Superimposition of ribbed moraines on a palaeo-ice stream bed: implications for ice stream dynamics and shutdown

Chris R. Stokes[1], Olav B. Lian[2], Slawek Tulaczyk[3],
and Chris D. Clark[4]

[1] Department of Geography, University of Reading, Reading, RG6 6AB, UK; c.r.stokes@reading.ac.uk

[2] Department of Geography, University College of the Fraser Valley, 33844 King Road, Abbotsford, British Columbia V2S 7MB, Canada;olav.lian@ucfv.ca

[3] Department of Earth Sciences, University of California, Santa Cruz, CA 95064, USA;tulaczyk@pmc.ucsc.edu

[4] Department of Geography, University of Sheffield, Sheffield, S10 2TN, UK; c.clark@sheffield.ac.uk

Abstract

The sediments and landforms preserved on palaeo-ice stream beds can provide important information about their subglacial conditions, flow mechanisms, and the processes accompanying their shut-down. In this paper, detailed observations of an intriguing subglacial landform assemblage of ribbed moraines superimposed on glacial lineations on the Dubawnt Lake Ice Stream bed (north-west Canadian Shield) are presented; including their morphometry, internal structure (from Ground Penetrating Radar (GPR) surveys and from glaciogeological analysis), and sedimentological characteristics (from sediment architecture and lithofacies analysis). The observations suggest an abrupt change in ice dynamics that correlates with two phases of glacial landform development. This hypothesis is based on evidence from a deformed lodgement till which subsequently underwent brittle deformation and developed prominent thrust (shear) structures and tension fractures. Tension fractures are observed in a sediment exposure and thrust structures are observed in GPR surveys, where they occur most prominently in the ribbed moraine ridge crests. The presence of the fractures, and their association with a population of clasts in the till that are orientated with their a-axes transverse to the inferred ice flow direction suggests a compressional

flow regime. It is therefore inferred that the glacial lineations were formed under an extensional flow regime during ice stream activity but that at some point, patches of till under the ice stream stiffened through dewatering. The subsequent increase in basal shear stress resulted in compressional flow and the development of subglacial thrusting and the building of ribbed moraines. We therefore suggest that ribbed moraines may form in areas of compressional flow under ice streams, i.e. sticky spots and/or at the transition between slow and fast ice flow along parts of an ice stream. The general absence of ribbed moraines on most other palaeo-ice stream beds suggests that either those ice streams continued operating during deglaciation or, processes other than the development of localised compressional flow (sticky spots) led to their shut-down (e.g. ice depletion).

Keywords: ice stream; ribbed moraines; subglacial bedforms; sticky spots

Introduction

Ice streams represent the most dynamic component of continental ice sheets and play a dominant role in ice sheet mass balance and their response to climate change (Bamber et al., 2007). For this reason, they are a major focus of research in Greenland and Antarctica, and recent observations have revealed rapid (annual to decadal) changes in their speed (Joughin et al., 2004) and their flow direction (Conway et al., 2002). The availability and distribution of subglacial meltwater is known to be a critical control on their activity (Parizek et al., 2002; Vogel et al., 2003) but their temporally and spatially variable behaviour represents a major challenge for predictive modelling. Moreover, the development of ice stream models is, to some extent, limited by a dearth of observations regarding how sediment behaves under ice streams, and a limited understanding of the details of the processes that drive the temporal and spatial changes in their activity.

The growing realisation of the importance of ice streams has stimulated many workers to search for their locations in regions formerly occupied by ice sheets (e.g. Shipp et al., 1999; Stokes and Clark, 2001; Ó Cofaigh et al., 2002, 2005; Winsborrow et al., 2004; Ottesen et al., 2005). Significantly, it has been recognised that detailed examination of the sediments and landforms preserved on a palaeo-ice stream bed can provide important information about their basal conditions, flow mechanisms, and the processes that trigger ice stream shut-down (e.g. Christofferson and Tulaczyk, 2003; Lian et al., 2003; Clark et al., 2003; Dowdeswell et al., 2004; Iverson and Hooyer, 2004; Ó Cofaigh et al., 2005; Stokes et al., 2007). Therefore, data collected from the exposed beds of palaeo-ice streams has the potential to provide valuable insights about subglacial conditions and processes, complementing those being obtained from under the bed of active ice streams (e.g. Tulaczyk et al., 2001; King et al., 2003; Vogel et al., 2005; Peters et al., 2007; Smith et al., 2007). The inferred strain history of sediments preserved on a palaeo-ice stream bed, for example, might assist in parameterising the basal conditions in ice stream models. Likewise, palaeo-ice stream beds that have remained largely unmodified since deglaciation might preserve data that could be used to test theories regarding ice stream shut-down (e.g. Christofferson and Tulaczyk, 2003)

In this paper, we present detailed observations of an unusual assemblage of landforms on a palaeo-ice stream bed, including their sedimentology and internal structure. The Dubawnt Lake Ice Stream operated immediately prior to deglaciation and produced a remarkably intact and well-preserved bed imprint on the north-western Canadian Shield (cf. Stokes and Clark, 2003). Previous work (Aylsworth and Shilts, 1989a, b; Stokes and Clark, 2003; Stokes et al., 2006) reported that in some places transverse ridges, known as ribbed moraines, occur superimposed on the mega-scale glacial lineations and drumlins that characterise the ice stream bed. Because ribbed moraines are not associated with rapid ice flow (cf. Hättestrand and Kleman, 1999; Dunlop and Clark, 2006), their patchy appearance on an ice stream bed is intriguing and their presence implies a marked change in ice dynamics following ice stream activity.

Our motivation for this investigation is that this landform-sediment assemblage might provide important insights regarding the basal processes that triggered or immediately followed ice stream shut-down. This is potentially significant given the interest in the processes that led to the shut-down of Ice Stream C in West Antarctica (Anandakrishnan et al., 2001), and the possibility that other presently active ice streams are slowing down and may shut down in the near future (e.g. Bougamont et al., 2003). More generally, the observations presented here are an important test of theories which seek to explain the genesis of subglacial bedforms, such as ribbed moraines (e.g. Fisher and Shaw, 1992; Bouchard, 1989; Dunlop and Clark, 2006).

Methods

The Dubawnt Lake Ice Stream imprint lies immediately to the north-west of the last inferred position of the Keewatin Ice Divide in the Laurentide Ice Sheet (Aylsworth and Shilts, 1989a; b; Stokes and Clark, 2003; McMartin and Henderson, 2004). McMartin and Henderson (2004) compiled an extensive dataset of ice flow indicators from this region and suggested that the complex arrangement of cross-cutting subglacial bedforms and striations resulted from the migration of the main ice divide by as much as 500 km during the build up and decay of the Laurentide Ice Sheet. The onset zone of the Dubawnt Lake ice stream extends very close (50-100 km) to the last inferred position of the ice divide (cf. McMartin and Henderson, 2004: Fig. 12). Moreover, the terminus of the ice stream is closely associated with the McAlpine moraine, dated to around 8.2 ka BP (Dyke, 2004). For these reasons, the ice stream is presumed to have operated very close to final deglaciation of the area and probably contributed to the final south-westward migration of the ice divide and its eventual demise (cf. Stokes and Clark, 2003; McMartin and Henderson; 2004)

Fields of ribbed moraines exist in a broad zone surrounding the last inferred position of the Keewatin Ice Divide (Aylsworth and Shilts, 1989a; b) and on the Dubawnt Lake Ice Stream bed they are clearly superimposed on the underlying ice stream lineations, covering approximately 7% of the bed, with individual fields of ribbed moraine ranging in extent from <1 km^2 up to >2,500 km^2 (cf. Stokes et al., 2006). Based on the mapping of Stokes et al. (2006), a small and relatively accessible field of ribbed moraines was selected for detailed field investigation in the presumed fast-flowing trunk of the ice stream, see Figure 1. The study area covers around 275 km^2 and is centred on 100° 40' 21'' W, 64° 32' 23'' N, immediately south-east of, and including, the area where the Thelon River enters the western end of Beverly Lake (Nunavut Territory).

The first task was to produce a detailed map of the glacial geomorphology in the study area and this was performed using Landsat ETM+ satellite imagery (panchromatic band = 15 m spatial resolution) and both digital and hard copy stereo-air photographs. These data were obtained from the Natural Resources of Canada GeoGratis website (http://geogratis.cgdi.gc.ca/) and the National Air Photograph Library of Natural Resources Canada (http://airphotos.nrcan.gc.ca/). Orthorectification and on-screen digitising of glacial geomorphology using break of slope mapping based on stereo-aerial photographs was performed using ERDAS Imagine 8.7 software and this mapping was also verified by field observations conducted in the summers of 2004 and 2005.

To identify any prominent sub-surface structures in the bedforms, Ground Penetrating Radar (GPR) was used to survey a number of the ribbed moraine ridges. PulseEKKO 100 GPR was deployed in a reflection-profiling mode with either 50 MHz or 100 MHz antennas. Additionally, one common-midpoint survey was performed in the study area to determine the vertical structure of radar wave velocity using semblance analysis. Topographic data were collected along profiles using a handheld Global Positioning System (GPS) and altimeter. The GPS unit was a 'GARMIN GPS 12 Personal Navigator' with an accuracy of 4-8 m in the horizontal. The vertical accuracy of the altimeter was ~1 m and this was calibrated to a known elevation on a 1:50,000 topographic map on each of the two days that profiles were being taken.

Interpretation of the GPR data benefited from correlation with a well-preserved exposure in an active river-cut bank within the study area. The face of the exposure is oriented transverse to the inferred ice stream flow-direction, and is situated immediately down-ice (< 200 m) from the area of surveyed ribbed moraine. This section was logged with regard to sedimentary structures, bed contacts and sediment architecture. The clast macrofabrics were measured by recording the azimuth and plunge of the long axis (*a*-axis) of stones (> 2 cm long) in the diamicton; four such *a*-axis fabrics were recorded, each consisting of about 50 measurements. Other characteristics that were recorded were the trend of stone stoss and lee ends of clasts (cf. Krüger, 1984); the orientation of the freshest striae on the upper surfaces of clasts; the position of keels on clasts; and the orientation of fractures in the diamicton. Stone keels can form from a curvilinear surface as the stone is ploughed at the ice/substrate interface prior to lodgement. After lodgement, a planar surface (facet) can develop opposite the keel. The position of the keel therefore reflects whether or not (and how) a clast's position has been modified during the till (de)forming process (cf. Lian et al., 2003). The freshest striae can be identified through careful examination of washed stones and cross-cutting relationships can be used to determine the youngest. These data were used together to infer the genesis of the diamicton (e.g. Krüger, 1984; Stanford and Mickelson, 1985; Hicock and Dreimanis, 1985; Dreimanis, 1993; Hart, 1994; Lian and Hicock, 2000; Lian et al., 2003).

Observations

Morphometry and Pattern of Subglacial Bedforms

Observations

The distribution of ribbed moraines on the Dubawnt Lake Ice Stream bed is shown in Figure 1, which also indicates the location of our study area within an area interpreted as the fast-flowing trunk of the ice stream. Figure 2 shows a detailed map of the glacial landforms in the study area

(overlaid on a digital elevation model) and a Landsat Enhanced Thematic Mapper Plus (ETM+) image of the same area. Four major glacier-related landform types are found in the study area: glacial lineaments, ribbed moraines, abandoned shorelines, and a large esker.

The glacial lineaments dominate the landscape and their exceptionally parallel orientation records an ice flow direction toward the north-west (orientation ~300º). It can be seen from Figure 2 that the glacial lineations have a variety of lengths, widths and elongation ratios. This is in contrast to those in other areas of the ice stream trunk, which are arranged in a remarkably coherent and densely packed manner and which are mega scale glacial lineations (see for example, Fig. 5 in Stokes and Clark, 2003). Given that there is persuasive evidence that this part of the ice stream bed was formerly submerged beneath a large proglacial lake (see Craig, 1964; Stokes and Clark, 2004), it is likely that the partially fragmented and subdued appearance of the glacial lineations in this location, compared to elsewhere on the ice stream bed, is due to wave-washing and some surficial erosion during lake formation and drainage. A suite of abandoned shorelines are also mapped in the study area (see Fig. 2), which record successive lowering of this lake during its waning stage.

The glacial lineations have also been modified and fragmented by the superimposition of ribbed moraines in discrete patches (Fig. 2). This situation is rare: most other reported ribbed moraines are found to be overprinted by glacial lineations (cf. Knight and McCabe, 1997; Hättestrand and Kleman, 1999; Dunlop and Clark, 2006). The ribbed moraine ridges lie perpendicular to the long axis of the underlying lineations and because the ribbed moraine ridge crests form transverse to ice flow, it can be inferred from their alignment that they too, were formed by ice flowing towards the north-west (~300º). The dimensions (lengths/widths/crest to crest spacing) of the ribbed moraines in this study area fall within the typical range of an extensive dataset of ribbed moraines mapped by Dunlop and Clark (2006), although in this location the mean wavelength (272 m) is somewhat less than the mean wavelength (505 m) reported by those authors, i.e. the ribbed moraines in our study area are more closely packed together than is commonly observed.

Observations in the field indicate that the ribbed moraine crests are typically on the order of 10-15 m above the inter-ridge depressions and they tend to exhibit asymmetric profiles with steeper distal slopes. This accords with the description of Aylsworth and Shilts (1989a: p.10) who described their form as resembling "inclined plates of sediment thrust one on top of another". The surface of some of the ribbed moraines has a wave-washed appearance (including beach deposits) and some of the ridges exhibit wave-cut benches. The wave cut benches are typically cut 2-3 metres horizontally into the ribbed moraines and appear to indicate only a few metres of surface lowering in the vertical direction, assuming that the ridge crest itself has not been lowered. We therefore estimate that the ridge crests have been lowered by a minimum of around a few metres since their formation. Although we have no way of verifying the maximum depth of lowering, we note that the heights (10-15 m) and form of the ribbed moraines in our study area are consistent with those from a much larger sample mapped by Dunlop and Clark (2000), which show a mean height of 17 m. This suggests that large-scale sediment re-working and modification has not substantially altered the overall form of the ribbed moraines. Aylsworth and Shilts (1989a; b) did not discuss the possibility that some the ribbed moraine crests on the Dubawnt Lake ice stream bed had been eroded but they did note the presence of boulders on many of the ridge crests and their general absence in between the ridges. We suggest that the boulder lag may be an artefact of the washing away of fines on the ridge crests and that their presence between ridges is likely obscured by subsequent bog infilling.

Although the ribbed moraines may have been superficially modified by lake occupation and drainage, they do appear to be reasonably intact, i.e. ridge crests can be traced continuously for several hundred metres and crest to crest spacing is not atypical. The crests in our study area are not anastomosing; nor do they show a predominant pattern of down-ice or up-ice curving ridges. Rather, they are similar in appearance to what Dunlop and Clark (2006: Figure 17) classify as 'broad arcuate ribbed moraines'. We also note that ribbed moraines and glacial lineations are not restricted to any particular topographic setting, appearing on both high and low terrain and this is true for the whole ice stream bed (cf. Stokes et al., 2006).

In addition to the glacial lineations and ribbed moraines, a large, prominent esker is present in the east of the study area running north-northwestward into Beverly Lake (see Figure 2). Inspection of satellite imagery and other published maps (e.g. Aylsworth and Shilts, 1989c) of the region immediately outside our study area indicate that this esker is fed by two large tributaries, one of which can be traced almost continuously for a further 100 km up-stream. Taken as a whole, this esker system, and those around it (see for example Aylsworth and Shilts, 1989a, b, c), are orientated towards the north-northwest and, assuming that they formed approximately perpendicular to the ice margin position, reveal a flow direction towards ~318º during deglaciation.

Interpretations

The glacial lineations in the study area form part of the much larger Dubawnt Lake ice stream flow-set reported in Stokes and Clark (2003) and we therefore interpret them as being formed by fast ice flow. The ribbed moraines are clearly superimposed on the glacial lineations and we therefore interpret that they formed after the generation of the glacial lineations. Their presence probably indicates an abrupt change in ice dynamics because ribbed moraines are not normally associated with fast ice flow (Dunlop and Clark, 2006). Although both sets of landforms have been modified by erosion from proglacial lake formation and drainage, they both record ice flow towards the north-west (~300º). In contrast, the large esker that terminates at the southern end of Beverly Lake is orientated towards the north-northwest (~318º: see Figure 2). We therefore suggest that only a short time period elapsed between the formation of the glacial lineations and the formation of the ribbed moraines but that a longer time may have elapsed between the formation of the ribbed moraines and deglaciation, thus allowing for the observed shift in ice flow direction. Hence, our working hypothesis is that the process responsible for the formation of the ribbed moraines occurred immediately after or during ice stream shut-down. Following this, there were no other major ice flow events (recorded by glacial lineations or even smaller streamlined landforms) prior to deglaciation, but a subtle shift in the flow patterns and the position of the Keewatin ice divide took

place, which accounts for the slight mismatch in the orientation of the overlying eskers (cf. McMartin and Henderson, 2004). This inferred change in flow direction is not just based on the single esker in our study area but on numerous other esker systems nearby that were mapped by Alysworth and Shilts (1989a, b, c).

Sub-Surface Structure of Ribbed Moraines

Observations

In order to investigate their internal structure, eight individual ribbed moraine ridges (ridges RM01-RM08) were investigated using Ground Penetrating Radar (GPR), see Figure 3 for their location. Both 50 and 100 MHz radar antennas were employed in the GPR surveys, with the former showing consistently good results in terms of a satisfactory balance between resolution and penetration. One 100 MHz survey will, however, be shown here to illustrate the variability of small subsurface structures.

A common-midpoint survey performed in the study area using 50 MHz antennas showed an upper layer, several metres thick, with relatively low radar velocity, ~0.1 m/ns, overlying material of higher velocity, ~0.14 m/ns. In all GPR sections we also observed a continuous, few-to-several-meter thick, near-surface layer: termed radar facies 'A' in Figures 4, 5, 6 and 7. This layer truncates all other radar facies.

At the left hand side of Figure 4 we identify radar facies 'A-1', which has consistent high-angle internal structures. In the field, the locality of facies 'A-1' was associated with a topographic step (shallow wave-cut bench?) cut into a ribbed moraine and other indications of erosive and depositional action (possible primitive shoreline?) along the edge of a proglacial palaeo-lake (Stokes and Clark, 2004).

We classify much of the subsurface material covered by our GPR sections as radar facies 'B'. This radar facies is characterized by relatively short, weak reflectors which are commonly oriented at

high angles to the horizontal (e.g. Figure 4 and 5). The lower boundary of this facies is, in places, sharp (e.g. as seen on the right section of Figure 4) and can be difficult to define, even in places where the signal strength was sufficient to 'see' the bottom of the facies (e.g. Figure 6). The thickness of radar facies 'B' tends to increase in places where ribbed moraines rise above the surrounding landscape and the facies thins, or maybe even disappears, in the depressions between the ribbed moraines. In transverse sections, facies B is commonly lenticular in shape (e.g. Figure 5). The presence/absence of this facies in depressions is difficult to ascertain because such depressions are also associated with the 'chaotic' radar facies 'D', which overprints all other structures (Figure 4). In places, facies 'B' is cut by medium-strong, low-angle reflectors. The direction of these reflectors is typically consistent with the inferred direction of ice flow (~300°), i.e., the hanging wall is displaced up and above the footwall in the direction of ice flow (examples are shown as dotted lines in Figures 4, 6, and 7). Radar facies 'C' is characterised by strong, relatively continuous reflectors that are mostly (sub)horizontal in orientation and occur below the more chaotic looking facies 'B'.

We differentiate between facies 'B' and facies 'C' on the basis that facies 'B' reflectors are shorter, discontinuous, and rarely form contiguous horizontal structures. They are more 'choppy' in appearance. In contrast, facies 'C' reflectors are longer and form continuous horizontal to sub-horizontal structures. This is most clearly seen towards the right hand side of Figure 4, although it is not always obvious. In places, the precise boundary between the two is difficult to define.

Radar facies 'D' is always associated with topographic depressions and the presence of vegetation characterised by higher water demand, such as shrubs and dwarf trees. The internal architecture of this facies is very consistent, comprising multiple parabolic reflectors which tend to obscure original subsurface layering and structures. In Figure 6 we have applied a migration procedure to filter out, with some success, the impact of these reflectors on the reflectors associated with the overall subsurface structure.

Interpretations

The common-midpoint survey using 50 MHz antennas showed an upper, several-meter-thick layer with relatively low radar velocity, ~0.1 m/ns, overlying material of higher velocity, ~0.14 m/ns. Whereas the former is a typical velocity for near-surface sediments, the latter is unusually high and is close to the velocity of radar waves in temperate ice. The simplest explanation of this situation, and our favoured interpretation, is that the upper layer represents an active layer, thawed in summer, with the underlying material representing permafrost. If correct, this would imply that the continuous, few-to-several-meter thick, near-surface layer (facies 'A') that truncates all other radar facies is the active layer. We therefore interpret facies 'A' as the active layer which experiences thawing and freezing on seasonal basis. The localised radar facies 'A-1' (left hand side of Figure 4) has consistent high-angle internal structures and because this locality is associated with a shallow (wave-cut?) topographic step in a ribbed moraine and other indications of erosive (and depositional) action we interpret it as a small wedge of wave-washed deposits that accumulated successively as a result of proglacial wave erosion into the ribbed moraine after deglaciation.

The bulk of the subsurface material covered by our GPR sections is classified as radar facies 'B', which is characterised by relatively short, weak reflectors oriented quite frequently at high angle to horizontal. In transverse sections, facies 'B' is commonly lenticular in shape and its thickness tends to increase in places where ribbed moraines rise above the surrounding landscape (see Figure 5 and right hand end of Figure 4). We therefore interpret facies 'B' as material of subglacial origin that is predominantly responsible for landform generation, i.e. it seems to form the bulk of the landforms.

Moreover, in places, facies 'B' is cut by medium-strong reflectors, which are consistent with the inferred direction of ice flow (Figures 4, 6, and 7). We interpret these reflectors to represent thrust structures. In Figure 6, these proposed thrust structures are accompanied by conjugate discontinuities, dipping in the direction opposite to the dip direction of the thrusts themselves. We infer that these structures played a key role in generation of the large ribbed moraine (Figure 6) by enabling glaciotectonic shearing and stacking of subglacial material. In the absence of identifiable

unique markers on both sides of these structures it is not feasible to unequivocally evaluate the amount of offset accommodated by them. However, the very fact that the structures produce recognizable disturbances in the 50 MHz radar sections suggests that the offsets exceed the limit of spatial resolution, which is ~1 m, or about a quarter of the radar wavelength. Note that the upper limit of facies 'B' is defined by the extent of facies 'A' and because facies 'A' is interpreted as the active layer (and is therefore unrelated to the sediment architecture of the ribbed moraines) it might be that the true upper limit of facies 'B' is stratigraphically higher.

Below facies 'B', radar facies 'C' is characterised by strong, relatively continuous reflectors that are mostly (sub)horizontal in orientation. Comparison of this radar facies to other radar profiles taken from other parts of the Dubawnt Lake palaeo-ice stream bed, suggest that facies 'C' probably represents a crudely-stratified sequence of fluvioglacial sediments, which we have observed in other sections on the ice stream bed below a regionally widespread reddish, highly-consolidated, matrix-supported diamicton, see Figure 8.

The depressions in between ribbed moraine crests are associated with the 'chaotic' radar facies 'D', which overprints all other structures in these locations (Figure 4). The architecture of this facies comprises multiple parabolic reflectors which tend to obscure original subsurface layering and structures. Although we have not observed this type of material in exposures, we infer that this facies represents a recent overprint of postglacial permafrost processes on the glacial sedimentary sequence. The numerous parabolic reflections are caused by subsurface discontinuities, point reflectors, such as ice lenses and/or ice wedges.

Sedimentology of Ribbed Moraines

Observations

Shilts et al., (1987) and Aylsworth and Shilts (1989a; b) noted that the sediment within the ribbed moraine ridges on the Dubawnt Lake ice stream bed tended to consist of stony till. They also noted,

however, that the ribbed moraines can develop from the same deposits as those which form the glacial lineations, including re-worked sand and gravel. Here we report on observations from a section in the bank of the Thelon River, shown in Figure 9, which lies immediately down-ice of the field of ribbed moraines which we surveyed using the GPR. The section is just a few hundred metres from the nearest prominent ridge crest (see location on Figure 3).

The section consists of 4–5 m of reddish, highly-consolidated, matrix-supported diamicton (unit 1), which is sharply overlain by ~3 m of light grey-coloured normally consolidated sand and gravel (unit 2); a lower contact with unit 1 is not observed (Figure 9). Unit 2 is poorly sorted, with a texture similar to that of unit 1, and it is nearly massive; crude bedding and small lenses of sorted sand are observed locally. Most stones in unit 1 are subrounded, and many are facetted, bullet shaped, and show stoss-lee forms (cf. Krüger, 1984). A relatively large proportion of these stones have developed keels which are commonly found opposite the most prominent facet. Multiple sets of fine striae are found on washed stones extracted from the section face and from the shore of the river, immediately below the section face.

Unit 1 contains a pair of prominent parallel fractures, separated from each other by about 30 cm; these fractures occur ~1.5 m above the base of the section (Fig. 9). Both fractures dip to the NW (~300˚). The fractures are filled with fine silt that shows no bedding. This silt does, however, have in places discontinuous boudin-like structures. Approximately 10–12 cm above the upper fracture occur a discontinuous line of bullet-shaped boulders (crude boulder pavement?) which are oriented (*a*-axes) roughly parallel to the strike of the fractures. These boulders show two main sets of striae: one set is oriented parallel to the stone *a*-axes (i.e. transverse to ice flow), while the other set, which consists of the finest striae, is oriented transverse to the stone *a*-axes (i.e. parallel to ice flow), see Figure 10. In the upper ~1 m of unit 1 are smaller discontinuous fractures, some of which have been locally eroded and filled with bedded fine sand and silt.

Stone *a*-axis fabrics were measured at four sites within unit 1 (Fig. 9). Fabrics C and D were collected near the base of the exposure, about 4 m apart. Fabrics A and B were collected about 4 m

and 2.5 m higher in the unit, respectively, fabric B being ~1 m below the pair of large fractures (Fig. 9). All of the stone fabrics, except fabric D, have spread unimodal shapes oriented roughly SE-NW; fabric D, on the other hand, is girdle-like, although a weak E-W mode is suggested, see Figure 11. Fabric A also has a spread-out transverse mode that plunges roughly to the S, and this trend is consistent with the *a*-axis orientation of the 'pavement' stones that occur near the pair of large fractures (Fig 11).

The majority of stone keels measured at all the *a*-axis fabric locations, except those at the location of fabric C, are found in-situ. That is, with the keel facing down and the opposing most prominent facet facing up (Fig. 11). At the location of fabric C, equal numbers of stones have their keels facing up, down, or sideways (Fig 11).

Interpretations

The lower unit at our exposed section (unit 1) consists of highly-consolidated matrix-supported diamicton containing striated stones, and this suggests a subglacial origin for this unit. This interpretation is supported by the shape of most of the stone *a*-axis fabrics which show a main spread-unimodal mode trending in the direction of ice flow that is also indicated by the surficial landforms (Fig. 11). Moreover, the occurrence of bullet-shaped striated and facetted stones with prominent (plucked) lee ends, and the presence of keels on many stones which have formed opposite the most prominent facet, suggests that unit 1 is till that was deposited initially by lodgement (cf. Lian et al., 2003: p. 104). The fact that the orientation of stone lee ends is diverse, however, indicates post depositional deformation of this unit, although the *in-situ* orientation of the majority of stone keels suggest that this post deposition deformation was minor and probably restricted, for the most part, to stone rotation about their *c*-axes.

We interpret the two prominent silt-filled fractures near the centre of unit 1 as tension fractures, and these indicate an episode of brittle deformation. The dip direction of these fractures is to the NW. This orientation is not in contradiction with the orientation of the thrust structures observed in the

ribbed moraines which tend to dip to the SE. Thrust structures are shear structures which rise in the direction of force (ice flow in this case), whilst the structures we observe in the exposure are tension fractures, which commonly dip down-ice. Tension fractures in subglacial sediments that dip down-glacier and thrust (shear) fractures that dip up-glacier in till, and in other subglacial sediment and in bedrock, have been used together as indicators of ice-flow direction and till rheology in several previous studies (e.g. Hicock and Dreimanis, 1985; Dreimanis, 1993). They tend to occur in situations where there is substantial drag at the sole of the glacier, i.e. as in the case of compressional glacial flow (e.g. Hicock and Dreimanis, 1985). The orientation of these structures is consistent with that of the landforms and the main stone *a*-axis fabric modes, which supports a glacigenic origin for them. The dip direction of the fractures is also consistent with the orientation of the finest/freshest (youngest?) set of striae on the upper surfaces of stones that form a discontinuous pavement adjacent to the fractures (Fig. 10). Boudin-like structures found in the silt that fills the tension fractures indicates subsequent ductile deformation within the fractures. This is not surprising because once the fractures had opened they would have acted as drainage routes for subglacial water.

Unit 1 is sharply overlain by ~3 m of light grey-coloured crudely-bedded, normally consolidated, sand and gravel (unit 2). We interpret unit 2 to have been reworked form unit 1 by mass movement and minor fluvial activity. Sediment comprising unit 2 would have moved downslope, away from the face of the exposure (to the NW) into Thelon River basin.

Reconstruction of Ice Sheet Dynamics

The Dubawnt Lake Ice Stream is thought to have operated for a relatively short time (<1000 years and probably <500 years) prior to 8.2 (uncalibrated) ka yr BP and was the last vigorous ice flow event in this region (cf. Stokes and Clark, 2003; McMartin and Henderson, 2004). After around 7 ka BP the Keewatin Ice Divide was dissected by marine incursion through Chesterfield Inlet and

through to the lower reaches of the Thelon River basin (McMartin and Henderson, 2004). Ice had finally disappeared from the area soon after 6.5 ka (Dyke, 2004). At some point after, or during, ice stream shut-down, ribbed moraines formed in distinct patches on the ice stream bed. It is impossible to know precisely when the ribbed moraines formed but we suggest that their genesis occurred relatively recently after the ice stream operated because they lie almost perfectly transverse to the ice flow direction indicated by the glacial lineations.

The ribbed moraines cover around 7% of the Dubawnt Lake Ice Stream bed, forming patches which range in area from <1 km^2 up to 2677 km^2 (Stokes et al., 2006). Their occurrence does not appear to correlate with any obvious topographic setting (e.g. depressions, stoss slopes, lee slopes) nor do they appear to be related to the underlying bedrock geology: they occur, for example, in areas underlain by both igneous and sedimentary rocks (see Donaldson, 1968). Aylsworth and Shilts (1989a: p. 1) suggested that "less deformable sediments formed the ribbed moraine while more deformable sediments form drumlins". The fact that ribbed moraines are commonly found in the same location as glacial lineations (Figure 2), however, would appear to falsify this hypothesis (see also Figure 13 in Aylsworth and Shilts, 1989; Stokes et al., 2006). Rather, the most logical explanation for their presence is a significant change in basal processes and ice dynamics which results in the reorganisation of subglacial sediment.

On the basis that the ribbed moraines in this area resemble inclined plates of sediment thrust one on top of the other, Aylsworth and Shilts (1989a) suggested that they might be formed under conditions of compressive ice flow. They further speculated that they may be relict englacial structures that had been transported along thrust planes in the basal ice, similar in composition and aspect to glaciotectonic moraines formed near the margins of polythermal glaciers (Bennett, 2001). This hypothesis was developed from an earlier explanation of ribbed moraine known as the 'shear and stack theory' (Bouchard, 1989), whereby glaciotectonic shearing and stacking of subglacial sediments results from compressive ice flow in topographic depressions, followed by basal melt out of englacial debris.

Related to these theories, Dyke and Morris (1988) and Dyke et al. (1992) describe ribbed moraines in the onset zone of a small ice stream on Prince of Wales Island (Canadian Arctic Archipelago) that clearly extend into an area of inferred cold-based ice. The ribbed moraines occur at the transition between the cold-based catchment and warm-based ice stream and Dyke et al. (1992) suggested that oscillations of the basal thermal regime between cold and warm-based created a patchwork of cold-based areas that led to acceleration and deceleration of ice velocity and attendant infolding and stacking of debris. The explanation by Dyke et al. (1992) differs from that of Bouchard (1989) and Alysworth and Shilts (1989a) in that they invoke compressional ice flow around cold-based ice. Hättestrand and Kleman (1999) also associate ribbed moraines with cold-based ice, but their theory of formation invokes thawing and extensional fracturing of a pre-existing frozen till sheet which reverts to warm-based ice. Their fracturing theory was developed from the idea that individual ribbed moraine crests can be compressed and be shown to fit together like a jig-saw puzzle (see Hättestrand and Kleman, 1999) but more recent analysis by Dunlop and Clark (2006) questions the ubiquity of this observation.

In this study, we present new insights regarding the subsurface structure and sedimentology of ribbed moraines which helps shed light on their possible origin. The lower unit at our exposed section (unit 1) is interpreted as a subglacial till that was deposited initially by lodgement and then underwent relatively minor post depositional deformation. The two prominent silt-filled fractures (extension cracks or tension fractures) near the centre of unit 1 are interpreted to have resulted from an episode of brittle deformation. The genesis of unit 1 may therefore be interpreted as follows. Ice advances into the site, and till is deposited by lodgement. As the till thickened, and as pore water pressure in the till matrix increased, the lodgement till was deformed (stones rotated, mainly about their c-axes, but their a-axes remained, for the most part, parallel to ice flow direction). The existence of stone a-axis fabrics that are parallel to the ice flow direction suggest that the ice (and the deforming till) was undergoing extensional flow at this stage. At some point the pore water pressure dropped and the till stiffened enough for the two prominent fractures to develop, which dip

in the direction of ice flow; the other smaller fractures at the top of the unit probably developed at that time as well. The presence of stones adjacent to the prominent fractures that are oriented with their *a*-axis transverse to the inferred ice flow direction suggests that the ice was undergoing compressional flow during this phase. Stone *a*-axis fabric transverse to the ice flow direction has been associated with compressive flow in till in several other studies (e.g., Boulton 1971; Stanford and Mickelson, 1985; Hicock, 1990, Hart, 1994; Lian and Hicock, 2000). Dreimanis (1988) also notes that that the presence of a transverse fabric mode in lodgement tills can be associated with folding and shearing of the till matrix (a consequence of compression) and that it can be used as a criterion for its identification. A change in ice flow direction as the cause for this transverse fabric is discounted because the youngest striae on these stones remain parallel to the dip direction of the fractures. It is likely that it was also during this phase that some of the main NW-SE trending *a*-axis fabrics modes recorded at most of the fabric sites became attenuated in a roughly N-S direction. This attenuation is most obvious in the fabric recorded at site A, where a weak (spread-out) mode can be seen plunging to the south. For the fabric recorded at site B this is less obvious, but there seems to be attenuation of the fabric into the SE quadrant. It is also possible that the rotation of a significant number stone *a*-axes into a position transverse to the ice flow direction is responsible for the girdle-like shape of the fabric recoded at site D.

It is likely that the phases of extensional and compressional flow inferred from the glacial geology at this section are correlative with the two phases of glacial landform development nearby. That is, the phase of extensional flow is consistent with the existence of the glacial lineations and ice stream activity, whilst the phase of compressional flow is consistent with the later formation of the ribbed moraine. The normally consolidated nature of unit 2, together with the presence of crude subhorizontal bedding, suggests that the sediments in this unit were likely produced by mass movement of unit 2 material into Thelon River basin

Our interpretation of the exposed sediments also sheds light on the origin of the radar facies identified in the GPR surveys, which provide constraints on the internal structure of ribbed

moraines at length scales of ~1-100 m. We suggest that the internal architecture of facies 'B' (weak, chaotic reflectors) is consistent with our interpretation of sediment Unit 1 as a deformed till, possibly derived/eroded from pre-existing fluvioglacial deposits (radar facies 'C') that were subsequently over-ridden during ice stream advance. Moreover, the three-dimensional geometry of the bodies containing facies 'B' implied by our GPR data is consistent with origin of the ribbed moraine ridges under compressional flow, i.e. pre-existing glacial sediments are stacked by thrusting to form the ridges (Figures 4 and 6). The fact that facies 'B' also looks lenticular in transverse direction (Figure 5) may be due to the fact that this facies used to be incorporated in the glacial lineations and is now being eroded from them through this subglacial thrusting process. We speculate that facies 'B' is likely formed by subglacial erosion of facies 'C' and deformation of the eroded material, first in the process of formation of glacial lineations and then in the process of 'recycling' facies 'B' from these lineations and into the ribbed moraine ridges. The latter process was associated with low-angle thrusting, indicating compressional subglacial and basal stresses. The incorporation of facies 'C' into facies 'B' during landform development might also explain why the interface between these two facies is not always clear-cut.

In summary, our observations of the sub-surface sediments and structures indicate a switch from subglacial (ex)tension to subglacial compression that accompanied a change from glacial lineations to ribbed moraine formation. This interpretation is also consistent with the conjecture that the change from glacial lineations to ribbed moraines implies ice stream stoppage. It is well accepted that fast-flowing ice streams are associated with horizontal extension (at least in ice) and the example of the UpC ice bulge in West Antarctica's Ice Stream C indicates that this stress state switches to compression after stoppage (Vogel et al., 2005).

The final question, therefore, is what triggered the switch from fast (extensional) ice flow to slow (compressional) ice flow that effectively shut down the ice stream? Our observations of prominent fractures and thrusting appear to suggest a stiffening of the till that previously contributed to the motion of the ice stream. One way that till could become stiffer would be to drain water from it, and

direct observation of contemporary ice stream beds using boreholes suggests that basal freezing is an effective mechanism to accomplish this process (Vogel et al., 2005). If this was the case at the Dubawnt Lake Ice Stream bed, the presence of ribbed moraines reveals the locations that underwent basal freezing, prior to warm-based deglaciation. A similar situation is also envisaged by Andreassen et al. (2004) who identified shear and thrust structures associated with mega-blocks and rafts of sediment on the former bed of the Bear Island Ice Stream in the Barents Sea. They suggest that the ice stream underwent periods of basal freezing and that frozen sediment rafts were entrained by the ice stream and transported along thrust planes to a higher position in the ice.

Alternatively, the Dubawnt Lake ice stream till could have been dewatered as a result of a switch in the basal drainage system from a distributed system of broad, shallow canals to a more efficient system of much larger low density R-channels. This switch is predicted by theory (Walder, 1982) and has also been invoked as a shut-down mechanism for Ice Stream C (Retzlaff and Bentley, 1993). Although we favour a mechanism of till dewatering as the cause of the switch in ice dynamics under the Dubawnt Lake Ice Stream, our observations are unable to discriminate the precise cause. We do note, however, the apparently strong association between ribbed moraines and areas of cold-based ice reported by other workers (e.g. Dyke et al. 1992; Hättestrand and Kleman, 1999).

Conclusions

This paper presents detailed observations of an intriguing landform assemblage of ribbed moraines superimposed on glacial lineations on the Dubawnt Lake Ice Stream bed. Mapping from satellite remote sensing reveals a landscape dominated by glacial lineations that range from around 400 to 5500 m in length, 60 to 350 m in width, and exhibit elongation ratios from 3:1 up to almost 40:1. In places, these glacial lineations are overprinted by ribbed moraines whose lengths (i.e. transverse to ice flow) range from around 190 m to 1100 m and whose widths range from around 90 m to 300 m.

The ribbed moraines are typically around 10-15 m higher than the surrounding terrain and their mean crest to crest wavelength (spacing) is around 270 m but ranges from around 120 to 500 m. The orientation of both sets of landforms suggests that they were formed by ice flow from the same direction but overlying eskers suggest a subtle change in flow direction after ribbed moraine formation but prior to deglaciation.

Based on our interpretation of GPR surveys of the sub-surface structure of the ribbed moraines and of sediments and glacioteconic structures exposed in a section ~100 m from the landforms, we suggest that they resulted from an abrupt change in ice dynamics, and that this may have accompanied/lead to ice stream shut-down. Evidence for this comes from a deformed till which subsequently underwent brittle deformation. Glaciotectonic structures indicative of this brittle deformation are observed in the sediment exposure and in the GPR surveys where, in the latter, they occur most prominently in the ribbed moraine ridge crests. The presence of tension fractures in the till and their association with stones that are orientated with their *a*-axes transverse to the inferred ice flow direction suggests a compressional flow regime during the final phase of till deformation.

We therefore hypothesise that the glacial lineations were formed under an extensional flow regime during the ice stream activity. At some point, patches of till under the ice stream stiffened (pore water drainage caused by basal freeze-on, or some other process) which resulted in a compressional flow regime that in turn resulted in subglacial thrusting and the building of the ribbed moraines. This sequence of inferred events fits with observations of contemporary ice streams which also show a switch from extensional to compressional flow regimes before and after ice stream stoppage (Vogel et al., 2005).

Recently, the imaging of active ice stream beds has improved such that we are beginning to identify individual subglacial bedforms (King et al., 2003; Smith et al., 2007). Over the next few decades, it is likely that technology will advance to a level where it is possible to view whole suites of landforms under the ice. The increasing body of work on palaeo-ice stream beds provides an excellent foundation to compare and interpret these landform assemblages. In this paper, we show

that the presence of superimposed ribbed moraines likely reflects an abrupt change in ice dynamics from an extensional to compressional flow regime on the Dubawnt Lake ice stream bed. We therefore predict that ribbed moraines may form in areas of compressional flow under ice streams (i.e. sticky spots) and might be preserved on palaeo-ice stream beds that shut down through the progressive development of such areas, prior to deglaciation. The general absence of ribbed moraines on most palaeo-ice stream beds might suggest that either they operated during deglaciation (i.e. without shutting down) or, processes other than the development of localised compressional flow (sticky spots) led to their shut-down (i.e. ice depletion).

Acknowledgements

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Figure Captions:
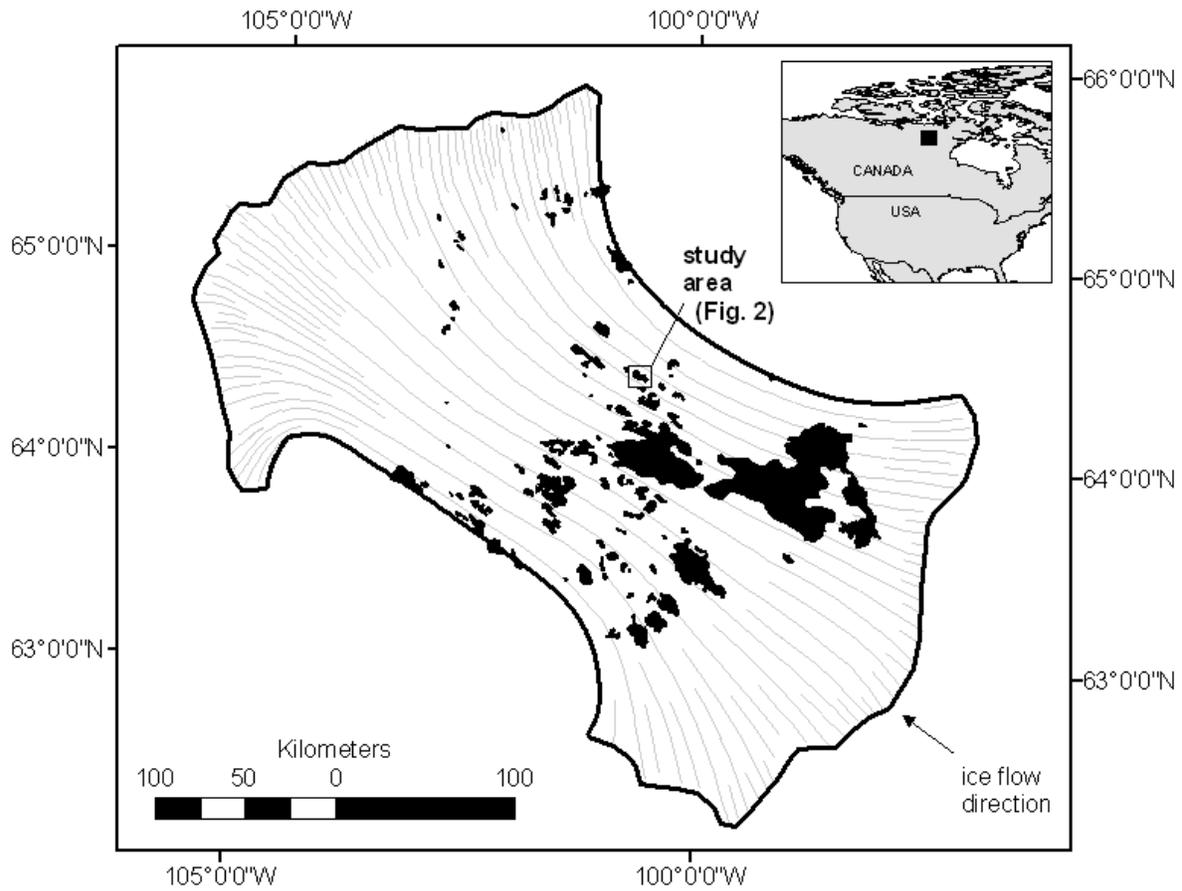
Figure 1: Spatial extent (thick black line) and flow-lines of the Dubawnt Lake Ice Stream imprint on the north-western Canadian Shield (see inset for location) showing the spatial extent of superimposed ribbed moraines (black polygons) and the location of the study area in this paper.
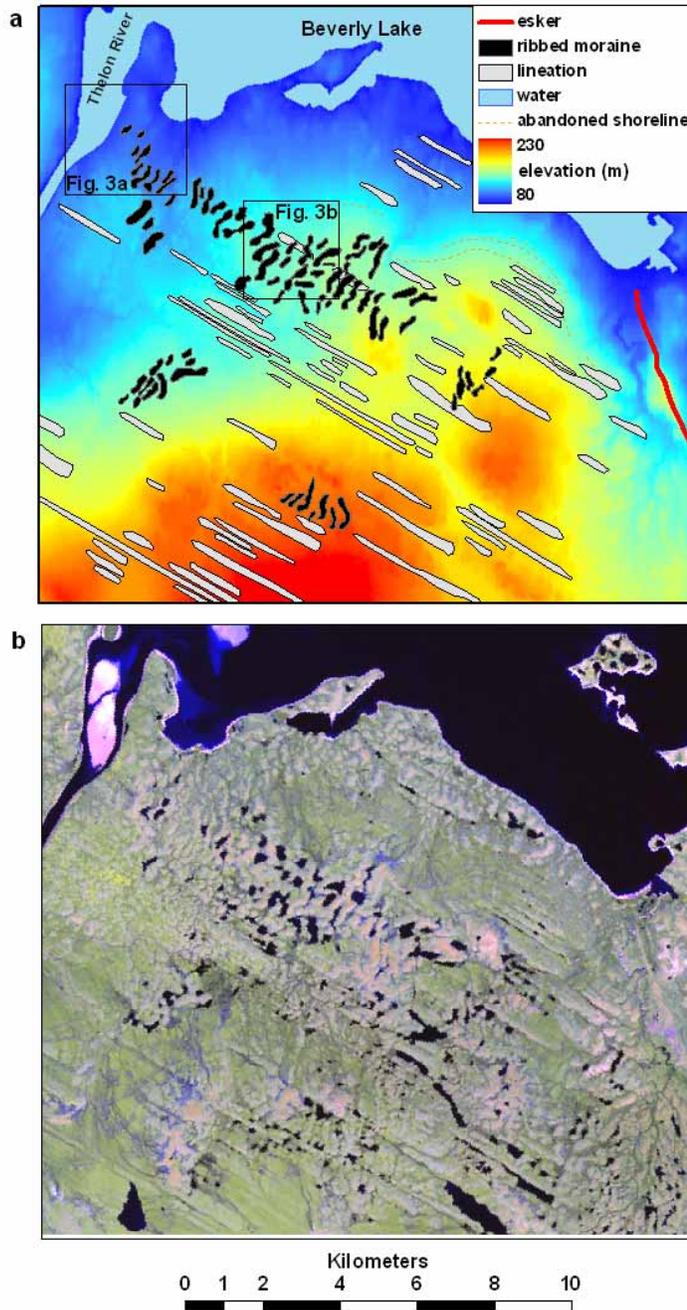
Figure 2: (a) Glacial landforms in the study area overlaid on a digital elevation model. (b) Landsat ETM+ image (R,G,B: 7, 5, 2) of the study area showing vegetated areas as green and surficial sediments/less vegetated areas as pink (water is shown as black). Note that the ribbed moraines predominantly appear as unvegetated surfaces between lakes/bogs in the centre of the image.
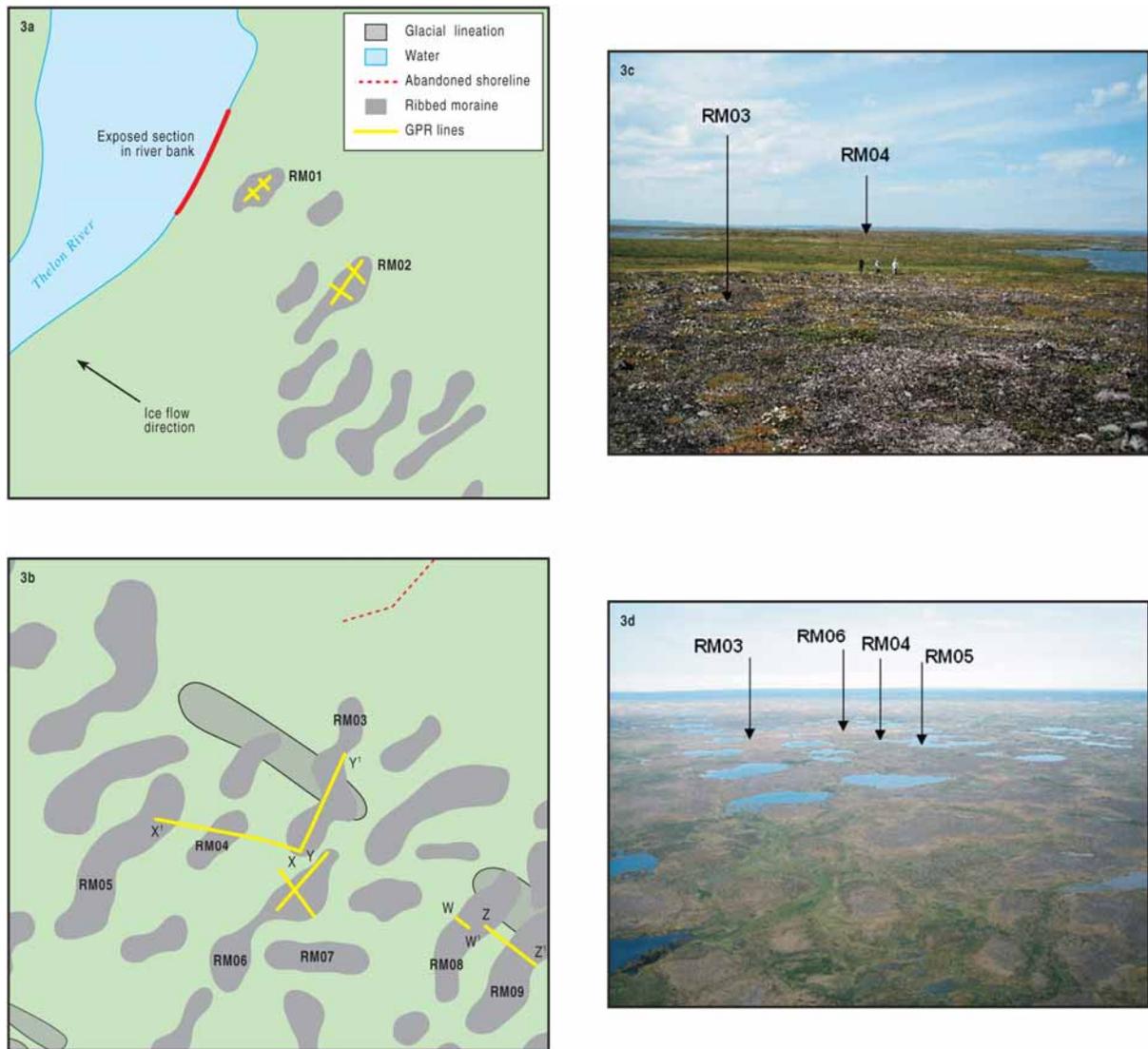
Figure 3: Detailed maps of ribbed moraines showing the location of the GPR transect lines on 9 individual ribbed moraine ridges (ridges RM01-09) in (a) and (b), and the location of the section shown in Figure 9. Note the close proximity of the section in (a), whose face is oriented transverse to the ice flow direction. Photograph in (c) is taken from ribbed moraine ridge 03 (ridge RM03) and looks towards crest of ridge RM04. Oblique aerial photograph in (d) shows spatial arrangements of ribbed moraine ridges (ridges RM03-06).
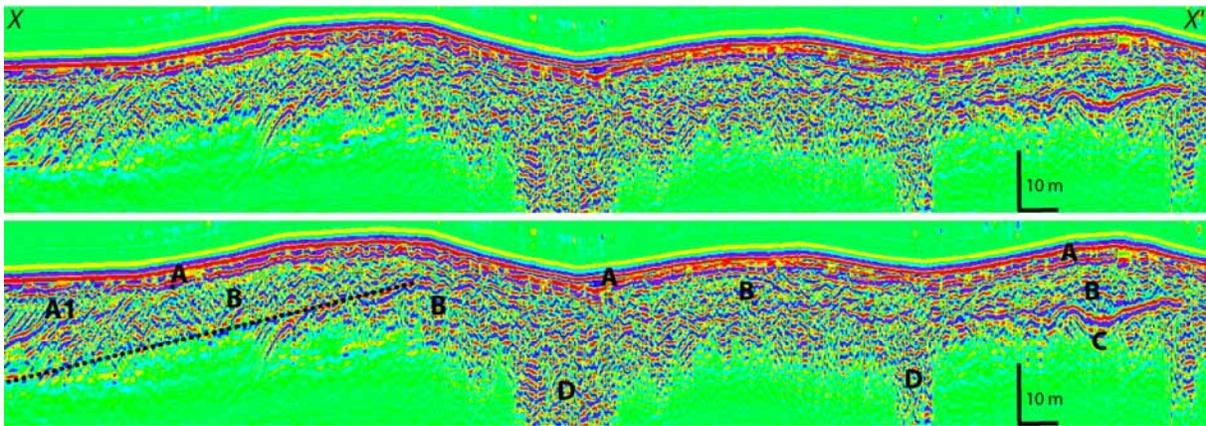
Figure 4: Radar section (50 MHz) through ribbed moraine ridges RM03, RM04 and RM05 (see Figure 3b for location). Ice flow was from left to right (NW). These radar data have been processed using a 'dewow' function and Automatic Gain Control. Vertical and horizontal scale given by black bars (10 m long in both directions). Letters designate individual radar facies (see text for details). Dotted lines give the locations of proposed thrust fault.
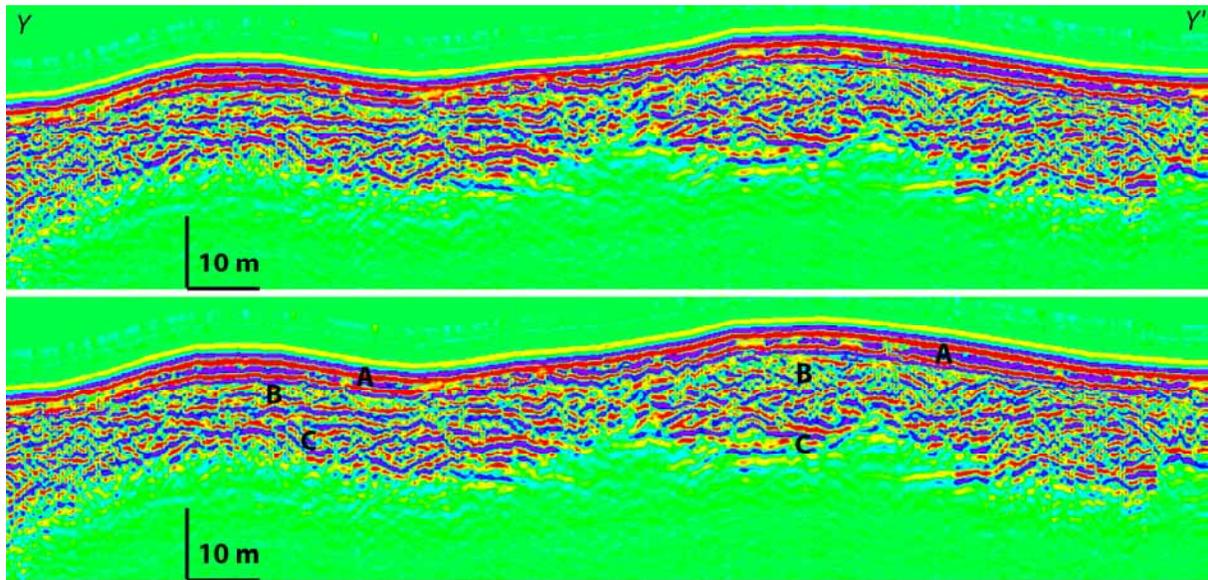
Figure 5: Radar section (50 MHz) through ribbed moraine ridge RM03 running approximately at right angle to the radar section shown in Figure 4 (see Figure 3b for location). Both sections (here and in Figure 4) share the left corner. Ice flow was into the plane of this figure. These radar data have been processed using a 'dewow' function and Automatic Gain Control.
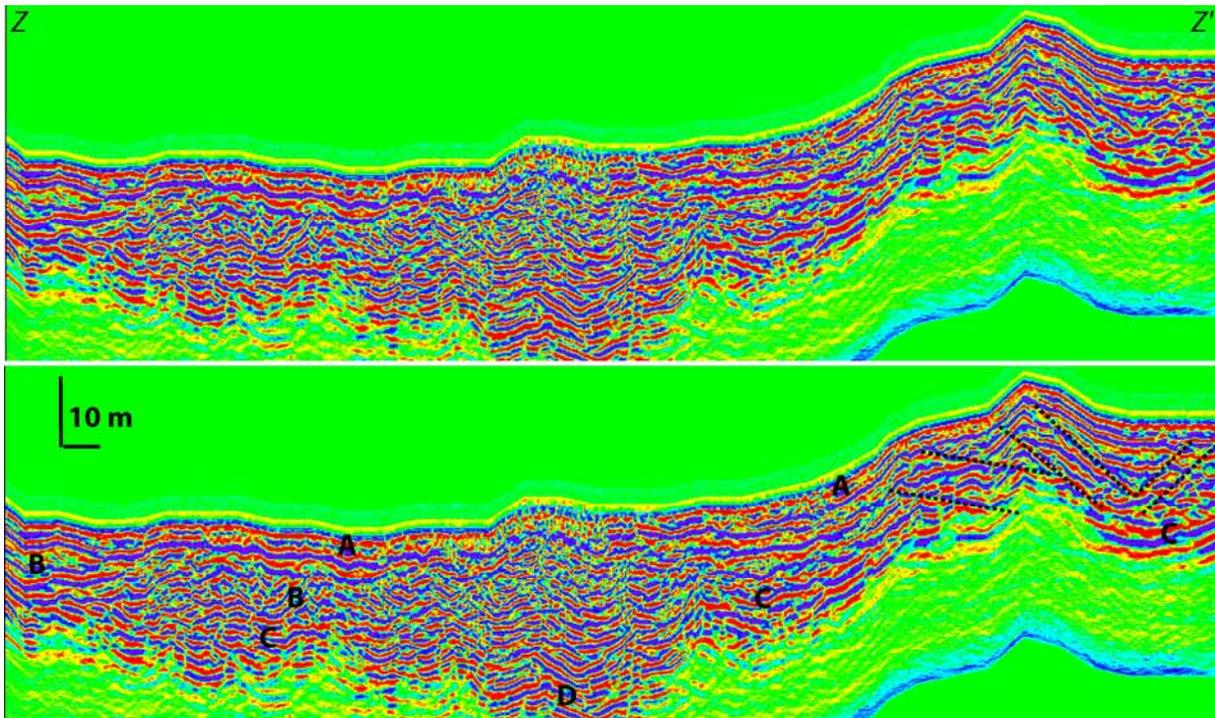
Figure 6: Image showing the GPR reflection survey (50 MHz) running (from right to left) across ribbed moraine ridge RM09, an inter-morainal depression, and onto the back slope of ridge RM08 (see Figure 3 for location). Ice flow is from right to left (NW). The data used in this figure has been 'dewowed', subjected to Automatic Gain Control, and a migration procedure using subsurface velocity of 0.14 m/ns.
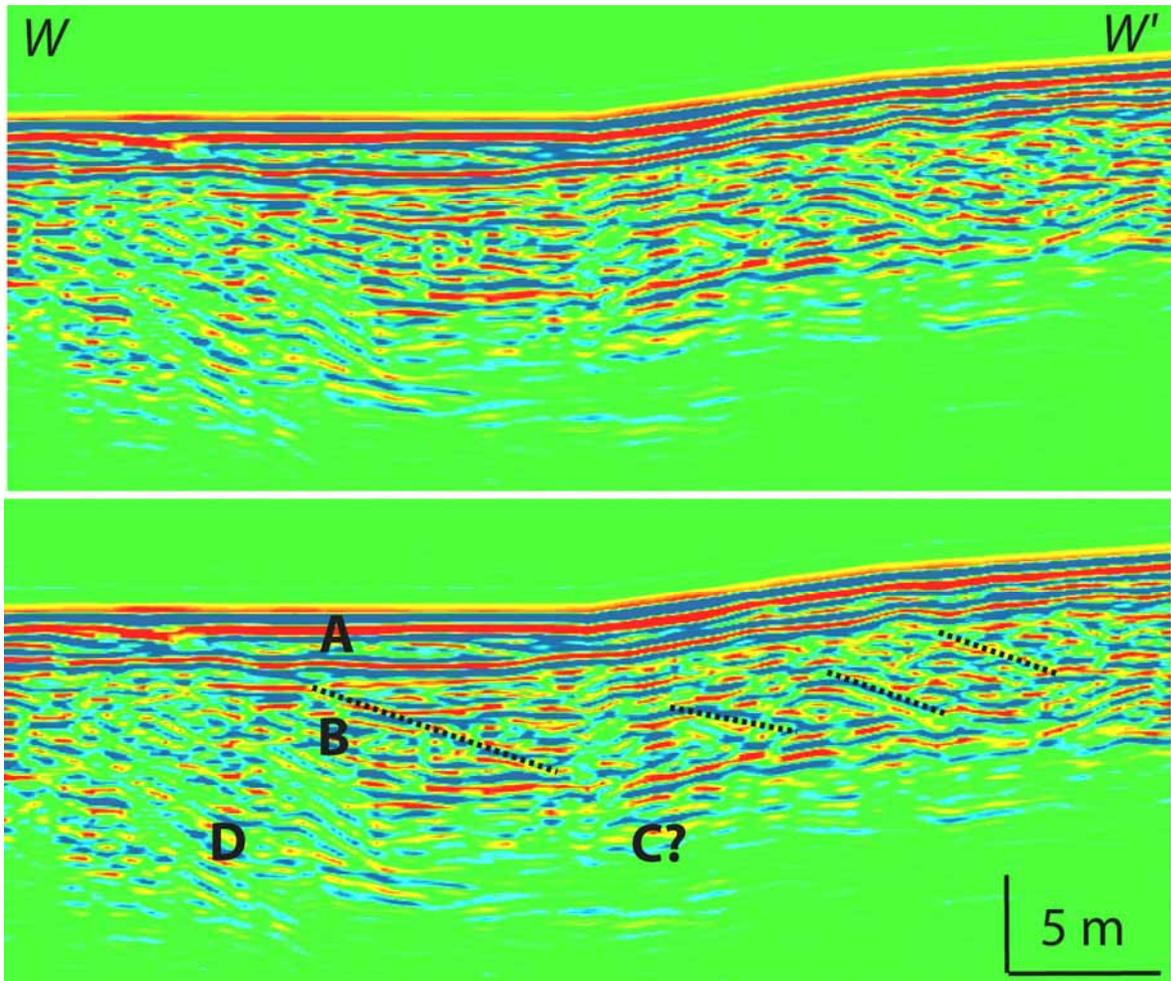
Figure 7: This section shows a line run a few metres to the left of the section presented in Figure 6 (the down-ice slope of ribbed moraine ridge RM08) using 100 MHz antennas, see Figure 3b for location. Data processing as in Figure 6. Ice flow is from right to left (NW).

Figure 8: Photograph showing 6-7 m of crudely stratified fluvioglacial sediments (pebbles/gravel/sand) below 2-3 m of sandy red diamicton containing striated and faceted clasts. The contact between the two units is clearly identifiable in the uppermost central part of the exposure. This section is exposed close to a mega scale glacial lineation along the Finnie River at 102° 25' 20" W, 64° 02' 27" N which lies in the main trunk of the ice stream. The upper red sandy diamicton is regionally extensive along the Finnie River and the lower Thelon River within the main ice stream trunk, including in our study area in this paper. We tentatively suggests that the strong (sub)horizontal facies 'C' in the radar survey could possibly correlate to the stratified sand and gravel unit observed here, and elsewhere on the ice stream bed.
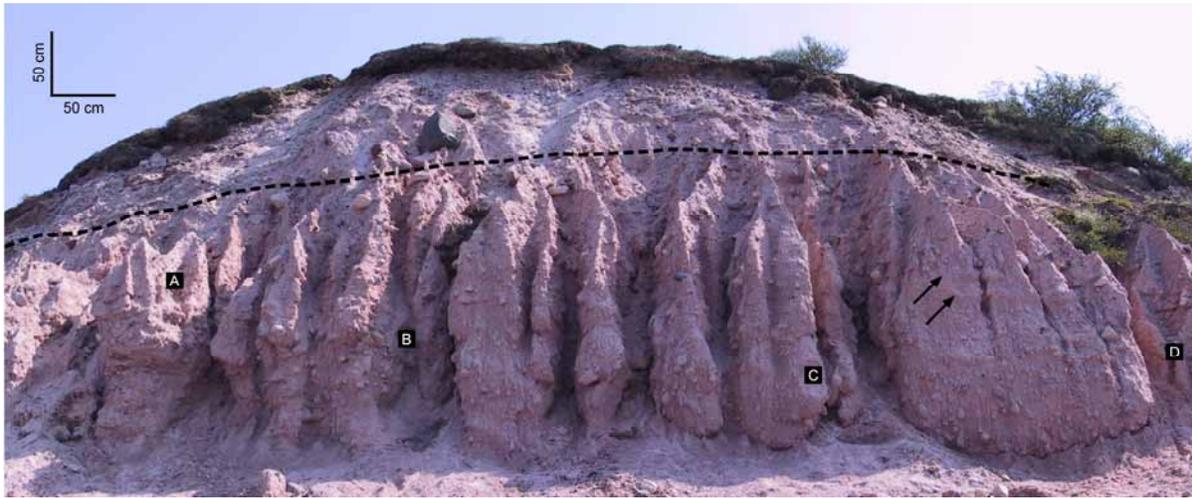
Figure 9: Photograph of the Thelon river bank section showing the locations where the stone *a*-axis fabrics (A, B, C and D), and associated data, were collected, see also Figure 3. The position of the two prominent infilled fractures is indicated by arrows. The exposure of unit 1 is 4–5 m high and consists of highly consolidated, matrix-supported diamicton interpreted to be deformed lodgement till, which is sharply overlain (contact at dashed black line) by approximately 3 m of normally consolidated, crudely bedded, sand and gravel (unit 2), interpreted to be a mass-movement deposit. The subtle stone pavement is not discernible at this scale. Note that the photograph is an oblique view and the 50 cm scale bars apply to the bottom centre of the photograph.

Figure 10: Photograph of one of the bullet-shaped stones in the discontinuous boulder pavement; a knife blade, ~2.5 cm wide, is shown for scale. The *a*-axis of this stone, and those of the other stones in the pavement, are oriented perpendicular to the ice flow direction inferred from the landforms, the dip of the tension fractures, and the trend of the main fabric modes at sites A–D. Note, however, that the fine/freshest (i.e. youngest) striae set on the upper surface of this stone are oriented parallel to ice flow direction (two striae from this set are indicated by white arrows), which weathered (older) striae can be seen oriented parallel to the stone a-axis.
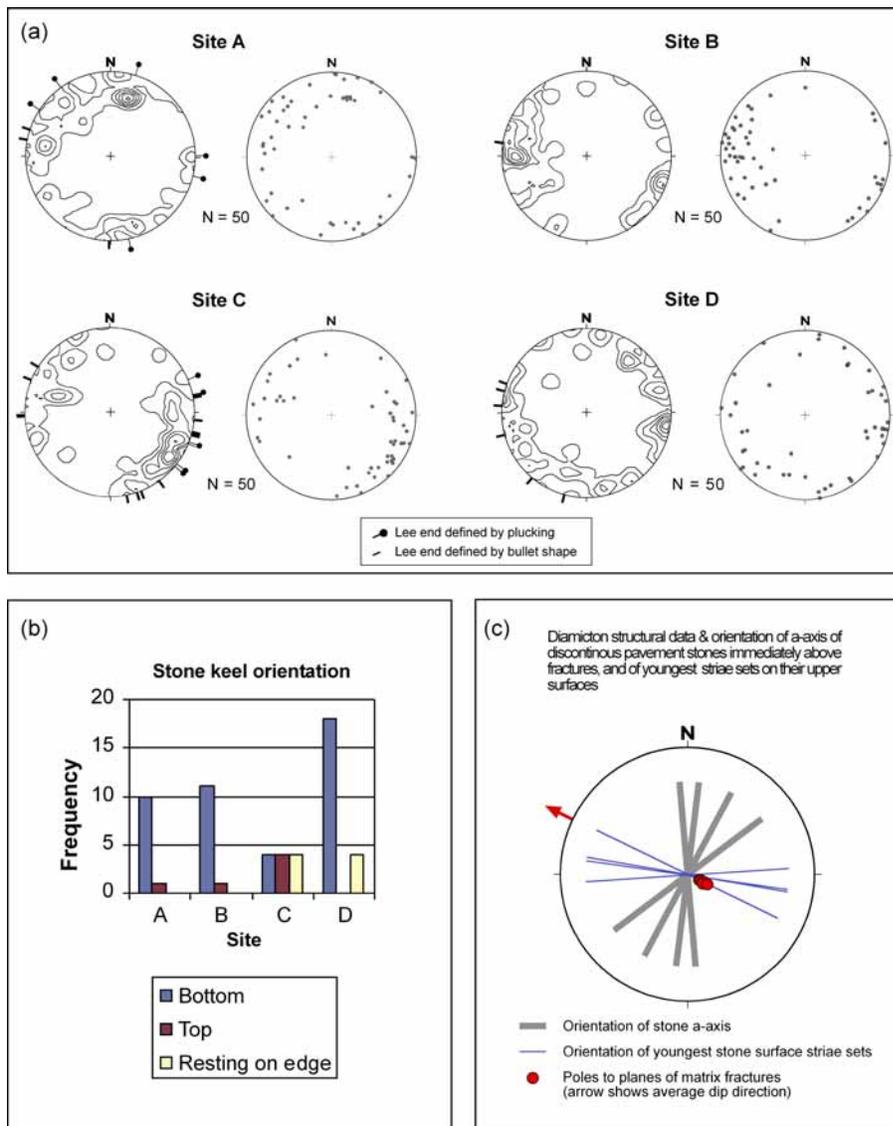
Figure 11: (A) Three-dimensional stone *a*-axis fabric data (scatter and contour plots) collected at the four sample locations (A, B, C, and D) shown in Figure 9. The plots are lower hemisphere Schmidt projections. The trends of stone lee ends are plotted as line segments on the circumference of each Schmidt diagram. (B) Frequency of position of stone keels at the locations of the *a*-axis fabric measurements. (C) Orientation of *a*-axes, and youngest striae, on upper surfaces for stones in a discontinuous pavement near the prominent fractures. Also shown is the strike and dip of the tension fractures (plotted as poles to planes); two measurements were made of each fracture at different positions laterally along the section.