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The Dubawnt Lake palaeo-ice stream: evidence for dynamic ice sheet behaviour on the Canadian Shield and insights regarding the controls on ice-stream location and vigour

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We report evidence for a major ice stream that operated over the northwestern Canadian Shield in the Keewatin Sector of the Laurentide Ice Sheet during the last deglaciation 9000–8200 (uncalibrated) yr BP. It is reconstructed at 450 km in length, 140 km in width, and had an estimated catchment area of 190 000 km^2. Mapping from satellite imagery reveals a suite of bedforms ('flow-set') characterized by a highly convergent onset zone, abrupt lateral margins, and where flow was presumed to have been fastest, a remarkably coherent pattern of mega-scale glacial lineations with lengths approaching 13 km and elongation ratios in excess of 40:1. Spatial variations in bedform elongation within the flow-set match the expected velocity field of a terrestrial ice stream. The flow pattern does not appear to be steered by topography and its location on the hard bedrock of the Canadian Shield is surprising. A soft sedimentary basin may have influenced ice-stream activity by lubricating the bed over the downstream crystalline bedrock, but it is unlikely that it operated over a pervasively deforming till layer. The location of the ice stream challenges the view that they only arise in deep bedrock troughs or over thick deposits of 'soft' fine-grained sediments. We speculate that fast ice flow may have been triggered when a steep ice sheet surface gradient with high driving stresses contacted a proglacial lake. An increase in velocity through calving could have propagated fast ice flow upstream (in the vicinity of the Keewatin Ice Divide) through a series of thermomechanical feedback mechanisms. It exerted a considerable impact on the Laurentide Ice Sheet, forcing the demise of one of the last major ice centres.

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Ice streams are rapidly flowing corridors of ice bordered by comparatively stagnant sheet flow. They are large features (>20 km wide and >150 km long) that dominate ice-sheet discharge and exert a major influence on ice-sheet configuration (Oppenheimer 1998). 'Topographic' ice streams flow through deep bedrock troughs, but 'pure' ice streams are not constrained by the underlying topography. Although these definitions are clear in principle, some ice streams may display characteristics of both along their length (Bentley 1987). Of particular interest is the idea that pure ice-stream flow may switch on and off and is often out of equilibrium with ice-sheet mass balance (Anandakrishnan & Alley 1997). This has fuelled the idea that pure ice streams may exhibit unsteady behaviour and has re-emphasized their role in determining ice-sheet stability. An understanding of pure ice streams is central to predicting the response of contemporary ice sheets to future climate change (see Joughin & Tulaczyk 2002) and yet their enigmatic nature pose two main questions: (i) what controls their locations within an ice sheet and (ii) what mechanisms facilitate their rapid velocity?

Ice streams also played a key role in the behaviour of the former northern hemisphere ice sheets and may have been instrumental in driving abrupt climate change at the end of the last glacial cycle (Hughes 1992; Bond *et al*. 1992). Investigating former ice streams holds great potential for advancing our understanding of ice-stream behaviour and their impact on ice-sheet configuration. Moreover, if we can confidently find a palaeo-ice-stream bed, we have an opportunity to glean information regarding their basal conditions and processes, including the spatial (geologic, topographic?) controls on their location and vigour (Stokes & Clark 2001). Obtaining similar data from contemporary ice streams is compromised by the inaccessibility of their basal environment.

In this article, we report outstanding evidence for a major terrestrial ice stream which operated in the Keewatin Sector of the former North American (Laurentide) Ice Sheet, north of Dubawnt Lake, Nunavut Territory, Arctic Canada. The significance of this ice stream is that: (a) it represents one of the most robust terrestrial ice-stream imprints yet identified, and (b) it operated in a surprising location on the Canadian Shield – a region often assumed to be incompatible for fast ice flow and dynamic ice-sheet behaviour because of the generally hard crystalline bedrock (see Clark 1992, 1994).

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Identifying former ice streams

Many workers have attempted to locate ice streams from a variety of former ice sheets (reviewed in Stokes & Clark 2001). Unfortunately, hypothesized locations have tended to outweigh meaningful evidence, due in part to a limited understanding of the landform assemblages they leave behind (Matthews 1991; Stokes & Clark 1999, 2001). As a first attempt to overcome this problem, Stokes & Clark (1999) constructed a theoretical glacial land system that an ice stream might be expected to produce. This land system incorporated several 'geomorphological criteria', predicted from the known characteristics of contemporary ice streams. These include (among others) bedform patterns which fit the characteristic shape and dimension of contemporary ice streams, display highly convergent onset zones, exhibit abrupt lateral margins and contain elongated subglacial bedforms such as mega-scale glacial lineations. Individually, none of the criteria are necessarily indicative of palaeo-ice-stream activity, but when several are found within the same glacial land system they may represent substantive evidence (cf. Stokes & Clark 1999; Clark & Stokes in press).

The Dubawnt Lake flow pattern: previous work

Many investigators have noted a distinct flow pattern trending in a northwesterly direction, north and east of Dubawnt Lake, Nunavut Territory, Arctic Canada (Bird 1953; Craig 1964; Aylsworth & Shilts 1989a–c; Boulton & Clark 1990). Figure 1 shows the location of this flow pattern, also represented on the Glacial Map of Canada (see Prest *et al*. 1968).

In their reconstruction of the Laurentide Ice Sheet, Boulton & Clark (1990) attributed the 'flow-set' to a late glacial event (*c*. 10 000 yr BP) relating it to a southeast shift in the position of the Keewatin Ice Divide. Kleman & Borgström (1996) interpreted the landform assemblage as a 'type-landscape' for a 'surge fan' in their glaciological inversion model for reconstructing former ice sheets. Surge fans exhibit a characteristic bottleneck pattern and subglacial lineaments are presumed to have been generated rapidly during deglaciation.

Notable characteristics of the Dubawnt Lake flow-set compared to adjacent flow-sets in the region are the highly elongated subglacial bedforms (Bird 1953; Craig 1964; Aylsworth & Shilts 1989a, b). This 'spectacular fluting' led Aylsworth & Shilts (1989a) to speculate on the possible role of ice streaming in shaping the lineaments, but they refrained from postulating an exact location.

The similarity between this flow pattern and an idealized palaeo-ice-stream imprint (Stokes & Clark 1999; Clark & Stokes in press) suggests that it might represent a suitable candidate for a major terrestrial ice stream and this article aims to demonstrate this. We extend previous work and (i) document evidence of ice-stream activity; (ii) describe and characterize the ice-stream bed; (iii) assess its impact on the Keewatin Sector of the Laurentide Ice Sheet; and (iv) speculate concerning the controls on its location and vigour.

Methodology and data sources

Fifteen Landsat Multi-Spectral Scanner (MSS) hard copy positives (band 5) were developed into photographs, providing complete coverage of the study area (Fig. 1) at a scale of 1:250 000. These images were used to map the regional ice-flow patterns shown in Fig. 1.

To investigate the Dubawnt Lake flow-set in more detail and assess its validity as an ice-stream track, three adjoining Landsat MSS images (180 km by 180 km, spatial resolution 80 m) were obtained in digital format (bands 1, 2, 3 and 4). This allowed a more refined mapping approach, enabling bedforms to be identified at a range of scales. Image processing techniques were used (where necessary) to preferentially highlight topography and increase detection of bedforms.

On-screen mapping was accomplished by digitizing a line along the main axis of each subglacial bedform (lineament) on the Landsat MSS imagery. The following measurements were taken and stored as ARC/INFO coverages in a Geographic Information System (GIS): lineament length, width (to calculate elongation ratio: length/width), orientation, parallel conformity (standard deviation of a sample of lineament orientations), density (number of bedforms per unit area) and packing (surface area of bedforms per unit area).

To analyse internal variations in bedform morphometry within the flow-set, three flow corridors were constructed and gridded at 20-km intervals in the centre of the flow-set but allowing for flow convergence and divergence up-ice and down-ice. Unfortunately, the coverage of the digital imagery did not extend to the northern margin of the flow-set. However, the flow-set is superimposed on older flow patterns of a similar orientation in these areas and precise measurement would have proved difficult (see Fig. 1). The three flow corridors closest to the southern lateral margin provide a largely uncomplicated pattern.

Additionally, one Landsat Enhanced Thematic Mapper Plus image (ETM+: spatial resolution up to 15 m) was obtained from the central region of the flow-set. For consistency, only lineaments identified on the MSS imagery were mapped. However, the ETM+ imagery provided detailed information from an important area and permitted a first order check on the accuracy of the mapping from the Landsat MSS imagery, which in most cases was entirely complementary.

A 30-arc second (*c*. 0.5 km) Digital Elevation Model (DEM) was obtained to visualize the regional topo-

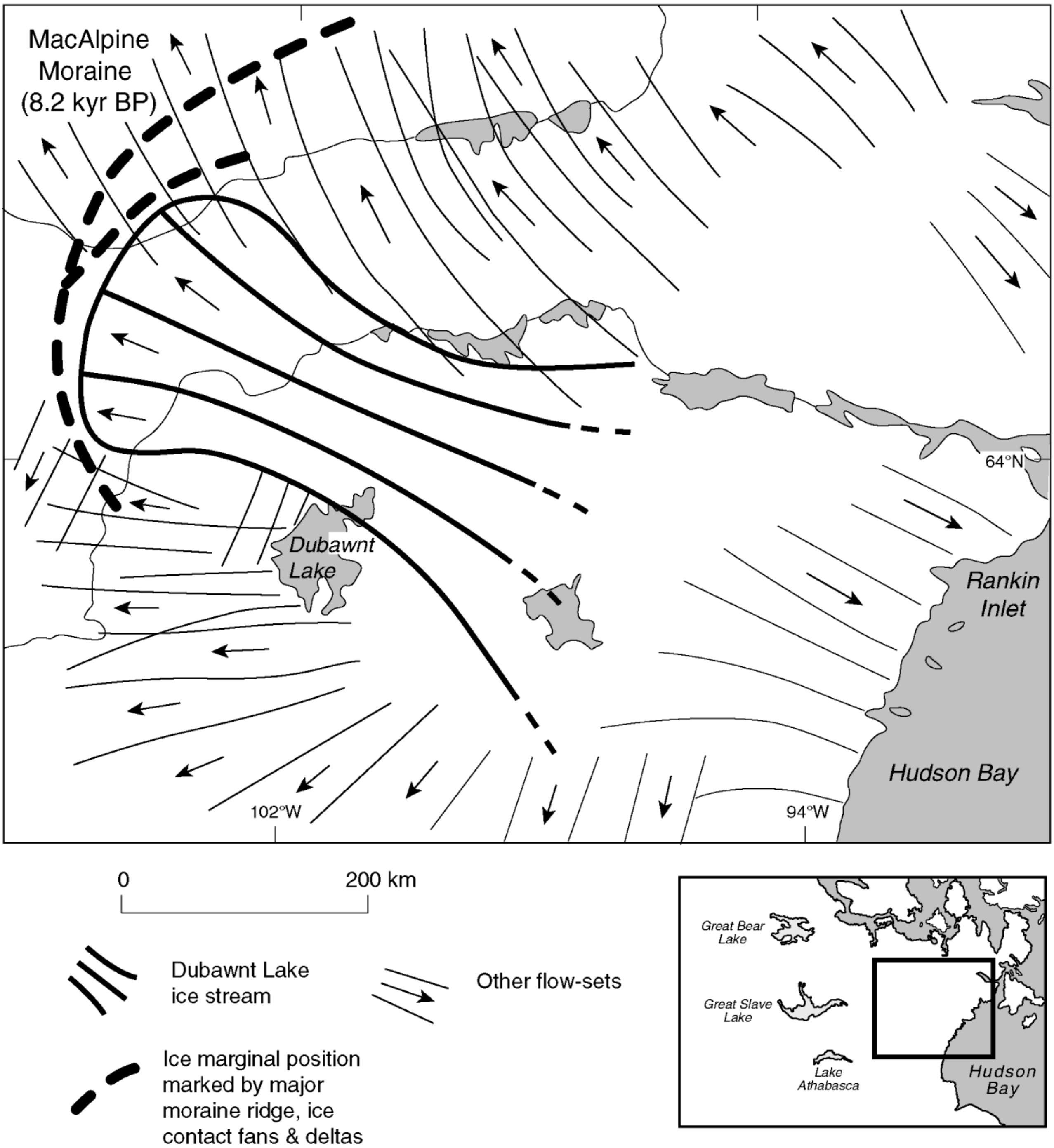


Fig. 1. Study area centred on Keewatin, Nunavut Territory, Arctic Canada, showing the regional ice-flow patterns and the location of the Dubawnt Lake flow pattern with respect to the MacAlpine moraine system. Modified from Aylsworth & Shilts (1989a, b) and Kleman & Borgström (1996).

Fig. 2. The Dubawnt Lake flow-set. Each lineament records the long axis of a drumlin or mega-scale glacial lineation indicating former ice-flow direction. Mapping from Landsat MSS imagery was concentrated on the southern half of the flow-set because of cross-cutting bedforms in the north producing a complication (see text). The flow-set is gridded at 20 km intervals in the centre of the flow-set but these broaden up-ice and down-ice such that each grid square contains a sample representative of the palaeo-flow line immediately up-ice or down-ice. Note the position of the three flow corridors 'A', 'B' and 'C'.

graphic influence on the ice-sheet flow patterns. This also allowed elevation transects to be calculated but more detailed elevation data were taken directly from topographic maps (1:250 000) with a contour interval of 20 m. Information regarding the sedimentary characteristics of the area was taken from published sources, the most comprehensive of which are provided by Aylsworth & Shilts (1989b, c).

Results

Landform assemblage of the Dubawnt Lake flow-set

Subglacial lineaments. – The digital imagery covered an area in excess of 91 000 km^2 and 11 825 lineaments were digitised, of which 8856 were from the Dubawnt Lake flow-set. Figure 2 shows all of the lineaments detected on the digital Landsat MSS imagery, the limits of the Dubawnt Lake flow-set and the 20 km^2 grid employed to sample the bedforms, including the three flow corridors 'A', 'B' and 'C'.

The flow-set is characterized by bedforms which

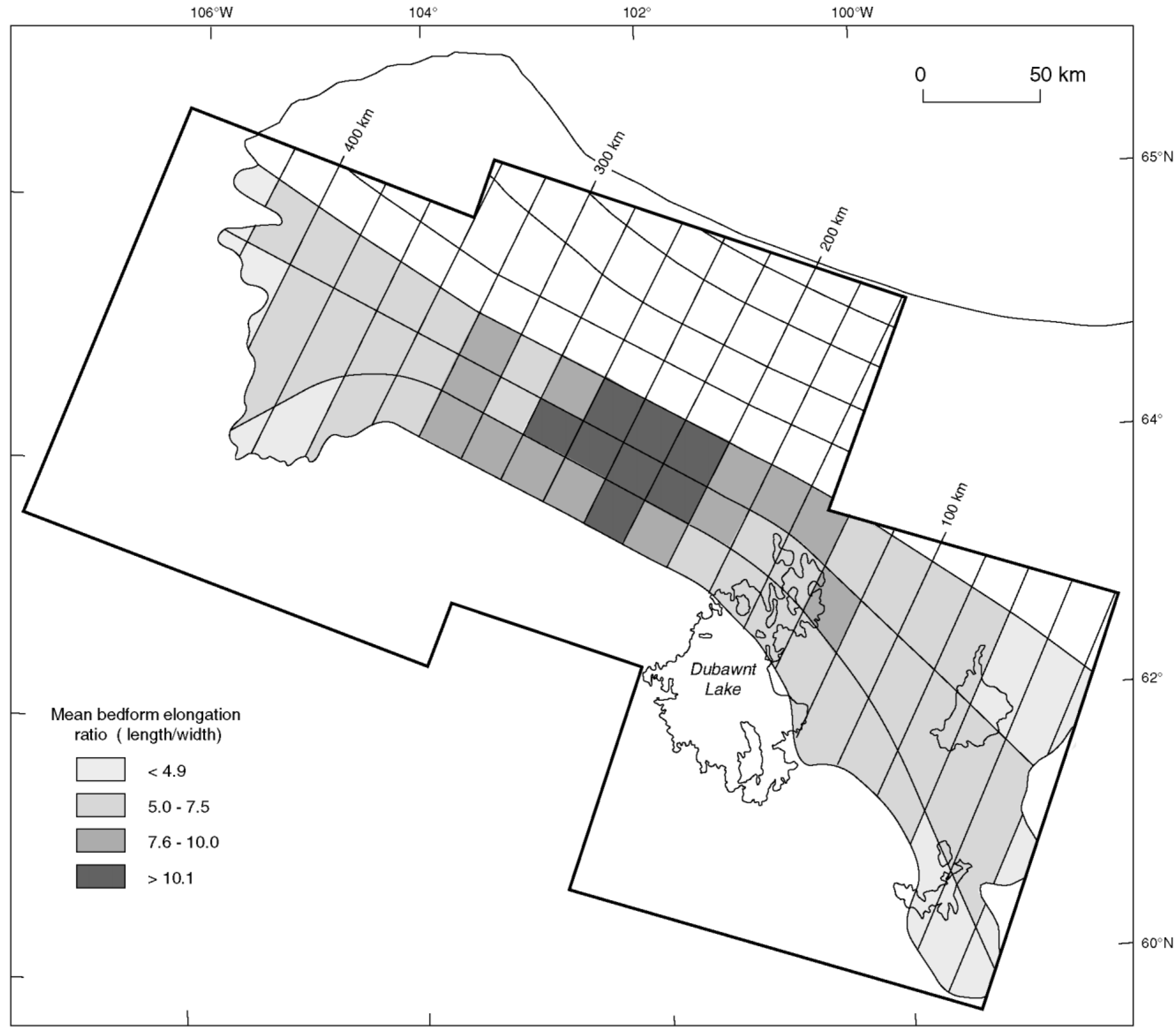


Fig. 3. Variations in bedform elongation ratio along the southern half of the Dubawnt Lake flow-set. Although there is a degree of 'patchiness', elongation ratios match the expected velocity pattern whereby ice speeds up in the onset zone, reaches a maximum in the main trunk and decreases towards the divergent terminus (from Stokes & Clark 2002).

are considerably longer (mean 1808 m, maximum 12 743 m), wider (mean 256 m, maximum 990 m) and display higher elongation ratios (mean 7:1, maximum 48:1) than other flow-sets in the region (Stokes 2000). It exhibits a remarkably coherent pattern (Fig. 2) with systematic downstream variations in bedform elongation ratios that match the expected pattern of ice velocity variations within the bottleneck flow pattern (Stokes & Clark 2002). Bedform elongation ratios increase in the convergent onset zone, reach a maximum in the narrowest part of the main trunk and thereafter decrease towards the lobate terminus. This is shown in Fig. 3, which depicts bedform elongation ratios across the southern half of the flow-set. Figure 4 illustrates variation of bedform lengths within the flow-set with the longest bedforms in the centre of the ice stream. Evidently, subglacial bedforms are longest and have highest elongation ratios approximately half way down the flow-set. In this region, lineament lengths consistently approach and occasionally exceed 10 km (maximum 12.7 km) with elongation ratios in excess of 40:1.

The lineaments in the main trunk (centre) also exhibit exceptional parallel conformity with neighbouring bedforms. Over an area of 720 km^2 in the centre of the flow-set, the standard deviation of lineament orientation does not exceed 3.8°. Figure 5 shows a Landsat ETM+ image of a sample of bedforms from the centre of the

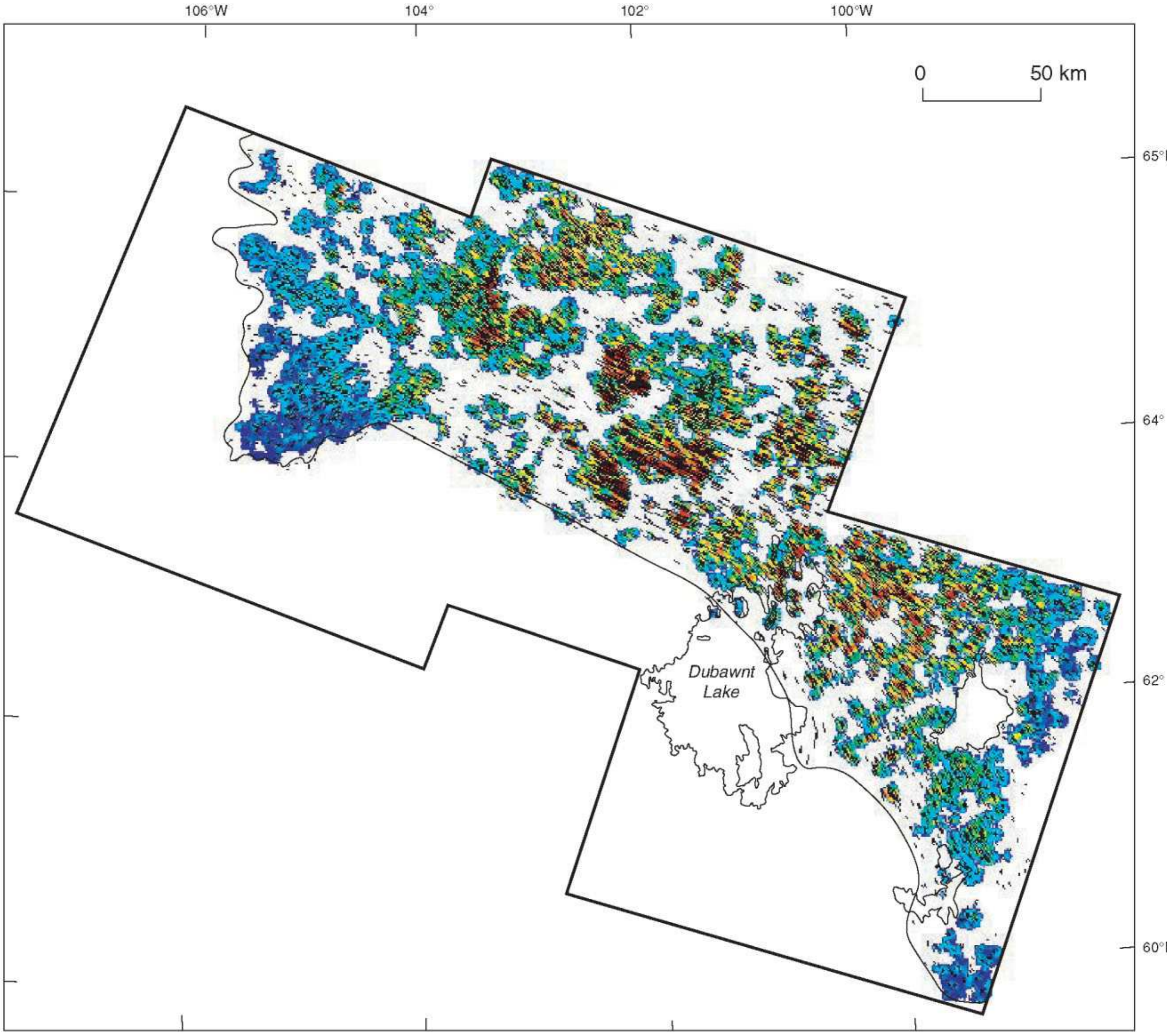


Fig. 4. Variation in bedform lengths across the Dubawnt Lake flow-set. The longest bedforms (red and orange) are most common approximately half way down the flow-set at its narrowest part, while shorter bedforms (blue and green) occur up-ice in the onset zone and down-ice towards the inferred terminus. This pattern closely matches the expected velocity field of ice experiencing convergence and then divergence. Note that interpolations of lineament length were restricted to data-rich areas to avoid interpolation artefacts. This explains the blank patches.

flow-set, approximately 200 km down-ice in flow corridor 'B' (location shown in Fig. 2). The great length and exceptional parallel conformity of the bedforms gives the appearance of a ridge/groove structure. This is in contrast to drumlins in the onset zone, which are shorter, wider, and have a much more rounded appearance (see (D) in Fig. 6).

Ribbed ('Rogen') moraines. – Ribbed moraines are ubiquitous throughout the study area and have been mapped in detail by Aylsworth & Shilts (1989a–c), who note their close association around the inferred Keewatin Ice Divide. The ridges trend transverse to ice flow and are typically less than 2 km long, around 150–300 m wide, and generally around 10 m high (Aylsworth & Shilts 1989a). Their small dimensions hinder individual ridge identification on the satellite imagery but large (>25 km^2) patches could be identified and their spatial relationship with the lineaments was noted. Only on the Landsat ETM+ imagery could individual ridge crests be identified with confidence. Figure 6 (location shown in Fig. 2) is a subset of this imagery showing the flow-set lineaments in close association with the ribbed moraines.

Ribbed moraines are most abundant up-ice towards the onset zone, where they cover around 50% of the

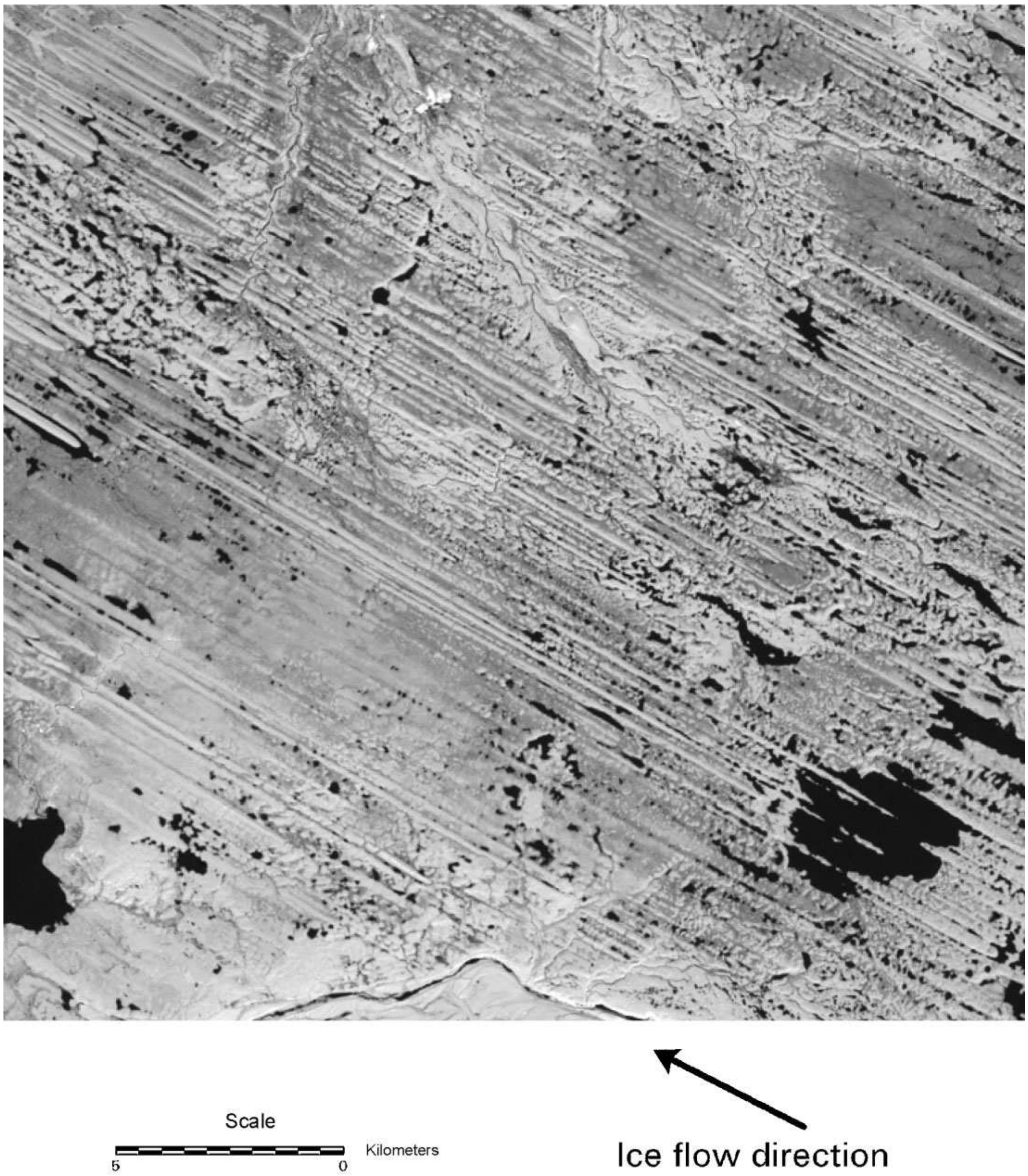


Fig. 5. Landsat ETM+ satellite image (band 5) of elongated drumlins and mega-scale glacial lineations from the centre of the Dubawnt Lake flow-set. The exceptional parallel conformity of the bedforms and high elongation ratio gives the appearance of a ridge groove structure. Image centred on 101″50 W, 64″05 N (northwest of Dubawnt Lake), approximate width is 22 km. For image location, see Fig. 2.

flow-set area (Fig. 6). Further down-ice their occurrence is scarcer but they still occur in discrete patches up to 320 km down-ice. In all cases, ribbed moraines lie superimposed on the lineaments and in some cases lineaments have been partially broken up by their formation; see Fig. 6 (cf. Aylsworth & Shilts 1989b). This is an extremely rare occurrence, because drumlins normally lie superimposed on ribbed moraines (cf. Hättestrand & Kleman 1999). Although they were not mapped in detail, they represent a major component of the land system and we comment on their possible significance in the discussion.

Eskers. – Eskers and other glaciofluvial features are common throughout the northwestern Canadian Shield forming a dendritic pattern radiating from the last inferred position of the Keewatin Ice Divide (cf. Aylsworth & Shilts 1989a–c). Throughout the study area, several large esker systems (10s km) are found superimposed on the streamlined bedforms. Towards

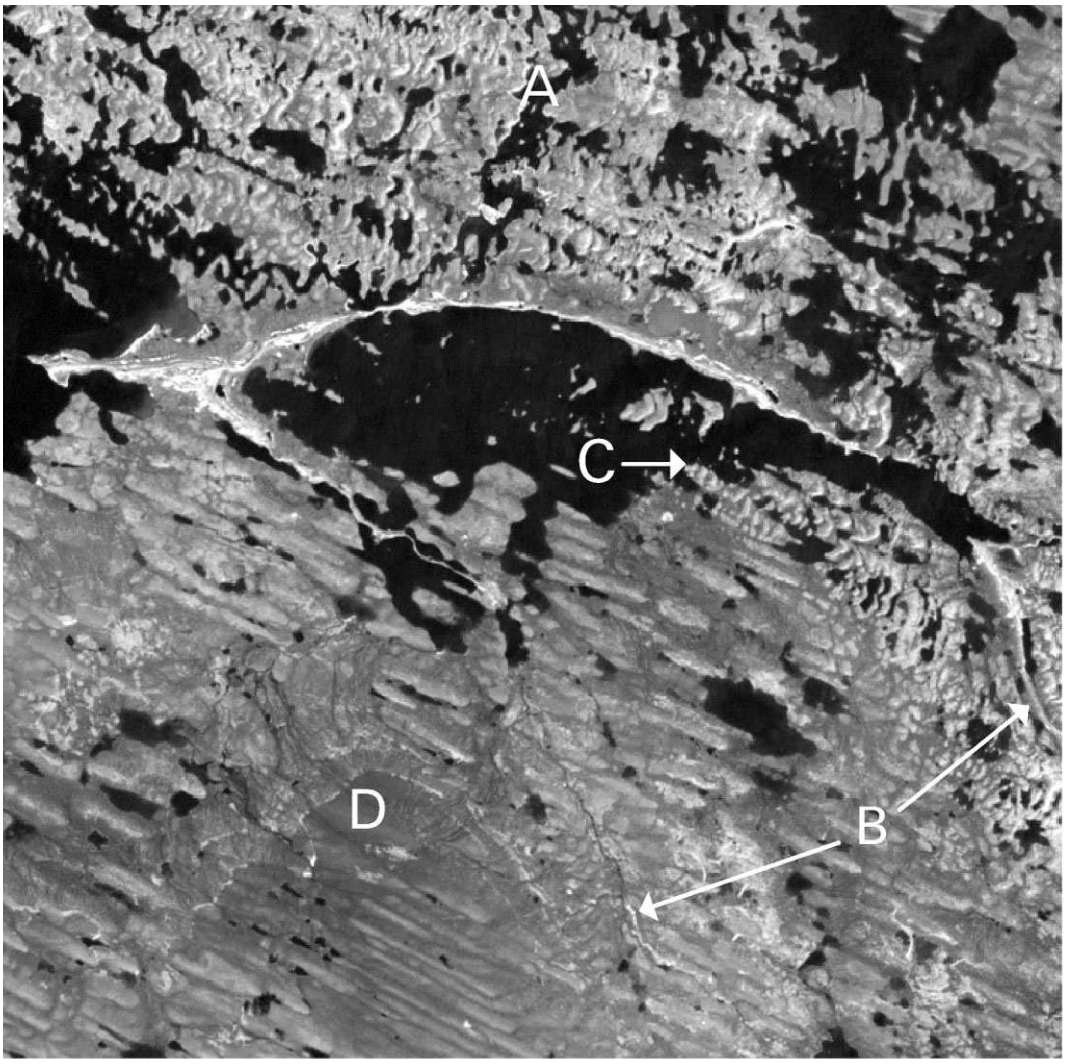


Fig. 6. Ribbed moraines (A) superimposed on the main flow-set lineaments (D). Reorganization of the lineaments into ribbed moraines is apparent and in places the lineaments have been partially broken up (C). Also note a large esker system (B) that cross-cuts the main ice-flow direction. Ice flow to the northwest. Landsat ETM+ subset image (bands 6, 3, 2) centred on 99″20 W, 63″55N. Image width approximately 22 km, location shown in Fig. 2.

the distal part of the flow-set, eskers are aligned roughly parallel to the streamlined bedforms but up-ice, towards the inferred Keewatin Ice Divide, some eskers obliquely cross-cut the pattern of ice flow (cf. Ayslworth & Shilts 1989a–c; Kleman & Borgström 1996). Figure 6 shows an esker system cross-cutting patches of both the lineaments and ribbed moraines in the onset zone of the flow-set.

Ice marginal features. – The distal part of the Dubawnt Lake flow-set occurs in close association with a discontinuous end moraine system composed of glaciofluvial deposits, including ice contact deltas and outwash fans; see Fig. 1 (cf. Aylsworth & Shilts 1989b, c; Kleman & Borgström 1996). Some of these outwash deposits could be identified on the digital imagery from their spectral signature. Although major end moraine systems are rare on the Canadian Shield, a complex of ridges (the MacAlpine moraine system) coincides with the glaciofluvial deposits (Blake 1963, 1966; Craig 1965; Falconer *et al*. 1965a, b). The till behind the MacAlpine moraine (and the moraine itself) is characterized by sediments from the Thelon Sedimentary Basin up-ice (see below) but outside of the moraine these sediments are rare (W. W. Shilts, pers. comm. 2002). This landform assemblage and the moraine are taken to record a major recessional position of the Keewatin Ice Sheet which correlates closely with the inferred terminus of the Dubawnt Lake flow-set.

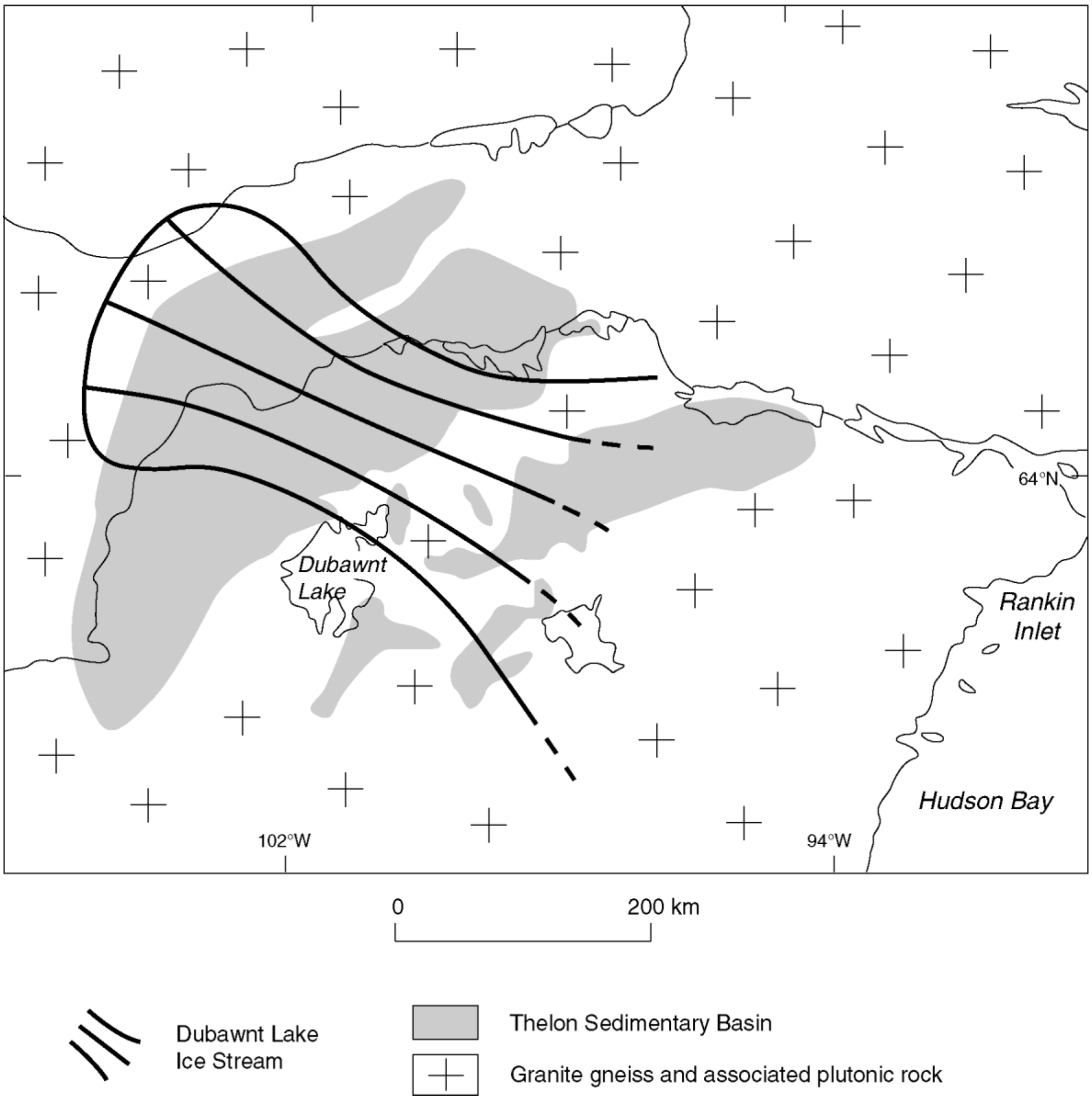


Fig. 7. Location of the Thelon Sedimentary Basin with the Dubawnt Lake flow-set superimposed. Although the softer rocks underlie a large part of the flow-set, there is no correspondence between the margins of the sedimentary basin and the margins of the flow-set. The sedimentary basin may have played a role in lubricating fast ice flow over the harder crystalline bedrock but we argue that it was not the primary control on the location of the ice stream within the ice sheet. Modified from Aylsworth & Shilts (1985).

Sedimentary characteristics

The study area is characteristic of the Canadian Shield, underlain by Precambrian bedrock, mainly massive granitoid gneisses. This bedrock produced a generally coarse grained, gravely and sandy till with a low content of clay-sized particles (Scott 1976). The widespread recrystallization related to metamorphism has produced rocks that are physically hard and yielded relatively little debris through glacial erosion (Aylsworth & Shilts 1989b). However, in places, plutonic and sedimentary strata outcrop. In the centre of the study area, the Thelon Sedimentary Basin is characterized by flat-lying unmetamorphosed sedimentary and volcagenic rocks (Fig. 7). This poorly consolidated outcrop yielded large volumes of disaggregated debris from the friable sandstones, which were subsequently dispersed down-ice over the adjacent crystalline terrain (cf. Aylsworth & Shilts 1989a, b). Compared to the crystalline bedrock, these dispersal trains cover large areas with till that is relatively thick and clay rich (W. W. Shilts, pers. comm. 2002).

Topography

Typical of the Canadian Shield, the topography of the study area is gently sloping, with no major valleys or escarpments. Figure 8 shows a Digital Elevation Model (DEM) of the study area with the location of the Dubawnt Lake flow-set superimposed. To investigate any topographic influence on the location of the flow-set, surface elevation transects were taken down-ice and orthogonal to the southern lateral margin. Figure 9 shows the down-ice changes in bed elevation along the southern lateral shear margin and 20, 40 and 60 km in-stream. A slight drop in elevation occurs between 200 and 300 km down-ice, coinciding with the location of the Thelon Sedimentary Basin. The surface profiles drawn orthogonal to the southern lateral margin are shown in Fig. 10 and it would appear that the margin coincides with a drop in elevation at some places. However, vertical exaggeration in Fig. 10 is extreme ($\times 40$) and an elevation change of 200 m over a distance of 50 km gives an average slope of only 0.004 degrees.

Interpretation of the Dubawnt Lake flow-set

Evidence for ice-stream activity

One of the most obvious clues to the existence of a palaeo-ice stream is the shape of the flow pattern it inscribed; in particular, highly convergent flow patterns (cf. Stokes & Clark 1999; Clark & Stokes in press). Almost without exception, contemporary ice streams are characterized by a large zone of flow convergence which feeds a main channel. This is exactly what is exhibited by the Dubawnt Lake flow-set, which has a convergent onset zone capturing ice from a large source area (Figs 1, 2).

Lineament mapping (Figs 2, 4) indicates that the Dubawnt Lake flow-set exhibits a remarkably abrupt southern margin, which can be traced for over 400 km from the onset zone to the terminus. Over a distance of as little as 1 km, elongated bedforms are bordered by areas with no bedforms, or bedforms with a different orientation (Fig. 2). It is suggested that the southern margin of the flow-set represents the abrupt lateral shear margin of an ice stream. It could be argued that the area outside the abrupt margin was ice-free, thus explaining the general absence of bedforms and the preservation of older flow patterns. We reject this hypothesis because eskers and ribbed moraines that are superimposed on the flow-set cross the ice-stream margin. This indicates that ice existed outside of the margin after the flow-set was produced (cf. Aylsworth & Shilts 1989c).

The northern part of the margin can be traced but with more difficulty because the flow-set cross-cuts bedforms of a similar orientation (Fig. 1). We expect that this margin is equally abrupt to that of the southern margin, but that imagery/aerial photographs with a finer resolution are required to confirm this.

Although there is no unequivocal method for reconstructing former flow velocities from geomorphological evidence, many studies report correlations between inferred (or modelled) fast ice flow and high bedform elongation ratios (e.g. Clark 1993; Hart 1999; Wellner *et al.*, 2001; Clark & Stokes 2001; Stokes & Clark 2002). In the case of the Dubawnt Lake flow-set, bedform lengths and elongation ratios from the centre of the ice stream are exceptionally high and we suggest that they can only have been formed by fast ice flow (cf. Stokes & Clark 2002). Down-ice variations reveal exactly what we would expect from a terrestrial ice stream whose velocity steadily increases in the onset zone, reaches a maximum in the main trunk, and then decreases as the ice flow diverges towards the terminus (Figs 3, 4).

Another indicator of fast ice flow relative to the adjacent portions of the ice sheet is the abrupt margin of the bedform pattern. To create such an abrupt margin requires a sharp discontinuity in either the basal thermal regime (warm/cold) or ice velocity, both of which are characteristic of ice-stream margins. Taken together, all these factors suggest that the Dubawnt Lake flow-set can be considered a terrestrial ice stream, here termed the 'Dubawnt Lake ice stream'.

Ice-stream dimensions

The ice stream is reconstructed as having a maximum length of 450 km with a width narrowing from ~300 km in the onset zone to around 140 km in the main trunk (Fig. 2). Towards the terminus, the ice stream widens to around 190 km, producing a splayed lobate pattern. We assume that velocities were below that considered as 'streaming' in this region. The surface area of the ice stream is calculated at around 72 000 km^2 with an estimated catchment area of 190 000 km^2. Length and estimated catchment area of the ice stream are comparable to contemporary ice streams in West Antarctica that feed the Ross Ice Shelf. For example, the estimated catchment areas of Ice Streams B and D are around 163 000 km^2 and 131 000 km^2, respectively (Rose 1979).

Isochronous or time-transgressive bedform record

We now consider whether the ice-stream lineaments were laid down by a single episode of ice flow, providing an isochronous ('snapshot') view of the bed immediately prior to shut down, or were produced time-transgressively, producing a 'smudged' imprint as the ice stream continually re-organized its imprint while operating during deglaciation (cf. Clark 1999; Stokes & Clark 2001).

If the flow-set were produced time-transgressively, it would be expected to display systematic discontinuities in the coherency of the pattern (Clark 1999). In contrast, we observe a remarkably coherent parallel pattern with relatively smooth down-ice changes in bedform morphometry (Figs 3, 4). In addition, it would be difficult to explain the abrupt southern margin of the ice stream if it

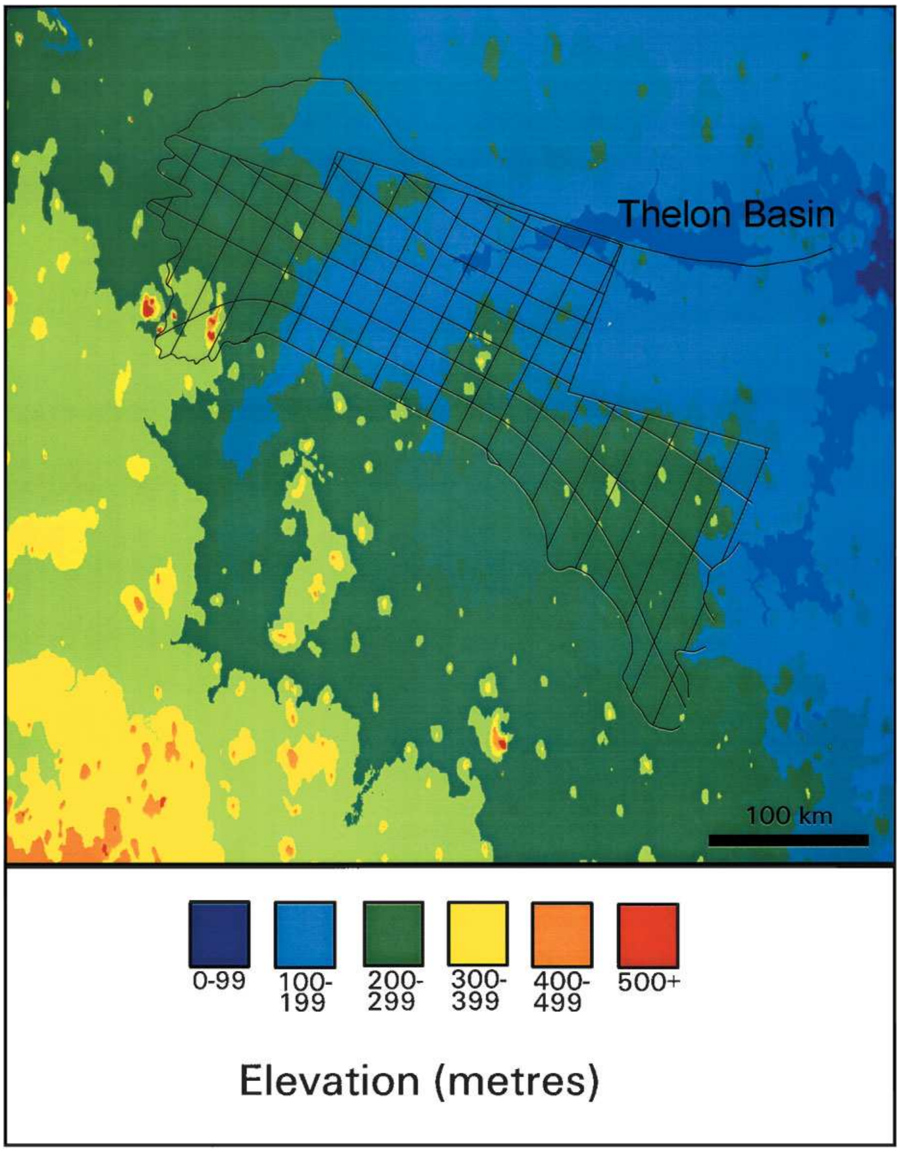


Fig. 8. The Dubawnt Lake flow-set in relation to topographic variation. Note that there is no correlation between the underlying topography and the location of the flow pattern.

were formed time-transgressively. We interpret the imprint as a snapshot view of the ice-stream bed immediately prior to shut-down which survived deglaciation without substantial modification. The retreat direction was not always parallel to the direction of ice-stream flow, as evidenced by eskers and outwash deposits that occur superimposed at oblique angles upon the well-preserved ice-stream landforms (Fig. 6).

Glacial history and deglaciation

The ice-stream flow-set occurs superimposed on all of the other flow-sets, indicating that it was the last major active ice flow through the area and was probably produced immediately prior to deglaciation (cf. Boulton & Clark 1990; Kleman & Borgström 1996). Dyke & Dredge (1989) provide a comprehensive reconstruction of the retreat of the Keewatin Ice Sheet sector based on an extensive review of the literature from the region, including dated ice marginal positions. From our focused mapping in a specific area, and without the benefit of additional dating constraints, it does not fall within the scope of this article to provide a full and revised reconstruction of this sector of the ice sheet. Rather, at this preliminary stage we simply attempt to bracket the ice-stream activity into a suggested time period.

The ice-stream terminus coincides with the MacAlpine moraine system (Falconer *et al*. 1965a, b). Blake (1963) acquired a ^{14}C date from shells that were recovered from the distal side of the moraine at an elevation of 183 m, within 16 m of the marine limit. An uncalibrated ^{14}C year age of 8160 ± 140 yr BP (GSC-110) was obtained and Blake (1963, 1966) suggested

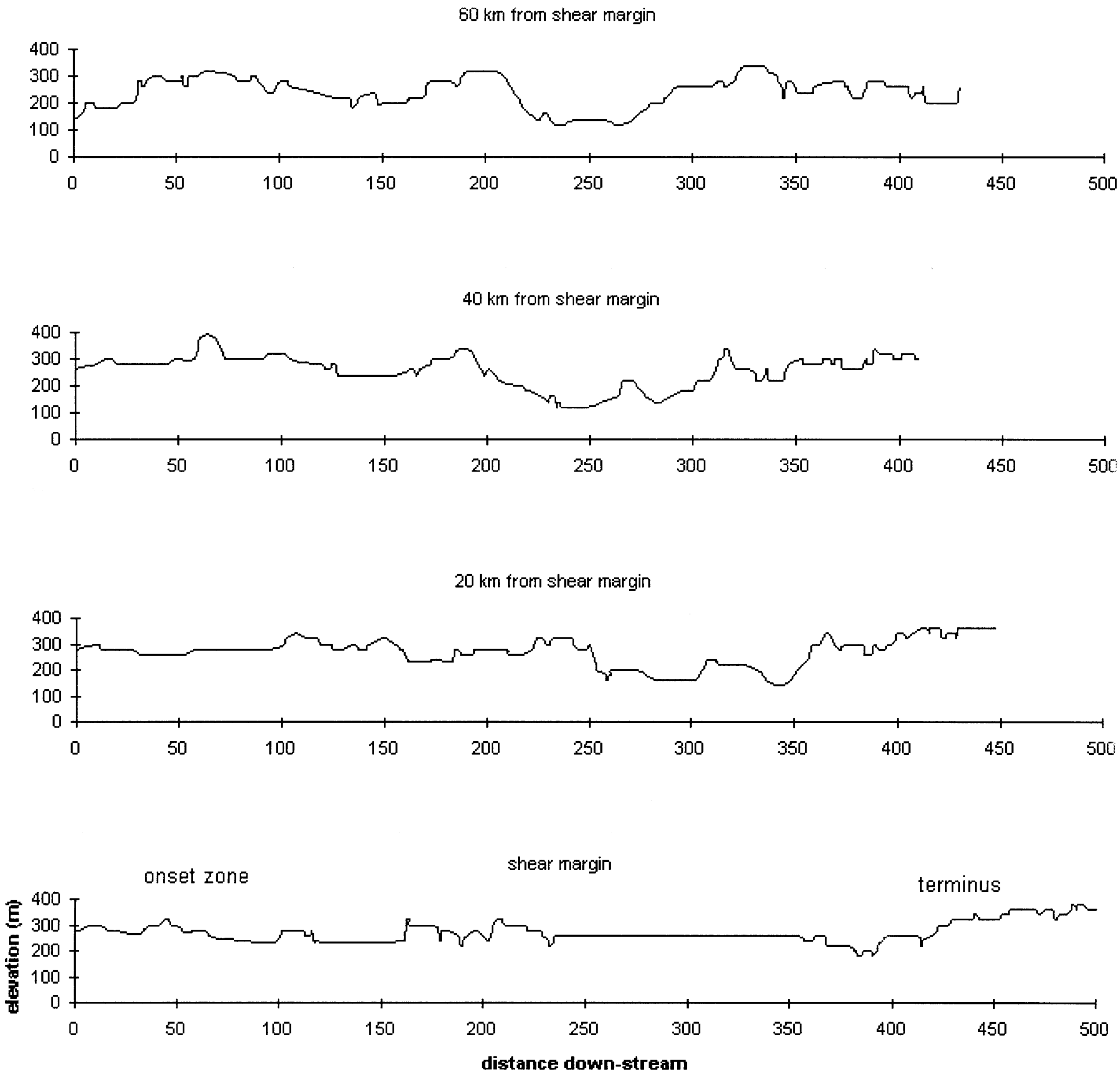
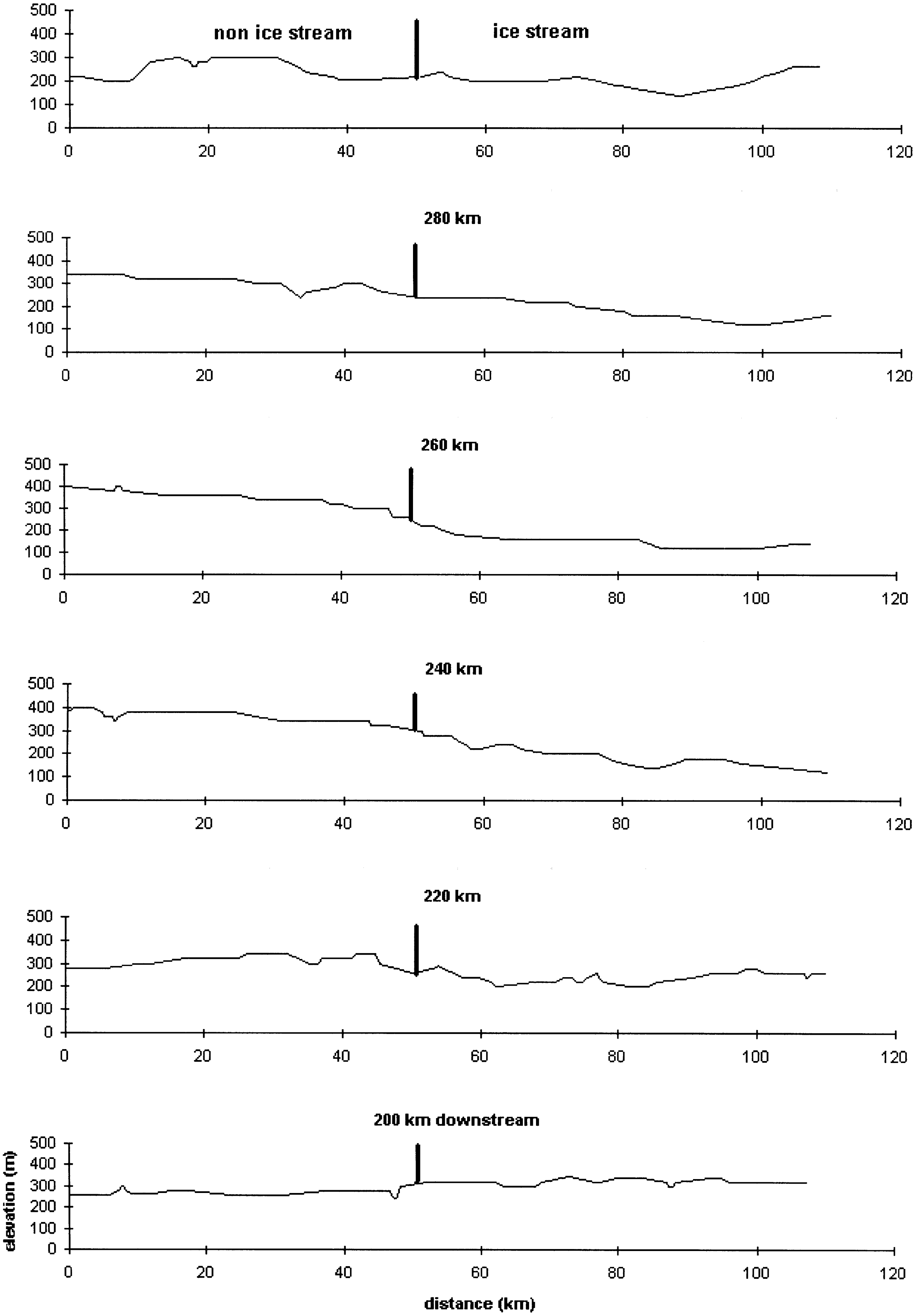


Fig. 9. Surface elevation transects down the inferred flow lines of the Dubawnt Lake flow-set. Note that there is relatively little elevation change in the down-ice direction. The Thelon Sedimenatary Basin occurs around 250 km downstream, coinciding with the lowest elevations on three of the transects away from the lateral shear margin.

that the ice margin rested against the proximal side of the moraine when the shells were deposited. This provides a minimum age for the ice-stream operation, which we tentatively bracket within the period 9000–8200 (uncalibrated) yr BP. This relates to the ice-stream imprint that we report in this article. The possibility of prior ice streaming at this location cannot be ruled out, although we note that widespread preservation of older flow patterns makes this unlikely.

Significantly, field investigations and aerial photographs have revealed nearly 2240 km of end moraines extending from eastern Keewatin (the MacAlpine

Fig. 10. Surface elevation changes orthogonal to the southern lateral shear margin of the Dubawnt Lake flow-set. In some places, the southern lateral shear margin appears to coincide with a slight drop in elevation but this would have been a negligible control on regional ice flow given the vertical exaggeration (×40).


non ice stream
ice stream
280 km
260 km
240 km
220 km
200 km downstream
elevation (m)
distance (km)
500
400
300
200
100
0
0
20
40
60
80
100
120

moraines) across Melville Peninsula and on to Baffin Island (Falconer *et al.* 1965a, b). Dating of these moraines demarcates the margins of the Laurentide Ice Sheet between 8000 and 9000 (uncalibrated) yr BP. This time period has been linked to a warmer climate with enhanced snowfall over the ice sheet from newly available moisture sources (Dyke 1984; Andrews 1985, 1989).

We speculate that as the margins of the Keewatin Sector retreated, enhanced snowfall on Keewatin ice may have led to a steep ice-sheet surface gradient with high driving stresses. In addition, as the western margin retreated, it led to the development of proglacial lakes impounded by the ice-sheet margin (cf. Craig 1964; Falconer *et al.* 1965a). Of most importance to this investigation is the development of a proglacial lake in the Thelon Basin (glacial Lake Thelon), thought to have formed late in the interval between 10 000 and 8400 (uncalibrated) yr BP (Craig 1964; Dyke & Dredge 1989). A steep ice-sheet profile with high driving stresses contacting this lake may have been a significant prerequisite for ice streaming. Although a proglacial lake in the Thelon Basin may have been important in triggering the ice stream, the splayed lobate terminus suggests that either the ice stream eventually advanced through the area formerly occupied by the lake, or the lake was drained or diverted prior to ice-stream shutdown. The effect of the ice stream was to thin and lower the ice-sheet profile below the equilibrium line, probably initiating widespread deglaciation and contributing to the rapid demise of the Keewatin Sector of the Laurentide Ice Sheet.

Surge event or ice-stream activity?

It could be argued that the Dubawnt Lake ice stream is not an ice stream but simply a surge of the ice-sheet margin. However, a terrestrial ice stream will nearly always manifest itself as a lobate advance of the ice-sheet margin. In this sense, a terrestrial ice stream is a surge event. Despite this, we refrain from calling it a surge because the term implies periodic or cyclical behaviour (cf. Meier & Post 1969). In addition, a glacier surge often refers to flow acceleration of the whole (or a large part) of the ice mass. In contrast, what we are dealing with here is a discrete zone of fast ice flow embedded within the ice sheet, i.e. an ice stream. This is evidenced by the abrupt lateral margins of the flow-set. We suggest that an 'ice-sheet surge' should be restricted to 'a dramatic flow acceleration and margin advance of the whole ice mass and which happens at regular cycles' (Stokes & Clark 2001).

Discussion: controls on ice-stream location and vigour

The Dubawnt Lake ice stream represents one of the most robust terrestrial ice-stream imprints yet identified. Its location on the Canadian Shield is surprising (see Clark 1994) and we now explore the possible controls on its location and vigour, fitting our interpretation within the wider spectrum of ice streams in general.

Close inspection of the surface slope across the inferred southern lateral shear margin (Fig. 10) reveals that although surface elevations are slightly higher outside of the margin, there is no discernible 'step' in topography that fixed its position along its entire length. Analysis of the regional topography (Fig. 8) supports the notion that this ice stream was not channelled by the underlying topography.

The location of many documented ('non-topographic') ice streams from the Laurentide Ice Sheet appears to be related to areas of the ice-sheet bed underlain by 'soft' sediments (Hicock 1988; Clark 1994; Patterson 1997, 1998; Clark & Stokes 2001). Thick (tens of metres) fine-grained sediments (e.g. carboniferous limestone, clayey tills) impede drainage of subglacial water, allowing them to become saturated. This results in increased subglacial pore-water pressures and reduced basal shear stress beneath the ice sheet (Clark 1994). Such conditions facilitate fast ice flow, either through pervasive deformation of the till layer or through basal sliding across the surface of the till layer (Engelhardt & Kamb 1998). Additionally, seismic observations from the up-glacier part of a contemporary ice stream in West Antarctica have led some workers to conclude that its position is dependent on a soft sediment bed (Anandakrishnan *et al.* 1998).

Although the Canadian Shield is characterized by granitic gneisses described as hard bedrock, large parts of the Dubawnt Lake ice-stream bed are underlain by unmetamorphosed sedimentary outcrops (Fig. 7). However, the margins of the sedimentary outcrops in no way correspond to the margin locations of the ice stream, either up-glacier or down-glacier (Fig. 7). It may be that those ice streams that appear to be constrained by subglacial geology (e.g. Anandakrishnan *et al.* 1998) are extreme end-members of a larger spectrum ranging from soft to hard bed geology. The variable geology beneath the Dubawnt Lake ice stream places it as an intermediate between the two. As such, it is unlikely that the soft sediments were the only control on the location of this ice stream. A possible contemporary analogue may be Rutford Ice Stream, which feeds the Ronne Ice Shelf in West Antarctica, although it lies in a topographic trough. Seismic reflection data from beneath this ice stream indicate that parts of the bed are characterized by soft, saturated sediments that may be pervasively deforming, whereas other areas are non-deforming and basal sliding may be a more important process there (Smith 1997).

Theoretically, the variable geology beneath the Dubawnt Lake ice stream could have facilitated a combination of basal sliding or till deformation. However, the lobate terminus of the ice stream is characterized by a thin or patchy till cover and there are no substantial sediment accumulations other than the MacAlpine Moraine system (cf. Aylsworth & Shilts 1989c). The termini of other terrestrial ice streams thought to have operated through a deforming bed mechanism are characterized by substantial zones of 'stagnation landforms', hummocky topography and considerable volumes of exported sediment, e.g. the Des Moines Lobe at the southern Laurentide margin (see Patterson 1997).

It could be argued that the lack of substantial sediment towards the terminus of the Dubawnt Lake ice stream is evidence against the deforming bed mechanism. While this explanation is appealing, the thickness of till at the lobate front would depend not only on the operation of a deforming bed, but also on the rate of advance and the longevity of the system. The presence or absence of abundant subglacial and proglacial meltwater streams near the ice front that might remove till is also important. Thus, if the ice stream was short-lived, rapid advance would produce a thin or discontinuous till even with a deforming bed. We speculate that the soft tills from the Thelon Sedimentary Basin were transported downstream across the harder bedrock and 'lubricated' the bed, but did not necessarily support a pervasively deforming subglacial till layer that contributed significantly to fast ice flow. Rather, we suggest that it is more likely that basal sliding was the predominant flow mechanism.

While the softer sediments permitted faster ice flow though the area, a number of other prerequisites (in combination) may have 'switched on' the ice stream and triggered rapid basal sliding. The presence of Glacial Lake Thelon (cf. Bird 1953; Craig 1964) may have been an important seed point for ice-stream initiation. Increased velocity at the margin may have led to greater basal friction, increased basal meltwater for lubrication and warmer less viscous ice, leading to a further increase in basal sliding velocity (see Payne & Dongelmans 1997). These thermomechanical feedback mechanisms would have propagated warm-based conditions inland. The lower surface profile of the ice stream would also draw-down ice from higher elevations.

We speculate that after fast ice flow was triggered the ice stream captured a considerable amount of ice from the vicinity of the Keewatin Ice Divide. The ice stream probably increased in velocity until (a) it ran out of ice and/or (b) conditions at the ice-stream bed led to a reduction in ice velocity. There is no evidence that the ice-stream margins migrated inwards prior to shut down (such as 'ghost' margin positions), i.e. no gradual decrease in velocity and reduction in width. We conclude, therefore, that whatever mechanism led to ice-stream shut-down it happened over a short time period and that the ice stream did not operate during the final phase of deglaciation.

The ribbed moraines, which lie superimposed on the ice-stream bedforms, may be related to the processes that led to or resulted in ice-stream shut-down. If the ice stream ran out of ice it would have thinned the ice sheet considerably, possibly leading to a return to cold-based conditions. Alternatively, ice-stream shut-down may have resulted from the advection of colder ice from up-glacier which promoted widespread basal freezing (Christoffersen & Tulaczyk 2003). It is impossible to say whether basal freezing was the cause or a result of ice-stream shut-down, but it seems likely that a reduction in basal shear heating played an important role. The ubiquitous ribbed moraines may record either the fracturing of a frozen till sheet after ice-stream shut-down (Hättestrand & Kleman 1999), or, we suggest, sticking and slipping over cold-based patches, leading to ice-stream shut-down. Given the paucity of information regarding processes of ice-stream shut-down, the Dubawnt Lake ice-stream bed may represent a fruitful venue for further investigations.

Conclusions and implications

The Dubawnt Lake ice stream is reconstructed at over 450 km in length, 140 km in width and had an estimated catchment area of around 190 000 km^2. Despite the convergent onset zone and divergent terminus, the ice-stream bedforms display a remarkably coherent pattern. In the main trunk of the ice stream, where flow was presumably fastest, bedform elongation ratios approach 40:1 and their exceptional parallel conformity gives rise to a ridge/groove appearance. Elongation ratios decrease up-ice and down-ice, matching the expected variations in ice velocity along a terrestrial ice stream. We take this imprint to record a single episode (isochronous) of fast ice flow and the superimposition of the lineaments on all other flow patterns in the region indicates that it was the last active flow prior to deglaciation. The terminus coincides with a discontinuous end moraine system composed of glaciofluvial deposits, ice-contact deltas, outwash fans and several ridges (the MacAlpine moraines) which have been dated to around 8200 (uncalibrated) yr BP (Blake 1963; Craig 1965).

The time period during which the ice stream is thought to have operated (8200–9000 (uncalibrated) yr BP) coincides with other periods of end moraine construction around the shrinking Laurentide Ice Sheet. Enhanced snowfall over the Keewatin Ice Divide coupled with margin retreat may have led to a steep ice-sheet profile with high driving stresses. We speculate that as the margin retreated inland, the development of Glacial Lake Thelon would have induced calving and led to a subsequent increase in ice velocity which propagated warm-based conditions up-ice through a series of thermomechanical feedbacks.

The location of the ice stream does not conform to a change in the underlying topography or geology.

Although the Canadian Shield is generally characterized by physically hard crystalline bedrock, a large part of the ice-stream bed is underlain by softer sedimentary rocks. It is unlikely that these softer rocks dictated the location of the ice stream within the ice sheet, but they helped lubricate the ice stream over the harder crystalline bedrock down-ice. We acknowledge that the variable geology beneath the ice stream could potentially have led to a combination of basal sliding and subglacial till deformation. However, we speculate that because of (i) the generally hard bedrock, (ii) the lack of thick deposits of soft sediments, and (iii) the lack of evidence of significant till transport, that basal sliding was the predominant flow mechanism.

We recognize that while bed properties are likely to be influential in determining the occurrence and vigour of some ice streams, soft-bed conditions are not an essential requirement. Some ice streams can arise purely by glaciological controls (Payne & Dongelmans 1997). In this case, the ice stream acted as a release valve on ice-sheet accumulation. An ice stream had to evolve somewhere in order to regulate the ice sheet and owing to the slightly lower elevation and presence of proglacial lakes, the strongest seed point was that of the Dubawnt Lake ice stream. Fast ice flow was possible (as evidenced by the bedform signature) and was probably achieved without a pervasively deforming bed.

We note that similar 'hard bedded' ice streams have been hypothesized for parts of the Scandinavian Ice Sheet as it deglaciated across the Baltic Shield (Punkari 1980). Such ice streams have been modelled by Payne & Dongelmans (1997), who concluded that they can arise purely from thermomechanical feedback mechanisms. An implication of their work, and the existence of the Dubawnt Lake Palaeo-ice stream, is that it challenges the view that pure ice streams are necessarily restricted to areas of an ice sheet underlain by soft, saturated, deformable sediments.

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References

Anandakrishnan, S. & Alley, R. B. 1997: Stagnation of Ice Stream C, West Antarctica by water piracy. *Geophysical Research Letters 24*, 265–268.

Anandakrishnan, S., Blankenship, D. D., Alley, R. B. & Stoffa, P. L. 1998: Influence of subglacial geology on the position of a West Antarctic ice stream from seismic observations. *Nature 394*, 62–65.

Andrews, J. T. 1985: Reconstruction of environmental conditions in the eastern Canadian arctic during the last 11,000 years. *In* Harrington, C. R. (ed.): *Climatic Change in Canada*, 423–451. Syllogeus, National Museums of Ottawa.

Andrews, J. T. 1989: Quaternary geology of the northeastern Canadian Shield. *In* Fulton , R. J. (ed.): *Quaternary Geology of Canada and Greenland*, 276–301. Geological Survey of Canada, Geology of Canada, no. 1.

Ayslworth, J. M. & Shilts, W. W. 1985: Glacial features of the west central Canadian Shield. *Current Research, Part B, Geological Survey of Canada, Paper 85-1B*, 375–381.

Aylsworth, J. M. & Shilts, W. W. 1989a: Bedforms of the Keewatin Ice Sheet, Canada. *Sedimentary Geology 62*, 407–428.

Aylsworth, J. M. & Shilts, W. W. 1989b: Glacial features around the Keewatin Ice Divide: Districts of Mackenzie and Keewatin. *Geological Survey of Canada Paper 88-24*, 21 pp.

Aylsworth, J. M. & Shilts, W. W. 1989c: Glacial features around the Keewatin Ice Divide: Districts of Mackenzie and Keewatin. *Geological Survey of Canada, Map 24-1987*, scale 1:1,000,000.

Bentley, C. R. 1987: Antarctic ice streams: a review. *Journal of Geophysical Research 92 (B9)*, 8843–8858.

Bird, J. B. 1953: The glaciation of central Keewatin, Northwest Territories, Canada. *American Journal of Science 251*, 215–230.

Blake, W., Jr. 1963: Notes on glacial geology, northeastern District of Mackenzie. *Geological Survey of Canada, Paper 63-28*, 12 pp.

Blake, W. Jr. 1966: End moraines and deglaciation chronology in northern Canada, with special reference to southern Baffin Island. *Geological Survey of Canada Paper 66-26*, 31 pp.

Bond, G., Heinrich, H., Broecker, W., Labeyrie, L., McManus, J., Andrews, J., Huon, S., Jantschik, R., Clasen, S., Simet, C., Tedesco, K., Klas, M., Bonani, G. & Ivy, S. 1992: Evidence for massive discharges of icebergs into the North Atlantic ocean during the last glacial period. *Nature 360*, 245–249.

Boulton, G. S. & Clark, C. D. 1990: A highly mobile Laurentide Ice Sheet revealed by satellite images of glacial lineations. *Nature 346*, 813–817.

Christoffersen, P. & Tulaczyk, S. 2003: Signature of palaeo-ice stream stagnation: till consolidation induced by basal freeze-on. *Boreas 32*, 114–129.

Clark, C. D. 1993: Mega-scale glacial lineations and cross-cutting ice-flow landforms. *Earth Surface Processes and Landforms 18*, 1–29.

Clark, C. D. 1999: Glaciodynamic context of subglacial bedform generation and preservation. *Annals of Glaciology 28*, 23–32.

Clark, C. D. & Stokes, C. R. 2001: Extent and basal characteristics of the M'Clintock Channel ice stream. *Quaternary International 86*, 81–101.

Clark C. D. & Stokes, C. R. In press: The palaeo-ice stream landsystem. *In* Evans, D. J. A. (ed.): *Glacial Landsystems.* Arnold, London.

Clark, P. U. 1992: Surface form of the southern Laurentide Ice Sheet and its implications to ice-sheet dynamics. *Geological Society of America Bulletin 104*, 595–605.

Clark P. U. 1994: Unstable behaviour of the Laurentide Ice Sheet over deforming sediment and its implications for climate change. *Quaternary Research 41*, 19–25.

Craig, B. G. 1964: Surficial geology of east-central District of Mackenzie. *Geological Survey of Canada Bulletin 99*, 1–41.

Craig, B. G. 1965: Notes on moraines and radiocarbon dates in northwest Baffin Island, Melville Peninsula, and northeast District of Keewatin. *Geological Survey of Canada Paper 65-20*, 7 pp.

Dyke, A. S. 1984: Quaternary geology of Boothia Peninsula and northern District of Keewatin, Central Canadian Arctic. *Geological Survey of Canada Memoir 407*, 26 pp.

Dyke, A. S. & Dredge, L. A. 1989: Quaternary geology of the northwestern Canadian Shield. *In* Fulton, R. J. (ed.): *Quaternary Geology of Canada and Greenland*, 189–214. Geological Survey of Canada, Geology of Canada, no. 1.

Engelhardt, H. & Kamb, B. 1998: Basal sliding of Ice Stream B, West Antarctica. *Journal of Glaciology 44*, 223–230.

Falconer, G., Ives, J. D., Løken, O. H. & Andrews, J. T. 1965a: Major

end moraines in eastern and central Arctic Canada. *Geographical Bulletin 7*, 137–153.
Falconer, G., Andrews, J. T. & Ives, J. D. 1965b: Late-Wisconsin end moraines in Northern Canada. *Science 147*, 608–610.
Hart, J. K. 1999: Identifying fast ice flow from landform assemblages in the geological record: a discussion. *Annals of Glaciology 28*, 59–67.
Hättestrand, C. & Kleman, J. 1999: Ribbed moraine formation. *Quaternary Science Reviews 18*, 43–61.
Hicock, S. R. 1988: Calcareous till facies north of Lake Superior, Ontario: implications for Laurentide ice streaming. *Géographie Physique et Quaternaire 42,* 120–135.
Hughes, T. J. 1992: Abrupt climatic change related to unstable ice-sheet dynamics: toward a new paradigm. *Palaeogeography Palaeoclimatology Palaeoecology 97*, 203–234.
Joughin, I. & Tulaczyk, S. 2002: Positive mass balance of the Ross Ice Streams, West Antarctica. *Science 295*, 476–480.
Kleman, J. & Borgström, I. 1996: Reconstruction of palaeo-ice sheets: the use of geomorphological data. *Earth Surface Processes and Landforms 21*, 893–909.
Mathews, W. H. 1991: Ice sheets and ice streams: thoughts on the Cordilleran Ice Sheet Symposium. *Géographie Physique et Quaternaire 45*, 263–267.
Meier, M. F. & Post, A. 1969: What are glacier surges? *Canadian Journal of Earth Sciences 6*, 807–817.
Oppenheimer, M. 1998: Global warming and the stability of the West Antarctic Ice Sheet. *Nature 393*, 325–332.
Patterson, C. J. 1997: Southern Laurentide ice lobes were created by ice streams: Des Moines Lobe in Minnesota, USA. *Sedimentary Geology 111*, 249–261.
Patterson, C. J. 1998: Laurentide glacial landscapes: the role of ice streams. *Geology 26*, 643–646.
Payne, A. J. & Dongelmans, P. W. 1997: Self-organisation in the thermomechanical flow of ice sheets. *Journal of Geophysical Research 102 (B6)*, 12219–12234.
Prest, V. K., Grant, D. R. & Rampton, V. N. 1968: *Glacial Map of Canada*. Geological Survey of Canada, Map 1253A, scale 1:5 000 000.
Punkari, M. 1980: The ice lobes of the Scandinavian ice sheet during the deglaciation in Finland. *Boreas 9*, 307–310.
Rose, K. E. 1979: Characteristics of flow in Marie Byrd Land, Antarctica. *Journal of Glaciology 24,* 63–75.
Scott, J. S. 1976: Geology of Canadian tills. *In* Legget, R. F. (ed.): *Glacial Till: An Interdisciplinary Study*, 50–66. Royal Society of Canada Special Publication 12, Ottawa.
Smith, A. M. 1997: Variations in basal conditions on Rutford Ice Stream, West Antarctica. *Journal of Glaciology 43,* 245–255.
Stokes, C. R. 2000: *The Geomorphology of Palaeo-Ice Streams: Identification, Characterisation and Implications for Ice Stream Functioning*. Ph.D. dissertation, University of Sheffield, 250 pp.
Stokes, C. R. & Clark, C. D. 1999: Geomorphological criteria for identifying Pleistocene ice streams. *Annals of Glaciology 28*, 67–75.
Stokes, C. R. & Clark, C. D. 2001: Palaeo-ice streams. *Quaternary Science Reviews 20*, 1437–1457.
Stokes, C. R. & Clark, C. D. 2002: Are long subglacial bedforms indicative of fast ice flow? *Boreas 31*, 239–249.
Wellner, J. S., Lowe, A. L., Shipp, S. S. & Anderson, J. B. 2001: Distribution of glacial geomorphic features on the Antarctic continental shelf and correlation with substrate: implications for ice behaviour. *Journal of Glaciology 47*, 397–411.