Multiple subglacial till deposition: a modern exemplar for Quaternary palaeoglaciology

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Abstract

The sedimentology of a vertical succession of alternating beds of massive and fissile diamictons on a Þorisjökull plateau icefield outlet foreland is employed to assess the evolution of subglacial traction tills at the margins of active temperate glaciers with deformable substrates. Lodged boulders display strong A-axes and surface striae alignments which parallel surface flutings, indicating that fluting construction and till emplacement was related to moulding by consistent glacier flow from the SSW during the Little Ice Age. In contrast, clast macrofabrics at the sub-boulder size, not unlike those reported from other Icelandic tills, are not as strong as would be expected in a subglacially sheared medium, indicating shear strains too low for a steady state strain signature. This separation of fabric data has isolated the strain signatures of the lodgement and deformation components of subglacial traction till, whereby the orientations of the largest, lodged clasts record high cumulative shear strains and those of the sub-boulder sized clasts record greater susceptibility to deformation of their enclosing matrix. This is likely due to the effect of clast collisions in clast rich till and the perturbations set up by the numerous large boulders, consistent with observations on till fabrics in flutings and around lodged clasts. A/B plane macrofabric data display unusually high degrees of isotropy, reflective of the more variable orientations of A/B planes, which are thought to reflect A/B plane susceptibility to dip parallel or anastomosing shear planes. A wide range of clast angularity values reflects the localized input of freshly plucked and hence relatively highly angular blocks to the deforming layer, a characteristic of stepped bedrock profiles beneath the snouts of mountain glaciers. Finally, we hypothesize that the massive and fissile units may represent A and B horizons of subglacial deforming layer couplets, and that each couplet could record seasonal emplacement and partial inter-couplet modification near the ice margin. If the latter is the case, then less than 1 m of subglacial till is advected to the glacier margin per (annual?) deformation event.

Introduction

Although thick and complex sequences of multiple subglacial tills have been reported from ancient depositional settings based upon interpretations of the sedimentology of predominantly massive diamicton stacks (e.g. Eyles et al. 1982; Hicock 1992; Evans 1994, 2000a; Hicock & Fuller 1995; Larsen & Piotrowski 2003; Larsen et al. 2004; Piotrowski et al. 2004; 2006), they are relatively rare in modern glacierized catchments. Therefore, where modern till stacks can be identified they are critical to advancing our understanding of how multiple till sequences have been emplaced in the past, especially if sedimentological process and form can be linked. In the latter case, Icelandic glacier forelands contain tills that can be directly related to subglacial processes, as determined by the benchmark process measurements of Boulton and Hindmarsh (1987) and Boulton et al. (2001). For example, Benn (1995) sampled subglacial tills from the location of the Boulton and Hindmarsh (1987) experimental site at Breiðamerkurjökull after it had been exposed by glacier recession and was able to identify, based upon the diagnostic sedimentological characteristics of dense and fissile diamicton overlain by massive and friable diamicton, the A and B horizons proposed by the earlier process study. This further enabled Benn (1995) to assign clast macrofabric signatures as well as textural characteristics to the different styles of subglacial sediment deformation. Similarly, Evans and Hiemstra (2005) reported sedimentological and clast macrofabric data collected from sub-marginal till wedges that had been stacked to produce a complex push moraine during a period of glacier readvance in southern Iceland during the 1990s. Both studies have proposed potentially diagnostic sedimentological criteria for the identification of subglacial processes such as brittle and ductile deformation and lodgement. This is a knowledge base that needs to be expanded and substantiated in order to establish greater confidence in the interpretation of ancient glacigenic sediments. At the same time field based observations thought to be indicative of deformation style and cumulative shear strain need to be reconciled with apparently contradictory laboratory based measurements of shearing style and magnitude (e.g. Iverson et al. 1994, 1998, 2008; Hooyer & Iverson 2000; Iverson & Iverson 2001; Iverson & Hooyer 2002; Hooyer et al. 2008).

A thick and complex sequence of fine-grained diamictons exposed on the recently deglaciated foreland of an outlet lobe of the Þorisjökull plateau icefield in west-central Iceland (Fig. 1) provides an ideal opportunity to assess the sedimentology of a potential multiple subglacial till stack in a situation where form can be reasonably confidently related to process (cf. Benn 1995). The sedimentary exposure has been revealed by the fluvial incision of a small gorge through a fluted till surface characterized by numerous lodged, bullet-shaped and striated clasts (Fig. 2). Therefore the vertical sequence of compact and alternately massive and fissile diamictons exposed in the walls of the gorge potentially represent subglacial till deposition by the emplacement of multiple glacier sub-marginal deforming layers.

Methods

Sedimentological investigations were principally focused on two main sections within the natural gorge exposure (sites A and B on Figure 2). Individual lithofacies can be traced along the gorge but are described in detail in two vertical sediment logs, which were compiled based on the identification of separate lithofacies according to bedding, texture, lithology and sedimentary structures. Lithofacies are described and classified according to the modified scheme of Eyles *et al.*, (1983) proposed by Evans and Benn (2004).

Clast macrofabrics and forms were measured based upon 50 clasts per sample where possible, although a minimum of 30 clasts were sampled in sedimentary units where clasts were more sparsely distributed and to ensure that data collection was confined to small areas and thereby reflected local variability in till properties (cf. Evans & Hiemstra 2005). Macrofabric measurements of the dip and azimuth (orientation) of the A-axis and A/B plane of clasts were taken using a compass clinometer, aiming to use predominantly clasts in the range of 30-125 mm (A-axis length) to allow comparison with other studies (Benn 1994a, b, 1995; Evans 2000b; Evans & Hiemstra 2005; Evans et al. 2007). The A-axes and A/B planes of clasts will tend to rotate to parallelism with the direction of shear in a shearing Coulomb plastic medium like till (cf. March 1932; Ildefonse & Mancktelow 1993; Hooyer & Iverson 2000). Evans et al. (2007) proposed that within thin subglacial shear zones, A/B planes will adopt up-ice flow-parallel dips more readily than A-axes, which can align transverse to flow and therefore display bi-modal orientation statistics. Fabric data were plotted on spherical Gaussian weighted, contoured lower hemisphere stereonets, using Rockware™ software. Statistical analysis was also undertaken using eigenvalues (S_1 - S_3), based on the degree of clustering around three orthogonal vectors (V_1 - V_3), presented in fabric shape ternary diagrams (Benn, 1994b). This identifies end members as being predominantly isotropic fabrics (S_1-S_2~S_3), girdle fabrics (S_1-S_2>>S_3) or cluster fabrics (S_l>>S_2~S_3). To further ascertain strain histories, fabric data has been classified according to five modal groups (un - unimodal, su - spread unimodal, bi- bimodal, sb - spread bimodal and mm - multimodal) and plotted against isotropy (S_3/S_1) in a modality-isotropy template, based on the modification of Hicock et als. (1996) modality-isotropy plot (Evans et al. 2007). The collection of macrofabrics based on A-axes as well as A/B planes further allows an independent assessment of both forms of clast fabric measurement, thereby addressing issues raised by previous studies that have promoted more regular use of both approaches (cf. Benn 1995, 2004a; Li et al. 2006; Evans et al. 2007).

In order to assess the operation and impacts of debris transport pathways contributing to the subglacial sediment (Boulton 1978), clast form was assessed on samples from site A only by measuring the three principal axes (A, B and C) of massive basalt clasts and the results plotted in ternary diagrams, based on the C:A axial ratio (blockiness) and B:A axial ratio (elongation). This facilitated the calculation of the C_{40} index (the percentage of clasts with a C:A axis ratio of <0.4), which determines the relative proportion of slabby to blocky clasts within a sample (Benn & Ballantyne 1993). The roundness of clasts was classified according to Powers (1953), and was used to calculate the RA summary index (percentage of angular and very angular clasts within a sample; Benn & Ballantyne 1993) and the mean roundness (cf. Spedding & Evans 2002; Evans 2010). These data are compared to available datasets on different glacigenic materials through the use of co-variance plots (Benn & Ballantyne 1994; Evans 2010), specifically the "Type 1" co-variance plot proposed by Lukas et al. (2013) for Icelandic glacigenic deposits in order to account for the low anisotropy basalt clast lithologies and ice cap outlet glacier setting. Finally the morphological characteristics of clasts indicative of subglacial transport, including striae, facets and stoss/lee forms, were noted (cf. Sharp 1982; Krüger 1984; Benn 2004b) and presented as percentages for each sample.

Bulk samples of 200 gm were taken from each lithofacies at site A only and were dry sieved to separate the <2mm fraction from the sample before being passed through a laser granulometer. This generated a grain size distribution histrogram, from which the percentage of clay, silt, sand and gravel in each sample was calculated in order to assess inter- and intra-sample variability.

The micromorphology of the sediments was analyzed based on 7 thin section samples from lithofacies 1-7. The micromorphology was assessed qualitatively and semi-quantitatively using thin sections of c. 55 x 75 mm in size, from which the sediment characteristics were described according to standard terminology (cf. van der Meer, 1993; Menzies, 2000; Carr, 2004).

Sedimentology of the Þorisjökull diamicton sequence

The foreland of the Þorisjökull case study outlet glacier contains a 30 m wide corridor of fluvially scoured diamicton which has been winnowed and cleared of loose surface clasts and morainic debris and appears to have resisted erosion because of its indurated nature (Figs. 2 & 3). After initially winnowing the diamicton, the glacial meltwater stream became confined to a narrow gorge which has been incised through a thick sequence of alternately massive and fissile diamictons containing numerous bullet-shaped, facetted and striated clasts which in places are organized into weakly developed intra-diamicton clast lines (Fig. 3). Extensive exposures are available through the diamictons and two representative sites were chosen for intensive logging and sampling, one in the downstream (Site A) and one in the upstream (Site B) portions of the gorge. The local flutings on the foreland are aligned 030° - 210°, recording former subglacial streamlining by glacier ice flowing from the south-southwest. Macrofabrics on the clasts protruding from the 30 m wide fluvially scoured till surface along the entire length of the gorge (Fig. 4a) reveal a strong NNE-SSW alignment, especially on A-axes (S_1 eigenvalue = 0.813), corresponding to the surface fluting orientations; an additional transverse orientation is apparent in the A/B plane data (S_1 = 0.561). A similar pattern is apparent in the data collected on clasts exposed along the entire length of the gorge section (Fig. 4b) and clast surface striae are also strongly aligned NNE-SSW (Fig. 4c).

Site A

Site A is a 1.95 m high section comprising grey-brown, massive, matrix supported diamicton overlying 0.1m of basal fine gravels and sands (Fig. 5). Eight separate lithofacies (LFs) can be distinguished within the section, seven of which consist of diamicton of 0.09 m – 0.45 m thick, primarily based upon the degree of fissility, which varies with depth. Lithofacies 2, 4, 6 and 8 are structurally massive to weakly fissile and indurated, with widely spaced partings at depth, although LF 8 displays a porous and open framework appearance. In contrast, LFs 3, 5 and 7 are densely fissile with closely spaced and anastomosing partings that commonly display slickensided or polished surfaces. Contacts between the lithofacies are generally conformable, but also sharp at the bases of LFs 6 and 7. The vertical pattern is thereby a repetition of alternate massive and fissile lithofacies. The basal massive to weakly fissile diamicton, LF 2, directly overlies gravel, sand and fine grained interbeds that display vertical clastic dykes and are locally deformed into open folds.

Clast A-axis macrofabrics (Figs. 5b & 5e) vary from weak to moderate clusters (S_1 = 0.44 - 0.63) and reveal consistently low mean dip angles (<12.4°). In general, clast orientations are aligned parallel to the local SSW-NNE ice-flow, as deduced from surface flutings, and display low isotropy indexes (S_3/S_1 = 0.1 - 0.4). Nonetheless, some samples display ice flow-transverse components (e.g. A6, Fig. 5b), a previously well documented characteristic of A-axis macrofabrics that has been interpreted as a product of the tendency for elongate clasts to roll in deforming or shearing media (cf. Jefferey 1922; Lindsay 1970; Hart et al. 2009), thereby reducing cluster strength and tending towards girdle-type fabric patterns. Although the highest S_1 eigenvalues occur in densely fissile diamicton (LF7 = 0.60 & 0.63 with a range of 0.44 –

0.63), the A-axis macrofabrics of massive diamictons are not significantly less well clustered (S_1 range = 0.47 - 0.59).

Clast A/B plane macrofabrics (Figs. 5b & e) are consistently weaker and show less clustering than A-axis data, with generally higher isotropy indexes (S_3/S_1 = 0.34 – 0.57) and slightly lower S_1 values (0.40 – 0.52), although the differences are not statistically significant. Mean dip angles are also significantly higher (13.9° - 42.6°) than those for A-axes. Visually A/B plane stereonets therefore display more girdle-like to isotropic fabric shapes than their A-axis counterparts, and orientation comparisons between techniques vary from very similar (e.g. A8) to almost incompatible (e.g. A4). The A/B plane macrofabrics of massive diamictons are not significantly less well clustered (S_1 range = 0.40 - 0.52) than those of densely fissile diamictons (S_1 range = 0.44 - 0.48).

Clast form varies significantly between samples A1 – A10, especially with respect to RA and average roundness (Fig. 5c & d). Clasts are predominantly blocky and sub-angular, as reflected in mean roundness values of 1.3 - 2.3, RA values of 10 – 80.1%, and C_{40} values of 6.7 – 33.3%. The narrow range of C_{40} (high blockiness and hence distinctly subglacial) and wide range of RA values is reflected in co-variance plots (Fig. 5f), a characteristic that has been recognized in Icelandic tills on the forelands of plateau icefield outlet glacier lobes (Evans 2010) but one that is not evident in the Type 1 co-variance plot for low anisotropy lithologies (Lukas et al. 2013). Within the thicker lithofacies 5 and 8, there is a tendency for RA values to increase markedly and average roundness to decrease with depth, although a minor reversal of this trend is apparent in LF7, another thick unit (Fig. 5c). In the thinner lithofacies 3, 4 and 6, the RA values distinctly increase and average roundness values decrease relative to lithofacies above and below them (Fig. 5c).

Finally, sediment grain size analysis displays a predominant composition of silt and sand within the diamictons (65-85% per sample), which can be compared to the grain size of the underlying stratified sand and gravel substrate as represented by sample A11 from LF 8 and indicating a silt and sand component of ≤65% (Fig. 5b). At the base of the sequence, LF2 (sample A10), when compared to the underlying sand and gravel substrate of LF 1 (sample A11), displays the abrupt fining in grain size indicative of the direct emplacement of glacigenic diamictons over stratified deposits. Above this, the intra-unit grain size patterns reveal vertical changes between massive and fissile diamictons. Between LF 3 (fissile) and LF 4 (massive), samples A9 and A8 reveal a vertical change to a finer matrix due specifically to an increase in the silt and sand components at the expense of gravel. The gravel component increases again in samples A7 and 6 from LF 5 (fissile), and further increases in sample A5 in the overlying LF 6 (massive). Gravel content again decreases in LF 7 (fissile) in favour of increasing silt in samples A4 and A3. This contrasts with overlying LF 8 (weakly fissile to massive) where the silt component increases at the expense of sand and gravel content rises between samples A2 and A1. Overall these grain size data display a repeat set of vertically fining trends between each pair of fissile and massive diamicton lithofacies with the exception of LFs 5 and 6 which vertically coarsen.

Site B

Site B is located 30 m upstream of Site A and comprises a more extensive exposure, 20 m long and up to 3.2 m high (Figs. 2b & 6a) from which a composite vertical profile log was compiled (Fig. 6b). The sequence is characterized by grey-brown, massive to fissile, matrix-supported diamictons that contain relatively few small clasts with an average A-axis length of 30 mm. The diamictons are classified as lithofacies (LF) 2-9 based upon their structural characteristics, which like Site A results in the identification of alternate indurated massive and fissile beds, with thicknesses varying from 0.1m – 0.8m.

At thin section scale all the diamictons appear grey in colour and massive but there are subtle structural elements in some lithofacies. Skeletal clast components are typically A-SR in form and up to 30mm in diameter. Sample TS1 (LF9) is matrix-supported with VA-SR skeletal clasts up to 5mm in diameter. There are very few grain to grain contacts though some skeletal clasts have silt/clay coatings. There are no clear structures in TS1. Sample TS2 (LF8) is a grey, matrix- supported, diamict with A-SR skeletal clasts up to 20mm in diameter and few grain to grain contacts. In contrast to TS1 it is clearly cross cut by a series of low angle lineations/partings which anastomose. There is evidence of grain long axes alignments along these partings (Fig. 7). Samples TS3 (upper LF6) and TS4 (lower LF6) also contain subtle low angle lineations, with TS4 exhibiting cross cutting patterns in places. The lineations/partings are typified by subtle changes in groundmass birefringence and grain alignments. Again there are few grain to grain contacts within the skeletal component of the matrix. Sample TS5 (LF4) is massive with no distinctive structure, but sample TS6 (upper LF2) does have subtle low angle lineations partially highlighted by plasmic fabric and grain alignments. Sample TS7 (lower LF2) also exhibits lineation/partings but in places these have a cross cutting pattern similar to TS4 (Fig. 7).

Clast A axis macrofabrics (Fig. 6b & 6c) are predominantly moderately clustered (S_1 = 0.45 – 0.60) with low mean dip angles of <15.1° and isotropy indexes of 0.23 – 0.48. Mean lineation azimuths range between 354.4° to 50.6°, consistent with the former ice flow direction of SSW-NNE recorded by surface flutings; more significantly the uppermost diamicton (LF9; sample B1) is strongly aligned SSW-NNE. Samples B5 from LF5 and B6 from LF4 display two distinct alignments, with SW-NE and SSW-NNE trends in the two samples being accompanied by subordinate N-S and E-W trends respectively; these appear to be manifest in the A/B plane macrofabrics of the two samples in bi-modal contour clusters (Fig. 6b), the two fabric measurement styles thereby recording slight deviations from a dominant SSW-NNE alignment rather than the influence of transverse elements. The A-axis macrofabrics of massive diamictons are not significantly less well clustered (S_1 range = 0.47 - 0.46) than those of densely fissile diamictons (S_1 range = 0.45 - 0.60).

A/B plane fabrics are moderate to weak (S_1 values 0.40 – 0.52) with higher mean dip angles (25.3°) relative to A axes. While mean lineation azimuths are both parallel and transverse to ice flow direction for a number of samples, weak clustering is apparent on contoured stereonets and displays a broad SW-NE alignment (Fig. 6b). The weakest S_1 eigenvalues for A/B plane macrofabrics occur in massive diamictons (0.40 - 0.51), indicating a tendency for densely fissile diamictons to be slightly more clustered.

Secondary analysis of clast macrofabric data

Clast macrofabric strengths are further employed here in comparisons with previously reported macrofabric data for subglacial deposits of known origin. This involves first the plotting of samples from this study onto clast fabric shape ternary diagrams (Fig. 8a), which visually categorize samples according to their isotropy and elongation and contain envelopes of fabric shapes for lodged clasts, subglacial traction tills and glacitectonites (*sensu* Evans et al. 2006b) from both modern Icelandic settings as well as ancient glacigenic deposits. Laboratory experiments on the shearing of till-like materials have prompted Iverson et al, (2008) to plot the influence of initial consolidation and then increasing shear strain on clast fabric shapes on ternary diagrams; this is represented by the arrows on Figure 8a depicting the changing fabric shape with increasing shear strain magnitude, from isotropic to girdle to cluster. Secondly, the data is plotted onto the modality-isotropy graph (Fig. 8b), modified by Evans et al. (2007) from Hicock et al. (1996), on which envelopes for lodged clasts, subglacial traction till and glacitectonite are identified, and plotted positions can be used to infer the cumulative strain recorded by the fabric at the time of deposition.

There are two apparent misunderstandings surrounding the use of these types of plot and hence prior to our interpretations of the sediments we provide some reflective discussion. First, they are often regarded as an illustration that glacial geologists use clast macrofabrics to infer genesis of sediment or a classification of till facies (e.g. Bennett et al. 1999; Iverson et al. 2008); this is not reflected in recent developments in till sedimentology. Instead, till classifications are based primarily upon lithofacies analysis and although some reports on clast macrofabrics have used the language of genetic interpretation they also caution against the exclusive use of such data (e.g. Benn 1995; Hicock et al. 1996). Macrofabrics are secondary data and are plotted on fabric shape diagrams for comparison with previous studies on tills of known genesis, thereby providing us with envelopes of fabric data with which to develop critical discussions on the strain history of till deposits. These discussions are tempered by the expanding knowledge base on physical process measurement and experiment (see Evans et al. 2006b for a review). Second, the continued employment of ternary and modality-isotropy plots by glacial geologists are often viewed as a vehicle for re-enforcing the tenet that fabric strength decreases with increasing cumulative strain, butfabric strength plots such as those represented in Figure 8 actually repeatedly demonstrate quite the opposite. Some early literature on bed deformation did indeed propose that thicker deforming layers produced weaker macrofabrics, potentially due to what was then regarded as the behaviour of a viscous medium (e.g. Hicock 1992; Hart 1994, 1997; Benn 1995; Benn & Evans 1996) but this does not equate to a belief or folklore as charged by Clarke (2005) for the glacial sedimentology community. Some glacial geologists have indeed highlighted the tendency for macrofabrics to weaken in A horizon deforming layers and have speculated on the role of ductile intergranular shear, localized dilatancy or even sediment flowage as an explanation of this characteristic, potentially controlled by the spatial and temporal history of solid state deformation (e.g. Roberts and Hart 2005; Evans et al. 2006b). However, studies of till sedimentology have consistently acknowledged the role of increasing cumulative strain in the development of stronger clast fabrics (e.g. Hicock et al. 1996; Evans et al. 1998, 1999; Hiemstra & Rijsdijk 2003), informed by physical experimentation and hence physical laws. Indeed we employ the experimental database of Iverson et al. (2008) in our following interpretations.

In both types of plot in Figure 8, the strongest clustering represents Iverson et als. (2008) "steady state fabric", where S_1 eigenvalues > 0.78 develop at shear strains of 7-30. Although macrofabrics from deposits in the field cannot be related to a specific strain magnitude in the same way as experimentally induced fabrics such as those reported by Iverson et al. (2008), shear box experiments provide us with a clear indication of the relationships between shearing and fabric strength and increase our confidence in the employment of fabric shape plots in determining the shear strain history of tills. Because of our previously acknowledged uncertainties surrounding the strain magnitude in till deposits we have used the term "cumulative strain" when interpreting fabric strengths. We continue to use this term here, even though Iverson et al. (2008) have demonstrated that shear strain magnitude is the most likely variable to be closely correlated to fabric development, simply because we do not know the strain magnitudes or deformation histories of the deposits being studied. In continuing to use the term "cumulative strain" we here qualify that it does not equate to shear strain magnitude or strain rate but does reflect the fact that a till is likely to contain a strain signature produced through multiple and complex deformational events.

The clast fabric shapes (Fig. 8a) indicate that A-axis fabrics are generally more strongly clustered than A/B plane fabrics, although samples occupy the middle of both ternary plots, suggesting generally weak to moderate clustering. Most of the A-axis fabrics lie within the envelope representing previously reported upper or A horizon subglacial tills, with only two samples (A7 & A8) lying on the weaker margins of lower or B horizon tills; only the stronger of these (A7) is from a diamicton (LF6) with the fissility typical of B horizon tills. The A-axes of the lodged boulders along and around the sections occupy the more clustered end of the Breiðamerkurjökull lower till envelope and lie just outside the lodged clast envelope of Evans and Hiemstra (2005), whereas the A/B plane fabrics are significantly more girdle-like.

General comparisons of the clast fabric shape ternary plots indicate a tendency for A-axes to become more girdle-like as their cluster strengths diminish, whereas A/B planes trend towards increasing levels of isotropy. This is particularly apparent in the modality-isotropy plots (Fig. 8b), which demonstrate an almost exclusive multi-modal signature in the A/B plane data, with a number of highly isotropic samples plotting outside the envelopes of previously reported subglacial tills (Evans et al. 2007) despite the boulder samples plotting as relatively strong and spread-unimodal in character. In contrast, A-axis fabrics display less multi-modality, plot within the envelopes of previously reported tills and are characterized by more spread-bimodality. Although boulder A-axis fabrics are stronger than those for A/B planes and plot near Evans and Hiemstra's (2005) envelope for lodged clasts, none of the A-axis macrofabrics display the unimodality, and only one the spread-unimodality, typical of relatively highly strained subglacial tills (cf. Hicock et al. 1996). With respect to the consolidation/shear strain pathway proposed by Iverson et al. (2008), the Þorisjökull data are indicative of very low strains. Additionally, the A axis samples in particular could be interpreted as being representative of strain signatures that occupy the various positions along the girdle to cluster pathway following on from consolidation, as identified by Iverson et als. (2008) experiments on ancient tills. If we are to interpret the A/B plane strain signatures in the same way, they predominantly reflect very immature/very low strain fabric development; alternatively the A/B planes are just not recording strain signature in the same way as A axes.

Interpretation of the Þorisjökull diamictons

The sedimentological characteristics of the alternating beds of massive and fissile diamictons at Þorisjökull are indicative of subglacial traction till deposition (*sensu* Evans et al. 2006b). The structural and textural appearance of the massive diamictons, such as their weak fissility and induration as well as a localized porous and open framework (presumably dilated) nature, are similar to A horizon type or "upper" tills previously reported from Icelandic glacier beds (cf. Boulton & Hindmarsh 1987; Benn 1995; Evans 2000b; Evans & Twigg 2002). In contrast, the density and locally slickensided or polished, closely spaced and anastomosing partings that characterize the strongly fissile diamictons are typical of B horizon or "lower" tills identified in Icelandic settings. Hence from hereon we refer to the alternating diamictons as potential A and B horizon tills.

This classification of the till stack as a repetitive sequence of A and B subglacial deforming horizons is, however, over-simplistic for three reasons: 1) the thickness of the diamicton units and their internal structure is not uniform; 2) the depth of deformation beneath glaciers varies (e.g. Truffer et al. 2000; 7m beneath Black Rapids Glacier, Alaska) and hence shearing may be activated along decollement planes within previously deposited tills or even at the till-bedrock contact (e.g. Kjær et al. 2006); and 3) the locus of failure within single deforming layers may migrate in response to temporal variations in porewater pressure and hence the thickness of dilatant horizons may change even on a diurnal basis (e.g. Boulton et al. 2001). For these reasons the occurrence of fissility within A horizons in particular may reflect the collapse of a predominantly dilatant till, likely at the later stages of its development, or it could have been superimposed on the A horizon by the development of an overlying B horizon. Late stage brittle shearing of this nature has been proposed by Menzies (1990) as a likely origin for brecciated diamictons. Micromorphological analysis clearly shows that both fissile and massive diamicts at this site are characterised by low angle micro-shears with anastomosing patterns indicative of multiple brittle shear events. However, there is little evidence to support dewatering of these sediments, or that their matrices have undergone deformation through ductile intergranular rotation (e.g. lack of turbates). Hence, the development of these microshears is more likely to have been a product late phase overprinting of A horizons by the emplacement of the next B horizon unit. B units are widely reported to acquire fissility as a result of microshears developing in response to multiple, discrete shear failure events within a solid state, relatively unsaturated, deforming bed environment (Evans et al., 2006a).

In addition to locally displaying <1 m thick alternating units of massive and fissile diamicton, similar to the A and B horizons reported from modern Icelandic glacier beds and recently deglaciated forelands (Boulton & Hindmarsh 1987; Benn 1995; Evans 2000b; Evans & Hiemstra 2005), the diamictons at Þorisjökull possess diagnostic subglacial characteristics, such as numerous bullet-shaped, facetted and striated clasts (cf. Sharp 1982; Krüger 1984; Clark 1991; Hicock 1991) whose A-axes and surface striae in particular record fabric development aligned 030° - 210° and parallel to surface flutings at the site (cf. Boulton 1976; Krüger 1979; Clark & Hansel 1989; Benn 1994a; Benn 1995; Benn & Evans 1996). Fluting and at least upper till emplacement was therefore related to moulding by glacier flow from the south-southwest during the last phase of ice advance, which geomorphology and surficial geology mapping (Evans et al. 2006a) indicates occurred during the historical Little Ice Age. In the absence of any depositional breaks recorded by non-subglacial sediments, we assume that the tills were emplaced sequentially during the period when the plateau icefield outlet glacier developed into a piedmont lobe and occupied the Little Ice Age maximum limit. Hence the complete sequence of tills likely records part of a sub-marginal thickening stack (*sensu* Evans & Hiemstra 2005; cf. Boulton 1987, 1996a, b) deposited on a northeasterly sloping foreland at least partially cloaked in glacifluvial outwash, as represented by LF 1 in section A. Open folding and clastic dykes in the LF 1 stratified sediments are likely related to glacitectonic disturbance and hydrofracture infills (Rijsdijk et al. 1999; Le Heron & Etienne 2005) respectively, developed during initial glacier advance, prior to the emplacement of the first subglacial till in the stratigraphic sequence.

Clast macrofabrics in Icelandic till A and B horizons normally display distinctly different shapes, whereby B horizons tend towards stronger clustering but this is not particularly well replicated in this study, with most A-axis fabric shapes being typical of those for upper or A horizons, and A/B plane fabric shapes revealing only a slight tendency for greater clustering in B horizons (Fig. 8). Nevertheless, there is some evidence for higher S_1 eigenvalues for both A axes and A/B planes in the densely fissile or potential B horizon diamictons, indicative of relatively higher cumulative strain and/or more constrained, brittle shearing. Variations between potential A and B horizons can be visualized by identifying the matching clast fabrics of A and B horizon couplets in a fabric shape plot and linking them with hysteresis-type curves (Fig. 8d). Generally these curves reveal that there are vertical trends in clast fabric strength but these are not consistent. For example, LFs 5 & 6 display decreasing A axis fabric strengths from the potential lower B to the upper B to the A horizons in the downstream section but the reverse trend in the upstream section. In contrast, LFs 3 and 4 display decreasing A axis fabric strengths from potential lower B to upper B to A horizons in the upstream section but a reverse trend in the downstream section. The most densely sampled vertical sequence is in downstream section LFs 7 and 8 where a weakening in both the A axis and A/B plane fabrics is apparent from potential B to A horizons and an additional trend of strengthening also appears towards the upper parts of each horizon. These intra-couplet patterns reflect the tendency for the strengthening of fabrics in B horizons but this is not a particularly convincing trend.

The larger and clearly lodged clasts, which were grouped into two sample sets representing a single near surface horizon and a combined set from various lithofacies throughout section B, display the strongest fabric shapes and, together with their surface striae orientations, replicate the former glacier flow direction most closely. This indicates that lodgement not only creates the strongest clast alignments but also is most effective on larger particles, as depicted by the modality-isotropy plots (Fig. 8b). It follows therefore that smaller clasts have a tendency to move more freely within the finer-grained matrices of subglacially shearing tills even though deforming layer thicknesses are small and even within brittle sheared B horizons. This may reflect the partitioning of dilatancy-driven deformation into the weaker matrices of the till, with larger clasts behaving independently (Evans et al. 2006b). Hence Iverson et als. (2008) proposal that grains are locked in as a result of consolidation prior to shearing appears to be inapplicable here. A/B plane fabrics clearly trend from relatively highly isotropic to girdle-like, with the strongest alignments being visible in large lodged clasts (Fig. 8). The unusually high isotropy for A/B planes indicates that they must in some way be more susceptible to subtle changes in shearing-induced localized pressures on clasts and so do not get particularly strongly locked in to an up glacier dipping imbrication in the thin shearing zones of the B horizons, as proposed by Evans et al. (2007) to explain the normally stronger A/B planes. This most likely reflects the influence of subtle changes in dip orientations, which are typically more variable on A/B planes than on A axes and additionally could more faithfully replicate the three dimensionally more variable dips of the anastomosing failure planes in the narrow shear zones in thin subglacial deforming layers.

The thickness of individual diamicton lithofacies is predominantly modest, ranging from 0.09 m – 0.80 m, but when grouped as potential couplets of fissile and massive diamicton range from 0.2 – 0.9 m thick. If the diamicton couplets represent A and B horizon deforming layers, as their structural appearance suggests, then they record the advection of less than 1 m of subglacial till per deformation event, the sharp or conformable contacts between couplets recording vertical stacking of couplets and/or the partial erosion of A horizon tops by subsequent B horizon development. Evans & Hiemstra (2005) demonstrated that such sub-marginal advection events were annual in active temperate glacier lobes and that thinner till packages and/or incomplete couplets, together with lodged clast lines, represent the up-ice end of sub-marginal thickening wedges where partial erosion of till packages take place and larger clasts become increasingly more lodged and striated. This model of till emplacement implies that A and B horizon couplets could reflect a seasonal sub-marginal depositional signal reflective of changing porewater pressure regimes.

The grain size analysis of the lithofacies stack at section A reveals that the till matrices are predominantly silty sand and that this represents the import of finer materials to the site by subglacial deformation when compared to the grain size of the basal stratified sand and gravel (LF 1, outwash). This trend is interrupted by the import of coarser materials in LF 4, although neither section contained any evidence of meltwater deposits between or within tills that could represent ice-bed separation events (canal fills) and so the origins of coarsening matrices are unclear; it is possible that up-ice patches of the basal outwash sand and gravel could have been cannibalized in the erosional zone and reworked into the subglacial deforming layer as it thickened towards the snout. This is reflected in the vertical coarsening between LFs 5 and 6 (fissile and massive diamictons respectively). This vertical coarsening between couplets is however the exception, as fining between other fissile and massive diamictons is more prevalent. The cause of this fining is unknown but could relate to the upward mobility of finer grain sizes during periods of high porewater pressures and the development of vertical water escape pathways, for example when A horizons lose their matrix framework (Evans et al. 2006b).

Although the blocky shapes and sub-angular clast forms of the tills at this locality are normally regarded as typical of mature subglacially modified materials, the co-variance plots (Fig. 5f) are unlike those from other subglacial samples, specifically because of a wide range of angularity values. This style of co-variance plot likely reflects the localized inputs of freshly plucked and hence relatively highly angular blocks to the deforming layer, a characteristic of stepped bedrock profiles beneath the snouts of mountain glaciers; bedrock steps beneath the nearby Þorisjökull outlet glacier lobe are evidenced by areas of extensive transverse crevassing. Evans et al. (1998) have previously reported that such plucking can take place even with the operation of a deforming layer, due to the injection of fine grained matrix into bedrock fractures and the concomitant elevation of shear stress in the fracture. The potential influence of clast shape on macrofabric development (i.e. elongate and slabby clasts should attain high A axis and A/B alignments respectively in deforming media) is not especially evident in the data, although there is a tendency for S_1 eigenvalues to increase with A axis C_{40} values (R^2 = 0.12) but to show no trend with A/B plane C_{40} values (R^2 = 0.03). Clearly, in order to deliver a more meaningful test of clast shape control on macrofabric strengths, a greater range of C_{40} values need to be displayed in the sample data.

Discussion: wider implications of the Þorisjökull till sequence

The architecture of subglacial deforming tills and associated sediments has been elucidated by empirical based theory (e.g. Boulton 1987, 1996a, b), process-form based observations (e.g. Evans & Hiemstra 2005) and numerical modelling (e.g. Leysinger-Vieli & Gudmundsson 2010), highlighting the importance of sub-marginal to marginal thickening of till to produce down-ice thickening wedges and linking subglacial deforming layers to moraine construction. The product of incremental advection and stacking of subglacial tills during multiple events and at a relatively stable glacier snout should therefore comprise complex vertical sequences of partial and/or complete individual deforming layers in the proximal core of complex push moraines (Evans & Hiemstra 2005; Fig. 9). Each deforming layer may develop and display A and B horizon couplets if the subglacial till deformation processes observed at Icelandic glacier snouts are operative. These till stacks represent the sub-marginal depositional zone created by sediment flux from the erosional zone located in the area of bed overdeepening below the equilibrium line (Boulton 1996a, b). An advancing or receding active temperate glacier snout, such as those that drain the ice caps of Iceland, will tend to produce only one sub-marginal till wedge and moraine per year (Price 1970; Sharp 1984; Boulton 1986), the moraine being produced by the advection of subglacial deforming sediment from erosional zones typically located ≤ 400 m (based on Breiðamerkurjökull; Boulton 1987) from the ice margin. Moraines that become overridden during periods of glacier advance can be recognizable in the landform record as subdued arcuate transverse ridges with fluted surfaces and cores of multiple subglacial tills (cf. Krüger 1994; Rose et al. 1997; Evans & Twigg 2002; Evans & Orton 2014; Evans et al. 2009), hence explaining how such complex till stacks can be found in former subglacial settings. The migration of the boundary between the erosional and depositional zone over time will result in the excavation/cannibalization of pre-existing tills during ice advance or till superimposition during recession. Features typical of this erosional and depositional overprinting are clast lines and incomplete deforming layer couplets (Hicock 1991; Boulton 1996a, b; Boyce & Eyles 2000; Evans & Twigg 2002; Evans & Hiemstra 2005). The location of the Þorisjökull till stack within 400 m of the Little Ice Age limit, together with the consistent orientations of surface flutings and clast fabrics throughout the sequence, strongly suggest that it represents a sub-marginal accretionary wedge of subglacial traction tills displaying multiple till emplacement events, potentially of annual scale (Fig. 9). Weakly developed boulder lines, in places associated with major partings or shear planes, are explained by Boulton (1996a, b) as the location of the A/B horizon interface during periods of localized erosion when the interface descends into the B horizon, but by Hicock (1991) as hiatuses between the emplacement of till units. As the clast lines at Þorisjökull are only weakly developed and do not occur at clear A-B horizon or till unit boundaries they most likely represent shear planes within deforming till developed during brittle shear in B horizons and during late stage compaction/collapse or brittle shear overprinting of formerly dilated A horizons (Krüger 1984; Benn & Evans 1996; Eyles & Boyce 1998; Fig. 9). The creation of an accretionary wedge versus a clast line or pavement has been related by Evans and Hiemstra (2005) to the position of the depo-centre relative to the glacier margin, whereby thicker ice results in preferential removal of finer grained matrices (creating the thin end of the accretionary wedge) and thinner snout ice results in reduced overburden pressure/bed shearing (creating the push moraine). Annual thinning and recession of the snout results in a reduction of overburden pressures and hence driving stress at a single location, so that subsequent sub-marginal deforming layers, as well as their component A and B horizons (particularly the A horizons), are increasingly better preserved in vertical sequence (Fig. 9).

The relatively weak S_1 eigenvalues (A axes = 0.44 - 0.62; A/B planes = 0.39 – 0.52), hence higher isotropy, of the Þorisjökull tills are slightly weaker but not unlike those previously reported from Icelandic subglacial deposits (e.g. A axes = 0.51 - 0.74, A/B planes 0.46 – 0.67; cf. Sharp 1984; Dowdeswell & Sharp 1986; Benn 1995; Benn & Evans 1996; Evans 2000b; Evans & Twigg 2002; Evans & Hiemstra 2005). Such fabric strengths are difficult to reconcile with the shear strain history that has been proposed for subglacial processes (Boulton & Hindmarsh 1987). More specifically, in order for a glacier to move mostly by bed deformation, Iverson et al. (2008) point out that shear strains are likely to be in excess of 100 and their laboratory experiments on till-like materials have demonstrated that strong fabrics (S_1 >0.78) are developed at lower strains of only 7-30 in response to shearing (Hooyer & Iverson 2000; Thomason & Iverson 2006; Hooyer et al. 2008; Iverson et al. 2008). These “steady-state fabrics” for sheared till are relatively strong compared to the S_1 eigenvalues reported in studies of ancient till deposits, which regularly report strong (e.g. 0.65 - 0.97, Larsen & Piotrowski 2003) but also a wide range of A axis macrofabric strengths (e.g. 0.47 – 0.95, Hicock et al. 1996; 0.44 – 0.83, Gentoso et al. 2012). The Þorisjökull till fabric data are therefore reflective of strains too low to represent a steady state strain signature and are at the weaker end of the strength range reported from ancient tills, despite displaying the sedimentological and structural characteristics of subglacial traction till emplacement and having been produced in a subglacial depositional environment.

The range of S_1 eigenvalues from Icelandic till A axis fabrics (0.44-0.74) indicates that the subglacial deforming layer at the position it is being sampled (i.e. sub-marginally) has predominantly not been subject to high levels of shear strain, at least as far as we can deduce at the macrofabric scale. This latter point is significant bearing in mind that a Coulomb-plastic medium such as till, unless it is dilated, will fail along narrow shear zones developed in its finer grained matrix, thereby imparting fissility and slickensided partings. Even in a dilated state the zone of failure may be narrowly confined and can migrate with changing porewater pressures; hence Iverson et als. (2008) explanation of weak flow-parallel fabrics as an outcome of strain that was focused in a zone too thin to be identified by the sampling density. Because the sampling area for clast macrofabrics can bridge such zones, not all the clast alignments measured will reflect the strain signature. Nevertheless, the macrofabric strengths of clasts with lodged characteristics are very strong both within till units (A axes S_1 = 0.81, A/B planes S_1 = 0.56) and between them (A axes S_1 = 0.77, A/B planes S_1 = 0.59). This reveals not only the consistent flow direction of glacier ice during the deposition of the till stack, not unexpected in a plateau outlet valley, but also that the tills can be classified as highly strained if we consider only the larger boulder-sized clasts as the passive strain markers. Additionally, the identification of lodgement entirely by sedimentary characteristics and independent of clast orientation measurements (cf. Evans & Hiemstra 2005) allows the macrofabric signature of that process to be isolated; hence our macrofabric data reflect the lodgement and deformation components of subglacial traction till (*sensu* Evans et al. 2006b) but are not used to identify depositional facies to a specific process level (i.e. lodgement till, deformation till, melt-out), a procedure identified as inappropriate by Evans et al. (2006b) and Benn and Evans (2010) for a variety of sedimentological reasons.

The influence of clast size on fabric development has been analysed and demonstrated as potentially significant by Kjær and Krüger (1998) and Carr and Rose (2003). At larger scales, the stronger macrofabrics of boulder sized clasts compared to smaller particles has been identified in previous studies of ancient tills (e.g. Evans & Hiemstra 2005) but laboratory experiments have not indicated any significant influence of grain size over fabric strength (e.g. Iverson et al. 2008), at least at diameters up to 8 mm. Boulders, however, are significant obstacles in thin deforming layers, and macroscale strain markers (e.g. boudins, faults and folds; e.g. Krüger 1979, 1984; Benn & Evans 1996) and microstructures (e.g. van der Meer 1993; Carr & Rose 2003) clearly show that such obstacles, once lodged, perturb the deforming matrix and its smaller clasts, often with leeside pressure shadows being created on their down flow side (Evans et al., 1995; Roberts and Hart, 2005; Fig. 9); striated surfaces on the boulders also record the passage of the deforming layer once lodgement has taken place (Benn & Evans 1996; Benn 2002). Both pressure shadows and striated boulder surfaces are clearly demonstrated by herringbone pattern macrofabrics and lodged stoss boulders respectively in the flutings that develop at the ice-deforming bed interface (Rose 1989, 1992; Benn 1994a). Hence, although some tills display fabrics that are well developed at all grain sizes, it is apparent that clast size alone does not control fabric development and moreover, it is the distribution of different clast sizes that has a major influence on the distribution of stress within a till.

Clast collisions may also be significant (Ildefonse et al. 1992), especially in tills with large numbers of variably sized clasts, such as those at Þorisjökull (Fig. 9); weaker fabrics in samples of smaller particles have been identified in tills and related to their greater susceptibility to collisions both with each other and with larger neighbours (Kjær & Krüger 1998; Carr & Rose 2003; Carr & Goddard 2007; Thomason & Iverson 2006). Consequently there is every reason to expect sub-boulder sized macrofabrics to be weakened, especially in tills that include numerous boulders such as those at Þorisjökull. Hiemstra and Rijsdijk (2003) have demonstrated an association between strong alignment of relatively large particles along shear planes and adjacent turbate structures produced through ductile shear in the finer matrix of tills. However, the micromorphology of the Þorisjökull sediments lacks evidence to support ductile intergranular shear, grain rotation or grain to grain interaction. Rather the micromorphological evidence indicates discrete brittle shear in both A and B units. The predominance of brittle microshears in the potential A horizons, and lack of structures relating ductile intergranular shear and grain rotation is somewhat unusual in these tills (c.f. Evans and Hiemstra, 2005; Evans et al. 2006a) and contradicts the fabric evidence that the sub-boulder element of the till matrix is subjected to variations in the local shear stress direction and magnitude in response to local perturbations with the deforming bed set up by large lodged boulders. A possible explanation for this is that subsequent emplacement of an overlying potential B horizon may control the late phase, final strain signature locked into an underlying A horizon. The depth of transferral of simple shear into the underlying units is difficult to estimate, but the development of a thin B horizon during the initial seasonal formation of a B/A couplet could in theory impart a penetrative shear signal into the underlying substrate (i.e. the top of the A horizon from the previous season). Under such a scenario, the shear strain signal with the finer matrix elements of both B and A horizons becomes controlled predominantly by subglacial conditions that prevail during the early part of the seasonal cycle when B horizons accrete.

Iverson et al. (2008) correctly point out that the poor constraints on shear strain magnitude in field based studies hamper accurate determinations of other variables affecting fabric development. Nevertheless a variety of field based observations on process-form relationships provide us with some clear indications of the operation of till emplacement in the sub-marginal depositional environment that cannot be replicated in laboratory experiments, hence macrofabric strengths from till deposits must be interpreted with those field based observations in mind. The deposits at Þorisjökull can be confidently interpreted as subglacial tills based upon their sedimentology and therefore their relatively weak sub-boulder sized macrofabrics, indicative of shear strains of <2 (Hooyer & Iverson 2000; Iverson et al. 2008), need to be reconciled with the predicted shear strains in such settings; these are at values of up to 100 if most basal motion occurs by bed deformation. The tills at Þorisjökull exhibit a range of sedimentological and clast fabric attributes that suggest different component parts of the till respond to applied stress in different ways and at different times in the accretionary cycle of the potential B/A units. The strong boulder fabrics clearly suggest the larger elements of the till matrix become lodged sub-parallel to ice flow early in the depositional cycle. The lodged boulders may then influence the distribution of applied stress through the deforming bed as it develops and thickens. Both A-axes and A/B plane fabric data in the sub-boulder fraction of the till suggest clast fabric strength is reduced as till is deformed and deposited in between and around lodged boulders, which effectively control the three dimensional distribution of applied stress on a local scale within the deforming bed. What is unusual about the micromorphological signature of the till matrices at this locality is that both potential B and A horizons are characterised by brittle microshears. It is unlikely these develop in response to ductile, viscous deformation as the till is deposited in between large lodged boulders, but we hypothesise this could be a late phase penetrative, deformational imprint as potential B units form at the start of each seasonal cycle.

Indeed, previous models of glacier sub-marginal till thickening (cf. Matthews et al. 1995, Krüger 1996, Evans & Hiemstra 2005) have emphasized the combined operation of various glacier sub-marginal processes over seasonal cycles. These processes include summer squeezing/flowage of subglacially deforming tills, as manifest in crevasse squeeze ridges and saw-tooth moraines (e.g. Price 1970), winter freeze-on of sub-marginal deforming till wedges, and spring melt-out and deformation of the till wedges leading to the liberation of porewater and its escape through the till matrix. Moreover, these processes operate in the marginal subglacial zone which is characterized by a gradual reduction in basal shear stress towards glacier snout. Tills created in these settings appear to possess strong boulder fabrics, moderate to strong sub-boulder sized clast macrofabrics and very weak shearing indicators at microscale, the latter being subordinate to water escape and sediment flowage features (Evans & Hiemstra 2005). None of these seasonally driven field conditions can be replicated in laboratory simulations of subglacial till shearing, reinforcing the argument that field based fabric strengths cannot be used as an indication of shear strain magnitude. It also demonstrates nonetheless that a variety of site-specific conditions and processes, including the influence of boulders and lee-side pressure shadows on sub-boulder sized macrofabrics, can impede the development of steady state fabrics.

Finally, the clast form data collected from the Þorisjökull tills provide us with important information on debris entrainment and transport pathways in temperate glaciers with deforming beds. The apparent role of bedrock plucking in the dilution of what should be mature subglacial clast form samples prompts us to propose that the co-variance plot for the data collected during this study (Fig. 5f) be employed as an exemplar for subglacial mountain tills, especially as the Type 1 co-variance plot proposed by Lukas et al. (2013) for Icelandic tills does not capture the influence of bedrock plucking in mountain glacier snouts traversing bedrock steps. The potential influence of clast shape on macrofabric development, specifically the extent to which elongate and slabby clasts (high C_{40} values) should create stronger S_1 eigenvalues, in A axis and A/B plane alignments respectively, in deforming media, cannot be meaningfully tested using the data from this study alone. This is because the clast shapes are predominantly blocky and hence a suitably large range of C_{40} values is not available. Future work on the testing of this clast shape-fabric strength relationship should employ data from other subglacial tills for inter-comparisons with the Þorisjökull site.

Conclusion

A vertical succession of alternating beds of massive and fissile diamictons on a Þorisjökull plateau icefield outlet foreland displays the characteristics of a thickening wedge of subglacial traction tills, each massive and fissile component resembling the A and B horizons previously identified in Icelandic subglacial deforming layers and potentially together representing a subglacial deforming layer couplet indicative of seasonal emplacement. This modern till assemblage demonstrates that it is possible to advect and stack tills and retain their internal structures, specifically in glacier sub-marginal locations as predicted by theory. The stratigraphic sequence indicates that less than 1 m of subglacial till is advected to the glacier margin per (annual?) deformation event (Fig. 9).

Numerous lodged boulders throughout the till sequence, in places arranged in weakly developed clast lines, display strong A-axes (S_1 0.76 - 0.81) and surface striae alignments at 030° - 210°, which parallel surface flutings and thereby indicate that fluting construction and till emplacement was related to moulding by consistent glacier flow from the south-southwest during the historical Little Ice Age. However, clast macrofabrics at the sub-boulder size, despite replicating the general SSW-NNE orientations of the lodged boulders, their surface striae and the flutings, are not as strong (S_1 <0.62 for A axes; S_1 <0.52 for A/B planes) as would be expected in a subglacially sheared medium, where shear strains up to 100 would not be unusual; laboratory experiments have demonstrated that shear strains of 7-30 generate S_1 eigenvalues >0.78. The Þorisjökull till macrofabric strengths are, however, not unlike those reported from other Icelandic tills (S_1<0.74 for A axes; S_1<0.67 for A/B planes), indicating shear strains too low to attain Iverson et als. (2008) steady state strain signature. If a steady state strain signature is a realistic postulate and all tills do tend towards this state when being sheared, then the Þorisjökull tills have not reached it. Our data indicate that this unexpectedly low measure of strain magnitude is most likely unrepresentative, because it is based upon only the sub-boulder sized strain markers, whereas the larger boulder-sized clasts display S_1 eigenvalues (0.76 – 0.81) representative of steady state strain magnitude. There are two, not necessarily mutually exclusive, explanations for these macrofabric trends:

1) By separating fabric data on lodged boulders from sub-boulder sized clasts, we have isolated the strain signatures of the lodgement and deformation components of subglacial traction till. These signatures indicate that the largest clasts are recording steady state strain and lodgement and hence the tills have been subject to cumulative shear strains of at least 7-30, whereas, in contrast, the deformation of matrix and sub-boulder sized clasts is recorded by weaker fabrics, which taken in isolation fail to provide a realistic reflection of the magnitude of the shear strain history or cumulative strain in the tills. The most likely cause of this dichotomy is the effect of clast collisions in clast rich till and the perturbations set up by the numerous lodged boulders (Fig. 9). This is consistent with observations on till fabrics in flutings and around lodged boulders and also questions laboratory based assumptions that clast size is not important in macrofabric development.
2) Because the emplacement of the tills has taken place in a sub-marginal environment, where basal shear stresses drop off but water and debris flux both increase, it is likely that till is susceptible to flowage, especially in colder upland settings where frozen-on till layers develop during winter and then melt out during summer to produce a more mobile matrix. If this process is indeed significant, we must be aware that sub-marginal tills are prone to small and intermediate scale strain markers becoming more mobile during emplacement and hence these are not appropriate locations to measure subglacial shear strain magnitude using macrofabrics. The micromorphology of these sediments strongly supports deformation via discrete brittle shear in both B and A type till units. Such a shear strain signal is commonly reported in B horizons, but not in A horizon tills where structures indicative of ductile intergranular shear and grain rotation are more common. At Þorisjökull this may be the result of late phase, penetrative overprinting of A horizon units as overlying B horizons accrete the following season.

An important corollary with respect to applications of this modern exemplar to interpretations of Quaternary glacial stratigraphy is that the range of clast macrofabric strengths on ancient tills, many of which display unexpectedly weak S_1 eigenvalues typical of very low strains (<2), is a reflection of the fact that they are sampled at former ice sheet and glacier sub-marginal settings, which are not representative of strictly subglacial processes and forms. Additionally, it is apparent that clast size is very likely influential in the development of steady state fabrics in response to subglacial shearing.

The A/B plane macrofabric data display unusually high degrees of isotropy, and their uniformly weaker clustering compared to partner A axis data sets indicates that A/B planes have not been locked in to an up glacier dipping imbrication. This could reflect the more variable dip orientations of A/B planes, which in the Þorisjökull tills are perhaps actually replicating the three dimensionally more variable dips of the anastomosing failure planes that develop in the narrow shear zones of a thin subglacial deforming layer. Consequently, A/B planes do not developing anything stronger than a girdle fabric. This is apparent even for boulders.

The wide range and unusually high angularity values displayed by the clast form data reflects the localized input of freshly plucked and hence relatively highly angular blocks to the deforming layer. This is a characteristic of stepped bedrock profiles beneath the snouts of mountain glaciers and therefore we propose the employment of the co-variance plot for the data collected during this study as an exemplar for subglacial mountain tills.

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Figure captions

Figure 1: Location map of the study site in west-central Iceland.

Figure 2: Glacial geomorphology of the of the foreland of the western outlet of the Þorisjökull plateau icefield where this study was undertaken: a) surficial geology and geomorphology map extract from Evans et al. (2006) based on 1999 aerial photography. Orange is till and moraine, yellow is glacifluvial, black lines are moraine ridges, green lines are flutings and red lines are ice-cored/controlled moraine ridges. North is towards the top of the map; b) annotated 2008 aerial photograph (Loftmyndir ehf) extract of the same area depicted in (a), showing inset recessional moraines and flutings relating to Little Ice Age snout expansion and retreat. The 1999 moraine and associated ice-cored moraine at the left of the image is located at the snout margin in the map extract depicted in Figure 2a.

Figure 3: Ground views of the study area, showing: a) the diamicton surface exposed at the top of the gorge by fluvial reworking of loose surface morainic debris and characterized by protruding bullet-shaped clasts aligned with their A-axes and striated A/B planes parallel to adjacent flutings; b) the cliff in the middle section of the gorge where the meltwater stream incision has revealed the stacked sequence of alternately fissile and massive diamictons and associated bullet-shaped clasts.

Figure 4: Macrofabric and striae data for boulders sampled over the study area independently from the vertical profile logs: a) A-axis (upper) and A/B plane (lower) macrofabric stereonets for boulders protruding from the 30 m wide fluvially scoured diamicton surface; b) A-axis (upper) and A/B plane (lower) macrofabric stereonets for boulders exposed at various depths along the sediment gorge; and c) rose plot of boulder surface striae orientations for boulders protruding from the 30 m wide fluvially scoured diamicton surface (upper) and exposed at various depths along the sediment gorge (lower). All A/B plane stereonets plot the dip direction of the A/B plane.

Figure 5: Site A stratigraphy and sedimentology: a) annotated photograph log; b) vertical profile log, showing locations of clast macrofabric and grain size samples as well as contoured stereoplots for both A-axis and A/B plane macrofabrics; c) clast form characteristics plotted with depth alongside the section log; d) clast form data for each lithofacies depicted as roundness histograms and clast shape ternary plots and plotted alongside the section log; e) clast macrofabric data plotted with depth alongside the section log; f) co-variance graphs for RA against C40 and mean roundness against C40, alongside the characteristics of the Type 1 co-variance plot from Lukas et al. (2013).

Figure 6: Site B stratigraphy and sedimentology: a) annotated photograph log, together with enlargements (i-iii) of details of the sedimentary structures indicative of fissile and massive diamictons; b) vertical profile log, showing locations of clast macrofabrics as well as contoured stereoplots for both A-axis and A/B plane macrofabrics; c) clast macrofabric data plotted with depth alongside the section log.

Figure 7: Thin sections

Figure 8: Secondary clast macrofabric analytical graphs: a) clast macrofabric shape ternary diagrams, depicting the positioning of glacial deposits of known origin as envelopes and samples from this study according to their isotropy and elongation (after Benn 1994); b) modality-isotropy plots, modified from Hicock et al. (1996) by Evans et al. (2007), showing envelopes for typical subglacial deposits and the samples from this study (left graph is for A-axis data and right graph for A/B planes data); c) clast macrofabric shape ternary diagrams plotting the positioning of samples from this study, colour coded according to their horizon and compared to envelopes of Icelandic till fabrics from previous studies; d) clast macrofabric shape ternary diagrams, plotting the positioning of samples from the same couplets (LF 3 & 4, LF 5 & 6, LF 7 & 8) but from different horizons. These hysteresis-type curves provide an assessment of the vertical progression in macrofabric strength through A and B horizons and from up-ice to down-ice locations. A = A horizon and B = B horizon, with suffixes "u" or "l" to indicate upper or lower parts of horizons respectively. Open, colour-coded arrows show the change from up-ice (U) to down-ice (D) curves on same couplets for LFs 3 and 4 and LFs 5 and 6.

Figure 9: Conceptual model to explain the development of multiple subglacial tills at Þorisjökull and incorporating the processes proposed by Boulton and Hindmarsh (1987), Evans and Hiemstra (2005) and Benn and Evans (1996) for active temperate glacier snouts with deformable substrates. The model assumes that seasonal conditions impact upon glacier sub-marginal processes and hence identifies the separation of spring-summer deformation events by a phase of winter freeze-on. During "deformation event 1" a subglacial traction till comprising A and B horizons develops over a glacitectonite of former glacifluvial outwash, within which hydrofracture fills are commonly produced by elevated groundwater pressures. The first till developed over a glacitectonite will be characterised by a basal zone of sheared inclusions. Plucked blocks derived from bedrock steps below the icefall are delivered to the deforming layer by meltout of debris-rich basal ice. "Deformation event 2" begins after winter freeze-on of the thin snout ice to the top of the A horizon, initiating a decollement plane and down-ice displacement of the top of the A horizon. This is followed by the advection of a new subglacial deforming layer in response to thawed conditions and elevated porewater pressures in the following spring-summer period. At this time the new B horizon is developed in the top of the old A horizon and deeper shear planes may develop in the older till units due to deformation partitioning. Specific processes identified widely in subglacial traction tills, including ploughing, clast lee-side matrix perturbations, lodgement and abrasion of large clasts, clast collisions and micro-shears (fissility) are also incorporated into the model. Note that the clast macrofabrics are examples from this study that are indicative of the various levels in the A and B horizons. The cumulative relative displacement curves are representative of the individual displacement events and therefore must be combined when assessing the total strain signature for a multiple till sequence. The impact of potential shearing at depth within a subglacial till is reflected in the alternative curves for deformation event 1.