Thrust faulting in glaciers? Re-examination of debris bands near the margin of Storglaciären, Sweden

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Thrust faulting has been suggested as a viable mechanism of debris transport at many glaciers, often inferred from the presence of up-glacier dipping bands of debris that emerge at the ice surface to form ridges of basally derived material. However, modelling indicates that the development of thrust faults is mechanically inhibited because stresses are much lower than that required for shear failure, a prerequisite for thrust faulting, and field measurements fail to detect thrust-related displacement. The mechanism for the emplacement of these ridges that appear at the surface of many polythermal valley glacier termini remains open to question. This study re-examines the origin of debris ridges on the surface of Storglaciären, a polythermal valley glacier in northern Sweden, using field observations, ice microstructural analyses, sediment grain size analysis, stable isotope composition of the ice, and modelling. We find no evidence of discrete displacement across the debris bands that produce the ridges, nor do we find evidence that folding might be responsible. We propose that the bands originate at the base of the glacier by one of two mechanisms, perhaps in combination: i) refreezing of meltwater near the thermal transition in basal ice, and ii) injection into tensile fractures periodically opened at the base due to high fluid pressure and then freezing. In either case, separation from the base occurs due to high fluid pressure and freezing introduces ice below the debris bands, which are then transported forwards due to basal shear and upwards due to longitudinal compression, and revealed by surface ablation.

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Thrust faulting has long been proposed as a mechanism for upward transport of basal debris in regions of strong longitudinal compression in glaciers (e.g. Chamberlin 1895; Salisbury 1896; Chamberlin & Salisbury 1904; Ward 1952; Bishop 1957; Boulton 1970; Hambrey & Huddart 1995; Alley *et al.* 1997; Kamb *et al.* 1985; Glasser *et al.* 1998, 1999, 2003; Hambrey *et al.* 2005), such as at the base of an icefall, or where ice flows against the reverse slope of a subglacial overdeepening (e.g. Goodsell *et al.* 2002; Cook *et al.* 2011; Swift *et al.* 2018). It has been invoked to explain bands of basally derived debris within the ice (e.g. Boulton 1970; Tison *et al.* 1993; Sharp *et al.* 1994; Hambrey *et al.* 1996), and exposed at the glacier surface by ablation to form ridges (Fig. 1; e.g. Lliboutry 1965, 2002a; Goldthwait 1973; Glasser *et al.* 1998; Glasser & Hambrey 2001; Hambrey *et al.* 2005). Evidence consistent with thrust faulting includes the up-glacier dip of these bands (Glasser *et al.* 2003), and the fact that many such bands appear to be continuous to the glacier bed (e.g. Jansson *et al.* 2000; Glasser *et al.* 2003).

Originally, it was envisaged that basal debris could be 'scraped' into the overlying ice along shear planes (Chamberlin 1895). Weertman (1961) was critical of this concept and proposed instead that the debris was entrained in the ice by the freezing of meltwater at the glacier bed. This is expected to occur in association with fluctuations of the positions of the basal thermal transition (BTT) from the refreezing of meltwater via loss of heat by conduction through the ice (Weertman 1961; Robin & De 1976). An alternative to entrainment by the freezing of meltwater, where basal effective stress is high, is regelation infiltration, as ice is forced into the substrate (Iverson 1993; Iverson & Souchez, 1996; Glasser *et al.* 2003; Iverson *et al.* 2007). A further possibility is the emplacement of debris subglacially or englacially by water flowing in englacial channels or fractures and then freezing due to supercooling following rapid decompression (Lawson *et al.* 1998; Ensminger *et al.* 2001; Roberts *et al.* 2002; Cook *et al.* 2006). Once entrained by one of these mechanisms, it is commonly assumed that basally-derived debris is elevated along thrust planes or shear planes (Fig. 1; e.g. Tison *et al.* 1993; Bennett *et al.* 1996a, b; Hambrey *et al.* 1996, 1999; Murray *et al.* 1997; Hambrey & Dowdeswell 1997; Glasser *et al.* 1998, 1999, 2003; Herbst & Neubauer 2000; Glasser & Hambrey 2001). Another possible explanation is that the compressional regime at the margins of ice masses, where superimposed ice is formed, will, upon glacial advance, lead to natural upward movement of debris-bearing ice toward the ice surface without the need for thrusting (Hooke 1973), and along trajectories that resemble the proposed thrust surfaces (Hooke & Hudleston 1978).

Folding has been proposed as an alternative to thrusting to explain the elevation of debris-rich bands into the glacier (Moore *et al.* 2013). It has also been proposed as a mechanism that occurs in association with thrusting (Alley *et al.* 1997; Hambrey & Dowdeswell 1997; Glasser *et al.* 1998, 2003; Glasser & Hambrey 2001). One concept is that folding serves as a precursor to thrusting (Clarke & Blake 1991; Glasser & Hambrey 2002; Goodsell *et al.* 2002; Swift *et al.* 2006), and more specifically Murray *et al.* (1997) suggested that thrusts develop from fold hinges formed in debris-rich ice at the glacier base in response to intense longitudinal compression. Other studies have suggested that the longitudinal compression required to generate a thrust fault in glacier ice could be experienced at the BTT, or slip to no slip transitions (SNST), which are typical of polythermal glaciers (Clarke & Blake 1991; Lliboutry 2002b; Moore *et al.* 2010). According to this concept, faster-flowing, warm-based ice from up-glacier flows against cold-based and perhaps stagnant ice close to the glacier terminus, generating sufficient stress to propagate a thrust fault and elevate basal material along the thrust (Clarke & Blake 1991; Lliboutry 2002b; Moore *et al.* 2010).

There is no argument about the fact that basally-worked sediment appears as bands within basal ice and ultimately may appear on the surface of glaciers and that there must therefore be a process or suite of processes that incorporate the sediment into the ice and transport it to the surface. Despite the appeal of the englacial thrusting hypothesis, however, the only unequivocal evidence for thrust faulting in glaciers was documented in the Variegated Glacier in Alaska during its 1982-83 surge (Kamb *et al.* 1985; Lawson *et al*. 1994; Eisen *et al.* 2005). The fastest strain rate measured during this surge was 2 x 10^{-6} s^{-1} (Kamb *et al.* 1985). By comparison, maximum strain rates in non-surging glaciers are on the order of 10^{-9}–10^{-8} s^{-1} (Raymond 1971; Moore *et al.* 2010). Peak compressive strain rates during the Variegated Glacier surge were reached as the glacier surge encountered stagnant marginal ice at the terminus of the glacier. This caused a combination of unequivocal imbricate style faulting, buckling and thrusting near the snout of the glacier, accounting for ~50% of the shortening (Moore *et al.* 2010; Lawson *et al.* 1994). Faults were striking parallel with the overall foliation, and many were rooted at the glacier bed. It is important to note, however, that calculations by Moore *et al.* (2010) indicate that even the highest compressive strain rates at the snout of the Variegated Glacier were theoretically an order of magnitude too small to induce fracture. The dips on the thrust faults in the Variegated ranged from 40° to 80° up-glacier. Lawson *et al.* (1994) suggested that this variability might be a result of faults originating as tensile fractures that were then activated as thrusts during the surge. Faulting becomes more feasible in pre-existing fractures that fill with water to a certain height corresponding to hydrostatic pressure, as this decreases the frictional resistance allowing slip (Moore *et al.* 2010).

The development of thrusting has been inferred, but not demonstrated, in many other surge-type glaciers (Raymond *et al.* 1987; Bennett *et al.* 1996b; Hambrey *et al.* 1996; Hambrey & Dowdeswell 1997; Murray *et al.* 1997; Hambrey *et al.* 1999; Woodward *et al.* 2003) that have brief periods of rapid ice motion with velocities reaching up to 100 times the normal rate of sustained flow (Sharp *et al.* 1988). The acceleration phase during surges is hydrologically controlled by high subglacial water pressure that reduces basal friction, which modulates the velocity (Sharp *et al.* 1988).

There are several physical conditions that are required for ice to fail by shear fracturing, which are not expected to be met in valley glaciers (Hooke & Hudleston 1978; Hooke 2005; Moore *et al.* 2010). Stresses that result in the failure of ice by shear fracture are much larger than stresses that have been inferred in glaciers (Moore *et al.* 2010). Application of fracture mechanics to the conditions found in polythermal glaciers indicate that peak compressive strain rates are six orders of magnitude too small to induce fracture in the ice (Moore *et al.* 2010). Additionally, while debris ice mixtures have been studied extensively, and debris concentration, particle size, temperature and solute concentration can impact the strength of ice, conflicting observations from field measurements and experimental work make it difficult to quantify these constitutive properties (Arenson *et al.* 2007; Moore 2014). Theoretically, frictional slip on preexisting fractures may be possible if the ice is thin, undergoes large horizontal compressive strain rates, and a large sediment-bearing fracture already exists and is in hydraulic communication with high water pressure at the glacier base (Moore *et al.* 2010). However, these conditions are extreme and probably uncommon in natural glacial systems. They are most likely to exist, if at all, over short time periods during glacial surges, as in the case of the Variegated Glacier.

Excellent examples of debris ridges are found on the surface of Storglaciären, a small, non-surging polythermal valley glacier in the north of Sweden (Fig. 2). Valuable information about these ridges has been obtained in a number of studies over the past twenty-five years, and several different explanations for their entrainment and elevation to the glacier surface have been advanced, including, for entrainment: regelation of ice into the substrate (Iverson & Semmens 1995; Iverson 2000; Glasser *et al.* 2003), freezing of meltwater near the BTT (Moore *et al.* 2013) and deposition in englacial channels or fractures (Iverson *et al.* 1995; Glasser *et al.* 2003; Moore *et al*. 2011, 2013); and for elevation: by thrusting (Glasser *et al.* 2003), shearing (Jansson *et al.* 2000), and folding (Glasser *et al.* 2003; Moore *et al.* 2013).

Differential stresses in glaciers are typically small, attaining maximum values of about 0.1 MPa (Cuffey & Patterson 2010; Hooke 2005). Under such stresses, ice behaves in a ductile manner under compression, but it fails brittlely in zones of extension at the surface of glaciers, with tensile fractures developing perpendicular to the maximum extending strain rate (Hambrey & Lawson 2000; Hudleston 2015). Velocity gradients in flow cause previously formed fractures to rotate and ultimately close as they enter compressive flow (Hambrey & Lawson 2000; Hudleston 2015). Healed fracture zones can be identified as coarse, clear veins, often known as “blue bands” or fracture traces that become part of the planar fabric as the ice continues to deform. Fracture traces can be used as markers to record cumulative strain since formation (Hooke & Hudleston 1978; Hambrey & Lawson 2000; Hudleston 2015). Displacement of pre-existing markers across fracture traces may result from rotation and healing of fractures, but the displacement is not a result of faulting due to shear failure (Hudleston 2015). Shear failure in ice by fracture occurs under strain rates orders of magnitude faster than those found in glaciers (Schulson 2001).

This study aims to address the debate about the possibility of thrusting in glaciers by re-assessing the origin of the debris on the surface of Storglaciären using field observations, microstructural analyses, sediment grain size analysis, stable isotope composition of the ice, and illustrative modelling. Specifically, we take advantage of the previous work by Jansson *et al.* (2000), Glasser *et al.* (2003), and Moore *et al.* (2013), and provide additional data and observations that may help better explain the origin of these features and that, in particular, allow critical evaluation of the possibility that thrusting played a role in their origin.

Debris ridges on Storglaciären

Storglaciären is located in the Tarfala valley of Swedish Lapland (67º55'N, 18º35'E) (Fig. 2). It is 3.2 km long, extending in an east-west direction and has a total surface area of 3.1 km^2. The glacier has been well studied since 1886 (Svenonius 1910), and has been subject to mass balance measurements since 1946 (Holmlund & Jansson 1999). These suggest a 28% volume loss between 1910 and 2015 (Holmlund & Holmlund 2019). Mass loss, however, began to stabilize as the terminus retreated towards a prominent riegel (Holmlund & Holmlund 2019), combined with cooling of the front part of the glacier due to thinning, decreasing its velocity (Delcourt *et al.* 2013).

Storglaciären is characterized as temperate (~0 ^{o}C) but with a cold (~ –4 ^{o}C) surface layer of variable thickness (Hooke *et al.* 1983b; Holmlund & Eriksson 1989; Pettersson *et al.* 2007). In the ablation area, this cold surface layer reaches a maximum thickness of 50 m along the southern margin, but is 20-30 m thick along the centreline; the northern margin of the terminus is cold-based (Hooke *et al.* 1983a; Holmlund & Eriksson 1989; Moore *et al.* 2011; Gusmeroli *et al.* 2012). Measurements undertaken by Moore *et al.* (2011) indicate that the cold surface layer comes into contact with the bedrock within ~50-150 m of the terminus. The thermal regime influences glacier dynamics: the glacier undergoes basal sliding and internal deformation at its centre, but primarily undergoes internal deformation at the terminus and margins where it is frozen to the bed and valley walls (Holmlund *et al.* 1996). Electrical resistivity measurements by Brand *et al.* (1987), made in a series of boreholes drilled to the base of the glacier, indicate the existence of a ~0.4-0.7 m-thick layer of basal till, which has undergone pervasive deformation (Brand *et al.* 1987; Iverson *et al.* 1995). This weak substrate limits longitudinal compression near the centre of the terminus where the bedrock slopes down-valley (Moore *et al.* 2011). Closer to the northern margin, the glacier encounters a small moraine mound, which increases longitudinal compression locally (Hooke *et al.* 1987). Moore *et al.* (2011) showed that the frozen toe of the glacier provides limited resistance to forward movement of the ice from up-glacier.

Seasonal variations in surface velocity on Storglaciären show consistent peaks in the early summer (Brzozowski & Hooke 1981; Hooke *et al.* 1983b, 1989). Diurnal changes in velocity directly correlate with changes in subglacial water pressure, which is highly variable at Storglaciären. The glacier's hydrological system during the melt season characteristically involves sudden inputs of rainwater (Hooke & Pohjola 1994; Iverson *et al.* 1995). Similar seasonal and short-term variations in velocity have been documented for many other glaciers (e.g. Meier 1960; Müller & Iken 1973; Iken 1981; Iken & Bindschadler 1986; Iken & Truffer 1997).

The overdeepened, upper part of the ablation zone of Storglaciären drains primarily through englacial conduits, which enlarge throughout the melt season (Hooke *et al.* 1988; Hock & Hooke 1993; Hooke & Pohjola 1994; Fountain *et al.* 2005a, b). The lower part of the ablation zone is characterized by faster subglacial drainage (Seaberg *et al.* 1988; Hock & Hooke 1993). Increased water pressure can cause a decoupling between the glacier sole and the bed (Brand *et al.* 1987; Iverson *et al.* 1995). Video observations by Hooke & Pohjola (1994) in boreholes in 1990 showed fine sediment can be consistently held in suspension near the bed due to high water pressure, and, at times of extreme pressure, water and fine sediment can extrude out of the surface of the glacier through preexisting fractures (P. Jansson, pers. comm. 2018).

Crevasse fields are associated with overdeepenings and are prominent above the upper two basins of the glacier and along its northern margin in the lower ablation zone (Glasser *et al.* 2003). Transverse fracture traces along the centreline of the glacier are semi-continuous, and have an arcuate shape in plan view due to the velocity gradient across the glacier. These traces become progressively more accentuated as the ice flows down-glacier. Deviations of fracture strikes from perpendicularity to the maximum extensional strain rate occur due to rotation of crevasses after formation and before closure. As fractures rotate with progressive deformation, they enter a compressive regime and close (Hudleston 2015). Healed fractures can be identified as coarse, clear veins, and rotate towards parallelism with the direction of maximum shear strain rate along the glacier margins and base, due to the velocity gradient within the ice, contributing to the overall fabric (Hooke & Hudleston 1978).

A debris band composed of diamicton was first identified in 1994 by Jansson *et al.* (2000) on the northern side of the terminus of Storglaciären (e.g. Fig. 3). This band, like subsequent bands that have appeared in the same general location (Glasser *et al.* 2003; Moore *et al.* 2013), stood out as a ridge due to differential ablation, with the debris once reaching the surface shielding the ice below from melting (Fig. 3). Ground penetrating radar (GPR) (Jansson *et al.* 2000) surveys indicated that this layer was continuous to some depth, projected to the base of the glacier, and appeared to originate at the transition between warm-based ice and the cold-based ice at the terminus (Jansson *et al.* 2000). Stakes at 50 m intervals indicated that the ridge coincided with a >50% decrease in surface velocity across a transect that crossed the debris ridge (Jansson *et al.* 2000). Glasser *et al.* (2003) found several debris-charged ridges, one of diamicton and two they identified as being glaciofluvial in origin and they interpreted all these as having been brought to the surface by thrust faults as a result of longitudinal compression across the slip-no slip transition at the glacier bed. Moore *et al.* (2013) described two types of debris bands, one of basally-derived diamicton, and the other of well sorted sand and gravel that they interpreted as having been deposited in an englacial channel. On the basis of isotopic analysis, they concluded that the diamicton band was younger than the underlying sand and gravel bands. The isotope data and the apparent lack of sufficient relative movement of the ice on either side of the ridge required by thrusting, led the authors to tentatively suggest that folding, based on modelling passive markers, combined with the upward flow in ice under compression at the toe, brought the bands to the surface and explained the inversion of the position of the two types of bands.

Material and methods

Fieldwork

Fieldwork included detailed descriptions of the surface expression of the debris ridges (Figs 3, 4, 5), mapping of structural fabric, including primary stratification, foliation, and fracture traces (Figs 5, 6), and the collection of ice samples in 2016 and 2018 (Fig. 2) for microfabric analysis. In addition, sediment and isotope analysis sampling was undertaken. The character of near-surface ice is affected by ablation, thus samples were collected from beneath this surface layer, which is typically up to ~20 cm thick. Damaged ice was removed and samples (blocks with vertical sides) were then cut using a chainsaw, oriented such that the top of the block was parallel to the glacier surface and the vertical edges parallel and transverse to the flow direction. Samples were immediately shaded with a tarp upon removal to avoid damage and reduce ablation. Ice underneath the ridges was accessed manually using ice axes, water buckets and the chainsaw.

Ice movement

Differential GPS was employed using the down-glacier movement of six stakes across two of the frontal debris ridges (ridge 1, the southernmost ridge near the centre of the glacier and ridge 2, 20 m north of ridge 1) (Figs 3, 4, 7) to assess the relative movement of ice above, within and below the debris ridges (e.g. Fig. 4). Positional measurement errors were limited to 1-2 cm on the stakes above and below the debris ridges. Two sets of three stakes were installed five metres vertically into the ice using a steam drill, and initial positional measurements were taken on August 11, 2018. We recorded a second set of positional measurements on September 7, 2018, and a third set on August 4, 2019 (Fig. 7). The stakes in the centre of each ridge (1C and 2C) were difficult to emplace vertically due to the slope of the ridge, making their displacement more difficult to measure accurately. Additionally, the outline of each ridge was traced using dGPS in August and September 2018 (Fig. 7), by walking the debris contact with the surface of the glacier. This step could not be repeated in August 2019 due to loss of ice and the deflation of the debris ridges. One metre of ice ablated during the one-month interval in 2018. Three positional measurements were taken around each stake and averaged.

Sediment grain size distribution

Sediment grain size analysis was undertaken on different sediment-laden ice facies in order to provide insight into the origin of the debris bands (Fig. 8). Specifically, the grain size distribution of sediment included within debris bands was compared to that of basal till and basal ice, which have both been suggested to have been the source of debris band materials in previous studies (Glasser *et al.* 2003; Hubbard *et al.* 2004; Cook *et al.* 2011; Swift *et al.* 2018). Samples of debris band, frozen basal till and basal ice were collected from the glacier using an ice axe after first having removed the top ~20 cm of ice affected by surface ablation, which may be affected by percolating surface meltwater and precipitation (Fig. 2; cf. Toubes-Rodrigo *et al.* 2016). Sediment particle size analysis was performed on the sediments extracted from the stable isotope ice/water samples through vacuum pump filtering (see below). After filtration, the samples were returned to the laboratory for wet-sieving from -3 Φ (8 mm) to 1 Φ (0.5 mm), with finer material collected in the pan at the base of the sieve stack; laser granulometry was then conducted on this finest material down to 12 Φ (0.24 μm).

Stable isotope composition

Oxygen and hydrogen isotope analysis ($\delta^{18}O$ and δD) has been undertaken in several studies of glacial sediment entrainment in order to aid interpretations of the origins of debris-bearing ice structures (Fig. 9; e.g. Iverson & Souchez 1996; Hubbard *et al.* 2004; Cook *et al.* 2010, 2011; Moore *et al.* 2013; Swift *et al.* 2018). For this study, samples were collected from glacier ice and snow, which represent meteoric waters, and the various sediment-laden ice facies observed around the glacier margin, including debris bands, basal ice facies, and frozen basal till (Fig. 2). These samples, of ~30 to 100 mL, were allowed to melt in sealed sample bags. These were then filtered within 24 hours of collection through 0.45 μm cellulose nitrate filter papers and bottled in 8 mL Nalgene bottles. The resultant water samples were kept refrigerated until their analysis at the National Environmental Isotope Facility at the British Geological Survey. Oxygen isotope ($\delta^{18}O$) measurements from water were made using the CO_2 equilibration method with an Isoprime 100 mass spectrometer plus Aquaprep device. Deuterium isotope (δD) measurements from water were made using an online Cr reduction method with a EuroPyrOH-3110 system coupled to a Micromass Isoprime mass spectrometer. Isotope measurements use internal standards calibrated against the international standards VSMOW2, VSLAP2, VPDB. Errors are typically ± 0.05‰ for $\delta^{18}O$ and ± 1.0‰ for δD. Statistical analysis of the difference between regression lines of meteoric waters and debris bands was conducted using an ANOVA test in MATLAB.

Microstructural analysis

Microstructural information, including crystallographic preferred orientations, was obtained from three orientated samples collected across one of the ridges: one up-glacier from the ridge (SG20), one from the ridge itself, just below the debris band (SG11-B), and one down glacier from the ridge (SG21) (Fig. 4). Thin sections were made from sample SG11-B (Fig. 10), and crystallographic c-axis orientations were measured (Fig. 10) for these using a Rigsby universal stage. This work was carried out at the University of Wisconsin, Madison. Complete lattice orientations (c-axes and a-axes) were subsequently measured for SG20 and SG21 using electron backscatter diffraction (EBSD) at the University of Otago, New Zealand, which has a facility for work on ice that can accommodate 4 x 6 cm thin sections, necessary for analysing the coarser grain-size typically associated with warmer ice. One advantage of EBSD over the Rigsby stage is that it has much higher accuracy, and a second is that crystallographic a-axes are determined in addition to c-axes. This allows for better assessment of deformation mechanisms (Monz *et al.* 2021). The Rigsby stage, however, accommodates larger sections (10 x 10 cm) and is better for analysing grain size and shape.

For all analyses, the data were plotted in lower hemisphere, equal-area projection, both in geographic coordinates and in a coordinate frame tied to the orientation of foliation in the ice samples.

For SG11-B, some idea of grain size and shape were examined directly from the thin section viewed under crossed polars. The two EBSD samples were first examined using low-angle light on polished surfaces of samples (Fig. 10) cut perpendicular to the planar fabric (foliation), the samples then being allowed to sublimate overnight (Monz *et al.* 2021). A newly developed sample preparation technique (Monz *et al.* 2021) allows for the measurement of bulk lattice orientations using a scanning electron microscope fit with an EBSD camera and a custom cryo-stage (Prior *et al.* 2015), and the production of plots (pole figures) to represent preferred crystallographic orientation (CPO) within a sample without requiring an excessive number of sections to establish a pattern (Monz *et al.* 2021). The pole figures for SG20 and SG21 (fig. 10) were plotted using the MTEX texture analysis software package that runs in MATLAB.

Flow modelling

It is important to take into account the pattern of flow in the frontal part of the glacier where the debris ridges emerge. Several models of flow in Storglaciären have been made. First by Hanson (1995), who utilized a coarse-mesh three-dimensional finite element model for the whole glacier, choosing flow-law parameters that best matched measured surface and borehole velocity measurements (Hooke *et al.* 1989; 1992; Jansson & Hooke 1989; Hanson & Hooke 1994) and taking basal topography as determined by radio-echo sounding (Holmlund & Schytt 1987; Eriksson 1990). Pohjola (1996) used measured velocities to track particle paths in order to estimate how crevasse traces in the upper part of the glacier were transported and later exposed in the ablation zone. Moore *et al.* (2011) utilized fine-mesh two-dimensional finite element models of the frontal part of the glacier and found that the model velocities did not match measured velocities if a transition from slip to no-slip occurred at the BTT near the cold front of the glacier, implying the frozen toe provided limited resistance to basal slip.

To illustrate how planar features in the ice, such as fracture traces and debris bands, are expected to behave towards the glacier front in the area where the debris bands are exposed, we utilize Hanson's (1995) model. The model is used mainly for illustrative purposes, recognizing that the grid used by

Hanson was coarse, with sparse control in this region and using basal conditions of no slip under the frozen toe that have been shown by Moore *et al.* (2011) to be questionable. By using it, we are not claiming to follow the actual flow history of the glacier up to the time that our observations were made in 2016 and 2018. The glacier is not in steady state on any time frame (Hooke *et al.* 1989; Holmlund & Holmlund 2019), yet the mapped outline of the glacier and the surface expression of structural features (bedding and fracture traces) have not changed significantly since the map that Hanson (1995) used was made, even though overall the glacier has been retreating since 1910 (Holmlund *et al.* 2005; Holmlund & Holmlund 2019). In particular, it should be noted that the location of the debris bands was very nearly at the same geographic coordinates and elevation in 2016 and 2018 as it was in 1994 (Jansson *et al.* 2000) and 2001 (Glasser *et al.* 2003). This observation implies a rough balance between flow and ablation, as noted by Jansson *et al.* (2000). The bands have, however, become closer to the glacier front as some overall frontal retreat and thinning have occurred. In our approach, we follow the expected tracks of particles in steady state, and thus of any planar features induced in the distal part of the glacier, and consider, in a general way, how these may help explain the observed debris bands and associated planar features at the ice surface.

We used a 2-D subset of Hanson's 3-D gridded velocities by following a profile through the debris ridges in the ablation zone of the glacier (Fig. 11). Lateral velocities at these nodes (in the y direction) were ignored, although their existence is reflected in their effect on the velocities in our domain in the xz plane through conservation of mass in 3D. Where the ice is less than 30 m thick, basal velocities are taken as zero, implying that ice is frozen to the bed. Within the selected domain, particle paths were derived using positional coordinates of each node and the associated velocities (Fig. 11) using methods outlined by Hudleston & Hooke (1980). We created several sets of initially planar features (linear in 2D) representing potential healed hydraulic fractures introduced at the base of the glacier in the region of the BTT: one with an initial vertical orientation and one at an initial orientation of ~45° to the shear plane. Each of these features was defined using four particles, which were tracked at three separate time intervals to identify how they changed during glacier flow (Fig. 11). Additionally, a sub-planar feature at the base of the glacier oriented initially approximately parallel to the bed was tracked to assess the possibility of passive folding within the ice. Due to model boundary conditions, the initial position of the feature had to be raised ~0.5 m above the ice—bed interface.

Results

Field observations

The appearance of the debris-covered and ice-cored ridges in 2016 and 2018 is shown in Fig. 3; surface debris insulates the underlying ice, leading to differential ablation and a protruding surface expression. The overlying debris is a remnant of the englacial diamicton bands, as defined by Moore *et al.* (2013). After ablating out of the ice, the debris remains on the surface, and is poorly sorted with a grain size ranging from sand to sub-rounded to rounded boulders. The up-glacier sides of the ridges are covered with ~10 cm of poorly sorted debris, and they are bounded on their upper sides by a sharp contact with clean ice (Fig. 4). The down-glacier sides of the ridges are covered by a thin layer (~3 cm) of poorly sorted sediment, and have a fan-shaped appearance, grading into the debris-free surface ice (Fig. 4). Most of the cobbles on the down glacier side have slipped to the base of the ridge. We did not find any bands with inverse grading from sand to gravel, as described by Moore *et al.* (2013: Fig. 5d). The ridges appear as a series of transverse, laterally discontinuous pockets, sometimes at two levels at the glacier surface (Fig. 3), with an apparent lateral continuation into the ice as a distinct surface between bands of relatively clear ice that dip up-glacier 30°-70°, decreasing in steepness along-strike towards the centre of the glacier. Individual pockets of debris appear as ice-cored ridges on the surface, which have abrupt, and rounded lateral boundaries. Their height ranges from 0.5-1.5 m, generally decreasing along-strike towards the centre of the glacier. The bands are approximately parallel to one another, and their strike is parallel to that of the local foliation. Thus, they are in accord with the overall arcuate expressions of foliation and fabric at the front of the glacier (Fig. 2).

The surface expression—length, width and height—of the ridges varies with ablation rate. In 2018, the exposed velocity stakes reflected removal of about one metre of ice due to ablation in one month (Fig. 3). Between 2016 and 2018, ridges 2, 5 and 6, which were <0.5 m high in 2016, were ~1.0 m high in 2018 (Fig. 3); ridge 1, which was not present in 2016, was <1 m high in 2018. Ridges 3 and 4 shrank in height, and ridge 7 had become a moraine at the front of the glacier, rather than a feature within the ice. Additionally, over a period of one month in 2018, the ridges appeared to grow in areal extent due to differential ablation (Fig. 7A). The northern ridge exhibited minor down-glacier expansion while the southern ridge exhibited significant lateral expansion and down-glacier expansion.

Subsurface investigation of the extent of the bands using a steam drill is possible because such drills are unable to penetrate through decimetre thick debris. Steam drill investigations ~1 m upglacier from the debris ridges 1 and 2 in 2018 (Fig. 3), however, encountered no resistance in drilling 5 m through the ice, indicating the absence of a debris barrier. This was consistent with additional investigations using a chainsaw on the upglacier side of the debris bands, in which debris was not found at any depth.

Ice facies

Following Glasser *et al.* (2003), we characterize ice facies based on physical properties, including bubble content, debris content, and crystal size. We identify three different types of ice that form the ridge beneath the debris, and also the ice at the front of the glacier: i) coarse-grained (>1 cm), bubble free ice, ii) coarse-grained (>1 cm) bubble-rich ice, and iii) localized fine-grained (<< 1 cm) bubble-rich bands (bubbles <1 mm in diameter; bands ~2 mm in width) that cross-cut foliation. Alternating layers of coarse-grained, bubble-free ice and coarse-grained bubble-rich ice define a planar fabric in the ice underlying the surficial debris bands that strikes SSE near the centre of the glacier, and dips gently up-glacier at ~15°-20°. This is consistent with the overall composite planar fabric that defines foliation at the front of the glacier (Figs 5, 6).

Ice movement

There was no significant difference in horizontal displacement of the debris free ice on either side of the two debris ridges monitored (stakes 1W and 1E and 2W and 2E), either for the short period in one summer or over a full year, giving an average annual velocity of 1.6 m a^{-1} for the north ridge and 2.1 m a^{-1} for the south ridge (Fig. 7). As expected the displacement decreases towards the north margin of the glacier. The data are below the precision that would allow us to determine longitudinal strain rate over this short distance. The central stakes in each of the two ridges yielded questionable results, especially within the first month, likely due to initial tilt of the stakes during emplacement, and subsequent tilting during ablation. The deviations in displacement between these and the outer stakes are therefore considered spurious.

Sediment grain size distribution

The particle size distributions of debris bands and the solid facies ice (i.e. the frozen basal till) share several similarities in that all samples show a pronounced peak at 4 Φ (very fine sand/coarse silt), show a tailing-off toward the finest grain sizes, and are moderately well-represented in the coarser end of the distributions, albeit with more variability for the coarsest grain sizes examined here (Fig. 8). Basal ice has much greater variability between debris-poor (i.e. clean and bubble-laminated ice) and debris-rich (i.e. laminated ice) ice types, and within those two broad classifications too; in short, there is no consistent pattern to the grain size distributions for basal ice, and most samples do not share a particle size distribution similar to the debris bands or frozen basal till.

Stable isotope composition

The $\delta^{18}O$ and δD of debris-laden ice facies (frozen basal till, basal ice, debris bands) and meteoric ice types (glacier ice and snow) are shown in Fig. 9. A least squares linear regression line is plotted through glacier and snow sample compositions as a local meteoric water line (LMWL) against which other samples can be compared; the LMWL has a slope of 8.16 with an r^2 value of 0.99, and meteoric samples also represent the highest and lowest compositions in the dataset; the lowest $\delta^{18}O$ and δD are from a snow sample. The slope of 8.16 is consistent with the expected co-isotope relationship of meteoric water, which conventionally has a slope of ~8 (Craig 1961; Cook *et al.* 2010). An ANOVA revealed that the regression lines for meteoric waters and debris bands in ice have a p-value <0.5, indicating that they are statistically different.

The debris-bearing ice facies plot along regression slopes of between 5.90 to 6.68 with high r^2 values (between 0.95 and 0.99). These regression lines are based on four samples collected from frozen basal till, and four samples collected from debris bands. The slopes for both of these facies suggest that some fractionation has taken place in the formation of frozen till and debris band ice. In general, both frozen till and debris bands have similar but higher isotope compositions than basal ice, although there is some overlap.

Microstructure

Crystals within the debris bands and in the adjacent ice measure up to 70 mm across in thin section, but vary significantly with bubble content, and they show no shape-preferred orientation. The true size of many crystals greatly exceeds 70 mm. Due to the irregularity of the grain shapes, individual grains likely intersect thin section planes in multiple places, thereby appearing as several separate grains (Monz *et al.* 2021). Grain shape in the 2D thin section plane is also irregular, and grain boundaries are interlocking and lobate-cuspate. Consistently, larger crystals have smooth grain boundaries and triple junctions and define the coarse clear ice. In the bubble rich ice, air bubbles exist as a secondary phase and reside along grain boundaries, at the grain boundary triple junctions and in the centres of grains. Typically, there is an increase in the overall sutured texture and a decrease in grain size associated with areas that have a high density of bubbles. Individual grains are entirely recrystallized, with no evidence for undulose extinction, therefore lacking significant internal lattice distortion (Fig. 10). The c-axis fabrics are similar in samples above, within and below the debris ridge, with several maxima that are displaced 20-40^{o} from orthogonality to the foliation plane (Fig. 10). Sample SG21, structurally below the debris ridge, did not have enough grains measured to be considered statistically significant, but appears to have a skeletal fabric consistent with SG11-B and SG20. To allow comparison with other work, we computed the eigenvalues (following Woodcock 1997) to provide some indicator of c-axis fabric strength and symmetry for each of the CPO distributions, noting however that the eigenvalues provide limited meaning for multimaxima fabrics, with the maximum eigenvalue providing a weaker clustering than any of the individual maxima and the associated eigenvector situated among the individual maxima. The values and location of the maximum eigenvector are given in Fig. 10.

Flow modelling

Flowlines along the profile shown in Fig. 11 suggest that ice exposed at the location of the debris bands originates near the glacier base, between columns c and d; that is in the vicinity of the BTT, as noted earlier by Jansson *et al.* (2000). Planar features introduced into the ice column approximately parallel to the bed will follow paths close to the flow lines on which they lie. The velocity gradient in the upper part of the ice column is not high, and therefore the particles in such a scenario move at similar velocities along similar trajectories. Introduced closer to the bed, sub-planar features will passively rotate and may become folded based on the velocity gradient, with increased drag along the substrate (Fig. 11D). If folds develop, their axial planes rotate towards parallel with the foliation. Planar features introduced either vertically or at a 45^{o} angle to the bed will be affected by the bed parallel shear and rotate towards the flow direction. They will also become increasingly curved with time, with steeper dips near the top, reflecting the downward gradient of increased shear.

Discussion

The surface expression of the debris ridges on Storglaciären changed substantially between 2016 and 2019 (e.g. Fig. 3). Originally documented as more-or-less continuous features in 2016, the debris ridges appeared only as concentrated pockets of material on the glacier surface in 2018, and by 2019, some individual segments had been reduced to dispersed debris on the glacier surface. Lateral boundaries that characterize individual ridges appear to be similar to those identified by Moore *et al.* (2013). Our observations differ from those made by Glasser *et al.* (2003) and Moore *et al.* (2013), who noted the ridges present at two distinct levels on the surface, as we did in 2016, but described the surface expression as continuous (Jansson *et al.* 2000; Glasser *et al.* 2003; Moore *et al.* 2013). Additionally, Glasser *et al.* (2003) described the up-glacier ridge as sandy gravel in composition, and the down-glacier ridge as diamicton, whereas Moore *et al.* (2013) described the opposite stratigraphic sequence. We did not notice a compositional difference between the two levels of ridges, both being diamicton in composition. This difference in appearance reflects a change in surface expression over time, a change in source material, and indicates a finite sediment supply to these features.

In addition to the change in lateral extent and continuity of the debris bands, their extent back into the glacier appears to have diminished over time. Jansson *et al.* (2000) traced a debris band back into the glacier about 50 m using GPR. Moore *et al.* (2013) traced the debris bands about 10 m back into the glacier using a steam drill and established that the dip of the bands decreased away from the surface. By contrast, based on our steam drill investigation in 2018, the bands do not extend more than a metre into the ice, suggesting a discontinuous nature to the bands both in strike and in dip directions. Overall, the continuity of the bands decreased from the time of the GPR surveys undertaken in 1995 (Jansson *et al.* 2000) to the time of our measurements in 2016-2019. By 2019, many ridges had become subdued and diffuse, and were no longer supplying much debris to the glacier surface. The southern extent of the ridges, however, increased as new pockets of debris appeared along strike (Fig. 3).

The sediment grain size distribution of included debris, and the $\delta^{18}O$ and δD data suggest that the debris bands are of basal origin, and seem to indicate that till has been frozen and then elevated to an englacial position, before melting out supraglacially. Specifically, debris bands and basal till have similar grain size distributions (Fig. 8A, B). Both differ from the grain size distribution of basal ice (Fig. 8C), which is often produced by processes (e.g. regelation, glaciohydraulic supercooling) that preferentially include or exclude grains of different sizes (e.g. Hubbard 1991; Lawson *et al.* 1998). The sedimentological similarity between debris bands and basal till indicates that till has been incorporated into the glacier as debris bands without modification. The $\delta^{18}O$ and δD data also show compositional similarity between debris band ice and frozen basal till (Fig. 9); both have higher $\delta^{18}O$ and δD compared to other water and ice types sampled, and both plot along a freezing slope (albeit with small sample numbers). The dissimilarity with basal ice samples (which have lower isotope composition and plot along a steeper regression line similar to the LMWL) indicates that the debris bands are not composed of basal ice. Some previous studies suggested that debris bands are composed of basal ice that has been elevated to higher positions within the glacier (e.g. Hubbard *et al.* 2004; Swift *et al.* 2018).

If a shear zone existed, we would expect it to be in the ice immediately below the debris band. Texturally, however, there are no characteristic changes in the microstructure in the ice above and just below the ridges that would suggest the development of a discrete shear zone. Rocks in shear zones often undergo grain size reduction during deformation, which is not the case here (Fig. 10). Hudleston (1980) documented a decrease in elongate bubble plunge, an increase in grain size, an increase in c-axis preferred orientation, a slight shape- preferred orientation, and the development of a sutured grain structure associated with a shear zone in the Barnes Ice Cap. These textural changes are not apparent in the ice above or directly below the debris bands (Fig. 10). Additionally, both in Hudleston (1980) and in experiments on ice subjected experimentally to simple shear (Duval 1981; Bouchez & Duval 1982; Budd & Jacka 1989; Budd *et al.* 2013; Qi *et al.* 2019; Journaux *et al.* 2019), the most significant change is in the degree of c-axis preferred orientation, which is weakly developed if shear strain is modest, but increases to become a strong single maximum c-axis fabric at high shear strain. By contrast, the intensity of c-axis fabric in the ice above and below the debris bands is a weakly to moderately defined fabric with several maxima and one dominant maximum (Fig. 10). The most well-defined of these fabric diagrams (Fig. 10A) is similar to those found in highly sheared marginal ice (Monz *et al.* 2021). The skeletal fabric in sample SG21 (Fig. 10B) is likely due to a grain sampling bias as discussed in Monz *et al.* (2021), and does not reflect a real difference in CPO.

In the bubble-free ice directly below the debris bands on Storglaciären, no evidence exists of brittle deformation or sharp offset of planar features. There is, however, an increase in grain size, but a decrease in the overall suturing texture and an associated increase in polygonal grains with smooth grain boundaries (Fig. 10). This highlights an inverse correlation between bubble concentration and grain size, and between bubble concentration and boundary smoothness reflecting the bubbles existence as a secondary phase that can stop or slow grain boundary movement during recrystallization. There is no apparent grain shape-preferred orientation characteristic of discrete shear zones in rocks (e.g. Simpson 1983) and in ice (e.g. Hudleston 1980; Journaux *et al.* 2019). These microstructural characteristics seem to define a representative texture of bubble-free ice throughout the glacier, and are not unique to the ice below the debris bands.

On the basis of tritium concentrations, Moore *et al.* (2013) argued that the ice within the upper debris band was younger than the ice in the underlying sand-rich layer as well as in the overlying glacier ice. Such a relationship would be expected if the debris band was emplaced by a thrust or ductile shear zone. If thrusting played any role in the elevation of the bands to the surface, however, no evidence exists from the current kinematics or texture and fabric of the ice in and adjacent to the bands to indicate this. Macro- and micro-scale observations from 2016-2019 indicate that no sharp discontinuity occurs in the ice fabric between debris-bearing ice and the adjacent glacial ice, as seen on the surface today, that might indicate thrust faulting (Figs 5, 6, 10). Our dGPS measurements in the ice on either side of the debris bands (Fig. 7B, C), although taken over a limited time frame, show no significant decrease in velocity over a distance of 10-15 m. Thus we found no evidence for differences in horizontal surface displacement that would be required for active thrusting. It should be noted, however, that velocities derived from our measurements at the debris ridges are considerably less than those recorded 24 years earlier by Jansson *et al.* (2000), who measured horizontal velocities at points about 25 m upglacier (9.1 m a^{-1}) and downglacier (3.6 m a^{-1}) from the debris bands. The horizontal velocity measured by Moore *et al.* (2011) about 20 m upglacier from the southern end of the debris ridge in 2008 was about 3.5 m a^{-1}. These measurements are consistent with an overall decrease in velocity near the terminus with time.

Additionally, it should be noted that the large decrease in velocity measured by Jansson *et al.* (2000) across a distance of about 50 m included the exposed debris bands. Jansson *et al.* (2000) suggested this sharp decrease in velocity was associated with localized zones of shear in the bands of clean ice below the debris bands. Whether or not this is the case, there is no evidence of localized slip or shear in the ice, including bubble-free and bubbly ice, within about 5 m up and down glacier from the debris bands. In addition, in 2008, Moore *et al.* (2013) could not detect any slip offset across 1.5 m of ice containing a debris band. Based on their estimate of maximum age, assuming a thrust dipping at 30^{o}, these authors calculated that a vertical displacement of 0.7 m per year of the hanging wall with respect to the footwall would be needed. This would be equivalent to 1.4 m difference in horizontal displacement and thus 1.4 m a^{-1} difference in velocity, which was not detected.

Irregularities at the base of the glacier likely combine with variable water pressure in creating conditions leading to debris incorporation and entrainment. Where the effective pressure is high, such as the on the stoss side of the riegel at the base of Storglaciären, continuous or discontinuous layers of sediment can be incorporated into the base of the glacier by regelation (Fig. 12; Hooke & Pohjola 1994; Iverson 1993). In addition, near the BTT, debris may be incorporated into the ice by basal freeze-on (Fig. 12; Moore *et al.* 2013); the frozen basal till observed and sampled at Storglaciären is evidence of this. This can create layers containing >40% debris, which is similar to the concentration in the debris bands present on the surface of Storglaciären (Lawson 1979; Ronnert & Mickelson 1992; Iverson 1993; Glasser *et al.* 2003). The sedimentological and oxygen and hydrogen isotope data all support a genetic link between the frozen basal till and the debris bands (Figs 8, 9).

Incorporating debris into the ice by either regelation or refreezing of meltwater, however, cannot alone raise debris bands over older debris free ice. One mechanism that can achieve the initial separation from the substrate depends on the variability of fluid pressure, for which there is ample evidence as already noted based on *in-situ* measurements of basal water pressure (see Hooke & Pohjola 1994; Iverson *et al.* 1995; Moore *et al.* 2013). Once sediment is incorporated into the base of the glacier, we postulate, near the front of the glacier where the cold surface layer is frozen to the till, periodic high water pressure decouples the base of the glacier (containing adhered basal debris due to regelation or freeze-on of meltwater) from the underlying till (that is, the effective vertical stress becomes negative). This decoupling is then followed by refreezing of meltwater below the uncoupled base (Fig. 12). It is likely that the zone of separation is propagated forward as a tensile fracture into ice ahead of the detached segment of frozen debris. The refreezing of water below the debris and in the fracture creates the band, and this feature extends some distance into the older ice above and below it (Fig. 12). This places younger, clear ice on top of older ice. The natural shear gradient in the basal ice will then transport the debris band on to older ice. This process could be enhanced by some strain localization in the clear, young ice just below the debris band, as envisioned by Jansson *et al.* (2000).

Once entrained, the bands of debris must be elevated to the surface. Direct measurement and flow models (Fig. 11) indicate that longitudinal compression and vertical uplift occur in the frontal part of the glacier, and the presence of the debris bands dipping moderately to steeply up-glacier demands that this is the case. The magnitude of this effect, however, depends on the degree to which the frozen toe impedes forward movement, coupled with the variable basal velocities (described above). The fairly abrupt decrease in horizontal velocity close to the front of Storglaciären was attributed by Jansson *et al.* (2000) to the glacier being frozen to its bed in this region, leading to increased compression where the faster flowing temperate ice encounters the cold, slow-moving or stagnant frontal ice. The modelling of Moore *et al.* (2011) shows that the velocity field in the frontal part of the glacier does not match the measured velocities if the ice is frozen to the base at the front. These authors determined that longitudinal compression at the terminus of Storglaciären, underlain by deformable sediment, is limited in magnitude because the frontal ice is not stagnant. Instead, they concluded that longitudinal strain rates correlate more directly with bedrock topography (Moore *et al.* 2011), and they estimated excess longitudinal stresses of 20-25 kPa in this region, less than half what would be predicted if there were no slip at a frozen base. In any plausible scenario, the longitudinal compression, a feature of all models and observations, coupled with vertical extension implies that particles will naturally elevate englacially at the terminus of the glacier (Figs 11, 12). In such a situation, the trajectory of the flow path becomes steeper, and the ice will respond more to coaxial flow as the debris band is elevated higher above the base. In contrast, when the debris is at or near the base of the glacier, the ice responds more to basal shear associated with non-coaxial flow. Any shear localization at the early stage of elevation of the debris band is therefore replaced by more homogeneous deformation as the surface is approached, resulting in the observed pattern of foliation and microstructural fabric in the ice above and below the debris bands (Figs 5, 6, 10).

As described above, debris entrained by regelation or freezing-on, once detached from the base, will track particle paths and ultimately appear at the surface. The bands will define a steepening upward orientation and tend towards sub-parallelism with foliation (Fig. 12). The actual orientation of the bands, however, will depend on the interplay between the base-parallel shear, tending to decrease the up-glacier dip of the bands, and longitudinal shortening, tending to increase their dip, and this will change as the band elevates. Perturbation in flow near the base, or non-planar geometry of the band as it is entrained, may result in the band becoming folded as it is transported towards the surface, in the manner described by Hudleston (1976). The amount of shear involved, however, is unlikely sufficient to produce an isoclinal recumbent fold, with both limbs parallel to the flow lines, such as would appear to be necessary to produce the inverted stratigraphy described by Moore *et al.* (2013). It is possible that the folding could be amplified if the viscosity contrast between debris-rich ice and debris-free ice led to a sufficient mechanical instability and rapid fold growth, although the wavelength of such folds would be just a few times greater than the debris band thickness (e.g. Hudleston & Treagus 2010) and much less than the wavelength implied by the inverted stratigraphy described by Moore *et al.* (2013). Additionally, towards the centre of the glacier, where new debris bands (e.g. debris bands 1 and 2 in Fig. 3) had started to outcrop in 2018, there appears to only be one structural level at which these bands exist. This highlights a lack of evidence in favour of folding. If the debris bands were folded structures, the two limbs of the fold should outcrop at two structural levels and potentially meet at the fold hinge. Glasser *et al.* (2003) identified small-scale, complex folding in the ice adjacent to the debris bands, or directly underneath the debris on the lee side of the ice-cored ridges. These folds are often defined by concentrations of bubbles, have axes sub-parallel to lineation, and axial planes transverse to flow, sub-parallel to foliation, all of which could be consistent with healed longitudinal fractures that subsequently folded in longitudinal compression.

The degree of horizontal compression and thus of horizontal shear and vertical elevation, will depend on the amount of basal slip in front of the BTT. We used Hanson's (1995) velocity field, which assumes no basal slip where the ice thickness is less than 30 m, and this enhances shear and elevation. The observations and models of Moore *et al.* (2011), however, lead us to expect limited resistance to basal shear in front of the BTT. Weakened shear implies less rotation than shown for the planar features illustrated in Fig. 11. Since the debris bands originate as bed-parallel features once entrained and emplaced by some shear near the base, it is the longitudinal shortening and vertical extension that play the major role in determining their orientation of increased up-glacier dip as they approach the surface. Consistent with findings from Moore et al. (2011), we note that the observed dip of the debris bands, increasing from the centre towards the margin of the glacier from about 30^{o} to about 70^{o}, implies significant northward increasing resistance to slip in the marginal ice and thus enhanced longitudinal shortening and vertical extension.

On the basis of measured surface velocities and geometry of the debris bands extended back into the ice, Jansson *et al.* (2000) estimated that the minimum time for the bands to be elevated to the surface was 14 years. This was based on surface velocities. The time would be much longer taking into account the decrease in velocity with depth. Moore *et al.* (2013) estimated a maximum time of 42 years for the debris bands to first appear in 1994, based on the tritium content of the ice in the debris bands, indicating formation by basal freeze-on since 1952, the year in which the first thermo-nuclear weapons testing began and tritium levels began to spike. Times associated with our model are much longer, up to 160 years, reflecting both lower surface velocities and no slip at the bed, which Moore *et al.* (2010) show is not the case. We thus place little significance on the times in our model.

In addition to playing a key role in the emplacement of the debris bands, high fluid pressure may also induce cross-cutting basal tensile fractures, which can subsequently fill with water and refreeze; water may carry varying amounts of sediment into the fractures, ranging in grain size from silt to sand

(Fig. 12). We note that video surveys in boreholes undertaken by Hooke & Pohjola (1994) in 1989 and 1990 exhibited sediment held in suspension due to upwelling of water from the bed, and in some cases water containing fine sediment overflowing at the surface. Moore *et al.* (2013) also intersected pressurized englacial conduits when mapping the subsurface extent of the debris ridges with a steam drill, which transported turbid water with silt and fine sand to the glacier surface. We suggest that basal tensile fracturing accounts for the high-angle, sandy and silty bands that cross-cut the overall planar fabric associated with the debris bands. Such fractures have been described elsewhere (e.g. Ensminger *et al.* 2001). They may not always fill with sediment, in which case they may appear as clear ice with a central line of bubbles (Hambrey & Müller 1978; Lawson *et al.* 1994; Voigt *et al.* 2003), typical of healed fractures elsewhere on the glacier. We argue that all planar features in the ice in the vicinity of the debris bands originated in this way as tensile fractures that subsequently healed and moved with the ice. While tensile fractures could also result in injection of diamicton, the distribution and nature of the sediment within the bands make this extremely unlikely.

Tensile fractures that develop in basal ice will do so at orientations that depend on the local principal stresses. After healing and with or without inclusion of injected debris, they will be transported to the surface as compressive flow and vertical uplift occur, accompanying ablation. Modelling shows that fractures that develop near the BTT, at a wide range of initial orientations, will eventually outcrop as traces with curved surfaces dipping up-glacier at angles approaching the particle paths (Fig. 11B), which will also be subparallel to foliation. This is consistent with field observations. The initial orientation of fractures may vary widely as the stress field responds to changes in compressive stress induced by up-glacier velocity fluctuations as well as bedrock topography (Hanson *et al.* 1998). The thin fracture traces at a high angle to the foliation and the debris bands (as in Fig. 5) must have formed closer to the glacier front and, therefore, are less affected by the shear that would otherwise rotate them towards parallelism with the foliation.

Conclusions

The debris bands that underlie the debris ridges present on the surface of Storglaciären and the ice just below them lack the characteristic elements typical of discrete shear zones in rocks and ice. Field observations, isotope data and grain size analysis confirm previous interpretations that these bands were likely entrained at the base of the glacier due to the freezing of meltwater within basal till, and then detached from the unfrozen bed by high water pressure lifting the base and then subsequent freezing of water below the frozen till. They were subsequently elevated englacially due to compressive flow at the glacier front. Unlike those described earlier (Jansson *et al.* 2000; Glasser *et al.* 2003; Moore *et al.* 2013), the bands exposed in 2016 – 2019 were laterally discontinuous, indicating a patchiness to their initial entrainment, likely due to undulations in bed topography. In accord with Moore *et al.* (2013), we found no evidence for discrete displacement across debris bands that would indicate thrusting, although we suggest that some localization of shear in the ice near the base of a band after initial emplacement was necessary in order to carry the band over older ice that now appears to lie beneath it.

We found no evidence of overturned stratigraphy or of large-scale recumbent folding as proposed by Glasser *et al.* (2003) and Moore *et al.* (2013) and doubt the existence of such a fold or folds. The otherwise fairly uniformly up-glacier dipping fabric in the ice above and below the bands is disrupted by thin high-angle bands of ice containing fine sediment or trains of bubbles, which we believe originated as tensile fractures at the base of the glacier during intervals of high water pressure. These features cross-cut the debris bands and older planar features.

This multi-proxy approach to studying debris ridges on Storglaciären provides the basis for future investigations of similar features on other glaciers where there is conflicting evidence to support thrusting. Modelling and microstructural analyses in particular should be expanded. As indicated by field measurements and modelling, upward trajectories occur naturally in regions of glaciers experiencing longitudinal compression. More sophisticated modelling accounting for variable basal conditions near the BTT with time would be informative, especially with attention given to fracture development.

Microstructural analysis can be particularly valuable in characterizing ice deformation and understanding kinematics. It would be helpful to obtain microstructural data on glaciers, such as the Variegated Glacier, where thrusting is unequivocal, to use as a comparison for situations like those on the surface of Storglaciären. Our hypothesis is that if a localized ductile shear zone is present, it would be active close to the base of the glacier where flow is non-coaxial, but would be unlikely to survive as an active feature as the surface is approached and flow becomes more coaxial. Microstructural analyses collected from Storglaciären for this study primarily reflected near-surface conditions. We should therefore strive to conduct microstructural analyses on extracted ice cores that intersect debris bands close to the base. Englacial debris bands that can be detected using GPR surveys, steam drills or visually confirmed in ice tunnels, could be particularly useful as targets for analysis.

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Data Availability. — Data in support of the findings in this study are available online at: https://doi.org/10.6084/m9.figshare.14415458.v1 (Monz *et al.* 2021) or from the corresponding author upon request.

Author Contributions. — This paper is part of the Ph.D. thesis of M. Monz, who conducted fieldwork in 2016 and 2018, completed all structural analyses (macro- and micro-), synthesized dGPS data, made all figures and wrote the first draft of the manuscript. P. Hudleston, as advisor for Monz, provided guidance and assistance with project design, data collection, interpretation and writing. T. Zimmerman assisted with fieldwork and microstructural analyses using the Rigsby universal stage. S. Cook completed the sedimentological analyses and helped with interpretation of sedimentology and isotopic results. M. Leng performed the isotopic analyses. All authors participated in editing the manuscript and approved it for submission.

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Figure Captions:

Fig. 1. Schematic diagram illustrating the concept of thrust faulting and debris transport from the base of the glacier to the surface in the terminus, and highlighting the velocity gradients, from the base to the surface and from the margins to the centre, that exist due to drag of the ice against the base and sides. The schematic also emphasizes the previously documented, laterally quasi-continuous nature of these debris-bearing features at the surface.

Fig. 2. A. Simplified structural map of Storglaciären, highlighting the area where debris bands are located, and showing trends of major structural elements and orientations of foliation, bedding and lineation near the terminus where the debris bands are exposed. The orientation diagrams are in geographic coordinates. B. Inset focused on the debris ridge outcrops visible in 2018. Locations of samples SG11-B, SG20, and SG21, the three samples discussed in this paper, are indicated.

Fig. 3. Images of the north front of Storglaciären in the summers of 2016 and 2018. Numbers are used for identifying and comparing the surface expression of debris ridges, and emphasize that ridge 7 disappeared and ridge 1 appeared between 2016 and 2018. Note that the surface expression of the debris ridges is approximately parallel to what we interpret to be healed fracture traces.

Fig. 4. Annotated image in 2018 of debris band segments 1 and 2 (Fig. 3) (view north) displaying the initial position of one set of velocity stakes, the traces of major structural features, and showing the sample locations of SG20 and SG21 (collected in 2018), and the approximate location on a ridge of SG11-B from the adjacent ridge (collected in 2016). Fracture traces in this image appear irregular at the surface due to ablation and debris cover. Fractures traces, however, are smoothly curving nearly planar features.

Fig. 5. Cross section (view south) in 2018 of one of the debris ridge segments that was cleared of debris and then planed off using a chain saw. The debris cleared planar surface represents the base of the debris band responsible for the underlying ridge. All major fabric elements are visible. This includes foliation and late stage healed fractures. One such fracture (coloured brown) is decorated with sediment and one (coloured red) is defined by a train of fine bubbles, in each case encompassed by bands of clear ice and cross-cutting the overall fabric.

Fig. 6. Debris ridge segment (view south) in 2016, of the southern segment of ridge 3 (Fig. 3) cleared of debris, with major structural elements labelled. Image shows both, on the right (coloured brown), a steeply dipping sediment decorated fracture cross-cutting foliation, and on the left (shaded yellow) an englacial debris band approximately parallel with foliation, exposed either side of a melt gulley. The debris band is traced out of the ridge where it has both a component of debris bearing ice and debris that has melted out, beginning the formation of an ice-cored debris ridge that barely extends south from this location. The debris band sub-parallel with foliation contains mm scale lenses of clear ice and debris ranging in size from sand to boulder.

Fig. 7. dGPS results from August and September 2018 and August 2019. A. Map view of outlines of the two measured ridges (ridges 1 and 2 from Fig. 3) traced from August and September 2018, and stake

positions from August and September 2018 and August 2019. Ridge segment outlines were not measured in 2019, because although the stakes were still in place, the ridges themselves had deflated and were hard to discern from the surrounding ice. B and C. Positional measurements from each stake and calculated surface amount of movement in metres over the course of one month and one year for each of the two measured ridge segments.

Fig. 8. Particle size distributions for sampled ice facies: (A) debris bands; (B) frozen basal till; (C) basal ice (where light red lines indicate sediment-rich facies, and dark red lines indicate sediment-poor facies). Both (A) and (B) show pronounced peak at 4 Φ and a tailing off towards the finest grain sizes.

Fig. 9. Stable isotope composition of meteoric source (glacier ice and snow) and debris rich ice types. A least squares linear regression plotted through glacier snow and ice compositions gives a local meteoric water line and a least squares linear regression gives a shallower slope for ice in debris bands and frozen basal till.

Fig. 10. Section images from (A) SG20 (EBSD orientations), (B) SG11-B (Rigsby stage, cross polarized light) and (C) SG21 (EBSD orientations) and associated c-axis pole figures and eigenvalues for entire samples. The red line on the pole figures represents the foliation, the red square its pole, and the blue circle is the location of the maximum eigenvector. Pole figures for SG20 and SG21 were plotted using MATLAB, and that for SG11-B using OSX stereonet and contoured using the Kamb method. Pole figures are in geographic coordinates where x is east or the flow direction, y is north, and z is vertical.

Fig. 11. A. Frontal surface nodes (each at the top of a column of nodes) from Hanson's (1995) model, with those used for the two-dimensional cross-section across the debris ridge area labelled. B. Vertical cross-section showing columns of nodes and associated velocity vectors, and flow lines (particle paths for steady state). C. Frontal part of the cross-section (highlighted in B) showing the positions of two initially

planar features at two time intervals. D. The same frontal area highlighting the movement of a feature originating sub parallel to the bed of the glacier at two time intervals.

Fig. 12. Schematic representation of debris entrainment into the base of the glacier based on variability in bedrock topography, and subsequent transport and elevation followed by differential ablation causing the protruding surface expression.