. 1) and the crystallographic fabric that develops as a result of this flow. The bulk of the paper is devoted to descriptions of the various structural and fabric elements found and mapped in glacial ice and a discussion of the processes involved in their development. The paper concludes with a brief section discussing the assembly and sequence of structures and fabrics developed in glacial 'tectonic' systems.

2. Rheology

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93 Data from experiments and theoretical considerations suggest that isotropic
94 polycrystalline ice deforms under secondary creep according to a relationship of the
95 form:

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97
$$\dot{\epsilon}_{ij} = AS_{eff}^{n-1} S_{ij} \quad 1)$$

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99 (cf., Cuffey and Paterson, 2010, eq. 3.22), where $\dot{\epsilon}_{ij}$ and S_{ij} are components of the
100 strain rate and stress tensors, respectively and S_{eff} is the effective stress - the
101 second invariant of the stress tensor. This is the generalized form (Nye, 1953) of
102 Glen's (1955) flow law relating shear strain rate to shear stress. The value of the
103 constant A is affected by various physical and chemical properties of the ice and
104 reported values vary widely (see Cuffey and Paterson, 2010, sect. 3.4.5). It primarily
105 incorporates a strong temperature dependence through an Arrhenius relationship.
106 Reported values of n also vary widely (Weertman, 1983; Cuffey and Paterson, 2010,
107 sect. 3.4.4). A value of $n=3$, reported in many studies, is consistent with theoretical
108 considerations of intracrystalline slip on the basal plane (Weertman, 1983) and is
109 commonly preferred for modeling ice flow. Values of $n < 3$ and as low as 1 are
110 reported at low stresses and for fine grain sizes and may reflect a change to grain-
111 size sensitive creep mechanisms (e.g. Goldsby and Kohlstedt, 1997; Montagnat and
112 Duval, 2000). Flow laws that include a term expressing a linear stress-strain

relationship have been successfully applied to low deviatoric stress regions such as near ice divides (e.g. Pettit and Waddington, 2003; Pettit et al., 2011).

Ice crystallographic fabric evolves during deformation of initially isotropic ice (e.g. Alley, 1992), and secondary creep is in fact transient rather than steady state. It appears that under a constant stress regime of either uniaxial compression or simple shear and with deformation accompanied by dynamic recrystallization, a steady state tertiary creep regime may be achieved after about 10% strain (Fig. 2, Budd and Jacka, 1989; Treverrow et al., 2012). The nature of the crystallographic fabric and stress-strain rate relationship depend on the kinematics of flow (Duval, 1981; Wilson, 2000; Budd et al., 2013). Tertiary creep is characterized by dynamic recrystallization (Duval et al., 2010).

As crystallographic fabric develops the viscosity becomes anisotropic, with the ice becoming softer in some stress configurations and harder in others. Equation 1 then becomes inadequate for describing the rheology. Various ways of modifying the flow law to take anisotropy into effect have been suggested. One way is to introduce an enhancement factor into the equation, which then becomes.

$$\dot{\epsilon}_{ij} = EAS_{eff}^{n-1}S_{ij} \quad 2)$$

(cf., Cuffey and Paterson, 2010, eq. 3.35), where E is the factor by which the strain rate deviates from that predicted by the flow law for isotropic ice. The value of E

depends on the nature and intensity of the fabric, and is about twice as large in simple shear as in uniaxial compression (Treverrow et al., 2012). It can reach a value of about 10 in simple shear (see Fig 3 and the discussion in Cuffey and Paterson, chap. 3).

In the past twenty years considerable attention has been paid to modeling flow in polycrystalline ice that treats deformation on multiple scales up to that of ice sheets. In these models crystallographic fabric evolves with deformation. Examples are Azuma (1994), Castelnau et al. (1996), Thorsteinsson (2002) and Pettit et al. (2007). A comprehensive review of the various approaches taken to the modeling of ice deformation and fabric evolution at different scales is given by Montagnat et al. (2014).

2.1 Effect of solid impurities

Glacial ice frequently contains debris, ranging in particle size from dust to boulder, usually well dispersed except near the base of glaciers, ice caps and ice sheets where debris rich layers are frequently found. Debris enters the ice in several ways. It does so as wind born particles on the surface in the accumulation zone, where it occurs in trace amounts that help create annual markers of stratification when the ice is later exposed in the ablation zone. Debris also may also be deposited on the ice surface near the sides of glaciers, as a result of mass wasting processes affecting the rock walls or deposition from supraglacial streams. Rock avalanches may produce

159 local concentrations of coarse debris. High concentrations of debris are found along
160 the base of glaciers or ice caps, where it may be entrained by several processes.
161 These processes have received much attention by glaciologists, since so much of
162 glacier movement occurs at the base of glaciers, either by slip on the bedrock or
163 shear within the basal ice or sediment below the ice. There is almost a complete
164 range in composition from 100% ice to close packing of debris particles with ice
165 filling the interstices. Basal ice is typically strongly layered, due either to shear
166 within the ice/debris mixtures or the process by which the debris is entrained, (e.g.
167 Knight, 1995; Samyn et al., 2008).

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169 A comprehensive review of the effect of debris on ice rheology and strength is given
170 by Moore (2014), who considers the deformational behavior of materials over the
171 complete range of debris-ice mixtures. As expected, the inclusion of solid particles
172 has a pronounced effect on rheology, most typically by strengthening the ice and
173 increasing the viscosity (e.g. Hooke et al., 1972). Under certain conditions, however,
174 as evidenced both in nature (e.g. Echelmeyer and Wang, 1987) and in experiments
175 (e.g. Song et al., 2005, 2008) the presence of dispersed solid particles weakens
176 rather than strengthen the ice, and the reason for this duality of behavior has been
177 much debated (see discussion in Moore, 2014). It appears mainly to be through the
178 influence of the particles on recrystallized grain size and grain-boundary sliding
179 (Durand et al., 2006). Weakening of ice-debris mixtures occurs at relatively high
180 temperatures when unfrozen water exists along crystal and rock-ice interfaces. At
181 concentrations of above about 40% debris by volume, frictional interaction among

particles becomes important and Mohr-Coulomb (pressure dependent) behavior occurs, modified by the presence of the interstitial ice (Moore, 2014).

The same mixture of ice and sediment can be stronger than pure ice under some conditions and weaker under others. For example, in experiments by Bray (2013), ice-rich silt at -1°C was weaker than pure ice at low stresses and stronger than pure ice at stresses above a threshold of about 220 kPa. Since this threshold stress is at the high end of the range of shear stresses expected for basal glacial ice (Cuffey and Paterson, 2010. Sect.8.2), it suggests interesting possibilities for structural development where such material is interlayered with pure ice.

Overall, it appears from the data available (Moore, 2014) that the presence of debris in ice may increase or decrease its resistance to shear by up to about a factor of about 10, which is also approximately the factor by which viscosity may be increased or lowered as a result of fabric-induced anisotropy (Fig. 3).

3. Flow Field and Kinematics

Surface velocities can be measured directly on glaciers and ice caps, and internal velocities measured by use of boreholes (see Hooke, 2005, ch. 5). Typical maximum horizontal velocities in valley glaciers are on the order of tens of meters per year, while in surging glaciers and ice streams in Antarctica and Greenland the velocities are one or more orders

of magnitude faster (see Cuffey and Paterson, 2010, table 8.2). Generalized flow patterns for valley glaciers and ice caps are illustrated schematically in Fig. 1.

Basal ice may slide over or be frozen to its substrate, and basal conditions are different under temperate glaciers than under polar glaciers and ice sheets. Sliding is inhibited if the ice is frozen to the substrate (Hutter and Olunyolo, 1980), although the effect may be limited if the substrate is unlithified (Echelmeyer and Wang, 1987; Moore et al., 2011). Sliding velocities can be measured directly in some instances and the phenomena of local pressure melting and regelation (refreezing) around obstacles observed (Kamb and LaChapelle, 1964; Cohen et al., 2000; Iverson et al., 2007).

Velocity profiles in valley glaciers are generally what would be expected for a material deforming according to a non-linear flow law, with high gradients near the margins and the base (Fig. 4, Raymond, 1971a, b), although there may be local complexity even when the overall glacier geometry is straightforward (e.g. Harper et al., 2001). In ice sheets, ice divides may move with time and ice streams change in position and intensity (Voight et al., 2003; Siegert et al., 2004; Jacobson and Waddington, 2005).

Knowing the velocity field allows strain rates to be calculated. Longitudinal strain rates are generally extensional in the accumulation zone and contractional in the ablation zone of a valley glacier (see Fig. 1). Shear strain rate is high along the margins and at the base of glaciers and in the margins of ice streams (e.g. Clarke, 1987) . As an example, the velocity profiles in Fig. 4 imply shear strain rates of about 0.2 a^{-1} ($7.0 \times 10^{-9} \text{ s}^{-1}$) and 0.1 a^{-1} ($2.5 \times 10^{-9} \text{ s}^{-1}$).

1) along the sides and base respectively of the Athabasca glacier. Longitudinal strain rates at the same location are negative and an order of magnitude smaller than these (Raymond, 1971b).

Estimates of shear stress at the base of glaciers and ice caps can be made from ice thickness and surface slope (Cuffey and Paterson, chap. 8). Typical values are 50-200 kPa (Cuffey and Paterson, 2010, table 8.3). Values an order of magnitude lower are found in water-saturated sediment below some ice streams (Joughin et al., 2004) .

Flow variations occur on all time scales, in part related to changes in weather and climate (e.g. Hooke et al., 1989; Keller and Blatter, 2012) and in part due to seasonal changes and episodic surges, which are discussed later. Steady state is never achieved, but may be approached over certain time intervals and on some scales. It is useful to consider steady state as a reference frame for understanding structural development.

Assuming steady state flow, measured or modeled velocities can be used to estimate finite strains within a glacier or ice cap (Figs. 5, 6, 7) (Milnes and Hambrey, 1976; Hambrey and Milnes, 1977; Hambrey et al., 1980; Hudleston and Hooke, 1980; Budd and Jacka, 1989; Hubbard and Hubbard, 2000). In ice caps and ice sheets, cumulative strain changes from uniaxial shortening at the centers of domes or pure shear at divides, with the minimum strain vertical, to being progressively more non-coaxial with depth and distance from the center or divide (Figs. 5,6) (Hudleston and Hooke, 1980). Strongly non-coaxial strain is found in basal ice and along the margins of glaciers and ice streams. In regions where flow

is two-dimensional, basal strain will accumulate under simple shear. In radial flow (as in Fig. 1a), cumulative strain will be of flattening type. In valley glaciers, in general the strains will be close to coaxial along the centerline and strongly non-coaxial along the base and sides, locally approaching simple shear (Fig. 7). The patterns described above are subject to modification when the flow deviates substantially from steady state. The consequences of this for structures are discussed in section 5.

4. Crystallographic Fabric

Considerable attention has been paid in recent years to the development of crystallographic fabric (lattice preferred orientation, or LPO), especially in ice sheets, with recognition of the significant effect fabric has on the flow of ice, imparting anisotropy to the viscosity (Azuma, 1994; Alley, 1988; Castlenau et al., 1996). A complete analysis of LPO includes a -axes, which can be determined using such techniques as X-ray Laue diffraction (e.g. Weikusat et al., 2011). An excellent treatment of crystallographic fabric, set in a broader discussion of microfabric deformation mechanisms and recrystallization, is given by Faria et al. (2014a, 2014b) in their review of microstructure of polar ice.

A brief description of typical c -axis fabrics is given here

Fabric anisotropy develops due to intracrystalline slip on lattice planes during deformation. In ice, slip on the basal plane is dominant, causing these planes in individual crystals to rotate towards being perpendicular to the maximum shortening direction (e.g. Alley, 1988, 1992). This produces a small circle distribution of c -axes around the axis of

uniaxial compression at the center of ice domes (Fig. 8a), a bimodal pattern symmetric about the compression axis in pure shear, as found at ice ridges (Fig. 8b), and an asymmetric bimodal pattern in simple shear, with the maxima lying in a plane perpendicular to the shear plane and parallel to the shear direction, as found in shear zones deep in ice sheets and near the base of polar glaciers (Fig.8c). Fabric strength intensifies with increasing strain (e.g. Hudleston, 1977a; Montagnat et al., 2012) and may achieve steady state with increasing deformation as fabric enhancing lattice rotations are balanced by fabric weakening recrystallization. In simple shear and at large shear strains the weaker maximum may disappear (Fig. 8d) as a result of recrystallization. Debris particles, if present even in small amounts, may weaken the development of a fabric (Song et al., 2005).

In valley glaciers and in the deeper parts of ice sheets, in coarse-grained ice that has undergone significant recrystallization and often subsequent annealing, multiple maximum fabrics, often with a four-fold diamond arrangement, are common (Fig. 8e) (Kamb, 1959; Budd, 1972; Gow and Williamson, 1976; Hambrey, 1976b). These may result from a modification of a girdle fabric, and several mechanisms have been proposed to account for them, with twinning likely playing a role (Matsuda and Wakahama, 1978; Faria et al., 2014b).

The planes and axes of symmetry of the crystallographic fabric may or may not be geometrically related to the foliation with which the fabric is associated (Budd, 1972; Hambrey, 1976b, 1977a). Where it is, the pole to the foliation typically is perpendicular to

a *c*-axis maximum or to the center of a multi-maxima fabric (e.g. Kamb, 1959; Allen et al., 1960; Hambrey, 1976b).

5. Mappable Structural and Fabric Elements in Glaciers

Glacial ice contains several distinctive fabric and structural elements that in part are imparted to the ice as it forms and in part develop and are modified as a result of deformation accompanied by recrystallization. These elements are defined by variations in grain size, bubble content, and impurity content. In addition, there may be a grain shape fabric (shape preferred orientation or SPO) and bubble fabric (bubbles with preferred shape and orientation). As in deformed rocks, there are both planar and linear fabric elements.

5.1 Primary Stratification

Primary stratification or bedding is present in the accumulation zones of all glaciers and ice caps. It is manifest by variations in crystal size and bubble content that reflect the changing conditions related to snow deposition and the transition through firn to glacial ice. The nature of the transition differs in temperate glaciers, where melting and refreezing occur (e.g. Kawashima and Yamada, 1997), and polar glaciers and ice caps, where the transition is largely in a dry state (e.g. Alley et al, 1982). In the latter case initial bubble concentration is related to temperature and snow accumulation rate during firnification (Spencer et al., 2006). Dust may be

seasonally concentrated on the surface and often serves to mark annual layering. Also, in valley glaciers debris carried onto the glacier from the valley sides may accumulate locally on the surface in the accumulation zone and this angular englacial material can provide evidence of bedding when later exposed, often folded, in the ablation zone (e.g. Hambrey and Dowdeswell, 1997; Hubbard and Hubbard, 2000; Roberson, 2008). Bedding is the only fabric element seen at the front of the smallest glaciers (e.g. Fig. 9a).

Besides deposition of layers in the upper parts of glaciers and ice caps, irregularities in topography and local conditions mean that net accumulation may occur in other places, giving rising to local unconformities, an example of which is shown in Fig. 9b.

Primary stratification may be traced throughout the length of small glaciers (e.g. Hambrey, 1976b), although not necessarily as continuous strata, and its orientation down glacier may be tracked using flow models and compared with surface exposures (Hubbard and Hubbard, 2000). In larger glaciers, it becomes difficult to distinguish primary stratification from foliation (see below).

5.2 Extensional fractures and veins

Brittle failure of ice under extension at the surface of glaciers and ice caps occurs in virtually all glaciers, producing crevasses if the fractures widen enough. These are predominantly mode I fractures (Lawn, 1993) that are initially perpendicular to the ice

surface and perpendicular to the direction of maximum extensional strain rate.

Crevasses are typically transverse to valley glaciers where developed at or close to the mid line in regions of longitudinal extensional strain rate and arranged en echelon to the glacier sides close to the ice margins or edges of ice streams (Marmo and Wilson, 1998), reflecting the dominance of shear stress at the glacier margins or edges of ice streams.

The classical patterns are illustrated in Fig. 10 (after Nye, 1952). Natural examples are shown in outcrop and on a map in Figs. 11 and 12., and in general they match Nye's predictions well. Deviations of crevasse orientations from perpendicularity to the maximum extensional strain rate, as is the case locally in Fig. 12, occur due to rotation of the crevasses once formed (and before closure). In addition, stresses are reoriented around crevasses once formed, such that further fracture development, if it occurs, may lead to complex patterns. This is especially true in icefalls (Fig. 11c) and in surging glaciers (Lawson et al., 2000). Estimates of extensional strain rates associated with tensile fracturing based on field measurements are quite variable, with values generally being higher in temperate glaciers than in polar glaciers. Meier et al. (1974) estimated values of 0.03 a^{-1} (ca. 10^{-9} s^{-1}) for the temperate Blue Glacier and Hambrey and Müller (1978) estimated values as low as 0.004 a^{-1} (ca. 10^{-10} s^{-1}) for the polar White Glacier, while noting that strain rates locally as high as 0.16 a^{-1} (ca. $5 \times 10^{-9} \text{ s}^{-1}$) were not associated with fracturing. Crevasse patterns have been used to characterize the flow and flow history in fast moving glaciers (Herzfeld et al., 2004)

There is still uncertainty about how deep crevasses penetrate in glaciers and ice caps (Hooke, 2005; Mottram and Benn, 2009; Cuffey and Paterson, 2010, p.449).

The vertical normal stress, zero at the surface, increases linearly with depth and thus the difference between the vertical and horizontal normal stresses (the latter being zero – or more correctly atmospheric pressure - perpendicular to a crevasse wall at all depths) also increases and at some depth must be sufficient to close the fracture by plastic flow. The particular circumstances will depend on the rate of propagation of the fracture and tensile strength of the ice. Water filled crevasses can theoretically extend to glacier beds provided water supply is maintained (Weertman, 1973). Evidence of fractures extending deep into glaciers is provided by the presence of veins of clear ice, presumed to be traces of old fractures, far downstream in the ablation zone where ice is exposed (Hambrey, 1976a; Goodsell et al., 2005).

From the moment of formation onwards, fractures in marginal ice are subject to displacement downstream and rotation induced by the velocity gradients of the flow (see Hubbard and Hubbard, 2000). This displacement or rotation will bring a fracture into a position or orientation such that the strain rate across it becomes compressional, which eventually causes closure and creation of a vein from transformed snow or freezing of melt water entering the fracture. Healed fractures, or ‘fracture traces’ in the terminology of Hambrey and Milnes (1977), filled with clear ice are the most likely origin for the “blue bands” seen so abundantly in valley glaciers. During rotation, and before the fracture finally heals and becomes a vein or fracture trace, the lack of shear stress along the crevasse walls and consequent modification of the stress and strain rate fields around the fracture causes fracture-

and strike-parallel displacement of one wall with respect to the other. Relative “strike-slip” movement of one wall of an open crevasse with respect to the other has been directly recorded (e.g. Kehle, 1964). It seems most unlikely that such crevasses originated as shear fractures.

Some fractures may not develop into open crevasses, but rather develop directly as veins in a manner similar to that described as crack-seal in rocks (Durney and Ramsay, 1973; Ramsay, 1980), with elongate crystals of bubble free ice perpendicular to the walls growing in continuity with crystals in the walls, sometimes with a train of bubbles along the center or along the vein margins (Hambrey and Müller, 1978; Hambrey et al., 1980; Lawson et al., 1994; Voigt et al., 2003).

In marginal ice, generation of new crevasses in ice containing older healed fractures, appearing as veins, leads to younger veins cutting across older ones and the kind of fracture and vein relationships shown in Fig. 13, similar to what is found in rocks (Passchier and Trouw, 1996, fig. 6.11). The crosscutting relationships are systematic, as older veins are rotated and cut by younger ones. Patrick et al. (2003) have described this phenomenon associated with en echelon arrays of sigmoidal fracture traces in the Sørsdal glacier in East Antarctica (Fig. 14). The distribution of orientations of fracture traces and veins in a straight segment of the marginal zone of Storglaciären in Sweden is shown in Fig. 15a. This can be compared with orientation distributions produced by a simple model of fracture generation and

subsequent passive rotation by simple shear. In this case, if the fractures originated either at 45° to the glacier margin (Fig. 15b) or angles up to 90° (Fig. 15c) in the region of extensional flow up glacier in the accumulation zone, significant displacement and shear would concentrate the traces at small angles to the glacier margin. The data (Fig. 15a) are consistent with most fractures having formed between 45 ° and 90 ° to the margin.

Steep transverse fractures that form along the mid lines of glaciers will tend to have a crescent shape in map view, initially convex up glacier (Fig. 10), and may close to form veins in regions of compressional strain, and become convex down glacier with inward dips that become more gentle downstream, the crescent shape and shallowing dips reflecting the cross-glacier and vertical velocity gradients, respectively. Numerical modeling may be used to predict the three-dimensional form of healed fracture traces down glacier (Hubbard and Hubbard, 2000).

Shear fractures (mode II or III fractures) (Lawn, 1993) are much rarer than extensional fractures in glaciers, and many of the structures described as faults or thrusts likely have an alternative origin. These will be discussed in a later section.

5.3 Foliation

The most prominent fabric element visible in the ablation zones of most valley glaciers is a foliation, defined largely by variations in crystal size, crystal shape, and

bubble concentration, often in the form of distinct, typically discontinuous, bands (Fig. 16a, b)(e.g. Allen et al., 1960; Hambrey, 1975; Hambrey, 1976b, 1977a; Hambrey and Milnes, 1977) and on a finer scale by bubble distribution (Fig. 16c, d, Fig.17) and sometimes bubble shape (e.g. Hudleston, 1977a). A crystallographic lattice preferred orientation (LPO) is commonly developed in foliated ice and may or may not have a well-defined geometrical relationship to the foliation (e.g. Kamb, 1959; Budd, 1972). The foliation is most strongly developed at the base and along the margins of glaciers and along sutures or septa formed by the joining of two ice streams. Along margins or along septa it is steep to vertical and parallel to the glacier margins. It is common in the literature to refer to foliation in valley glaciers as either being 'longitudinal' or 'transverse,' the latter typically developing at the base of an icefall (Allen et al., 1960; Ragan, 1969; Hambrey and Milnes, 1977; Hambrey et al., 1980). Overall patterns of transverse foliation in valley glaciers have been described as arcuate at the surface with dips inward and up-glacier (e.g. Hambrey, 1977b), with a general appearance of one or more sets of "nested spoons" (Allen et al., 1960; Rutter, 1965). Cross-cutting foliations, often of several generations, have been described in many glaciers (e.g. Ragan, 1969; Hambrey and Milnes, 1977; Hambrey et al., 1980).

As in rocks, foliation may take different forms in glacier ice, and a number of processes contribute to its formation. In all cases however, and by definition of how the term 'foliation' is used, deformation is involved, with the most intense foliation found where the deformation is greatest. Early workers looked for a relationship

between foliation and the state of stress in the ice where the foliation was observed; that is, they expected foliation to form perpendicular to the direction of maximum compressive stress (Glen, 1956; Meier, 1960) or parallel to the direction of maximum shear stress (Ragan, 1969). Allen et al. (1960) considered foliation to be a wholly secondary structure formed as a result of ice deformation, and in the Lower Blue Glacier, Washington, they found it to be developed in a vertical orientation in regions of high compressive stress at the base of an ice fall and then to be modified passively to take up the arcuate nested spoon pattern observed downstream. Pfeffer (1992) considered transverse foliation in the terminal lobe of the Variegated glacier, to have formed by bubble migration as a result of high compressive stress during the 1982-1983 surge. Ragan (1969) considered the foliation to be continually reformed parallel to the plane of maximum shearing stress as earlier foliations rotated away from this orientation, producing serial cross-cutting relationships.

Observations support the view of Hambrey and Milnes (1977), Hooke and Hudleston (1978) and Hambrey et al. (1980) who argued that total strain and strain history, rather than stress, is the most important factor controlling the development of foliation, and they identified several structural or fabric elements that taken together form foliation (Fig. 18), with a composite foliation becoming parallel to the XY plane ($X \geq Y \geq Z$) of the cumulative strain after large deformation. Hambrey (1975) considered that foliation was often developed from sedimentary layering, and a number of subsequent studies have borne this out (e.g. Roberson, 2008; Jennings et

al., 2014). Hooke and Hudleston (1978) argued that bedding and veins or fracture traces formed from the closure of fractures were the two principal components that contribute to foliation. Bedding and veins are both manifest largely in the distribution of bubbles, with variations in bubble concentration, often associated with variations in grain size. An example of ice with bedding well preserved and with a foliation expressed on the grain scale by grain shape and bubble concentration, and these both cut by later veins, is shown in outcrop in Fig. 16d. In thin section and on a fine scale the foliation may appear as lenses and diffuse patches of varying bubble concentration. In Fig. 17a such a foliation is cut by a clear ice vein, but deformation and recrystallization accompanied by some reorganization of bubbles since vein formation have modified the vein edges and made them ragged. Azuma et al. (2012) documented the effect of bubbles impeding grain growth and being swept by moving grain boundaries. In crossed polars, a shape preferred orientation of the ice crystals, parallel to the foliation, can be seen (Fig. 17b). In this sample the bubbles themselves are small and remain spherical. In other cases, especially in cold glaciers, the bubbles themselves may be elongate and their alignment may contribute to the foliation (Gow, 1968; Hudleston, 1977a).

If different fabric elements contribute to foliation, cross-cutting foliations may indicate: 1) a strain insufficiently large to bring all elements into parallelism (e.g. Fig. 16d); 2) additional cross-cutting elements being introduced after a strong fabric has already developed (e.g. Fig. 16c); or 3) in non-coaxial deformation, the existence of fabric elements that are sensitive to only the most recent part of the strain history

(see section 5.5 on ductile shear zones) and thus forming a foliation oblique to those elements reflecting the total strain. Another way of expressing this is to say that different fabric elements will reflect different segments of the strain history. Primary stratification or bedding is a feature, as long as it is preserved, whose orientation reflects the total strain experienced by the ice, whereas veins record only the strain accumulated since vein formation. Ice crystal size and shape (SPO) and small-scale distribution of bubbles, being modified continuously by dynamic recrystallization, reflect the latest part of the strain history, and thus may form fabric elements oblique to a foliation defined otherwise by variations in bubble concentration due to stratification and vein fillings. In addition, as a result of recrystallization, bubbles may be redistributed on the grain scale by grain boundary migration (Azuma et al., 2012), and this may lead to an alignment of bubble clusters parallel to the SPO and oblique to the dominant early foliation formed by a combination of bedding and rotated veins (e.g. 16c). It is possible also that bubbles may become reorganized by migration in response to stress, creating a foliation perpendicular to the maximum compressive stress (Pfeffer, 1992). In the case studied by Pfeffer at the terminus of the Variegated glacier, after the 1982-83 surge, bubble migration might also have been associated with reorganization of bubbles due to grain boundary migration during dynamic recrystallization associated with the final increment of compressional strain, i.e., incremental strain related rather than directly related to stress.

Bubble elongation develops as ice deforms, as has been demonstrated experimentally (Nakawo and Wakahama, 1981) and in natural ice (Hudleston, 1977a; Russell-Head and Budd, 1979; Russell-Head, 1982) and elongate bubbles may constitute a fabric element, especially in polar glaciers (e.g. Hambrey, 1977a; Hudleston, 1977a, and section 5.5). Bubble elongation, however, may not always reflect a coherent fabric component. Hambrey (1976b) reported bubbles in coarse-grained ice flattened in the plane of the foliation in a Norwegian valley glacier. Viewed in the plane perpendicular to foliation, however the bubbles were elongate in different directions in adjacent crystals, reflecting control by lattice orientation. Hambrey attributed this as an effect related to recrystallization under stagnant ice conditions (annealing).

5.4 Folds

Primary stratification, veins, and foliation all locally become folded, and folds have been described in all types of glaciers. Fold shape and style are quite variable (e.g. see Fig. 5 in Hambrey, 1977a), reflecting different mechanisms responsible for their formation. The most common folds visible at the surface have similar (Class 2, Ramsay's 1967 classification) or sub-similar shapes (Fig. 19) (Hambrey, 1977a; Calvet and Santanach, 1992), indicating little or no rheological contrast among the layers involved. These occur on many scales from kilometer (Fig. 19a) to centimeter (Fig. 19d). They are the consequence of non-uniform flow on different time scales and spatial scales, resulting in the passive transformation of planar markers such as

bedding, veins, or the banded form of foliation into folds. Alternatively, they may result from a more homogenous strain affecting stratification with originally non-planar form (see Hudleston, 1977b), for example inherited from the topography of the accumulation zone (Figs. 20, 21) (Fig. 10 in Hambrey and Glasser, 2003). In either case the folds are amplified kinematically by intense horizontal shortening or shear strain at various positions within the glacier. The upright folds with hinges plunging mostly gently at various angles up glacier are examples of this. They are commonly found in the ablation zone of valley glaciers, especially in the strongly foliated ice along glacier margins or internal longitudinal septa (resulting from the joining of two ice streams) and are typically associated with an axial planar foliation (Figs. 19c, 20-22, 23a) (Figs. 5,6 in Allen et al., 1960; Fig. 16 in Hambrey et al., 1999; Fig. 10 in Hambrey and Glasser, 2003). Significant horizontal shortening across the glacier suggested by these folds and associated foliation is to be expected considering the reduction in width of glaciers formed by the merging of ice streams of collectively much greater width (Fig. 20) (Lawson et al., 1994; Fig. 15 in Hambrey et al., 1999; Goodsell et al., 2005). Similar or sub-similar (Class 2) folds with steeply plunging hinges affecting the foliation (Fig. 19d, 23b) are younger than the folds just described and have a different origin, connected to the creation of the axial planar ice veins with which they are associated. They are a variety of flanking structure, discussed below.

Recumbent isoclinal or near isoclinal folds with similar (Class 2) geometry are often exposed at the termini of valley glaciers and ice caps (Fig. 19b). The planar fabric

element folded in this situation is a composite foliation produced by intense shear near then base of the ice, and these folds can be explained as follows (Hudleston, 1976, 1977b). Under steady state flow this planar fabric will tend to become parallel with the particle paths and folds will not form. If flow is non steady over sufficiently long time periods, such as is produced by secular variations in accumulation and ablation of ice along the glacier profile, the stress state and resulting strain rate may change sufficiently such that slight changes in the particle paths result in gentle undulations in foliation over bedrock irregularities becoming kinematically amplified following ice movement down glacier to form recumbent folds, as shown in Fig. 24. This way of producing folds requires no rheological contrast in the ice; the foliation can behave in a completely passive way. In the example shown in Fig. 19b that provided the impetus for the model fold of Fig. 24, the crystallographic fabric is the same in the two fold limbs and the hinge (Hudleston, 1976), consistent with homogenous flow on the scale of the fold. This is an example of non-steady laminar flow. At any instant in time and location near the glacier bed the deformation is very close to simple shearing. If there are small rheological contrasts produced by variations in bubble content, crystallographic fabric or dirt content, these may introduce secondary weak perturbations that result in non-perfect similar geometry, as in Fig. 19b. The total strain associated with these folds is large; the strain and strain rate variation around the folds small, and thus the crystallographic fabric is expected to be independent of position around them (Hudleston, 1976).

In passive folding where the perturbation is non-two dimensional, under large displacements and high strains the folds will tend to become sheath like. In such folds the hinges will rotate towards parallelism with the flow direction. The folds in the white Pleistocene ice at the base of the Barnes Ice cap are likely examples of this. In profile, parallel to the flow, they are most likely as illustrated in Fig. 24, similar to the style of folding seen in the superimposed ice in Fig. 9b. In plan view, they appear as in Fig. 25, an aerial photograph of the ice cap margin. The distinctive band of white ice is repeated by folding, in places up to three or four times. The process by which fold hinges rotate towards the shear direction is illustrated schematically in Fig. 26. In the case of the white ice folds in Fig. 25, the rotation likely has not been great, as the closed outcrops of white ice are drawn out parallel to the ice cap margin. Although no inner closed loops of layering can be discerned within the white loops of Fig. 25, these folds certainly belong to the “cat’s eye” category of sheath folds (Alsop et al., 2007) given the geometry of folds exposed elsewhere in marginal ice. This is expected where rheological contrast is small and where the kinematics is on the flattening side of plane strain, as is the case for the ice cap margin in Fig. 25.

A similar process of fold production to that operating near glacier beds, caused by annual or longer variations in ice flow in valley glaciers and their tributaries, produces folds with initially presumably steep plunges near glacier sides where tributaries join, clearly marked by medial moraines (e.g. Rutter, 1965) and especially pronounced in periodically surging glaciers (Post, 1972; Driscoll, 1980;

Lawson et al., 1994; Lawson, 1996; Hambrey and Lawson, 2000). This phenomenon is responsible for the spectacular folds seen on aerial photographs of many valley glaciers, and folds of this type have are found most dramatically in the piedmont Malaspina glacier (Fig. 19a) (Sharp, 1958). Folds in the Malaspina have been modeled effectively by Ramberg (1964).

Folding of passive type may also develop deep in ice caps under ice divides, even under steady-state flow. Such flow produces low-amplitude folds over highs and lows in a basal decollement (Whillans and Jezek, 1987) and a “Raymond bump” (Jacobson and Waddington, 2005), an anticline in layering centered at the divide reflecting vertical velocities being lower close to the divide than several ice thicknesses horizontally away from it (Raymond, 1983). The bump, however, does not develop if the flow is Newtonian, as might be the case if deviatoric stresses are sufficiently low (Pettit and Waddington, 2003) and is less pronounced if there is sliding at the base (Pettit et al., 2003). It is enhanced with the development of a vertically oriented *c*-axis fabric (Pettit et al., 2007), and in this case. the anisotropic rheology may cause the bump to develop a central syncline and flanking synclines (Martin et al., 2009). If a bump develops and there is migration of the divide under non-steady flow, a common situation, it can be transformed into a recumbent structure (Jacobson and Waddington, 2005), not unlike that illustrated in Fig. 24. Significant folding in ice deep within ice cores of Antarctica and Greenland has been inferred from borehole observations and from radio echo sounding images (Dahl-Jensen et al., 2013). These folds may originate from the effect described by Jacobson

and Waddington, by anomalies in accumulation rate (Vaughan et al., 1999), or from lateral flow over irregular topography, with enhancement due to the significant difference in viscosity between ‘cold’ glacial-period ice and ‘warm’ interglacial ice. Note that there is only a weak folding instability in a layered sequence of varying viscosity due to simple shear alone parallel to the layering, and a strong instability if there is combined compression and shear (Fletcher, 2005). Small-scale folding of kink style is observed in the highly anisotropic ice of the GISP2 ice core in Greenland (Alley et al., 1997b), consistent with Fletcher’s theoretical results, although it appears that the longitudinal strain rate in this case may be extensional rather than compressional. Thorsteinsson and Waddington (2002) presented an analysis to indicate that the tendency to produce passive recumbent folding of irregularities in layering under the combined influence of pure shear and simple shear near a divide is enhanced by the development of *c*-axis anisotropy. Recumbent passive folds produced by flow associated with migration of a dome have the potential to be of sheath type.

Low amplitude flow-parallel folds in stratigraphy noted on satellite images across fast flowing ice streams with high basal flow velocities in Antarctica are similar in some respects to sheath folds, likely being caused by flow over irregularities in the bedrock or anomalies in accumulation rate, often accentuated by constrictional flow (Gudmundsson et al., 1998; Siegert et al., 2004; Campbell et al., 2008). Unlike sheath folds, which rotate to become parallel to the flow direction, these folds initiate in the flow direction, affect much of the ice column and are effectively

transmitted by advection to the surface, where they are called flow stripes (Gudmundsson et al., 1998; Campbell et al., 2008). These large scale stripes are not the same as grain scale features described as stripes by Alley et al. (1997b).

Another type of passive folding is produced in association with the opening and closing of extensional fractures developed in foliated ice in the marginal zones of valley glaciers. In this case, the perturbation of the stress field around a crack from the time of opening until closure and sealing causes folding of the adjacent foliation (Figs. 27, 28) (Hudleston, 1989). Such folds are called *flanking structures* in current terminology (Passchier, 2001). A characteristic of the type developed here is that the sense of offset of a marker layer is opposite to the sense of shear in the glacier associated with the glacier flow or by the vergence of the folds (Fig. 28). Closely spaced crevasses may produce folds of more complex geometry. Fig. 23b is an example of this.

The examples of folds discussed so far do not necessitate any rheological differences between layers; they are all examples of (mostly) passive folding. By contrast, there are examples of folds in glaciers clearly produced by a buckling instability, in which perturbations in the layering become dynamically amplified when subjected to layer-parallel stress (Fletcher, 1974; Smith, 1975, 1977; Johnson and Fletcher, 1994). Characteristics of buckling are the development of folds with a characteristic wavelength that depends on the rheological contrast between the buckled layer and its matrix and the tendency of the buckling layer to maintain constant thickness

during folding (e.g. Hudleston and Treagus, 2010). High concentrations of debris or volcanic ash may produce a sufficient rheological contrast to lead to buckling. Excellent small-scale examples have been described by Ximenis et al. (2000) and Fitzsimons (2006), as illustrated in Fig. 29. Buckling may play a role in the folds developed in sediment layers in ice found near the terminuses of many glaciers (e.g. Glasser and Hambrey, 2002). Small-scale folds with buckling characteristics have been produced experimentally in ice in which an initial anisotropy due to grain size and layer-separating thin plastic sheets was employed (Wilson and Russell-Head, 1979; Wilson, 1981).

Increased concentration of debris in medial moraines may produce sufficient rheological contrast to enhance the kinematic amplification of perturbations in the folding of moraines during strong longitudinal shortening. Ramberg (1964) demonstrated experimentally this effect in modeling folding in piedmont glaciers. Any such amplification implies a departure from similar fold geometry; modest evidence of this is seen in the dark bands (medial moraines) in Fig. 19a and in the darker debris rich bands in the hinge of the recumbent fold in Fig. 19b.

5.5 Ductile Shear Zones

Zones of high shear strain are found along the margins and at the base of all glaciers and ice caps where the velocity gradient is highest. They are also found bounding ice streams in ice sheets - regions of faster flow than in the surrounding ice (e.g.

Marmo and Wilson, 1998). Zones of high finite strain – typically manifest as strong longitudinal foliation - are also associated with medial moraines in valley glaciers downstream of where two tributaries merge. In such cases the finite strain will be large, but the strain rate small once the two marginal zones merge to become more centrally located in the glacier. In the case of the Urumqi Glacier No. 1 in China, Echelmeyer and Wang (1987) documented enhanced deformation within the frozen subglacial drift and discrete shear zones within the frozen drift or at the drift-ice interface. Within basal ice, strain localization may occur at rheological interfaces and strain rate, the nature of recrystallization and crystallographic fabric may all vary from layer to layer (Samyn et al., 2008). Hambrey et al. (1999) described steeply up-glacier dipping shear zones near the terminus of the surging glacier Bakaninbreen in Svalbard. Shear localized in alternating limbs of folds likely enhances folding in sediment interlayered with basal ice, as in the folds described by Glasser and Hambrey (2002).

Other locally discrete zones of high strain also exist within the body of glaciers. These have been described deep in the Antarctica and Greenland ice caps, and may be characterized by finer grain size and a strong single-maximum vertical *c*-axis fabric (e.g. Gow and Williamson, 1976; Gow and Meese, 2007). The vertical shortening strains developed near the center of an ice cap lead to the basal planes rotating towards being perpendicular to the shortening direction and a vertical *c*-axis maximum. This effect is better developed in glacial-period ice than in interglacial ice. Glacial-period ice is thus in a more favorable orientation to deform

by horizontal shear, and shear zones and folds may develop under these conditions (Dahl-Jensen et al., 2013).

Examples of such zones on a small scale are provided in the margin of the Barnes Ice Cap where superimposed ice is overridden during an ice advance (Fig. 9b), Hooke, 1973). The shear zones are marked by a sigmoidal pattern of elongate air bubbles and bubbles (Fig. 30) and the development of a strong single-maximum preferred *c*-axis orientation (Fig. 31).

5.6 Linear Fabric Elements

Fold hinges are discrete linear fabric elements in glaciers, and their orientations depend on the position of the folds in the glacier and the strain history of the ice containing the fold. The recumbent folds formed near the base of a glacier or ice cap will generally have hinges that are initially perpendicular to the flow direction. If the topography in the bedrock surface seeding these folds is irregular, the fold hinges will locally be drawn towards parallelism with the flow direction and sheath folds will form, as described above (see Figs. 24-26).

Folds in foliation formed by variations in horizontal flow velocity in the plane parallel to the ice surface in marginal ice, or as flanking folds associated with fracturing, will initially have vertical hinges. These in general will become reoriented to plunge up-glacier at increasingly lower angles, due to the vertical

gradient in horizontal down glacier velocity. Sheath folds could potentially form here, too, with hinges horizontal and parallel to the flow direction.

Along the midline of valley glaciers, variations in attitude of primary stratification inherited from the accumulation zone may become passively amplified to form folds down glacier as a result of longitudinal shortening strain. The hinges of these folds will tend to be horizontal or gently plunging and perpendicular to the flow direction if the longitudinal strain rate is compressional, or parallel to the flow direction if there is convergent flow where ice streams merge (Figs. 19c, 22, 23a).

Locally, bubble elongation forms a penetrative linear fabric in glacial ice, with the direction of maximum elongation of the bubbles indicating a direction of maximum stretch in the enclosing ice (Hudleston, 1977a), modified by effects of surface energy, which tends to return bubbles to a spherical shape. The intensity of this effect depends on temperature and bubble size. The smallest bubbles maintain a spherical shape regardless of strain (Hudleston, 1977a). In coarse-grained temperate ice, bubble elongation may vary from crystal to crystal, either reflecting a difference in strain in adjacent crystals or lattice control of the bubble elongation (Hambrey, 1976b). At best, bubbles can be used as qualitative indications of strain and not as strain markers with shapes reflecting the strain ellipsoid (see Hudleston, 1977a).

5.7 Boudinage

777

778 Boudinage has been described in glaciers at different scales and with different
779 characteristics of the ice in which it is developed (Hambrey and Milnes, 1975, 1977;
780 Hambrey and Dowdeswell, 1994). The most distinctive examples are ones in which
781 the boudinaged layer is either of debris-rich ice (or frozen sand layers) embedded in
782 cleaner ice, or of clear ice embedded in debris rich ice (Fig. 32). This contrast in
783 behavior is interesting for two reasons. The first is that, depending on several
784 factors including debris concentration, temperature, and the presence or absence of
785 water at the ice particle interfaces, the debris rich ice can act either as the stiffer
786 layer or as the softer layer, behavior consistent with other observational data and
787 experimental data (see section 2.1; Moore, 2014). The second is because the pinch
788 and swell phenomenon that may precede boudinage is a weak dynamic instability
789 (Smith, 1977) with the dynamic growth rate exceeding the kinematic dampening
790 effect associated with the extension only for strongly non-linear viscosity (Smith,
791 1977; Schmalholz et al., 2008). Rheological data suggest that variations in debris
792 content may lead to differences in viscosity of up to about a factor of around 10 (Fig.
793 3). The pinch and swell instability is very weak with $n \leq 5$ for such a viscosity
794 difference. Strain softening significantly enhances the growth rate of the instability
795 (Neurath and Smith, 1982) and this may well be a factor in the development of
796 boudinage in ice, with the softening in this case being a result of fabric development.
797 A combination of theoretical and numerical modeling considerations led Schmalholz
798 et al. (2008) to suggest that values of $n \geq 5$ and viscosity ratio of ≥ 20 are sufficient for
799 the development of pinch and swell without the need for other enhancing

mechanisms such as strain softening. Such values are at the high end of what might be expected for glacial ice containing debris versus debris free ice.

Some of the structures described by Hambrey and Milnes (1975, 1977) as foliation boudinage in strongly foliated ice appear to be the same as flanking folds associated with veins described by Hudleston (1989) (Fig. 27), and likely have the same origin, especially since the sense of offset associated with the vein or shear zone that separates the two parts of the structure is opposite to that suggested by the symmetry of the folds. Such flanking folds are in fact one end member of types of foliation boudinage described by Arslan et al. (2008). In contrast, the symmetric structures also described by Hambrey and Milnes as foliation boudinage (and similar structures described as confluent structures by Taylor, 1963) could not have originated as flanking folds and must represent foliation-parallel extension, with localized necking or fracture to produce the boudinage.

The most unambiguous examples of boudinage occur in basal ice with layers of strongly contrasting debris content (Fig. 32) (Hambrey and Fitzsimons, 2010, Fig. 13B). Less unambiguous examples have been invoked to explain or predict structures in ice in which variation in debris content is not a factor. Cunningham and Waddington (1990) applied the analysis of Smith (1977) and Neurath and Smith (1982) to predict that pinch and swell instability would lead to localized enhancement of variations in the thickness of the Sangamon interglacial ice below the Summit divide of the Greenland ice cap. This is by assuming the interglacial ice

to have an effective viscosity about three times that of the glacial-age ice above and below. Theory indicates the kinematic dampening effect on initial thickness variations is just about balanced by the dynamic growth of these variations due to the pinch and swell instability with this viscosity contrast, leading to potentially significant enhancing of the *relative* total thickness variations in the interglacial ice, since the overall thinning exceeded 90%. In another situation, Marmo and Wilson (2000) described segmented veins defined by coarse-grained columnar ice infilling tensile fractures in an outlet glacier of the Framnes Mountains, east Antarctica (Fig. 32d). They attributed the segmentation to boudinage of an originally continuous ice vein under the influence of stresses during subsequent ice flow and used a finite difference numerical model to explore the conditions under which this might occur. The initial crystallographic fabric of the ice forming the veins is one of *c*-axes parallel to the vein walls, which put this ice in an orientation unfavorable for basal glide during extension of the vein and thus made the vein ice stiffer than the surrounding ice, which had a more isotropic fabric, by a factor of up to about five. The dynamic growth rate of perturbations in the vein margins is very small under these conditions such that a strong softening mechanism would be needed to produce the observed structures through a pinch and swell instability. An alternative origin for these structures seems possible; they may have originated as en echelon fractures rather than as a single continuous fracture, with the segmentation being primary rather than the result of boudinage.

5.8 Faults

846

847 Many examples of thrusts or shear planes in glaciers have been described in the
848 literature (e.g. Boulton, 1970; Glasser et al., 1998; Hambrey et al., 1999; Herbst and
849 Neubauer, 2000; Herbst et al., 2006), and the general concept and a specific example
850 are shown in Fig. 33. However, the evidence for discrete surfaces across which
851 there is a displacement discontinuity produced by faulting is indirect, except in the
852 case of surging glaciers (Sharp et al., 1988). In some studies, structures with clear
853 sense of offset of markers have been interpreted as crevasse traces reactivated as
854 thrusts (e.g. Goodsell et al., 2005; Appleby et al., 2010).

855

856 There is a serious problem in interpreting the structures described as faults or
857 thrusts as the result of shear fracturing, in that the stresses required to produce
858 shear fractures in ice are much larger than the stresses usually inferred throughout
859 glaciers (Moore et al, 2013). Estimates of shear stresses near the base of glaciers
860 are typically in the order of 50-200 kPa (Cuffey and Paterson, 2010, table 8.3). In
861 compressional tests, shear failure of ice other than by plastic flow is of Mohr-
862 Coulomb type, with failure stress dependent on pressure. In uniaxial compressive
863 tests on polycrystalline fresh water ice of about a millimeter grain size, the
864 transition in failure mode from ductile to brittle occurs at an applied stress of about
865 9 MPa (Schulson, 1999), about an order of magnitude greater than the uniaxial
866 tensile strength of about 1 MPa. In compression tests at strain rates of less than
867 about 10^{-3}s^{-1} , ice fails by ductile flow before shear fracture occurs under confining
868 pressures of greater than about 1 MPa (Rist and Murrell, 1994; Schulson, 1999).

869 The fastest strain rate measured in the Variegated glacier under surge conditions
870 was about $2 \times 10^{-6} \text{s}^{-1}$ (0.2 per day) (Kamb et al., 1985). Thus even in the case of a
871 surging glacier where the highest stress are encountered, conditions generally
872 appear not to favor thrusting (Moore et al., 2010).

873
874 Evidence cited for thrusting and more generally of faulting in glaciers includes the
875 offset of markers across planar structural features (Fig. 34) (Hambrey, 1976;
876 Hambrey et al., 1996; Goodsell et al., 2005). Many of these features likely arise from
877 the same process that leads to the flanking structures described earlier (Figs. 27, 28).
878 That is, the 'fault' originates as a *tensile* fracture that is later closed to become a vein,
879 and that between its opening and closing there is rotation of the fracture due to the
880 velocity gradient in the embedding ice. The walls of the fracture can sustain no shear
881 stress until the fracture is closed and sealed. The strain in the surrounding ice that
882 must accommodate the rotation in a varying stress field results in relative
883 displacement of markers cut by the fracture and leads to the appearance of faulting,
884 as in Fig. 34. Such structures might still be called faults, but they are of a peculiar
885 kind and are not indicative of shear failure. In regions of heavy crevassing, the
886 process just described can lead to a complicated pattern of veining and folding (Fig.
887 34b) (see also Ragan, 1969, Fig. 7).

888
889 The structures most commonly interpreted as thrusts or shear planes in glaciers are
890 located near the termini where the ice is subjected to longitudinal compression.
891 What is indisputable in such cases is that coarse and fine-grained debris from the

base of ice caps and some valley glaciers finds its way to the surface in ablation zones upstream from the glacier front (Hambrey et al., 1999). There must therefore be some mechanism or mechanisms by which basal sediment becomes entrained in glaciers and then moved upwards, and possibilities for this have been much discussed in the literature. Historically, one of the most popular such processes involves shearing or thrusting originating from the bed (Goldthwait, 1951; Swinzow, 1962; Tison et al., 1993; Alley et al., 1997a). Not all structures so described involve discrete planes of movement. The structure envisioned by Tison et al. is not actually a thrust fault, but rather a ductile shear zone containing debris, localized between active glacial ice and “dead” ice near the ice-sheet margin. Thus shear failure is not required in this case. Enhanced deformation in the debris-rich basal ice, as described by Echelmeyer and Wang (1987), would facilitate development of such a shear zone.

Several mechanisms to account for sediment entrainment that do not involve thrusting, and that may eventually lead to emergence of the sediment at the surface along what appears to be a thrust, have been described. These include regelation associated with flow around obstacles in wet-based glaciers (Weertman, 1964) and other mechanisms associated with freezing of water at the base of wet-based or polar glaciers (e.g. Hooke, 1973, 2005; Hubbard and Sharp, 1989; Shoemaker, 1990; Iverson and Semmons, 1995). In polar glaciers, such mechanisms may be facilitated by the glacier being frozen to a deformable bed of sediment at the ice front, and unfrozen some distance upstream, with the boundary between frozen and unfrozen

915 shifting with time. The upward vertical component of velocity in the zone of
916 longitudinal compressive strain at the terminuses of glaciers may bring ice
917 containing debris to the surface once entrained. The overriding of clean
918 superimposed ice by debris-bearing ice during an ice advance provides a good
919 explanation for the ice-cored moraines at the margin of the Barnes Ice Cap (Hooke,
920 1973, 2005) and an alternative to a thrust origin for the material that outcrops at
921 the surface.

922
923 Folding originating at the junction between frozen and unfrozen bed has been
924 suggested as a possible alternative to thrusting as a mechanism for bringing debris
925 to the surface (Clarke and Blake, 1991; Moore et al., 2013). In general the complex
926 interplay of glacier sliding with melting and regelation over an irregular topography
927 produces highly variable stresses and complex structural features including folds,
928 boudinage and localized zones of shear, in particular in cold-based glaciers (e.g.
929 Hubbard and Sharp, 1989; Sharp et al., 1994; Fitzsimons et al., 1999; Hambrey and
930 Fitzsimons, 2010.) Elevated pore pressure is likely to be important in these
931 processes and may be sufficient to induce extensional fracturing.

932
933 For a specific example of evaluating the hypothesis of thrusting, consider the case of
934 Storglaciären, where the sediment in an up-glacier dipping debris band near the ice
935 front of has a basal source (Jansson et al., 2000). Thrusting and folding are two
936 possible explanations for this feature (Glasser et al., 2003). By use of numerical
937 modeling and application of fracture mechanics, informed by empirical data on the

rheology and fracture of ice, Moore et al. (2010) showed that the strain rates are six orders of magnitude too small to induce shear fracturing and thus to produce a thrust that could bring debris to the surface along a thrust plane. They did, however, show that frictional slip on a preexisting fracture is possible, if the fracture is large enough, properly oriented, and there is high fluid pressure along the fault surface. They further considered these conditions most likely to be met on surfaces containing sediment. That is, if a sediment-bearing thrust already exists, movement on it by frictional slip is possible under the right conditions. Emplacement of the debris into the ice and elevation above the base, however, must be by some other mechanism. Moore et al. (2013) suggest folding as the most likely process for this. If fractures are involved, they are most likely to have originated as tensile cracks and not by shear. In the case of Variegated glacier, the observed thrusts during the 1982-1983 surge were likely to have been the result of folding and tensile fracturing as the surge passed (Raymond et al., 1987; Sharp et al., 1988; Moore et al., 2010).

6. Surging Glaciers

Surging glaciers are ones that periodically go into a phase of movement in which velocities increase by up to two orders of magnitude over their non-surging state, triggered by a build up of water pressure at the base of the ice (Kamb et al., 1985; Raymond et al., 1987; Eisen et al., 2005). They result in a net transfer of a large volume of ice from the accumulation zone to the ablation zone over a period of a year or two. They may be preceded by mini surges last a day or two (Kamb and

Engelhardt, 1987). The fluctuations in flow from different tributary ice streams associated with surges typically leads to folding of the medial moraines, with the folds becoming tight to isoclinal downstream along the valley sides and in the region of high longitudinal compressive strains near the glacier front (Post, 1969, 1972; Clarke, 1987). The geometry of the folds suggests this folding is largely passive.

The surge of the Variegated glacier in Alaska was closely monitored from start to finish in 1982-83 (Kamb et al., 1985). The surge started in the upper reaches of the glacier and moved downstream in a kinematic wave. Ice thickened following the passage of the surge front, with extremely high compressional strain rates as the front passed, as large as 0.2 per day ($2 \times 10^{-6} \text{ s}^{-1}$). The velocity gradient was small after passage of the front and longitudinal stresses locally changed from compressional to extensional as the wave passed. This behavior led to a distinctive suite of structures (Fig. 35) (Lawson et al, 1994; Lawson, 1997; Sharp et al., 1988; Lawson et al., 2000), including both lateral and frontal faults. During the surge, the surface velocity profile changed from the classical parabolic shape to almost plug flow, with two strike-slip fault zones containing brecciated ice on either side (Kamb et al., 1985). Folding and thrusting developed at the front of the glacier as the surge encountered stagnant ice. Folding of the ice in the case of the Variegated glacier was observed at the surface, involving both ductile and brittle behavior, the latter producing what Sharp et al. (1988) called exfoliation buckle folds. Sharp et al. (1988) traced some thrust faults to roots at the glacier bed – which therefore acted as a decollement – and to lateral terminations in bulge folds, suggesting a ductile

precursor to the passage of the shear fracture and development of the structures in the style of fault-propagation folding (Suppe, 1985). Although clearly active as thrust faults, Lawson et al. (1994) suggested that because of the considerable variation in dip, they may have originated as transverse tensile fractures and later become activated as thrusts. Such an origin would be consistent with the calculations of Moore et al. (2013) who found the compressional stresses too small to initiate shear fractures during the surge. In their propagation as thrusts, the Variegated structures developed in classic piggyback style and slickensides developed on the fault surfaces.

Thrusting has been inferred to occur at the fronts of other surging glaciers (e.g. Hambrey et al., 1996; Hambrey and Dowdeswell, 1997; Porter et al., 1997; Hambrey et al., 1999), although the only case in which thrusting appears to have been directly observed is that of the Variegated. Hambrey and Dowdeswell (1997) found foliation in the surging Hessbreen glacier in Svalbard to be overall less intense than that in non-surging glaciers, indicating lower strain over most of the ice and slip or concentration of strain in narrow zones at the glacier sides and base. The identification of debris-filled planar structures as thrusts by previous workers in surging glaciers in Svalbard has been questioned by Woodward et al. (2002), who considered most of these features to have originated as sediment-filled crevasses that become later deformed and reoriented into attitudes suggesting thrusting.

Crevassing was intense during the surge of the Variegated glacier and followed an interesting pattern (Fig. 35). At the surge front, longitudinal compression and divergent flow produced longitudinal crevasses, and subsequent longitudinal extension produced transverse crevasses, leading to a blocky crevassed surface (Lawson et al., 1994). In general, fracturing was much more extensive in the surge phase than in the quiescent phase. At the end of the surge, a threefold pattern of crevassed regions was noted; an upper region with transverse crevasses, a long mid region with superimposed longitudinal and transvers crevasses, and a lower region with longitudinal crevasses (Fig. 35, Lawson, 1996). This pattern also apparently followed earlier surges also (Lawson, 1996).

7. Progressive deformation

A particle of glacial ice may experience a complex history of stresses during its passage from formation in the accumulation zone – or accretion at the base, front or elsewhere on a glacier - to its eventual removal by ablation in the frontal zone or melting elsewhere, and a glacier as a whole might be considered its own tectonic entity. As such, glaciers provide possible analogs for some tectonic settings in rocks. For instance, Coward (1982) interpreted parts of the Moine thrust zone as representing local surges, analogous to glacial surges, and Herbst and Neubauer (2000) compared structural development in the Pasterze glacier in Austria with structural development in an extensional allochthon.

Two different modes of ice ‘tectonism’ might be distinguished, one in which the glacier is more or less in steady state, and the other in which there is great variation with time, as in surging glaciers. In the first, ice particles follow their given paths through the glacier and experience their particular stress histories depending on the initial starting point of each particle. Measurements of ice velocity and structures and fabrics at any fixed geographic point on the glacier would be similar from year to year. Schematic examples of this are shown in Figs. 5 and 6. In the second mode, as in surging glaciers, the deformational history experienced by the ice varies greatly both in space and time. In this case measurements taken at a fixed geographic point will show great variation depending on when they are taken and overall structural development will change with time (Fig. 35, Lawson, 1997).

There are similarities in the structure developed near the fronts of surging glaciers with structures in fold-and-thrust belts, as noted by Kamb et al. (1959) and Sharp et al. (1988). The role of fluids is critical in both cases, and the sequential development of linked folds and thrusts is also similar in the toe regions of strong compressional strain.

In both steady and non steady state glaciers, structures and fabrics can be mapped following the classical techniques of structural geology, identifying bedding (S_0) and then sequentially developed planar fabrics, designated, S_1 , S_2 , etc. and generations of folds, F_1 , F_2 etc. (Hambrey and Milnes, 1977; Lawson et al., 1994; Hambrey and Dowdeswell, 1997; Roberson, 2008; Jennings et al., 2014). The sequence generally follows the progression of ice down glacier or down flow lines, where particles of ice

will be subjected to different stress states and structural regimes. As in rocks, as nicely demonstrated for example in the Pennsylvanian Appalachian Valley and Ridge province by Gray and Mitra (1993), the designation D_1 and D_2 (or S_1 and S_2) does not mean that D_1 is everywhere earlier than D_2 , only that in one location this is the case. D_2 structures may develop at one location a glacier at the same time as D_1 structures are forming upstream.

In all glaciers and ice caps, moving from source to terminus involves moving from an accumulation zone of generally longitudinal tensile stresses to an ablation zone of generally longitudinal compressive stresses, with local fluctuations depending on valley geometry. The structures that develop reflect this overall change in longitudinal stress regime, as well as the shear stresses developed along the base and sides. As climate changes and many glaciers recede, the dynamics of ice flow may change and with it the nature of structural development (Hambrey et al., 2005).

In valley glaciers, extension in the accumulation zone leads to extensive transverse crevassing, providing the likely principal source for transverse foliation downstream following closure of the crevasses. Shear along the sides of valley glaciers gives rise to en echelon crevasses and subsequent strong near vertical longitudinal foliation as the crevasse rotate and become veins. Depending on where in the glacier the foliation develops, transverse foliation may crosscut longitudinal foliation or vice versa. Hambrey and Milnes (1977) mapped an early longitudinal S_1 ,

cut by a transverse S_2 , with a later longitudinal S_3 developed near the glacier terminus. Folds developed with axial surfaces parallel to both S_1 and S_2 .

Folds develop in bedding and in foliation and frequently are associated with an axial planar foliation (Figs. 19c, 21, 22 and 23b) (e.g. Hambrey et al., 1996; Hambrey et al., 1999; Goodsell et al., 2005). The orientation of folds within even a simple glacier can be highly variable. This is illustrated with folds in either bedding or foliation in Storglaciären (Fig. 36). Depending on location, fold hinges can be considered to have three 'end member' orientations: sub horizontal perpendicular to the flow direction (with vertical axial surface); sub vertical (with axial surface parallel to the flow direction); sub horizontal parallel to flow (with variable axial surface dips). Two end members are shown in Fig. 23. Intermediate orientations are also found. This is all in a small region of overall tectonic simplicity with a single eastward flow direction.

8. Conclusions

Deformation in glaciers and ice caps often leads to a complex suite of structures, even when the flow field is apparently simple. This is due to the combination of brittle fracture and plastic flow affecting ice with inherited and induced heterogeneities, largely visible through variations in bubble content. Primary stratification, S_0 , is preserved to large strains and becomes folded, largely passively, for various reasons, including irregularities in the accumulation zone, changes in flow responding to climate, localized shear, and deformation accommodating the

opening and closing of tensile fractures, often involving fracture rotation. In the most strongly deformed ice along the margins of valley glaciers and near the base of all glaciers and ice caps S_0 contributes to a composite foliation in which other fabric elements, in particular fracture fillings, play an essential role. This foliation can itself become folded. Dynamic recrystallization on the grain scale also contributes to foliation, leading to a reorganization of bubbles and grain boundaries. The different fabric elements contributing to foliation commonly result in crosscutting relationships, with a younger foliation (S_2) cross-cutting an older one (S_1), with a possibility of several generations progressing down glacier. Rheological contrasts, with viscosity variations of up to about an order of magnitude, due to grain size and bubble content, crystallographic-fabric induced anisotropy or debris content, may give rise to dynamic instabilities of buckle folding and boudinage, usually on a small scale. Faulting by shear failure in ice is theoretically unlikely and has been directly observed only in surging glaciers. The commonly observed fault-like discontinuities in glaciers are mostly the result of displacement occurring between the opening and closure of extensional fractures. Structures and fabrics at any location in a glacier or ice cap must be interpreted in terms of the flow history the ice has experienced.

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1679 **Figure captions**

1680

1681 Fig. 1 Schematic diagrams to illustrate the flow patterns in ice caps and valley

1682 glaciers in map view (a, c) and in cross section (b, d). Net accumulation exceeds

1683 ablation in the accumulation zone (and longitudinal strain rate, $\dot{\epsilon} > 0$), and net

1684 ablation exceeds accumulation in the ablation zone ($\dot{\epsilon} < 0$). If basal temperature is at

1685 the pressure-melting point there will be a component of basal slip (not shown). Ice

1686 sheets (not shown) are much larger than ice caps and possess ice divides separating

1687 flow basins; they possess outlet glaciers and ice streams at their margins in which

1688 the velocity is much higher than in the surrounding ice (see Rignot et al., 2011).

1689 Otherwise, the basic flow pattern in an ice sheet resembles that of an ice cap.

1690

1691 Fig. 2. a) Curves of log strain rate versus log strain for creep tests on polycrystalline
1692 ice under unconfined compression, illustrating the difference between starting with
1693 an isotropic *c*-axis fabric and an initially compatible anisotropic fabric – a small
1694 circle of *c*-axes centered on the compression axis. In the initially isotropic case, the
1695 strain rate passes through a minimum (secondary creep) and tends towards a stable
1696 value (tertiary creep) as a small circle *c*-axis fabric develops. The fabric after about
1697 10% strain is similar in both cases and similar to the starting anisotropic fabric. b)
1698 Same data on a traditional creep plot of strain against time. After Treverrow et al.
1699 (2012) and experiments of Gao and Jacka (1987).

1700

1701 Fig. 3. Numerical model results of stress-strain rate relationships in a
1702 polycrystalline aggregate to show softening or hardening of ice due to lattice
1703 preferred orientation of *c*-axes (after Cuffey and Peterson, 2010, fig. 3-20). The plot
1704 shows enhancement of shear strain rate in simple shear and of compressive strain
1705 rate in uniaxial compression (vertical axis) against strength of fabric for simple
1706 shear and cone angle of girdle distribution for uniaxial compression (horizontal
1707 axis). The characteristic angle delimits the distribution of *c*-axes as shown. The
1708 width of the band in compression is 30°.

1709

1710 Fig. 4. Measured and calculated (based on measurements) velocities in a cross
1711 section of the Athabasca Glacier, from Raymond (1971a, 1971b). a) Contours of
1712 longitudinal velocity, in m a^{-1} . Positions of boreholes are indicated. b) Difference

between surface longitudinal velocity and longitudinal velocity at depth at position 1A. c) Longitudinal velocity variation across the glacier, same location. Positions of markers or boreholes shown by black dots. Velocity at open circle estimated from adjacent stakes and tape measure.

Fig. 5. Schematic particle paths (light solid blue lines), isochrons (red dashed lines) and strain pattern in an ice cap at steady state (from Hooke and Hudleston, 1978). E.L. is equilibrium line; D is ice divide. If there is extension in the third dimension, as in radial flow, the strains will all be of flattening type; otherwise this is plane strain.

Fig. 6. Flow and strain regimes in a cross section of the Law Dome, Antarctica, from the summit to Cape Folger (after Budd and Jacka, 1989). Bedrock topography from radio echo sounding. a) Particle paths and isochrons in thousands of years; locations of boreholes shown by grey lines. b) Contours of cumulative compressive strain. c) Contours of cumulative shear strain.

Fig. 7. Estimated surface finite strain field accumulated in flow starting below an icefall in Griesgletscher, Switzerland based on displacement of stakes over several years and assuming steady state (from Hambrey and Milnes, 1977). Five separate particle paths are followed, with the time interval in years associated with each strain state noted. The estimates are realistic in as much as flow does not vary much

with time and ice velocities do not change significantly with depth. See Hambrey and Milnes for discussion.

Fig. 8. Schematic crystallographic *c*-axis patterns in found in ice caps and in experiments (after Budd, 1972; Hudleston, 1977a; Duval, 1981; Bouchez and Duval, 1982; Budd and Jacka, 1989). Axes are with reference to ice caps or experimental tests: *x*, flow direction, *z*, vertical, *y*, perpendicular to the 'flow plane.' a) Small circle girdle about the vertical at domes and in uniaxial compression. The opening angle of the cone gets smaller as strain increases. b) Bimodal pattern at a divide, and in pure shear experiments. c) One strong and one weak maximum in simple shear at low shear strain. d) A strong single maximum in simple shear at higher strain. e) Four-maxima 'diamond' pattern found in coarse-grained ice in valley glaciers and deep in ice caps.

Fig. 9. a) Sedimentary layering is the only fabric element evident (apart from a few fracture traces at the left) at the terminus of Southeast Kaskasatjåkkaglaciären, a small cirque glacier in northern Sweden. The field of view is about 100m. b) Sedimentary layering in superimposed (post Pleistocene) ice at the edge of the Barnes Ice Cap, Canada, with older layers folded, overturned and unconformably overlain by younger ice. The cliff is about 15 m high.

Fig. 10. Classic pattern of crevasses in a valley glacier according to Nye (1952). Marginal crevasses form at 45° to the valley sides in the absence of longitudinal

stress ($\sigma_x = 0$), and at angles either less than or greater than 45° in the presence of a longitudinal stress, depending on whether this is compressive or tensile. Transverse crevasses form under longitudinal tensile stresses and curve as the longitudinal stress is modified by shear stress approaching the margins. A second set of crevasses may form (dashed lines) if both principal stresses become tensile. Arrows represent directions, sense and magnitude of principal stresses. a) compressive flow, b) no longitudinal stress, c) extensional flow

Fig. 11. Natural crevasse patterns in three glaciers in Swedish Lapland. a) En echelon marginal crevasses, as in Fig. 10b, affecting ice of Rabotsglaciären with an arcuate trace of primary stratification, the center of the arc corresponding to the position of maximum horizontal velocity. (Distance from the near margin to the center of the glacier ca. 200m). b) En echelon crevasses with the pattern shown in Fig. 10a, just below the equilibrium line in Storglaciären, with primary stratification as in a). In this case the active crevasses cross cut older rotated crevasse traces. . (Distance from the near margin to the center of the glacier ca. 400m). c) Isfallsglaciären, with transverse crevasses, for which the pattern becomes more complex with flow over rough bedrock topography. There are marginal en echelon crevasses in the superimposed snow/ice in the foreground. (Width of the glacier ca. 800 m).

Fig. 12. Pattern of crevasses (red lines) in a stretch of the Blue Glacier, Washington, and relationship to surface principal strain rates (from Meier et al., 1974).

1781

1782 Fig. 13. Pattern of closed crevasses, veins or fracture traces, along the south margin
1783 of Storglaciären, Sweden, with cross-cutting relationships consistent with formation
1784 and subsequent rotation before older crevasses are crosscut by younger. The
1785 arcuate trace of sedimentary layering, combined with closed transverse crevasse
1786 traces, is prominent in the center of the glacier, which is about 1 km wide. There is a
1787 secondary partial arc of sedimentary layering in the near part of the glacier,
1788 originating from a subsidiary accumulation basin. Flow is to the right.

1789

1790 Fig. 14. Sequence of structures developed in the marginal ice of Sörsdal Glacier, East
1791 Antarctica as a result of crevassing, from Patrick et al. (2003). a) Open crevasse
1792 cutting ice with numerous veins/fracture traces (shaded blue) formed by closure of
1793 older crevasses and showing cross-cutting relations consistent with rotation after
1794 formation in a zone of dextral shear, as indicated by the inset, with veins in S_1
1795 orientation being the oldest, and S_5 being the youngest. b) Sequence of deformation
1796 of vein after formation from a crevasse, involving closure, rotation and differential
1797 shear to form folds.

1798

1799 Fig. 15. a) Histogram of orientations, defined by angle, α , of ice veins/fracture traces
1800 formed from closed crevasses along the north margin of Storglaciären, Sweden, in
1801 the ablation zone just below the equilibrium line b) Simulated distribution of
1802 orientations of fracture traces forming at regular intervals (values of shear strain)
1803 and then being rotated passively by subsequent flow, modeled as simple shear;

initial orientation of fractures is 135° . The average shear strain, γ , required to produce the rotations for each 20° range of orientations is shown above the histogram. c) The same as b) except that the initial orientation of the fractures is 90° . The initial fractures sub-vertical; this is also true of the natural fracture traces and veins.

Fig. 16. a) Strong steeply dipping longitudinal foliation between two ice streams from separate accumulation zones on Rabotsglaciären, Sweden. b) Strong composite longitudinal foliation along the south side of Storglaciären, Sweden, associated with a large shear strain. The relatively bubble-free ice is interpreted to have originated as veins filling fractures, now parallel to primary stratification and bubble clustering. c) Strong longitudinal foliation (marked by the pencil) near the front of Storglaciären cross cut by clear ice veins forming a second planar fabric element, with significant deformation and rotation post vein emplacement. d) Three distinctly different planar features near the center of Storglaciären: black: primary stratification (S_0), blue: two veins, one composite, infilling fractures, and red: a near vertical longitudinal foliation defined by slight variations in grain size and bubble clustering on the grain scale. The foliation affects both S_0 and, to a lesser extent, the veins.

Fig. 17. a) Vertical slab of ice about 4 mm thick in a transverse orientation from the north side of Storglaciären to show steeply dipping longitudinal foliation defined by bubble clustering, and a vein of clear ice, with irregular edges due to deformation. b)

The same slab, thinned to 2 mm, under crossed polarizers to show large recrystallized grains with a shape-preferred orientation generally parallel to the foliation.

Fig. 18. Schematic diagram to illustrate how a) various components of foliation in ice become modified by b) pure shear and c) simple shear. d) Strain alone for shear strain of $\gamma = 10$. (after Hooke and Hudleston, 1978). With very large strains, especially in simple shear, often exceeding $\gamma = 10$, a composite planar fabric will form from these components.

Fig. 19. Folds at very different scales seen on the surface (a,c,d) and in an ice cliff (b) of different glaciers. a) Similar style folds in medial moraines on the east edge of the Seward Glacier lobe of Malaspina Glacier (from Sharp, 1958). The field of view is several km. b) Recumbent fold in debris-containing superimposed ice exposed in an ice cliff at the edge of the Barnes Ice Cap, Canada (see Hudleston, 1976). Dirt coats the surface of the cliff and the debris content of the ice is much smaller than appears from the picture. Ice axe for scale in center of picture. c) Sharp-hinged folds with axial planar foliation in primary stratification midre Lovénbreen, Svalbard (Fig. 5B in Hambrey and Glasser, 2003). Ice axe (circled) lies along the fold axial surface trace. d) Fold of subsimilar style affecting composite foliation at the edge of Storglaciären, Sweden. Axial surface is parallel to the glacial margin and the hinge plunges steeply. This fold does not have an axial planar foliation. Penknife (circled) for scale is 9 cm long.

1850

1851 Fig. 20. Convergent flow from an accumulation basin produces folds in irregularities
1852 in primary stratification and a longitudinal foliation (as in Fig. 19c), most intense
1853 along the valley sides due to combination of shear with lateral shortening. Based on
1854 Fig. 7 of Lawson et al. (1994) for structures developed in Variegated Glacier, Alaska.
1855 Note that the deformation history as the glacier narrows is basically one of
1856 transpression, with opposite sense of shear on either side.

1857

1858 Fig. 21. Schematic diagram to show progressive development of folds in the
1859 situation illustrated in Fig. 20, as seen in a transverse section, with horizontal
1860 shortening across the glacier eventually producing transposition of primary
1861 stratification and development of a strong longitudinal foliation in marginal ice
1862 (modified from Hambrey and Lawson, 2000).

1863

1864 Fig. 22. Schematic illustration of structures and fabrics developed in valley glaciers
1865 due to lateral shortening associated with flow from two or more accumulation
1866 basins. Folds in stratification have sub horizontal hinges parallel to the flow
1867 direction and an axial planar foliation. The figure is after Hambrey et al. (1999),
1868 based on observations on several glaciers in Svalbard.

1869

1870 Fig. 23. Examples of folds with different origins and orientations in Storglaciären,
1871 Sweden. a) Folds of the type illustrated in Fig. 21 and 22 exposed in the walls of
1872 closely spaced closed crevasses in marginal ice of Storglaciären, Sweden, with sub

horizontal hinges and vertical longitudinal axial planar foliation. b) Steeply plunging fold of foliation in marginal ice, with vertical ice veins parallel to the axial plane and at a small angle to the glacier margin.

Fig. 24. Numerical model of passive recumbent folding in stratification at the base of the Barnes Ice Cap. The fold is generated by a small perturbation of flow over bedrock topography (to the right of what is shown here) and amplified kinematically. The direction and magnitude of the maximum extensional strain accumulated since the perturbation occurred are indicated by the short lines with numbers beside them (after Hudleston, 1977).

Fig. 25. Aerial photograph of the SE margin of the Barnes Ice Cap to show band of white Pleistocene ice repeated by folding of the kind illustrated in Fig. 24. See Hooke (1973) for explanation of how the Pleistocene age is both overlain and underlain by younger ice. Original photograph supplied by Department of Energy, Mines and Resources, Ottawa, Canada. Field of view is 5 km wide.

Fig. 26. Schematic diagram to show how recumbent folds of the kind shown in Fig. 24 can become sheath folds, with conical closures drawn out in the flow direction (after Hudleston, 1986). The layer designated by heavy black line outcrops on the glacier surface (shaded plane) as shown by the dashed lines.

Fig. 27. Flanking fold in marginal ice of Storglaciären associated with the creation, closure and rotation of a crevasse, now a vein, shown in blue (after Hudleston, 1989).

Fig. 28. Simple model, based on experiments on clay, to show development of a flanking fold (after Hudleston, 1989), with shear strain values associated with rotation of the fracture as it fills in – indicated with blue shading - and becomes a vein, with initial orientation at 45° to the shear plane (glacier margin). An initially continuous planar marker cut by the fracture is shown to be folded and displaced to form a flanking structure – double hook shaped fold - while the rotation occurs. The strain ellipse at $\gamma = 10$ is truncated.

Fig. 29. Examples of small-scale buckle folds in glacial ice. a) In a thin tephra layer near the terminus of Johnsons Glacier, Livingston Island, Antarctica, with hinges parallel to the flow direction (from Ximenis et al., 2000). The folds indicate shortening perpendicular to this direction. The difference in shape of the inner arcs of the anticlines and synclines in the tephra layer indicates that the ice above and below the layer have different viscosities, with that above being higher than that below. b) In a layer of gravelly sand about 1m above the base of Suess Glacier, Antarctica (from Fitzsimons, 2006).

1916 Fig. 30. Small-scale shear zone in superimposed ice at the margin of the Barnes, Ice
1917 Cap, Baffin Island, Canada, defined by stretched bubbles. Seen in a vertical slab of
1918 ice 0.3 cm thick. See Hudleston (1980) for location and setting.

1919

1920 Fig. 31. Small-scale shear zone from the same location as Fig. 30, in crossed
1921 polarizers to show the grain shape fabric and extremely strong *c*-axis preferred
1922 orientation in coarsely recrystallized ice within the shear zone, with the maximum
1923 perpendicular to the shear zone boundaries. In both this and Fig. 30 the shear zones
1924 are sub horizontal and shear direction is perpendicular to the ice cap margin. a)
1925 Oriented with shear zone grains showing brightest. b) Oriented with nearly all shear
1926 zone grains in extinction. Section about 1mm thick. Centimeter grid in a) gives
1927 scale.

1928

1929 Fig. 32. Examples of boudinage in glaciers. a) Line drawing from photograph by
1930 Lawson in Hubbard and Sharp (1989, Fig. 5) of boudins in debris-free ice (blue)
1931 embedded in debris-rich ice at the base of Variegated Glacier, Alaska. b) Boudins in
1932 sand layers within clean and bubbly ice on Sandy Glacier, Victoria Land, Antarctica
1933 (Dort, 1967, Fig. 5). c) Boudins in 10mm thick layer of sand near the base of Suess
1934 Glacier, Antarctica (Fitzsimons, 2006, Fig. 65.3). Penknife blade for scale. d) Results
1935 of numerical model to simulated boudins developed in ice-filled crevasses observed
1936 on the surface of an outlet glacier of the Framnes Mountains, Antarctica (Marmo and
1937 Wilson, 2000). The individual modeled boudins are about 1 m long and have this

1938 form about 1000 m from the crevasse field, with estimated strains since flow over
1939 that distance of $\epsilon_y = 1.6$ and a shear strain, γ_{xy} , of 0.6.

1940

1941 Fig. 33. a) Schematic illustration of the common concept of thrust faults bringing
1942 basal debris to the surface at the front of glaciers (after Moore et al., 2010). b)
1943 Interpretation of structures including thrusts and folds at the front of Bakaninbreen,
1944 Svalbard, following a surge in 1985-86; based on ground-penetrating radar (after
1945 Murray et al., 1997).

1946

1947 Fig. 34. a) Fold in longitudinal foliation with steep plunge, cut by ice vein, in
1948 marginal ice of Storglaciären. There is no new axial planar foliation here. The fold is
1949 displaced across the ice vein by an apparent strike-slip fault, but not by shear along
1950 a fault or shear zone. b) Folds in primary stratification and multiple axial planar
1951 fracture traces with offsets of stratification, exposed on a crevasse face on
1952 Kebnepakteglaciären, Sweden. Displacement was not caused by shear fracturing.

1953

1954 Fig. 35. Schematic diagram to illustrate the structures at the end of the 1982-1983
1955 surge of the Variegated Glacier, Alaska, showing strain rate histories during the
1956 surge at three locations with different histories (after Lawson et al., 1994).

1957

1958 Fig. 36. Map of the ablation zone of Storglaciären, Sweden, to show generalized
1959 variation in orientation of fold hinges depending on location in ice flowing with a
1960 single “tectonic transport” direction. Horizontal hinges are for folds in primary

1961 stratification, with axial planar foliation in the marginal ice; at the front of the
1962 glacier the hinges are for folds in a composite foliation that may be of sheath type.
1963 Vertical hinges are for folds affecting composite foliation in marginal ice. The
1964 foliation is everywhere steep to vertical. Based on data collected in 1979, 1980.
1965

Fig. 1

Fig. 2

Fig. 3

b Decrease in longitudinal velocity with depth, ma^{-1}

c

Fig. 4

Fig. 5

Fig. 6

Fig. 7

Fig. 8

Fig. 9

Fig. 10

Fig. 11

Fig. 12

Fig. 13

Fig. 14

Fig. 15

Fig. 16

Fig. 17

Fig. 18

Fig. 19

Fig. 20

subhorizontal
primary stratification

undulose stratification
amplified to form folds

folds tighten and start
to become dismembered
as foliation develops

stratification
mostly transposed
into foliation

increased shortening

Fig. 21

Fig. 22

Fig. 23

Fig. 24

Fig. 26

Fig. 27

Fig. 28

Fig. 29

Fig. 30

Fig. 31

Fig. 32

Fig. 33

Fig. 34

Fig. 35

Fig. 36