The Geomorphology of Palaeo-Ice Streams: Identification, Characterisation and Implications for Ice Stream Functioning.

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Abstract.

Ice streams are the dominant drainage pathways of contemporary ice sheets and their location and behaviour are viewed as key controls on ice sheet stability. Identifying palaeo-ice streams is of paramount importance if we are to produce accurate reconstructions of former ice sheets and examine their critical role in the ocean-climate system.

Many workers have invoked palaeo-ice streams from a variety of former ice sheets, despite a limited understanding of their glacial geomorphology. This thesis addresses the problem by predicting several diagnostic geomorphological criteria indicative of ice stream activity. These are developed objectively from the known characteristics of contemporary ice streams and can be summarised as: large flow-set dimensions (>20 km wide and >150 km long), highly convergent flow patterns, highly attenuated subglacial bedforms (length:width >10:1), Boothia-type dispersal plumes, abrupt lateral margins (<2 km), ice stream marginal moraines, evidence of pervasively deformed till, and submarine sediment accumulations (marine-terminating ice streams only). Collectively, the criteria are used to construct conceptual landsystems of palaeo-ice stream tracks. Using satellite imagery and aerial photography to map glacial geomorphology, identification of the criteria is used to validate the location of a previously hypothesised ice stream and identify a hitherto undetected palaeo-ice stream from the former Laurentide Ice Sheet. Implications for ice stream basal processes are explored and their ice sheet-wide significance is assessed.

On Victoria Island (Arctic Canada) five of the geomorphological criteria are identified and the extent of the marine-based M'Clintock Channel Ice Stream is reconstructed at 720 km in length and 140 km in width. The ice stream (operating between 10,400 and 10,000 yr BP) was located within a broad topographic trough, but internal glaciological processes, rather than properties of the bed controlled the margin locations. It eroded into pre-existing unconsolidated sediments and left a spectacular pattern of subglacially-produced landforms, recording a snapshot view of the bed prior to ice stream shut-down. Sediment availability appears critical to its functioning (deformable bed?) and the debris flux of the ice stream is inferred to have been high. Frictional shut-down occurred once down-cutting through sediments reached hard bedrock close to the terminus.

The presence of four of the geomorphological criteria are used to identify a terrestrial ice stream which drained the Keewatin Sector of the Laurentide Ice Sheet between ca. 10,000 and 8,500 yr BP. Its size is reconstructed at over 450 km in length and 140 km in width, and it left behind a subglacial bedform pattern consisting of highly attenuated drumlins (length:width ratios up to 48:1) displaying exceptional parallel conformity. This represents an isochronous bedform pattern and variations in lineament elongation ratio are thought to be a useful proxy for ice velocity. Highest elongation ratios occur immediately downstream of a topographic step where the ice stream entered a sedimentary basin. It is inferred that the ice stream was triggered by climatic warming which altered the ice sheet configuration and the thermal state of the bed. A switch from cold to warm-based conditions probably triggered rapid basal sliding. The ice stream (and a tributary) shut down when it ran out of ice, causing widespread thinning of the ice sheet and subsequent deglaciation.

These ice streams denote considerable ice sheet instability over both hard and soft (deformable) beds and emphasise the enormous effects that ice streams had in controlling the deglaciation of the Laurentide Ice Sheet.

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Contents.

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Chapter 1: Introduction.

1.1. The Significance of Ice Sheets.

Contemporary ice sheets cover around 10% of the earth's surface. These phenomena accommodate over 33 million km^3 of freshwater (Benn and Evans, 1998), constituting around 70-80% of the earth's total, the majority of which (89%) is locked up in Antarctica's ice mass (Bindschadler, 1998). As Bindschadler (1998) noted, early attempts at modelling the earth's climate ignored these polar regions, partly for simplicity, but partly because their importance was significantly underestimated. Now, it is known that ice sheets are critical to our understanding of the ocean-climate system. They regulate the meridional transfer of heat from the tropics to the poles and their significant topographic effect influences large-scale atmospheric flow. In addition, their growth compresses the crust of the earth and their subsequent disappearance causes substantial rebound of the surface.

Of greater significance, however, is the fact that a 1% change in their volume can raise sea levels by some 70 cm (Van der Veen, 1987). If contemporary ice sheets were to disappear completely, global sea levels would be increased by some 70 m (Benn and Evans, 1998). In contrast, when ice sheets reached their maximum extent during the last major glaciation (around 20,000 years ago), sea level has been estimated to be around 125 m lower than today (Denton and Hughes, 1981).

The relevance of this for today's society are the potential changes occurring in contemporary ice sheets which may take place over shorter (possibly decadal) timescales. During the last 20,000 years, sea levels have been shown to have risen abruptly, possibly at rates of over 50 mm a^{-1} (Fairbanks, 1989). The most likely cause of these abrupt shifts are profligate influxes of freshwater into the oceans from waning ice sheets. Indeed, recent evidence for rapid ice sheet collapse has been substantiated by the ocean sedimentological record which records increased pulses of iceberg rafted debris into the North Atlantic from former northern hemisphere ice sheets (Heinrich, 1988). Such influxes of fresh water affected ocean salinity and temperature over very short timescales. For example, a sea surface temperature change of 5°C occurred in as little as 40 years during the last deglaciation (Lehman and Keigwin, 1992). Moreover, such changes had the ability to alter ocean circulation, which in turn, altered hemispheric climate.

Of particular concern is the current debate surrounding the stability of contemporary ice sheets. Since the industrial revolution, the anthropogenic output of greenhouse gases, such as carbon dioxide, has increased at an alarming rate. This has fuelled speculation that significant increases in global temperatures are imminent and this could have a profound effect on contemporary ice sheets. Of greatest threat is the West Antarctic Ice Sheet, because the majority of its bed is grounded below contemporary sea level. This makes it particularly vulnerable to possible changes in sea level and some workers have argued that it is inherently unstable (e.g. Mercer, 1978).

The possible instability of the West Antarctic Ice Sheet has also been evidenced by the non-uniform flow velocity of the ice sheet. While much of the ice sheet flows at typical velocities below 50 m a^{-1}, there are distinct arteries which flow at least an order of magnitude faster, usually in excess of 400 m a^{-1}. These features are known as *ice streams* and have been the focus of intensive research because they are thought to be the key controls on the ice sheet's discharge and stability.

1. 2. The Significance of Ice Streams.

An ice stream is a region in an ice sheet which flows much faster than the surrounding ice. A unique characteristic of ice streams is that they are bordered by slower-moving ice. This creates a zone of intense crevassing at their margins and allows them to be identified in ice sheets. Figure 1.1 shows a schematic diagram of the West Antarctic Ice Sheet, where several ice streams are being studied.

Most ice streams are over 20 km wide and over 150 km long. Because of their size and profligate ice flux, they have a profound effect on ice sheet configuration and are thought to represent key controls on ice sheet stability. Therefore, it is very important to ascertain the spatial and temporal controls on their location and behaviour. Why are they located where they are, and what enables them to move so fast? Answering these questions may be able to help us predict the future behaviour of contemporary ice sheets and in particular, the West Antarctic Ice Sheet. However, as Burckle (1993) commented, the glaciological community are "charged with predicting the future behaviour of this ice sheet while knowing very little about its past behaviour".

It is thus necessary to broaden our understanding of ice streams by investigating their location and behaviour over longer timescales than our contemporary observations permit. This thesis aims to contribute to our understanding of ice streams by investigating their basal characteristics and behaviour in former ice sheets.

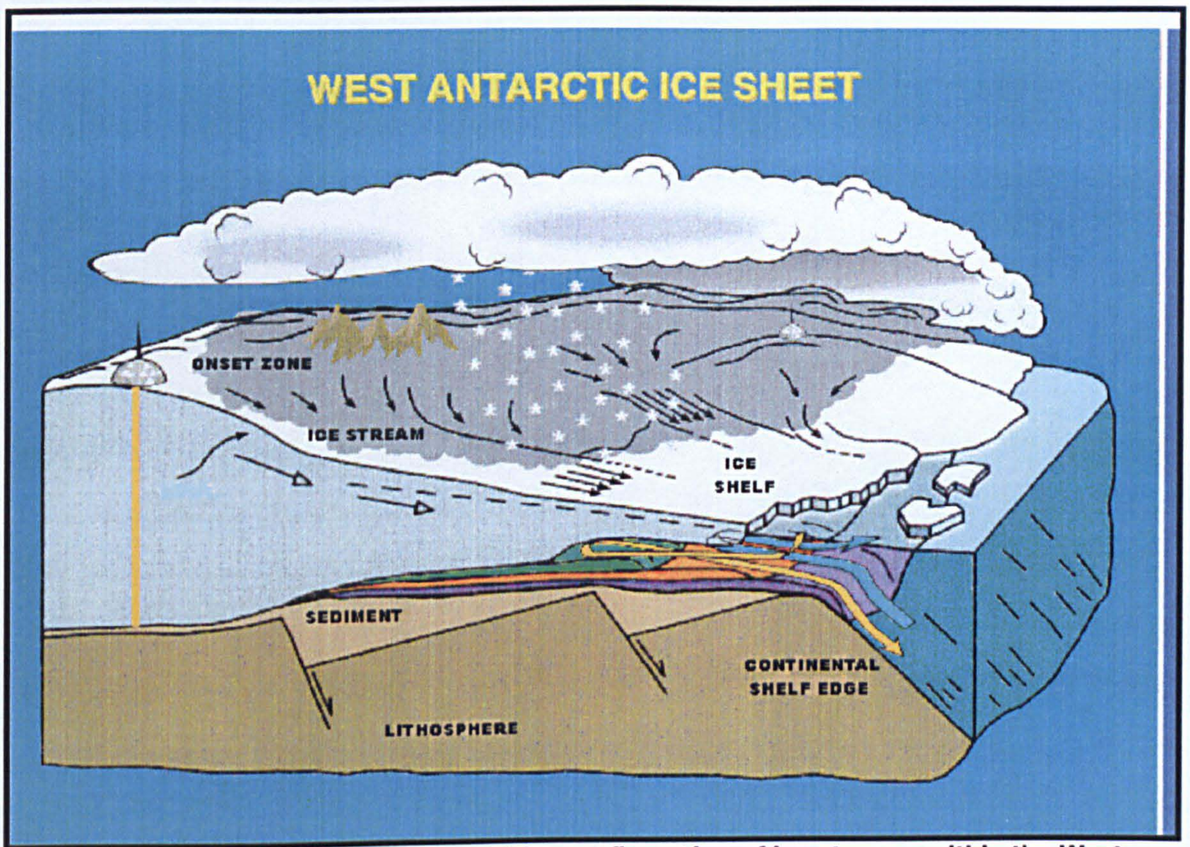

Figure 1.1. Schematic representation of the configuration of ice streams within the West Antarctica Ice Sheet (from Bindschadler *et al.*, 1998)

1.3. Specific Aims of the Research.

The specific aims of this research are as follows;

1. To provide the first ever literature review of palaeo-ice streams.
2. To identify the main inhibitors to palaeo-ice stream research.

3. To develop and predict several geo(morpho)logical criteria to aid the identification of palaeo-ice streams.

4. To use the criteria in (3) to validate and then investigate the location of a previously hypothesised marine-terminating ice stream and;

(a), use remote sensing to map and characterise the ice stream bed.

(b), use the bedform imprint to glean information about ice stream behaviour and functioning.

5. To use the criteria in (3) to postulate on the location of a hitherto unidentified terrestrial-terminating palaeo-ice stream and;

(a), use remote sensing to map and characterise the ice stream bed.

(b), use the bedform imprint to glean information about ice stream behaviour and functioning.

6. To use the case studies in (4) and (5) to provide an appraisal of the geomorphological criteria (in 3).

7. To explore the glaciological implications and ice sheet-wide significance of the palaeo-ice streams identified in (4) and (5), and to discuss the role of ice streams in ice sheets and the controls on their location, initiation and functioning.

These aims inevitably introduce an element of circularity. This unavoidable problem is discussed in more detail at the end of Chapter 4 (Section 4.5.2).

1.4. Thesis Structure.

This chapter has briefly outlined the significance of ice sheets and ice streams. **Chapter 2** provides a detailed description of the characteristics of contemporary ice streams from the Antarctic and Greenland ice sheets. This includes a discussion of their flow mechanisms, their resistive forces, the transition to ice stream flow, and outlines the problems of modelling ice streams. The significance of ice streams and climate change is presented and it is concluded that to fully understand ice stream behaviour, we must also investigate those ice streams which operated in former ice sheets.

Chapter 3 represents the first ever review of palaeo-ice stream hypotheses in the literature and it is clear that there are several inhibitors to palaeo-ice stream research. Of these, the most pertinent is a lack of objectivity concerning their geomorphological products. A huge variety of evidence has been used to cite former ice stream locations but this has rarely been scrutinised in detail. **Chapter 4** addresses this problem by predicting several geomorphological criteria for identifying palaeo-ice streams. These criteria are based entirely on the known characteristics of contemporary ice streams (described in Chapter 2) and can be grouped into a glacial landsystem which provides an observational template upon which ice stream hypotheses can be better based.

Following the theoretical predictions in Chapter 4, the second part of the thesis investigates the bedform record of two palaeo-ice streams which drained the former North American (Laurentide) Ice Sheet. Satellite imagery provides an excellent tool to map glacial geomorphology over large areas relatively quickly, and **Chapter 5** outlines the basic methodology used in this thesis

Chapter 6 describes the basal characteristics of the previously hypothesised M'Clintock Channel Ice Stream which drained the north-western margin of the Laurentide Ice Sheet. Having established that it represents a valid candidate for a marine-terminating palaeo-ice stream track, its bedform imprint is described and its extent is reconstructed. Implications for ice stream functioning and basal processes are explored.

Using the criteria outlined in Chapter 4, **Chapter 7** presents new evidence for a large terrestrial ice stream which drained the Keewatin Sector of the Laurentide Ice Sheet. Its bedform imprint is described and its extent is reconstructed. Implications for ice stream functioning are explored and its role in the deglaciation of the Laurentide Ice Sheet is assessed.

Chapter 8 draws together several themes of the thesis and provides a discussion of the wider implications of the research. The geomorphological criteria developed in Chapter 4 are appraised in the light of the two palaeo-ice stream case studies in Chapters 6 and 7. The glaciological implications of the two palaeo-ice streams are also explored and their ice sheet-wide significance is assessed. There then follows a discussion of the role of ice streams in ice sheets and a conceptual model of ice stream initiation and functioning is presented.

Concluding remarks outlining the key discoveries of this thesis are found in **Chapter 9**. Contributions of this thesis have also been written up in the following publications; Stokes and Clark (1999), Stokes (2000), Clark and Stokes (in press), Stokes and Clark (submitted[a]), and Stokes and Clark (submitted[b]).

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Chapter 2: Characteristics of Contemporary Ice Streams.

2.1. Introduction.

The aim of this chapter is to introduce and describe ice streams, arguably the most dynamic component of contemporary ice sheets. It begins by defining ice streams and explaining their simple classification with respect to the underlying topography. The location of *pure ice streams* is irrespective of underlying topography whereas *topographic ice streams* are constrained by the bedrock troughs in which they flow. The characteristics of both types of ice stream are therefore examined, including their flow mechanisms, resistive stresses and the transition from ice sheet to ice stream flow. There then follows a brief discussion of the problems of modelling ice streams, including basal modelling, ice stream/ice shelf modelling and the treatment of ice streams within ice sheet models. This chapter ends with a discussion of the importance of understanding ice streams and their critical interaction with climate. It is concluded that to fully understand the behaviour of ice streams over century to millennial timescales, we must investigate the activity of palaeo-ice streams which drained the northern hemisphere ice sheets during the last major glaciation.

It is anticipated that this chapter will provide a summary of the characteristics of ice streams which will serve to emphasise their dynamic role in both contemporary and Quaternary ice sheets.

2.2. Definition and Classification of Ice Streams.

An ice stream is defined as "a region in a grounded ice sheet in which the ice flows much faster than in the regions on either side" (Paterson, 1994). If the fast flowing ice is bordered by rock it is generally considered to be an outlet glacier. Morgan *et al.* (1982) estimated that 90% of Antarctica's drainage is accounted for by outlet glaciers and ice streams. The relative importance of these features is further emphasised by the fact that they only comprise around 13% of the Antarctic coastline (Paterson, 1994). Similarly, much of the discharge of the Greenland ice sheet is channelled through ice streams and outlet glaciers. A single ice stream in Greenland, Jakobshavns Isbræ, is thought to account for as much as 6.5% of the annual mass lost from the ice sheet (Echelmeyer *et al.*, 1991). It is now widely acknowledged then, that the discharge from ice sheets is largely governed by fast-flow through outlet glaciers and ice streams.

2.2.1. Pure and Topographic Ice Streams.

Most ice streams can be loosely categorised as either 'pure ice streams' (i.e. those which do not lie in bedrock troughs) and 'topographic ice streams' (i.e. those which are in some way controlled by the underlying topography). This simple classification is particularly important because pure and topographic ice streams are believed to operate by different flow mechanisms (see Sections 2.3.2 and 2.4.2). Furthermore, because topographic ice streams are fixed in space, their behaviour is thought to be far more predictable than pure ice streams.

It is often far more difficult to make a distinction between a topographic ice stream and a large outlet glacier. The confusion arises because many topographic ice streams display characteristics of both, especially those in East Antarctica and Greenland (Bentley, 1987). This is illustrated by Jakobshavns Isbræ, which begins as an ice stream surrounded by slower moving ice but becomes bordered by rock walls as it flows towards the coastline (Clarke, 1987). Although the distinction between an outlet glacier and topographic ice stream is clear in principle, they are often difficult to distinguish in practice (Bentley, 1987).

2.2.2. Marine-Based and Terrestrial Ice Streams.

It is also possible to categorise ice streams as either marine-based or terrestrial. Marine-based ice streams terminate in a marine environment, usually into an ice shelf (e.g. the Siple Coast Ice Streams of West Antarctica) or into open water conditions, perhaps as an ice tongue (e.g. Thwaites and Pine Island Glacier which flow directly into the Amundsen Sea, West Antarctica). In contrast, terrestrial ice streams terminate on land, although proglacial lakes may produce a terminal environment largely analogous to that of a marine-based ice stream.

All contemporary ice streams are marine-based, but this can be considered a geographical coincidence. It is important to acknowledge that ice streams almost certainly drained the terrestrial margins of extinct ice sheets, e.g. the Laurentide Ice Sheet in North America. It should also be noted that it is quite conceivable that, given time, an ice stream can alternate between marine-based and terrestrial, either as a result of an advancing or retreating ice mass, or as a result of a change in relative sea level.

2.3. Characteristics of Pure Ice Streams.

2.3.1. Location of Pure Ice Streams.

The location of pure ice streams is not controlled by their underlying topography. A major challenge therefore, is to ascertain what does facilitate and confine their fast flow. This has resulted in pure ice streams attracting a disproportionate amount of interest in the literature.

The pure ice streams which flow towards the Siple Coast of West Antarctica and feed the Ross Ice Shelf are by far the most widely studied ice streams. Figure 2.1 shows the five main ice streams, named A-E, which have been delineated by their heavily crevassed shear margins. They range in width from 30-80 km and reach lengths of between 300 and 500 km. These ice streams have been monitored closely and much data are available in terms of their configuration (e.g. Shabtaie and Bentley, 1987; Jacobel *et al.*, 1996), morphology (e.g. Shabtaie *et al.*, 1987), velocity (e.g. Whillans and Van der Veen, 1993), strain rates (e.g. Jackson and Kamb, 1997) and basal characteristics (Engelhardt and Kamb, 1998). Table 2.1 shows a selection of published data collected from the Siple Coast ice streams.

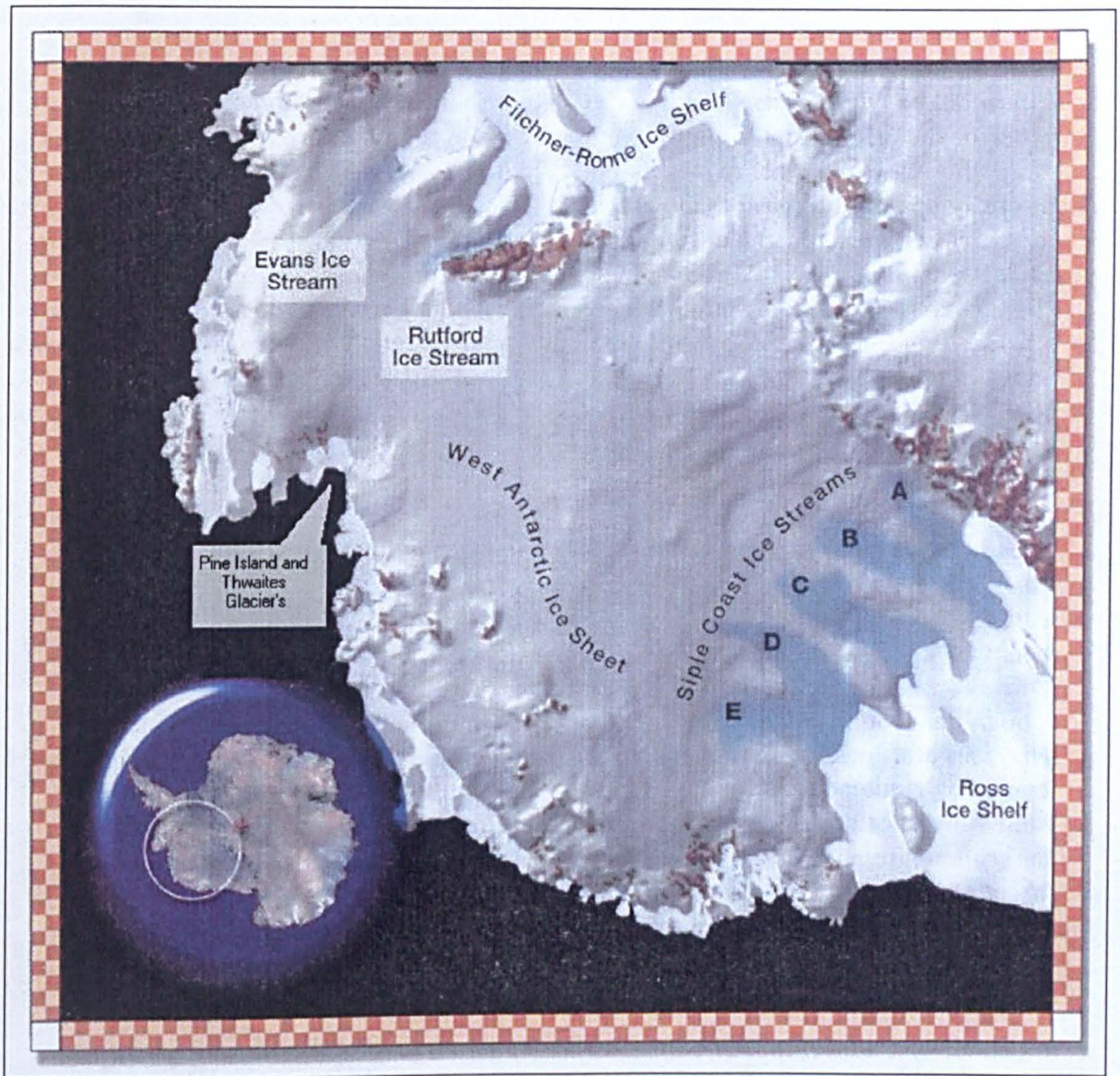

Figure 2.1. Location map of the Siple Coast ice streams and other topographic ice streams in the West Antarctic Ice Sheet (from Walker , 1999, New Scientist, No. 2182, p. 57).

It can be seen from Table 2.1 that Ice Stream C is characterised by the slowest velocities. Radar investigations of buried crevasses have revealed that this ice stream shut down approximately 130 years ago (Retzlaff and Bentley, 1993). This is of interest because it emphasises the temporally variable nature of ice streaming on the Siple Coast (described in more detail in Section 2.3.5).

Table 2.1. Selection of published data from pure and topographic ice streams in Antarctica and Greenland.

Ice Stream type	Name	Length (km)	Width (km)	Thickness (m)	Velocity (m a^{-1})	Drainage basin area (km^2)	References
Pure Ice Streams	Ice Stream A	>200	~50	~1000	217-254	No data	Echelmeyer *et al.* (1994); Rose (1979) Shabtaie *et al.* (1987)
	Ice Stream B	500 300	35 50 30-80	1000->2000	450 >500 ~500	163,000	Alley *et al.* (1989); Engelhardt and Kamb (1998); Rose (1979); Scambos and Bindschadler (1993); Shabtaie and Bentley (1987); Whillans and Van der Veen (1993)
	Ice Stream C	<400	50-60	200-1500	1-13 40 50 5	122,000	Fastook (1987); Rose (1979); Shabtaie and Bentley (1987); Whillans and Van der Veen (1993); Whillans *et al.* (1987)
	Ice Stream D	550	46-59 30-82	800 927-1432	420-670 700 100-370	104,000	Hodge and Doppelhammer (1996); Rose (1979); Scambos and Bindschadler (1993); Scambos *et al.* (1994)
	Ice Stream E	~320	75-100	975-1091	400-550 ~400	131,000	Bentley (1987); Rose (1979); Scambos and Bindschadler (1993)
Topographic Ice Streams	Pine Island Glacier	200 ~300	26 30	1564 ~2000	1300-2600	220,000	Jenkins *et al.* (1997); Lucchitta *et al.* (1995); Bentley (1987)
	Thwaites Glacier	ca. 300	ca. 83	~2000	2200-3400	121,000	Rosanova *et al.* (1998); Ferrigno *et al.* (1998); Bentley (1987)
	Rutford Ice Stream	>150	18-26	1640-2250	360-400 302-377	~36,000	Frolich and Doake (1988); Doake *et al.* (1987)
	Jutulstraumen Ice Stream	ca. 300	ca. 40	2000	443	124,000	Høydal (1996)
	Jakobshavns Isbræ	70-80 85-90	~6	2500 1900-2600	800-7000 8360	10,000	Clarke and Echelmeyer (1996); Echelmeyer and Harrison (1990); Echelmeyer *et al.* (1991); Fastook *et al.* (1995); Lingle *et al.*(1981)

Until recently, it was thought that the Siple Coast ice streams were the only population of pure ice streams in contemporary ice sheets. However, a 'rapid flow feature' has been identified in north-east Greenland (Fahnestock *et al.*, 1993) and this may represent the only pure ice stream outside of the Siple Coast system.

Pure ice streams are characterised by rapid velocities (typically 300-400 m a^{-1}), highly convergent onset zones and extremely abrupt lateral margins. They are unique in that they usually have low driving stresses and extremely low basal shear stresses and this produces a characteristic concave ice surface profile (Bentley, 1987). These fundamental characteristics pose a number of important questions, not least of which is; what enables pure ice streams to attain their rapid velocities? In order to elucidate their exact flow mechanisms, much research has concentrated on the basal environments of pure ice streams.

2.3.2. Mechanisms of Flow.

It was suggested by Rose (1979), that the ice streams on the Siple Coast experienced basal melting, whereas the beds of the intervening ridges were frozen. Subsequent seismic evidence from Ice Stream B has been used to infer the presence of a layer of saturated subglacial till, averaging 5-6 m in thickness (Blankenship *et al.*, 1986; Alley *et al.*, 1986). Perhaps more importantly, high water pressures of only ~50 kPa less than the ice overburden pressure have been found. Indeed, evidence relating to the till porosity, the force balance and the water balance have indicated that the till is deforming. This is because a highly porous layer at a high water pressure would be too weak to support the shear stress exerted by the overlying ice.

These investigations led to the hypothesis that till deformation, rather than basal sliding, accounts for most, if not all of the fast flow of Ice Stream B (Blankenship *et al.*, 1986; Alley *et al.*, 1986). Further analyses presented by Alley *et al.* (1987a; b) suggested that Ice Stream B primarily moves by pervasive deformation of a several metres thick layer of subglacial till and the deforming bed flow model developed by Alley *et al.* (1987b) could explain all of the available data, e.g. surface velocity.

Following the work of Alley *et al.* (1987a; b), Engelhardt *et al.* (1990) provided the first direct measurements of the physical conditions at the base of Ice Stream B. The use of hot water drills confirmed the existence of a melting bed with a basal water pressure near to that of the ice overburden pressure, and a saturated deformable till layer. However, it remained difficult to ascertain whether the dominant flow mechanism was by basal sliding along the surface of this till or by subglacial till deformation.

To investigate the nature of the subglacial drainage system Engelhardt and Kamb (1997) drilled boreholes in Ice Stream B. Their results suggested that the basal water system does not conform to a simple model. The data indicated that observed pressure variations at the bed were spatially random beneath the ice stream, whilst the ice stream velocity pattern represented a much smoother picture. This would appear to suggest that the observed pressure variations were averaged out over the whole bed and integrated to support the ice stream motion (Engelhardt and Kamb, 1997).

Developing this work, Engelhardt and Kamb (1998) employed 'tethered stake' apparatus in boreholes. Results were revealing in that they suggested that basal sliding may be predominant over subglacial deformation. Although shear deformation did take place, it occurred in a thin, centimetres thick layer at the sole of the ice stream. This is in contrast to the model results of Alley *et al.* (1987b) and it is suggested that subglacial till deformation may not be as pervasive and deep as previously thought (Engelhardt and Kamb, 1998). This has major implications for ice stream basal modelling where basal sliding is usually omitted. However, the conclusions are largely dependent on the accuracy of the 'tethered stake' apparatus, which Engelhardt and Kamb (1998) acknowledged can only be resolved by further borehole investigations.

The above findings led some workers to hypothesise that the location of pure ice streams may be controlled by the nature of the underlying geology, and in particular, the availability of 'soft' sediments (e.g. Tulaczyk *et al.*, 1998). Using aerogeophysical data, Bell *et al.* (1998) have recently discovered that the onset zone of Ice Stream D coincides with a sedimentary basin and similar work by Anandakrishnan *et al.* (1998) has indicated that the margin of this ice stream lies immediately above the boundary of the basin.

In summary, the exact flow mechanism of pure ice streams remains elusive but several pioneering studies would appear to indicate that the presence of a weak subglacial till layer at high basal water pressures can account for their rapid velocity through a combination of till deformation and basal sliding. If this is the case, then the availability of soft sedimentary basins may be a key control on their location.

2.3.3. Resistive Stresses on Pure Ice Streams.

In simple terms, ice streams flow in response to the weight of the ice and gravity. This component of their force balance is expressed as the driving stress and can be readily calculated from ice slope profiles. Several resistive forces oppose the driving stress, but it is much more difficult to quantify and partition these components of the force balance.

In outlet glaciers and most topographic ice streams, driving stresses are high (~100 kPa) and their width to thickness ratio is low (Paterson, 1994). The resistive force is provided by friction exerted by the bedrock at the sides and beneath the glacier. However, pure ice streams have low driving stresses (< 50 kPa), low basal shear stresses, and are bordered at the sides by slower moving ice. More importantly, their width to thickness ratio is unusually high. Ice Stream B is at least 30 times wider than it is thick. This is much higher than typical valley glaciers whose width to thickness ratio is usually around 6:1 (Paterson, 1994). Put simply, given the extremely 'slippery bed', it is difficult to ascertain what slows down the ice stream and prevents it from accelerating further.

Traditionally, it was thought that the major resistive force beneath ice streams was provided by the bed. However, one measured sample of till beneath Ice Stream B had a yield stress as low as 2 kPa (Kamb, 1991) and such a layer of till is too weak to support the ice stream motion. Nevertheless, localised patches of high basal drag may be important (Alley, 1993), and evidence indicates that side drag (Jackson and Kamb, 1997) and the back-pressure provided by ice shelves (MacAyeal, 1989) may also be significant. In addition, it should also be remembered that the thermal regime of the ice sheet outside of the ice stream margin will also affect the resistive stresses exerted on the ice stream

2.3.3.1. Basal Drag (Sticky Spots).

Alley (1993) was one of the first to speculate on the role of localised patches of basal drag and coined the term 'sticky spots'. Using the available data from Siple Coast ice streams, Alley (1993) proposed several theoretical mechanisms which may lead to the formation of sticky spots. He postulated that bedrock bumps protruding into the base of the ice stream are the most likely source of sticky spots but variations in ice surface topography may also be significant because that influences variations in the subglacial water flow. In contrast, it is suggested that discontinuities in the basal till layer will only provide negligible 'stickiness', because areas with less till have a tendency to collect a layer of lubricating water (see Alley, 1993).

Developing the largely theoretical work of Alley (1993), MacAyeal *et al.* (1995) used sequential satellite imagery to derive surface velocity fields on Ice Stream E and estimated spatial variations in basal friction. It was calculated that approximately 15% of the ice stream bed was comprised of sticky spots and their observations supported the work of Alley (1993) because areas of high basal drag appeared to be clustered around regions of the bed affected by bedrock ridges. It would therefore appear that sticky spots are an important resistive force on ice streams.

2.3.3.2. Side Drag.

Side drag is the dominant resistive force on outlet glaciers whose margins are bordered by rock, but its influence on pure ice streams is far from clear.

In order to determine the role of side drag in the force balance of Ice Stream B, Echelmeyer *et al.* (1994) surveyed a transverse velocity profile across the margin. Using detailed analytical and numerical models, they concluded that the marginal drag is equal to or greater than the basal drag. However, more recent work by Jackson and Kamb (1997) indicated that marginal shear stresses may be twice that predicted by Echelmeyer *et al.* (1994). Jackson and Kamb (1997) assimilated strain data by tracking crevasses and performed laboratory creep tests on ice marginal samples. It was found that 63-100% of the driving stress is supported by shear stresses at the margins. These data provide a strong argument that side drag is the dominant resistive force on ice streams, although it should be assumed to vary along their length.

2.3.3.2. Ice Shelf Back-Pressure.

The resistance provided by ice shelves is poorly understood and it could be argued that because the resistive stresses on ice shelves themselves are negligible, the buttressing effect that they have is insignificant. However, Whillans (1987) identified back-pressure from ice shelves as a potential resistive stress on ice streams. To examine this, Vornberger and Whillans (1986) mapped the surface features (mainly crevasse patterns) of Ice Stream B and concluded that ice shelf back-pressure is only significant in the lower reaches, near the grounding line.

This viewpoint is in contrast to the traditional concept that suggests that ice streams would rapidly speed up if ice shelves were removed (see Thomas, 1979). Indeed, MacAyeal (1987) argued that 'ice shelf back-pressure' is a significant component of the force balance and that the conclusion of Vornberger and Whillans (1986) stems from an imprecise definition of the term. He used numerical models to describe the mass balance and flow of the Ross Ice Shelf and demonstrated that ice stream activation and decay may be controlled by ice shelves. If this is the case, the Ross Ice Shelf is of great importance when attempting to understand the Siple Coast ice streams because it may play a potentially vital role in halting their collapse.

In summary, the main resistive forces on ice streams are partitioned between, sticky spots, side drag and ice shelf back-pressure, with side drag probably dominant. Quantifying the relative role of each of these components is not straightforward however, and they almost certainly vary, both along and between individual ice streams.

2.3.4. Transition from Ice Sheet to Ice Stream Flow.

It is widely recognised that the majority of the discharge of the West Antarctic Ice Sheet takes place through the Siple Coast ice streams. If this ice sheet is going to collapse, it will presumably involve the headward propagation of the ice streams into the ice sheet (Hodge and Doppelhammer, 1996). It is, therefore, critical that we understand the onset zones of the ice streams and in particular, the transition from ice sheet to ice stream flow.

The simplest explanation for the transition from slow sheet-flow to fast stream-flow is strain heating leading to faster flow. Localised flow acceleration leads to an increase in strain heating which in turn, encourages faster flow. This positive feedback mechanism is thought to lead to the initiation of ice stream onset zones and may be triggered by a topographic ridge beneath the ice stream. This is exactly what Hodge and Doppelhammer (1996) discovered on imagery from the onset zones of Ice Streams C and D. They discovered a relatively sharp transition between ice sheet flow and ice stream flow which occurred immediately downstream of bedrock ridges. Although bedrock steps have been associated with the onset zones of topographic ice streams such as Thwaites Glacier (McIntyre, 1985), this is the first documented evidence from pure ice streams.

Other workers have argued that the zone between the ice sheet flow and ice stream flow is a gradual transition. One of the tributaries of Ice Stream D has been found to show no sharp transition between velocities of <60 m a^{-1} to >120 m a^{-1} and Scambos and Bindschadler (1993) suggested that the onset zones of ice streams can extend over great distances. Furthermore, they used this evidence to infer that the bed characteristics which initiate ice streaming must also occur gradually. This would appear to suggest that topographic features do not have a major influence in initiating the fast flow of some pure ice streams.

Recent investigations using satellite radar interferometry suggest that ice streams are fed by fast flowing tributaries which, unlike the main ice streams, coincide with subglacial valleys (Joughin *et al.*, 1999). They attributed the faster flow within the valleys to higher concentrations of sediment and subglacial water, and warmer ice conditions. Moreover, evidence from balance velocities in ice stream onset zones suggest that some tributaries may extend up to 1000 km into the ice sheet interior (see Bamber *et al.*, 2000).

It has also been hypothesised that individual tributaries from different ice streams share the same source areas (Joughin *et al.*, 1999). This represents a major finding because shared source areas complicate the calculation of individual ice stream mass balances and increase the likelihood that relative contributions to neighbouring ice streams changes over time. This strengthens the hypothesis that Ice Stream C may have shut down due to competition from Ice Stream B (Anandakrishnan and Alley, 1997a), see Section 2.3.5 below.

In summary, the transition from sheet flow to stream flow is characterised by a large zone of convergence at the head of the ice stream. The spatial extent of this transition is still under debate but it is known that in some cases at least, subglacial valleys may act as tributaries.

2.3.5. Transient Behaviour of Siple Coast Ice Streams.

Since the discovery that pure ice stream behaviour is variable on both spatial and temporal scales, some workers have suggested that this may be an intrinsic property of ice streaming (e.g. MacAyeal, 1989).

The first major evidence of significant variations in ice stream discharge were detected on Ice Stream B by Stephenson and Bindschadler (1988). They used repeated satellite tracking at the mouth of the ice stream and estimated that the velocity had decreased by around 20% over a ten year study period. The reason for the deceleration is unclear, but its possible causes included downstream changes in the configuration of an ice rise (Crary Ice Rise) or increases in the ice stream width. The second explanation proved prophetic, because more recent evidence from satellite photography has demonstrated that Ice Stream B has widened by 4 km in less than 30 years (Bindschadler and Vornberger, 1998).

Evidence of ice stream margin migration has also been found by Jacobel *et al.* (1996) who used ice penetrating radar and discovered a buried curved feature in the ice. This scar had been detected previously on satellite imagery and it is thought to represent the former south-western margin of an ice stream flowing from Ice Stream C towards the lower reaches of Ice Stream D, near the Ross Ice Shelf. The burial depth of this feature has been dated to around 1,300 years ago and it represents clear evidence that ice stream configurations are far from stable.

Much research has concentrated on Ice Stream C because it is the only ice stream on the Siple Coast which is currently inactive. Comparing the velocities of Ice Streams B and C, Whillans and Van der Veen (1993) found differences of as much as two orders of magnitude. The lower reaches of Ice Stream C are stagnant, but further upstream, velocities increase to between 1 and 13 m a^{-1} and may be as high as 40 m a^{-1} in the upper reaches. This longitudinal velocity pattern led Retzlaff and Bentley (1993) to infer a wave of stagnation whereby the lower regions shut down first and stagnation propagated upstream. They also suggested that the timing of Ice Stream C shut-down was between 130 and 100 years ago, and that it may be related to changes in the water pressure at the bed. This is exactly what Anandakrishnan and Alley (1997a) hypothesised in their 'water piracy' theory, describing how the lubricating water in the lower reaches of Ice Stream C was diverted into neighbouring Ice Stream B, producing localised sticky spots at the bed.

In conclusion, the transient nature of pure ice streams represents yet another major challenge to ice stream research and a successful dynamic model of the Siple Coast ice streams must be able to explain why Ice Stream C is currently inactive (Bentley, 1987).

2.4. Characteristics of Topographic Ice Streams.

2.4.1. Location of Topographic Ice Streams.

This group of ice streams refers to Antarctic and Greenland ice, excluding those on the Siple Coast of West Antarctica and other potential pure ice streams elsewhere (e.g. Fahnestock *et al.*, 1993). These ice streams occupy fjord-like channels with beds which may lie up to 2 km below sea level. If the bedrock walls are visible at the surface then they are termed outlet glaciers. Unfortunately, topographic ice streams are often intermediate in character between ice streams and outlet glaciers (see Section 2.2.1). For this reason the two terms are often interchanged in the literature.

Although topographic ice streams drain large parts of contemporary ice sheets, data from these drainage features are few in comparison with pure ice streams. Nevertheless, a handful of topographic ice streams have been investigated and data from these are included in Table 2.1

Two of the more widely studied topographic ice streams are Pine Island Glacier and Thwaites Glacier. Their dimensions are very similar to the Siple Coast ice streams with Pine Island Glacier around 30 km wide and 200 km in length (Jenkins *et al.*, 1997). Both ice streams terminate as ice tongues in Pine Island Bay and drain a huge amount of ice from Ellsworth Land in West Antarctica, see Figure 2.1. Pine Island Glacier alone drains approximately 200,000 km^3 of the West Antarctic Ice Sheet (Jenkins *et al.*, 1997) and Thwaites Glacier is the fastest flowing ice stream in West Antarctica, reaching velocities of up to 3400 m a^{-1} (Rosanova *et al.*, 1998).

These two ice streams have attracted the attention of glaciologists for two main reasons. Firstly, they are thought to be particularly vulnerable to potential instabilities because they do not nourish an ice shelf. Secondly, they occupy an area which some workers have suggested may be subject to regional climate warming (Hughes, 1981).

Rutford Ice Stream is another topographic Ice Stream which drains the West Antarctic Ice Sheet, see Figure 2.1. It flows into the Ronne Ice Shelf and is of concern because much of its bed lies below sea level (Doake *et al.*, 1987). It is also of interest because a neighbouring ice stream (the Carlson Inlet) is almost stagnant and this represents a situation largely analogous to neighbouring Ice Streams B and C on the Siple Coast. Like Pine Island Glacier and Thwaites Glacier, Rutford Ice Stream has a similar configuration to the pure ice streams of the Siple Coast. Its width varies between 18.5 and 25.5 km and its length is well over 150 km (Doake *et al.*, 1987). Its thickness and velocity are also comparable to Siple Coast ice streams, see Table 2.1.

Other topographic ice streams drain both the East and West Antarctic Ice Sheets, but many of these closely resemble outlet glaciers. The characteristics of some of these outlet glacier/ice streams are reviewed by Bentley (1987), to which the reader is referred for further information.

Perhaps the most widely studied topographic ice stream is Jakobshavns Isbræ in south-west Greenland. The drainage basin of this ice stream reaches 550 km inland, but its dimensions are much smaller than the Siple Coast Ice Streams. Inland from the grounding zone it extends for about 90 km and is only around 6 km wide. The ice stream terminates as an ice tongue confined by a deep fjord but most of the ice stream is bordered by heavily crevassed shear margins.

Jakobshavns Isbræ differs from the Siple Coast Ice Streams because some of the surface of the ice stream melts. Despite this, ice thickness of between 1.9 and 2.6 km are comparable, and much of the bed is below sea level. However, unlike Siple Coast Ice Streams, surface slopes are much steeper and as a consequence, exceptionally high driving stresses (200-300 kPa) dominate its movement. This results in high velocities, varying from 4 to 7 km a^{-1} along its length (Echelmeyer *et al.*, 1991). Jakobshavns Isbræ represents the fastest flowing ice stream in the world and has been measured to be flowing at up to 8360 m a^{-1} (Lingle *et al.*, 1981). Moreover, this ice stream drains around 6.5% of the total Greenland Ice Sheet, a remarkably high figure given its relatively small size.

In summary, topographic ice streams differ from large outlet glaciers because they are generally faster and because they are bordered by slower moving ice. They are similar to pure ice streams in terms of their dimensions and velocities, but their location is controlled by the bedrock troughs in which they lie. They differ from pure ice streams in that they generally have higher driving stresses (50-200 kPa) and steeper surface slopes. For these reasons, their flow mechanisms may be different to those operating beneath pure ice streams.

2.4.2. Mechanisms of Flow.

Paterson (1994) suggested several reasons why basal sliding is the predominant flow mechanism under most topographic ice streams, particularly those in East Antarctica. If ice deformation was important, velocity would be expected to correlate closely with the pattern of driving stress. However, velocity nearly always increases all the way to the grounding line, whereas driving stress maxima usually lie approximately 100 km from the coast (see Bentley, 1987). Alternatively, the high driving stresses of topographic ice streams could not be supported by water saturated till (Paterson, 1994).

In support of these arguments, basal sliding has been shown to account for the majority of the flow of Jutulstraumen, a topographic ice stream in Dronning Maud Land, East Antarctica. Høydal (1996) carried out a force balance study on this ice stream and concluded that high basal shear stresses and warm basal temperatures imply that basal sliding is the dominant flow mechanism. However, it has been demonstrated that basal sliding does not account for the fast flow of all topographic ice streams, and it is more likely that the exact flow mechanisms vary both along and between individual ice streams. For example, ice deformation probably contributes to the flow of topographic ice streams in the upstream reaches, whereas basal sliding may be predominant in the lower regions. This has been shown to be true for

Jakobshavns Isbræ where basal sliding may be important near the grounding line but evidence suggests that internal deformation of the basal ice layer accounts for the majority of the flow under this ice stream (Iken *et al.*, 1993). This process is known as 'thermally enhanced creep'.

The arguments outlined above would appear to indicate that the dominant flow mechanisms beneath topographic ice streams are partitioned between basal sliding and ice deformation. However, seismic studies beneath Rutford Ice Stream indicated that a deforming till may be contributing to the forward motion of this ice stream, analogous to the situation inferred from beneath the Siple Coast ice streams. Data from seismic reflection lines both along and across Rutford Ice Stream indicated that the characteristics of the bed material were highly variable (Smith, 1997). In some areas, the seismics revealed a highly saturated till which was probably experiencing pervasive deformation. Other regions of the bed are not underlain by soft sediments and it was speculated that basal sliding may be the dominant flow mechanism in those areas, particularly upstream. The unusually low basal shear stress (40 kPa) below Rutford Ice Stream (Frolich and Doake, 1988) indicated that internal ice deformation can be eliminated as a possible flow mechanism beneath this topographic ice stream.

In summary, the characteristics of topographic ice streams vary to a great extent and a variety of flow mechanisms have been inferred to account for their rapid velocity. The majority of topographic ice streams are characterised by high driving stresses (50-200 kPa) and ice deformation and basal sliding can be assumed to be predominant. However, some topographic ice streams (such as Rutford Ice Stream) appear to have unusually low driving stresses. Here, ice deformation can be discounted and basal sliding or till deformation may be dominant.

2.4.3. Resistive Stresses on Topographic Ice Streams.

The resistive stresses on topographic ice streams have often been assumed to be dominated by friction from the bedrock valleys in which they lie (McIntyre, 1985). This has been shown to be true for Pine Island Glacier where measurements indicated that the velocity is significantly slower towards the margins (Lucchitta *et al.*, 1995). However, a few studies would appear to indicate that the force balance of other topographic ice streams is far more complex.

Frolich and Doake (1988) surveyed three transverse velocity profiles across Rutford Ice Stream. Their results indicated that the majority of shear deformation occurs in distinct boundary layers at the margins, whereas shear stress gradients in the centre of the ice stream are close to zero. This led them to conclude that friction at the sides of the ice stream is restricted to narrow zones whereas the basal shear stress supported most of the driving stress in the centre of the ice stream.

Back pressure is viewed as negligible on those ice streams which terminate in open water conditions but Echelmeyer *et al.* (1991) detected a pinning point 3 km upstream from the calving front of Jakobshavns Isbræ. This pinning point can be identified by an ice rumple at the surface of the ice tongue. Velocity profiles indicated a marked decrease in fast ice flow around this rumple and Echelmeyer *et al.* (1991) speculated that it provided considerable back stress to the ice stream. This is supported by evidence which indicated that the ice stream retreated 30 km in 100 years up until the early 1960's but has since remained relatively stable (see Sohn *et al.*, 1998).

In summary, the resistive stresses acting on topographic ice streams are far more complex than traditionally thought. The assumption that the majority of the resistance is provided by the bedrock troughs in which they lie may be a crude oversimplification.

2.4.4. Transition from Ice Sheet to Ice Stream Flow.

To investigate the transition from sheet flow to stream flow in many of Antarctica's outlet glaciers, McIntyre (1985) compared data from aircraft altimetry and satellite imagery with radio echo-sounding. He found that the transition from slow flow to fast flow is abrupt and often coincides with a bedrock step. While this may be true for a large number of outlet glaciers and some topographic ice streams (such as Thwaites Glacier), it does not hold true for all topographic ice streams.

Echelmeyer *et al.* (1991) found that the longitudinal profile of Jakobshavns Isbræ showed no oversteepening at the transition from sheet flow to stream flow. Rather, the transition from sheet to stream flow in topographic ice streams probably varies greatly between ice streams. More importantly, nearly all topographic ice streams attain their high velocities as a result of the bedrock troughs in which they lie. These subglacial troughs preferentially channel thicker ice and lead to large zones of convergent ice flow, warmer ice and hence rapid velocities.

2.5. Ice Stream Modelling.

Ice sheet modelling experiments are stimulated by the need to predict the response of contemporary ice sheets to future climate scenarios and the need to understand past ice sheet configurations and their role in the climate system. Geological and geomorphological evidence of former ice sheets can provide useful information as to the extent and timing of growth and decay, but often provide little information pertaining to glacial dynamics or ice thicknesses (Siegert, 1997). The importance of modelling lies in its potential to deliver reconstructions of ice sheets which can provide information on ice thicknesses, velocities, dynamics and the timing of their growth and decay over whole (100,000 year) glacial cycles.

Since the realisation that ice streams have a profound effect on ice sheet configurations, the focus of ice sheet modelling has begun to address their behaviour. However, modelling ice streams is hampered by three main problems. Firstly, there is no valid sliding law to accurately describe the basal conditions of an ice stream. Secondly, it is difficult to model the transition between an ice stream and an ice shelf. Thirdly, the incorporation of ice streams has often been beyond the resolution of most ice sheet models. The aim of the following sections, is to use selected examples from the literature to briefly emphasise the problems of modelling ice streams. A comprehensive review of ice stream modelling falls outside the scope of this chapter. A useful overview of ice sheet modelling is provided by Hindmarsh (1993) and modelling the West Antarctic ice streams is the focus of several papers in Van der Veen and Oerlemans (1987), to which the reader is referred.

2.5.1. Modelling the Basal Processes that Facilitate Fast Flow.

2.5.1.1. Basal Sliding.

Basal sliding is thought to occur when the water melted at the ice sheet bed accumulates to a sufficient depth as to provide substantial lubrication. Most early models of the Siple Coast ice streams incorporated sliding laws generally considered applicable to fast glacier flow (Bentley, 1987). Weertman (1957) described how fast sliding would result when the subglacial water layer became thick enough to overcome obstacles at the bed. Since this seminal work, a variety of sliding laws have been developed in the literature concerning the basal modelling of ice streams.

Weertman and Birchfield (1982) prescribed a basal water thickness of around 10 mm at the grounding line, calculated from the combined effects of the geothermal heat flux and frictional warming. Their velocity results were consistent with observed velocities from the Siple Coast ice streams. However, the relevance of this model is questioned by the fact that it is only applicable to the grounding line region because subglacial water thicknesses probably decrease substantially upstream (Bentley, 1987).

In short, a variety of different sliding laws have been developed but none appear to be valid over a range of contexts at both the spatial and temporal scale. Moreover, the discovery of a deforming till beneath Ice Stream B led Alley *et al.* (1987b) to present a model based on entirely different physical principles, omitting basal sliding altogether.

2.5.1.2. Till Deformation.

Having inferred a layer of till of several metres in thickness beneath Ice Stream B, Alley *et al.* (1987b) attempted to model the flow using a one dimensional model which assumed that the till deforms like a Newtonian viscous material. It is also assumed that the ice moves as a block with no internal shearing and that there is no sliding between the ice and till. The mass conservation of ice and till is a fundamental component of the model and so it is suggested that shear deformation transports the till downstream, forming a till delta at the grounding line (see Alley *et al.*, 1989). An essential assumption is that the till is continually being replenished at the bed. Little is known about this process and until there are more data available on the mechanical properties of water saturated till, the relevance of a flow model based on till rheology is doubtful (Paterson, 1994). Furthermore, recent laboratory tests have suggested that some tills display plastic rather than viscous behaviour (Iverson *et al.*, 1998). If this is the case, the relevance of a flow model based on viscous behaviour may have to be re-assessed.

2.5.1.3. Thermally Enhanced Ice Deformation.

It has been inferred that ice streams which lie in deep topographic troughs and have high driving stresses may flow by thermally enhanced deformation in a thick basal ice layer (Section 2.4.2). This mechanism has been inferred from borehole observations and temperature measurements beneath Jakobshavns Isbræ in south-west Greenland (Iken *et al.* 1993). In the centre of the ice stream, the minimum temperature lies at a remarkably shallow depth and it has been invoked that the whole of the bed is at melting point.

In order to investigate how this unusual temperature distribution may have developed, Funk *et al.* (1994) modelled the flow fields and temperature distribution within the ice stream. The model accounted for the actual topographic, dynamic and climatic controls along a central flowline of the ice stream but it ignored the three-dimensional flow of ice. This produced 'undistorted' temperature profiles which could be compared with the actual profiles within the ice stream and allowed an estimate of the variation in vertical straining at different depths.

Model results from Funk *et al.* (1994) indicated that vertical straining plays a major role in the fast flow of Jakobshavns Isbræ, confirming that a basal layer of temperate ice exists which results from the large ice deformation heat production near the bed. Calculations suggest this layer may be as deep as 270 m in the ice stream, as opposed to only 30 m in the adjacent slower flowing ice (Funk *et al.*, 1994).

If plausible, this mechanism may play an important role in the fast flow of other ice streams which lie in over-deepened bedrock troughs. However, a lack of observations from other topographic ice streams means that we are still unsure as to how representative Jakobshavns Isbræ is. Furthermore, direct verification of this flow mechanism is required and this may only be resolved by more advanced three-dimensional models (Funk *et al.*, 1994).

2.5.2. Modelling Ice Stream/Ice Shelf flow.

The transition between floating and grounded ice plays a major role in ice stream functioning and the grounding line position is viewed as a key control on ice sheet stability. However, it is very difficult to couple ice stream movement with ice shelf movement and this represents the second major inhibitor to ice stream modelling. This is because the physics behind the transition from ice stream to ice shelf flow are poorly understood and assumptions are often made which oversimplify the actual processes.

Most models assume that downstream of the grounding line ice movement is primarily controlled by ice shelf spreading, whereas upstream of the grounding line, movement is largely governed by basal shear stresses (Bentley, 1987). However, observations suggest that the force balance of an ice stream is far more complex (see Section 2.3.3) and a characteristic of the models which ignore this complex behaviour is that they all predict unstable retreat in the absence of an ice shelf (Bentley, 1987).

Using a two dimensional flow model, Van der Veen (1985) examined the coupling of an ice sheet and ice shelf. It was assumed that there was a linear decrease in basal shear from a transition zone prescribed a known distance upstream of the grounding line. The model results indicated that when the ice shelf was removed, the grounding line advanced concomitant with a lowering of the surface profile. However, the reduction in ice volume was negligible and the ice sheet returned to a new steady state. The results of this model are strongly dependent on the choice of width of the transition zone, which varies between 0 and 150 km Herterich (1987). In Heterich's (1987) model, basal sliding was included and the transition zone was much longer. Alternatively, with ice deformation alone, the change from ice sheet to ice shelf was relatively abrupt. However, as Paterson (1994) noted, unrealistic mass balances are required to maintain a constant ice thickness profile in this model and as yet, an accurate model of the potentially critical zone between an ice stream and ice shelf remains elusive.

2.5.3. Ice Streams in Ice Sheet Models.

Traditionally, the spatial resolution of ice sheet models, and in particular, the more comprehensive three dimensional models, prevented the incorporation of ice streams. On the other hand, two dimensional models are generally simpler to compute and permit a greater degree of freedom to perform a wide range of simulations.

In an attempt to understand the dynamics of the ice streams and outlet glaciers in a cross section from Pine Island Glacier to Ice Stream B, McInnes and Budd (1984) used a two dimensional model which was parameterised with contemporary surface and bedrock elevations, accumulation and temperature data. By incorporating a combination of internal deformation and appropriate sliding velocities (with prescribed strain rates), ice thickness profiles were successfully simulated and results indicated that the dynamics of the ice streams were strongly influenced by ice thicknesses and thinning near the grounding line. This simple model provided the basis for several other experiments investigating the dynamics of the West Antarctic Ice Sheet (e.g. Budd *et al.*, 1985) but was limited by the two-dimensional realm.

Computational advances and modelling adaptations have now begun to permit the incorporation of ice streams in three-dimensional models. In order to gain a better understanding of the mechanisms which govern the location and behaviour of the Siple Coast ice streams, Payne (1998, 1999) used a three-dimensional thermomechanical model of the West Antarctic Ice Sheet. This model employed a 20 km grid which permitted the incorporation of ice streams whose widths typically exceed 30 km (see Table 2.1). The model was forced by contemporary climate conditions, snow accumulation and bedrock topography and these boundary conditions were held constant to isolate the effects of internal feedback mechanisms. For the same reason, grounding line migration was ignored. The model was simplified so that basal temperatures control the amount of basal slip with respect to the driving stress. Furthermore, till deformation and subglacial hydrology are not included in the model.

The success of Payne's (1998) model lies in the fact that it can accurately replicate much of the behaviour of the Siple Coast ice streams. As Payne (1998) concluded, their behaviour can be related to purely thermomechanical interactions and topography. A small perturbation in velocity is amplified by strain-heating to produce initiation of an ice stream. However, the model suffers from two fundamental problems which Payne (1998) readily acknowledged. Firstly, the physics of ice stream flow are poorly represented. The modelled ice streams arise solely from the interaction between ice temperature and ice flow which reduces ice viscosity and increases basal melting. However, the Siple Coast ice streams have exceptionally low driving stresses and a very well lubricated bed, and this would suggest that frictional heating from internal deformation and basal sliding are much more limited than the model predictions. Secondly, Payne's (1998) model predicted that ice stream shut-down can only occur through the process of basal freezing. This conflicts with field evidence from the stagnant Ice Stream C which suggests that its bed is not frozen.

In conclusion, although good mechanical models of ice sheets exist, the incorporation of ice streams poses a major technical challenge. The physics of ice streaming is poorly understood and it is difficult to simulate processes which govern the temporal variations in their activity.

2.6. Ice Streams and Climate Change.

The main objective of much of the research outlined in this chapter lies in understanding the operation and controls on ice streaming so that we may predict the response of contemporary ice sheets to future climate perturbations (see Chapter 1). Of particular concern is the possible instability of the West Antarctic Ice Sheet (WAIS) to potential global warming scenarios. Indeed, it was Mercer (1978) who first hinted at the potential severity of the response of the WAIS and the "threat of disaster" it could pose. Mercer's (1978) argument suggested that the removal of the Ross and Ronne-Filchner Ice Shelves (Figure 2.1) which fringe the WAIS could lead to grounding line retreat and drawdown of the marine-based ice sheet. In a sobering conclusion he speculated that even conservative model estimates of global warming will result in a rapid disintegration of the WAIS with an associated 5 m rise in sea level over the next 50 years. However, the predicted response of the WAIS and its ice streams is a controversial subject and opinions vary widely.

The problem is that our understanding of ice stream behaviour is still rather limited. It is still unclear whether ice stream behaviour is; (a), a system of inherent instability, (b), a transient system which suggests that they may switch off prior to a total collapse, or (c), a system which results from a steady state ice sheet linking the interior ice to ice shelves (Oppenheimer, 1998). This lack of understanding was emphasised by Bindschadler (1997) who commented that "opinions differ widely about the probability of even partial collapse" of the WAIS.

It has been argued that there is no compelling evidence (theoretical or observational) to suppose a major collapse of the WAIS more than once every 100,000 year glacial cycle (Bentley, 1997). In contrast, Bindschadler (1998) cited evidence of at least one previous collapse since the ice sheet formed (see Scherer, 1993) and argued that the WAIS is far from stable. Further support for this argument lies in the recent observations which depict a highly dynamic Siple Coast ice stream system (discussed in Section 2.3.5).

Recent evidence from the Ross Sea embayment presented by Conway *et al.* (1999) led them to hypothesise that the Siple Coast ice streams have been retreating steadily since the early to mid-Holocene. This retreat pattern was revealed by the dating of grounding line positions and it is suggested that it may not be related to anthropogenic warming but may be a predetermined retreat triggered during the early Holocene. This implies that the WAIS could collapse without external forcing.

The problem with most of these theoretical arguments is that they are often speculative. This is because we have never been able to observe the collapse of an ice sheet. Instead, we must turn to the often fragmentary evidence left behind when the ice sheets in the northern hemisphere disintegrated at the end of the last ice age. This evidence from the last deglaciation supports the notion that ice streaming and climate are inextricably linked. To examine this link over longer time scales than contemporary ice streams permit, involves the investigation of those ice streams which drained the now extinct ice sheets from the last ice age. Once identified, these ice stream beds provide an unprecedented opportunity to glean information about their functioning over whole cycles of operation.

2.7. Summary and Conclusions.

Ice streams are arguably the most dynamic component of contemporary ice sheets. They are arteries of fast flowing ice bordered by slower moving ice and can be broadly classified as either pure or topographic ice streams, depending on the influence of the underlying topography. Although this definition is clear in principle, some ice streams are intermediate in character. Furthermore, topographic ice streams often resemble large outlet glaciers and the two terms are often interchanged in the literature.

Our current understanding suggests that pure ice streams are limited to the Siple Coast of West Antarctica, although a "rapid flow feature" has been identified in north-west Greenland (Fahnestock *et al.*, 1993). These ice streams are characterised by low driving stresses, low basal shear stresses and concave ice surface profiles. Seismic and borehole observations have shown that at least one of these ice streams is underlain by a layer of highly saturated till which is deforming (Blankenship *et al.*, 1986; Engelhardt *et al.*, 1990) and it has been inferred that their fast flow may be attributable to a deformable bed. However, the subglacial drainage system beneath these ice streams is complex and basal sliding may also be important (Engelhardt and Kamb, 1998).

The resistive stresses along these ice streams are poorly understood but are thought to be composed of basal drag (including localised sticky spots), side drag and the back pressure provided by ice shelves. A further characteristic of pure ice streams is that they appear to be dynamic features, displaying both spatial and temporal variations in activity.

In contrast to pure ice streams, topographic ice streams are fixed within the bedrock troughs in which they lie. They generally have higher driving stresses, higher basal shear stresses and often attain higher velocities compared to pure ice streams. Topographic ice streams drain large parts of the East and West Antarctic Ice Sheets and the Greenland Ice Sheet. Traditionally, topographic ice streams were thought to flow by rapid basal sliding but evidence suggests that at least one of these ice streams (Jakobshavns Isbræ in Greenland) flows by thermally enhanced ice deformation. Furthermore, a layer of till similar to that found under Ice Stream B has been inferred from seismic evidence under a topographic ice stream in West Antarctica (Rutford Ice

Stream). Like pure ice streams, the resistive stresses on topographic ice streams (side drag, basal drag and ice shelf back-pressure) differ both along and between individual ice streams.

In summary, the basic characteristics of ice streams in the broadest sense are;

- large dimensions (>20 km wide, > 150 km long),
- highly convergent onset zones feeding a main channel,
- rapid velocities (>300 m a^{-1}),
- abrupt lateral shear margins,
- deformable bed conditions and/or high basal water pressures or enhanced ice deformation

Modelling ice streams has proved a major challenge. This is because, as yet, there is no valid sliding law which accurately represents the conditions at the base of an ice stream. At the ice sheet scale, ice streams have traditionally been overlooked because of their small size in relation to typical modelling grid sizes. More recent models have incorporated ice streams but they suffer from a lack of understanding of the physics of ice stream flow and a major technical challenge lies in modelling the transition from ice stream to ice shelf flow.

The link between ice streams and climate change is the driving force behind contemporary ice stream research. If the West Antarctic Ice Sheet (WAIS) is going to collapse, ice streams will play a key role. Many attempts have been made to model the response of the WAIS to future climate perturbations, particularly increased global temperatures. Unfortunately, these modelling experiments are limited by both a lack of data on basal processes, and a lack of suitable ice stream behavioural data at appropriate timescales with which to test the model functioning. For contemporary ice streams, we are limited to direct observations over decadal timescales. Attempts to model their response over longer timescales (100-1000 years) suffer from a paucity of data. Part of this stems from the fact that we have never been able to observe the behaviour of ice streams during ice sheet collapse. To fully understand the response of ice streams over these timescales, we must rely on the often fragmentary geological and geomorphological evidence of their activity during the last ice age.

In his benchmark paper on Antarctic ice streams, Bentley (1987) concluded that field projects on the West Antarctic ice streams will help us to better understand the dynamic principles which control them and "only then will it be possible to predict the future of the Antarctic and Greenland Ice Sheets, and understand the behaviour of the great ice sheets of the past" (p. 8856). While this certainly holds true, this thesis will offer the paradox that a better understanding of contemporary ice streams can be gained by investigating those ice streams which operated in the last great ice sheets. This approach has been undertaken by a large number of workers and the following chapter will introduce the significance of palaeo-ice streams followed by the first ever comprehensive review of hypothesised palaeo-ice stream locations.

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Chapter 3: Palaeo-Ice Streams: A Review.

3.1. Introduction.

This chapter begins by outlining the significance of palaeo-ice stream research. Because ice streams have a profound effect on ice sheet configuration, we need to know where and when they operated in order to accurately reconstruct former ice sheet histories. In addition, episodes of palaeo-ice streaming have been shown to have a considerable impact on ocean circulation and climate and their profligate sediment flux has implications for drift prospecting and potential geohazards.

If we can find former ice streams, we have an opportunity to advance our understanding of these dynamic glaciological phenomena. Many workers have recognised this and the main aim of this chapter is to review the evidence used to identify a large number of the most pertinent palaeo-ice streams in the literature. This represents the first ever synthesis of palaeo-ice stream research.

Evidence of palaeo-ice stream activity is often obscured or modified and there are several, often inherent, problems in identifying their former locations. These problems are summarised towards the end of the chapter and it is concluded that a lack of objectivity is the main inhibitor to palaeo-ice stream research. Put simply, we have no clear criteria to identify palaeo-ice streams.

3.2. The Importance of Palaeo-Ice Streams.

3.2.1. Palaeo-Ice Streams and Ice Sheet Reconstruction.

The large ice flux of ice streams has a profound effect on ice sheet configuration, including drainage basin and ice divide location and local and regional ice sheet topography. Therefore, we need to know where and when they operated in order to accurately reconstruct ice sheet histories. This was only fully recognised as late as

1981 when Denton and Hughes (1981) incorporated numerous ice streams into their reconstruction of the former northern hemisphere ice sheets, see Figure 3.1.

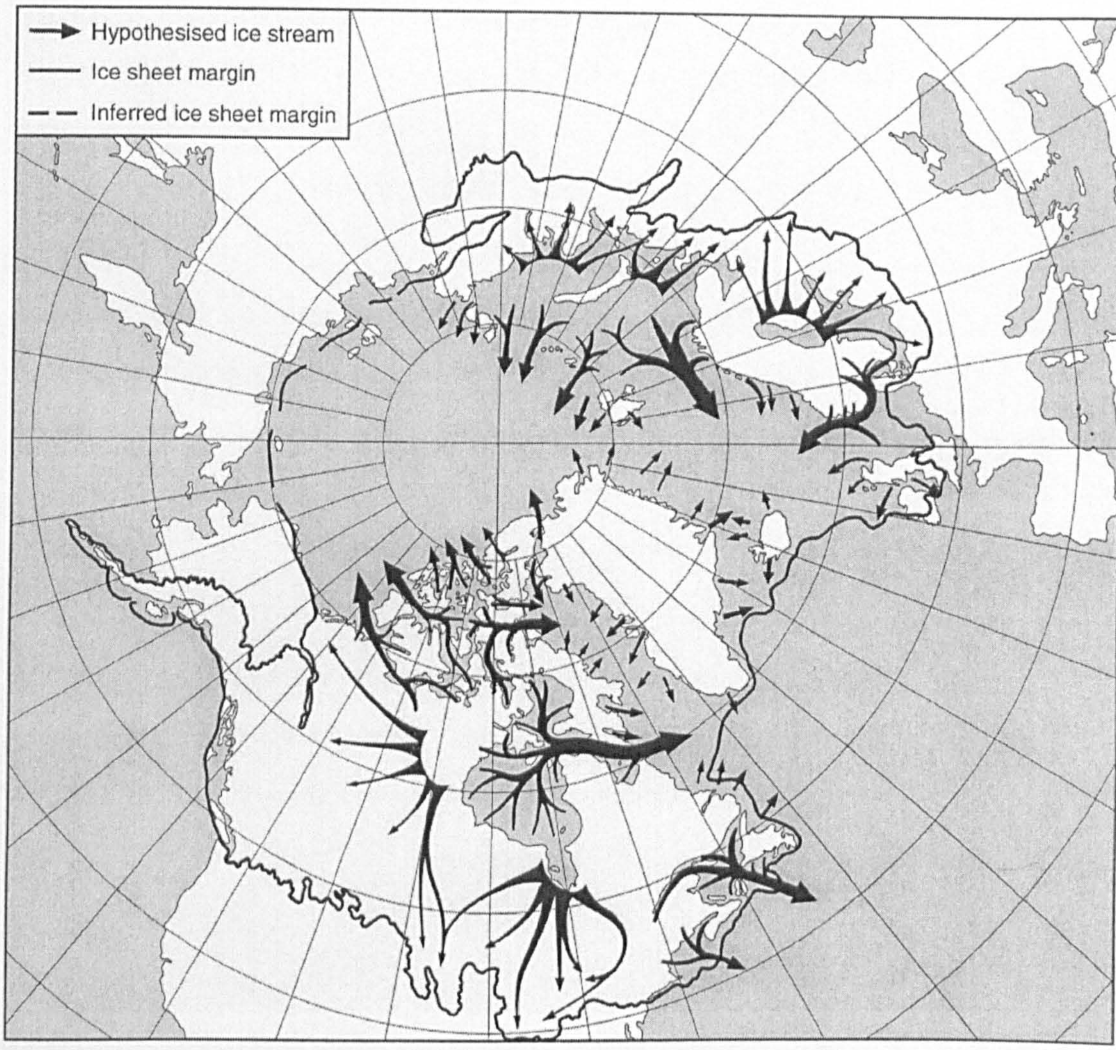

Figure 3.1. Potential locations of major ice streams (black arrows) in northern hemisphere ice sheets, as predicted by Denton and Hughes (1981). These were prescribed mainly from marine troughs and ice sheet geometry, (from Denton and Hughes, 1981).

While this realisation represented a major advance, the validity of many of the locations in Figure 3.1 remains open to question. Accurately locating the positions of former ice streams is of paramount importance when reconstructing former ice sheets and their elucidation is likely to radically alter our views of ice sheet geometries and their rapidity of change.

3.2.2. Palaeo-Ice Streams and Climate Change.

The marine geological record has provided important evidence of the role of ice streams in forcing 'abrupt' (decadal to millennial) climate change. Episodes of ice streaming, particularly from the eastern margin of the former North American (Laurentide) Ice Sheet, were responsible for large iceberg discharge events into the North Atlantic between 10 and 60 thousand years ago (Bond *et al.*, 1992; Broecker *et al.*, 1992). Evidence for these events comes from layers of ice-rafted debris found in ocean cores, known as Heinrich events.

Andrews and Tedesco (1992) have specifically linked the carbonate detritus associated with the two most recent Heinrich events to sedimentary rocks eroded by an ice stream in Hudson Strait. It has been postulated that the profligate influx of freshwater resulting from this ice stream was sufficient to cause changes in sea surface temperature and salinity, which had a considerable impact on the ocean circulation and northern hemispheric climate (Broecker, 1994).

Although the trigger for the ice streaming itself remains unclear, it is now widely recognised that ice streams are instrumental in driving some of the most abrupt changes in high latitude climate and ocean circulation (MacAyeal, 1993; Bond and Lotti, 1990; Andrews, 1998). Finding former ice streams is of great importance in terms of elucidating their role in the ocean-climate system.

In addition, the topographic effect of ice sheets can also influence climate by their ability to deflect atmospheric circulation patterns. Manabe and Broccoli (1984) have demonstrated the important coupling between ice sheet elevation and atmospheric circulation with respect to the Laurentide Ice Sheet, and Shin and Barron (1989) have demonstrated that North Atlantic climate is highly sensitive to ice sheet elevation. Because ice streams have the ability to rapidly drain large portions of ice sheets, they play a critical role in controlling overall ice sheet thicknesses and therefore, may indirectly alter atmospheric circulation.

3.2.3. Palaeo-Ice Stream Beds and Basal Processes.

Current research in West Antarctica is striving to ascertain the basal characteristics and processes of the ice streams present (see Section 2.3.2). While providing invaluable insights into the nature of the ice stream bed, borehole investigations are limited by the scale of the investigation and seismic studies suffer from a lack of detailed resolution. Ideally, it would be possible to view the whole ice stream bed at a variety of scales. Studying former ice stream tracks allows us to do just that, and if we can confidently find a former ice stream bed, we have a perfect opportunity to investigate the basal characteristics on a variety of scales, from large scale mapping of geomorphology to micromorphological till analysis. For example, palaeo-ice stream beds can provide information with regards to the basal topography, the bed roughness, the hard bed to soft bed fraction and the geology and lithology, all of which are thought to be critical with respect to ice stream location and functioning.

3.2.4. Palaeo-Ice Streams and Sediment Transport.

Rapid velocities make some ice streams extremely powerful erosional agents. As such, marine-terminating ice streams may be the most important mechanism by which sediment is delivered to continental margins and their fluxes are comparable with the efficacy of the largest fluvial systems, despite the far shorter duration of operation (Elverhøi *et al.*, 1998). A consequence of this is that ice stream location and vigour determines the distribution and volume of major accumulations of sediments, or fans. The locations, deposition rates, sediment volumes and facies architecture have strong practical implications for mineral exploration and geohazard management because gravity driven slumping within these fans may lead to tsunamis. This is particularly relevant in the Arctic Ocean, 70% of which was bounded by major ice sheets during the last glacial and into which numerous ice streams drained, some of which have already been identified (described in Section 3.3.5).

It is clear then, that locating former ice streams holds much potential in terms of advancing our understanding of these dynamic glaciological phenomena. Such research feeds directly into our knowledge of ice stream behaviour in the Pleistocene Ice Sheets and in doing so, may help us to better understand the pivotal role of the ice streams that are currently governing the stability of contemporary ice sheets. For these reasons, much work has been undertaken in order to postulate on the locations of palaeo-ice streams in former ice sheets. The following section provides a review of the most pertinent palaeo-ice stream hypotheses in the literature.

3.3. A Review of Palaeo-Ice Stream Hypotheses.

3.3.1. Ice Streams in the Laurentide Ice Sheet.

The Laurentide Ice Sheet is one of the most widely studied of all palaeo-ice sheets and many former ice streams have been inferred. Marshall *et al.* (1996) constructed an ice stream likelihood map of Canada and North America based on regional topography and areas underlain by soft (deformable) sediments. This indicated that large parts of the Laurentide Ice Sheet appeared to have been predisposed to ice streaming, particularly around the southern Laurentide Ice Sheet margin.

Matthews (1974) estimated that many of the southern Laurentide lobes had extremely low surface gradients presumed to have arisen from low basal shear stresses. The deforming bed model presented by Boulton and Jones (1979) predicted fast ice flow in these lobes. Following this work, several studies have focused on locating the exact positions of ice streams, both spatially and temporally, along the southern Laurentide Ice Sheet margin. The evidence used to postulate these palaeo-ice streams is shown in Table 3.1 and Figure 3.2 maps their locations.

Hicock (1988) was one of the first to speculate on the role of ice streams within the southern Laurentide Ice Sheet lobes, specifically from the Great Lakes region, an area known to have been associated with deformation tills. The location of these ice streams was based on streamlined depositional landforms and areas of till distributed in belts or plumes (see Table 3.1). The till plumes are thought to coincide with the main trunk of the ice stream and splayed landform patterns are thought to reflect the divergent marginal areas.

Table 3.1. Hypothesised palaeo-ice streams of the Laurentide Ice Sheet and the main lines of evidence used in their identification.

Reference	Location of Ice Stream(s)	Dimensions	Timing	Principal Evidence for Ice Stream Activity.
1. Andrews *et al.* (1985)	Hudson Strait	not given explicitly	not given explicitly	Evidence for an ice stream based on; (a) bedrock trough in relation to ice sheet geometry, (b), areas of intense to moderate glacial scour, (c), areas composed of distinct till transported from known source areas, and (d), areas of extensive striations.
2. Dyke and Morris (1988)	Prince of Wales Island, Canadian Arctic	20 km wide and 100 km long	Last glacial maximum	'Boothia-type' dispersal plume indicated that ice flowing in the centre of the ice stream flowed much faster than the surrounding ice. Lateral shear moraine thought to have resulted from an abrupt lateral margin. Large drumlin field with highly convergent flow patterns at the head of the ice stream appear to feed the main channel; longest drumlins formed in the central part of the plume where presumably ice flow was fastest. Morphological zonation of streamlined landforms both across and along the ice stream depict the characteristic lateral variations in ice stream velocity. Drumlins at the head of the ice stream overprint older drumlins indicating a more erosive ice flow and hence higher velocities.
3. Hicock (1988)	Albany Valley, James Bay Lowlands	ca. 50 km wide ca. 600 km long	Last glacial maximum	Broad bedrock trough conducive to ice stream flow; Boothia-type dispersal train indicated that the ice stream margin was abrupt and that ice flowing on either side of the dispersal train was slower-moving. Drumlins with a splayed flow pattern depict a broad lobate stagnation zone. Calcareous lodgement tills show evidence of deformation which probably aided ice streaming.
4. Boyce and Eyles (1991)	Simcoe Lobe (Ontario)	not given explicitly	less than 13,000 yr BP	Based primarily on bedform geomorphology. Systematic downstream decrease in drumlin length/width ratios are thought to reflect decreased duration of subglacial sediment deformation towards the outer limit of the lobe. Fast flow was facilitated by the fine grained sediments with high pore water pressures and the presence of a high level water body at the terminus was conducive to removing ice.
5. Laymon (1992)	Hudson Strait	not given explicitly	Last glacial maximum	Large marine trough with submarine moraine features indicated the boundaries of a large ice stream. Abundant crag and tail and or stoss and lee features indicated that ice flowing into Hudson Strait changed direction abruptly and was incorporated into the rapid eastward flowing ice stream Convergent ice flow patterns at the head of Hudson Strait appeared to feed the main ice stream channel.
6. Kaufman *et al.* (1993)	perpendicular to the mouth of Hudson Strait	max. of 200 km wide and over 300 km long	9,900 to 9,600 yr BP	Ice-contact fans record the position attained during short-lived and dramatic margin advance across the axis of Hudson Strait. The dating of marine molluscs and foraminifera provided limiting ages for the timing of the advance which is constrained to around 300 years. Rapid ice flow is inferred from the extensive (200 km long) calving margin which could only have been maintained by fast ice flow. Striations and small-scale erosional bedrock features were used to infer the ice flow direction and the dispersal of erratics delimited the ice provenance. Seismic stratigraphic records from a shallow marine basin also identified a sedimentary wedge linked to two small end moraines which further delimit the ice stream advance.
7. Hodgson (1993; 1994)	Victoria Island, Canadian Arctic Archipelago	at least 80 km wide and 200 km long	between 10,400 and 9,600 yr BP	Drumlin field several hundred kilometres long containing highly attenuated bedforms thought to be indicative of fast ice flow (length width ratios of drumlins up to 20:1, lengths up to 10 km with heights of 10-25 m). Very abrupt margin of drumlin field thought to delimit the abrupt lateral margin of the ice stream. Massive extended drumlin ridges interpreted as lateral shear moraines thought to have resulted from fast ice contacting slower moving marginal ice. Rapid advance and hence high velocities constrained by the dating of marine deposits.

Table 3.1. continued.

Reference	Location of Ice Stream(s)	Dimensions	Timing	Principal Evidence for Ice Stream Activity.
8. Veilette (1997)	James Bay Ice Stream	ca. 200 km wide and 500 km long	ca. 9,000 yr BP	Ice flow indicators such as striations and streamlined landforms in addition to glacial transport data indicated an abrupt change in ice flow direction thought to mark the beginning of rapid ice flow down the James Bay corridor. This change in ice flow direction is evidenced by cross-cutting landforms. The northward migration of glacial Lake Ojibway may also have provided an environment conducive to the rapid removal of icebergs.
9. Patterson (1997; 1998)	Des Moines Lobe, Minnesota	max. 200 km wide and 900 km long	between 14,000 and 12,000 yr BP	Highly convergent topography in the onset zone. Broad zones of stagnation landforms, especially hummocky topography, in the outer 20-50 km indicative of ice stagnation and downwasting during retreat. Area of rapid flow inferred from ice sheet profile reconstruction and low driving stresses. Rapid advance of ice constrained by radiocarbon chronology.
10. Kaplan *et al.* (1999)	Cumberland Sound Ice Stream	ca. 75 km wide and 150 km long.	not given explicitly	Numerical modelling of the Baffin Island region indicated that given the necessary (but realistic) boundary conditions, the topographic trough provides an ideal location for ice streaming. A critical but reasonable ice sheet surface elevation is required to initiate ice streaming which is facilitated by prescribing basal sliding in key areas, specifically the onset zone of the ice stream. The modelled ice stream has a low surface slope comparable to contemporary West Antarctic ice streams and can be accounted for by field based Interpretations of the glacial history of the area.

Table 3.2. Hypothesised palaeo-ice streams of the Cordilleran Ice Sheet and the main lines of evidence used in their identification.

Reference	Location of Ice Stream(s)	Dimensions	Timing	Principal Evidence for Ice Stream Activity.
11. Hicock and Fuller (1995)	Skeena Valley/continental shelf, British Columbia	ca. 12 km wide and ca. 100 km long	Late glacial	Data on stone striae, clast pavements, fabrics, morphology, provenance, and the matrix grain size of diamictons provided substantial evidence for pervasively deforming bed conditions; a large valley, comparable in size to Antarctic valleys containing ice streams is thought to have channelled flow into this area of soft sediments and hence rapid ice flow.
12. Evans (1996)	Lillooet and Cheakamus Valleys, southern Coast Mountains, British Columbia.	not given explicitly	between 20,000 and 15,000 yr BP	In deep topographic troughs, the distribution of abraded rock landforms (whalebacks) and drumlins are though to have formed under deep, rapidly moving ice streams which slide over bedrock, exploiting rock weaknesses to produce streamlined features. The dimensions of these topographic ice streams are consistent with contemporary ice streams in East Antarctic and Greenland and it is hypothesised that the flow mechanism may have been analogous to Jakobshavns Isbræ in south-west Greenland whose movement was dominated by enhanced thermal creep deformation in a basal ice layer.

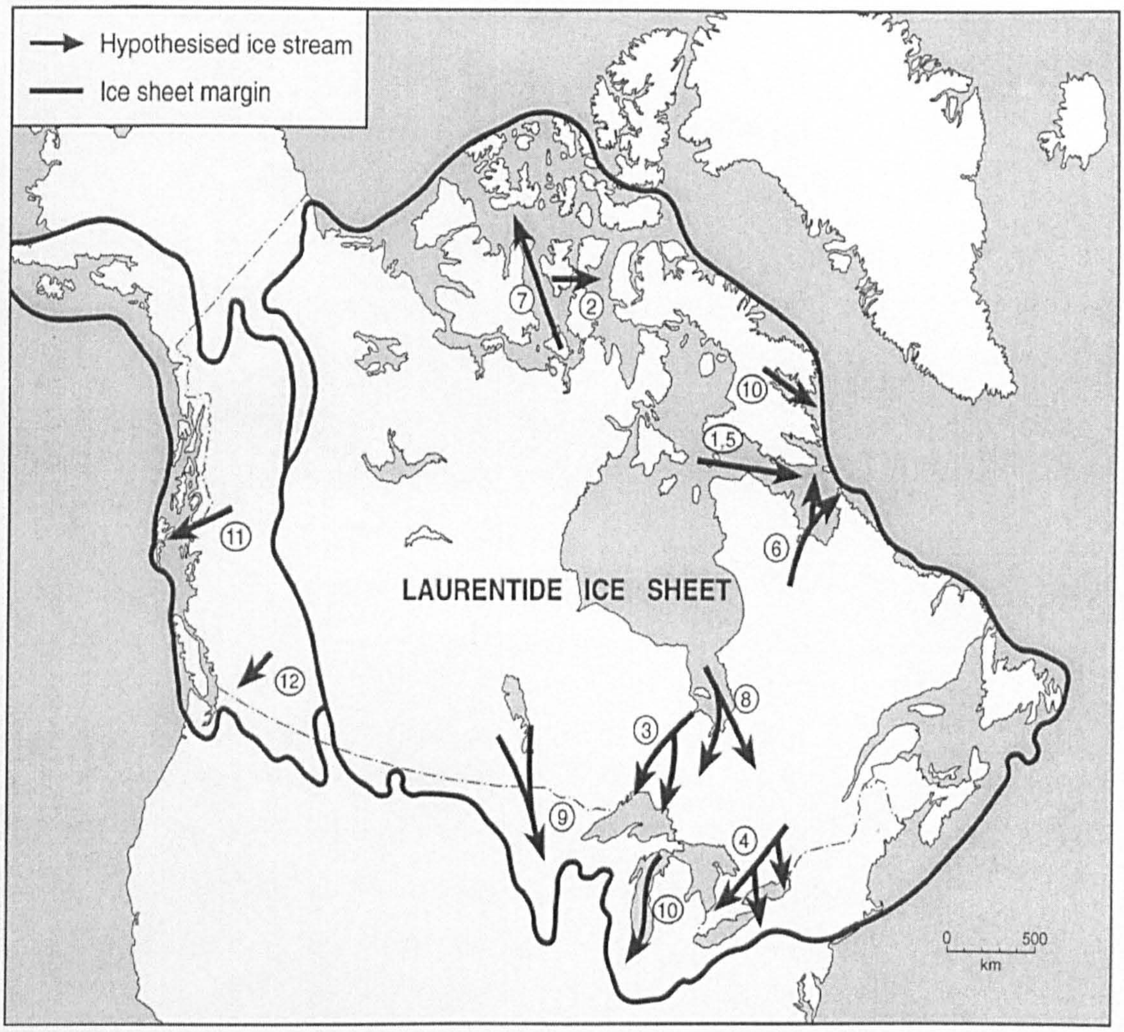

Figure 3.2. Hypothesised locations of selected ice streams from the Laurentide and Cordilleran Ice Sheets. Approximate locations of ice streams (not to scale) are depicted by a black arrow and the number refers to the corresponding citation and evidence outlined in Tables 3.1 and 3.2, respectively. Note that the ice sheet margin is depicted at full glacial conditions but that ice streams operated at various times. Ice sheet outline simplified from Dyke and Prest (1987b).

It can be seen from Table 3.1 that the presence of a deformable till within an extensive drumlin field also led Boyce and Eyles (1991) to infer an ice stream within the Simcoe Lobe of the southern Laurentide margin. They used borehole data coupled with morphometric analyses of the bedforms to infer that the ice stream rested on a deformable bed. Systematic decreases in drumlin elongation ratios towards the margins of the lobe were thought to reflect a decreased duration of sediment deformation. Nearby, further ice stream activity has been invoked to account for cross-cutting landforms and an abrupt shift in ice flow direction during deglaciation (Veillette, 1997), see Figure 3.2.

While several lobes have been cited as candidates for terrestrial ice streams, a fundamental problem is the lack of modern analogues. All contemporary ice streams are marine-based and this makes it difficult to conceptualise the configuration and behaviour of land-terminating ice streams. However, recent geomorphological mapping of the Des Moines lobe led Patterson (1997; 1998) to hypothesise that it may be the nearest terrestrial analogy to contemporary West Antarctic Ice Streams. She argued that the glacial geomorphology of the Des Moines lobe largely resembled the foregrounds of modern surging glaciers, (see Table 3.1). Close to the inferred ice stream terminus, transverse ridges are interpreted as crevasse squeeze features which grade laterally into hummocky moraine. Several of these marginal zones are identified, and it is postulated that the Des Moines lobe may have advanced and retreated several times, reaching its maximum extent between 14,000 and 12,000 yr BP (Patterson, 1997).

The northern and eastern sectors of the Laurentide Ice Sheet are very important because the present day straits and sounds would have preferentially channelled fast ice flow. Furthermore, the marine nature of the ice sheet margin was largely analogous to the margins of contemporary ice sheets drained by ice streams (e.g. the West Antarctic Ice Sheet). For example, Hudson Strait is thought to have a contained a huge ice stream (800 km x 150 km) which had a profound impact on the configuration of the Laurentide Ice Sheet (Andrews *et al.*, 1985; Laymon, 1992).

Three main criteria (shown in Table 3.1) were used by Andrews *et al.* (1985) to identify ice streams from the north-eastern Laurentide Ice Sheet. The Hudson Strait Ice Stream fulfilled all of these criteria and since the more recent work of Laymon (1992), the notion of an extremely large ice stream in Hudson Strait up to Late Glacial times is widely accepted. Laymon carried out extensive field investigations on the glacial geology of western Hudson Strait (the inferred onset zone of the ice stream) and found that ice flowing obliquely into the strait appeared to be diverted within less than 10 km of the coast. This indicated extreme flow convergence and analysis of glacial erratics supported the idea that ice was captured from a wide source area.

The presence of the Hudson Strait Ice Stream also has corroboratory evidence from sediment cores in the North Atlantic (see Section 3.2.2). Constraining the timing of delivery of the associated debris layers (Heinrich events) has allowed estimates of the ice stream's velocity which ranges from 4 to 37 km a^{-1} (see Dowdeswell *et al.*, 1995).

However, deglaciation of this area was not as simple as a calving front retreating along the Strait, because Kaufman *et al.* (1993) invoked further ice stream activity perpendicular to the mouth of Hudson Strait (see Figure 3.2).

Further north, a numerical model has demonstrated that another topographic trough, Cumberland Sound, could also have supported a marine-based ice stream (Kaplan *et al.*, 1999). Using geophysical, terrestrial and marine geological evidence to fix the boundary conditions of the ice sheet model, basal sliding was prescribed in specified areas and an ice stream developed within the channel. Although this ice stream would have been considerably smaller than the neighbouring Hudson Strait Ice Stream, it would have represented an important conduit linking the interior of the ice sheet to its marine margins.

Hypothesised former ice streams have also been proposed at the northern and north-western margins of the Laurentide Ice Sheet (Table 3.1). Glacial geomorphology and geology on Prince of Wales Island in the Canadian Arctic records a spectacular dispersal plume and Dyke and Morris (1988) provided several additional lines of evidence to suggest that this plume was formed by an ice stream flowing off the east coast of the Island (Table 3.1).

To the west of Prince of Wales Island (across M'Clintock Channel), Hodgson (1994) provided good evidence of ice stream activity on Storkerson Peninsula, Victoria Island. Detailed mapping of the surficial geology of the area (Hodgson, 1993) revealed a remarkable field of highly attenuated drumlins with an extremely abrupt margin. The ice stream bedforms trend in a south to north direction and represent some of the longest drumlins in the literature, with length to width ratios approaching 20:1.

3.3.2. Ice Streams in the Cordilleran Ice Sheet.

Hicock and Fuller (1995) suggested that several valleys may have acted as conduits for ice streams which drained the western Cordillera and debouched onto the continental shelf. One such ice stream is thought to have issued out from a valley, comparable in size to Antarctic troughs which carry contemporary ice streams and outlet glaciers (see Section 2.4.1). Table 3.2 shows the main lines of evidence used to postulate this ice stream and its location is shown on Figure 3.2.

Evans (1996) also invoked ice stream activity from topographic troughs in the Cheakamus and Lillooet valleys north of Vancouver. Table 3.2 shows the evidence used by Evans to identify these ice streams and their approximate location is shown on Figure 3.2.

3.3.3. Ice Streams in the British and Irish Ice Sheets.

Many questions concerning the configuration and behaviour of the British and Irish Ice Sheets remain unanswered but several possible ice stream locations have been identified. Table 3.3 shows the evidence used to postulate on the existence of ice streams within the British and Irish Ice Sheets and Figure 3.3 maps their locations.

Of particular relevance to palaeo-ice stream research is the debate concerning the nature and configuration of the North Sea lobe which is thought to have flowed down the east coast of England. The early modelling work of Boulton *et al.* (1977) could not explain the extension of ice down the east coast using steady-state assumptions and they invoked surging behaviour in this portion of the ice sheet. This led Eyles *et al.*, (1994) to postulate a North Sea Ice Stream which flowed down the eastern side of northern England and periodically surged onshore along the Holderness coast. They found evidence for a deformation till and suggested that it resulted from the periodic surging of an ice lobe over muddy marine sediments. Further evidence for ice stream activity is invoked from an arcuate belt of hummocky topography which largely resembled the foregrounds from contemporary surging glaciers in Spitsbergen.

On a smaller scale, several terrestrial ice streams are thought to have emanated from the Scottish Highlands during the Last Devensian glaciation (Merrit *et al.*, 1995). One such ice stream is thought to have resulted from the coalescence of several of these terrestrial ice streams to form a grounded, marine-based glacier; the Moray Firth Ice Stream. The evidence used to identify this ice stream is shown in Table 3.3 and its location is shown on Figure 3.3.

Table 3.3. Hypothesised palaeo-ice streams of the British and Irish Ice Sheets and the main lines of evidence used in their identification

Reference	Location of Ice Stream(s)	Dimensions	Timing	Principal Evidence for Ice Stream Activity.
1. Eyles *et al.* (1994)	North Sea Ice Stream	not given explicitly	not given explicitly	Till facies and sediments indicated deformation tills resulting from periodic surging of an ice lobe over muddy marine sediments. Arcuate belts of hummocky topography are thought to be indicative of a stagnation zone of the lobe and its landform assemblage is shown to be similar to the foregrounds of modern surging glaciers in Spitsbergen. Further evidence is provided by the fact that some ice sheet models can not explain this portion of the ice sheet using steady state conditions (e.g. Boulton *et al.*, 1977).
2. Merrit *et al.* (1995)	Moray Firth Ice Stream (Scotland)	not explicitly given	ca. 13,000 yr BP	Raised glaciomarine deposits and sediments provided evidence for an oscillating tidewater glacier in the Moray Firth. This is thought to be the outlet of one of several terrestrial ice streams emanating from the Scottish highlands. The rapid disintegration of this ice stream was punctuated by several stillstands caused by pinning points in the Moray Firth. Evidence for this comes from prograded till deltas deposited at the grounding line of the ice stream.
3. Knight and McCabe (1997); Knight *et al.* (1999); McCabe and Clark (1998)	Irish Ice Streams (north central Ireland) and Irish Sea Basin Ice Stream	up to 70 km long, width not given	between 15,000 and 14,000 yr BP	Ice streams are inferred from the differential modification of transverse (Rogen) moraine. Well preserved Rogen moraine are found outside the ice stream margins whilst ice stream activity has cross-cut, streamlined and drumlinised them in places. Some ice streams, e.g. the Armagh Ice Stream, were also responsible for the deposition of sediment in the form of sub-aqueous moraines. McCabe and Clark (1998) suggested that convergence of these ice streams in the Irish Sea basin led to a large oscillating marine-based ice stream and dating of foraminifera indicates that after reaching its maximum extent at around 14,000 yr BP deglaciation was rapid.

Table 3.4. Hypothesised palaeo-ice streams of the Scandinavian Ice Sheet and the main lines of evidence used in their identification

Reference	Location of Ice Stream(s)	Dimensions	Timing	Principal Evidence for Ice Stream activity.
A – 0 (Figure 4); Punkari (1993; 1995; 1997); Kleman *et al.* (1997)	11 ice streams in the Scandinavian Ice Sheet including the Baltic Sea and Norwegian Channel Ice Streams	10's of kms wide, 100's of kms long	During deglaciation	Analysis of satellite imagery permitted identification of ice streams on the basis of fan shaped flow patterns, abundant flow parallel features and eskers, and areas of effective glacial erosion. These lobes were delimited by glaciofluvial deposits known as interlobate complexes which were thought to be foci of meltwater activity. Cross-cutting bedform patterns were inferred to have been produced time-transgressively during retreat of the ice sheet margin as the ice streams remained active.
B. Holmlund and Fastook (1993)	Baltic Sea Ice Stream	not given explicitly	ca. 11,000 yr BP	Developed a numerical model to investigate the response of the Scandinavian Ice Sheet to the Younger Dryas climate oscillation. Model results indicated that in order to fit the geological evidence of maximum extent, high flow rates are required throughout the Baltic basin. It is argued that the modelled ice stream is morphologically and glaciologically similar to contemporary ice streams in Antarctica.

Table 3.4. continued.

Reference	Location of Ice Stream(s)	Dimensions	Timing	Principal Evidence for Ice Stream Activity.
B - G. Dongelmans (1996)	Central and South Finland, Russian Karelia and northern Scandinavia	ca. 40 km wide 200-250 km long	During deglaciation	Areas of streamlined bedforms (mainly drumlins) bordered by areas with little or no streamlined bedforms are thought to represent areas of fast ice flow. Inter-stream areas dominated by hummocky and glaciofluvial deposits where ice flow was much slower. Lateral changes in ice stream velocity illustrated by the central areas of the ice stream having the highest density of bedforms, decreasing towards the marginal areas; ice streams also inferred from ice sheet configuration.
A. King *et al.* (1996); Sejrup *et al.* (1998)	Norwegian Channel Ice Stream	not given explicitly	multiple glaciations; last deglaciation up until 15,000 yr. BP	A large sediment accumulation fan in the North Sea is attributed to periodic inputs from an ice stream in the Norwegian Channel. This fan is the largest such feature detected off the coast of Norway, comprising approximately 20,000 km^3 of sediment deposited during the Quaternary. Janocko (1997) found terrestrial evidence of this ice stream in the form of elongate ridges at Jæren in south-west Norway. Morphological, sedimentological and glaciotectonic analysis suggested that the ridges were formed in the marginal zone of the Norwegian Channel Ice Stream but were later overrun by a south-westerly flowing glacier.

Table 3.5. Hypothesised palaeo-ice streams of the Eurasian Arctic and Icelandic Ice Sheets and the main lines of evidence used in their identification

Reference	Location of Ice Stream(s)	Dimensions	Timing	Principal evidence for Ice Stream Activity.
X and Z on Figure 4. Punkari (1995b)	Novaya Zemyla / Barents Ice Sheet	not given explicitly	not given explicitly	These ice streams were postulated along the same lines of evidence used by Punkari (1993, 1995a, 1997) for identifying Scandinavian Ice Streams, see Table 3.4.
Figure 5. Vorren and Laberg (1997); Dowdeswell *et al.* (1998)	Barents Sea-Svalbard Ice Sheet	not given explicitly	operated at glacial maxima throughout the Pleistocene	Several 'trough mouth fans' up to 215,000 km^2 are indicative of a deforming bed depositing huge amounts of sediment at the grounding line. Radiocarbon dates indicated that the sediment delivery was episodic and thus attributable to fast ice flow events. Source area the adjoining shelf, linked by a large submarine trough conducive to fast ice flow. Glacigenic sediments thought to be deformation till with a high water content which facilitated fast ice flow.
Fig 6. Bourgeois *et al.* (2000)	Icelandic Ice Sheet	20 km wide, 150-200 km long	LGM (21,000 yr BP)	Several ice streams were identified using the distribution of mega-scale glacial lineations and flutes, and fast ice flow was confirmed by simple mass balance calculations. The ice streams were characterised by convergent onset zones feeding a number of topographic troughs. Field evidence strongly suggested that fast ice flow was facilitated by a deforming layer of saturated subglacial till which was deposited offshore in the form of 2 km thick sedimentary wedge. Several of the ice streams are thought to have been predisposed by geothermal heat anomalies which played an important role in producing excess meltwater for enhanced

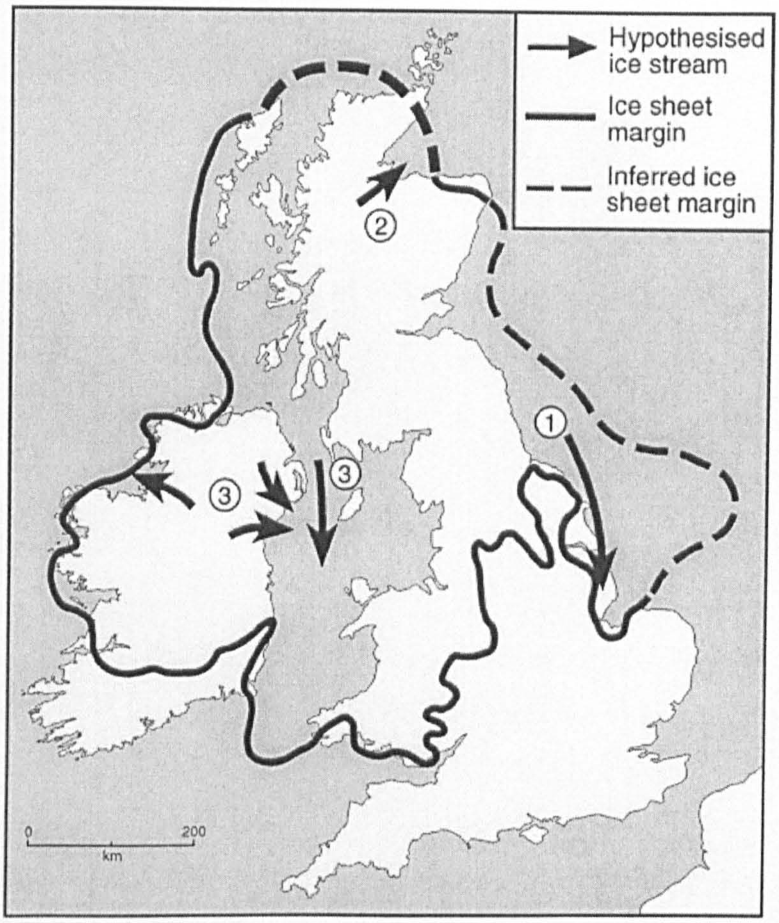

Figure 3.3. Hypothesised locations of ice streams from the British and Irish Ice Sheets. Approximate ice stream locations are depicted by black arrows (not to scale) and the number refers to the corresponding citation and evidence outlined in Table 3.3. Note that the ice sheet margin is depicted at full glacial conditions but that ice streams operated at various times. Ice Sheet outline simplified from Boulton *et al.* (1977).

Until recently, Irish Ice Sheet models were based on immobile ice centres assumed to have undergone an orderly pattern of retreat. However, recent evidence presented by Knight and McCabe (1997) and Knight *et al.* (1999) suggested that dynamic changes in the basal thermal regime of the ice sheet took place during deglaciation. Knight and McCabe (1997) invoked ice stream activity in order to account for cross-cutting landforms, bedform streamlining, drumlinisation and high sediment fluxes, see Table 3.3. Transverse ridges (Rogen moraine) are thought to be indicative of cold-based ice, but in places, these ridges have become drumlinised. Knight and McCabe (1997) argued that a shift in ice dispersal centres resulted in a change in the basal thermal regime and hence, ice stream activity.

McCabe and Clark (1998) also presented evidence for an ice stream which resulted from convergent flow into the northern Irish Sea and represented a major conduit of the British Ice Sheet. This ice stream is thought to have re-advanced at least five times between 22,000 and 14,000 yr BP. In particular, it is hypothesised that its final advance and associated iceberg discharge around 14,000 yr BP may have participated in Heinrich event 1. They suggested that the discharge of icebergs from the Laurentide Ice Sheet during Heinrich event 1 triggered a climate reversal in the North Atlantic, and that the activity of the Irish Sea Basin Ice Stream may have been the rapid response of a small Ice Sheet to this cooling.

3.3.4. Ice Streams in the Scandinavian Ice Sheet.

Punkari (1993; 1995a; 1997) specifically cited the locations of eleven ice streams within a broad marginal zone of the Scandinavian Ice Sheet, extending several hundreds of kilometres inland. The locations of these ice streams is shown in Figure 3.4 and Table 3.4 shows the evidence used to identify them. Analysis of glacial lineations revealed a complex pattern of ice flow, thought to have resulted from the time-transgressive retreat of the ice margin as the ice streams remained active (Punkari, 1997). During deglaciation, individual ice streams responded asynchronously, and some ice streams propagated upstream and captured the drainage basin of their neighbours, possibly leading to ice stream shut-down. This may be analogous to the situation inferred to have led to the shut-down of Ice Stream C in West Antarctica, see Section 2.3.5. Indeed, this theory is supported by Dongelmans (1996), who augmented the location of many of Punkari's ice streams and also suggested that several of them were regulated by the glacio-dynamic processes internal to the ice sheet, i.e. pure ice streams.

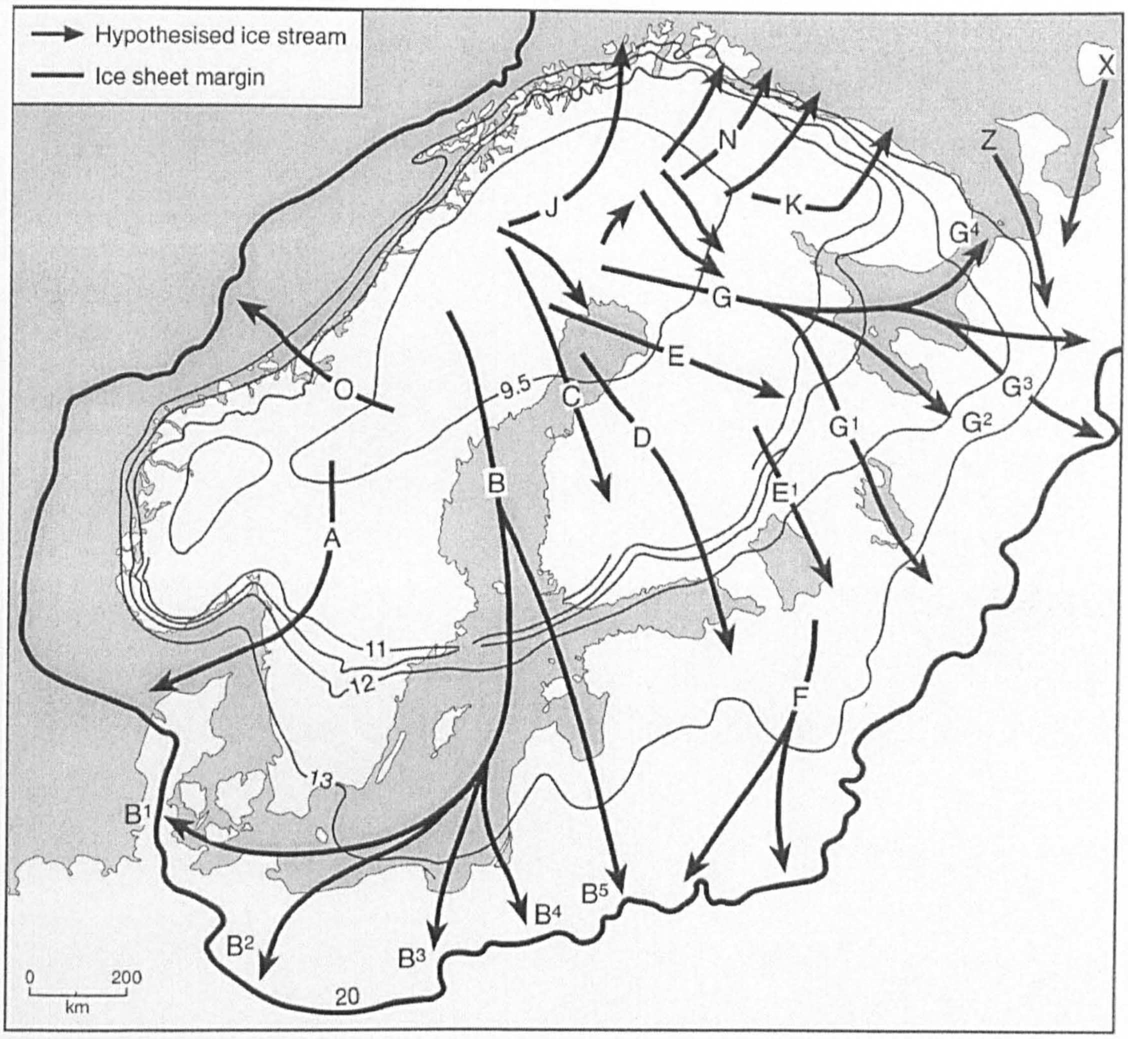

Figure 3.4. Ice streams of the Scandinavian and Novaya Zemyla Ice Sheets identified by Punkari (1993; 1995a; 1997). Ice streams (arrows, not to scale) are thought to have remained active during the retreat of this ice sheet and some marginal positions are shown. Of particular importance are the Baltic Sea Ice Stream (B) and the Norwegian Channel Ice Stream (A), which have also been suggested from other studies (see Table 3.4). Also note the coalescence of the Scandinavian Ice Sheet with two ice streams from the Novaya Zemyla Ice Sheet (X and Z), (from Punkari, 1997).

It is interesting to note that some of the ice stream locations (Figure 3.4) were also identified by Kleman *et al.* (1997) who used a series of 'fans' to construct a glacial geological inversion model of the Fennoscandian Ice Sheet. A fan is composed of geological features which depict a spatially coherent pattern of ice flow and may include striae, flutes, till fabrics and glaciotectonic folds etc. In all, six types of fan are recognised and of these, two fans may have been produced by ice streams. A 'synchronous' or event fan is characterised by abundant flow traces which lack aligned meltwater traces (Kleman *et al.*, 1997). In theory, such fans could be produced by former ice streams, but they may also form under conditions of slow ice sheet flow. However, a 'surge' fan can be related exclusively to enhanced ice flow. These are characterised by a distinctive bottleneck pattern which reflects a near synchronous period of formation. Kleman *et al.* (1997) identified several surge fans which display divergent patterns, terminating in end moraine systems and they are inferred to have been formed isochronously by ice streams.

Of the eleven ice streams identified by Punkari (1993; 1995a; 1997), it is thought that the Baltic Sea Ice Stream (labelled 'B' on Figure 3.4) was most important in terms of its impact upon the ice sheet (Punkari, 1997). Holmlund and Fastook's (1993) numerical modelling of the Scandinavian Ice Sheet indicated that in order to fit the geological evidence of the maximum extent of the Weichselian Ice Sheet, high flow rates typical of ice streaming are needed throughout the Baltic Basin. The modelling suggested that by prescribing zones of belting melting throughout the Baltic Sea Basin, an ice stream develops which is glaciologically and morphologically similar to Antarctic ice streams, see Table 3.4.

Like the Baltic Sea Ice Stream (because of its size and marine nature), the Norwegian Channel Ice Stream (labelled 'A' on Figure 3.4) is also thought to have played an important role in the dynamics of the Scandinavian Ice Sheet. Recent marine investigations have strengthened the evidence for the existence of this ice stream and both King *et al.* (1996) and Sejrup *et al.* (1998) have attributed a large sediment accumulation in the North Sea to its periodic activity (Table 3.4).

In a recent review of Quaternary glaciations in southern Fennoscandia, Sejrup *et al.* (2000) discussed the role of the Norwegian Channel Ice Stream based on records from the North Sea and southern Norway. Its behaviour can be related to periods of maximum glaciation, with the oldest dated activity occurring in the Early Quaternary, around 1.1 M yr BP, and its youngest dated activity occurring during a re-advance between 18,000 and 15,000 yr BP. Repeated advances of the ice stream during the Middle to Late Pleistocene correspond to an easterly located dispersal centre over Scandinavia, thus implying that its sustenance requires a large supply of ice (Sejrup *et al.*, 2000).

3.3.5. Ice Streams in the Eurasian Arctic and Icelandic Ice Sheets.

In the zone of confluence between the Scandinavian and Novaya Zemyla Ice Sheets, Punkari (1995b) has found evidence of two palaeo-ice streams marked 'Z' and 'X' on Figure 3.4. These ice streams were identified using the same evidence as the Scandinavian ice streams (see Table 3.5) and although the ice streams may have been active throughout the glaciation, their time-transgressive retreat during deglaciation is best preserved.

The identification of palaeo-ice streams in the former Eurasian Arctic Ice Sheets relies heavily on marine investigations because terrestrial evidence is scarce. Using seismic and sediment coring techniques, Vorren and Laberg (1997) and Dowdeswell *et al.* (1998) described large sediment accumulation fans in the Norwegian and Greenland Sea. These features occur in front of glacial troughs or channels on the continental shelf and are thus referred to as 'trough mouth fans'. They argued that these focused accumulations of sediment are produced by ice streams draining the north-western part of the Scandinavian Ice Sheet and the Barents Sea-Svalbard Ice Sheet. Using measured sediment amounts and assuming similar conditions to contemporary ice streams (i.e. a deforming bed layer 4 m thick etc.), Vorren and Laberg (1997) estimated that the Bear Island Ice Stream flowed at around 2.5 km a^{-1} and discharged around 200 km^3 of ice on a annual basis. Table 3.5 shows the evidence for these ice streams and Figure 3.5 their approximate location.

Based on glacial directional features at a range of scales, Bourgeois *et al.* (2000) reconstructed flow-lines of the Icelandic Ice Sheet at the Last Glacial Maximum (LGM) around 21,000 yr BP. Ice streams were identified by the presence of mega-scale glacial lineations and flutes, and flow was found to be channelled down several major valleys. Table 3.5 shows other lines of evidence used in their identification and Figure 3.6 maps their reconstructed location in the Icelandic Ice Sheet.

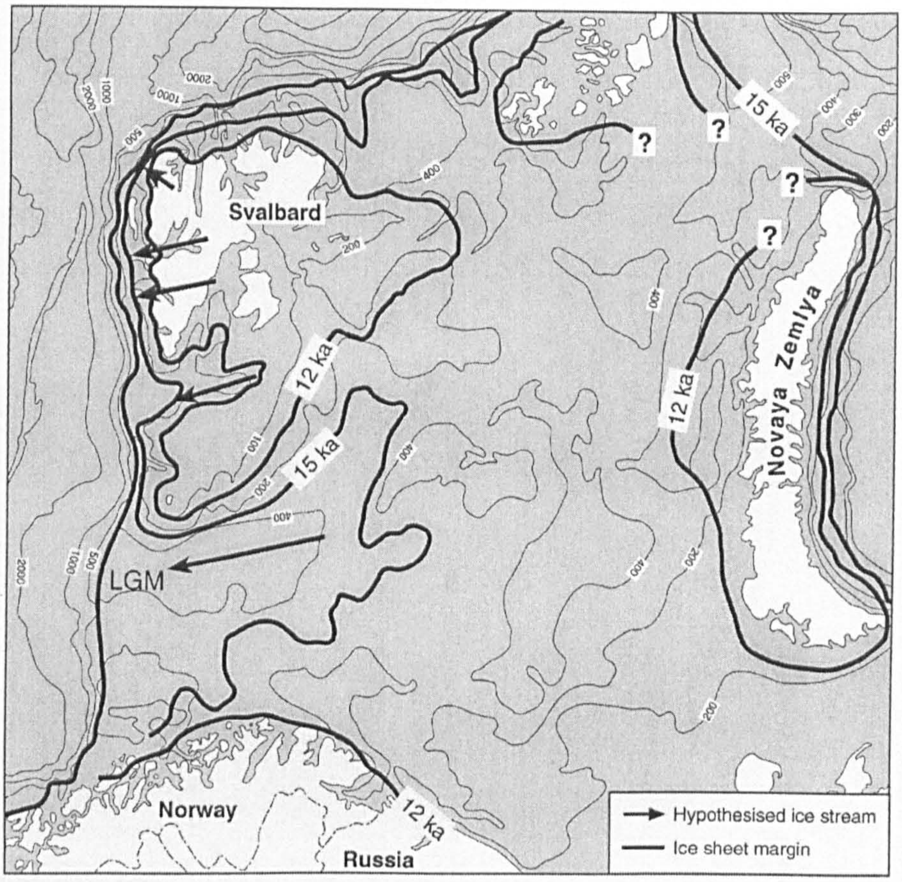

Figure 3.5 Hypothesised ice streams (black arrows, not to scale) from the Svalbard-Barents Sea Ice Sheet. Modified from Landvik *et al.* (1998) and Vorren and Laberg (1997).

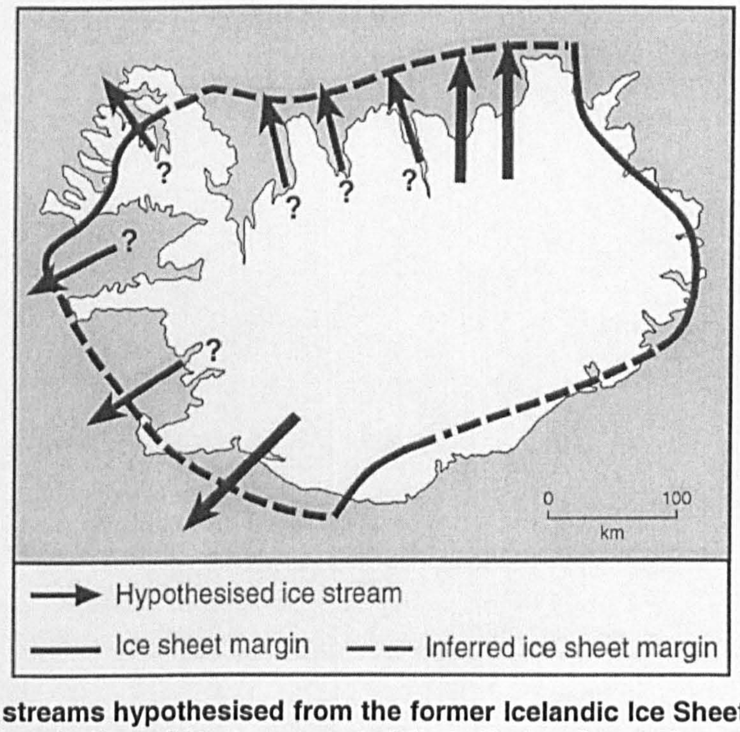

Figure 3.6. Ice streams hypothesised from the former Icelandic Ice Sheet (Bourgeois *et al.*, 2000). Although the evidence suggested that all nine former ice streams may be valid candidates, three of the ice streams have been postulated with far greater certainty (modified from Bourgeois *et al.*, 2000).

3.3.6. Summary.

Many workers have recognised the importance of palaeo-ice streams and a variety of evidence has been used to infer their existence (Tables 3.1 to 3.5). Table 3.6 is a summary of the main lines of evidence used to identify the ice streams outlined in Tables 3.1 to 3.5. The likelihood of each of these ice streams and the strength of the evidence used for their identification is assessed later in chapter 8 (Section 8.2.4).

The purpose of Table 3.6 is to illustrate the diversity of the evidence used to identify palaeo-ice streams. Perhaps more importantly, it highlights the most commonly used and/or more commonly available evidence, as opposed to that evidence which is rarely used or available. For example, striae and bedform evidence is cited for most palaeo-ice streams and a large number of ice streams were located in topographic troughs. In contrast, ice sheet modelling has been used to support only a handful of palaeo-ice stream locations.

More importantly, some palaeo-ice streams, such as the Hudson Strait Ice Stream, left behind a range of evidence to be scrutinised whereas the validity of others remains open to question. Only a small number of palaeo-ice streams are well constrained in both space and time whereas most are far more speculative. This is due to a number of problems inhibiting palaeo-ice stream research, the most important of which are discussed in the following section.

Table 3.6. Summary of the main lines of evidence used to identify a selection palaeo-ice streams.

Ice Stream	Reference(s)	Striae, bedforms/ bedform patterns	Till analysis	Erratics/ dispersal plumes	Ice sheet configuration	Trough topography	Marine evidence, e.g. sediment fan	Ice sheet modelling	Absolute dating constraints
Hudson Strait	Andrews *et al.* (1985); Laymon (1992)	✓		✓	✓	✓	✓	✓	✓
Prince of Wales Island, Canadian Arctic	Dyke and Morris (1988)	✓		✓					
Albany Valley, James Bay Lowlands	Hicock (1988)	✓	✓	✓		✓			
Simcoe Lobe (Ontario)	Boyce and Eyles (1991)	✓	✓						
perpendicular to the mouth of Hudson Strait	Kaufman *et al.* (1993)	✓		✓			✓		✓
Victoria Island, Canadian Arctic Archipelago	Hodgson (1993; 1994)	✓	✓			✓			✓
James Bay Ice Stream	Veilette (1997)	✓	✓						
Des Moines Lobe, Minnesota	Patterson (1997; 1998)	✓			✓	✓			✓
Cumberland Sound Ice Stream	Kaplan *et al.* (1999)					✓		✓	
Skeena Valley/continental shelf, British Columbia	Hicock and Fuller (1995)	✓	✓	✓		✓			
Lillooet and Cheakamus Valleys, southern Coast Mountains, British Columbia.	Evans (1996)	✓				✓			
North Sea Ice Stream	Eyles *et al.* (1994)	✓	✓					✓	
Moray Firth Ice Stream (Scotland)	Merrit *et al.* (1995)						✓		

Table 3.6. continued.

Ice Stream	Reference(s)	Striae, bedforms/ bedform patterns	Till analysis	Erratics/ dispersal plumes	Ice sheet configuration	Trough topography	Marine evidence, e.g. sediment fan	Ice sheet modelling	Absolute dating constraints
Irish Ice Streams (north central Ireland) and Irish Sea Basin Ice Stream	Knight and McCabe (1997); Knight *et al.* (1999); McCabe and Clark (1998)	✓				✓			✓
13 ice streams in the Scandinavian Ice Sheet including the Baltic Sea and Norwegian Channel Ice Streams	Punkari (1993; 1995; 1997); Holmlund and Fastook (1993); King *et al* (1996); Sejrup *et al* (1998); Kleman *et al.*(1997)	✓			✓		✓	✓	
Novaya Zemyla / Barents Ice Sheet	Punkari (1995b)	✓			✓				
Barents Sea-Svalbard Ice Sheet	Vorren and Laberg (1997); Dowdeswell *et al.* (1998)				✓	✓	✓	✓	✓
Icelandic Ice Sheet	Bourgeois *et al.* (2000)	✓	✓		✓	✓			

3.4. Discussion: Problems in Identifying Palaeo-Ice Streams.

3.4.1. Terminology.

Unfortunately, the term ice stream is often used very loosely to describe a range of fast ice flow phenomena. What are the differences, for example, between an ice stream, a surge and a surging ice stream? Does a transient ice stream qualify as a surge? This has led to many ambiguities in the literature. Ice streams have been used to account for terrestrial lobes protruding from the ice sheet margin (e.g. Patterson, 1997), and to describe zones of fast flow within a lobe protruding from the margin (e.g. Boyce and Eyles, 1991). The word ice stream has also been interchanged with the term 'tidewater glacier' (e.g. Merritt *et al.*, 1995) and it has been used to describe short-lived surge behaviour in a marine environment (e.g. Kaufman *et al.*, 1993) and surging lobes on land (Eyles *et al.*, 1994).

The merit of individual former ice stream hypotheses is commented upon in Section 8.2.4, but it is recommended here, that the definition of an ice stream given by Paterson (1994) should provide the foundation upon which former ice stream hypotheses should be based. That is, "a region in a grounded ice sheet in which the ice flows much faster than in regions on either side" (see Section 2.2). This carries no implication as to whether the flow is transitory or continuous.

The term 'surge' is usually taken to describe flow acceleration of a temporary (and non-steady state) nature, which in an ice sheet context is potentially confusing. It is probably best to restrict the use of the term 'surging' just to glaciers (whereby the whole glacier experiences flow acceleration), and to use the term 'ice stream' for zones of fast flow within an ice sheet. These ice streams may be in continuos operation or transitory. An 'ice sheet surge' would thus be reserved for a dramatic flow acceleration and margin advance of the whole ice mass.

3.4.2. Lack of Diagnostic Evidence.

Little theoretical work has been carried out to predict the geomorphology and geology that we should expect an ice stream to produce. In essence, we have no clear criteria on which to base our assumptions, and this point was raised by Mathews (1991) who emphasised the need for diagnostic criteria when attempting to find palaeo-ice streams. This oversight has resulted in two main problems; (a), a huge variety of evidence has been used to infer former ice streams (Tables 1-5), and (b), such evidence has rarely been scrutinised in detail, despite its often subjective nature. Because ice streams are distinct features within an ice sheet, it seems logical to suggest that they will, in general, leave behind a distinct imprint of their activity, a 'geomorphological signature'. However, it seems that in some cases the identification of a distinct flow pattern (i.e. of drumlins) is regarded as ample justification for postulating the location of an ice stream. This problem is addressed in Chapter 4.

3.4.3. Isochronous and Time-Transgressive Ice Stream Imprints.

Many palaeo-ice stream hypotheses are based on distinctive flow patterns which have been reconstructed from ice flow indicators such as subglacial bedforms. While it is imperative to demonstrate why the imprint was produced by an ice stream, a further complication involves the synchroneity and context of bedform generation beneath an ice sheet. We may consider two extremes. At one extreme, the whole ice stream imprint may have been produced 'isochronously' leaving behind a snapshot of ice stream activity at a given time (this can be thought of as a 'rubber-stamped' imprint). This may be produced by an ice stream that operates, then switches off, with the imprint remaining preserved during deglaciation. On the other hand, an ice stream may operate throughout several cycles of advance and retreat, or be continuously operating during margin retreat, whereby the earlier imprints are modified and overprinted by the younger ice flow patterns. The other extreme therefore, is a 'time-transgressive' imprint that is continuously re-organised over time and may appear discontinuous and complex (a 'smudged' imprint). Unfortunately, most palaeo-ice stream hypotheses overlook this distinction, which may lead to erroneous ice stream definition or misinterpretations regarding the glacial history and temporal activity of the ice stream.

3.4.4. Lack of Modern Analogues.

It is a geographical coincidence that all contemporary ice streams happen to be marine-based, leaving us with the problem that we do not actually know what a terrestrial ice stream looks like. This poses a number of important questions when trying to reconstruct the configuration of a terrestrial ice stream, not least of which is; how does the ice stream rapidly evacuate ice? Because contemporary ice streams feed ice shelves or terminate in open water conditions, the removal of ice at the terminus is rapid and this maintains a high velocity. As such, marine-based ice streams can maintain a high velocity without advancing. In contrast, a terrestrial ice stream has a much less effective method for removing ice and presumably therefore, has to advance producing a large splayed ice lobe at its terminus. This explains why many former ice streams have been postulated for the southern margin of the Laurentide Ice Sheet whose lobate configuration is thought to represent the distal portion of a number of palaeo-ice streams (see Section 3.3.1 and Figure 3.2).

Contemporary ice streams are bordered by slower moving ice all the way to the grounding line and have a predictable lateral velocity gradient often described as 'plug flow', although basal drag is also important (see Section 2.3.3). This presents further questions with regards to the configuration of terrestrial ice streams. Are terrestrial ice streams characterised by plug flow all the way to the ice margin? Is fast flow restricted to a narrow zone along the central axis of the lobe, or does the divergence of flow result in a more uniform but slower regime of flow velocities? These questions are addressed in Section 8.3.1.

3.4.5. Inaccessible Locations.

An obvious yet unavoidable problem when trying to identify palaeo-ice streams is the difficulty in getting to them. Remote Arctic regions are costly to visit and it is not a surprise that the southern margin of the Laurentide Ice Sheet has experienced more scrutiny than the numerous possible ice stream tracks of the northern sector of the ice sheet. Of greater difficulty is that many palaeo-ice streams were located in topographic lows whose subsequent marine submergence hinders the detailed observation of their geomorphology and geology.

3.4.6. Glaciological Problem Solving.

With the advent of ice sheet modelling comes another problem associated with the locations of former ice streams. That is, ice streams have sometimes been hypothesised in order to partly overcome a glaciological problem. For example, Holmlund and Fastook (1993) inferred the location of the Baltic Sea Ice Stream from numerical modelling which could not explain the maximum extent of the ice sheet without a large fast flow zone. The notion of an ice stream in the Baltic Sea basin may be a valid assumption, but the model relied heavily on the basal parameters which were prescribed on a rather *ad hoc* basis. Furthermore, the present day Baltic Sea hinders any detailed evidence being scrutinised.

Similarly, although Eyles *et al.* (1994) described detailed till analysis, their hypothesised ice stream in the North Sea was underlined by the fact that steady-state conditions in an ice sheet model could not explain this portion of the ice sheet. Again, present day sea level hinders any detailed analysis of the geomorphology and we may want to question the validity of the model simulation. While modelling has provided an invaluable tool in understanding ice sheet behaviour at the large scale, we should be wary of relying on it to such an extent that the geomorphological record is neglected. Rather, ice stream geomorphology when available, should be used to test the model outputs.

3.5. Summary and Conclusions.

Palaeo-ice streams are important for a number of reasons. They provide an unprecedented opportunity to observe the bed of an ice stream, thus alleviating one of the main inhibitors to contemporary ice stream research. Because of their profound impact on ice sheet configuration we need to know when and where they were in order to accurately reconstruct former ice sheet histories. Added to this, their behaviour is inextricably linked to the climate system where it has been demonstrated that not only do ice sheets respond to climate change but that ice streams are responsible for forcing some of the most abrupt climate changes ever identified. Furthermore, because ice streams are powerful erosional agents, they are responsible for rapid landscape evolution and for producing some of the world's largest sedimentary fans.

A large number of workers have recognised the significance of finding palaeo-ice stream tracks in formerly glaciated environments. However, hypothesised locations have tended to outweigh meaningful evidence because our understanding of ice stream geomorphology is limited. To compound this problem, palaeo-ice streams have been confused with terrestrial ice sheet lobes, tidewater glaciers and surge behaviour. Indeed, because all contemporary ice streams happen to be marine-based, the configuration of terrestrial ice streams proves somewhat problematic. A further problem is that the evidence of former marine-based ice streams is often obscured by present day marine submergence.

It is apparent from the literature that a multitude of evidence has been used to identify palaeo-ice streams, with different researchers using a variety of criteria. To evaluate hypothesised palaeo-ice stream tracks and to search for further examples, requires the establishment of a theoretical framework that relates ice stream processes and patterns to their geological and geomorphological products. Put simply, what evidence should an ice stream leave behind? If the West Antarctic Ice Sheet were to disintegrate, we could examine its bed and easily answer this question. Until such time however, we will have to build a conceptual model based on glimpses of contemporary ice stream beds derived from geophysics and boreholes, coupled with the inferences we can draw from formerly glaciated terrain.

Coupling the known characteristics of contemporary ice streams (described in Chapter 2) with traditional theories of glacial geomorphology provides a sound theoretical basis for predicting diagnostic criteria of palaeo-ice stream activity. The following chapter predicts several geomorphological criteria for identifying former ice streams.

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Chapter 4: Geomorphological Criteria for Identifying Palaeo-Ice Streams.

4.1. Introduction.

Ice streams are discrete features within an ice sheet and so we may expect them to leave behind a unique suite of glacial landforms. By coupling the known characteristics of contemporary ice streams (described in Chapter 2) with traditional theories of glacial geomorphology, it is possible to predict several geomorphological products of palaeo-ice stream activity. While individual criteria are not necessarily exclusive to ice stream activity, *collectively* they can be viewed as an idealised palaeo-ice stream track. The criteria are used to construct conceptual landsystem models of the beds of former ice streams, and it is envisaged that such models can provide an observational template upon which hypotheses of former ice streams can be better based.

4.2. Rationale: The Need for Geomorphological Criteria.

A large number of palaeo-ice stream hypotheses are based on the specific geomorphological products of an ice sheet. This can be seen in Table 3.6 where the majority of palaeo-ice stream hypotheses are based on some combination of bedforms or bedform patterns.

The most obvious and abundant landforms are subglacial bedforms, which have been defined as "longitudinal or transverse accumulations of sediment formed below active ice" (Benn and Evans, 1998). Examples of the most common longitudinal bedforms are flutings, mega-flutings, drumlins and mega-lineations, all of which are streamlined parallel to ice flow. These bedforms often occur in discrete sets which when viewed as a whole, are thought to reflect the regional ice flow pattern.

In theory, subglacial bedforms can be related to the position of ice divides and ice streams etc., and they can also provide information pertaining to local stress variations, sediment supply, ice flow and sediment deformation histories (Benn and Evans, 1998).

The main inhibitor to palaeo-ice stream research are that the geomorphological evidence of their activity is poorly understood. Little theoretical work has been carried out to predict what subglacial bedforms or patterns are produced by ice streams. A consequence of this is that hypothesised former ice streams have often outweighed meaningful evidence for their existence, see Section 3.4.2. Indeed, Mathews (1991) emphasised the need for diagnostic criteria; "still other candidates for ice stream tracks may be found within the network of channels between the islands of the Canadian Arctic. Though I doubt if many of these will ultimately qualify, they nevertheless merit investigation with this possibility in mind *once the criteria for identifying ice stream tracks become better established*". Because glacial geology and geomorphology are the most widely used (and available) evidence for reconstructing former ice sheets (see Table 3.6), the criteria presented in this chapter will relate ice stream processes and patterns to their geological and geomorphological products. These criteria are not definitive, but merely aim to introduce a more objective methodology to research on former ice streams.

A logical foundation for the development of such criteria can be found in the basic characteristics of contemporary ice streams (summarised in Section 2.7). Some of these characteristics may be manifest in the glacial geomorphological record produced by an ice stream.

4.3. Basic Characteristics of Contemporary Ice Streams.

4.3.1. Large Dimensions.

Ice streams are considered regional-scale phenomena, greater than 20 km wide and over 150 km in length. Their dimensions are much larger than valley, surge-type and outlet glaciers. The most widely studied ice streams are those on the Siple Coast of West Antarctica which range in width from 30-80 km and reach lengths of 300-500 km, see Table 2.1. Most ice streams fall within these dimensions, but there are exceptions. For example, Jakobshavns Isbræ in south-western Greenland is only around 6 km wide and 85-90 km long (Echelmeyer *et al.*, 1991). Although individual ice stream dimensions vary, the length is usually around 10 to 15 times the width.

4.3.2. Characteristic Shape.

As well as having characteristic dimensions, ice streams also display a characteristic shape, and simple analogies can be made with fluvial systems. Ice streams are often fed by small tributaries or 'channels' at or near the head of an ice stream which feed a main ice stream channel (see Section 2.3.4). It is very rare for a tributary to join an ice stream downstream from the onset of streaming flow (Whillans and Van der Veen, 1993). Most of the streaming flow is concentrated in a large, main channel (the trunk) and in this way, an ice stream may be likened to a second order fluvial system. The main channel is characterised by rapid velocities bordered by slower moving ice and shows little sinuosity. Figure 4.1 shows a simplified diagram of the theoretical ice stream shape described here.

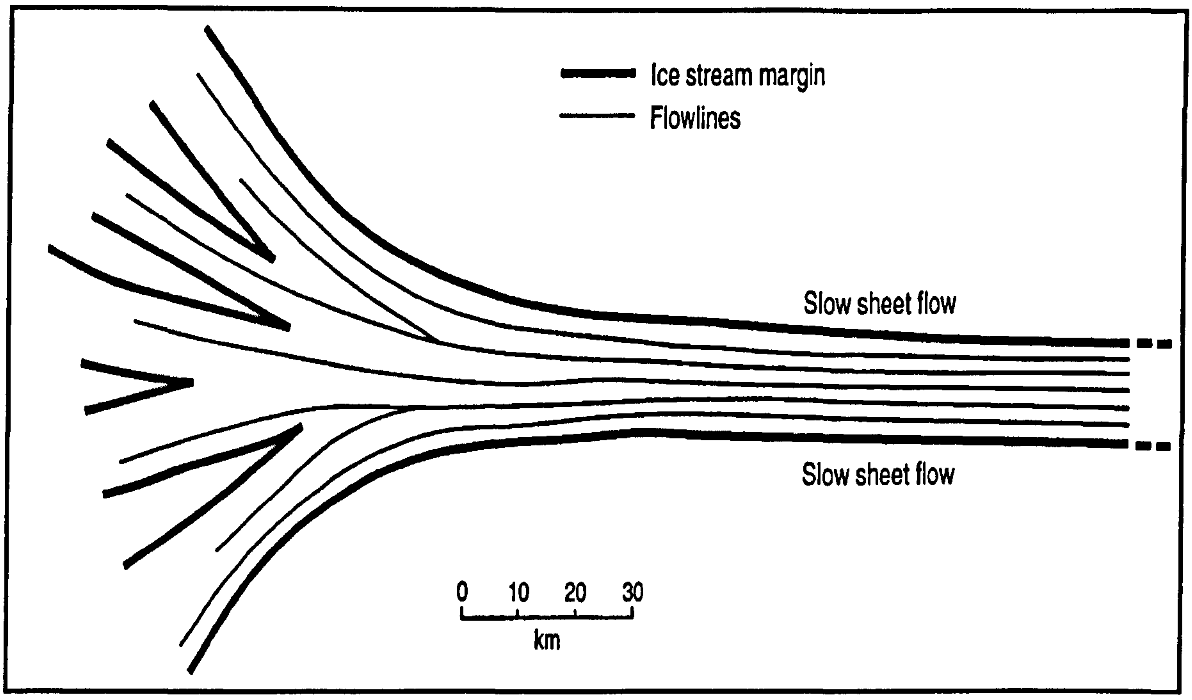

Figure 4.1. Simplified theoretical shape of an ice stream, characterised by convergent flowlines in the onset zone feeding the main channel (from Stokes and Clark, 1999).

It should be noted however, that ice streams display variations in shape. For example, streaming flow is often diverted around ice rises or pinning points such as bedrock protrusions.

4.3.3. Rapid Velocities.

Ice streams are characterised by their extremely rapid velocities. Table 2.1 shows the surface velocities of several selected ice streams. It can be seen that most ice streams flow at rates greater than 300 m a^{-1}, although some may flow considerably faster. In south-western Greenland, the fastest ever recorded ice stream velocity was measured on Jakobshavns Isbræ where the velocity was estimated at 8360 m a^{-1} (Lingle *et al.*, 1981). In terms of a minimum velocity of ice stream flow, Whillans and Van der Veen (1993) defined the onset of streaming flow as reaching speeds of about 100 m a^{-1} and this is consistent with most ice streams in the literature.

4.3.4. Characteristic Velocity Patterns.

Ice stream velocities have two main characteristics. Firstly, the velocity of a marine-based ice stream usually steadily increases all the way to the grounding line. Secondly, ice stream velocities remain high all the way across the ice stream until there is an abrupt decrease in the marginal areas. This unique characteristic is described as plug flow and is exhibited by all ice streams. Figure 4.2 illustrates this pattern of surface velocity across an ice stream, from the centre line to the slower-moving ice.

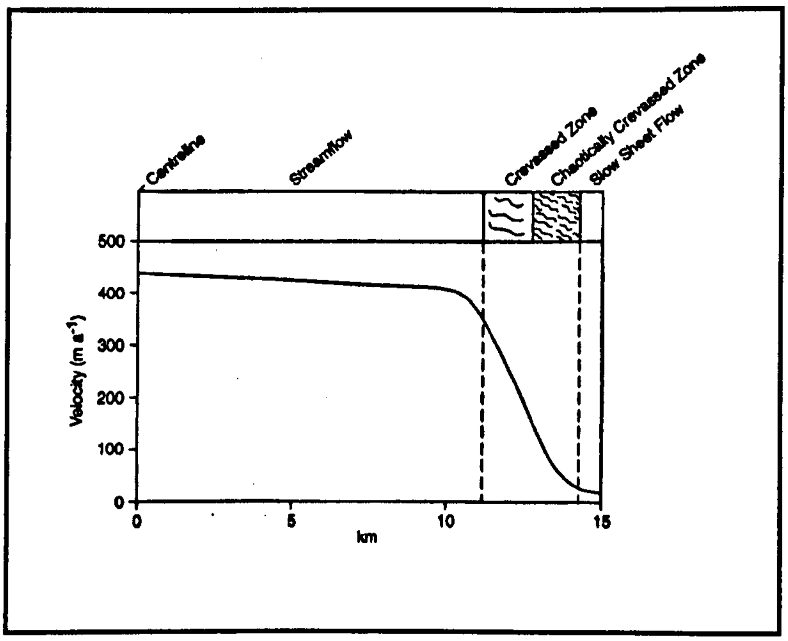

Figure 4.2. Lateral variation in ice stream velocity, simplified from Echelmeyer *et al.* (1994), (from Stokes and Clark, 1999).

4.3.5. Abrupt Lateral Shear Margins.

The fact that ice streams are bordered by slower-moving ice gives them a unique characteristic. Where the faster-moving corridor of streaming ice meets the slower-moving sheet ice, it creates intense shear strains at the margin. This leads to a highly crevassed shear zone, usually around 0.25-2.0 km in width (Vornberger and Whillans, 1990) but no greater than 5 km even on the largest ice streams. Figure 4.2 shows the abrupt marginal area compared to the width of the ice stream. Such heavily crevassed zones are a key feature of contemporary ice streams and are often used to delineate them, especially on satellite imagery (e.g. Bindschadler and Vornberger, 1998).

It should also be noted, that it is possible that ice stream margin locations are governed by shear-related fabric changes at basal thermal boundaries (i.e. frozen/thawed). For example, Kleman and Borgström (1994) suggested that ice crystal re-orientation in the shear zone between cold and warm-based ice leads to the development of an easy glide fabric. This implies that ice stream margins could be self-stabilising.

4.3.6. Evidence of Pervasively Deformed Sediment.

Although not necessarily exclusive to ice stream activity, evidence of pervasively deformed till has in some cases been associated with fast ice flow, especially on pure ice streams. The first direct evidence of deformable till beneath pure ice streams was facilitated by borehole investigations on Ice Stream B (Engelhardt *et al.*, 1990) and it is thought that the mechanism of subglacial till deformation may account for the rapid velocity of the Siple Coast ice streams, at least to some extent (see Section 2.3.2). Thus, a further characteristic of contemporary pure ice streams is that they appear be underlain by deformable sediment. However, caution must be exercised since this mechanism does not account for the fast flow of all ice streams, especially topographic ice streams.

4.3.7. Offshore Sediment Accumulation.

If some ice streams operate by deforming and transporting their underlying till, then till continuity is important. The till will have to be continually replenished upstream and subsequently deposited in the terminal area. Thus, a fast moving marine-based ice stream resting on a deforming bed has the ability to transport and deposit huge amounts of terrigenous sediment onto discrete areas of the continental shelf. Such a till delta has been detected where Ice Stream B enters the Ross Ice Shelf (Alley *et al.*, 1989).

It should also be remembered here, that topographic ice streams have also been associated with sediment accumulation fans. Section 3.3.5 identifies a number of palaeo-ice stream hypotheses from the Eurasian Arctic which were based on the architecture of sediment accumulation fans at the mouths of glacial troughs. Although the exact flow mechanisms and transport processes are difficult to ascertain, it is clear that focused accumulations of sediment can be deposited by both pure and topographic ice streams and that a deforming till layer is not an essential prerequisite for such features to occur.

4.4. Geomorphological Criteria of Ice Stream Activity.

The characteristic features of contemporary ice streams (described above) may be manifest in the glacial geomorphology and geology produced by former ice streams. It is argued that the following criteria are key geomorphological criteria for identifying palaeo-ice streams.

4.4.1. Characteristic Shape and Dimensions.

When inferring former ice stream activity, the most obvious clue to their existence are the shape and dimension of the subglacial bedform pattern. Ice streams are large features and are characteristically greater than 20 km wide and 150 km long. Clearly, their is plenty of scope for an ice stream to be bigger or even smaller, especially if the shape is consistent with ice stream activity (see Figure 4.1).

It is acknowledged that contemporary ice streams may not necessarily be a representative sample of the population of ice streams that have ever existed. However, they do provide the only analogues on which to base our assumptions. It is worth considering whether there is a process which limits the dimensions of ice streams, both past and present, and this is discussed in Section 8.3.2.1. In summary, the subglacial bedform pattern or geomorphic imprint produced by a former ice stream should represent a large flow pattern, characteristically greater than 20 km wide and 150 km long.

4.4.2. Highly Convergent Flow Patterns.

The second clue to the existence of former ice streams are highly convergent flow patterns. In contemporary ice streams, the onset of streaming flow (or onset zone) is characterised by a large convergence zone (see Figure 4.1), whereby slower moving ice is gradually incorporated into the ice stream (probably by tributaries). Thus, subglacial bedforms produced by an ice stream should exhibit a large degree of convergence in the onset zone. This has been observed at the head of a palaeo-ice stream on Prince of Wales Island by Dyke and Morris (1988). Here, the orientation of the drumlins represents a splayed pattern of convergence feeding into the main ice stream axis, see Section 3.3.1.

4.4.3. Highly Attenuated Subglacial Bedforms.

A manifestation of the rapid velocity of an ice stream may be highly attenuated, streamlined bedforms. Thus, extensive swarms of highly attenuated drumlins and mega-scale lineations could record the flow direction and spatial extent of former ice streams.

Elongation ratio (length divided by width) is a useful way of quantifying the degree of attenuation of subglacial bedforms (such as flutes, drumlins and mega-lineations). Although there is no unequivocal method for reconstructing former flow velocities from geomorphological evidence, there are many studies that report correlation's between inferred fast ice flow and high elongation ratios (Boulton, 1987; Boyce and Eyles, 1991; Clark, 1994; 1997; Hart, 1999). For example, Hart (1999) suggested that patterns in bedform elongation ratio may provide a suitable proxy for ice velocity. This hypothesis is supported by variations in elongation ratios in a drumlin field in New York State whose observed transverse and longitudinal patterns matched velocity profiles from contemporary ice streams in Greenland and West Antarctica.

It could be argued that highly-attenuated bedforms may form by slow flowing ice over long time periods. However, the discovery that the flow geometry of ice sheets are known to vary rapidly (cf. Bouton and Clark, 1990a; b) suggests that this scenario is highly unlikely, although theoretically possible.

4.4.4. Boothia-Type Erratic Dispersal Trains.

Dispersal trains "are long and relatively narrow belts of glacial debris of some distinctive composition transported down-ice from a source area" (Dyke and Morris, 1988). Dyke and Morris (1988) recognised two types of dispersal train, schematically represented in Figure 4.3.

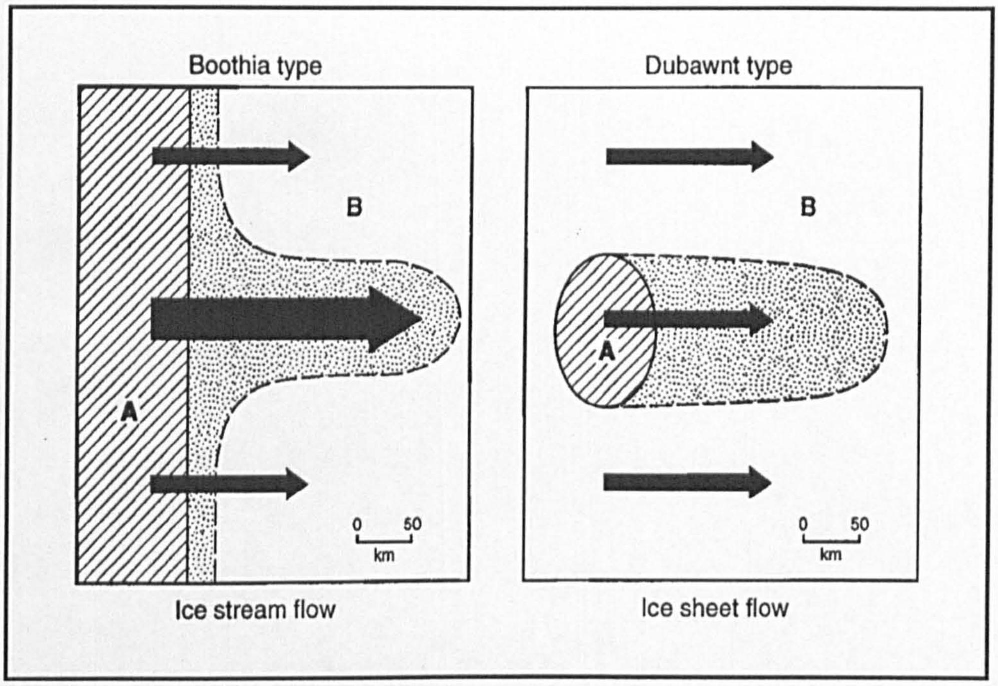

Figure 4.3. Simplified diagram of Boothia and Dubawnt-type dispersal trains, modified from Dyke and Morris (1988).

The Boothia-type dispersal train forms when an abrupt lateral variation in ice velocity transports distinctive sediment from a large source area. Such a lateral variation in ice velocity is a unique characteristic of ice streams (see Section 4.3.4) and Boothia-type dispersal plumes may be a product of their activity.

In contrast, the Dubawnt-type dispersal plume implies no lateral variation in velocity. Although it may appear similar to a Boothia-type dispersal plume, the source area of the sediment is the key control on the pattern and it can be formed by slow sheet-flow. Hence, it is very important to identify the spatial extent of the source area from which the distinctive till is transported. There is not necessarily a blatant connection between ice streams and dispersal trains, but when found in conjunction with other criteria, it may be highly suggestive of ice stream activity.

A Boothia-type plume was identified on Prince of Wales Island in the Canadian Arctic by Dyke and Morris (1988) who suggested that it was formed by the central part of an ice stream. It is also interesting to note that the more attenuated drumlins were formed in the central part of the plume, where presumably the flow was quickest. This provides further support for the ideas addressed in Section 4.4.3, above.

4.4.5. Abrupt Lateral Margins.

Ice streams are characterised by their abrupt lateral margins bordered by slower-moving ice (Figure 4.2). The characteristic geomorphology inscribed by a former ice stream would be expected to exhibit an abrupt margin or an abrupt zonation of subglacial bedforms at the margin. For example, Hodgson (1994) noted a well-defined margin of a drumlin field when postulating the existence of an ice stream flowing northwards over the eastern portion of Victoria Island in the Canadian Arctic, see Section 3.3.1. Dyke and Morris (1988) also noted an extremely abrupt margin to a bedform pattern formed by an inferred ice stream on eastern Prince of Wales Island (see Section 3.3.1).

In addition, both Kleman and Borgström (1994) and Kleman *et al.* (1999) suggested that an abrupt margin is produced at the transition between cold and warm-based ice. Given the fact that some ice stream margins may well be characterised by such a transition, then abrupt marginal areas (<2 km) could be used to de-limit the width of a palaeo-ice stream (see Kleman *et al.*, 1999).

4.4.6. Ice Stream Marginal Moraines.

An abrupt lateral margin associated with a change in ice velocity (as described above) may also be conducive to the generation of characteristic landforms. On south-eastern Prince of Wales Island, Dyke and Morris (1988) identified a single, narrow ridge of till which delineated the western side of a drumlin field. This ridge can be traced for up to 68 km but is less than 1 km in width (Dyke *et al.*, 1992). In a discussion of its origin, the ridge was interpreted as being a 'lateral shear moraine', marking a shear zone at the side of an ice stream, separating fast flowing ice from slower flowing cold based ice (Dyke and Morris, 1988). A similar ridge was also identified by Hodgson (1994), flanking an hypothesised ice stream on eastern Victoria Island. The mode of origin of this feature may be similar, and so it appears that an ice stream marginal moraine may be a unique bedform product of an ice stream.

4.4.7. Evidence of Pervasively Deformed Sediment.

Since the discovery of a deforming till layer beneath Ice Stream B (Englehardt *et al.*, 1990), evidence of pervasively deformed till has been linked to former ice stream activity, e.g. Hicock (1988). Indeed, areas of 'soft' sediments may pre-dispose an ice sheet to fast ice flow. Moreover, the presence of a deformation till may indicate former fast flow zones if additional criteria are also evidenced in the local glacial geomorphology. For example, Boyce and Eyles (1991) have attributed the formation of the Peterborough drumlin field in Ontario (Canada) to deforming till streams thought to record the flow of an ice stream within the Simcoe Lobe of the southern Laurentide margin (Section 3.3.1).

4.4.8. Offshore Sediment Accumulation.

Although not necessarily indicative of ice stream activity, focused accumulations of sediment on a continental shelf may complement, and indeed strengthen terrestrial evidence for ice stream flow. Recently, Vorren and Laberg (1997) have identified huge submarine till deltas, thought to have been produced by ice streams draining the northwestern part of the Fennoscandian Ice Sheets, and the Barents Sea Ice Sheet (see Section 3.3.5).

Because a stable ice sheet margin (i.e. without an ice stream) would not deliver such concentrated accumulations of sediment to the continental shelf, it is clear that offshore sediment accumulations can provide a valuable clue to the existence of marine-based ice streams. Clearly, if terrestrial evidence is available which suggests ice stream activity, the identification of offshore sediment accumulations serves to support such hypotheses. Furthermore, such fans can also be used as proxies for former ice velocities and discharges (see Section 3.3.5).

In summary, the bedform pattern of a palaeo-ice stream needs to have the characteristic shape and dimensions, and display a large degree of convergence in the onset area. Indicators of the rapid ice velocity may be evidenced by highly attenuated bedforms and 'Boothia type' dispersal trains. A manifestation of the sharply delineated margin found on ice streams may be an abrupt lateral margin to the bedform pattern and ice stream marginal moraines. Evidence of a deforming till layer and sediment accumulation fans are not necessarily indicative of ice stream flow but may provide substantial supporting evidence when found in conjunction with the other criteria. Table 4.1 shows how the criteria relate to the basic characteristics of contemporary ice streams.

Table 4.1. Geomorphological criteria for identifying palaeo-ice streams.

Contemporary ice stream characteristic	*Proposed geomorphological signature*
A. Characteristic shape and dimensions	**1.** Characteristic shape and dimensions (>20 km wide and >150 km long)
	2. Highly convergent flow patterns
B. Rapid velocity	**3.** Highly attenuated subglacial bedforms (length:width >10:1)
	4. Boothia-type erratic dispersal trains (Dyke and Morris, (1988)
C. Sharply delineated shear margin	**5.** Abrupt lateral margins (<2 km)
	6. Ice stream marginal moraines
D. Deformable bed conditions	**7.** Glaciotectonic and geotechnical evidence of pervasively deformed till.
E. Focused sediment delivery	**8.** Submarine till delta or sediment fan (only marine terminating ice streams)

4.4.9. Topographic and Pure Ice Stream Geomorphology.

A key question that has to be answered in the context of this chapter is; do topographic and pure ice streams leave behind different geomorphological evidence? A logical answer would be yes, since form and process are inextricably linked, and flow mechanisms appear to differ considerably between pure and topographic ice streams (compare Sections 2.3.2 and 2.4.2). Following on from this; can we differentiate between them in the geological record?

It would appear that several authors have already made this distinction. For example, Evans (1996) suggested that whalebacks and rock drumlins develop under ice streams of Greenland and East Antarctic type, i.e. topographic ice streams. It is suggested that topographic ice streams are erosional agents, sliding and plucking their beds, a consequence of which may be the continual deepening of the bedrock trough in which they lie. In contrast, Marshall *et al.* (1996) argued that topographic ice streams leave little or no geological imprint, but instead, should be easily identifiable on topographic maps of suitably high resolution, provided ice thicknesses can be readily reconstructed. However, this assumes that all topographic ice streams flow by basal sliding or thermally enhanced ice deformation under very high basal shear stresses. This may not be the case (as discussed in Section 2.4.2). For example, some authors have identified topographic ice streams whose motion may well be achieved by a layer of deforming till, e.g. contemporary Rutford Ice Stream (Smith, 1997) and a palaeo-ice stream from the former Cordilleran Ice Sheet (Hicock and Fuller, 1995). Thus, it should be remembered that although the criteria are primarily aimed at objectively identifying pure palaeo-ice streams whose locations are far more unpredictable, the criteria may also be applicable to some topographic ice streams.

4.5. A Landsystems Approach to Ice Stream Geomorphology.

It would be highly unlikely that all of the geomorphological criteria should be found in one location, produced by a single ice stream. This is because not all ice streams will leave a complete geomorphological signature and because preservation and modification scenarios often obscure the complete picture. However, the criteria outlined above can be thought of as comprising a characteristic 'landsystem' produced by a former ice stream. This illustrates the perfect, or unaltered, geomorphological signature of ice stream activity. In essence, it portrays a "fantasy ice stream record".

A landsystem is an area of common terrain attributes. In terms of glacial geomorphology, the concept of landsystems introduces a holistic approach (Benn and Evans, 1998). Here, we are dealing with the subglacial landsystem, and the bedforms produced are genetically related to the processes involved in their development, in this case, ice stream activity. The individual landforms, such as drumlins or flutes, are the lowest classification in the hierarchy, known as *land elements* and subsets or *land facets* may be represented by drumlin fields, for example. For an in depth discussion of the landsystems approach, see Benn and Evans (1998).

4.5.1. Classification of Ice Stream Landsystems.

Ice streams can be broadly categorised as either terrestrial or marine-based, depending upon the environment in which they terminate (see Section 2.2.2). All contemporary ice streams are marine in nature but former ice sheets would also have been drained by terrestrial ice streams. The difference between these two classes of ice streams will be reflected in the geomorphological signature of their activity. For example, terrestrial ice streams cannot produce an offshore sediment accumulation fan. It must also be remembered at this point, that a marine based ice stream will become a terrestrial ice stream once it retreats from the grounding line and progresses inland. Similarly, it is quite plausible that a terrestrial ice stream may also become a marine-based ice stream, should the ice sheet build up to such an extent that it reaches coastal areas, or as a result of sea level rise. These scenarios have important implications for the geomorphological record left behind by an ice stream.

The geomorphology of a palaeo-ice stream can be also be classified as either time-transgressive or isochronous, depending on the synchroniety of the bedform generation beneath the ice stream, see Section 3.4.3. At one extreme there is a simple isochronous (rubber stamped) imprint and at the other, a more disjointed (smudged) imprint, generated time-transgressively. By grouping the criteria together, these ideas can be represented by four simple landsystem models of former ice stream activity, shown in Figure 4.4.

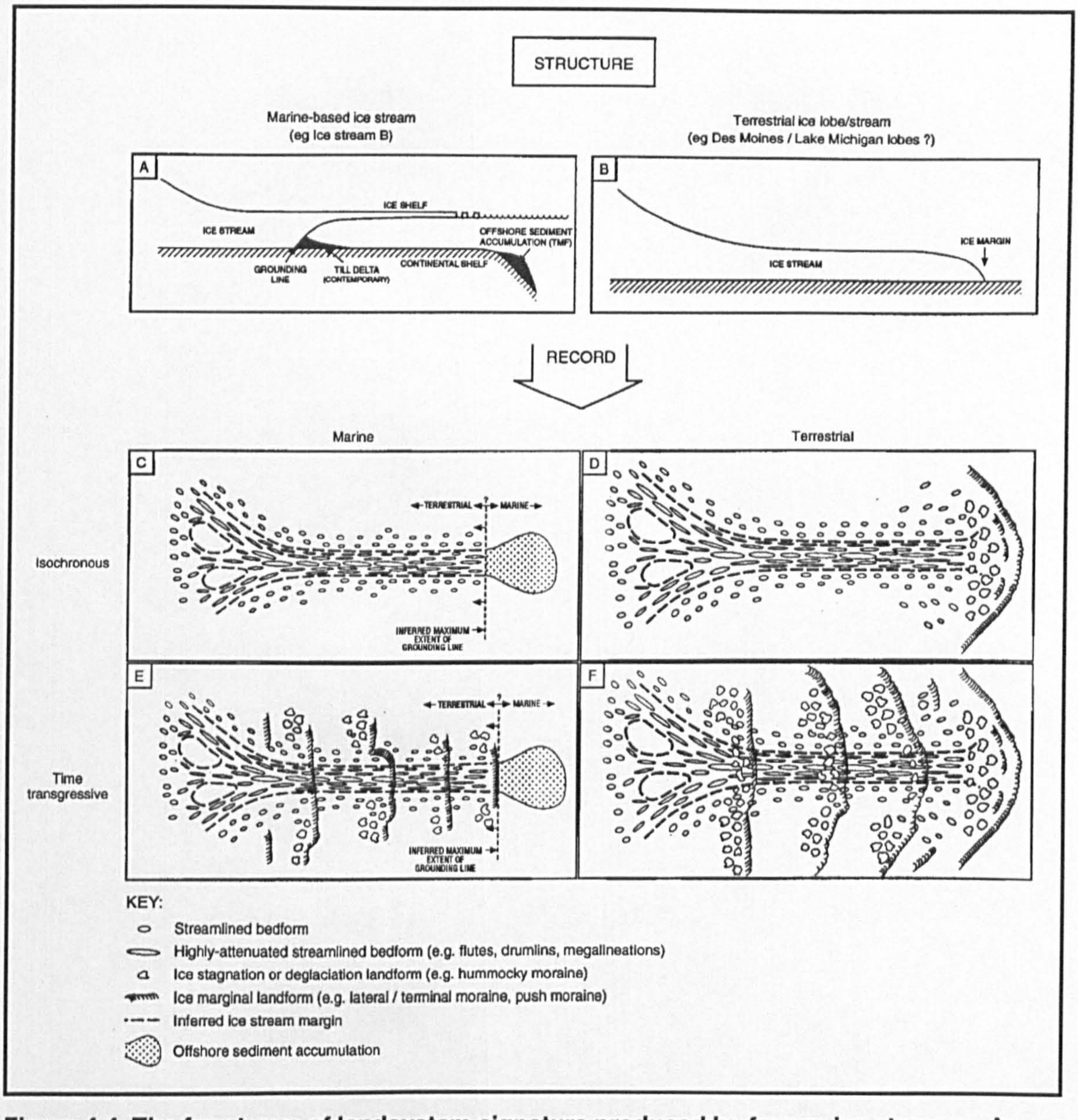

Figure 4.4. The four types of landsystem signature produced by former ice streams. An ice stream may be broadly classified as either marine (a), or terrestrial (b), depending on the environment in which it terminates. The recorded geomorphological evidence (i.e. its signature) can be classified as being isochronous (i.e. a rubber stamped imprint), or time-transgressive (i.e. a smudged imprint). Thus, four possible landsystems may result from former ice stream activity (c, d, e and f). Note that time-transgressive terrestrial ice streams may also superimpose (i.e. cross cut) a splayed pattern of drumlins onto the earlier bedforms as the lobate marginal area retreats. This has been omitted from the diagram for clarity but is illustrated in detail in Figure 4.5 (from Stokes and Clark, 1999).

Either type of ice stream may produce a time-transgressive or isochronous record of their activity and these are shown in Figure 4.4 (c), (d), (e) and (f), respectively. A time-transgressive imprint would also be characterised by cross-cutting bedforms, which have been omitted from Figure 4.4 for clarity. These complex cross-cutting relationships are shown in Figure 4.5 where ice stream operation during retreat creates several populations of bedforms which may appear similar or be cross-cut by younger bedforms. It is anticipated that these conceptual models will help in deciphering the bedform record of palaeo-ice streams.

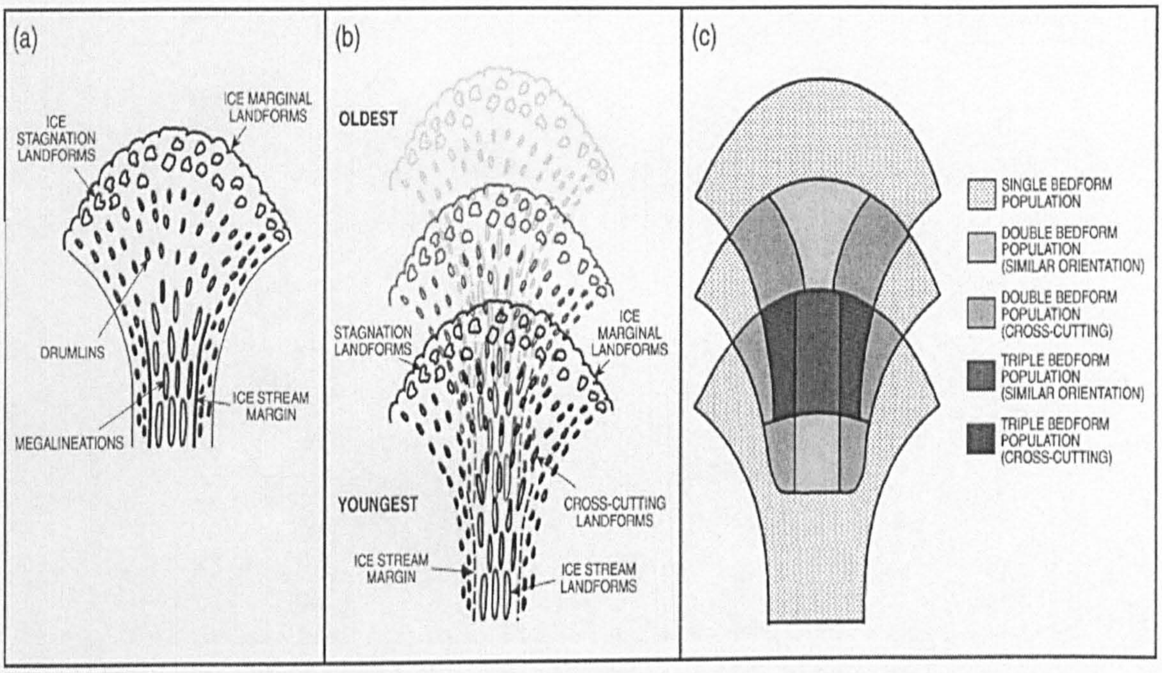

Figure 4.5. Contrasting bedform populations imprinted by isochronous (a) and time-transgressive (b and c) ice stream activity.

4.5.2. Validation of the Geomorphological Criteria.

In an ideal world, the validity of the criteria should be tested prior to their application. However, until we are able to make detailed examinations of the beds of contemporary ice streams, this will not be possible. Instead, we have to rely on glimpses of the bed afforded by seismic and borehole investigations. Alternatively, there is unrivalled access to the beds of former ice sheets. Their geo(morpho)logical products can be used to validate the criteria. The problem here of course, is that ice stream geomorphology has often been modified in a number of ways, and sometimes completely obscured. However, a far greater pitfall when attempting to validate the criteria in this context, are the problems associated with circular reasoning.

Due to the inaccessible nature of contemporary ice stream beds, some of the criteria were partly developed from previous studies on palaeo-ice streams. For example, ice stream marginal moraines have never been identified under contemporary ice streams. However, we may expect subglacial marginal moraines to form at the boundary between the fast and slow flowing ice. More importantly, ice stream marginal moraines *have* been identified from the margins of two inferred palaeo-ice streams (see Dyke and Morris, 1988; Hodgson, 1994). For this reason, ice stream marginal moraines were included as a diagnostic criterion.

In the same way, it is impossible to view a dispersal train beneath a contemporary ice stream. However, a Boothia-type dispersal train has been associated with strong evidence for an inferred ice stream on Prince of Wales Island (see Dyke and Morris, 1988). Because some of the criteria are developed from known palaeo-ice stream tracks, we have no objective way of validating the criteria. It is illogical to develop criteria from the best preserved palaeo-ice stream evidence and then validate their use by investigating the same ice streams.

While a true validation of the individual criteria is impossible, the landsystems models can be compared with known ice stream tracks. Although individual criteria are not necessarily indicative of a palaeo-ice stream, when found in conjunction with other criteria, the strength of the evidence increases. For example, highly attenuated bedforms are not necessarily indicative of ice streams, but when they comprise a convergent flow-set with abrupt lateral margins and ice stream marginal moraines, they provide very strong evidence for ice stream activity. Therefore, the landsystems models or clusters of four or five criteria, can be partly validated by known ice stream locations. This offsets (but does not negate) the problems associated with circular reasoning, because the landsystems models are purely theoretical and have never been identified previously.

4.6. Conclusions.

Several geomorphological criteria have been predicted to aid the identification of palaeo-ice streams. These criteria are developed from a convergence of knowledge gained from contemporary ice stream research, coupled with traditional theories of glacial geomorphology. It is postulated that the following six criteria are fundamental to former ice stream signatures which remain largely unaltered by later ice flow;

1. Characteristic shape and dimension (>20 km wide and >150 km long).
2. Highly convergent flow patterns.
3. Highly attenuated subglacial bedforms (length:width >10:1).
4. Boothia-type erratic dispersal trains (cf. Dyke and Morris, 1988).
5. Abrupt lateral margins (< 2 km).
6. Ice stream marginal moraines.

A further two additional criteria may be present. A former ice stream may also leave evidence of: (7) pervasively deformed sediment, and if the ice stream entered a marine basin, it may deposit a sub-marine till delta or sediment fan (8).

Taking a landsystems approach, the criteria can be grouped to illustrate a perfect, or unaltered palaeo-ice stream track. Marine-based and terrestrial ice streams will produce slightly different signatures, but a more fundamental difference lies in the synchroniety of bedform generation. A distinction is made between an isochronous (rubber stamped) imprint of ice stream activity and that producing a time-transgressive (smudged imprint). These simple conceptual models provide an observational template upon which theories of former ice streams can be better based.

4.7. Application of the Geomorphological Criteria.

The application of the criteria to former ice sheet beds is two-fold. Firstly, the criteria allow us to objectively validate previously cited palaeo-ice stream locations. The absence of individual criteria is just as important as their presence and it is hoped that this will lead to a more objective interpretation of the bedform record whereby preservation and modification is taken into account. Explaining the absence of criteria is just as important as citing their presence. Secondly, the criteria allow us to more objectively identify glacial geomorphology produced by an ice stream and can be used to aid the search for other hypothetical palaeo-ice streams.

With these two applications in mind, the second half of the thesis will concentrate on investigating the exposed beds of two hypothesised palaeo-ice stream locations. The most suitable and efficient way to map glacial geomorphology over large areas incorporates a range of remotely sensed data including satellite imagery, air photographs and Digital Elevation Models. The following chapter outlines the advantages of the various data sources and provides a brief overview of the basic methodology adopted in this thesis.

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Chapter 5: Methodology: Remote Sensing of Ice Stream Geomorphology.

5.1. Introduction.

The principal tool used to investigate the geomorphological signature of palaeo-ice streams in this thesis is satellite remote sensing. The aims of this chapter are to outline the advantages of using remote sensing, to provide an overview of the satellite sensors used, and to introduce the other data sources utilised in this thesis. A further aim of this chapter is to briefly explain the basic methodology used in the identification, interpretation and analysis of glacial lineaments. It should be noted that Chapters 6 and 7 will often elaborate on the basic methodology presented here. For example, the spatial coverage of the imagery differs between the two ice streams and different sampling procedures are adopted. Where necessary, any modifications made to the basic methodology outlined here are specified in the relevant chapter.

5.2. Advantages of Using Satellite Remote Sensing.

The use of satellite remote sensing has revolutionised the identification and mapping of glacial landforms and has greatly improved our knowledge of the dynamic behaviour of former ice sheets. This is because it has the potential to identify regional scale patterns of glacial landforms from which it is possible infer the glaciodynamic context of generation. For example, Punkari (1980; 1985; 1993) used Landsat imagery of Finland and Soviet Karelia to identify the time-transgressive retreat of several lobate ice streams in the Scandinavian Ice Sheet (Section 3.3.4).

The significance of this work lies in the identification of several phases of ice flow which provided information about the dynamic behaviour of the ice sheet during deglaciation. Such patterns were previously undetected because of the limited coverage of aerial photographs and field investigations. Indeed, remote sensing has revealed the highly dynamic nature of whole ice sheets throughout complete growth and decay cycles.

Using imagery from the mainland Laurentide Ice Sheet, Boulton and Clark (1990a; b) detected large-scale patterns of ice sheet flow and their changes through time. This provided comprehensive evidence of a highly mobile ice sheet whereby ice divides migrated considerable distances throughout the glacial cycle. Significantly, this remote sensing investigation also revealed a previously undetected glacial landform, namely mega-scale glacial lineations (Clark, 1993, 1994). For a thorough review of remote sensing of glacial geomorphology the reader is referred to Clark (1997).

There are a number of reasons why satellite remote sensing enables us to advance our understanding of the geomorphological products of ice sheets. This is largely due to the advantages it holds over other mapping techniques of which Clark (1997) outlined five main benefits:

1. Satellite imagery covers large areas and this allows a single user to map geomorphology over very large regions and even whole ice sheets.
2. It is often quicker to map glacial landforms using satellite imagery than by hard copy aerial photography or fieldwork.
3. Some glacial landforms can be detected far more easily on satellite imagery than by aerial photography or fieldwork.
4. Satellite imagery allows the user to work at a range of scales.
5. The large area view of satellite imagery allows the user to discover previously unseen landforms and patterns.

These points illustrate why remote sensing techniques were adopted to map and analyse palaeo-ice stream geomorphology over large areas, and provide ample justification as to why it is the primary technique employed in this thesis. The following section introduces the basic characteristics of the satellite sensors used.

5.3. Satellite Sensors Used in This Thesis.

The three main satellite sensors used in this thesis are Landsat Thematic Mapper (TM), Landsat Multi-Spectral Scanner and Synthetic Aperture Radar (SAR). The spatial coverage and resolution of these three sensors is shown in Table 5.1 where comparison is made with aerial photographs.

Table 5.1. Spatial Resolution and Coverage of Landsat TM, MSS and SAR Imagery Compared to Aerial Photographs.

Sensor	*Spatial Coverage (km)*	*Spatial Resolution (m)*
Landsat TM	185 x 185	30
Landsat MSS	185 x 185	80
SAR	100 x 100	25
Aerial Photographs (1:50,000)	13 x 13	1.5

5.3.1. Landsat Thematic Mapper and Multi-Spectral Scanner.

Landsat TM and Landsat MSS satellites sense wavelengths in the visible portion of the spectrum and are thus referred to as optical sensors. The MSS sensor provides 4 wavebands (1, 2, 3 and 4) at a spatial resolution of 80 m (see Table 5.1). The TM sensor provides another three bands (5, 6 and 7) providing information from the mid infra-red and thermal end of the spectrum. In addition, Landsat TM has a spatial resolution of 30 m and it is more expensive than MSS imagery.

For the purpose of mapping glacial geomorphology, the data is usually displayed as a monochrome image because it is well known that colour distracts the eye and often masks structure and pattern (cf. Clark, 1997). When using TM imagery, bands 5 and 7 are the most useful for detecting vegetation and soil moisture content and for discriminating between rock and mineral types. For this thesis, it was found that a monochrome image of band 5 was most suitable for detecting glacial lineations on TM imagery and band 4 was most suitable on MSS imagery. This displays land on a grey scale and water appears black.

Unfortunately, the wavebands of TM and MSS also detect cloud and so it is imperative to acquire cloud free scenes. For Arctic areas , such as northern Canada, this is often quite a challenge and it is further compounded by snow covered ground and the short length of the summer. The images used were restricted to the summer months when cloud was less likely and snow cover scarce or absent. However, an advantage of the Arctic summer is that the sun angle is low and this aids the identification of topography. For mid-latitude regions it would be necessary to acquire winter images to obtain the same effect.

The major advantage of satellite imagery is its digital format, but it is also possible to obtain photographic products of Landsat TM and MSS images. These hard copies are often supplied in the form of mosaics whereby several scenes are joined together to cover a geographic zone. It is also possible to obtain 12" negatives (or positives) of the imagery to develop personalised hard copies at a chosen scale. The main advantage of these products is that they are extremely cheap compared to the digital formats. Thus, it is possible to map flow patterns over large areas before selecting the most suitable digital imagery. For this reason, several mosaics and negatives were used in the study.

5.3.2. Synthetic Aperture Radar.

Synthetic Aperture Radar (SAR) utilises microwave energy and operates using entirely different principles compared to the Landsat sensors outlined above. It sends out microwave energy at an oblique angle and records the returned echo from the ground. The advantage of the oblique angle is that it preferentially highlights topography. Because of this, and the fact that the spatial resolution is good, SAR imagery is often seen as more favourable for mapping geomorphology than optical imagery. Indeed, SAR imagery is ideal for gently undulating terrain such as drumlin fields and because SAR emits its own radiation, it does not rely on illumination from the sun. This means it can be used both night and day and because the microwaves pass through cloud, it eliminates the problem of having to acquire cloud free images. However, there are difficulties associated with the use of SAR.

A fundamental problem stems from the oblique viewing geometry of the sensor, but this also inhibits optical imagery. This means that landforms orientated perpendicular to the sensor are preferentially highlighted whereas those which are more parallel are often under-represented. Hence, the look direction of the sensor is of paramount importance and the user should be familiar with the bias it creates.

Another problem with SAR imagery is known as 'speckle' which often appears as noise in the image. This is a manifestation of the fluctuation in back scatter but can often be partially rectified using image processing techniques. A thorough review of the use of SAR imagery for landform mapping is provided by Vencatasawmy *et al.* (1998), to which the reader is referred for a more in depth discussion.

In this study, different sensors were acquired to map different areas but in some cases the imagery overlapped. This proved advantageous in that it provided a first order check of the mapping. The actual spatial coverage of the specific imagery used is outlined in more detail in the relevant chapters to follow (6 and 7).

5.4. Aerial Photographs.

To complement the data sources outlined above, aerial photographs were obtained for selected areas. The main purpose of this was to provide an invaluable proxy for 'ground truthing' the lineaments identified on the imagery. Indeed, it was sometimes necessary to recourse to aerial photography to establish the relative chronology of cross-cutting relationships, for example.

The extremely good spatial resolution of air photographs also permitted the identification of features which could not be adequately resolved on the satellite imagery. Such features were often seen as a luxury for the purpose of the regional mapping of flow-sets, but the identification of drift exposures and small eskers, for example, often proved very useful.

A product of aerial photography are photographic maps (or photomaps) at a scale of 1:50,000. The advantage of these maps over aerial photographs is that they are geometrically corrected and display elevation data at 10 m contour intervals. The primary use of these photomaps (where available) was to provide enough information to construct topographic transects and crude digital elevation models from the study areas. Moreover, the supplementary data provided by air photographs and photomaps meant that there was a scale-unbiased data set where it was possible to map features at a resolution of 1.5 m up to >20 km.

5.5. Digital Elevation Models.

Digital Elevation Models, or DEM's, are raster-based models of topography where individual grid cells represent the average elevation within that given area. High resolution DEM's (metres) are extremely advantageous for landform mapping but they are generally not available for most previously glaciated terrain. This explains why crude topographic profiles and transects were often constructed from the photomaps outlined in Section 5.4. However, it was often necessary to view the topography at the regional scale and it was possible to do this using the best resolution DEM available free of charge to the public.

GTOPO30 is a DEM produced by the US Geological Survey's EROS Data Centre. Its primary aim is to provide regional and continental scale topographic data from all over the globe. Once downloaded, it provides a raster layer composed of elevation data regularly spaced at 30-arc seconds which represents a spatial resolution of around 0.5-1 km. This permitted rapid, and relatively accurate surface profile information as well as providing a general visualisation of regional topography.

5.6. Methodology: Mapping Glacial Lineaments.

5.6.1. Terminology and Assumptions.

This section outlines the basic methodology adopted to map glacial lineaments on the satellite imagery. For the purpose of this thesis, a *glacial lineament* is defined as a distinguishable linear expression formed by the action of overriding ice (cf. Clark, 1993). Glacial lineaments can often be linked together and assigned to a *flow-set* which Clark (1990) defines as "integrated groups of flow lines revealing a widespread pattern of flow". The term *flow-event* is used here to describe the glacio-dynamic behaviour which produced the flow-set, e.g. an ice stream.

When mapping glacial lineaments the following assumptions are made;

- lineaments are formed parallel to ice flow, (excepting transverse ridges, e.g. ribbed moraine),
- lineaments are formed under warm-based ice,

- coherent lineament patterns (or flow-sets) are related to individual flow-events,
- superimposition or cross-cutting lineaments can be used to reconstruct relative ice flow chronologies.

An inherent problem when mapping glacial lineaments on satellite imagery is that geological features may be mis-interpreted as drift lineaments formed by ice. Exceptional care must be taken to avoid this. In general, drift lineaments have a much smoother appearance on imagery whereas geological lineaments have a more granular appearance. Furthermore, geological lineaments are often far more irregular in their spacing and more angular in appearance. Although, these general rules are very useful, care was taken to compare all lineament mapping with published geological maps from the areas under investigation. This eliminated geological lineaments, relict strandlines and raised beaches from the data-set, all of which can appear similar to glacial lineaments.

5.6.2. Methodology.

The basic methodology for analysing glacial lineaments in this thesis involves three main steps;

1. The *identification* of glacial lineaments on satellite imagery.
2. The *interpretation* of these lineaments into distinct flow-sets.
3. The *measurement and analysis* of the morphometry of the lineaments within each flow-set.

Figure 5.1 shows the three steps involved when mapping glacial lineaments on satellite imagery and the following section describes each step in detail.

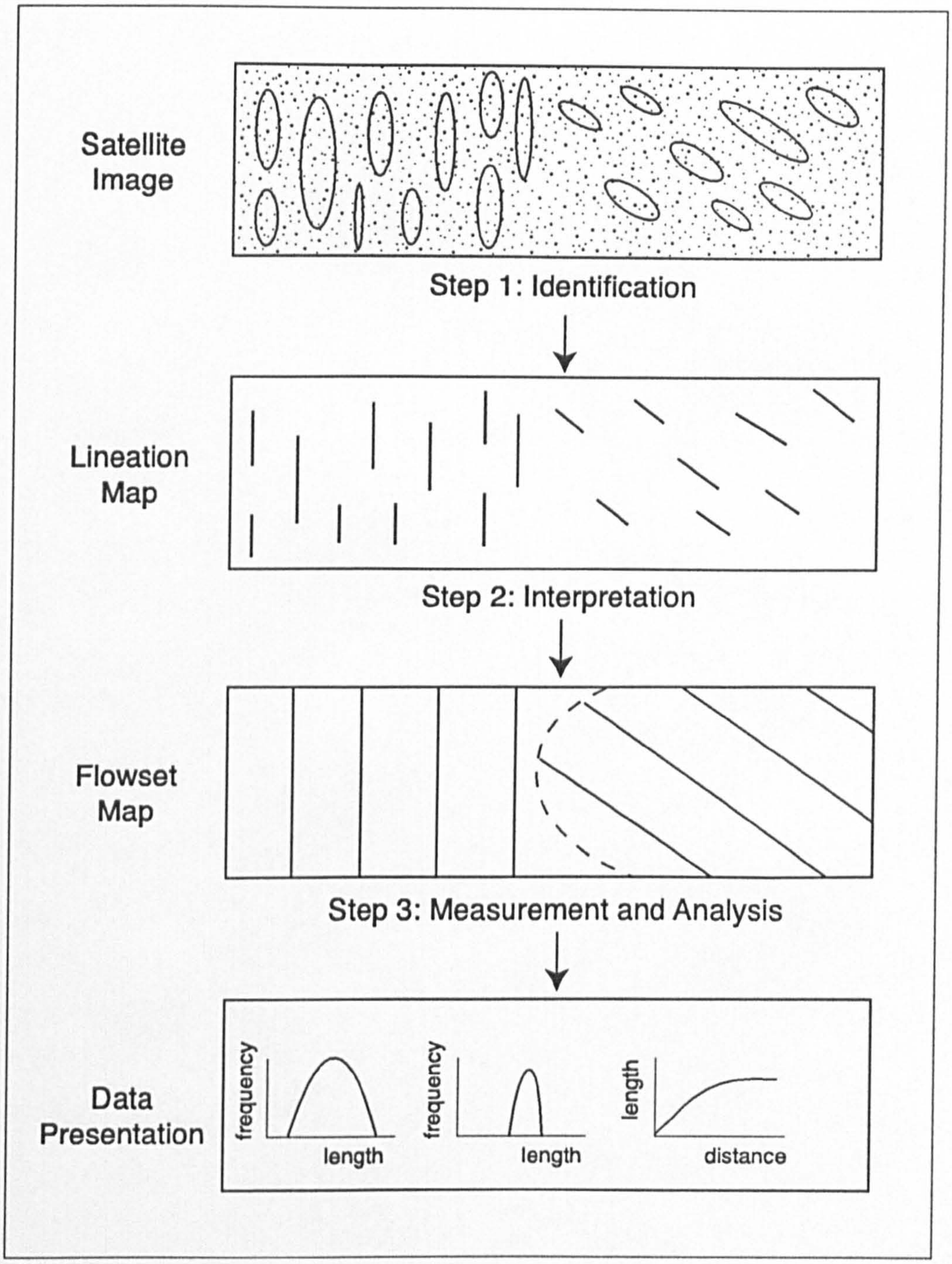

Figure 5.1. Schematic representation of the methodology used for the detection, mapping and analysis of glacial lineaments on satellite imagery, see text for details.

5.6.2.1. The Identification of Glacial Lineaments.

The identification of lineaments (Step 1, Figure 5.1) was undertaken using the digital satellite imagery. This allowed bedforms to be mapped at a variety of scales (1:30,000 up to 1:500,000). Using image processing techniques where necessary, it was also possible to preferentially highlight topography to increase the level of detection. In some cases, it was also possible to use more than one source of imagery and this allowed complimentary data sets to be constructed.

Satellite imagery were displayed and manipulated on Sun work-stations using ERDAS Imagine 8.3 software. Bedforms were digitised on-screen by drawing a line along the long axis of the lineament and all data were stored as ARC coverages. This produced several lineation maps in a Geographical Information System (GIS) which provided the basis for the next step.

To verify the lineation maps, aerial photographs and maps of surficial geology were obtained, together with previous studies in the field area. In most cases it was found that the satellite imagery agreed very well with the published sources and it was usual for more lineaments to be identified on the imagery.

5.6.2.2. The Interpretation of Flow-sets.

This is arguably the most subjective step in the methodology and involved the grouping of lineaments into distinct flow-sets (Step 2, Figure 5.1). Ideally, it may be possible to group bedforms into flow-sets using their statistical properties. For example, it may be possible to enter the dimensions of several hundred bedforms and perform some form of cluster analysis whereby groups of similar landforms are objectively grouped together. For several reasons, this approach has often proved fruitless in the quantitative analysis of glacial geomorphology. This is because statistical analysis may group together bedforms with identical lengths, widths and densities, but the bedforms may have subtle differences in orientation which proves glacially implausible (e.g. cross-cutting landforms).

With this in mind, it was decided to construct flow-sets on the basis of visual properties and then test the inference that different flow-sets exhibit different morphological properties using statistical analysis. Although this technique is subjective, several criteria were used and the grouping of lineaments into flow-sets was based on some combination of the following criteria (cf. Clark, 1999);

(a), parallel concordance (i.e. similar orientations),

(b), close proximity to each other,

(c), similar morphometry (length, width, elongation ratio),

(d), similar spacing (i.e. density),

(e), glacial plausibility.

These five criteria were used to group the raw lineament data into distinct flow-sets.

This second step could again be verified by published maps and papers and in particular, investigations concerning the glacial history of the study area. The construction of flow-sets was always done prior to consulting the literature so that bias was avoided. Indeed, it was sometimes the case that the flow-set chronology contradicted published interpretations and in such cases it was necessary to provide substantial supporting evidence from the measurements and analyses (step 3) to support the re-interpretation.

5.6.2.3. The Measurement and Analysis of Lineament Morphometry.

The third step involved measurement and analysis of the morphometry of the bedforms within each flow-set (Step 3, Figure 5.1). This was done for two main reasons. Firstly, it made it possible to use morphometric data to statistically test the inferences that different flow-sets display different characteristics. Secondly, by measuring the characteristics of the bedforms it may also be possible to identify trends *within* a flow-set. This may provide further information regarding the glacio-dynamic processes which formed it.

Lineaments could be measured on screen, or by printing out hard copies. This could be done for all bedforms in a flow-set or for samples of bedforms within a flow-set. In general, the following data were collected: lineament length, width, elongation ratio, orientation, parallel conformity (standard deviation of a sample of bedform orientations; 0 = parallel), density (number of bedforms per unit area) and packing (surface area of bedforms per unit area).

For large flow-sets, grids could be overlaid which allowed the systematic sampling of bedforms. More importantly, sampling grids could be prescribed to reflect flow-bands within the flow-set and this allowed within flow-set variations in lineaments to be identified. These sampling procedures (where necessary) are explained in more detail in Chapters 6 and 7.

5.7. Summary.

Remote sensing is the most efficient way to map glacial geomorphology at the regional scale. This is because of the numerous advantages it holds over fieldwork and aerial photography, specifically the wide spatial coverage, the high level of detection and the ease and cost of use. Optical sensors such as Landsat TM and MSS are easy to use and provide data from several wavebands. Synthetic Aperture Radar (SAR) uses microwave energy and has the advantage of producing cloud free scenes capable of highlighting undulating topography.

To complement these data sources, aerial photography and photomaps can be used to provide detail at a spatial resolution of up to 1.5 m. On a larger scale, Digital Elevation Models provide valuable information with respect to the regional topography and allow rapid surface profile transects to be calculated.

The basic methodology adopted to map glacial lineations can be considered a three step process:

(1). The identification of glacial lineaments on satellite imagery.

(2). The interpretation of these lineaments into distinct flow-sets.

(3). The measurement and analysis of the morphometry of the lineaments within each flow-set.

The following two chapters investigate the geomorphology of two palaeo-ice stream beds using the techniques and basic methodology outlined here.

❄ ❄ ❄

Chapter 6: The M'Clintock Channel Ice Stream.

6.1. Introduction and Rationale.

In this thesis, several geomorphological criteria have been identified from the literature as being indicative of ice stream activity (Chapter 4). Application of these criteria to published accounts of palaeo-ice stream hypotheses (reviewed in Chapter 3) revealed that the one with possibly the most compelling terrestrial evidence is that described by Hodgson (1993; 1994), and here called the M'Clintock Channel Ice Stream. According to Hodgson (1993; 1994), it would appear that the ice stream displays at least four of the criteria; characteristic shape and dimensions, highly attenuated bedforms, an abrupt lateral margin, and ice stream marginal moraines, see Table 3.1.

This chapter contributes to our understanding of ice streams by investigating the M'Clintock Channel Ice Stream. The chapter begins by outlining the previous work on the ice stream and critically assesses the validity of the evidence used in its identification. Having established that it represents a valid candidate for a palaeo-ice stream track, its basal characteristics are described and its extent is reconstructed. Implications for ice stream functioning and basal processes are explored.

6.2. Previous Work.

Several palaeo-ice streams have been postulated from the Canadian Arctic, where topographic channels are thought to have preferentially captured fast-flowing ice. One such ice stream has been postulated by Hodgson (1994) in the M'Clintock Channel, which drained the north-western margin of the Laurentide Ice Sheet, see Table 3.1. Figure 6.1 shows the location of the M'Clintock Channel, the generalised bedrock geology and the reconstructed margin of the Laurentide Ice Sheet at around 10,000 yr BP.

Hodgson (1993; 1994) identified and dated three diverse flow patterns on Storkerson Peninsula, Victoria Island. The coverage of Hodgson's (1993; 1994) fieldwork mapping is indicated by the box on Figure 6.1 and Figure 6.2 shows the reconstructed ice flow patterns of which Flow-2 is the inferred ice stream.

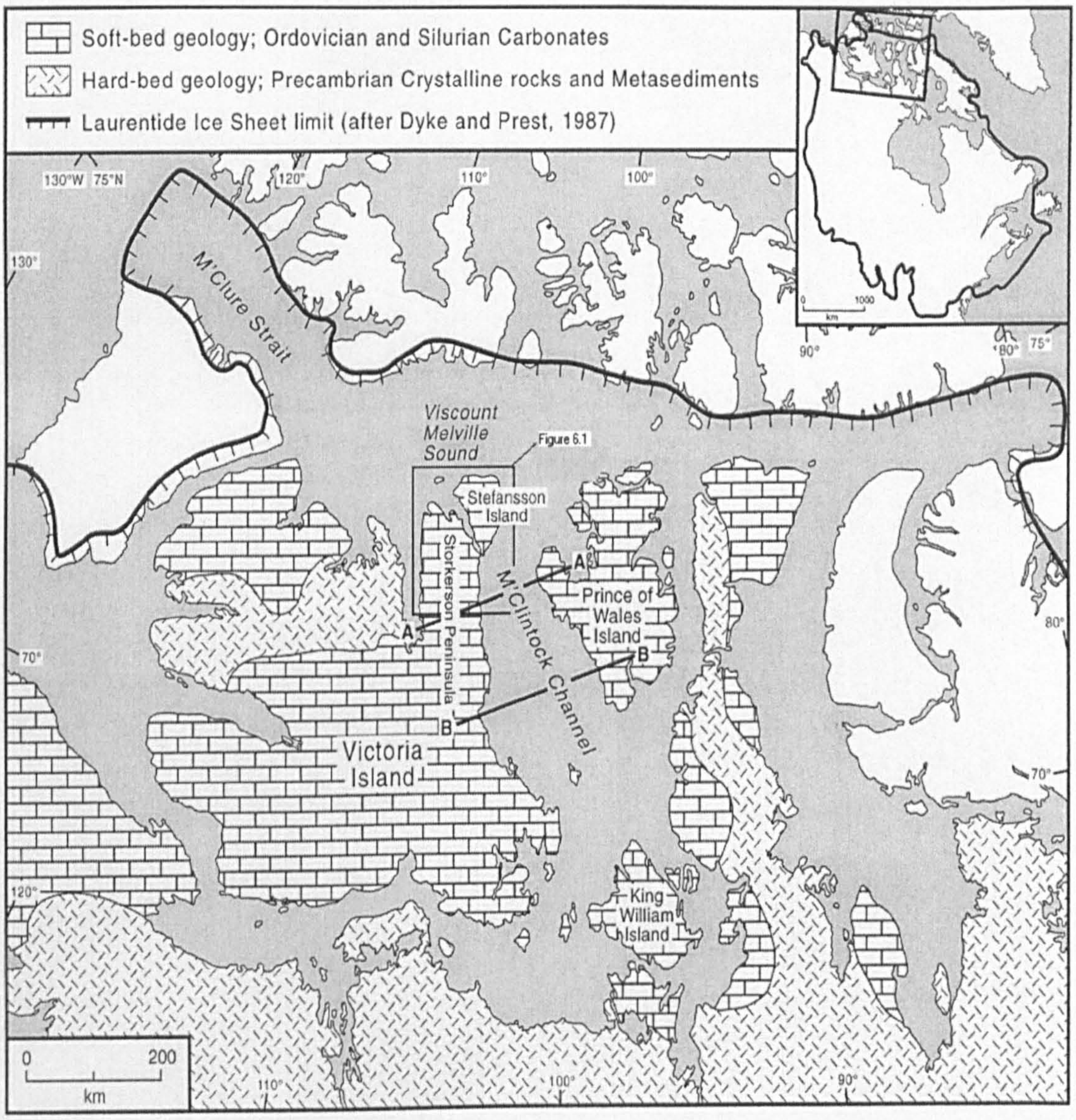

Figure 6.1. Location map of the M'Clintock Channel, Canadian Arctic Archipelago, showing generalised bedrock geology and the area investigated by Hodgson (1993; 1994).

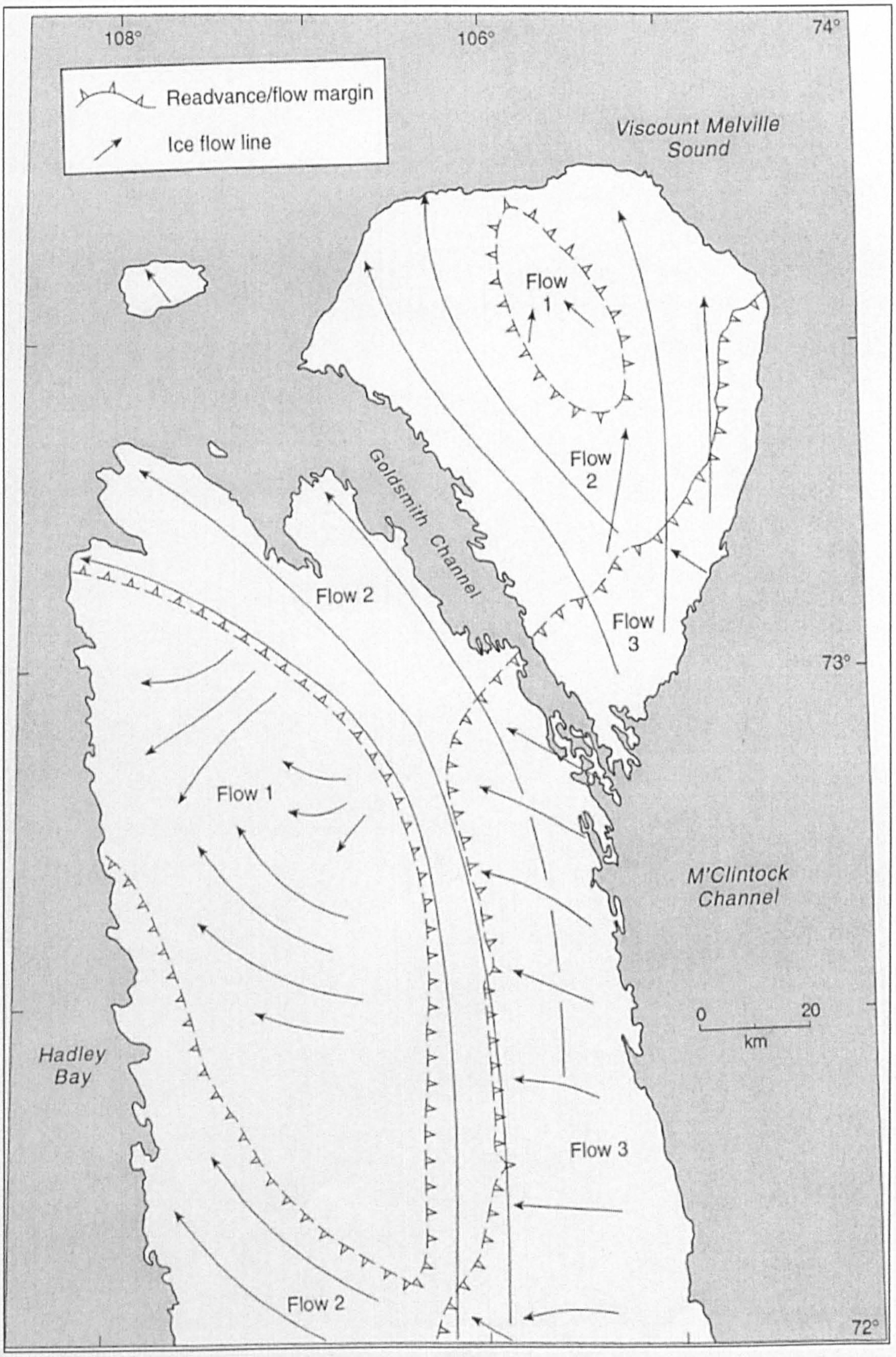

Figure 6.2. Reconstructed ice flow patterns on Storkerson Peninsula, Victoria Island, after Hodgson (1994). Flow-1 represents the oldest landforms which were cross-cut by the Flow-2 (ice stream). Flow-3 is that youngest ice flow and had a modifying influence on Flow-2 (modified from Hodgson, 1994).

A combination of relative age dating between overlapping flows, radiocarbon dating of marine molluscs and consideration of relative marine limits allowed Hodgson (1994) to reconstruct the ice flow chronology. The oldest landforms are represented by Flow-1 on western Storkerson Peninsula (see Figure 6.2). These relatively small drumlin fields are separated by hummocky till and it is speculated that they record minor surges of the ice margin into Hadley Bay before 10,400 yr BP.

On northern Storkerson Peninsula, the Flow-1 drumlins have been cross-cut by Flow-2. Flow-2 landforms are highly attenuated and reach length:width ratios of up to 20:1. This drumlin field displays a remarkably abrupt margin and Hodgson (1994) postulated that these landforms mark the western edge of what was probably a much larger ice stream extending some way into M'Clintock Channel. The eastern branch of the Flow-2 bedforms indicated that the ice stream overtopped Stefansson Island from which Hodgson inferred that ice thicknesses must have been in excess of 400 m and probably more to account for the lack of topographic deflection. The timing of Flow-2 (ice stream activity) was constrained to between 10,400 yr BP and 9,600 yr BP.

On eastern Storkerson Peninsula, the Flow-2 landforms have been cross-cut by Flow-3 drumlins. Flow-3 represents a readvance for a short distance on-shore (westward) which made minor modifications to the Flow-2 (ice stream) landforms. Hodgson (1994) speculated that Flow-3 bedforms were laid down between 9,600 and 8,800 yr BP by a buoyant tidewater glacier in M'Clintock Channel which thrust marine sediments on land.

6.2.1. Evidence for Ice Stream Activity.

In an effort to introduce a more objective approach to studying ice stream geomorphology, several geomorphological criteria have been predicted to aid the identification of palaeo-ice streams (see Chapter 4; Stokes and Clark, 1999). If the M'Clintock Channel Ice Stream is a valid candidate for a former ice stream track we would expect to find several of these criteria manifest in the glacial geomorphology of Hodgson's (1993; 1994) Flow-2 drumlin field. The documented evidence indicates that at least four of the criteria are present, possibly five.

Firstly, the ice stream is the characteristic shape and dimension (i.e. similar to contemporary ice streams). Hodgson (1994) suggested that this ice stream was probably around 80 km wide and several hundreds of kilometres long. This clearly places it within the characteristic shape and dimensions of contemporary ice streams (see Table 2.1). Secondly, the inferred ice stream bedforms are highly attenuated. According to Hodgson's (1993; 1994) fieldwork, Flow-2 landforms approach 10 km in length and possess elongation ratios of up to 20:1, some of the highest recorded in the literature. Thirdly, Hodgson (1994) noted that the Flow-2 (ice stream) drumlin field exhibits a very abrupt margin, a unique characteristic of ice streams. Fourthly, Hodgson (1994) identified "massive extended drumlin ridges" which are very similar to that found by Dyke *et al.* (1992) on Prince of Wales Island, who described it as a lateral shear moraine formed at the boundary between fast and slow flowing ice. In addition to these criteria, Hodgson (1994) speculated that Flow-2 landforms deformed and eroded a widespread cover of till. Thus, deformable sediments may be associated with this portion of the ice sheet.

The M'Clintock Channel Ice Stream fulfils four (possibly five) criteria for palaeo-ice stream activity. Apart from a Boothia-type dispersal train, there appears to be no evidence to suggest that the other criteria (convergent flow patterns and offshore sediment accumulations) may be absent, they just haven't been identified yet. In conclusion, the evidence suggests that the M'Clintock Channel Ice Stream is an excellent candidate for a former marine-based ice stream.

6.2.2. Specific Aims.

This ice stream imprint merits further study for a number of reasons. The landforms in the Canadian Arctic Archipelago are remarkably well-preserved and this 'fresh' topography permits extremely detailed mapping of the glacial geomorphology. It thus provides an opportunity to investigate the within ice stream variations in bedform morphometry with a view to explaining the basal processes which produced them. Also, the M'Clintock Channel is comparable in size to Hudson Strait. If an ice stream occupied the whole of this channel, it would have had a profound effect on the north-western portion of the Laurentide Ice Sheet, with likely implications for the stability of the whole ice sheet. Furthermore, the islands in the Canadian Arctic (particularly Victoria Island and Prince of Wales Island) display a diverse range of flow patterns. It is very important to investigate the role of ice stream activity to better understand the glacial history of this very complex portion of the Laurentide Ice Sheet.

The aims of this chapter are to extend the work of Hodgson (1993; 1994) by answering the following key questions concerning the ice stream;

1. What is the nature of its bed in terms of geomorphology, soft-hard bed fraction, lithology, roughness, and slope?
2. What were its spatial dimensions, thickness, and potential ice flux?
3. For how long did it operate and when?
4. What controlled its overall position and the location of its margins?
5. What controlled its activation and deactivation?
6. Do the characteristics of the subglacial bedforms reveal anything about ice stream functioning?
7. Can anything be inferred about the subglacial processes that promoted fast ice flow?
8. What effect did the ice stream have on Laurentide Ice Sheet behaviour?

6.3. Methodology and Data Sources.

Hodgson's field mapping was restricted to north of 72° and west of 104° (see Figure 6.2). To cover a much wider area, a variety of satellite imagery was obtained, which are well suited for mapping glacial geomorphology (cf. Chapter 5). One digital Landsat TM (30 m resolution, map orientated 112 by 56 km) and four SAR (ca. 25 m, 100 x 100 km) images were acquired, geocoded and processed to highlight geomorphology. In addition, 30 Landsat MSS (80 m, 185 x 185 km) hard copy positives were obtained and developed into 1:250,000 photographs. Together, this imagery covered the whole of Storkerson Peninsula and M'Clintock Channel, including the neighbouring islands, such as Prince of Wales Island and King William Island (see Figure 6.1).

To complement the satellite imagery, aerial photographs (scale 1:30,000) were obtained for the ice stream bedforms on Storkerson Peninsula. In addition, eight photomaps (1:50,000) with 10 m contour intervals were also obtained for the whole of Storkerson Peninsula. This source of information proved an invaluable proxy for 'ground truthing' the data on the images and also permitted the identification of extremely small-scale features which could not be adequately resolved by the satellite data.

To gain a broad indication of the general topography on and around Victoria Island and M'Clintock Channel, a 30-arc second (*ca* 0.5-1 km) Digital Elevation Model (DEM) was obtained (see Section 5.5). This permitted visualisation of topography and surface roughness and was used to extract surface profiles along transects. More accurate elevation data were obtained directly from 1:50,000 topographic maps (contour interval 10 m), and for a part of the ice stream margin these were used to construct a DEM with a grid size of 1 km. The Geological Survey of Canada's Map 1817a (Hodgson, 1993) provided detailed information regarding the geology of Storkerson Peninsula

The first task was to independently map the glacial geomorphology and then compare it against Hodgson's (1993) results from field mapping. This provided validation of the procedure, the potential to add further large-scale information, and characterisation of the ice stream bed. The second task was to map a much wider area to define the overall extent of the ice stream.

Mapping was accomplished by visual interpretation and on-screen digitising of landform information directly into a Geographical Information System (following the basic methodology outlined in Section 5.6). The following morphometric measurements of glacial lineaments were conducted: length, width, elongation ratio, orientation, parallel conformity (standard deviation of the orientation of a sample of bedforms), degree of packing (bedform surface area per unit area) and density (bedform number per unit area).

To analyse variations in bedform morphometry within the ice stream, a 10 x 10 km sampling grid was overlaid. This permitted unbiased sampling of bedform populations and allowed flow-bands to be reconstructed in the downstream direction and binned at 10 km intervals. Furthermore, the 10 x 10 km grid squares ensured that the longest ice stream bedforms would be incorporated and that each grid square contained enough bedforms for meaningful analysis.

6.4. Results.

6.4.1. Morphometric Bedform Characteristics of Flow-sets.

Figure 6.3 shows a Landsat TM image of a part of Storkerson Peninsula and the reconstructed flow patterns of Hodgson (1994). Even at this scale, the remarkable contrast in geomorphology is evident and Hodgson's three flow events are clearly visible. Of greatest note is the extremely abrupt margin between the Flow-2 ice stream drumlins and the Flow-1 hummocky terrain to the west. In the eastern part of the TM image and especially in the north-east, overprinting of the Flow-3 readvance on lineaments of Flow-2 can be seen.

Results of mapping from the TM image (Figure 6.3) and the SAR imagery, which covered a greater area, are presented in Figure 6.4. The most striking feature of the lineament map is the prominence of the ice stream bedform signature which is traceable for over 120 km in length and up to 40 km in width, without any apparent discontinuities. Moreover, the ice stream lineament are characteristically longer, closer together and more parallel, compared to other flow patterns.

The digital mapping picked out more lineaments than those mapped by Hodgson (1993) and it can be seen that Flow-1 landforms represent several patterns of divergent ice flow. Although they all relate to the same broad pattern of flow towards Hadley Bay (west), these bedforms were grouped into nine different flow-sets (Flow-1a to 1i), largely on the basis of different orientations. Similarly, Flow-3 landforms appeared to be represented by three separate flow-sets, again, represented by subtle differences in orientations. On the other hand, Flow-2 landforms represent only one coherent flow pattern traceable over the whole of Storkerson Peninsula. Figure 6.5 shows the locations of all of the flowsets identified on Storkerson Peninsula. Table 6.1. shows the basic morphometric characteristics of all of the flow-sets sampled from three transects across Storkerson Peninsula.

(a)

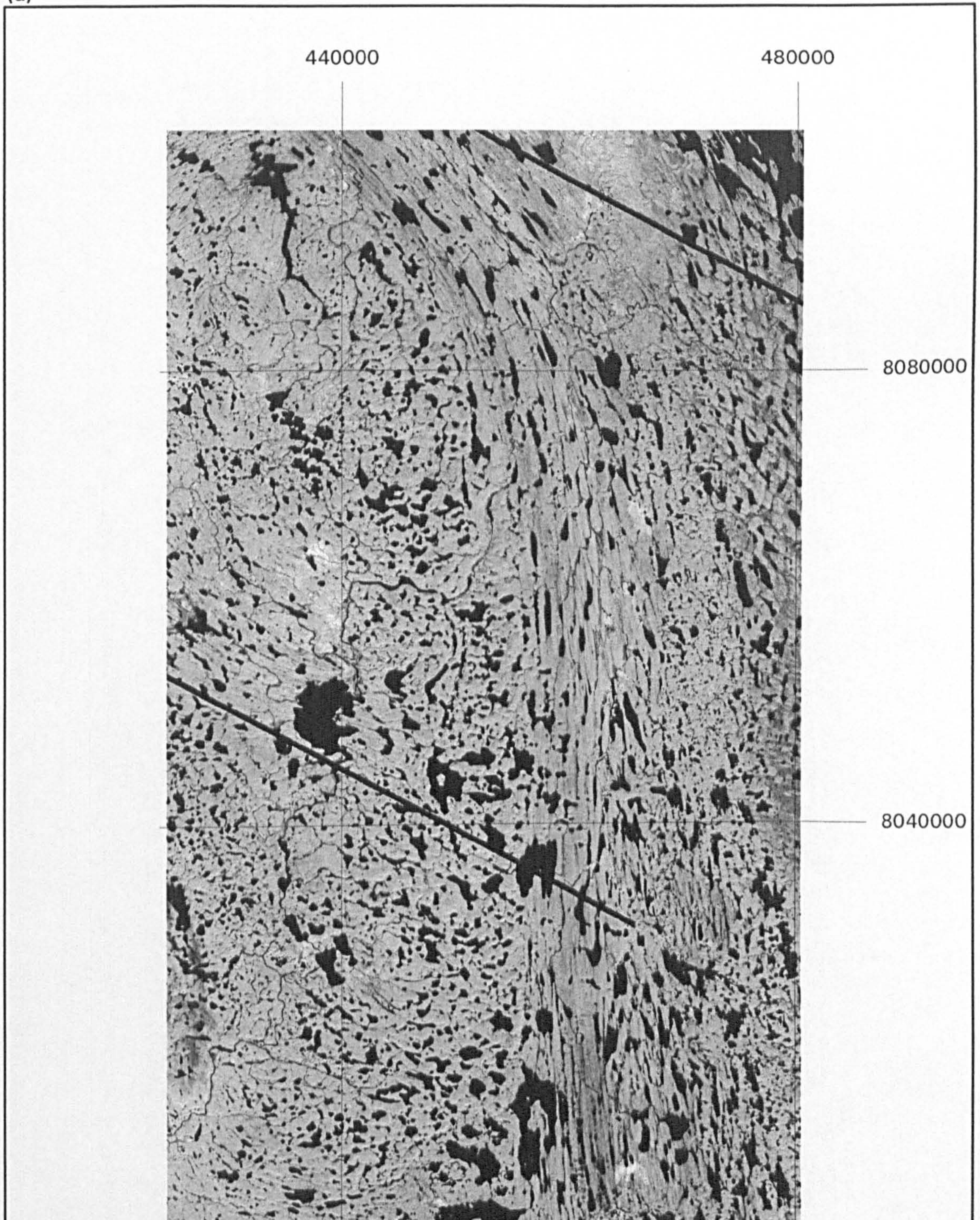

Figure 6.3a. Landsat TM image (band 5) of Storkerson Peninsula, Victoria Island. Note the extremely abrupt margin of the Flow-2 (ice stream) drumlin field.

(b)

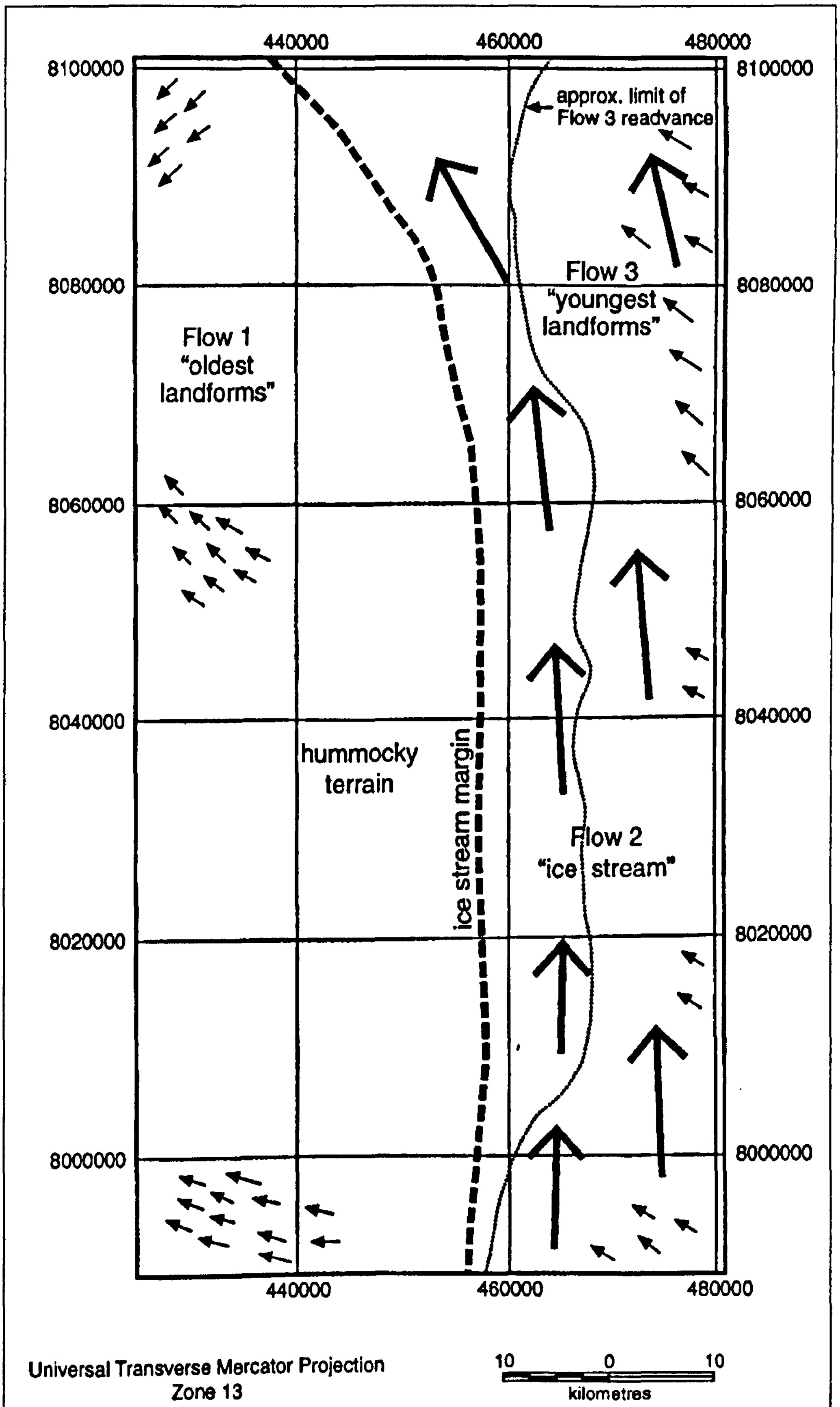

Figure 6.3b. Hodgson's (1994) reconstructed flow patterns on Storkerson Peninsula.

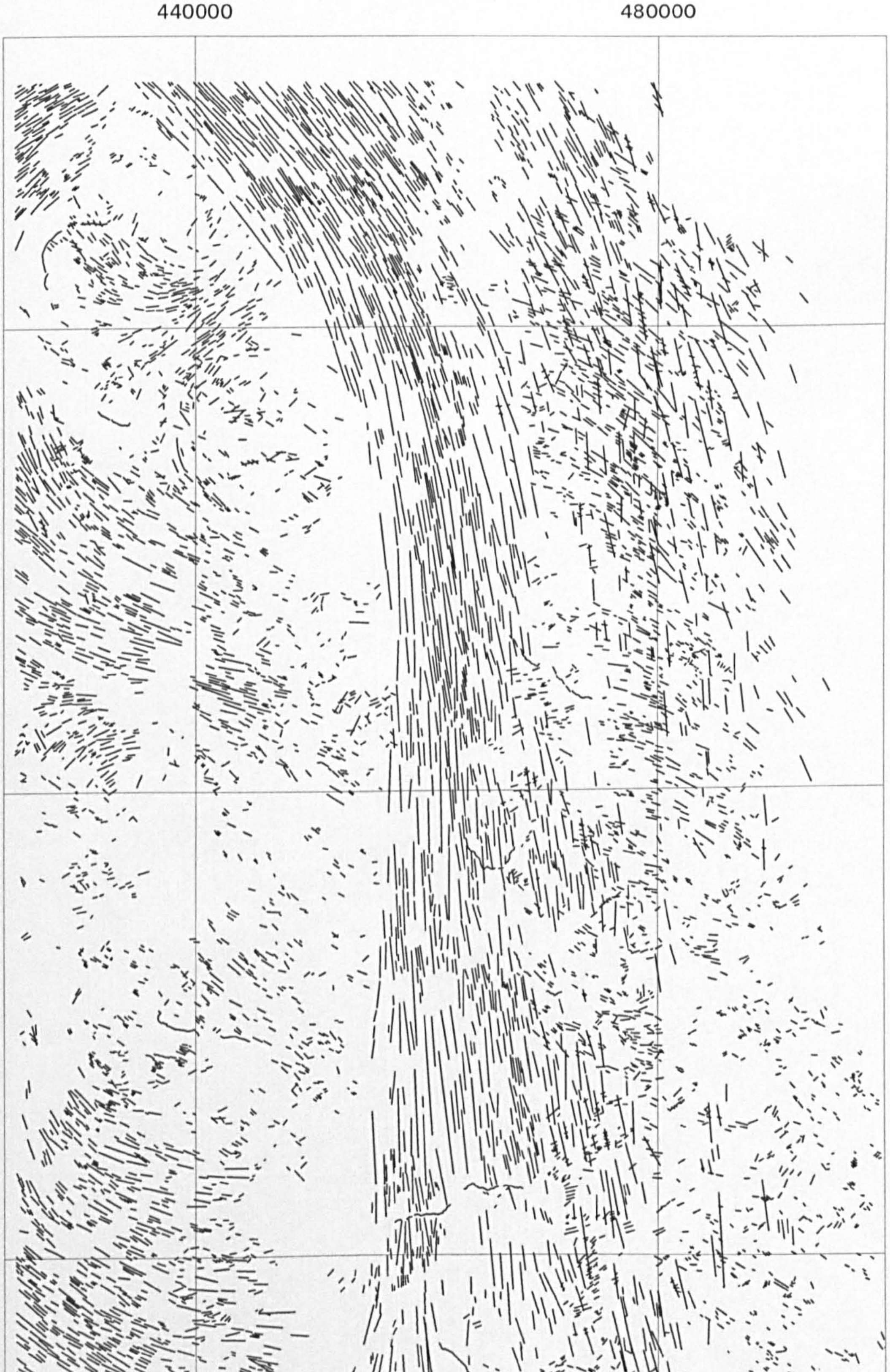

Figure 6.4. Lineaments detected on Landsat TM and SAR imagery of Storkerson Peninsula, Victoria Island (see Figure 6.3a for TM image coverage).

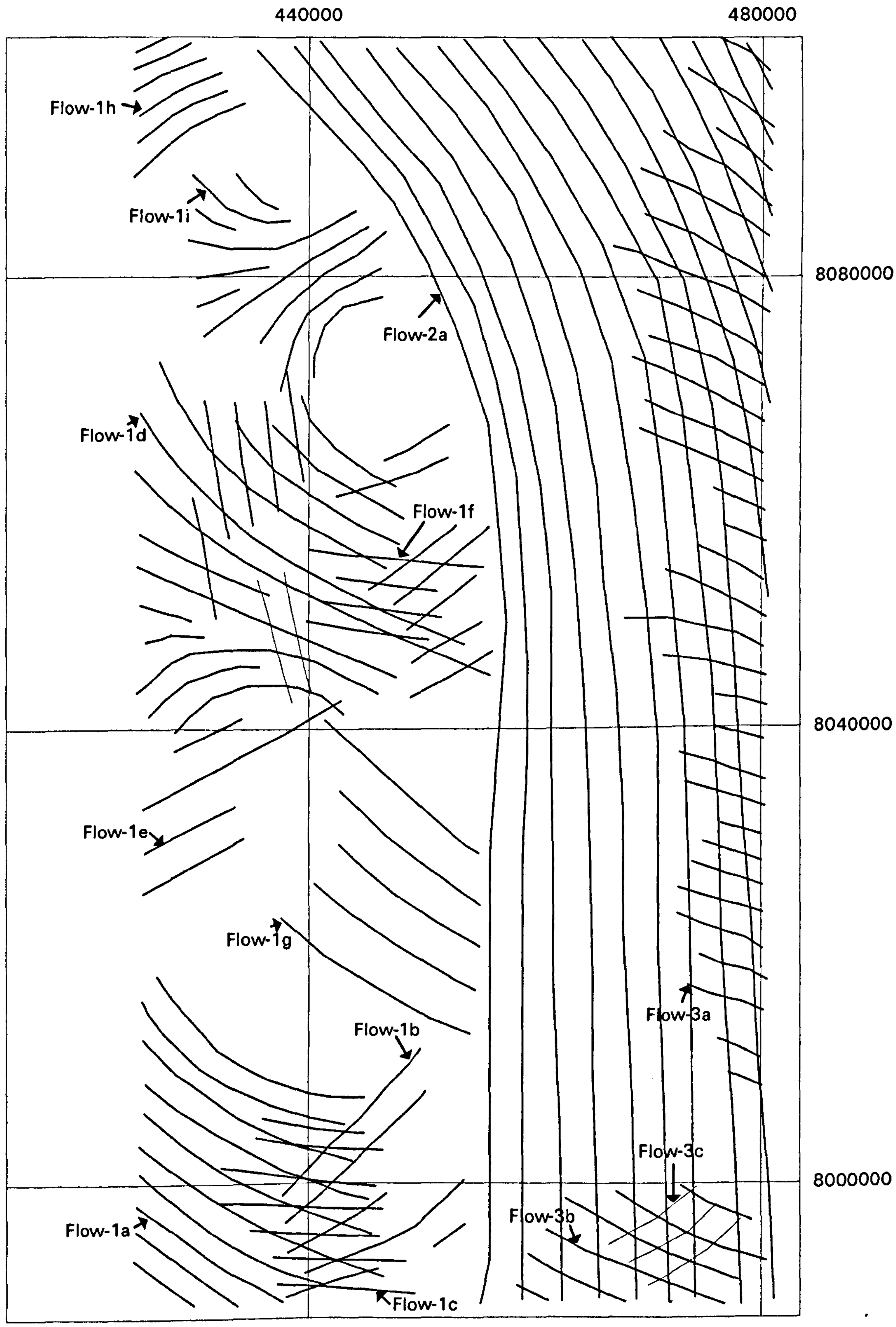

Figure 6.5. Flow-sets compiled from lineament data on Storkerson Peninsula.

Table 6.1. Simple dimensions and orientations of all flow-sets covered in the three transects (see Figure 6.5 for flow-set locations).

Flow-set	n	Mean Orientation degrees	Mean length (max.) metres	Mean width (max.) metres	Mean l:w ratio (max.)
1a	110	332 (NNW)	938 (2500)	183 (400)	4.9 (11.8)
1b	34	207 (SSW)	749 (1800)	165 (225)	4.5 (8)
1c	19	270 (W)	1182 (2300)	184 (250)	6.7 (15.3)
1d	109	336 (NNW)	860 (2950)	209 (525)	4 (9.8)
1e	20	227 (SW)	578 (1000)	173 (300)	3.3 (4.8)
1f	8	269 (W)	609 (1125)	138 (175)	4.3 (7.5)
1g	4	301 (NWW)	619 (800)	194 (225)	3.2 (3.6)
1h	9	225 (SW)	575 (800)	189 (250)	3.1 (3.7)
1i	7	325 (NNW)	500 (675)	211 (325)	2.4 (2.9)
2	**486**	334 (NNW)	**1709 (7950)**	**247 (750)**	**6.9 (27)**
3a	125	323 (NNW)	793 (2800)	196 (650)	4.1 (10.4)
3b	328	322 (NNW)	471 (1400)	122 (250)	3.9 (12.5)
3c	83	218 (SSW)	496 (1350)	128 (225)	3.9 (8.5)

It can be seen from Table 6.1 that the ice stream bedforms (Flow-set 2) are by far the most numerous (n=486). In addition, the ice stream bedforms are substantially longer, wider and more elongated than any of the other flow-sets. Maximum values of length (7,950 m), width (750 m) and elongation ratio (27:1) are also represented by the ice stream bedforms.

In addition to the obvious visual differences, statistical analyses support the inference that different flow-sets exhibit different characteristics. For example, Flow-1a, 2 and 3a are statistically different (according to t-tests) in terms of lengths, widths and elongation ratios. Furthermore, by plotting mean length versus mean width for all of the flow-sets, it can be seen that the ice stream flow-set falls into a distinct cluster, see Figure 6.6.

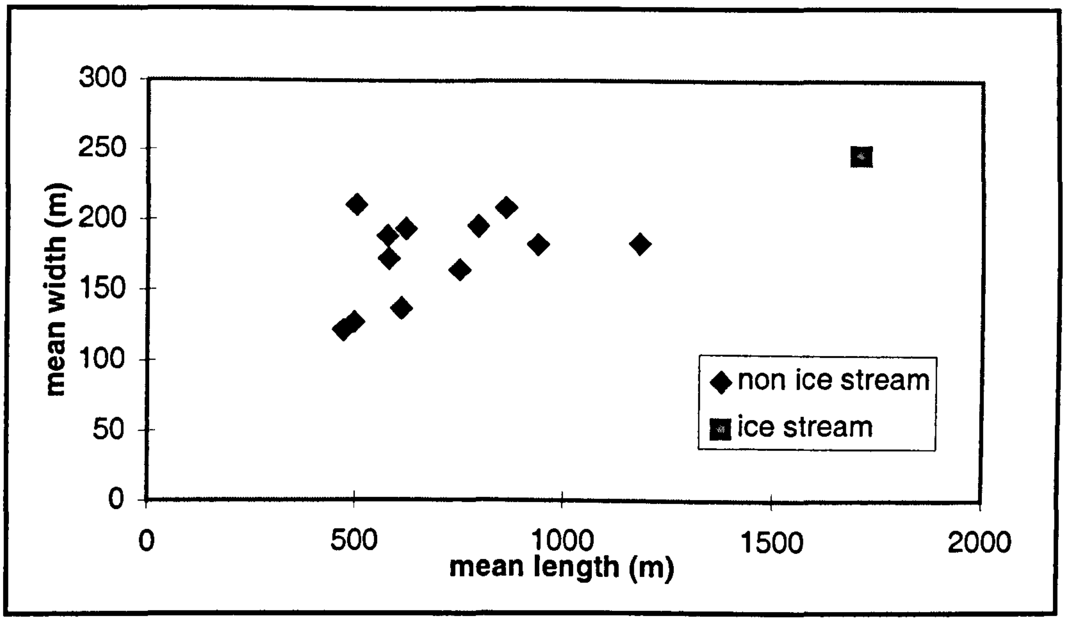

Figure 6.6. Plot of mean length against mean width for all of the flow-sets identified on Storkerson Peninsula.

Even though drumlin fields have frequently been observed to have abrupt lateral margins, the case here is exceptionally marked. The western edge of the ice stream appears extremely abrupt on the Landsat image (Figure 6.3a) and remains so, even at the scale of 1:50,000 aerial photographs. The margin can easily be defined at this scale and there is a change from north-south orientated ice stream lineations to hummocky terrain over a distance of around 100 m. This can be seen on Figure 6.7 which shows a part of the abrupt margin on the Landsat TM imagery.

Although there is no unequivocal method for reconstructing former flow velocities from geomorphology, the weight of evidence clearly points towards fast ice flow. Glacial lineations of the bed are some of the longest and most attenuated observed anywhere, and there are many studies that report correlations between fast flow and high elongation ratios, see Section 4.4.3.

Another indicator of fast ice flow relative to adjacent portions of the ice sheet (a key indicator of ice streams) is the extremely abrupt margin of the ice stream drumlin field (Figure 6.7). To create such a well-defined margin requires a sharp discontinuity in the overlying ice which may have been due to a contrast in basal thermal regime (cold/warm) and/or, abrupt differences in ice velocity, both of which are characteristic of ice stream margins.

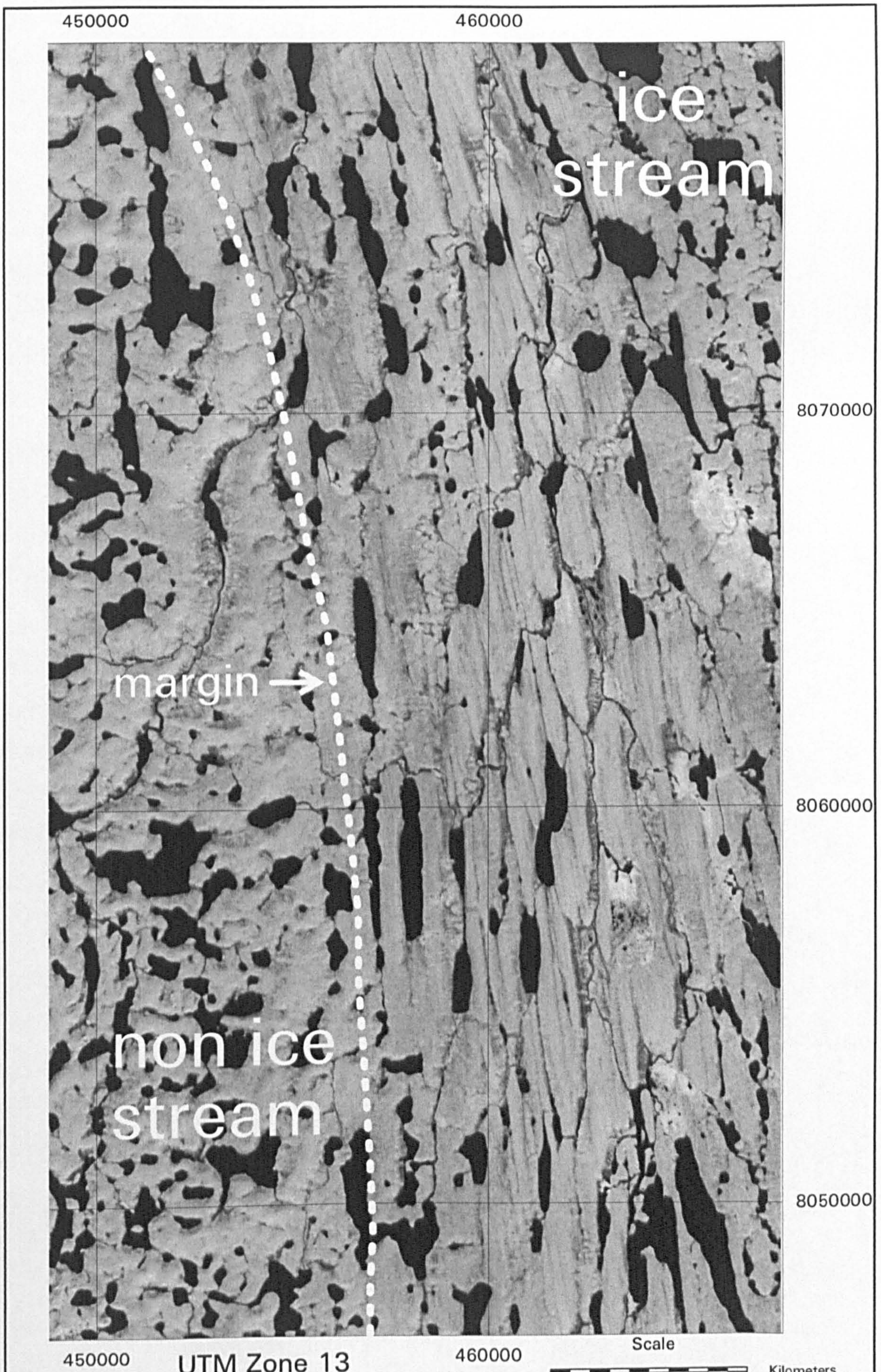

Figure 6.7. Landsat TM image (band 5) showing the abrupt margin of the ice stream drumlins.

Taken together, all these factors strongly suggest that the flow was fast enough to be considered an ice stream. Indeed, these results are the strongest terrestrial evidence that has so far been found for a palaeo-ice stream.

6.4.2. Isochronous or Times-Transgressive Ice Stream Activity?

With substantial evidence to suggest that Flow-2 was laid down by an ice stream, the next task is to ascertain the temporal context of bedform generation, see Section 3.4.3. Were the bedforms laid down isochronously (rubber stamped imprint) or were they generated time-transgressively (smudged imprint)?

It is argued that the high parallel conformity between individual lineaments and the abrupt margin could not have been generated time-transgressively. The standard deviation of lineament orientation with 100 km^2 of grid squares yields a value of only 4.5° along a flow-band. In such cases of exceptionally high parallel conformity, it has been argued that the lineament-generating episode must have been of short duration (Clark, 1999). If this were not the case, we would expect lineaments to display a variety of orientations, arising from fluctuations of flow directions, which fail to match well with their neighbours (low parallel conformity). Rather, it is inferred that the bedforms were created rapidly and that the imprint represents a snapshot view of the bed, which has survived deglaciation with a minimum of modification.

Clark (1999) outlined several characteristics of an isochronous bedform pattern. Criteria included parallel flow patterns, negligible correspondence to local topography, no cross-cutting relationships, high parallel conformity between lineations, gradual variations in lineament morphometry and no landform associations with eskers or moraines. It is argued that the evidence presented in this chapter fulfils all of these criteria and that the M'Clintock Channel Ice Stream signature represents an isochronous imprint.

Evidence left behind by ice retreat, such as eskers, and outwash accumulations, are found superimposed over the bedforms indicating that in places the retreat direction was different to the orientation of the bedforms (Hodgson, 1993). It is as if the ice stream simply turned off, and retreat occurred without erasing or modifying the bed in any significant way.

6.4.3. Variations in Ice Stream Bedform Morphometry.

Having established that the bedform imprint of the M'Clintock Channel Ice Stream represents a snapshot view of the bed prior to shutdown, variations in lineament morphometry within the ice stream were investigated. Across-stream analyses revealed no clear pattern for the three transects analysed. Two transects revealed an increase in lineament length away from the margin, but the third showed the opposite. It is safest not to draw conclusions here as the transects were too limited in length, yielding low sample sizes. Unfortunately, it was not possible to use longer transects because of the modifying influence of the Flow-3 landforms on eastern Storkerson Peninsula (see Figures 6.2 and 6.3).

For down-ice variations in lineament morphometry, it was possible to use a much larger sample size (n=722). Parameters of lineament morphometry were binned into 10 km samples running for 100 km along an approximate flow-band within the ice stream, and positioned close to the terminus and away from the modifying influence of Flow-3. Downstream change in lineament length is shown in Figure 6.8, where it can be seen that there is a slight decrease over a distance 90 km. Similarly, elongation ratio was found to exhibit a decrease in the downstream direction towards the terminus, see Figure 6.9.

Of greatest note was the downstream change in lineament packing (percentage lineament area per unit area). This is shown in Figure 6.10 where it can be seen that the percentage unit area of the bed occupied by lineaments increases from around 15% up to almost 40% along the 90 km flow-band. The lineaments at the down-ice end, and close to the inferred terminus of the ice stream, are much more tightly packed against each other than at the up-ice end. This is further evidenced by the downstream change in lineament density (number of bedforms per km^2), which increases from around 0.35 up to a maximum of almost 1, see Figure 6.11. This demonstrates that the lineaments are fitting closer together, rather than just increasing their surface area. This analysis accords with the visual observations of the area which appear to show individual lineaments merged or nested together rather than as separate entities (see Figure 6.7).

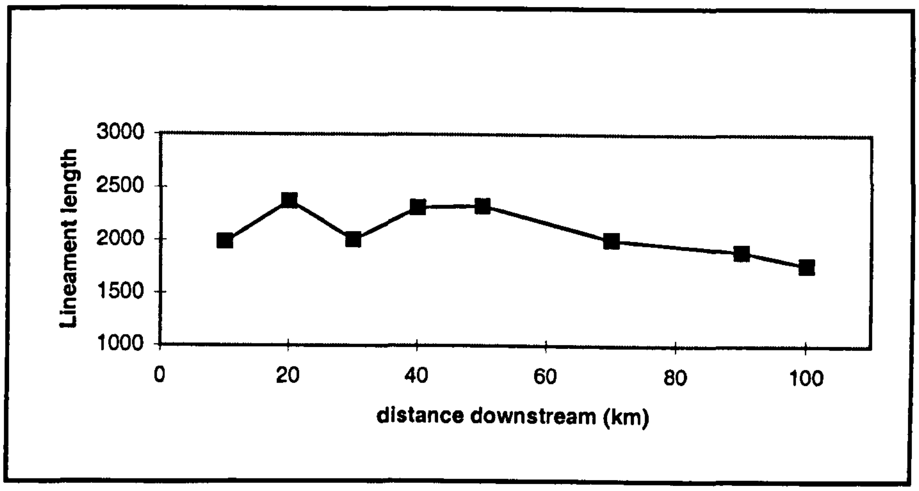

Figure 6.8. Downstream variation in lineament length.

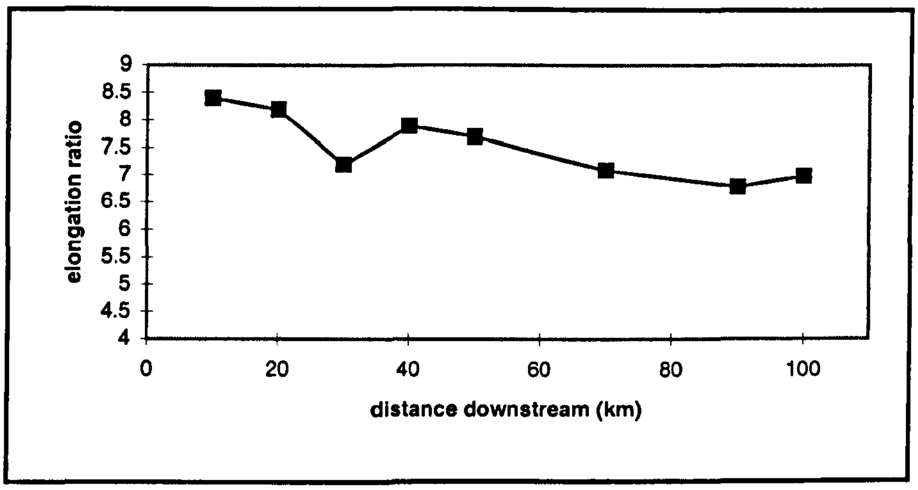

Figure 6.9. Downstream variation in elongation ratio.

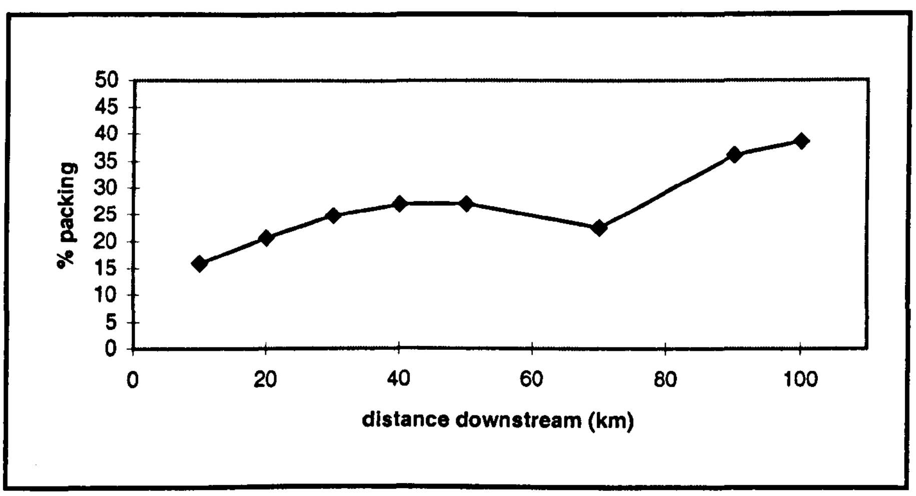

Figure 6.10. Downstream variation in lineament packing (% unit area occupied by lineaments).

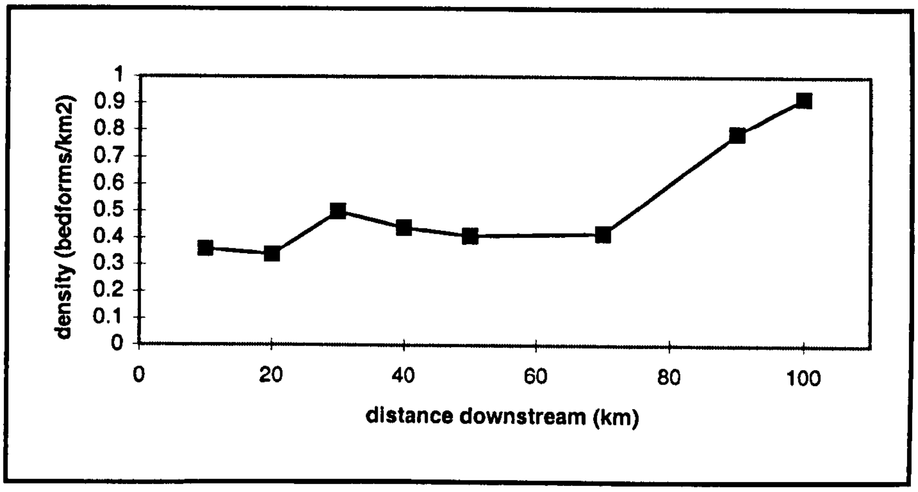

Figure 6.11. Downstream variation in lineament density (number of bedforms per unit area.

6.4.4 Ice Stream Marginal Moraines.

At or close to the western edge of the main drumlin field are a number of elongate ridges that at first appear similar to the ice stream bedforms. Figure 6.12 shows a TM image and an annotation of one of these ridges from the southern end of Storkerson Peninsula. From closer inspection of the imagery and aerial photographs, these ridges differ from the ice stream lineaments in a number of respects. Although suffering dissection from subsequent fluvial activity, the ridges are longer than any of the mega-lineations within the ice stream. The southern-most ridge, shown in Figure 6.12, is traceable for over 23 km.

Unlike the adjacent mega-lineations, which typically taper in a downstream direction, the ridges are 500-800 m wide and this width is maintained along their full length. Furthermore, although the ridges are mostly straight, they do display some sinuosity and this too is in contrast to the ice stream mega-lineations. Finally, the relief of these ridges, which stand between 10 and 40 m above the surrounding terrain, tend to be higher than the adjacent drumlins and megalineations, see Figure 6.13. For these reasons, it is inferred that the ridges are not merely heel-to-toe mega-lineations at the edge of the bedform suite but are distinctive ridges formed by another process.

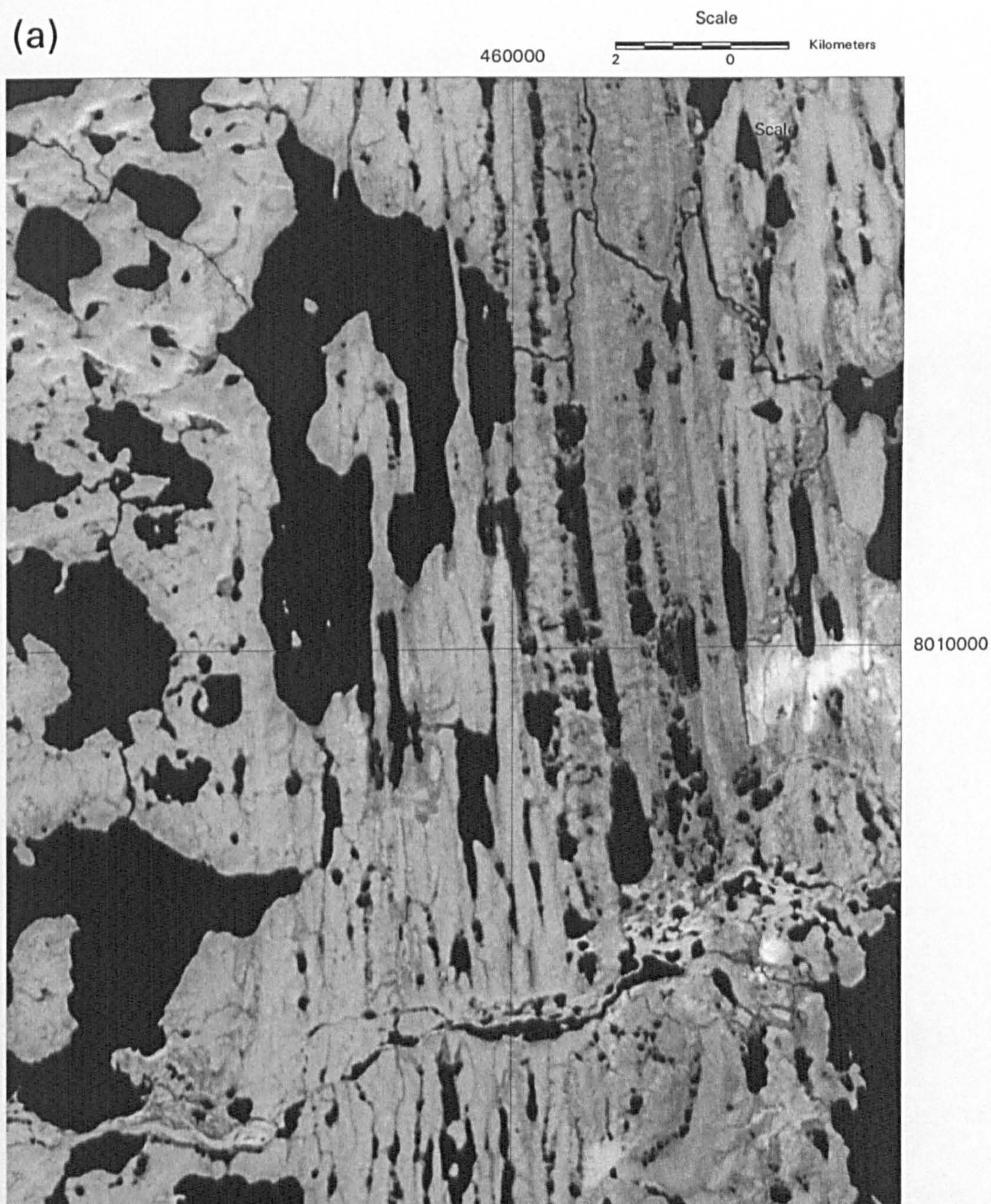

Figure 6.12. TM image (band 5) of ice stream marginal moraine on southern Storkerson Peninsula (a) and annotation (b) overleaf.

(b)

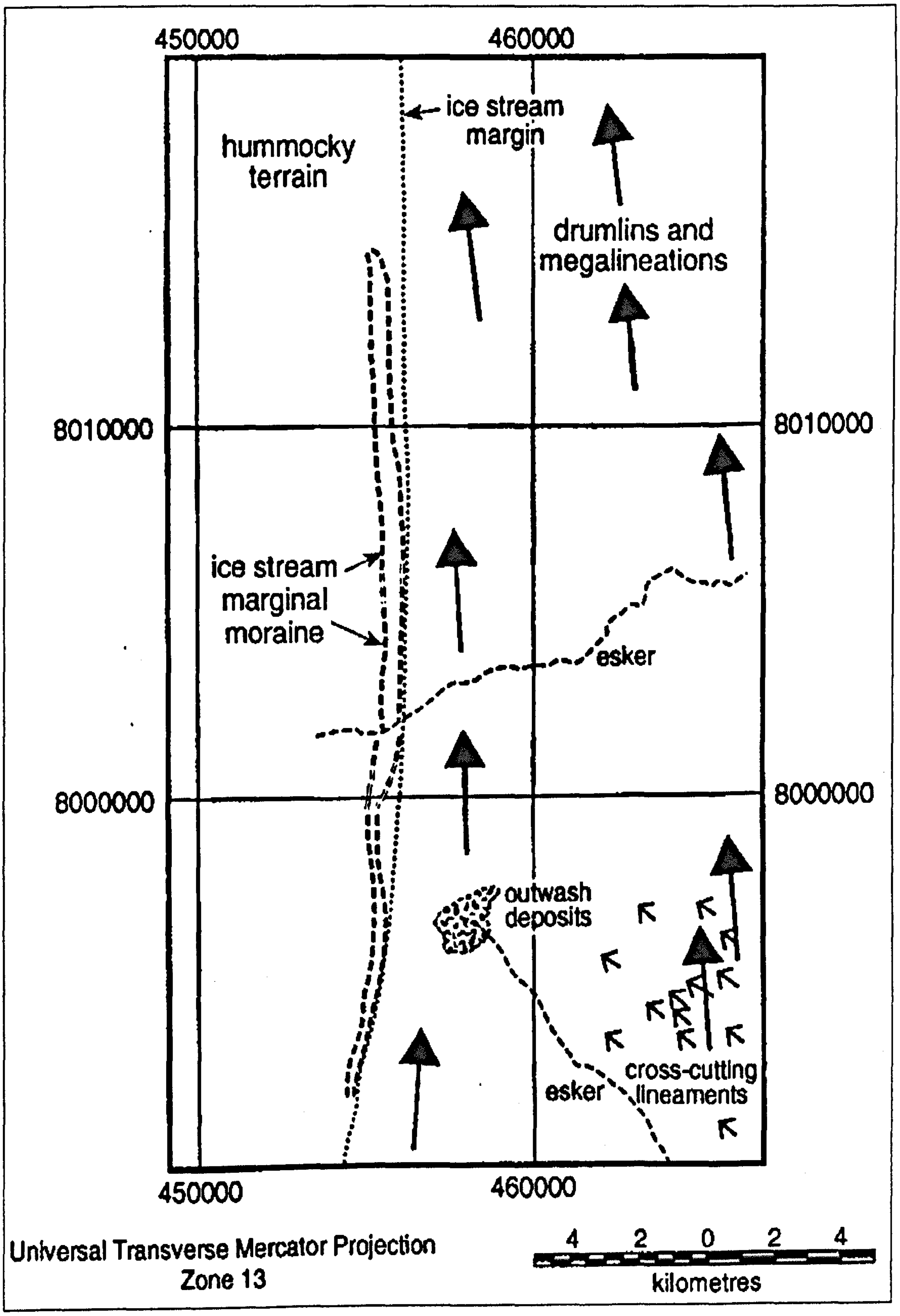

Figure 6.12b.

The geological map of the area does not depict any change in geology close to the margin and the notion that these features are geological structures can be dismissed. Rather, the area is composed entirely of drift, which Hodgson (1994) estimated may be as thick as 50 m on western Storkerson Peninsula. This negates the idea that the ridges represented a topographical obstacle to the ice stream's westward migration, because unconsolidated drift would have provided negligible resistance. This is further emphasised by the fact that the southern-most ridge has been eroded by meltwater activity and an esker can be seen cutting through it on Figure 6.12.

It is suggested that these ridges represent ice stream marginal moraines, which are subglacial accumulations of sediment formed in the shear zone between fast moving stream-ice and adjacent slow moving sheet-ice. Similar landforms have been observed on Prince of Wales Island in a similar context and have been described as lateral shear moraines marking the boundary of warm and cold-based ice (Dyke and Morris, 1988). Kleman and Borgström (1994) also found evidence of a lateral shear moraine marking the lateral-sliding boundary between cold and warm-based ice at Arvestuottar in northern Sweden.

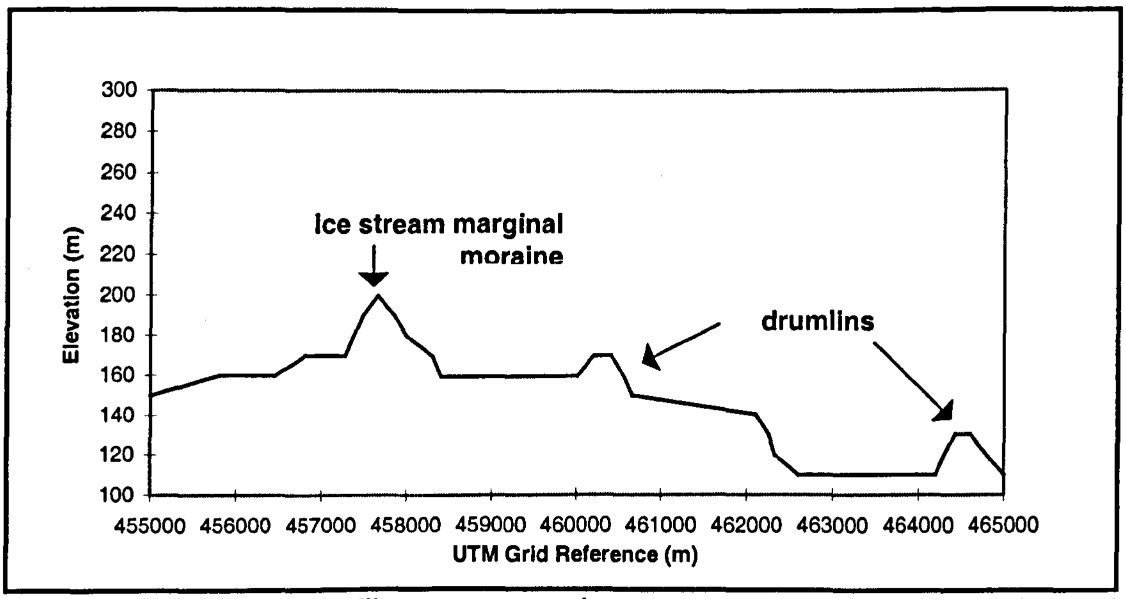

Figure 6.13. Surface profile across an ice stream marginal moraine (vertical exaggeration x25).

On northern Storkerson Peninsula, the ice stream marginal moraines (and their fragments) display a lateral offset of around 2 km from the last inferred ice stream margin. Figure 6.14 shows the position of these moraines in comparison to the ice stream margin.

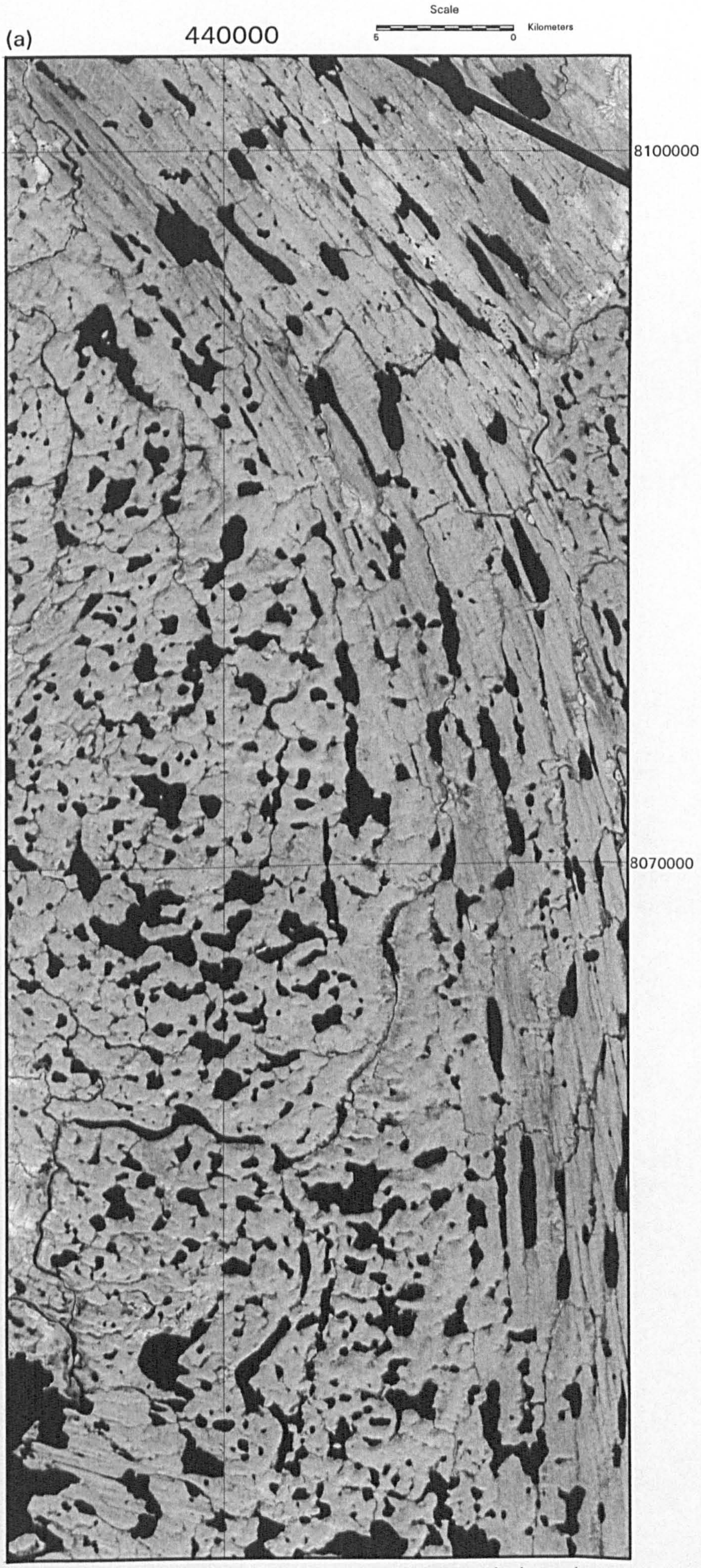

Figure 6.14. TM image (band 5) of relict ice stream marginal moraines on northern Storkerson Penisula (a) and annotation in (b), overleaf.

(b)

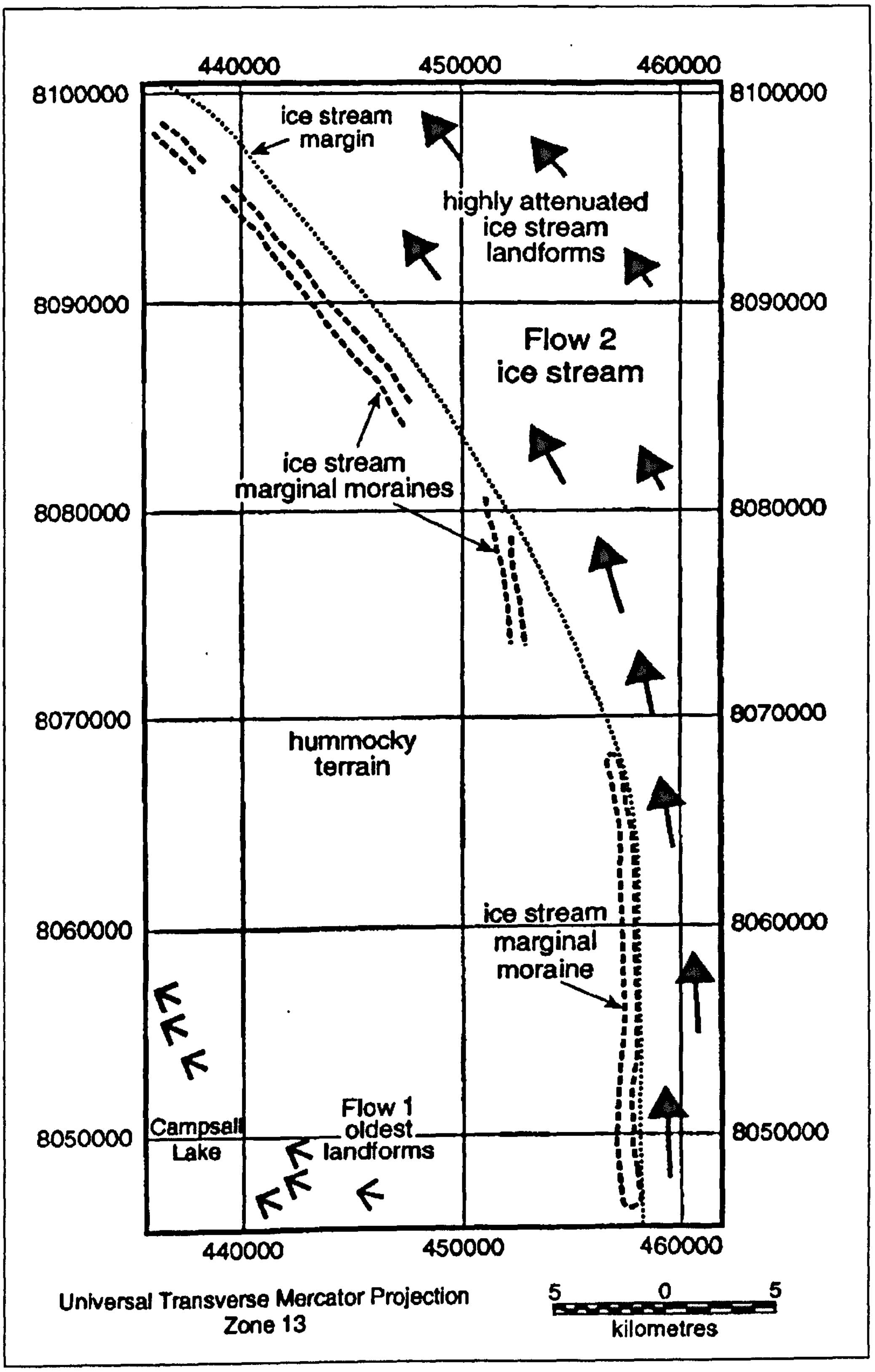

Figure 6.14b.

The southern-most of these three ridges is very similar to the ice stream marginal moraine further upstream and shown in Figure 6.12. However, further north (downstream) the ridges are far more fragmented, and their continuity can only be fully appreciated at a large scale on the satellite imagery. In close up, the ridges are quite broad and composed of several peaks reaching elevations of up to 150 m a.s.l.

Of most importance is the lateral offset of these features and it is suggested that either the actual ice stream margin had offsets of this nature, or more likely, the margin experienced minor migrations and the moraines are a record of these different positions. The apparent ruggedness of the features may be a cumulative result of a shorter period of generation or longer period of modification.

6.4.5. Geology and Topography.

6.4.5.1. Sediment Thicknesses.

Glacial drift in the area comprises fine-grained tills which are fairly homogenous, reflecting the Palaeozoic carbonate rocks of eastern Victoria Island (Figure 6.1), with about a 4% erratic content of shield rocks that derive from at least 400 km upstream (Hodgson, 1994).

There is a noticeable contrast in sediment thickness between the ice stream and non-ice stream areas. To the west, the hummocky terrain is uniformly thick, estimated to be around 50 m (Hodgson, 1994). In contrast, within the ice stream, the till is much thinner, with bedrock exposed in places, particularly at the northern end of Storkerson Peninsula where individual drumlins can be seen resting on bedrock. The observed contrast in sediment thickness is also apparent in the surface topography and Figure 6.15 shows a clear step between the ice stream and hummocky topography at the ice stream margin.

Immediately outside the ice stream the deposits and landforms of Flow-1 are characterised by hummocky terrain whose surface has been drumlinised in places. These drumlin patterns are truncated by the ice stream, demonstrating that the Flow-1 landforms and 50 metres of sediment thickness must predate the activity of the ice stream. It is reasonable to assume that this till thickness extended at least some distance across Storkerson Peninsula, and that the step in topography and sediment thickness must therefore be a direct result of erosion by the ice stream.

Without field measurements of drift thickness, it is hard to quantify the amount of erosion by the ice stream. In order to make a preliminary assessment of sediment erosion by the ice stream, two techniques are described below.

For areas where bedrock is exposed between drumlins and regional topography is gentle, drumlin area and relief can be used to estimate their approximate sediment volume. Sampling along a flow-band, it is possible to redistribute drumlin sediments to make an homogeneous till thickness. In the downstream portions of the ice stream, where drumlins are small and bedrock is exposed, a value of the order of 5 m of sediment is calculated. Given an initial thickness of 50 m, this indicates 45 m of sediment lowering.

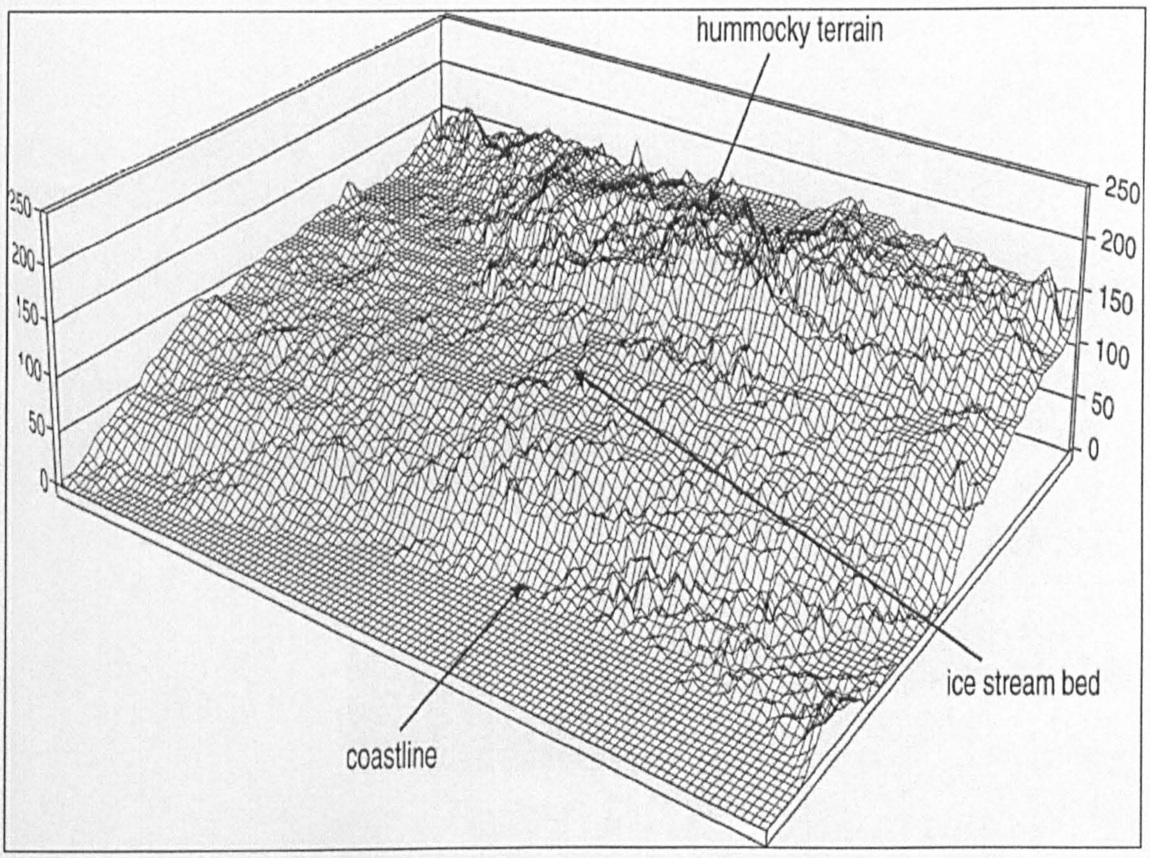

Figure 6.15. Digital Elevation Model of a part of Storkerson Peninsula (looking south-west) showing the drop in elevation between the ice stream bed and hummocky terrain to the west.

Further upstream, bedrock is not exposed between drumlins and the till thickness beneath and between drumlins is not known. However, it is possible to use the elevation drop at the margin to estimate the surface lowering. Preliminary estimates suggest that there is an approximate drop in elevation of about 35 m within the ice stream (see Figure 6.13). These approximations would appear to suggest that a maximum surface lowering (erosion) of 45 m declined to 35 m at 60 km upstream and presumably continued to do so further upstream.

6.4.5.2. Soft Bed to Hard Bed Fraction.

Concomitant with a decrease in till thickness in the downstream direction, a higher proportion of bedrock is progressively exposed. For a flow-band (varying in width between 12 and 25 km to allow for convergence), extending for 140 km downstream towards the terminus, data describing exposed bedrock were compiled at 10 km intervals. The Geological Survey of Canada's surficial geology map (1:250,000) was used to provide this information (Map 1817A; Hodgson, 1993). Results are presented in Figure 6.16 where it can be seen that for at least half the flow-band the bed is entirely soft and thereafter the proportion of hard bed varies between 5 and 25%, reaching a maximum of 40 % close to the inferred terminus.

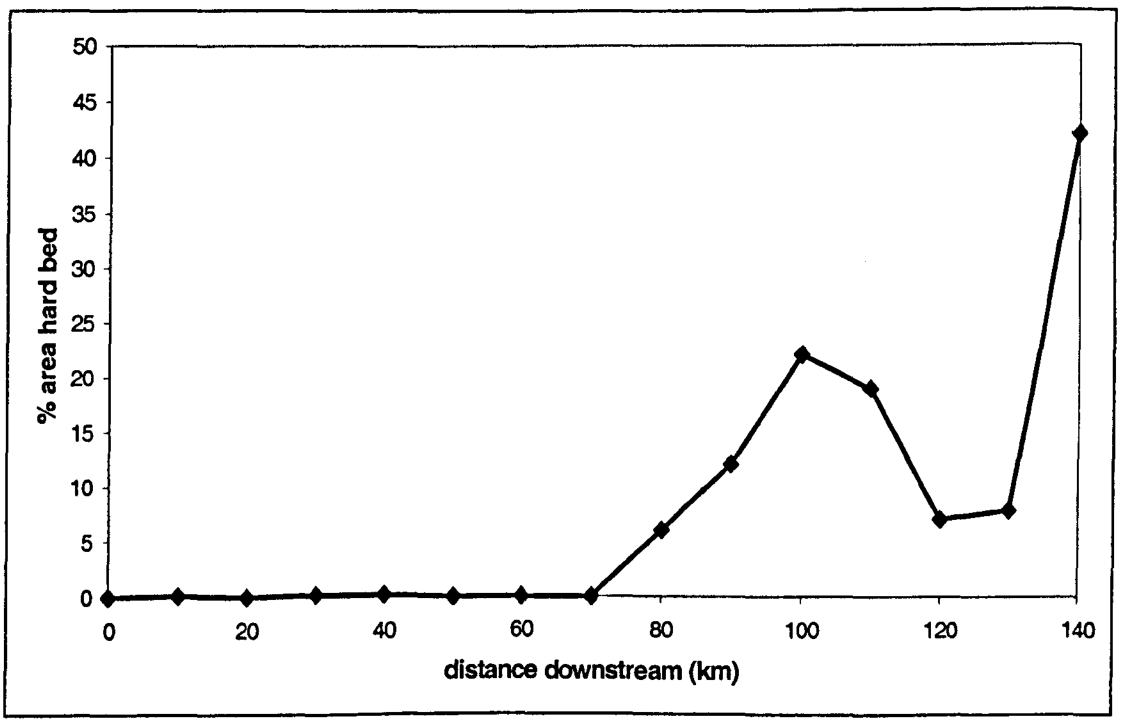

Figure 6.16. Downstream variation in the percentage of the bed occupied by hard bedrock.

6.4.5.3. Regional Topography.

Storkerson Peninsula is low lying, rising to only 200-300 m in its 80 km width, and exhibits a gentle and smooth slope upwards from its eastern shoreline. Topographic profiles computed from DEM's across Storkerson Peninsula did not reveal any continuous major topographic obstacles, such as a bedrock scarp for example, which may have contained the ice stream. Figure 6.17 illustrates the relationship of the margin to local topography.

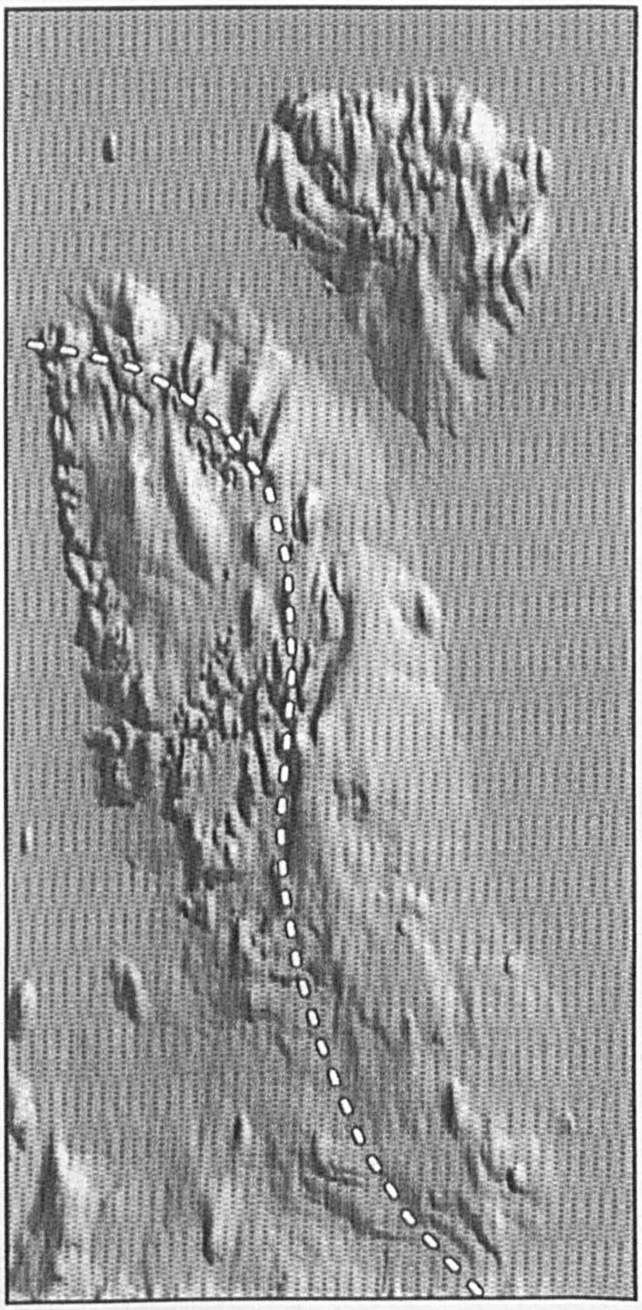

Figure 6.17. Digital Elevation Model of Storkerson Peninsula and Stefansson Island showing the location of the ice stream margin in relation to regional topography. Note the smoother topography within the ice stream, compared to the hummocky terrain to the west.

The margin rises in elevation from south to north and crosses topographic obstacles with no deflection. In places the terrain even slopes away from the margin and it is clear that there is no topographic control on its position.

6.4.6. Reconstruction of Ice Stream Extent.

Using Landsat MSS and SAR images from adjacent islands, it was possible to determine the overall extent of the ice stream. All areas adjacent to M'Clintock Channel (see Figure 6.1) were examined for lineament patterns that match with the distinctive signature of the ice stream bed, described above. South of 72° latitude (the limit of Hodgson's mapping) the ice stream signature was strong, permitting it to be traced upstream for a further 250 km stretching to southern Victoria Island. On King William Island, which occupies the southern end of the Channel (Figure 6.1), an extensive pattern (>10,000 km^2) of lineaments were mapped which also match the ice stream orientation, and which indicate flow into the central axis of the channel.

Along the eastern shore of the M'Clintock Channel, extending 50 km inland, on the Arrowsmith Plain (Prince of Wales Island), are a suite of mega-lineations displaying a smooth and straight flow pattern that parallels the main axis of the channel. These features were mapped by Dyke *et al.*, (1992) who reported huge mega-flutings (i.e. mega-lineations) with lengths approaching 20 km, widths between 1 and 3 km and heights of 10 m. Digital imagery (SAR) of these mega-lineations was obtained.

Although the SAR imagery appeared grainy, it was possible to measure the dimensions of a number (n=83) of the larger Arrowsmith Plains mega-lineations. Unfortunately, no detailed photomaps of the area are available. Table 6.2 is a comparison of the bedform dimensions on Prince of Wales Island and Victoria Island.

Table 6.2. Comparison of bedform dimensions on Storkerson Peninsula and western Prince of Wales Island.

Area	n	Mean Orientation. degrees	Mean length (max.) metres	Mean width (max.) metres	mean l:w ratio (max.)
Victoria Island	486	334 (NNW)	1709 (7,950)	247 (750)	6.9 (27)
Prince of Wales Island	83	335 (NNW)	2958 (11,775)	434 (1475)	6.8 (20.3)

It can be seen from Table 6.2 that the mean lengths of the Arrowsmith Plains mega-lineations are higher but this can be explained by the fact that it was difficult to detect smaller bedforms on the SAR imagery without the aid of aerial photography. However, it is clear that the Arrowsmith Plains mega-lineations are large enough to have been formed by fast ice flow. Of greater importance is the remarkable similarity in elongation ratio between the megalineations on Prince of Wales Island (6.8:1) and those on Victoria Island (6.9:1).

To demonstrate the similarity between the two bedform populations, Figure 6.18 shows the histogram frequencies of length (a) and elongation ratios (b) for each flow-set.

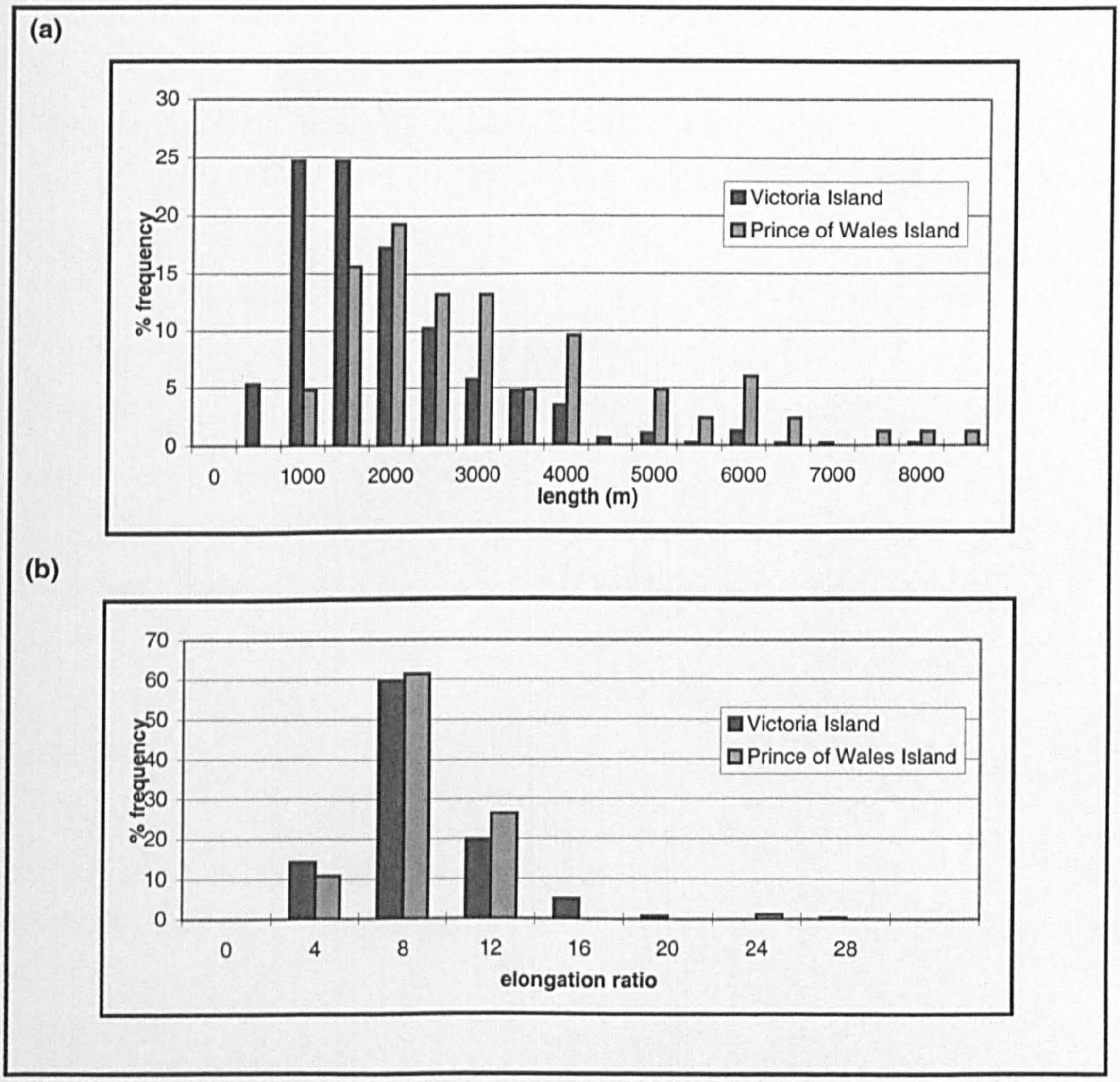

Figure 6.18. Comparison of lineament lengths (a) and elongation ratios (b) between bedform populations on Victoria Island and Prince of Wales Island.

It can be seen that despite the variable sample sizes, the bedform populations appear quite similar. Moreover, if elongation ratio is taken as a crude proxy for ice velocity or cumulative strain, then this suggests that both sets of mega-lineations could belong to the same population and may have been formed by the same flow event. In addition, the geographical proximity of these two sets of bedforms and the intervening channel strongly suggests that it was in fact the same ice flow event and the almost identical orientations of each flow-set provides further support for this hypothesis.

To the east, the Arrowsmith Plain mega-lineations have been cross-cut and truncated by a drumlin field with slightly different orientation and of a later age, the Crooked Lake drumlin field (Dyke *et al.*, 1992). Four radiocarbon dates associated with marine deposits that overlie the Arrowsmith mega-lineations indicate that the ice flow episode must have occurred prior to 10,000 yr BP. In the reconstructed history of glacial events, Dyke *et al.*, (1992) inferred that the Arrowsmith Plain mega-lineations originally extended across the whole of Prince of Wales Island, recording south to north flow from a mainland ice divide. They presumed that most of them were subsequently eroded (by the Crooked Lake ice flow event) leaving the remnants that we observe today adjacent to M'Clintock Channel.

In light of Hodgson's (1994) work on Victoria Island and the results presented in this chapter, this hypothesis appears unlikely and it is suggested that the limited eastern extent of the mega-lineations is a function of the extent of the flow that produced them rather than an artefact of differential erosion/preservation. In favour of this argument is the fact that because mega-lineations are extremely large features, they could reasonably be expected to be visible beneath the Crooked Lake drumlin field (and elsewhere on Prince of Wales Island) if they existed there. Many such relationships between mega-lineations and superimposed drumlins have been observed elsewhere (Clark, 1993). Furthermore, the orientation, extent and context of the Arrowsmith Plains mega-lineations are entirely consistent with an ice stream in M'Clintock Channel which extended for a short distance onto Prince of Wales Island in the same manner as it did on Storkerson Peninsula. The extra dating constraints of Dyke *et al* (1992) imply that if the ice stream produced the Arrowsmith Plain mega-lineations, its duration of operation must have been prior to 10,000 yr BP. Together with Hodgson's (1994) dates (Section 6.2), the ice stream operation is now bracketed to between 10,400 and 10,000 yr BP.

The life-cylcle of the ice stream (400 years) outlined above is heavily reliant on the dating constraints provided by Hodgson (1994) and so it is necessary to briefly outline the details of his radiocarbon chronology. Hodgson (1994) utilised 14 radiocarbon dates from northern Victoria Island and Stefansson Island (Table 1, Hodgson, 1994) together with 23 "representative" radiocarbon ages for the deglaciation of western Parry Channel, excluding Victoria Island and Stefansson Island (Table 2, Hodgson, 1994).

Flow-1 drumlins and hummocky till (see Section 6.2) were overlain with eskers and kames which drained to undated (possibly marine) deltas now at 120 to 140 m above sea level (a.s.l). Hodgson presumed that these features were older than shells and whalebones deposited on the same coast in a sea around 80 to 90 m a.s.l. Three sites give ages of 10,000 ± 110, 9,935 ± 190 and 9,820 ± 100 yr BP. Therefore, Flow-1 is inferred to be older than 10,000 ± 110 yr BP. In terms of the geomorphology, Flow-2 (ice stream) bedforms truncate Flow-1 drumlins and therefore, must be younger.

Following Flow-1, marine transgression permitted the migration of molluscs onto northern hummocky till on Storkerson Peninsula. Fragments of these shells supply one date of 10,350 ± 80 yr BP. Although these shells could have been transported by advancing ice, Hodgson suggested that their location lies down-ice from extensive uplands and this scenario is unlikely. It does, however, provide only a tentative maximum age for Flow-2 (ice stream). Indeed, Flow-2 cannot be directly dated but is bracketed between the older Flow-1 and the younger Flow-3 which cross-cuts the ice stream (see Section 6.2). Therefore, the minimum age of the ice stream is taken from shell deposits which overlie the ice stream. Two dates indicate 9,640 ± 110 yr BP and 9,560 ± 100 yr BP. The second date is taken from a shell within a ridge thought to have been thrusted onshore by Flow-3. Thus, Hodgson brackets the lifecycle of the ice stream to around 800 years but speculates that it could have been much shorter.

The reliability of Hodgson's dates and those of Dyke *et al.* (1992) should be questioned with respect to the timing of ice stream operation. Clearly, the suggestion in this thesis that the mega-lineations on Prince of Wales Island were deposited by the ice stream deserves further attention, because this brackets the ice stream operation to a maximum of 400 years. Furthermore, this is a relatively small window given the error estimates of the radiocarbon dates, which typically amount to between ± 80 and 190 years. However, these discrepancies should not detract from the approximate dating constraints and the main evidence that a relatively short-lived (i.e. hundreds of years) ice stream existed within this complex portion of the Laurentide Ice Sheet.

If the Arrowsmith Plain mega-lineations were formed by the M'Clintock Channel Ice Stream, as is suggested, then is it possible to find the eastern margin of the ice stream? A search for ice stream marginal moraines on imagery and in the work of Dyke *et al.* (1992) revealed nothing that matches with the pattern of megalineations. Although the mega-lineations can be detected for 40 km inland and appear to stop at a distinct line, this cannot be taken to record the ice stream margin. This is because this line, and a lateral shear moraine along part of it, was produced by the later Crooked Lake drumlin field which overprinted the Arrowsmith mega-lineations. It is impossible to discern the extent of the overlap, and therefore it is not possible to define the eastern margin with great confidence. It is safest to infer that the margin lay approximately 40 km inland from the present day coast.

A key question here, is whether the four spatially separate lineament patterns on NE and SE Victoria Island, King William Island and Prince of Wales Island, all belong to the same flow event. Given that the intervening area is now at sea, it is not possible to trace bedforms between these areas. For NE Victoria Island and Prince of Wales Island, the mean elongation ratios are similar and it is assumed that they come from the same bedform population and flow event (discussed above).

The morphometry of the lineaments on south-east Victoria Island, and King William Island have not been quantitatively examined, but their visual appearance in terms of size, shape and orientation match well with those of Storkerson Peninsula and Arrowsmith Plains. It is also important to note that the characteristics of these lineaments are significantly different to surrounding lineaments on the islands. Furthermore, their topographic association with the M'Clintock trough is significant and it is concluded that they all belong to the same flow event. Good dating control only exists for the Storkerson Peninsula lineaments, so the possibility of other lineament patterns being of different age and possibly recording different episodes of ice streaming cannot be excluded. This is unlikely however, because of the excellent match in patterns and morphometry.

Taken together, these arguments suggest that the ice stream filled the whole of M'Clintock Channel and the evidence suggests that it infringed upon a number of neighbouring islands, not just Storkerson Peninsula. Figure 6.19. shows the reconstructed flow patterns of the ice stream.

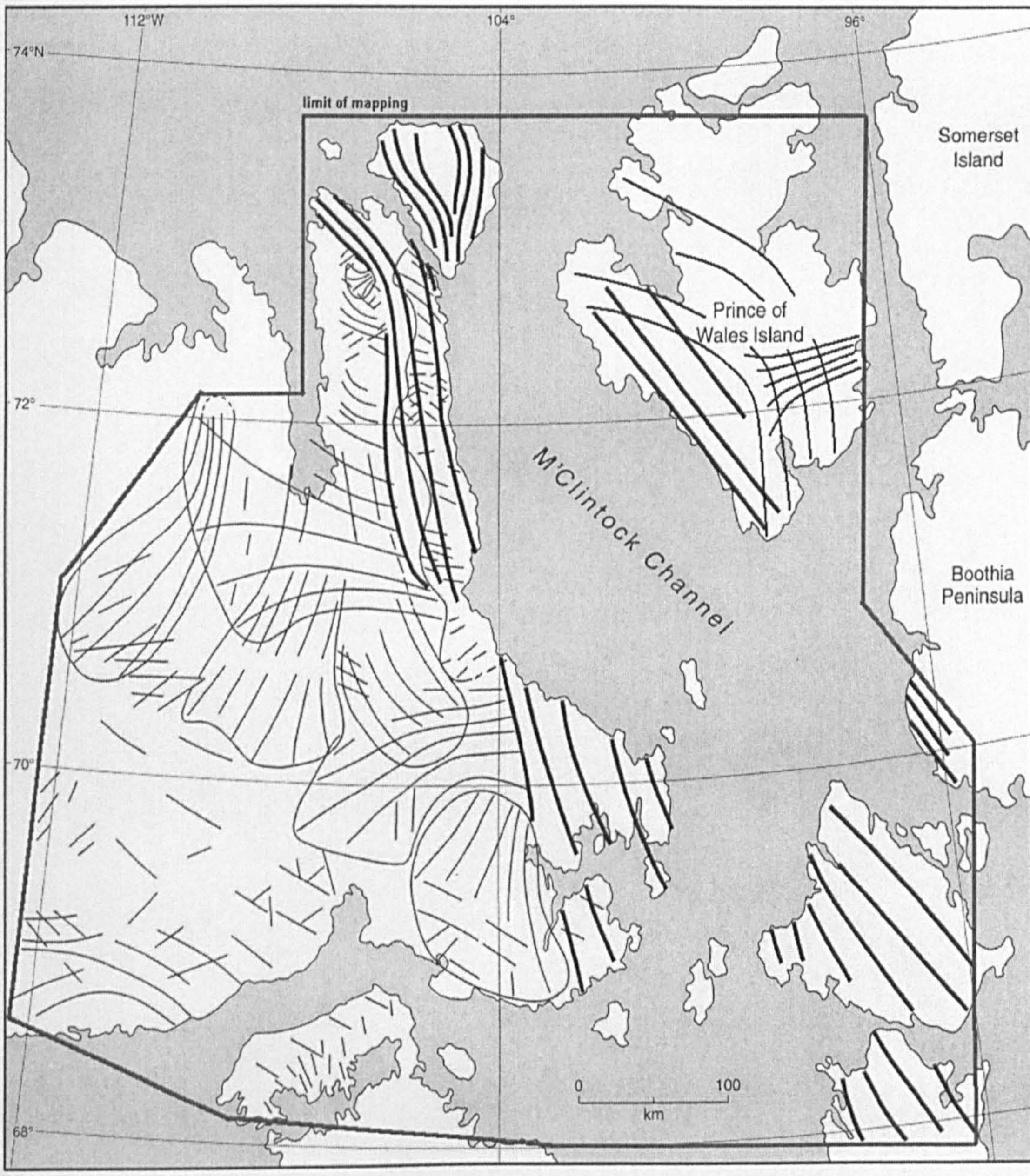

Figure 6.19. Reconstructed flow patterns of the M'Clintock Channel Ice Stream (bold). Flow-sets on Prince of Wales Island also mapped by Dyke *et al.* (1992).

The ice stream is reconstructed as being at least 720 km in length and between 140 and 330 km wide, displaying a highly convergent flow pattern. Figure 6.20 shows the estimated cross-sectional area of the ice stream from two transects across M'Clintock Channel (see Figure 6.1). This demonstrates the channelling effect of the trough and the downstream convergence of flow. The discovery of this convergent flow pattern adds a fifth diagnostic criteria to those already documented by Hodgson (1994), see Section 6.2.2.

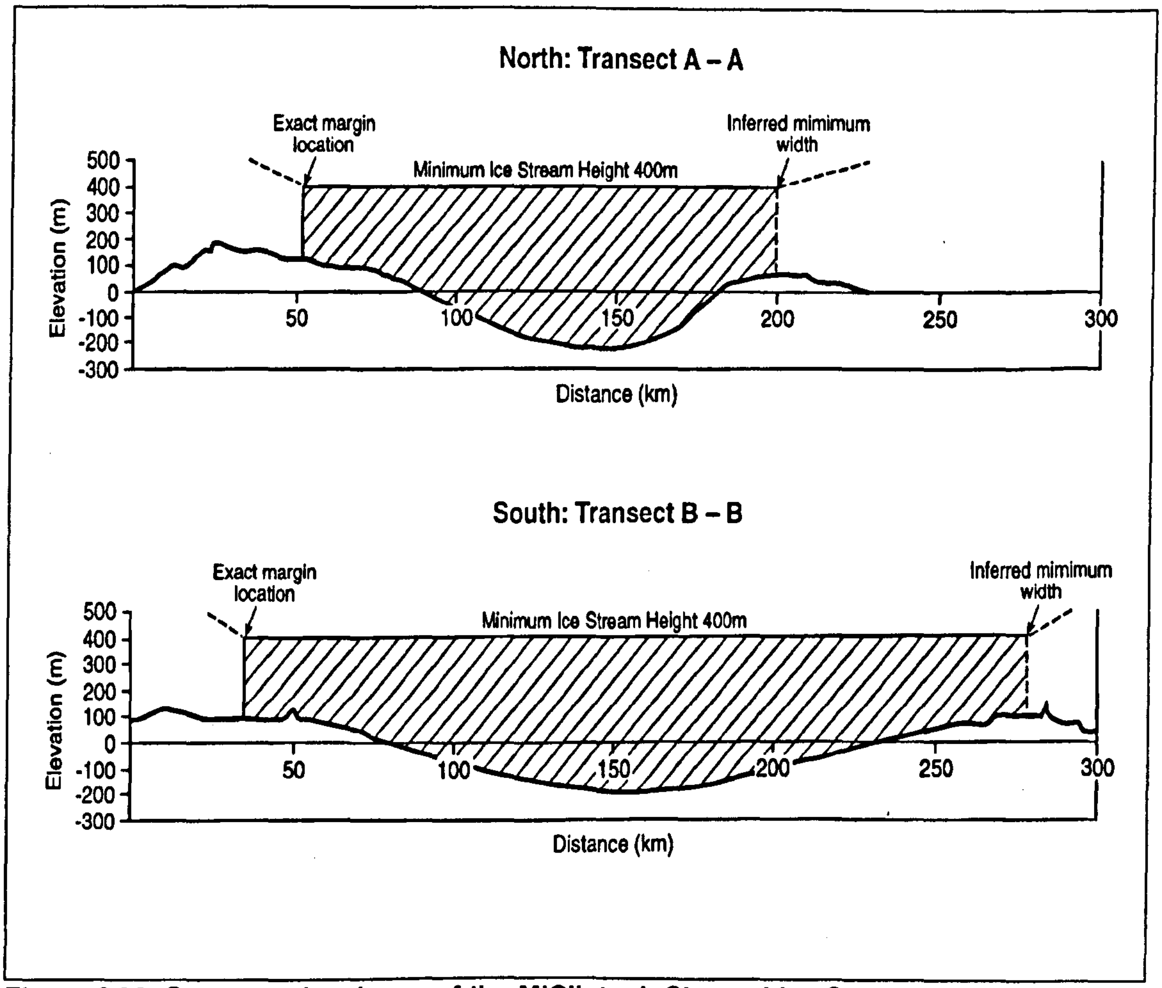

Figure 6.20. Cross-sectional area of the M'Clintock Channel Ice Stream upstream (B-B) and downstream (A-A), see Figure 6.1 for transect locations.

It is emphasised that the ice stream is not simply demarcated by selecting features drawn from a large population of lineaments, but is a discrete unit bounded by margins that have a strong impression in the bedform record. It has not been possible to define the upstream end (onset zone) of the ice stream as this was likely a gradational margin and left no clear signature. The ice stream terminus is presumed to be in the vicinity of Viscount Melville Sound where depths of over 500 m are reached.

6.5. Discussion.

In Section 6.2.2 several questions were posed concerning the behaviour of the M'Clintock Channel Ice Stream. Having established the nature of its bed in terms of geomorphology, soft to hard-bed fraction, lithology, roughness and slope, and reconstructed its spatial extent, the questions relating to its functioning, basal processes and significance will be addressed.

6.5.1. Controls on Ice Stream Location and Margin Position.

It is clear that the main control on the location of the M'Clintock Channel Ice Stream was that of topography. This is because the ice stream was centred within a major trough, see Figure 6.20. Given that the Laurentide Ice Sheet margin never extended much beyond Victoria Island (Dyke, 1987) it can be assumed that ice was relatively thin over this area (<800 m?) and that a trough of considerable depth (500 m) would have easily influenced flow patterns, preferentially focusing ice flow along its axis and capturing ice from an extensive catchment area.

The bedrock geology of the area (Figure 6.1) falls into two contrasting types in terms of basal shear strength; 'soft-bedded' carbonates on which the ice stream lay and 'hard-bedded' crystalline rocks on the mainland, upstream of the ice stream. It has been argued that soft bedded geology can promote lubrication at the base of an ice sheet by a metres thick layer of deforming sediment, thus facilitating fast flow velocities (see Section 2.3.2). The location of the ice stream conforms with this view. It is suggested that the easily-eroded soft-bedrock supplied thick accumulations of drift during pre-ice stream margin fluctuations and that it was these deposits rather than direct erosion of bedrock during ice streaming that helped promote fast flow.

It is concluded that the major trough of the M'Clintock Channel controlled the ice stream location and the occurrence of pre-existing unconsolidated drift encouraged fast flow velocities. Although it is not possible to define the onset zone of the ice stream with any certainty, the flow convergence pattern suggests it was near the mainland coastline (see Figure 6.1). It is interesting to note that this boundary coincides with the change from soft to hard-bedded terrain (compare Figures 6.1 and 6.19). In contrast, there are no changes in geology, or topographic obstacles, that fix the exact position of the ice stream margin and it is concluded that glaciological regulation, rather than the nature of the bed controlled the lateral extent of the ice stream. This would also suggest that the location of the ice stream margins were temporally variable and this notion is supported by the identification of relict ice stream marginal moraines on northern Storkerson Peninsula (Section 6.4.4), which demonstrate that switches in the margin position probably occurred there.

6.5.2. Controls on Ice Stream Shut-Down.

Given that the landforms and sediments represent the nature of the bed immediately prior to ice stream shut-down (i.e. an isochronous record), then what do they tell us about the mechanism that controlled shut-down? Analysis of the percentage of hard bed beneath the ice stream (Section 6.4.5.2) shows that erosion of pre-existing sediment left a sediment wedge, thinnest at the ice stream terminus and thickening in the upstream direction. In addition, ice stream bedforms display a notable downstream change in closeness of packing and density, see Figures 6.10 and 6.11. It is suggested that these characteristics are not merely how the bed happened to be at the time the ice stream shut-down, but are the geomorphological product of the mechanism of shut-down.

It is postulated that erosion beneath the ice stream progressively thinned the underlying sediment, to such an extent that eventually bedrock was exposed in places producing zones of much higher friction, (i.e. sticky spots; see Section 2.3.3.1). The change in basal drag from that supplied by a layer of unconsolidated drift to that of bedrock was enough to retard fast ice flow. Therefore, it is suggested that the principal control on ice stream cessation was by sediment exhaustion leading to frictional shut-down. It would seem that a necessary condition for the activity of this ice stream was a pre-existing thickness of unconsolidated sediment, and that its life-cycle was dictated by initial sediment thickness.

The dating constraints of Hodgson (1994) and Dyke *et al.* (1992) bracketed the life-span of the ice stream to a maximum of 400 years but Hodgson suggested that it probably operated for only a few hundred years. If this is the case, then 50 m of sediment was required to permit ice stream activity for a minimum of 200 yrs and a maximum of 400 years.

The variation in till thickness and hard-to-soft bed fraction can be used to infer the actual seed point for ice stream shutdown. In the ice stream's final 70 km the till becomes so thin that bedrock is exposed (see Figure 6.16) with the highest hard-bed ratio (45%) occurring close to the terminus. As this was the zone where the friction of the bedrock was first felt by the down-cutting ice stream it is inferred that this is where flow retardation started. This propagated throughout the rest of the ice stream, resulting in complete de-activation of fast flow.

It is somewhat surprising that the ice stream ran out of sediment in its lower reaches first. Taking till continuity into account, we would expect sediment exhaustion to occur first in the upstream portions of the ice stream. Clearly, this is not what is observed. Rather, it is evident that debris from the upstream catchment was not supplied in sufficient quantities to maintain steady state. This is indicated by the small incorporation of (Shield) erratic material in till samples (~4 %) with the majority being comprised of the local carbonate lithology. Clearly, some other mechanism was operating which limited the supply of sediments from upstream.

It cannot be ruled out that the ice stream may have shut down for reasons external to the basal system, such as the catchment area becoming exhausted of ice, or sea level fluctuations altering conditions for easy ice evacuation. However, these scenarios are not favoured, as it requires the distinctive geomorphology and ice stream shut-down to be a chance coincidence, and yet there seem good reasons to indicate causality. Examination of further ice stream beds that have shutdown may resolve this uncertainty, and it is interesting to note that on southern Wollaston Peninsula, Victoria Island, a possible ice stream (Sharpe, 1992) displays similar characteristics.

6.5.3. Sediment Discharge.

It would appear that the availability of soft sediments was an essential pre-requisite for the M'Clintock Channel Ice Stream. This has implications for the sediment discharge beneath the ice stream and it is possible to calculate crude estimates of the magnitude of this flux.

Surface lowering (erosion) near the terminus is around 45 m and 60 km upstream this lowering is reduced to 35 m. For the purpose of estimating a sediment flux, this surface lowering is attributed to one period of ice stream operation and it is assumed to decrease steadily to 0 m at the upstream limit of the ice stream (720 km). This assumes a wedge of sediment beneath the ice stream which is thickest in the onset zone and decreases towards the terminus. It is then possible to calculate the volume of sediment for one flow band which is a metre wide at the terminus but increases upstream to allow for flow convergence.

Using these estimates of surface lowering and applying them to a flow-band for the length of the ice stream (720 km), it produces a sediment volume discharge of 14.6 x 10^6 m^3 for the flow-band ((estimated sediment thickness) x (length) x (width for consecutive segments of the flow-band)). This gives a transport rate of 73,000 m^3 a^{-1} m^{-1} over 200 years of ice stream operation and 36,500 m^3 a^{-1} m^{-1} over 400 years of operation (according to Hodgson's (1994) dating constraints and the new data presented in Section 6.4.6). These figures represent a much higher transport rate than has been estimated for other ice streams in the literature, shown in Table 6.3.

Table 6.3. Estimates of sediment discharge rates from contemporary and palaeo-ice streams.

Ice Stream	Reference	Sediment Discharge Rate, Per Metre Width of Ice Stream Terminus; m^3 a^{-1} m^{-1}
M'Clintock Channel	this chapter	36,500 - 73,000
Ice stream B	Alley *et al.*, (1989)	100 - 1000
Lake Michigan Lobe	Alley (1991), Jenson *et al.*, (1995)	100 - 400
Isforden, Spitzbergen	Hooke and Elverhøi (1996)	560 - 980
Hudson Strait	Dowdeswell *et al.*, (1995)	800 - 37,300

Given the large size of the M'Clintock Channel Ice Stream, and its probable higher velocities, then it should be expected that its discharge would be much higher than contemporary Ice Stream B in Antarctica (Table 6.3).

For a large sediment package thought to have been transported from a Spitsbergen fjord to the shelf edge by a former ice stream, Hooke and Elverhøi (1996) used numerical modelling to estimate the likely range of sediment discharge rates required to effect this transfer. Their calculated sediment flux ranged from 560 to 980 $m^3 a^{-1} m^{-1}$. The flux for the M'Clintock Channel Ice Stream is orders of magnitude higher. However, their calculated fluxes should be regarded as minimum estimates, because there is no guarantee that the ice stream was actually functional as a sediment conveyor for the entire time that the fjord was occupied by ice. If the ice stream operated for shorter intervals during the glaciation then sediment transfer rates would have been considerably higher.

For the Hudson Strait Ice Stream, Dowdeswell *et al.*, (1995) used ice flux results from numerical models of the Laurentide Ice Sheet and likely sediment loads in the base of the ice stream to estimate the rate of sediment discharged from the terminus (Table 6.3). Such high sediment fluxes are required in order to explain the wide deposition of Heinrich layers in the North Atlantic but relate only to debris incorporated within basal ice and not for an underlying deforming layer. Inclusion of sediment yield from a deforming layer could drastically escalate these estimates.

Alternatively, the high sediment yield of the M'Clintock Channel Ice Stream may be a heavily biased by the assumption that sediment lowering *gradually* decreases to 0 m at 720 km upstream. If 0 m of sediment lowering is assumed to be at around 300 km upstream, sediment discharge decreases to between 37,035 $m^3 a^{-1} m^{-1}$ and 18,518 $m^3 a^{-1} m^{-1}$, respectively. This is more comparable to the estimates beneath the Hudson Strait Ice Stream (see Table 6.3).

If either of the estimates of the sediment discharge rate are correct, then the M'Clintock Channel Ice Stream was a major exporter of debris and likely resulted in accretion of a till delta in Viscount Melville Sound along the lines described by Alley *et al.*, (1989), and as found in Spitsbergen (Hooke and Elverhøi, 1996). Sediment carried in suspension or by ice rafting would have found its way into Viscount Melville Sound and down M'Clure Strait to be deposited on the continental slope. At the distal end of M'Clure Strait, bathymetry of the continental margin (Zarkhidze *et al.*, 1991) indicates the characteristic bulge of a sediment fan. Marine geophysical surveys and ocean coring could fruitfully answer questions of sediment supply, rates and timing of the M'Clintock Channel Ice Stream.

6.5.4. Implications for Ice Stream Basal Processes.

Given the characteristics of the ice stream bed and adjacent areas, what can be inferred about processes that operated at the base of the ice stream? A mechanism is required that:

- promotes high flow velocities,
- operates on fine-grained materials,
- erodes and transports large quantities of sediment,
- produces, at least during the final stages, a suite of drumlins and mega-lineations, with a downstream increase in packing and has an off-switch triggered by sediment exhaustion whereby the rate of removal exceeds supply and so down-cutting reaches bedrock.

The deforming bed model of fast glacier motion (Alley *et al.*, 1986; Boulton and Hindmarsh, 1987) is capable of accounting for all of the above conditions. Rapid ice flow may have been accomplished by shear within saturated fine grained carbonate sediments, which would require a large sediment flux and could produce subglacial bedforms (Boulton, 1987; Hart, 1995; Boyce and Eyles, 1991). The hypothesis developed earlier for ice stream shut-down (deformation steadily depleting the sediment pile until it reaches bedrock, thus increasing basal friction and retarding streamflow) by no means proves the existence of the deforming bed model, but it does demonstrate how the geomorphological record and glacial history of the ice stream can be explained by it.

It has been supposed that drumlins are formed within a deforming layer of till, by deformation around more resistant cores, and that individual drumlins can become 'uprooted' and migrate downstream (Boulton, 1987; Hindmarsh, 1998). If retardation of ice stream flow initiated in a downstream position with deforming bed processes ceasing here first, then drumlin migration would be expected to decrease at this point also. Meanwhile, drumlins formed up-ice continued to migrate downstream and as they moved into the decelerating zone they were forced to progressively pack together more closely and eventually stop. This is directly analogous to the development of a 'traffic jam' on a multi-laned high-speed road with cars packing more closely together and coming to a stop at the point of flow retardation. This is exactly the arrangement we find in the ice stream bed; significant increase in packing and density at the retardation area (Figures 6.10 and 6.11), and slight downstream decreases in length and elongation ratio (Figures 6.8 and 6.9).

Once the supply of unconsolidated sediments had been exhausted, the deforming layer was not self sustaining (i.e. Cuffey and Alley, 1996) because the processes of bedrock erosion were too slow to provide enough lubricating sediment and there was not sufficient sediment input from up-ice. Thus, the life-cycle of this ice stream was pre-determined by an existing sediment thickness, with 50 m of sediment able to sustain fast flow for ~200 years, possibly longer.

Of possibly greater significance to the basal processes beneath the ice stream is the divergence of flow around Stefansson Island (Figure 6.2 and 6.19), which likely acted as a pinning point, imparting significant basal drag. The island was completely submerged by ice but was large enough to deflect flow around it and there is a zone of unmodified geomorphology laterally adjacent to its summit (cf. Hodgson, 1994). This patch reveals no erosion or remoulding of the till surface even though it was beneath the ice stream, and it is thus inferred that it was a sticky spot (see Section 2.3.3.1).

An unresolved factor of ice stream functioning is how enough drag is produced in order to prevent them from excessive acceleration (cf. Murray, 1997). Current views indicate apportioning between side drag from slower moving ice and basal drag from sticky spots (see Section 2.3.3). Further examination of the exposed parts of the M'Clintock Channel Ice Stream may be able to reveal the proportion of the bed that potentially acted as sticky spots, retarding basal flow.

If the sediments were not subject to deforming bed processes, then a mechanism must be found that accounts for the points mentioned above. Based on laboratory determinations that reveal till to have a plastic rather than viscous rheology, the deforming bed model with pervasive shear has been questioned (Tulaczyk *et al.*, 2000). It has been argued that failure should only occur in a shallow layer of sediment rather than pervasively over depth, and that fast flow could be facilitated by basal uncoupling under high porewater pressures (e.g. Piotrowski and Tulaczyk, 1999) and, or, by ploughing (Iverson *et al.*, 1995).

The difficulty in using these models to explain the observed geomorphology is that they do not seem to explain the (albeit inferred) very high sediment discharges or drumlin formation. However, Tulaczyk and Scherer (*in press*), propose that with a plastic rheology, ploughing of sediment by large bumps of ice (>10's m) protruding from the base of an ice stream may provide an explanation for fast flow, a degree of sediment evacuation, and creation of certain types of bedform.

After the ice stream shut down, preservation of landforms in the area was widespread. Flow-1 terrain has been preserved in close proximity to the fast flowing stream ice. It is tempting to ascribe the sharp contrast at the ice stream margin and the landform preservation to a basal thermal control (i.e. cold and warm-based ice (cf. Dyke, 1993; Kleman, 1994). However, in this case, final ice retreat across the area draped a suite of deglacial landforms (eskers and outwash fans) characteristic of warm-based ice. Thus, rather than invoking rapid changes in thermal regime, it is simplest to interpret that the whole portion of the ice sheet was warm-based which retreated without much energy for modification. If so, then the sharp margin was a function of differential velocity, and landform preservation is a function of the short duration of flow during retreat, which was not capable of completely erasing the subglacially-produced landforms (cf. Clark, 1999).

6.5.5. Ice Stream Magnitude and Significance.

The M'Clintock Channel Ice Stream was clearly a major feature of the Laurentide Ice Sheet and of comparable size to the Hudson Strait Ice Stream (770 by 150 km) which drained the north-eastern margin. Both these ice streams are considerably larger than contemporary ice streams (see Table 2.1), but this should be no surprise given the rapidity of deglaciation after 11,000 yr BP.

Given the position and size of the M'Clintock Channel Ice Stream, its catchment area is likely to have approached 400,000 km^2 and may have been much larger. If the ice stream (cross-sectional area 52 km^2, see Figure 6.20) operated for 200 years with a velocity of 4 km a^{-1} (Dowdeswell *et al's.*, 1995, estimated velocity of the Hudson Strait Ice Stream), then it would have drained around 45,000 km^3 of ice from its catchment area. Ice losses of this magnitude were probably great enough to drive the Keewatin Ice Divide southwards and affect the mass balance and flow structure of the whole ice sheet. Indeed, the known south-eastern migration of the Keewatin Ice Divide (Shilts, 1980; Dyke and Dredge, 1989), may have been a direct response to this ice stream.

The M'Clintock Channel and Hudson Strait ice streams were of a similar size, and so it is reasonable to expect their ice fluxes and iceberg discharges to be broadly similar for times when they were functioning. Palaeoceanographic evidence has revealed the dramatic impact that the Hudson Strait Ice Stream had on North Atlantic temperatures and climate (see Section 3.2.2). It is predicted that the M'Clintock Channel Ice Stream also imparted a significant impact on oceanography and climate. Whether the activity of the M'Clintock Channel Ice Stream is related to known meltwater events, (or Heinrich events) in the Arctic ocean (Poore *et al.,* 1999) deserves further investigation.

There are a number of problems in trying to accommodate the ice stream into existing glacial reconstructions of the area around the M'Clintock Channel. This is because the northward flow of the ice stream is inset within westward flows onto Victoria Island and eastern flows onto Prince of Wales Island and Boothia Peninsula (see Figure 6.19). The westward and eastward flow patterns either side of the M'Clintock Channel (and palaeoshoreline isobases) led Dyke (1984) to infer significant ice thickness and hence a dispersal centre *within* the Channel. This was called the M'Clintock Ice Divide (Dyke, 1984) and has featured in succeeding reconstructions of ice sheet geometry in the area (e.g. Dyke and Prest, 1987a, b; Dyke and Dredge, 1989).

This ice divide is thought to have migrated from the west to the east side of the channel between 13,000 and 10,000 BP. However, the existence of this ice divide is problematic because Hodgson's (1994) dating constraints have since demonstrated that this was the time when the ice stream operated. Given the weight of evidence regarding the ice stream and the probability that a major trough was more likely to encourage development of an ice stream rather than an overlying ice divide, it is concluded that a major ice divide could not have existed in this location during the Late Glacial (ca. 10,000 yr BP). Therefore, a challenge lies in building a reconstruction that satisfies both the northward ice stream flow patterns and the westward and eastward flows emanating from the channel.

It is tentatively suggested that an ice divide lay in the channel and fed the easterly flow patterns on Victoria and but later collapsed to feed a northward flowing ice stream. During deglaciation, thick ice would have created an oversteepened ice sheet profile at the mouth of the M'Clintock Channel. This would have become unstable as the ice sheet margin retreated during a time of rapid sea level rise. Thus, marine drawdown may have triggered the ice stream which obliterated the ice divide within the channel. Whether this led to the complete disintegration of the ice divide is doubtful because of the younger westerly flows on Victoria Island (Flow-3) which cross-cut the eastern-most ice stream bedforms.

6.6. Summary and Conclusions.

The M'Clintock Channel Ice Stream is reconstructed as 720 km in length, converging downstream from 330 to 140 km in width, with an approximate surface area of 162,000 km^2. It was fed by ice from the Keewatin Sector of the ice sheet, and its onset zone is inferred to be in the vicinity of the current mainland coast. The ice stream probably terminated in Viscount Melville Sound, just beyond Stefansson Island, where water depths drop rapidly to 500 m. The ice stream was at least 400 m thick 50 km upstream from its terminus and the cross-sectional area is estimated as ~56 km^2 (Figure 6.20). It is comparable in size to the Hudson Strait Ice Stream, and is much larger than any contemporary examples from Antarctica or Greenland.

The location of the ice stream was controlled by topography with the occurrence of pre-existing unconsolidated drift encouraging fast flow velocities. Glaciological regulation, rather than properties of the bed, controlled the location of the margins and hence ice stream width.

The ice stream bed is characterised by subglacially-produced landforms, which record a snapshot view (isochronous record) of the bed just prior to ice stream shutdown.

Whilst some subsequent modification has occurred in places, this valuable record has remained preserved close to its original state.

The ice stream bed comprises drumlins and megalineations of great length (up to ~20 km) and elongation ratios (approaching 30:1), and which show a systematic downstream increase in packing and density. This flow pattern is characterised by an extremely abrupt margin (< 1 km) and ice stream marginal moraines (23 km in length) have been identified. Till thickness decreases in the downstream direction with the proportion of bedrock exposed at the bed increasing. There is also a strong contrast in drift thickness between the ice stream bed and adjacent terrain, indicating that the ice stream eroded into pre-existing unconsolidated sediments, producing a surface lowering of up to 45 m in the downstream portion. A first order estimate of sediment discharge along a flow-band just inside the stream margin produces a maximum rate of 73,000 $m^3\ a^{-1}\ m^{-1}$ (i.e. per metre width of the terminus).

From these observations of a portion of the ice stream bed, the following inferences about basal processes are made:

- Debris discharge from the ice stream may have been exceptionally high over a short period of time (200 years).
- The combination of high sediment flux, fast flow velocities and creation of bedforms is most easily explained by the operation of a deformable layer of sediment beneath the ice stream.
- The ice stream experienced frictional shutdown as a result of cutting down through pre-existing unconsolidated sediments until it reached bedrock, which produced higher levels of basal friction. Somewhat surprisingly, it reached bedrock firstly at its terminus, retarding the flow there, which then propagated upstream.
- The ice stream was not self-sustaining in that it did not generate its own lubricating sediment in sufficient quantities, but merely depleted pre-existing sediment. Its life-cycle was thus predetermined by initial sediment thickness. In this case, a thickness of 50 m was required for a minimum of 200 years of ice streaming, i.e. you need a lot of sediment to run an ice stream.

Given the extra timing constraint from Prince of Wales Island (Dyke *et al.,* 1992) the ice stream operated for a maximum of 400 years, between 10,400 and 10,000 BP, but more likely for only 200 years prior to 10,000 BP (Hodgson, 1994). If it operated for 200 years with a velocity similar to that thought for the Hudson Strait Ice Stream, then it could have drained around 45,000 km^3 of ice from its catchment. This would have significantly affected the dynamics of the shrinking Keewatin Ice Sheet and drive its ice divide south-eastwards.

The ice stream was active at a time of great ice sheet readjustment (Clark, P.U., 1994) particularly around the north-east margin, with Heinrich Event-0 (Hudson Strait ice streaming) occurring at 11–10 ka, and the Gold Cove readvance between 9.9–9.4 ka (Andrews, 1998; Miller and Kaufman, 1990). Whether the M'Clintock Channel Ice Stream is part of a pan-ice sheet destabilisation or it acted independently is discussed in Chapter 8.

The concept of the M'Clintock Ice Divide in the northwestern part of the Laurentide Ice Sheet needs re-evaluating in the light of the evidence for an ice stream at this location and time. A reconstruction is required that satisfies both the complex flow patterns and the isobases on palaeoshorelines that indicate areas of glacial loading. It is tentatively suggested that an ice divide lay in the channel prior to ice stream activation which may have been triggered by an oversteepened ice sheet profile and rising sea levels. The ice stream would have obliterated the last vestiges of this ice divide.

The M'Clintock Channel Ice Stream was a major component of the Laurentide Ice Sheet during deglaciation, accelerating its demise, and probably having far-reaching effects on sediment delivery to the continental shelf and Arctic oceanography. The possibility of the ice stream operating on more than one occasion should not be ruled out. It is predicted that a large till delta, fed by the ice stream, exists in Viscount Melville Sound, and that there is a strong ice stream signature in the form of a sedimentary fan on the continental shelf.

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Chapter 7: The Dubawnt Lake Ice Stream.

7.1. Introduction and Rationale.

The M'Clintock Channel Ice Stream (Chapter 6) was marine-based and drained the north-western margin of the Laurentide Ice Sheet. This ice stream was analogous to contemporary ice streams in West Antarctica because of its configuration, setting and inferred flow mechanism (deformable bed?). All contemporary ice streams are marine-based but this can be considered a geographical coincidence (Section 2.2.2). It should be assumed that during previous glaciations, some ice streams terminated on land but these have no modern analogues. What does the configuration of a terrestrial ice stream look like? How does it evacuate ice rapidly? To answer these questions (and many more) we must search for evidence of terrestrial ice streams from the now exposed beds of extinct ice sheets. Moreover, it could be argued that terrestrial ice streams may leave a more complete bedform record of their activity, because their evidence is (in most cases) more likely to be available for scrutiny. This is in contrast to marine-based palaeo-ice streams whose evidence is often partly obscured by subsequent marine inundation.

This chapter provides a detailed investigation of a previously undetected terrestrial ice stream from the Keewatin Sector of the Laurentide Ice Sheet. It begins with a synopsis of the previous work, which concentrated on explaining a number of conflicting ice flow patterns in relation to a major ice divide centred in Keewatin. Of these, several authors noted a spectacular drumlin field north-west of Dubawnt Lake (District of Keewatin). Using satellite imagery to map glacial lineaments, several lines of evidence suggest that the flow pattern was produced by an ice stream. The extent of the ice stream is reconstructed and the timing and context of bedform generation is ascertained. Analysis of within stream variations in subglacial bedforms provide several insights regarding the functioning of the ice stream. The influence of regional topography and geology are investigated and controls on ice stream location, initiation and shut-down are explored.

7.2. The Keewatin Sector of the Laurentide Ice Sheet.

7.2.1. Significance.

The Keewatin Sector of the Laurentide Ice Sheet lies to the west of Hudson Bay and is crucial to our understanding of the inland ice divides and domes. It is thought to have represented one of the three major domes of the Laurentide Ice Sheet, namely the Keewatin, Baffin and Labrador domes. While the marginal changes of the Laurentide Ice Sheet are relatively well constrained during deglaciation, the inland sectors of the ice sheet have often been neglected. As such, flow patterns during glacial maxima and the growth and decay of major ice domes throughout glacial cycles are poorly constrained (Shilts, 1980). In addition, changes in ice sheet geometry left behind a series of contrasting ice flow indicators which has often led to controversial hypotheses regarding the location of ice divides and domes. This is most clearly illustrated by the single-domed versus multi-domed hypotheses for Laurentide Ice Sheet dispersal centres. The location and duration of the Keewatin Ice Divide has been subject to debate for over 100 years (see discussion below, Section 7.2.2).

The Keewatin sector is also important because of its proximity to Hudson Bay and its role in the final deglaciation of the Laurentide Ice Sheet. It is thought that this area may have harboured the last vestiges of the continental ice sheet. However, the general lack of data from the interior has prevented objective reconstructions of its configuration. As Shilts (1980) noted, most inferences from this area have been drawn indirectly from geophysical data, from limited isostatic data from raised shorelines, and from broad large-scale reconnaissance mapping of glacial features.

7.2.2. Previous Work.

7.2.2.1. Multi-Domed Versus Single-Domed Ice Sheet.

The early work of Tyrell (1898) suggested that the Keewatin sector of the North American (Laurentide) Ice Sheet served as one of a number of dispersal centres around Hudson Bay. In contrast, Flint (1943) challenged this hypothesis and instead proposed a single domed ice sheet that became thickest over Hudson Bay. Flint further suggested that any other dispersal centres were restricted not only to limited times, but were also limited in extent. As a consequence, the general conclusion in the 1950's was that a major ice dome sat over Hudson Bay, with ice radiating out from it in all directions. This configuration was also advocated by the more recent steady state numerical modelling of Hughes *et al.* (1981) which is shown in Figure 7.1.

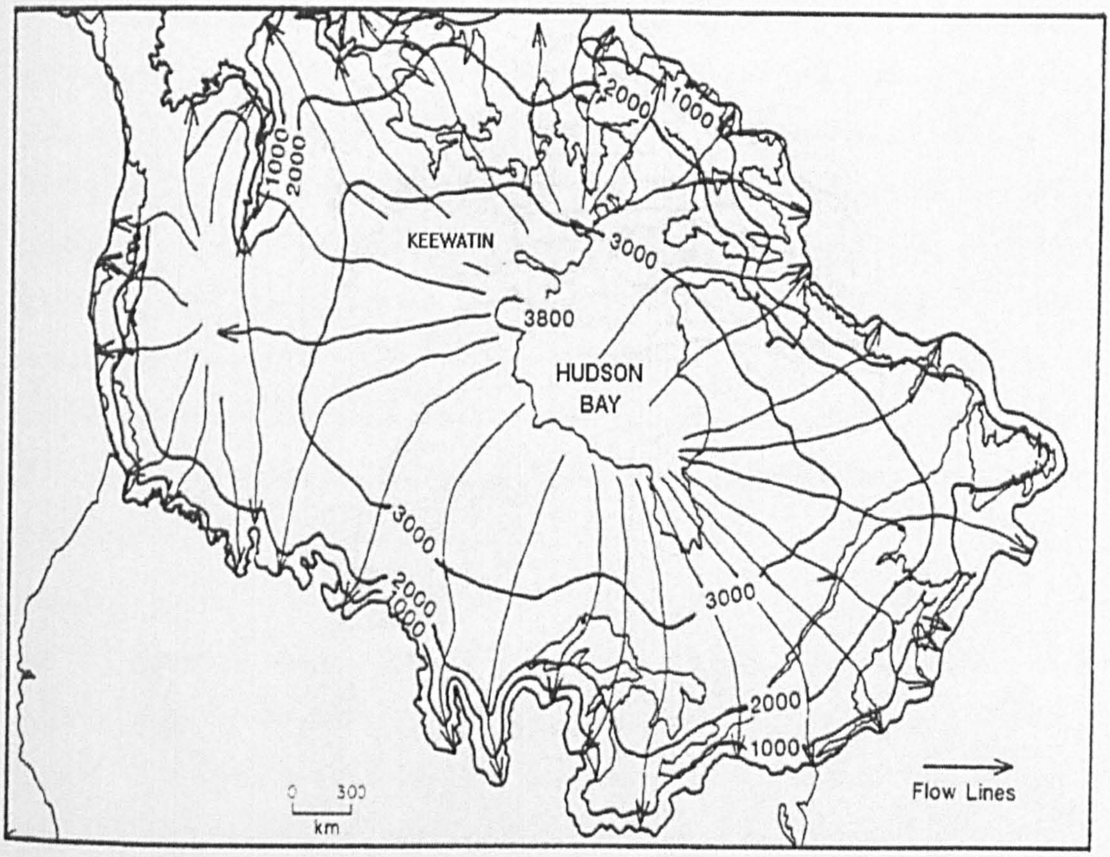

Figure 7.1. Single-domed model of the Laurentide Ice Sheet at 18,000 yr BP showing ice thicknesses and flow lines, after Hughes *et al.* (1981), taken from Andrews (1982).

This ice sheet configuration (Figure 7.1) implied that the Keewatin sector (to the west of Hudson Bay) was subjected to ice flow originating from the east and is supported by a number of field investigations in the area. Bird (1953) identified three distinct flow directions based on striae and drumlin orientations in central Keewatin. It was hypothesised that the earliest of the three ice flows was in a south-westerly direction, followed by a westerly flow. After the centre of outflow over Hudson Bay shifted to the south, Bird suggested that the Keewatin area was subjected to ice flow in a north-westerly direction. In conclusion, he stated that a major ice divide must have been present over the Hudson Bay area. This did not totally negate the idea of a Keewatin Ice Divide, but rather, relegated it to either the onset of glaciation or to the final stages of collapse when the ice sheet began to split up into discrete centres. Taylor (1956) used similar evidence from north-central Keewatin to identify the three principal flow patterns identified by Bird (1953) and also agreed that the oldest ice flow was to the south-west, followed by a westerly and then north-westerly ice flow.

Although Lee *et al.* (1957) and Lee (1959) found conclusive evidence of eastward and south-eastward flow towards Hudson Bay, they attributed this to a late glacial event when marine incursion into Hudson Bay split the dome into two dispersal centres (Labrador to the east and Keewatin to the west). This led Lee *et al.* (1957) and Lee (1959) to define the Keewatin Ice Divide as "the zone toward which the western remnant of the Laurentide Ice Sheet shrank". On the basis of ^{14}C dates of marine shells, Lee (1959) speculated that the Keewatin Ice Divide was in existence sometime between 7,000 and 8,000 years ago. The same evidence was used to conclude that ice had completely left Keewatin around 6,000 years ago. Although the Keewatin Ice Divide was recognised as a discrete dispersal centre its longevity was restricted to a few thousand years during deglaciation.

Following the work of Lee *et al.* (1957) and Lee (1959), Cunningham and Shilts (1977) mapped the surficial deposits of the Baker Lake area (District of Keewatin) in order to ascertain the complex ice flow indicators associated with a Keewatin Ice Divide. They argued that the conflicting ice flow directions can be explained by the southward and eastward migration of the ice divide. Cunningham and Shilts (1977) confirmed the earlier work that suggested that the oldest ice flows were in a southerly direction and attributed this to an ice divide located in northern Keewatin. The ice divide is then thought to have migrated southward and eastward, thus subjecting the area to north-westerly ice flow while also accounting for flow eastward into Hudson Bay. Unfortunately, no dating constraints are put on this migration and the duration of each of the flow patterns is not described. The longevity of the Keewatin Ice Divide remained elusive.

7.2.2.2. The Keewatin Ice Divide.

It was not until the work of Shilts *et al.* (1979) that the importance of the Keewatin Ice Divide was fully recognised. They studied several thousand till samples and glacial deposits in order to investigate dispersal trains to the west of Hudson Bay. The results were revealing in that they discovered 300 km long dispersal plumes emanating from central Keewatin near Yathkyed Lake and Baker Lake (see Figure 7.4 for locations), which were traceable all the way to Hudson Bay. This implied an eastward ice flow from an ice divide in Keewatin towards Hudson Bay. Although this had been hinted at in the earlier literature, it was always assumed to be a Late Glacial flow. However, Shilts *et al.* provided compelling evidence to suggest that it was of a far longer duration, thus implying a stable Keewatin Ice Divide.

The dispersal trains were not thought to be formed by several glacial advances because of their location near the centre of the ice sheet. More importantly, Shilts *et al.* (1979) argued that the dispersal trains were not late glacial. If the Keewatin Ice Divide only came into existence after marine incursion into Hudson Bay, the time allowed for transport at this time is constrained to between 1,000 and 3,000 years. This would imply minimum transport rates of between 60 and 300 m a^{-1} for a distance of over 300 km. However, relatively fast ice flow is not anticipated to have formed the dispersal trains, especially when it is considered that numerous esker systems deposited in the area are indicative of a massive stagnation of the ice sheet. Shilts *et al.* (1979) favoured an interpretation of the dispersal trains as a period of sustained ice flow lasting between 17,500 and 30,000 years respectively (assuming a minimum average flow rate of 10 m a^{-1}).

In their conclusion, Shilts *et al.* (1979) postulated that "the Keewatin Ice Sheet [divide] grew and remained on land west of Hudson Bay during the whole Wisconsin glaciation". Clustering of several centres of dispersal around Hudson Bay could leave behind an imprint which may be easily misinterpreted as radial flow lines emanating from Hudson Bay itself. Indeed, Shilts (1980) followed this work by suggesting that the main portion of the Laurentide Ice Sheet was made up of at least two land-based domes, one from Keewatin and one in Labrador. This provided considerable support for the multi-domed hypothesis, which is shown in Figure 7.2.

Some modelling experiments also support the multi-domed hypothesis for the Laurentide Ice Sheet, especially those which introduce soft deformable sediments beneath portions of the ice sheet, such as around Hudson Strait and the Prairie region of North America. Both Fisher *et al.* (1985) and Boulton *et al.* (1985) incorporated low yield stresses (deformable beds) into their numerical models of the Laurentide Ice Sheet. The introduction of deformable beds beneath the ice sheet resulted in areas of low surface profiles and abrupt changes in ice flow direction, but more importantly, created multiple domes and ridges, as shown in Figure 7.2. Moreover, these models agree well with the erratic dispersal patterns around Hudson Bay identified by Shilts *et al.* (1979) and Shilts (1980).

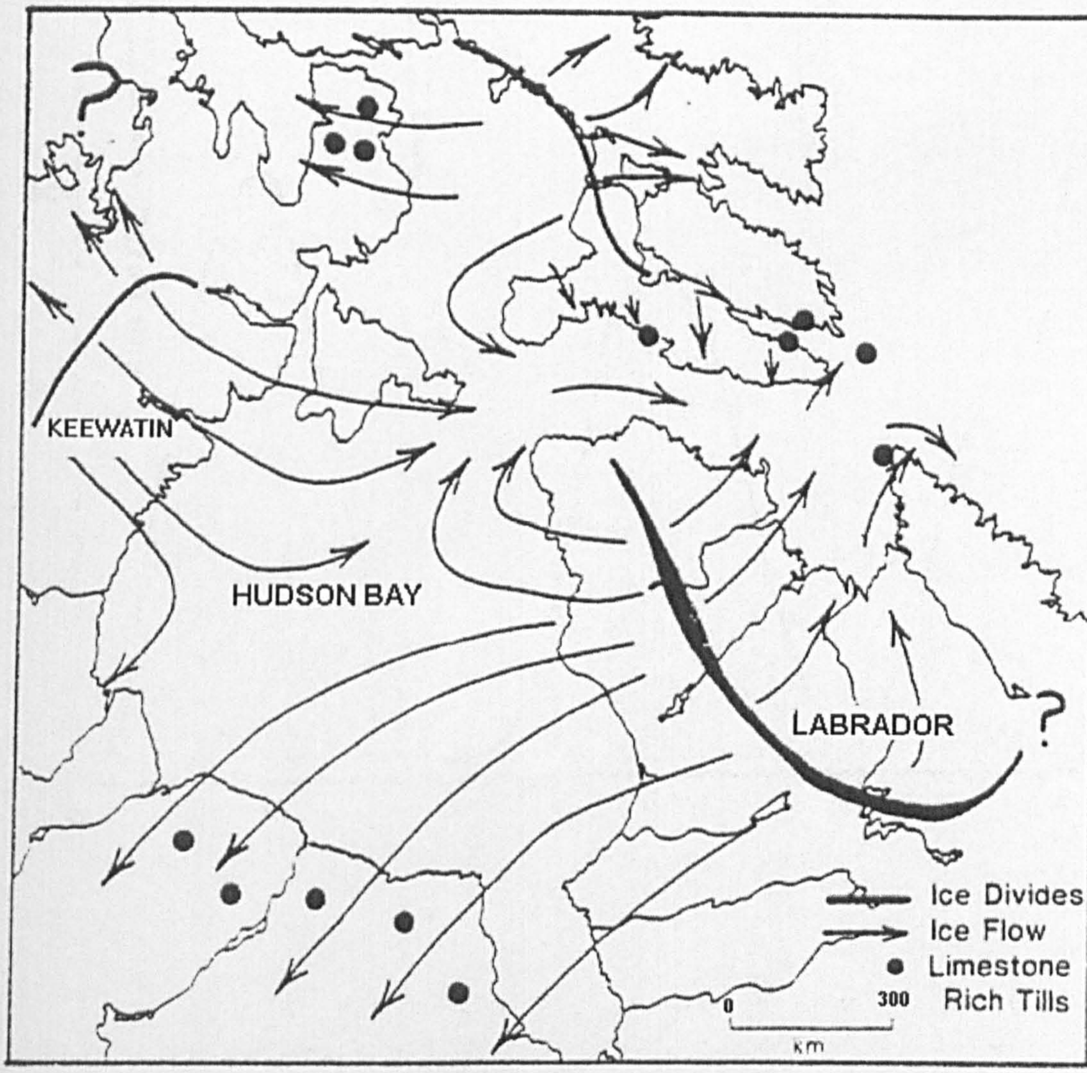

Figure 7.2. Multi-domed Laurentide Ice Sheet configuration in the Hudson Bay region, taken from Andrews (1982).

More recently, Boulton and Clark (1990a; b) have used geomorphological evidence to infer a highly dynamic Laurentide Ice Sheet, again invoking a multi-domed configuration. They discovered widespread patterns of lineations superimposed upon older flow patterns and reconstructed the sequential growth, migration and decay of several major domes. It is argued that such patterns of cross-cutting ice flow indicators can not be explained by a single domed ice sheet over Hudson Bay.

7.2.2.3. Subglacial Bedforms and the Keewatin Ice Divide.

Aylsworth and Shilts (1989a; b) focused on the bedform patterns of the Keewatin sector, centred on the Keewatin Ice Divide for an area of about 1,190,000 km^2. It was found that four broad zones were associated with the ice divide. Figure 7.3 shows the four zones in relation to the inferred location of the Keewatin Ice Divide.

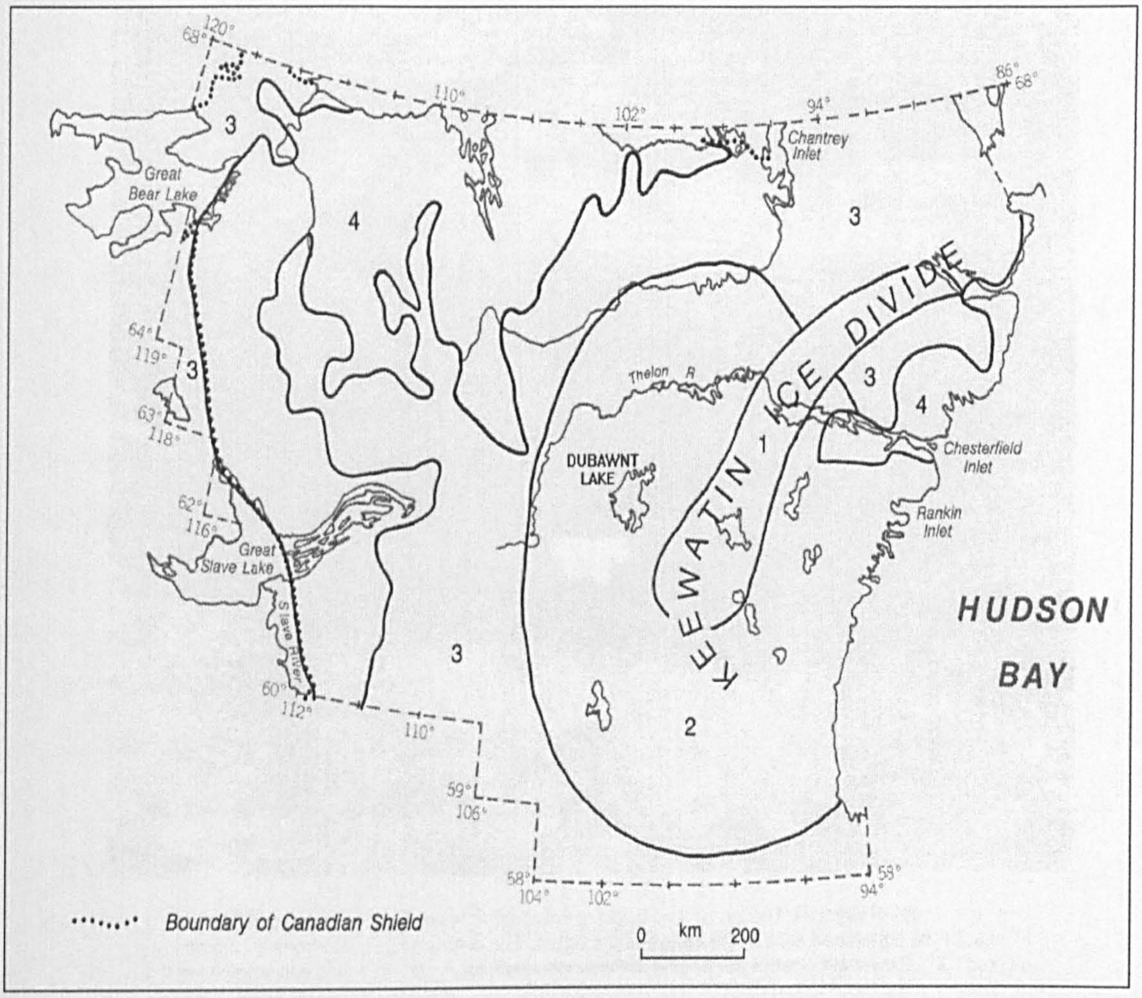

Figure 7.3. Glacial landform zones identified around the Keewatin Ice Divide after Aylsworth and Shilts (1989a), see text for explanation.

Zone 1 (which includes the ice divide) is characterised by low relief hummocky moraine and a marked absence of eskers. No orientated landforms are obvious and this landform assemblage is thought to be indicative of a very thin ice sheet during deglaciation. Zone 2 is characterised by extensive drumlin fields and esker systems but it is principally delimited by the occurrence of ribbed moraine, which commonly display a close lateral relationship with the drumlin fields. In Zone 3, the fairly continuous drumlinised drift thins out and the occurrence of eskers decreases. Zone 4 comprises exposed bedrock and the lack of glacial landforms is thought to reflect a general lack of available sediment. Although it may seem logical to assign each zone to different glacial dynamics, Aylsworth and Shilts (1989) stressed that the absence or presence of landforms may simply be a reflection of the sediment characteristics beneath the ice sheet and not the glaciological dynamics.

7.2.2.4. The Dubawnt Lake Drumlin Field: Ice Stream?

Much of the work specific to the Keewatin sector of the Laurentide Ice Sheet identified and mapped a distinct flow pattern approximately parallel to the northern shores of Dubawnt Lake (Figure 7.3) and trending in a north-westerly direction (e.g. Bird 1953; Lee, 1959; Shilts *et al.*, 1979; Aylsworth and Shilts, 1989a; b). The pattern is also represented on the Glacial Geological Map of Canada (Prest *et al.*, 1968) and from here onwards, will be referred to as the 'Dubawnt Lake flow-set'.

This flow-set was also mapped by Boulton and Clark (1990a; b); flow-set 3 on figure 10 of Boulton and Clark (1990b). They attributed it to a late glacial event (ca. 10,000 yr BP) related to a south-east shift in the Keewatin Ice Divide. In addition, Kleman and Borgström (1994) interpreted the flow pattern as a 'type landscape' for a surge fan in their glaciological inversion model for reconstructing former ice sheets. Surge fans are thought to display a characteristic bottleneck pattern, with eskers aligned parallel to flow in the distal part but running more obliquely in the proximal part. Lineations are thought to have been generated rapidly and surge fans are related to the decay stages of an ice sheet (Kleman and Borgström, 1994).

Bird (1953) was the first to note the unique nature of the bedforms in the Dubawnt Lake flow-set; "this region is a vast sandy drumlinoid plain with an area of nearly 10,000 square miles; throughout the plain, except in the few parts where rock ridges stick through the till cover, the shape of the relatively few lakes and the pattern of the stream drainage is totally controlled by the drumlinoids". This spectacular fluting led Aylsworth and Shilts (1989a) to speculate on the possible role of ice streaming in shaping the drumlins. However, they refused to postulate the exact location of an ice stream, adding that it is "difficult to relate drumlin formation to patterns caused by ice streams, partially because the concept has been so recently accepted that possible areas of drift deposited by ice streams have not been investigated".

7.2.2.5. Summary.

Early work in the Keewatin sector was aimed at resolving the debate concerning the multi-domed versus single-domed ice sheet. This work resigned the Keewatin Ice Divide to either early or late glacial times but the contrasting ice flow indicators could not be resolved. This paradigm was questioned by the work of Shilts *et al.* (1979) and Shilts (1980), who provided substantial evidence for a long-lived Keewatin Ice Divide based on erratic dispersal trains. This notion became widely accepted and the contrasting ice flow indicators could be adequately explained by the southward and eastward migration of the divide.

Recent work by Aylsworth and Shilts (1989a; b) investigated the landform assemblages associated with the ice divide and identified four broad zones emanating from its inferred final position. The second of these zones (zone 2) is characterised by extensive drumlin fields and Aylsworth and Shilts (1989b) speculated on the role of ice streams in shaping them. In particular, one flow pattern, characterised by elongated drumlins (the Dubawnt Lake flow-set) stands out as a potential candidate for ice streaming.

This chapter examines the evidence for an ice stream draining the Keewatin sector of the ice sheet and producing the Dubawnt Lake flow-set.

7.3. Methodology and Data Sources.

Satellite remote sensing is an ideal tool for identifying and mapping glacial landforms over large areas (cf. Chapter 5). Figure 7.4 is a location map of the study area showing several locations referred to in the text. Fifteen Landsat Multi-Spectral Scanner (MSS) hard copy positives (band 4) were developed into photographs, providing complete coverage of the study area (and the surroundings) at a scale of 1:250,000. The resolution of these images allowed the depiction of large (>1 km long) bedforms and the coverage permitted the identification of several regional flow patterns, which included the Dubawnt Lake flow-set.

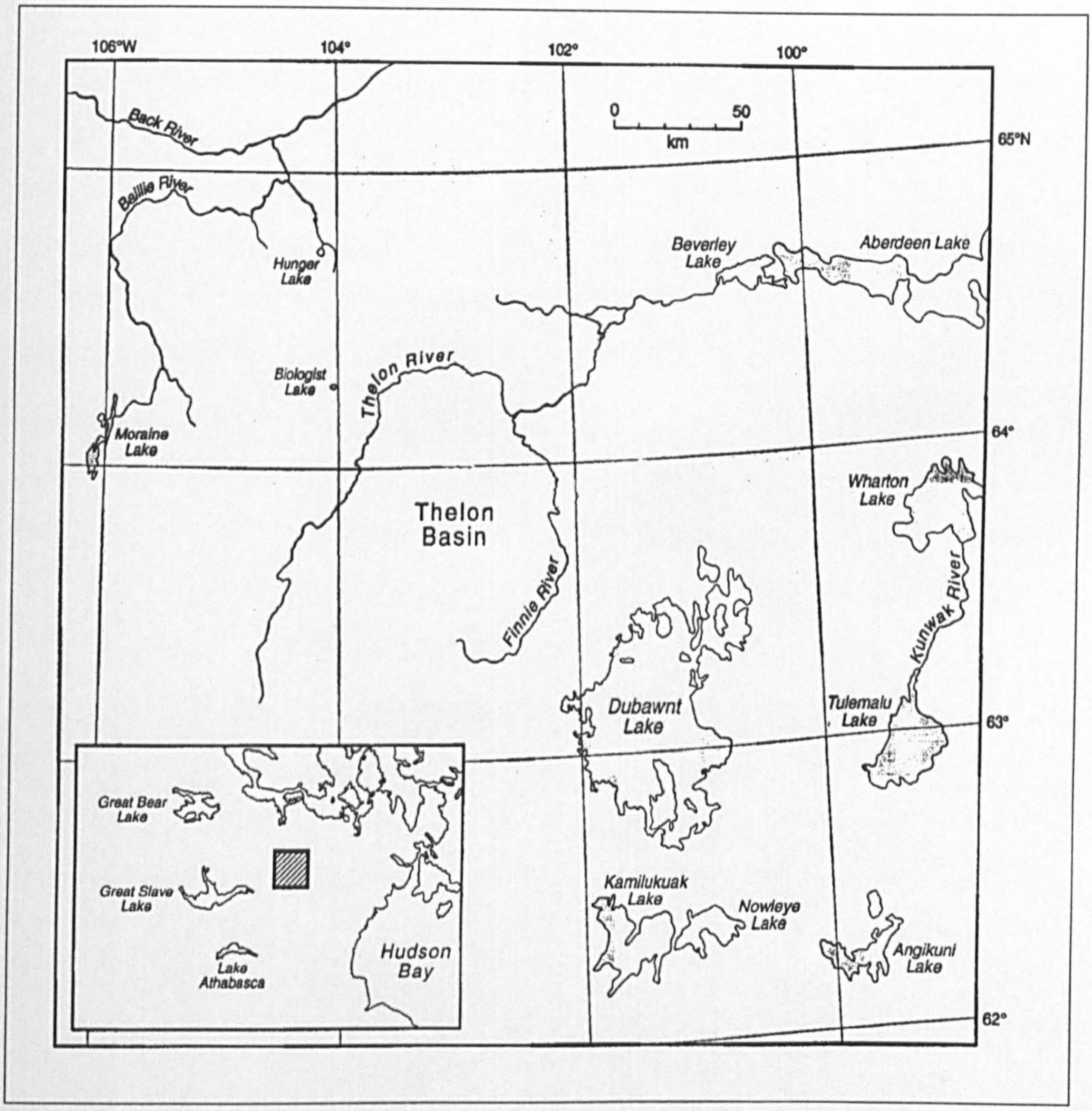

Figure 7.4. Location map of the study area covered by the Dubawnt Lake flow-set.

To investigate the Dubawnt Lake flow-set in detail and assess its validity as an ice stream track, three adjoining Landsat MSS images were purchased in digital format (bands 1, 2, 3 and 4). This allowed a more refined mapping approach. Each image had a spatial resolution of around 80 m and covered an area of 180 km by 180 km (Section 5.3.1). On screen mapping was accomplished following the basic methodology outlined in Section 5.6. The following bedform characteristics were measured; length, width, elongation ratio, 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 variations in bedform morphometry within the Dubawnt Lake flow-set, three flow-bands were constructed and gridded at 20 km intervals. Unfortunately, the coverage of the digital imagery did not extend to the northern margin of the flow-set. However, the flow-set is superimposed upon older flows in these areas (discussed below) and measurement of the bedforms would have proved problematic. In contrast, the three flow-bands closest to the southern margin of the flow-set provided a largely unmodified pattern.

In addition to the digital imagery, a 30-arc second (ca. 0.5 km) Digital Elevation Model was obtained from the study area (Section 5.5). This permitted the visualisation of the regional topography and allowed rapid transects of elevation to be calculated. To produce detailed transects, elevation data was taken from topographic maps (1:250,000) of the area with a contour interval of 20 m.

Information regarding the geology was obtained from the numerous (commonly field) investigations in the literature. Although many older maps of geology and geomorphology were available, most information regarding the surficial geology was obtained from the Geological Survey's Map 24-1987 (Aylsworth and Shilts, 1989a).

7.4. Results.

7.4.1. Regional Flow Patterns and Flow-Sets.

Figure 7.5 shows the regional flow patterns identified from the hard copy imagery at a scale of 1:250,000. All of these flow patterns matched well with those described in the literature and no new flow patterns were identified at this scale.

The oldest lineations depict ice flow in a south-westerly direction and have been best preserved immediately south-west of Dubawnt Lake (cf. Bird, 1953; Taylor, 1956). In this area, there is a cross-cutting relationship with a westerly ice flow direction which is presumably younger because of the superimposition. By far the most prevalent flow pattern on Figure 7.5 is that produced by the Dubawnt Lake flow-set (north-westerly flow pattern in bold). This pattern is the youngest in the area because it cross-cuts and is superimposed on all of the other flow-sets in the area.

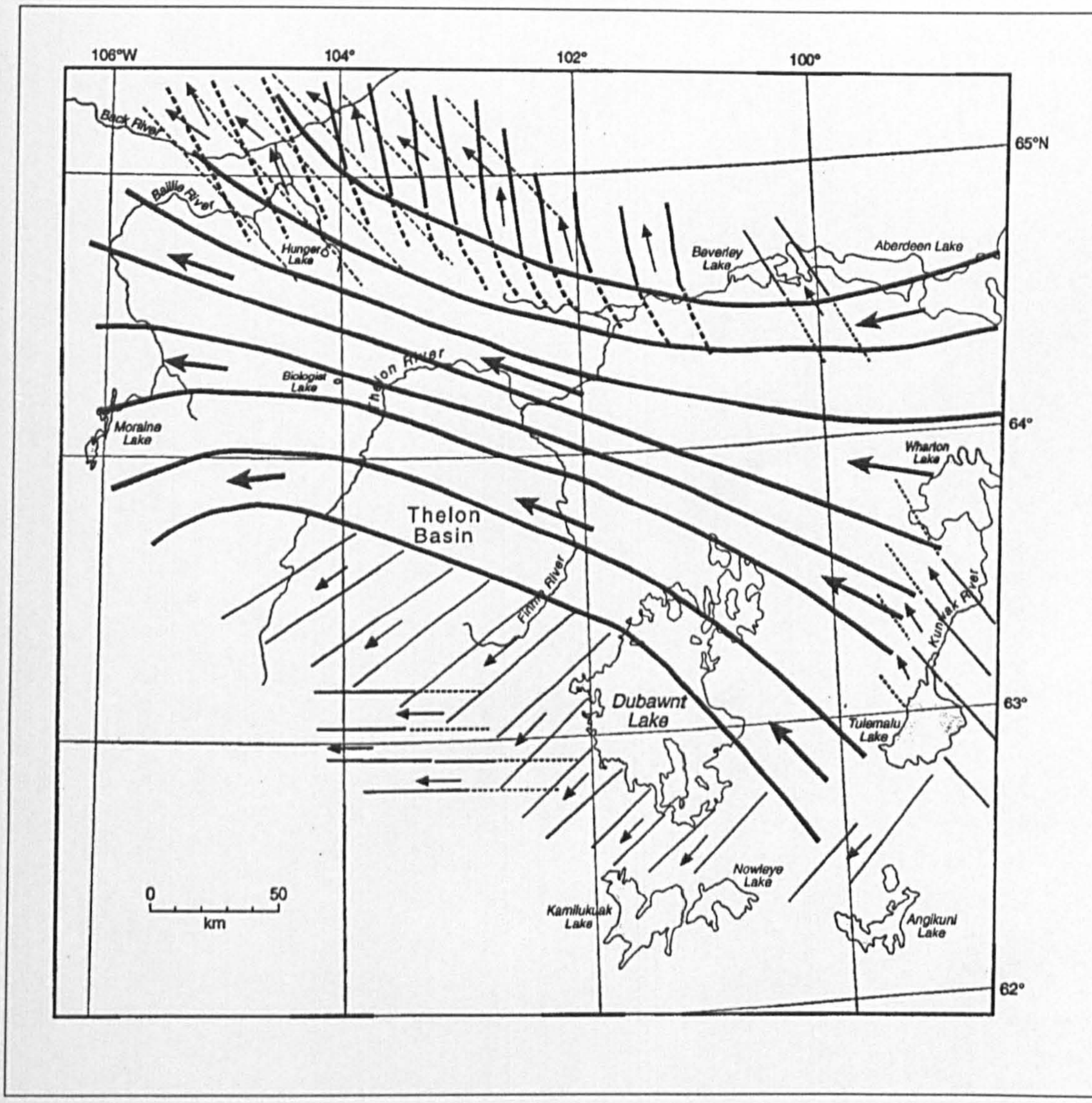

Figure 7.5. Regional flow patterns in western Keewatin reconstructed from Landsat MSS photographs.

Figure 7.6 shows the coverage of the three digital MSS images with the outline of the Dubawnt Lake flow-set superimposed, including reconstructed flow-bands. These images covered a total area of 91,135 km^2 and 11,825 lineations were identified and grouped into 14 possible flow-sets. Figure 7.7 shows all of the lineations detected on the imagery and Figure 7.8 shows the extent and location of each of the flow-sets identified (and named arbitrarily), not including the Dubawnt Lake flow-set. This represents the maximum number of flow-sets because although some flow-sets may have been produced by the same flow event, they were not grouped unless it was beyond reasonable doubt.

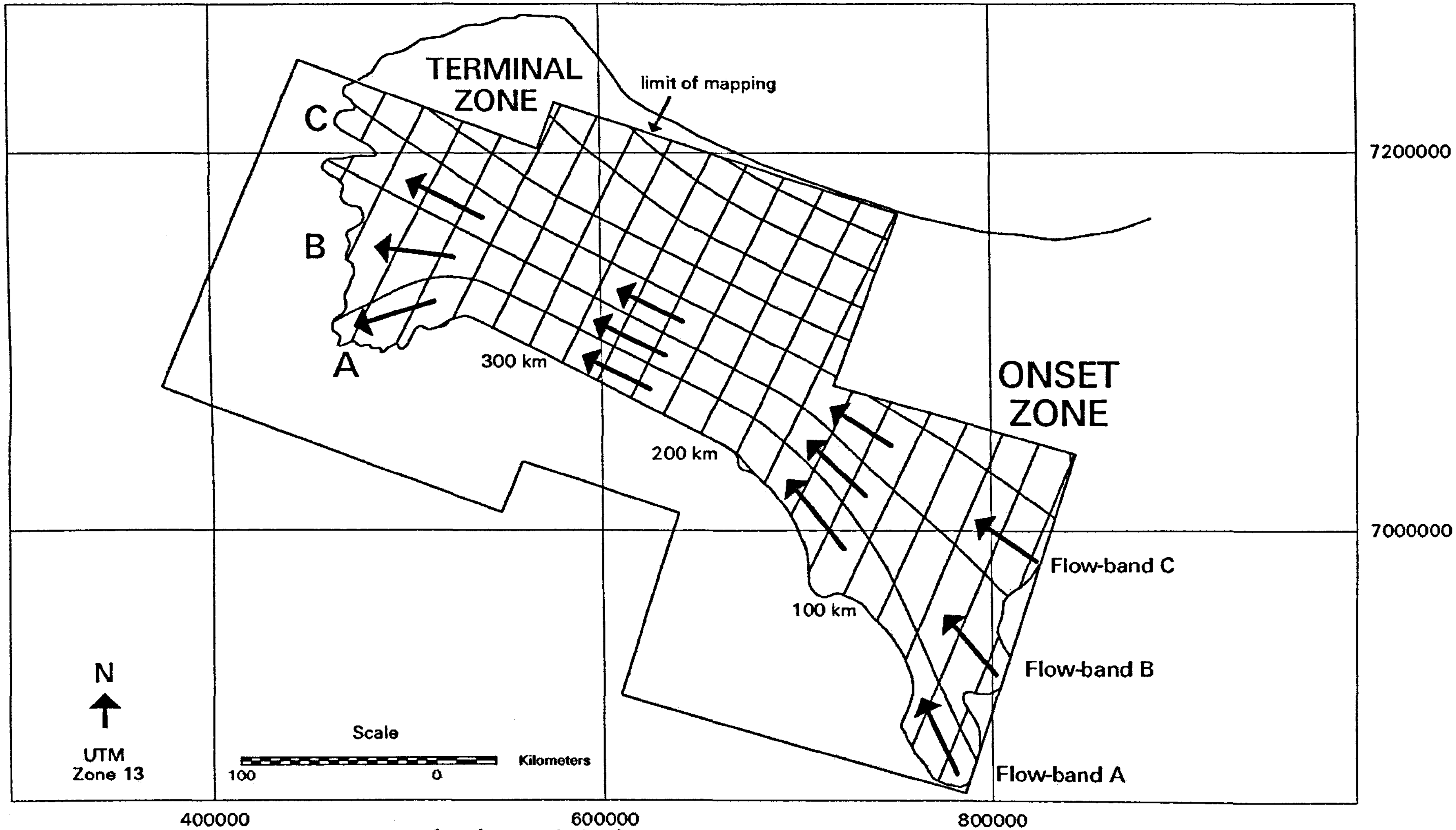

Figure 7.6. Digital satellite coverage of Dubawnt Lake flow-set and sampling grid, showing the location of flow-bands A, B and C.

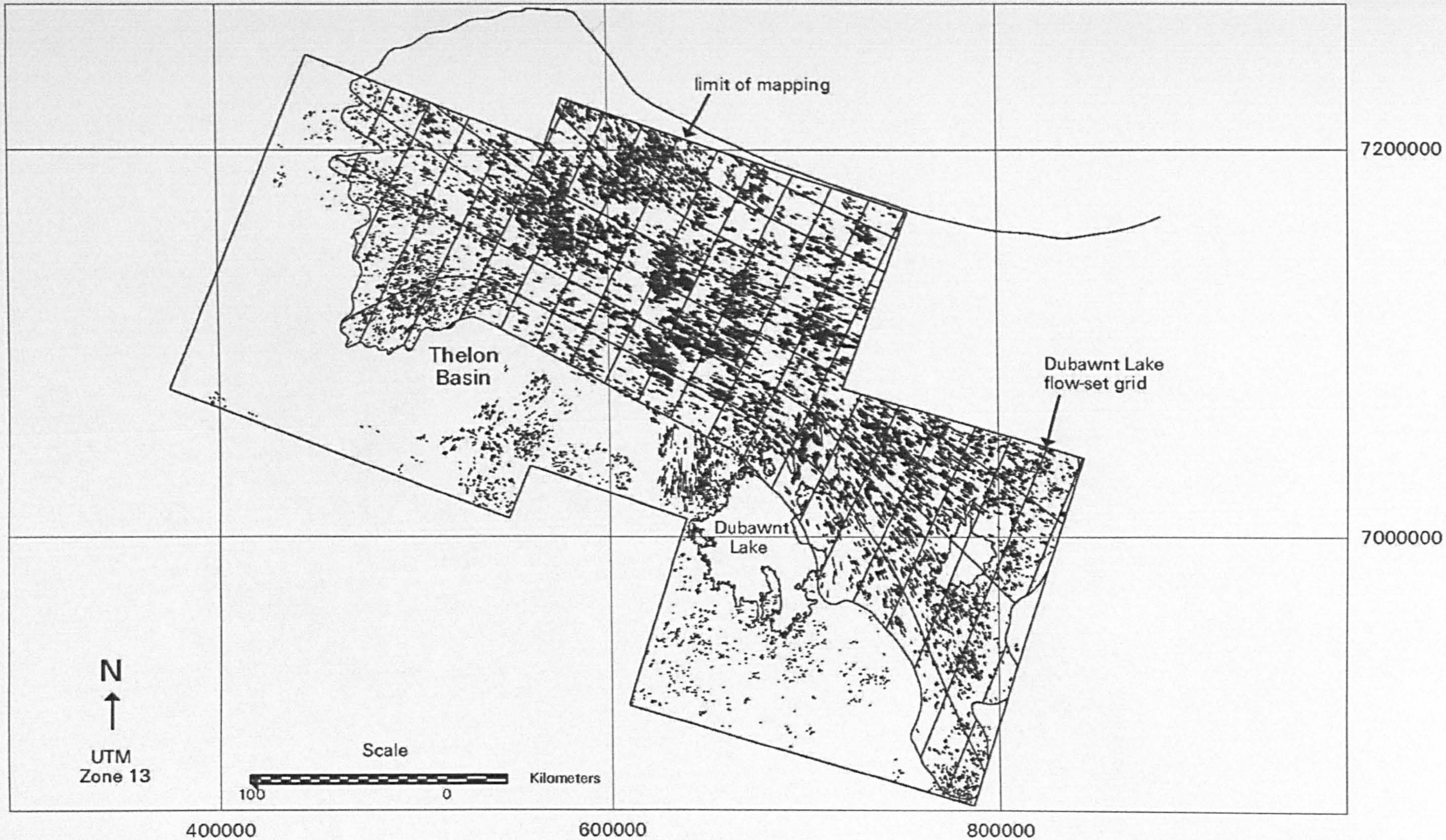

Figure 7.7. All lineations detected on the Landsat MSS imagery.

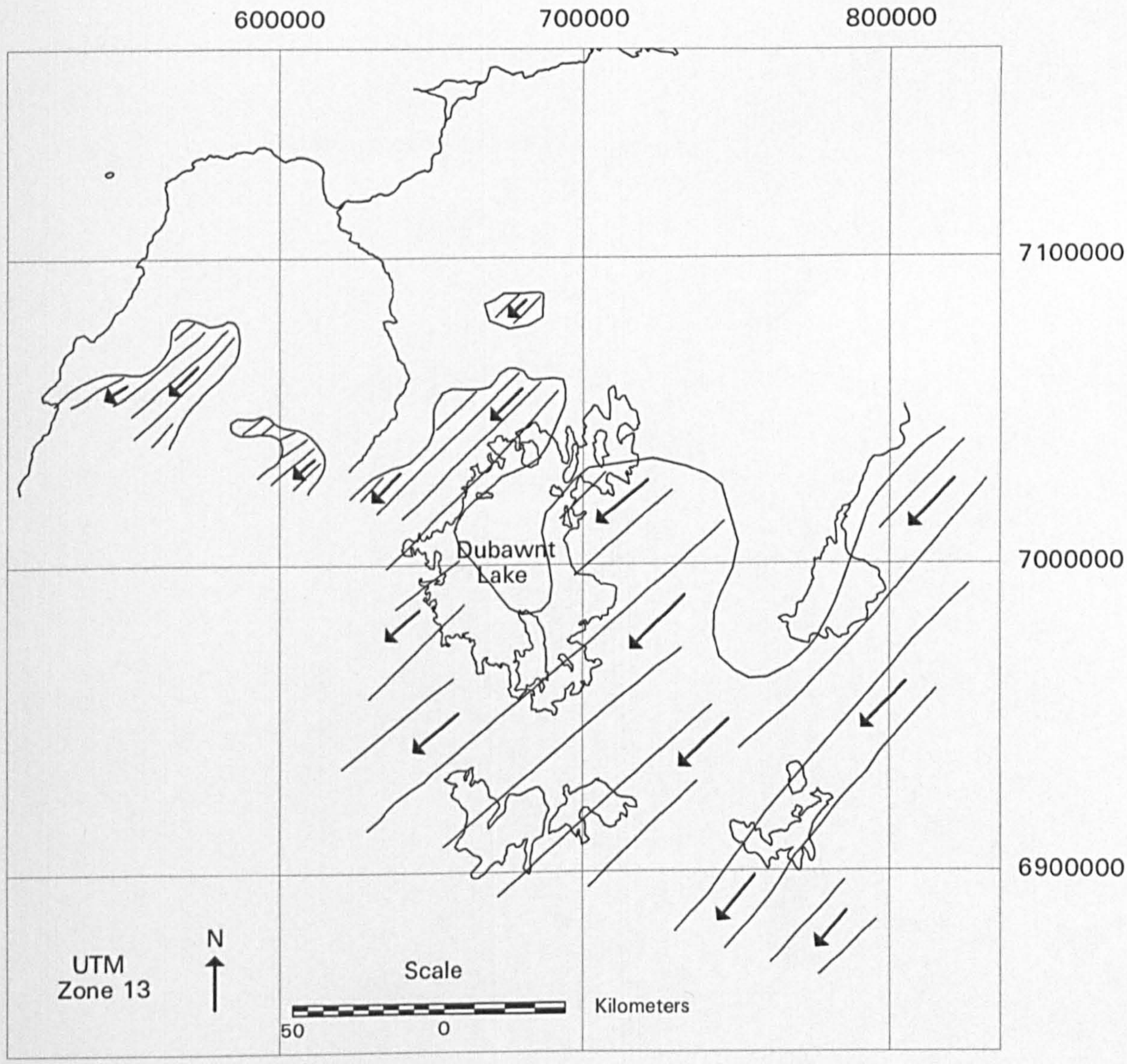

Figure 7.8a. Areal extent of Flow-set 1.

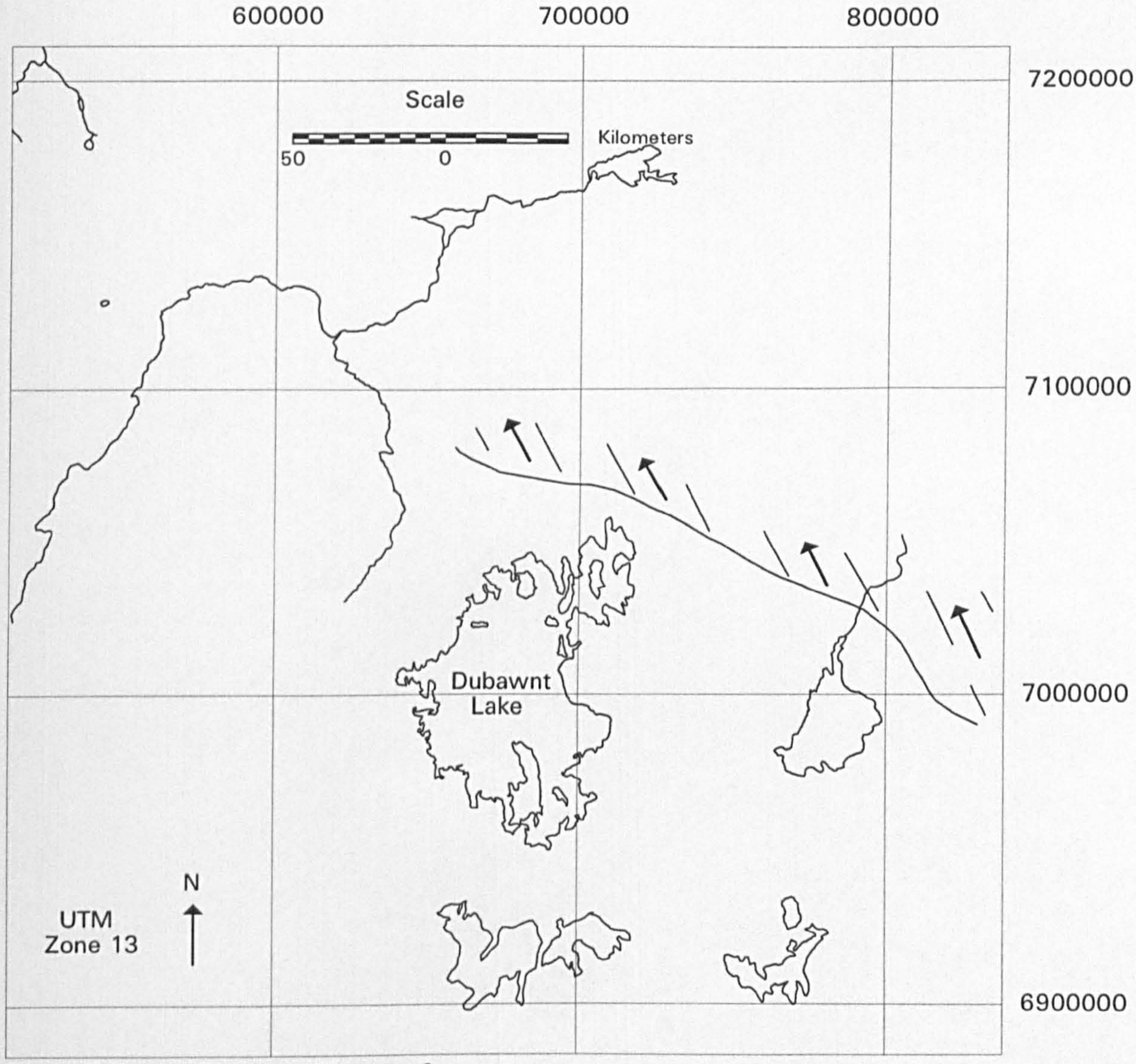

Figure 7.8b. Areal extent of Flow-set 2.

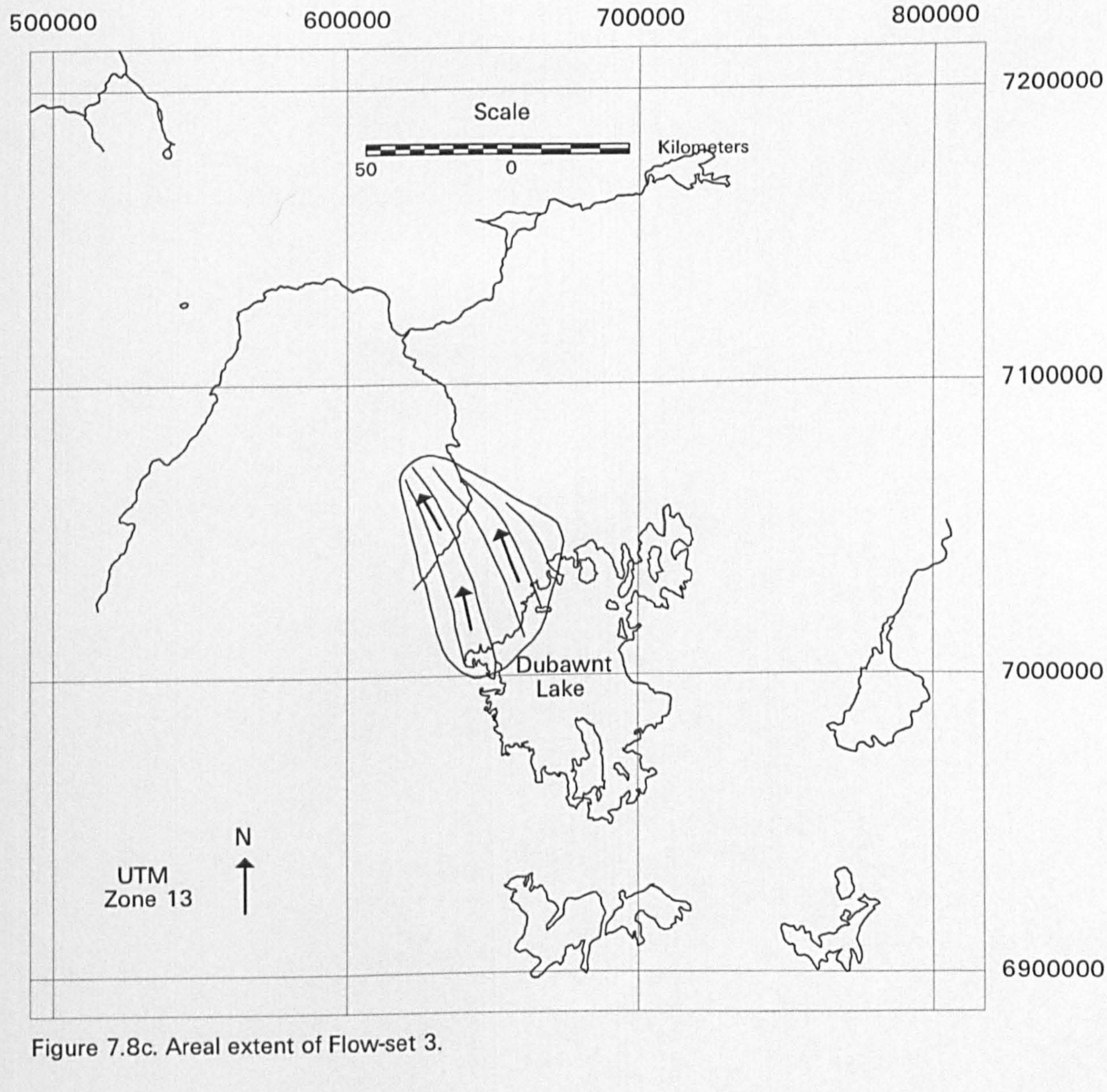

Figure 7.8c. Areal extent of Flow-set 3.

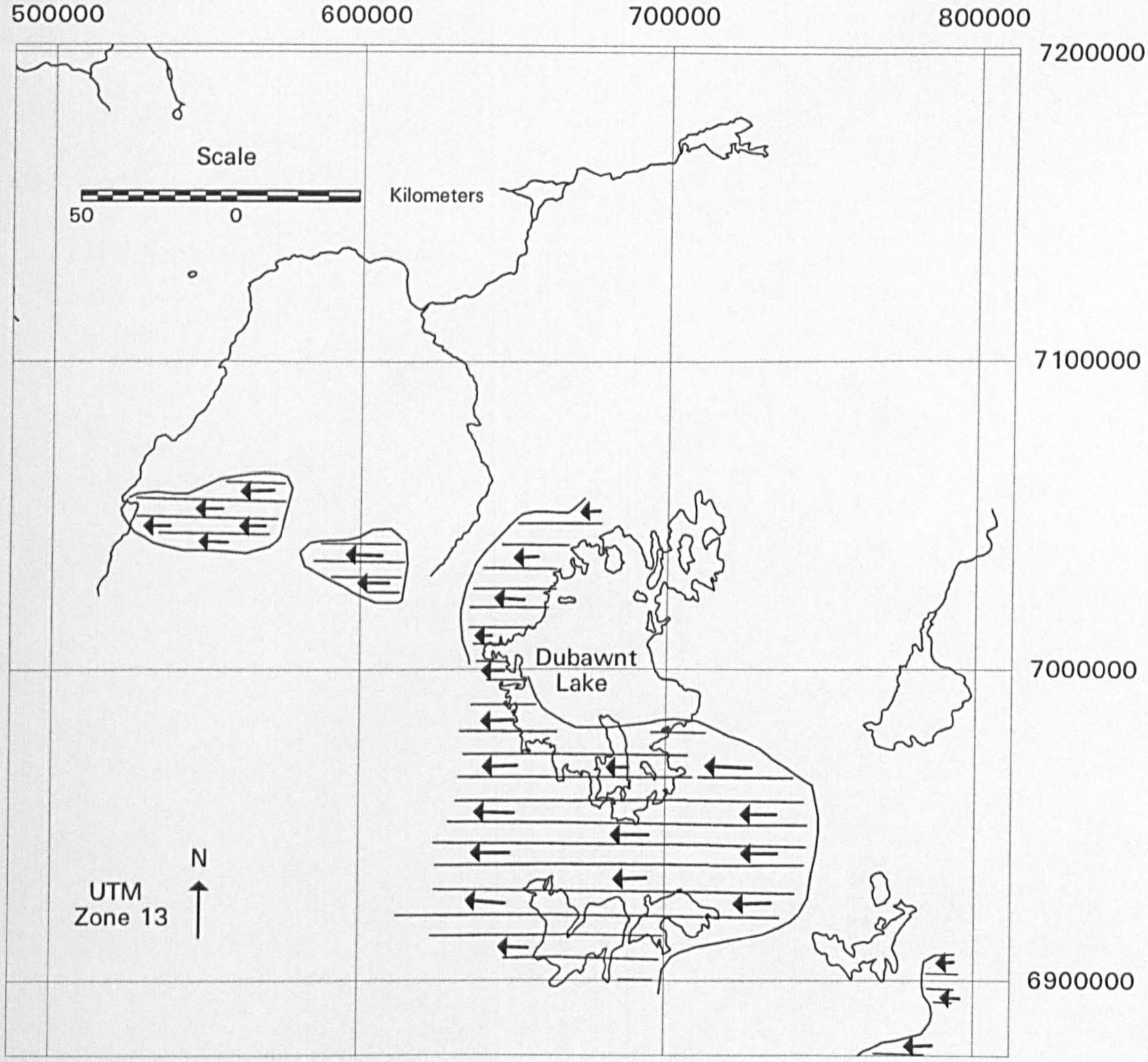

Figure 7.8d. Areal extent of Flow-set 4.

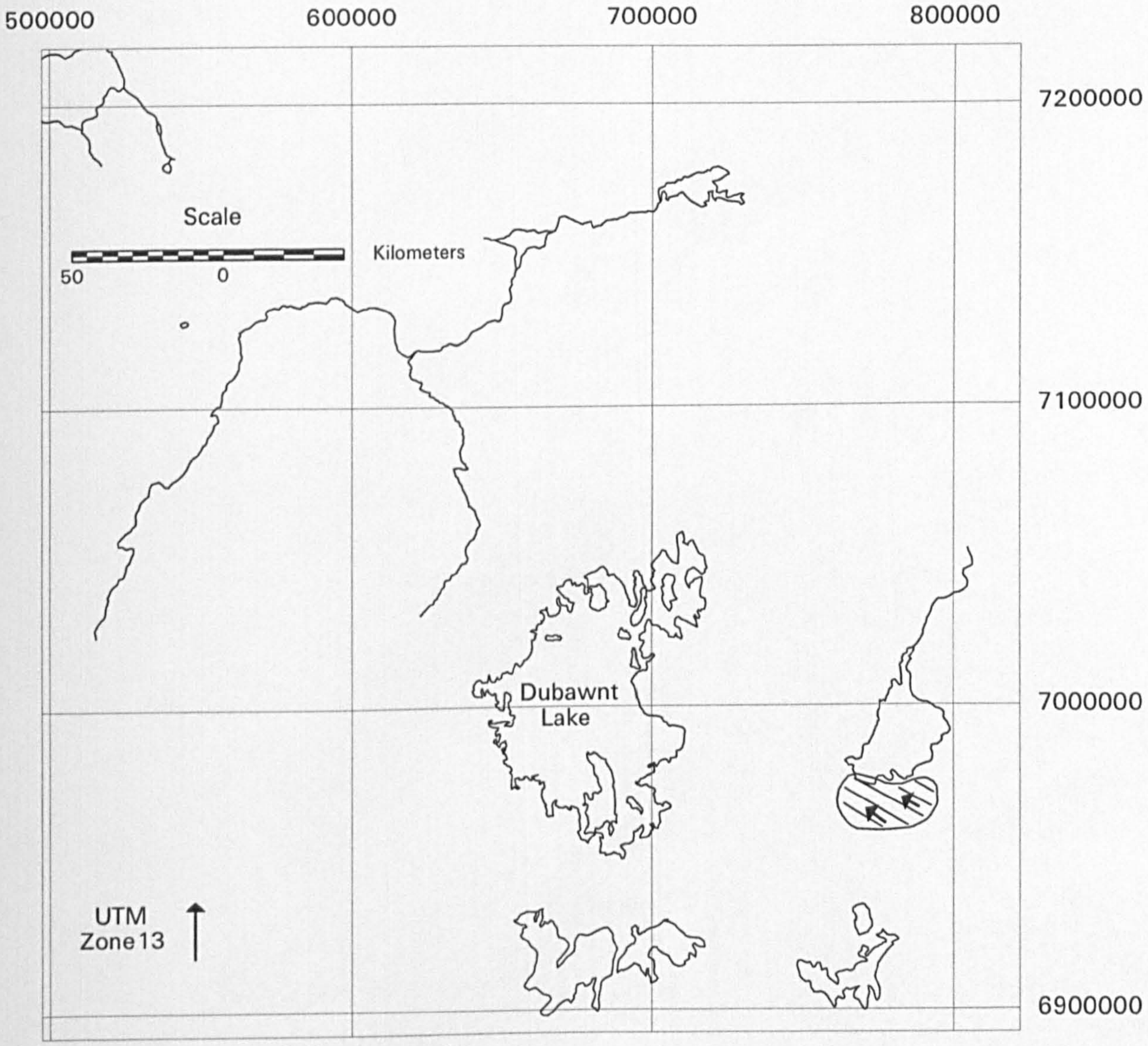

Figure 7.8e. Areal extent of Flow-set 5.

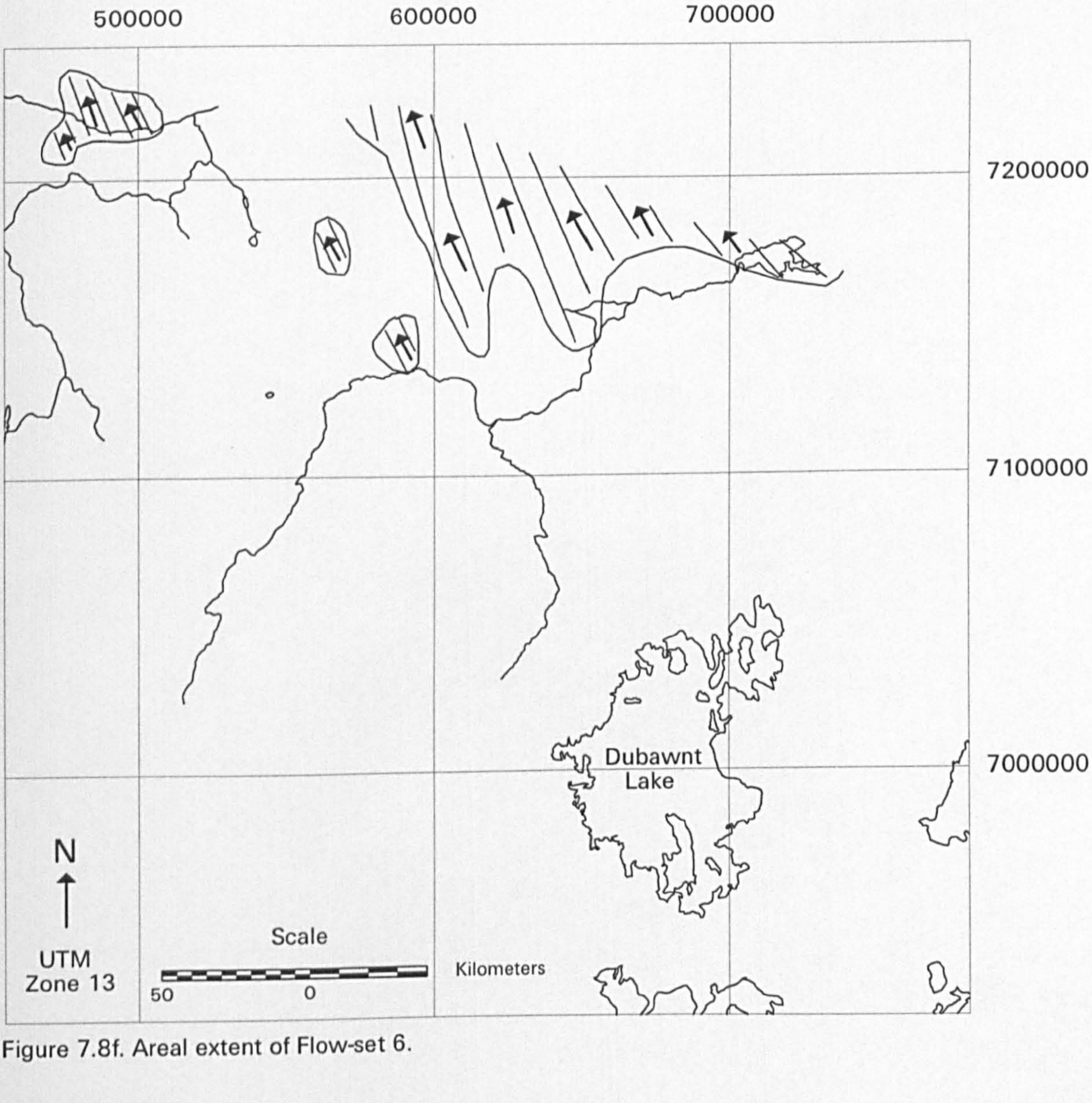

Figure 7.8f. Areal extent of Flow-set 6.

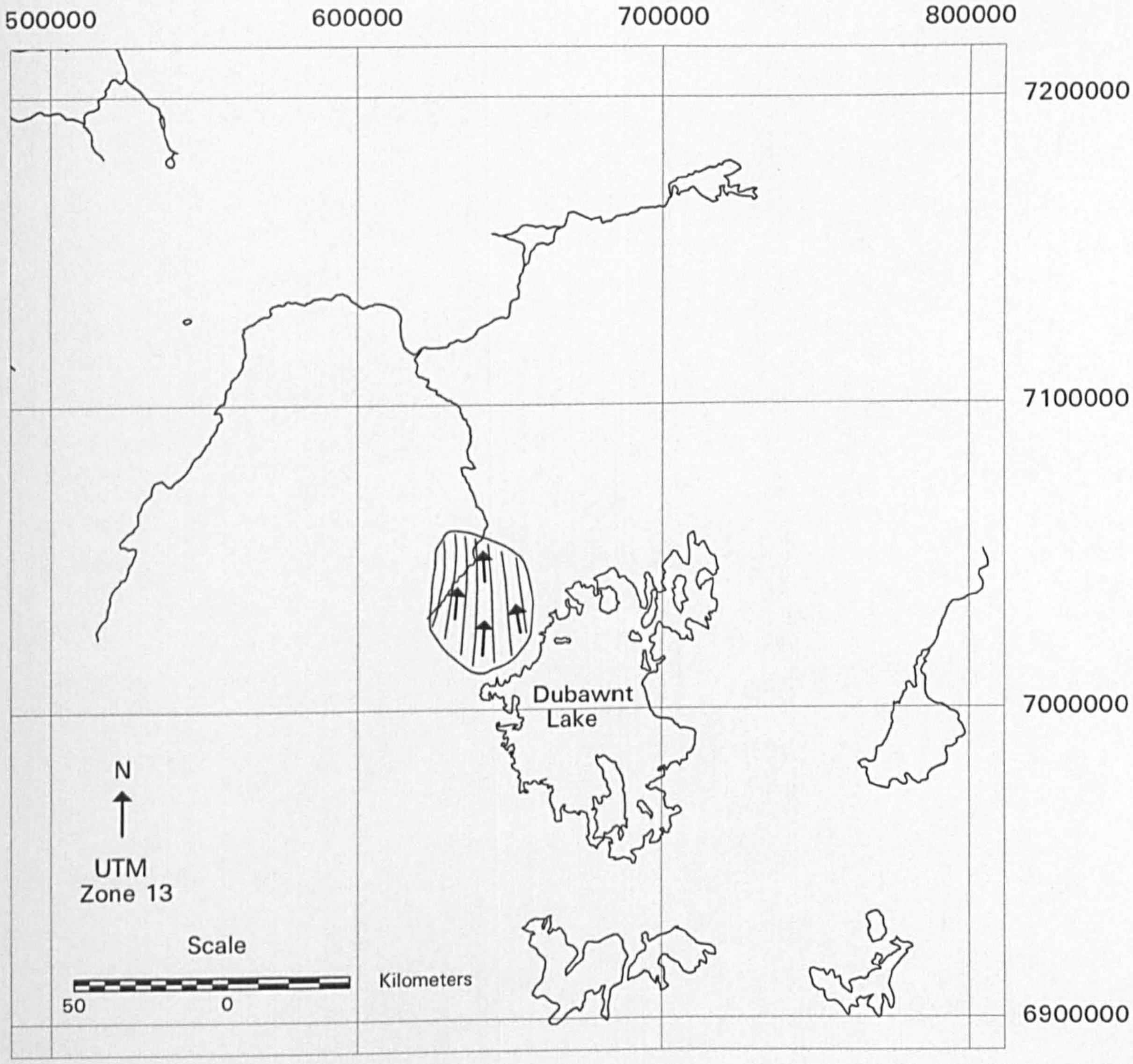

Figure 7.8g. Areal extent of Flow-set 7.

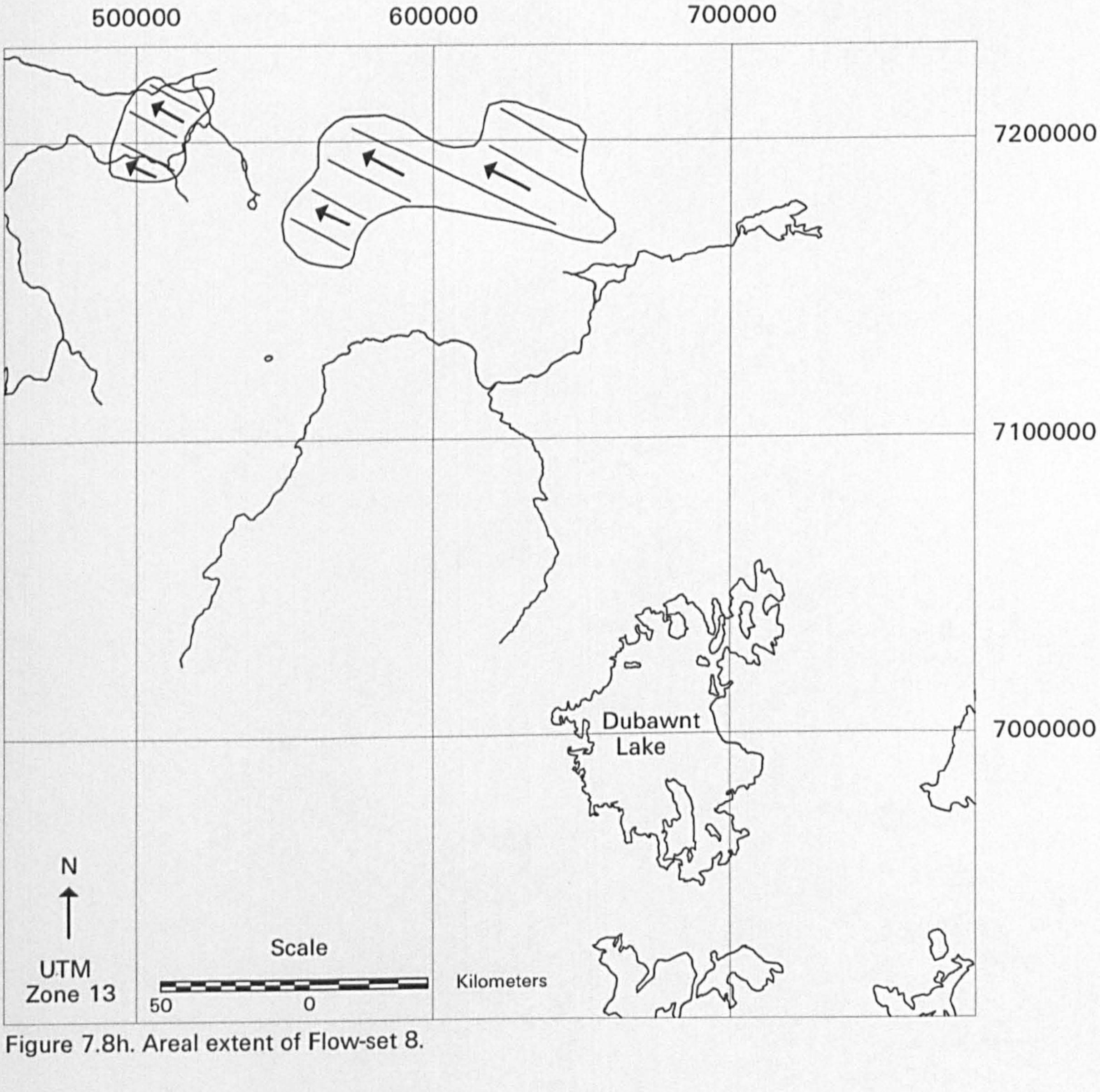

Figure 7.8h. Areal extent of Flow-set 8.

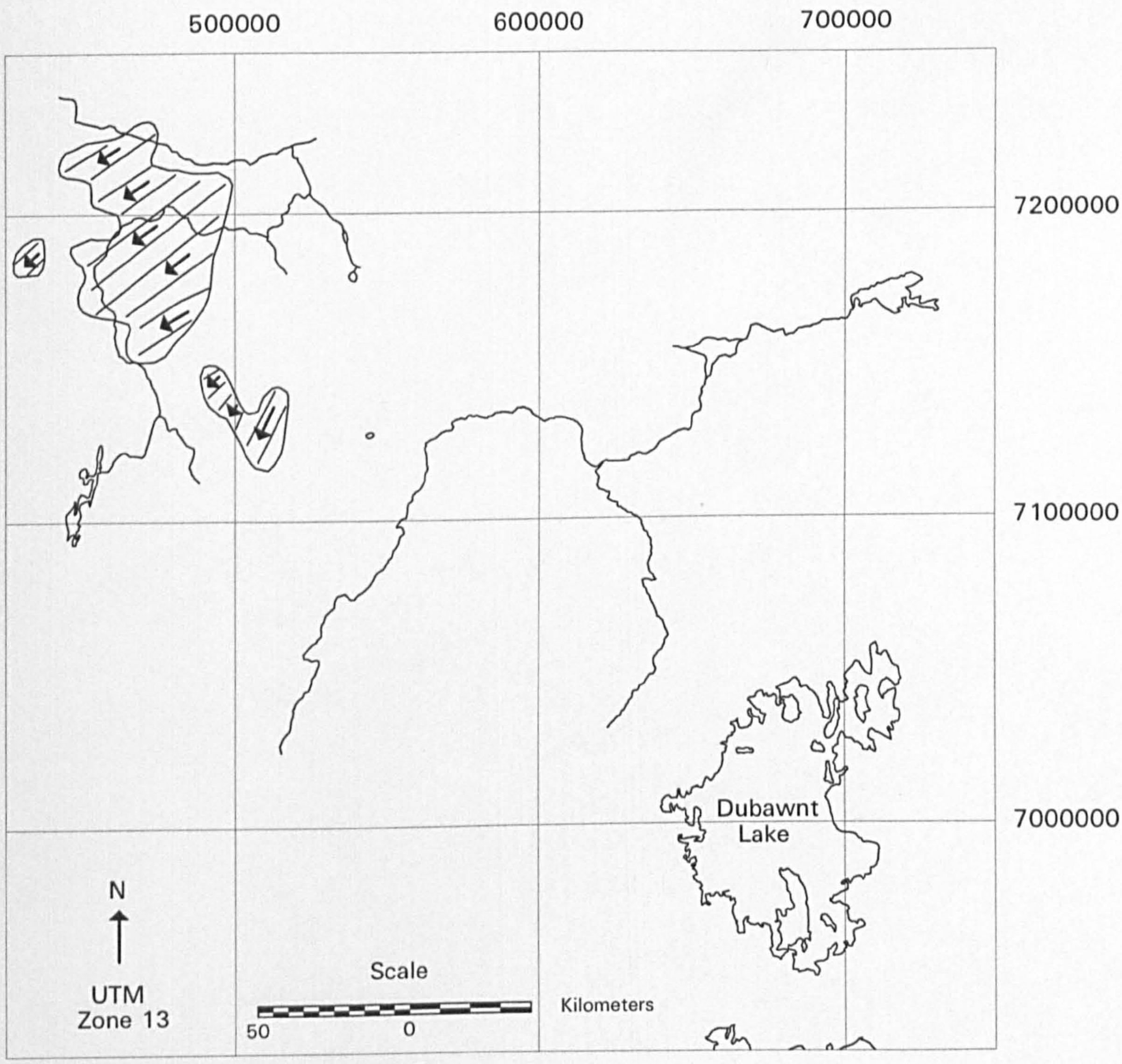

Figure 7.8i. Areal extent of Flow-set 9.

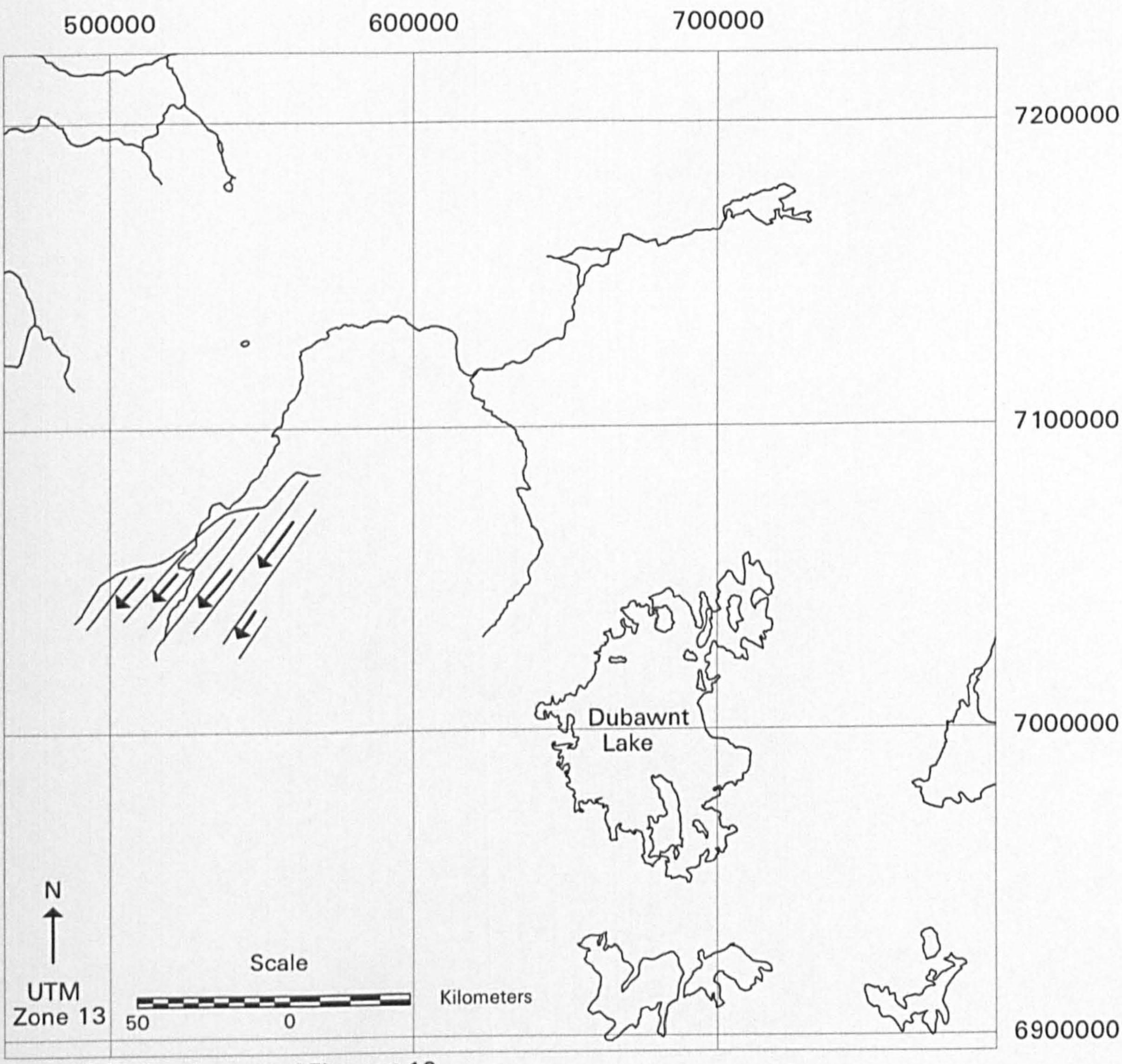

Figure 7.8j. Areal extent of Flow-set 10.

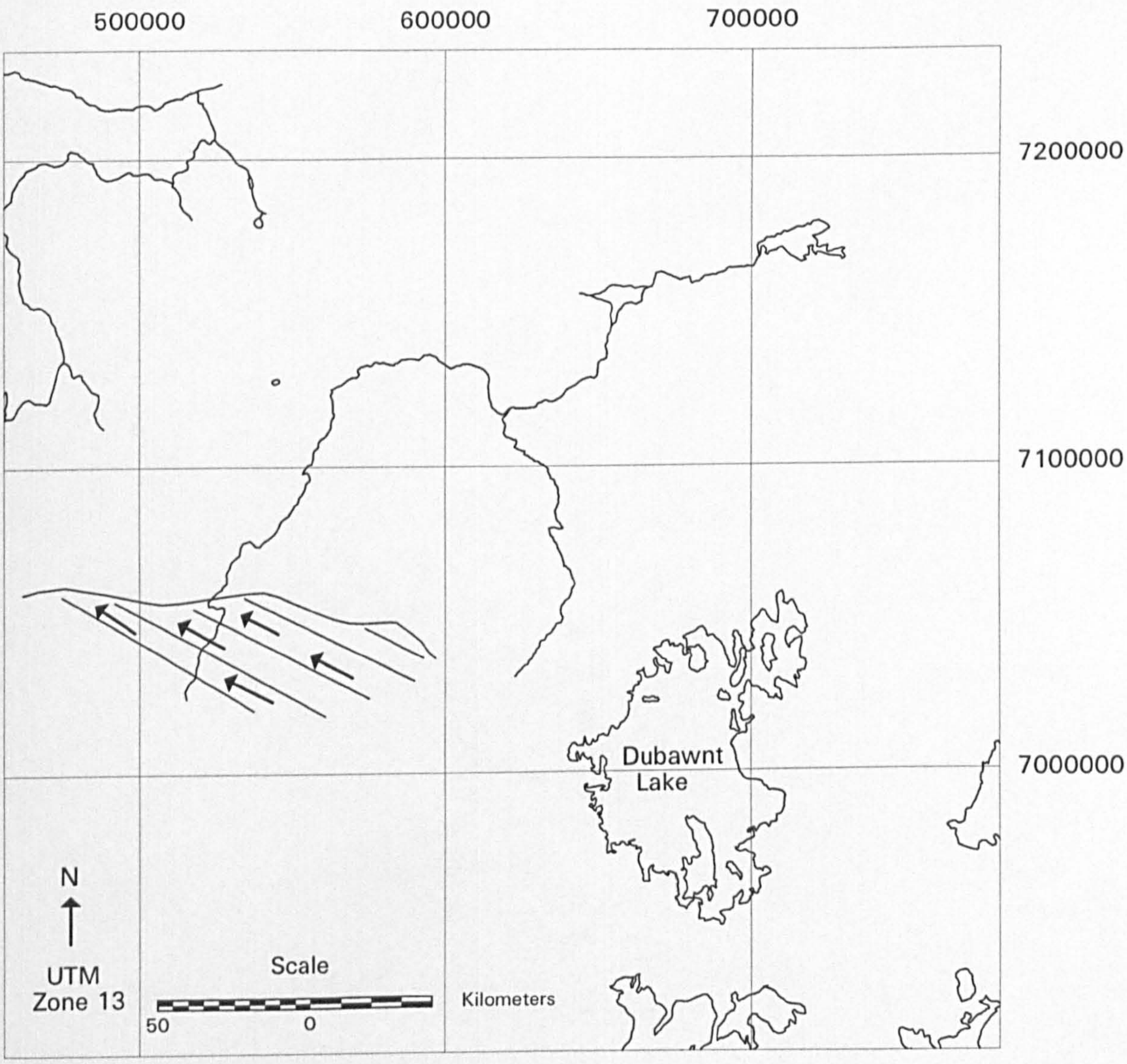

Figure 7.8k. Areal extent of Flow-set 11.

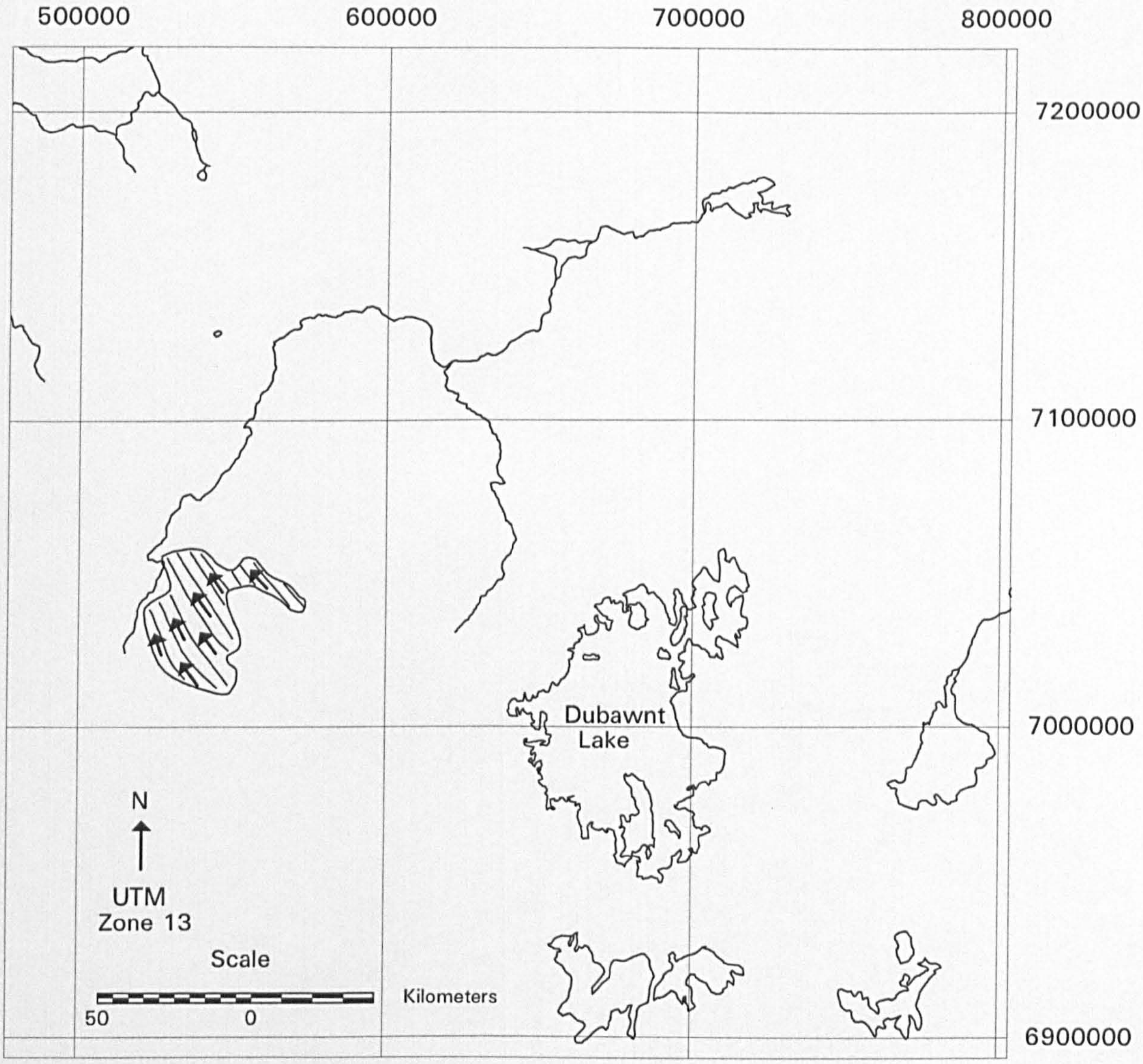

Figure 7.8l. Areal extent of Flow-set 12.

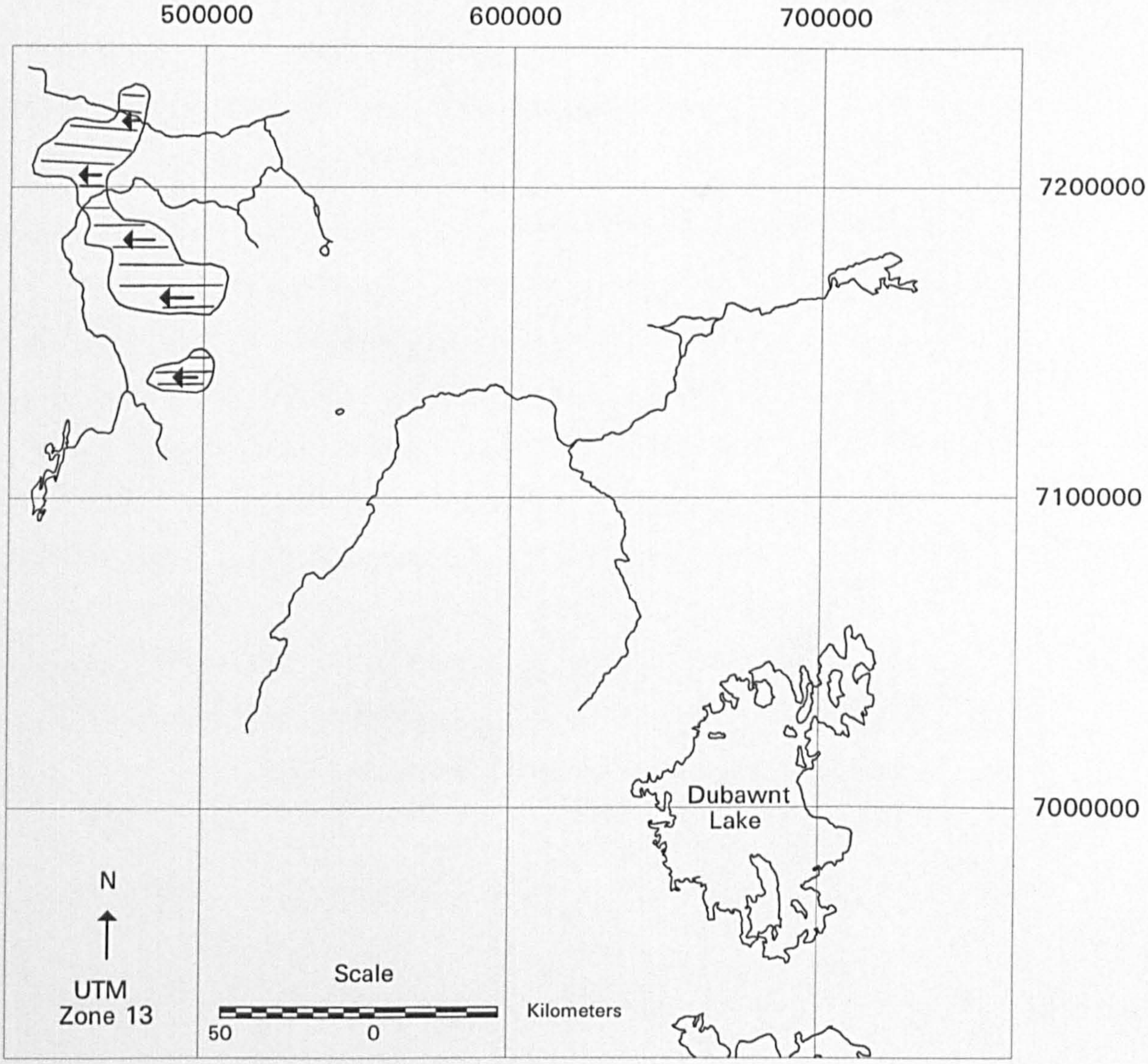

Figure 7.8m. Areal extent of Flow-set 13.

The Dubawnt Lake flow-set (Figure 7.7) contained 8,856 lineations, representing 62% of all those identified. The other flow-sets relate to the broad regional patterns identified on the hard copy imagery and shown in Figure 7.5. Flow-sets 1, 9 and 10 relate to the oldest south-westerly ice flow. These have been cross-cut (in places) by Flow-sets 4, 11 and 12 which relate to the later westerly and north-westerly ice flows. Most of the other flow-sets (2, 5, 6, 8 and 13) are similar in orientation to the Dubawnt Lake flow-set but have been cross-cut. They reflect a westerly-north-westerly ice flow prior to the Dubawnt Lake flow-set. It is difficult to accommodate Flow-sets 3 and 7 into the previous interpretations of flow patterns in the area. Their significance is discussed in Section 7.5.2.

The mapping and flow-set identification were compared with the Geological Survey of Canada's glacial features map of the area (Map 24-1987; Aylsworth and Shilts 1989a). Most of the flow-sets matched well, although the digital mapping consistently picked out more lineaments. Indeed, the Geological Survey of Canada's map represents somewhat of an oversimplification of the flow patterns, often grouping distinct flow-sets into the same flow pattern. An example of this can be seen in the northern part of the study area, near Beverly Lake (Figure 7.4). The digital mapping clearly revealed a cross-cutting relationship between Flow-set 6 (Figure 7.8F) and the Dubawnt Lake flow-set and detailed analysis reveals that the Dubawnt Lake flow-set is superimposed, thus representing the younger flow-set. However, the Geological Survey of Canada's map appears to erroneously groups both flow-sets into the same pattern.

Ribbed moraine are ubiquitous throughout the study area and have been mapped in detail by Aylsworth and Shilts (1989a; b). These 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. In most places, the ribbed moraine are superimposed upon lineaments and occasionally, drumlins have been broken up into individual ribbed moraine (cf. Aylsworth and Shilts, 1989a). Their dimensions and the irregular crest height hampered their identification on the imagery. However, large areas of ribbed moraine fields could be identified and their spatial relationship with the lineations was noted. Because of the difficulty in mapping the ribbed moraine and because detailed maps of their occurrence are readily available, they were not analysed in detail. Nevertheless, their occurrence is a major component of the glacial geomorphology of the study area and their significance is discussed in Section 7.5.4.

7.4.2. Morphometric Bedform Characteristics of Flow-sets.

Table 7.1 shows the basic morphometric characteristics of all the flow-sets identified and shown in Figures 7.7 and 7.8. It can be seen that the Dubawnt Lake flow-set comprises the most lineaments in the area. Maximum values in Table 7.1 show that it contains the longest and widest bedforms and also includes bedforms with the highest elongation ratio.

Table 7.1. Simple dimensions and orientations of all flow-sets mapped on the digital imagery.

Flow-set	n	Mean orientation degrees	Mean length (max.) metres	Mean width (max.) metres	Mean l:w ratio (max.)
F1 (Figure 7.8A)	915	ca. 229 (SW)	579 (1491)	ca. <100	no data
F2 (Figure 7.8B)	38	335 (NNW)	1093 (3474)	217 (596)	5 (6.9)
F3 (Figure 7.8C)	140	332 (NNW)	1649 (4762)	274 (599)	6 (9.3)
F4 (Figure 7.8D)	812	ca. 272 (W)	424 (1387)	ca. >100	no data
F5 (Figure 7.8E)	45	297 (NWW)	996 (1628)	202 (348)	5 (7.4)
F6 (Figure 7.8F)	232	336 (NNW)	1674 (4497)	331 (839)	5 (8.2)
F7 (Figure 7.8G)	120	3 (N)	**3037** (9119)	**402** (897)	**7.7** (14.4)
F8 (Figure 7.8H)	59	291 (NWW)	1506 (3618)	240 (348)	6.5 (12.7)
F9 (Figure 7.8I)	142	233 (SW)	663 (1296)	173 (313)	4 (7)
F10 (Figure 7.8J)	138	215 (SSW)	1067 (2194)	221 (499)	4.9 (7.4)
F11 (Figure 7.8K)	102	291 (NWW)	860 (1779)	135 (283)	6.5 (9.9)
F12 (Figure 7.8L)	131	322 (NW)	915 (1876)	171 (352)	5.3 (7.7)
F13 (Figure 7.8M)	95	269 (W)	606 (1254)	124 (267)	5 (8.1)
Dubawnt Lake (Fig's 7.6 & 7.7)	**5513***	301 (NWW)	1808 (**12,743**)	256 (**990**)	6.8 (**48.3**)

* 5513 lineaments were measured out of a total of 8856.

The highest mean values are not necessarily represented by this flow-set because it includes a large proportion of smaller drumlins up-ice and down-ice of the longest bedforms, see Figure 7.7. If the Dubawnt Lake flow-set was produced by an ice stream it represents the complete geomorphological record of a terrestrially terminating ice stream and could be one of the largest and most complete ice stream signatures available for scrutiny.

7.4.3. Evidence for Ice Stream Activity.

In this thesis, several geomorphological criteria have been identified to aid the identification of palaeo-ice streams (Chapter 4; Stokes and Clark, 1999). The eight criteria are shown in Table 4.1 (Chapter 4). If the Dubawnt Lake flow-set was produced by an ice stream it would be expected to exhibit several of these criteria. Evidence is presented to show that the Dubawnt Lake flow-set fulfils at least four out of a possible seven criteria.

7.4.3.1. Characteristic Shape and Dimensions.

In Section 4.4.1 it is argued that the most obvious clue to the existence of a palaeo-ice stream is its shape and dimensions. Ice streams are large features, characteristically greater than 20 km wide and 150 km long (see Table 2.1). The shape and dimensions of the Dubawnt Lake flow-set is shown in Figure's 7.5 and 7.7. The maximum length of this flow-set is 450 km and the width varies from 140 km at its narrowest point to around 190 km in the lobate terminus. Clearly this distinct flow pattern is large enough to be considered an ice stream.

7.4.3.2. Highly Convergent Flow Patterns.

Another clue to the past existence of an ice stream are highly convergent flow patterns (Section 4.4.2). In contemporary ice streams the onset area is characterised by a large zone of flow convergence (see Figure 4.1). It can be seen from Figures 7.5 and 7.7 that the Dubawnt Lake flow-set displays a large zone of convergent ice flow in the upstream portion of the flow-set which feeds a narrower zone. It is suggested that this convergent flow pattern represents the onset zone of an ice stream.

7.4.3.3. Highly Attenuated Bedforms.

In Section 4.4.3 it is argued that highly attenuated bedforms may be a manifestation of fast ice flow and that elongated drumlins and megalineations may record the flow direction and spatial extent of palaeo-ice streams. It can be seen from Table 7.1 that the Dubawnt Lake flow-set contains remarkably long bedforms with exceptional elongation ratios. The maximum length of these bedforms reaches over 12 km with elongation ratios as high as 48:1. Figure 7.9 shows a colour composite Landsat MSS image of a sample of ice stream bedforms from the centre of the ice stream. The high attenuation and exceptional parallel conformity gives the appearance of a ridge/groove structure.

If we take only the bedforms from the main trunk of the flow-set between 100 and 300 km downstream (i.e. excluding the shorter bedforms from the onset and terminal zones) and plot mean elongation ratio against mean length for all of the flow-sets it can be seen that the Dubawnt Lake flow-set is a distinct cluster, see Figure 7.10. Furthermore, the longest bedforms are longer than those discussed in Chapter 6 and Clark and Stokes (in press) relating to the M'Clintock Channel Ice Stream. It is thus suggested that the bedform dimensions of the Dubawnt Lake flow-set are a valid indicator of fast ice flow and provide further evidence for the existence of an ice stream.

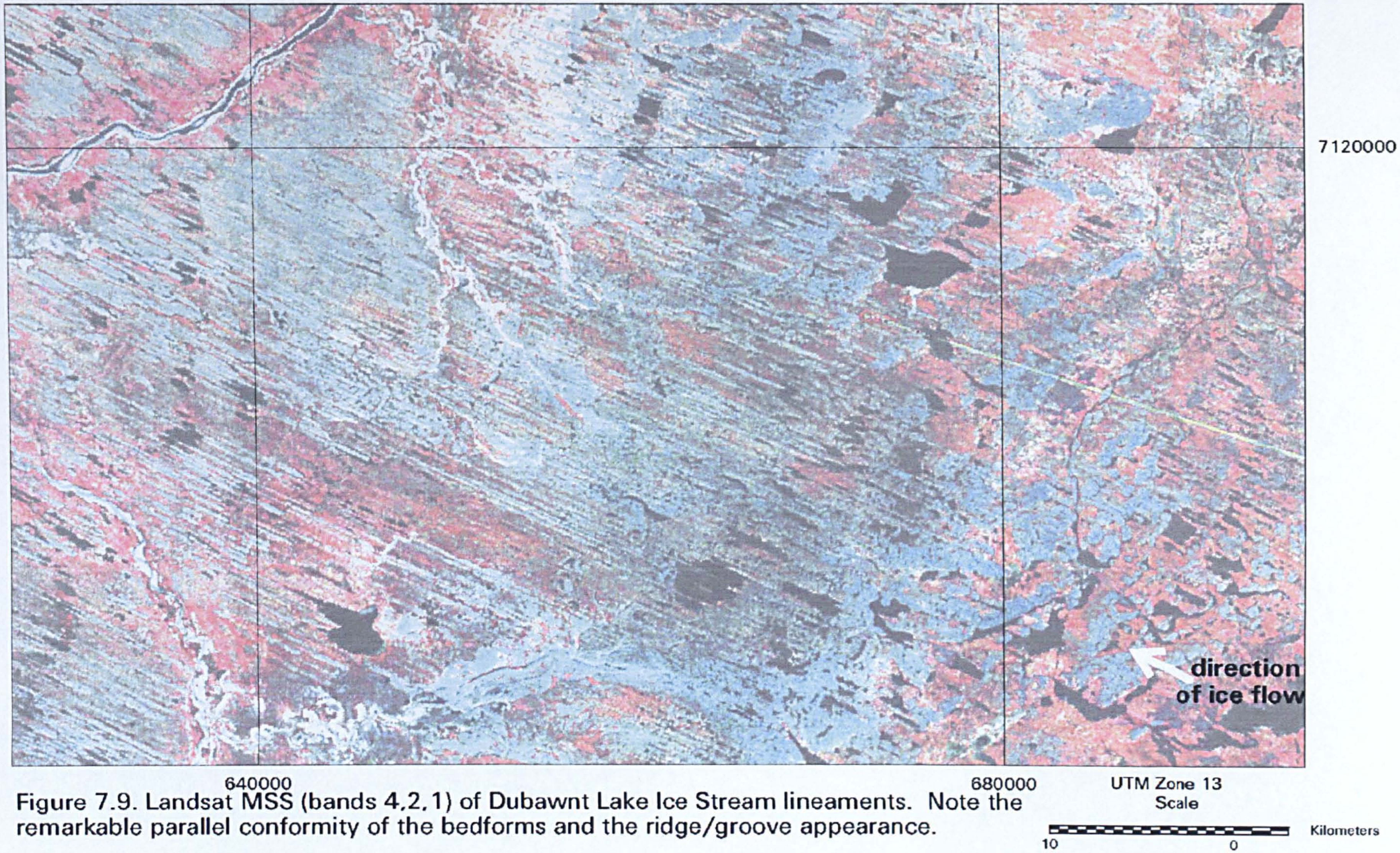

Figure 7.9. Landsat MSS (bands 4,2,1) of Dubawnt Lake Ice Stream lineaments. Note the remarkable parallel conformity of the bedforms and the ridge/groove appearance.

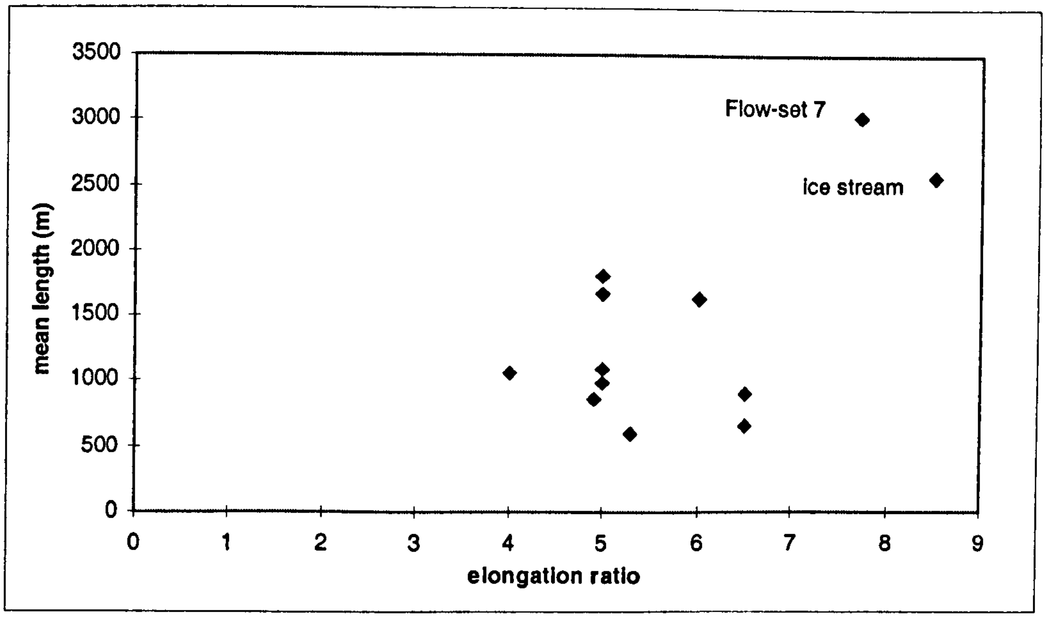

Figure 7.10. Plot of mean elongation ratio against mean length for all of the flow-sets identified on the digital imagery. The significance of Flow-set 7 is discussed in Section 7.5.2.

7.4.3.4. Abrupt Lateral Margin.

Contemporary ice streams are characterised by abrupt lateral margins where the fast flowing ice meets the neighbouring slow ice (Section 2.3.1). In Section 4.4.5 it is argued that the geomorphology of an ice stream should reflect this and that there may be an abrupt zonation of landforms at the margin. This pattern is reflected in the landsystems models shown in Figures 4.4 and 4.5. It can be seen in Figure 7.7 that the Dubawnt Lake flow-set displays a remarkably abrupt southern margin which can be traced along a distance of over 450 km from the onset zone to the terminus. This margin is bordered by the Dubawnt Lake flow-set on one side but within a distance of less than 1 km (and even less in places) there are no bedforms outside the margin. It is suggested that the abrupt southern margin to the bedform pattern represents the margin of an ice stream.

The northern margin of the ice stream is not covered by the digital imagery. However, on the hard copy MSS photographs the margin can be traced for over 300 km towards the terminus. At this point, it overprints two flow-sets (6 and 8) of a similar orientation and it becomes almost impossible to discern the Dubawnt Lake flow-set from the older flow-sets beneath (see Figure 7.5). This partly explains why the Geological Survey of Canada Map 24-1987 (Aylsworth and Shilts, 1989a) depicted this pattern as a single flow-set.

7.4.3.5. Summary and Specific Aims.

Of the eight criteria outlined in Chapter 4, the Dubawnt Lake flow-set fulfils at least four. It has the characteristic shape and dimensions, displays highly convergent flow patterns, contains highly attenuated drumlins and mega-lineations, and displays an extremely abrupt lateral margin. There appears to be no evidence of ice stream marginal moraines nor Boothia-type dispersal trains, and evidence of pervasively deformed sediment is not associated with this part of the ice sheet bed. Because this ice stream terminated on land, the presence of an offshore sediment accumulation fan is not applicable.

It is argued that there is ample evidence to invoke ice stream activity and it is suggested that the Dubawnt Lake flow-set was formed by a terrestrially terminating ice stream. This flow-set will from here on be termed the 'Dubawnt Lake Ice Stream' and represents the complete bedform record of a terrestrial ice stream. Examining the within-stream variations in the morphometry of these bedforms and the characteristics of the ice stream bed may provide fresh insights regarding its operation and functioning and the following pertinent questions will be addressed;

1. What is the nature of the ice stream bed in terms of geomorphology, soft-hard bed fraction, lithology, roughness, and slope?
2. Do the characteristics of the subglacial bedforms reveal anything of ice stream function?
3. Can anything be inferred about the conditions and processes that promoted fast flow?
4. What controlled its overall position in the ice sheet and the location of its margins?
5. For how long did it operate and when?
6. What controlled its activation and de-activation?

7. What effect did the ice stream have on Laurentide Ice Sheet behaviour?

7.4.4. Ice Stream Extent.

The ice stream is reconstructed as having a maximum length of around 450 km. In the onset zone the ice stream approaches widths of 305 km and this narrows into the main ice stream trunk which has a fairly constant width of around 140 km, see Figure 7.7. In the terminal zone the ice stream produced a lobate margin whose width broadened to around 190 km. The surface area of the ice stream is estimated at around 72,000 km^2 of which 53,000 km^2 is covered by the digital imagery (approximately 74 % of the ice stream bed is mapped in detail).

7.4.5. Isochronous or Times-transgressive Bedform Pattern?

Figure 7.7 shows the ice stream lineaments mapped from the MSS digital imagery. It can be seen from this pattern that the ice stream bedforms can be demarcated into three broad zones; a highly convergent onset zone, a main ice stream channel or trunk, and a splayed lobate terminal zone. Despite the convergence and divergence of this flow pattern, the ice stream lineaments depict a remarkably coherent pattern whereby neighbouring bedforms display a high degree of parallel conformity. For the central portion of the ice stream over an area of 720 km^2 (in the main trunk), the standard deviation of lineament orientation does not exceed 3.8° (see also Figure 7.9).

If the flow pattern were produced times-transgressively it would be expected to display systematic discontinuities in the coherency of the flow pattern. It is clear from Figure 7.7 that such discontinuities are not apparent. Furthermore, it is difficult to explain the abrupt southern margin of the ice stream if it were formed time-transgressively. In addition, the ice stream flow-set fulfils all of the criteria postulated by Clark (1999) as being indicative of an isochronous bedform pattern (outlined in Section 6.4.2). It is thus argued that the evidence presented here (and later) suggests that the Dubawnt Lake Ice Stream fulfils all of these criteria and that it represents a snapshot view of the bed prior to ice stream shut-down.

7.4.6. Variations in Ice Stream Bedform Morphometry.

Analysing the within stream variation in bedform morphometry may provide information regarding the functioning of the Dubawnt Lake Ice Stream, particularly ice velocity. It can be seen from Table 7.1 and Figures 7.9 and 7.10 that the ice stream bedforms display exceptionally high lengths and elongation ratios. Traditionally, it has been assumed that high bedform elongation ratios are indicative of fast ice flow (see Section 4.4.3). If this is the case, what do the downstream variations in lineament morphometry reveal about flow velocities within the Dubawnt Lake Ice Stream?

Figure 7.11 shows the downstream variation in mean length for each of the three flow-bands 'A' to 'C' (shown in Figure 7.6). A distinctive pattern reveals that bedforms are longest approximately half way down the ice stream between 200 and 220 km. The lineations increase from an average length of around 750 m in the onset zone to a maximum of up to 4,000 m in the main trunk of the ice stream, after which they steadily decrease in length towards the terminus.

Downstream variations in lineament width are shown in Figure 7.12. It can be seen that the maximum widths of the bedforms do not coincide with the maximum lengths. Rather, maximum widths peak between 100 and 160 km. This indicates that the longest bedforms are not necessarily the widest bedforms.

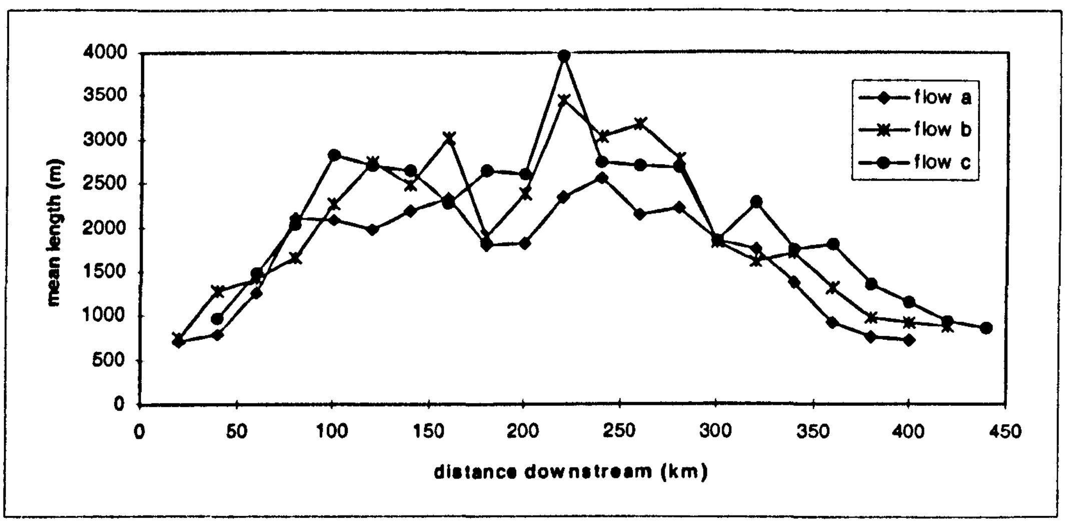

Figure 7.11. Downstream variation in lineament length within the Dubawnt Lake Ice Stream, see Figure 7.6 for flow-band locations.

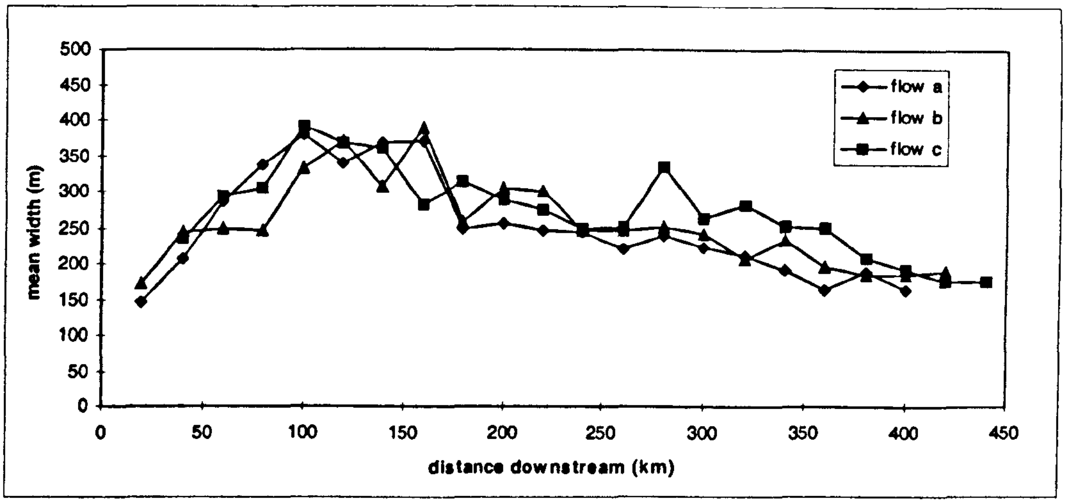

Figure 7.12. Downstream variation in lineament width within the Dubawnt Lake Ice Stream, see Figure 7.6 for flow-band locations.

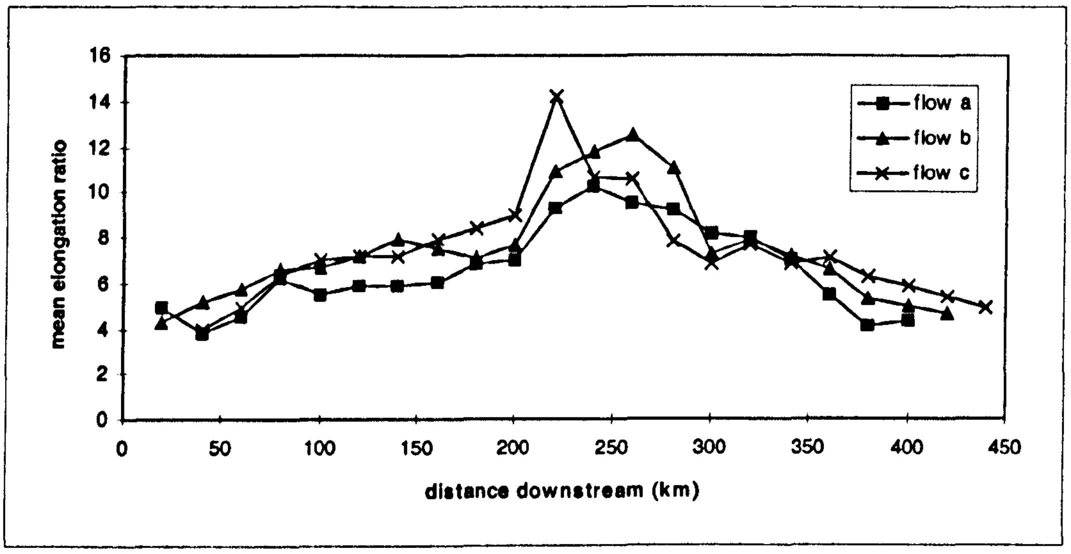

Figure 7.13. Downstream variation in lineament elongation ratio within the Dubawnt Lake Ice Stream, see Figure 7.6 for flow-band locations.

Instead, what is observed is a sharp increase in streamlining in a downstream direction. This is confirmed by plotting the downstream variation in elongation ratio for each of the three flow-bands, see Figure 7.13. Elongation ratio steadily increases and then jumps to a peak in the main ice stream channel (200 km downstream) before decreasing steadily towards the terminus.

What implications do these downstream trends have for ice stream velocity? It is assumed that these patterns reflect the basal flow velocities of the ice stream and that bedform morphometry, specifically elongation ratio, are a useful proxy for ice velocity (see discussion in Section 4.4.3). Unfortunately, we have no contemporary terrestrial ice streams on which to measure velocity and test these assumptions. However, the bedform pattern reveals exactly what we would expect from a terrestrial ice stream whose velocity increases in the onset zone, passes through a maximum in the main channel and slows down as the ice diverges towards the terminus.

7.4.7. Ice Stream Bed Elevation.

What is particularly interesting about the downstream variation in elongation ratio is the sharp increase between 200 and 220 km for each of the three flow-bands (see Figure 7.13). Figure 7.14 shows the downstream changes in bed elevation within the ice stream. Three transects in Figure 7.14 (excluding the margin transect) show a drop in elevation somewhere between 150 and 200 km downstream, immediately prior to the abrupt increase in elongation ratios. This drop occurs as the ice stream enters the Thelon Basin, which represents the lowest elevations of the ice stream bed. Thereafter, bed elevation increases into the terminal zone.

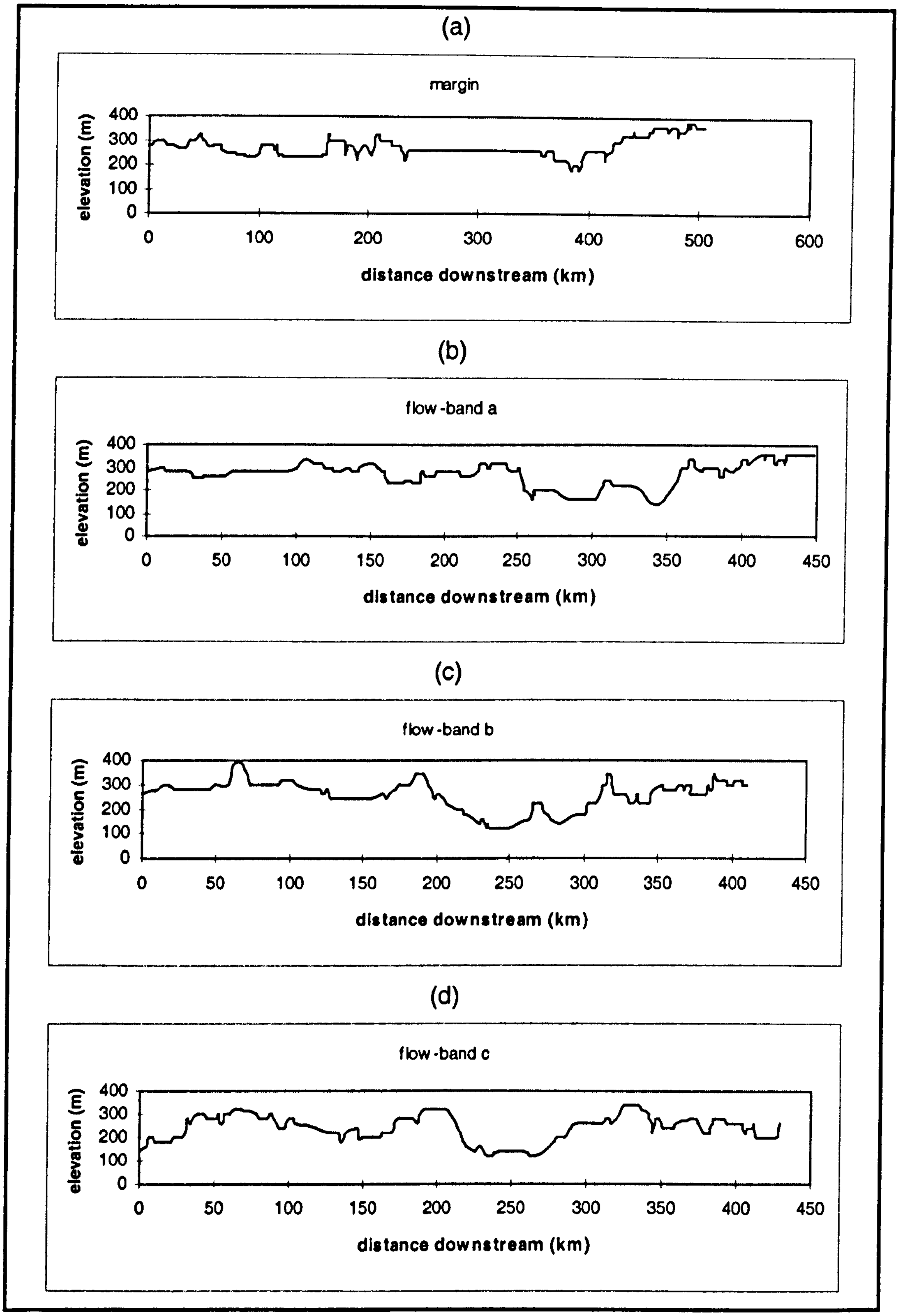

Figure 7.14. Surface elevation transects down the ice stream, (a) along the southern margin, (b, c and d) along three flow-bands (vertical exaggeration ca. x165). See Figure 7.6 for flow-band locations.

7.4.8. Ice Stream Sediment Characteristics.

The Canadian Shield is characterised by granitic gneiss, interrupted in places by sedimentary and volcanic strata (Aylsworth and Shilts, 1989a). In general, this geology can be described as 'hard rock' and is largely resistant to glacial erosion. However, in the study area (and beneath the inferred ice stream) there are exceptions, and large areas yielded abundant glacial debris which almost certainly played a role in influencing glacier dynamics (Aylsworth and Shilts, 1989a).

One such area, the Thelon Sedimentary Basin, was particularly susceptible to glacial erosion and produced relatively large volumes of sediment eroded from poorly consolidated sandstone outcrops. The location of the sedimentary basin is shown in Figure 7.15 with the ice stream extent superimposed.

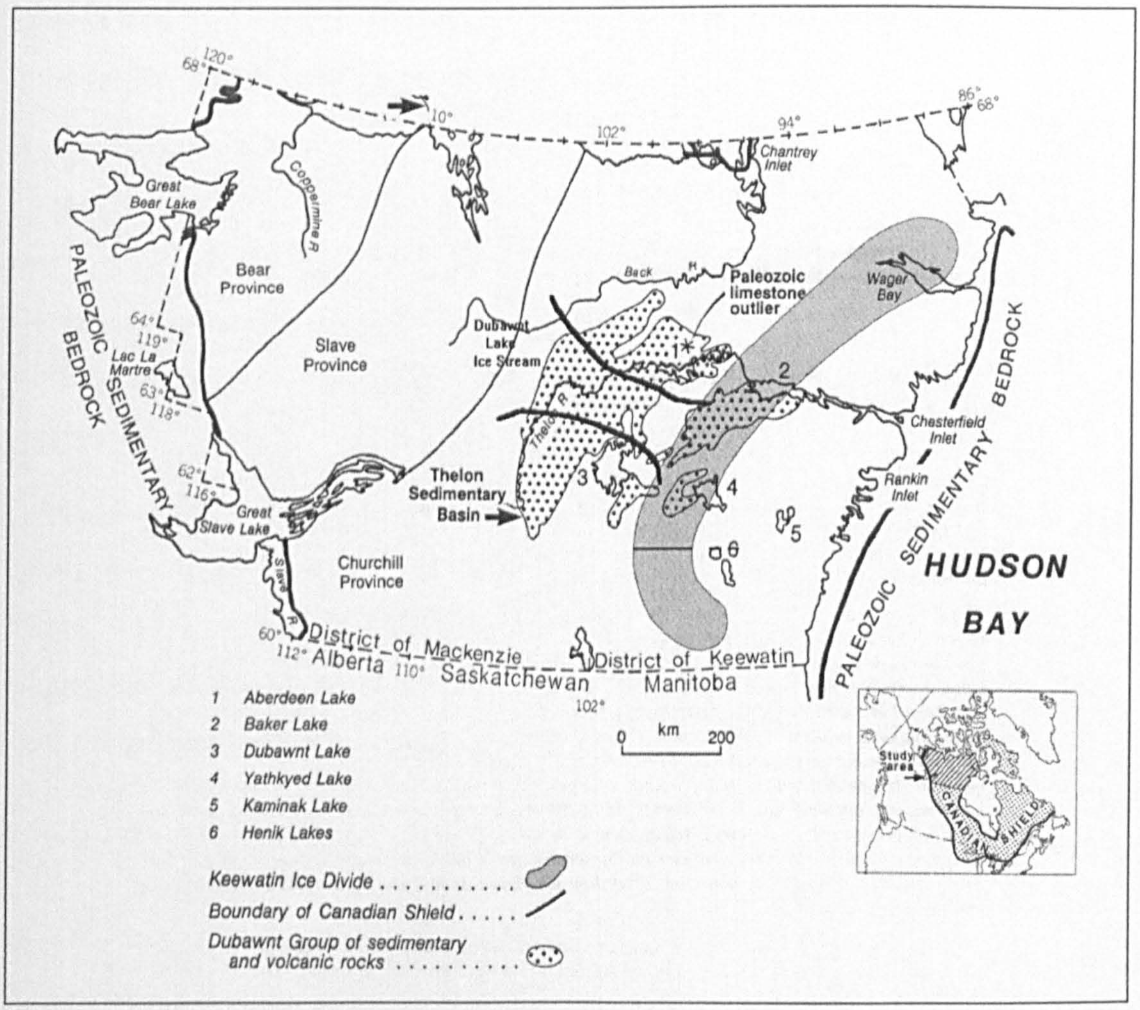

Figure 7.15. The location of the Thelon Sedimentary Basin (stippled) in relation to the Dubawnt Lake Ice Stream (adapted from Aylsworth and Shilts, 1989a).

The Thelon sedimentary basin can be seen in the centre of the ice stream beginning at around 220 km and stretching for 100 km downstream. This basin coincides with generally lower elevations (see Figure 7.14) but also spans the higher ground outside of the ice stream margin to the south.

7.4.9. Regional Topography.

To ascertain whether the Dubawnt Lake Ice Stream was channelled by the bedrock topography, a 30-arc second (ca. 0.5 km) Digital Elevation Model of the study area was inspected. Figure 7.16 shows the regional topography of the study area with the Dubawnt Lake Ice Stream superimposed with the digital imagery coverage.

The onset zone is characterised by high elevations in the south (up to 250 m) and low elevations (~175 m) in the north, near Baker Lake. Between 180 and 220 km downstream, the southern half of the ice stream traversed a topographic step before entering the Thelon Basin (elevations around 150- 200 m), see Section 7.4.8.

A part of the central trunk of the ice stream between 220 and 300 km downstream borders a topographic high immediately to the south, see Figure 7.16. To examine this relationship in more detail, surface profiles were drawn up perpendicular to the ice stream and starting 50 km outside of the margin. Figure 7.17 shows six of the surface profiles (between 200 and 300 km downstream) with the ice stream margin indicated.

It would appear that parts of the ice stream margin lie inside a topographic step, and could have been channelled by the ridge. However, the vertical exaggeration in Figure 7.17 is extreme (x 40) and when it is considered that a drop in elevation of around 200 m (maximum) occurred over a distance of around 50 km, any topographic effect should be viewed as negligible.

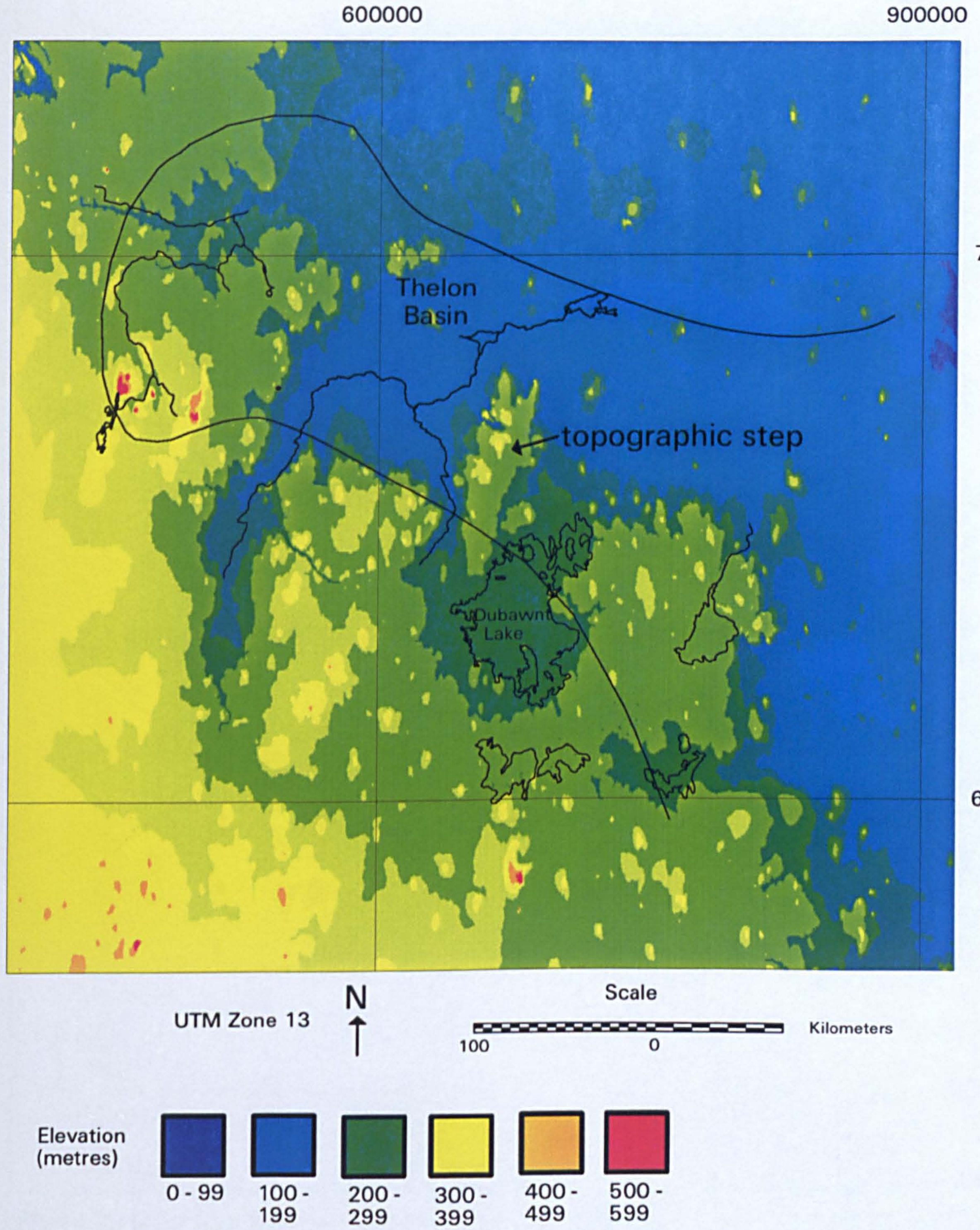

Figure 7.16. Digital Elevation Model showing surface elevations of the Dubawnt Lake Ice Stream (black outline) and surrounding region.

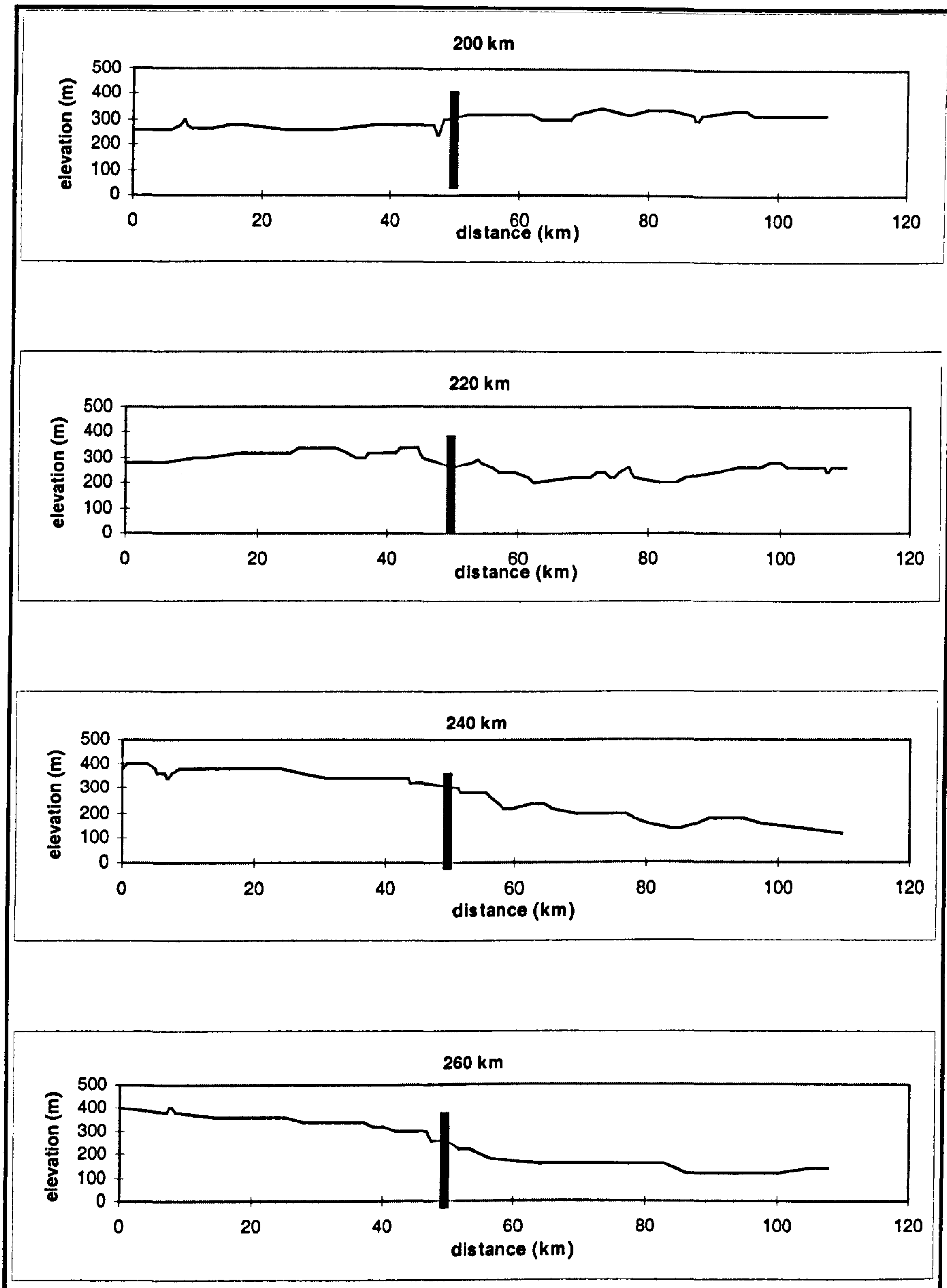

Figure 7.17. Surface elevation changes across the inferred ice stream margin (black line) between 200 and 300 km downstream (vertical exaggeration x40).

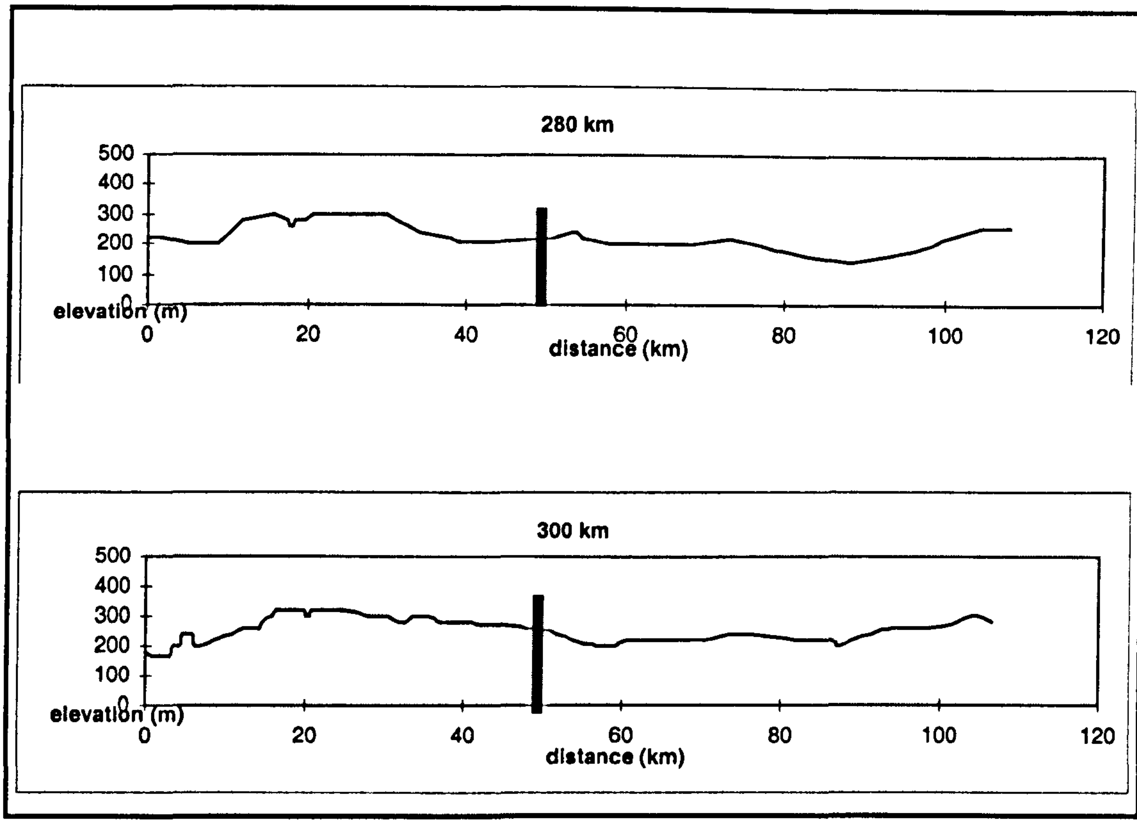

Figure 7.17. continued.

7.5. Discussion.

7.5.1. Controls on Ice Stream Location and Margin Position.

Aylsworth and Shilts (1989a) commented that the most spectacular streamlined drumlins in the Keewatin area occur immediately downstream of sedimentary basins. The results presented here would appear to support this observation. It has been shown that the most attenuated ice stream bedforms occur on, or downstream of a sedimentary basin, compare Figures 7.7 and 7.16. Aylsworth and Shilts (1989a) argued that this spectacular fluting is a result of the sediment availability and is not necessarily related to the dynamic conditions of the ice sheet, i.e. fast ice flow.

It is clear that the availability of softer sediments exerts a major control on bedform generation but it is argued here, that the spectacular fluting is also a result of rapid flow velocities. This inference is supported by the evidence suggesting fast ice flow through the Thelon Basin (i.e. highly convergent flow patterns, numerous highly attenuated bedforms and an abrupt margin to the flow-set). Perhaps more importantly, the margins of the ice stream lie irrespective of the sedimentary basin, which continues for a considerable distance outside of the ice stream. No elongated bedforms are found here despite the availability of sediment. Rather, it is suggested that the soft sediments actually encouraged fast ice flow through the area. This effect was also enhanced by a topographic step approximately half way down the ice stream, see Figure 7.16.

Figure 7.18 shows a composite diagram along flow-band 'B' of the ice stream. It can be seen that the topographic step and the onset of the sedimentary basin coincide with an abrupt increase in lineament elongation ratio. Although this increase in elongation ratio is less marked in flow-bands 'A' and 'C', and the topographic step is less clear, they also show an increase in attenuation between 200 and 220 km (see Figure 7.13).

Topographic steps have been associated with ice streams and outlet glaciers in Antarctica (cf. McIntyre, 1985) and it is logical to assume that flow velocities increased at this point. This is because the drop in elevation would have caused localised extensional ice flow and strain heating. This strain heating would have led to warmer ice, less viscous ice and hence, faster ice flow. This process has been inferred in the onset zones of contemporary ice streams (Section 2.3.4) and ice stream evolution has been successfully modelled by Payne (1998; 1999) based on purely thermomechanical interactions and bed topography (Section 2.5.3).

It is tempting to ascribe the location of this ice stream to the soft sedimentary basin and topographic step. However, it is suggested that the location of the ice stream within this portion of the ice sheet was not triggered by these geological factors. It is difficult to demonstrate that the change in topography (which only extends across half of the ice stream, see Figure 7.16) initiated ice stream flow. Furthermore, this change occurs around half way down the ice stream. Although its effect may have propagated upstream, the isochronous bedform record would appear to suggest that fast ice flow was triggered further inland.

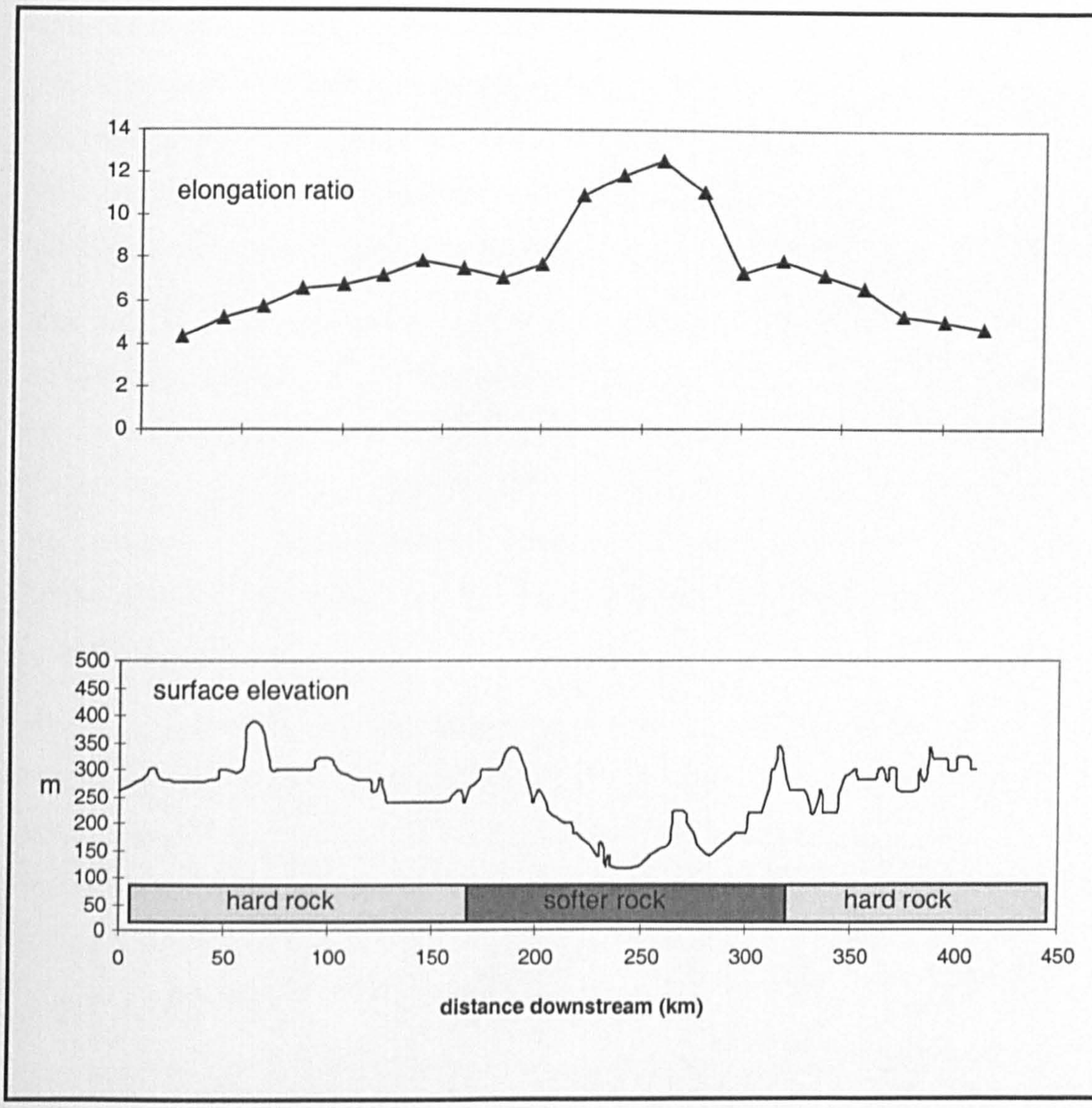

Figure 7.18. Composite diagram showing the relationship between sedimentary characteristics, bed elevation and bedform elongation ratio in the Dubawnt lake Ice Stream. Although the topographic step and softer rocks would have increased ice flow velocities, they did not control the location of the ice stream within the ice sheet.

It should also be remembered that although the sedimentary basin provided a less resistant bed (compared to hard bedrock) for a part of the ice stream, its importance should not be overestimated. This material (sedimentary rocks e.g. sandstone) was too coarse to deform, it simply provided less resistance to basal sliding. It is therefore concluded, that while the topographic step and sedimentary basin encouraged fast ice flow, they were not the trigger for ice streaming.

Because the margin of the sedimentary basin extends considerably beyond the ice stream margin, it is also inferred that the underlying geology had no influence on the exact margin positions of the ice stream. Likewise, there is no consistent or appreciable change in topography at the ice stream margins (Figure 7.17) and it can also be inferred that this was not a topographic ice stream. Rather, it is suggested that the location of the ice stream was forced by climatic warming which altered the ice sheet configuration and the thermal state of the bed. A switch from cold to warm-based conditions probably triggered rapid basal sliding.

This portion of the ice sheet was dominated by the Keewatin Ice Divide positioned immediately upstream of the ice stream onset zone. This ice divide would have provided a considerable catchment area for the ice stream, an essential pre-requisite for ice streaming. Citing evidence from two large dispersal trains emanating away from inferred ice divides on Somerset Island and Boothia Peninsula, Dyke and Dredge (1989) noted that ice streams may be generated by convergent ice flow in the vicinity of an ice divide.

Based on several dated marginal positions from the literature, Dyke and Dredge (1989) reconstructed the retreat of the Keewatin Ice Sheet and Figure 7.19 shows their reconstructed flow patterns, resulting from shifts in the ice divide location between 11,000 and 8,400 yr BP. It can be seen that at 9,000 yr BP, a horseshoe shaped ice divide is associated with a zone of convergence feeding north-westerly flowing ice through the study area.

Unfortunately, this is not independent evidence for triggering an ice stream. It is just as likely that the ice stream produced the convergent ice flow as it is that the convergent ice flow produced the ice stream. This represents a chicken and egg situation; which came first, the ice stream or the convergent flow? Added to this, the accuracy of the ice sheet reconstruction is unknown.

Both Andrews (1989) and Dyke (1984) noted that this period of deglaciation (around 8,500 yr BP) may have been accompanied by a climate that was as warm or warmer than today and led to the development of several proglacial lakes at the ice sheet margin, see Figure 7.19. The occurrence of such lakes, coupled with significant marine inundation, represented newly available sources of moisture and probably led to increased accumulation in the inland portions of the ice sheet (i.e. the Keewatin Ice Divide). At the same time, the warmer temperatures also led to the rapid deglaciation of the ice sheet margin and increased amounts of meltwater at the bed would have encouraged rapid basal sliding under warm-based ice. In addition, increased temperatures may have triggered a switch from a cold- to warm-based ice sheet, thus facilitating rapid basal sliding. The overall result of this, was a relatively steep, warm-based ice sheet which fed proglacial lakes.

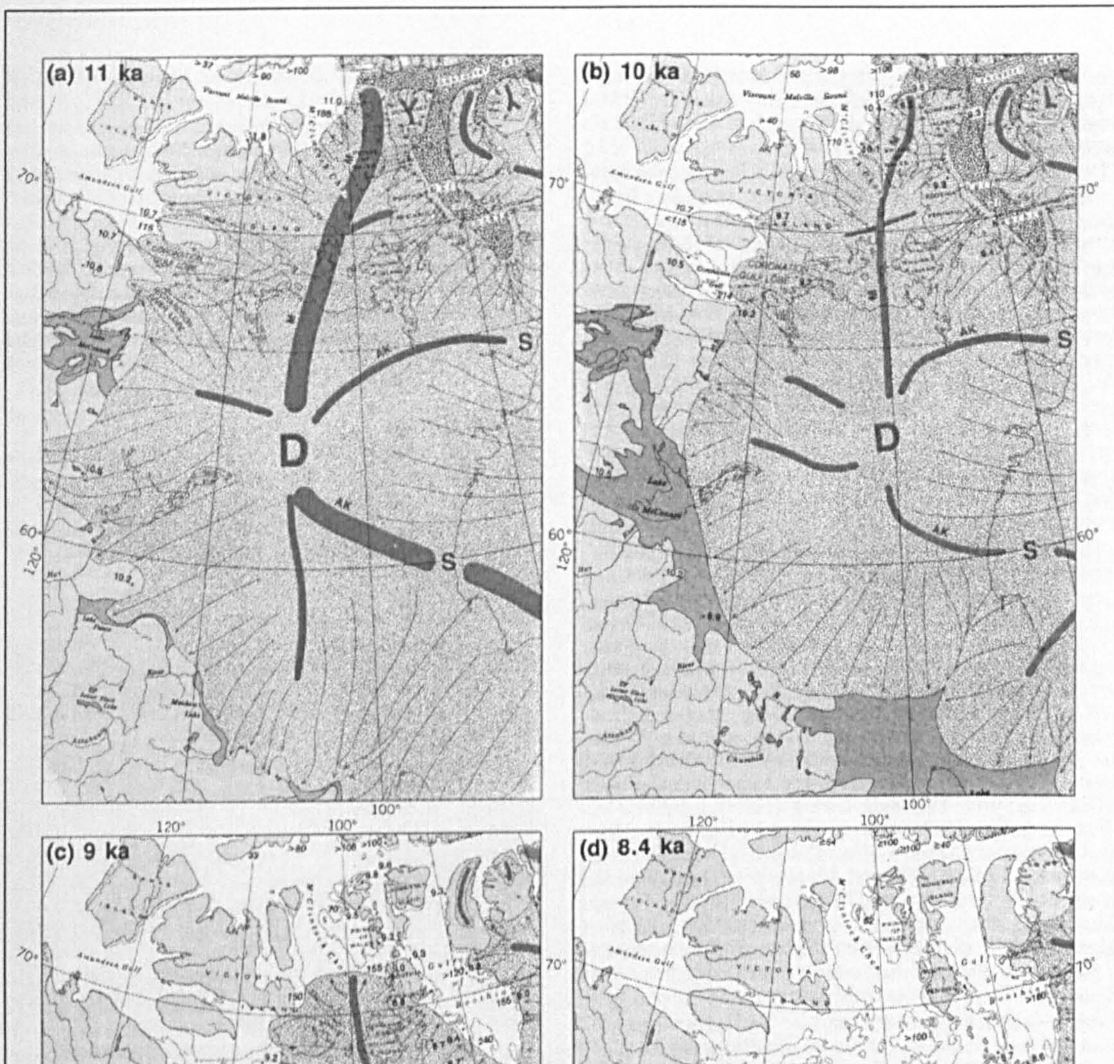

Figure 7.19. Reconstructed ice sheet configuration on the north-western Canadian Shield between 11,000 and 8,400 yr BP (from Dyke and Dredge, 1989).

This represented an unstable ice sheet and it is likely to have been the 'on switch' for the rapid basal sliding of the Dubawnt Lake Ice Stream. In addition, the development of proglacial lakes would have enhanced the removal of ice, thus allowing an ice stream to discharge disproportionate amounts, without the margin advancing. In the same way, although basal sliding may not transport as much sediment downstream compared to a deformable bed (see Chapter 6), small amounts of material that were transported or 'grooved out' would have been effectively drained away by a pro-glacial lake and river system. However, due to the nature of the underlying geology and the inferred flow mechanism (basal sliding), this ice stream would not have transported substantial amounts of sediment.

Between 9,000 and 8,400 yr BP, glacial Lake Thelon is thought to have formed at the ice margin (see Figure 7.19). Evidence for this lake is well documented by Bird (1953), Taylor (1956), Lee (1959) and Aylsworth and Shilts (1985). This lake would have provided an effective environment for removing ice and would presumably have accentuated ice stream drawdown and enhanced flow velocities. Although the lake may have helped trigger ice streaming, its long term influence on ice stream configuration was not substantial. This is because the terminus of the ice stream displays a divergent bedform pattern and this would not be expected if the ice stream continually calved into a lake throughout its life-cycle.

7.5.2. Ice Stream Tributaries.

Referring back to Table 7.1 it can be seen that Flow-set 7 is characterised by the highest values of mean length, width and elongation ratio. The large size and high elongation ratios of these lineations clearly groups them into a similar cluster as the ice stream bedforms and this is indicated in Figure 7.9. The location of Flow-set 7 is shown in Figure 7.8G where it can be seen that it lies very close to the ice stream margin (compare with Figure 7.7). It can also be seen from Figures 7.8C that Flow-set 3 also appears to feed into the main ice stream trunk north-west of Dubawnt Lake. Figure 7.20 shows an MSS digital image of Flow-sets 3 and 7 where they appear to join the main ice stream, and Figure 7.21 is a detailed map of all the lineations in the area. These form a remarkably coherent pattern and it is suggested that Flow-sets 3 and 7 may represent a tributary of the ice stream. This tributary would have been over 50 km long and varied in width between 20 and 30 km.

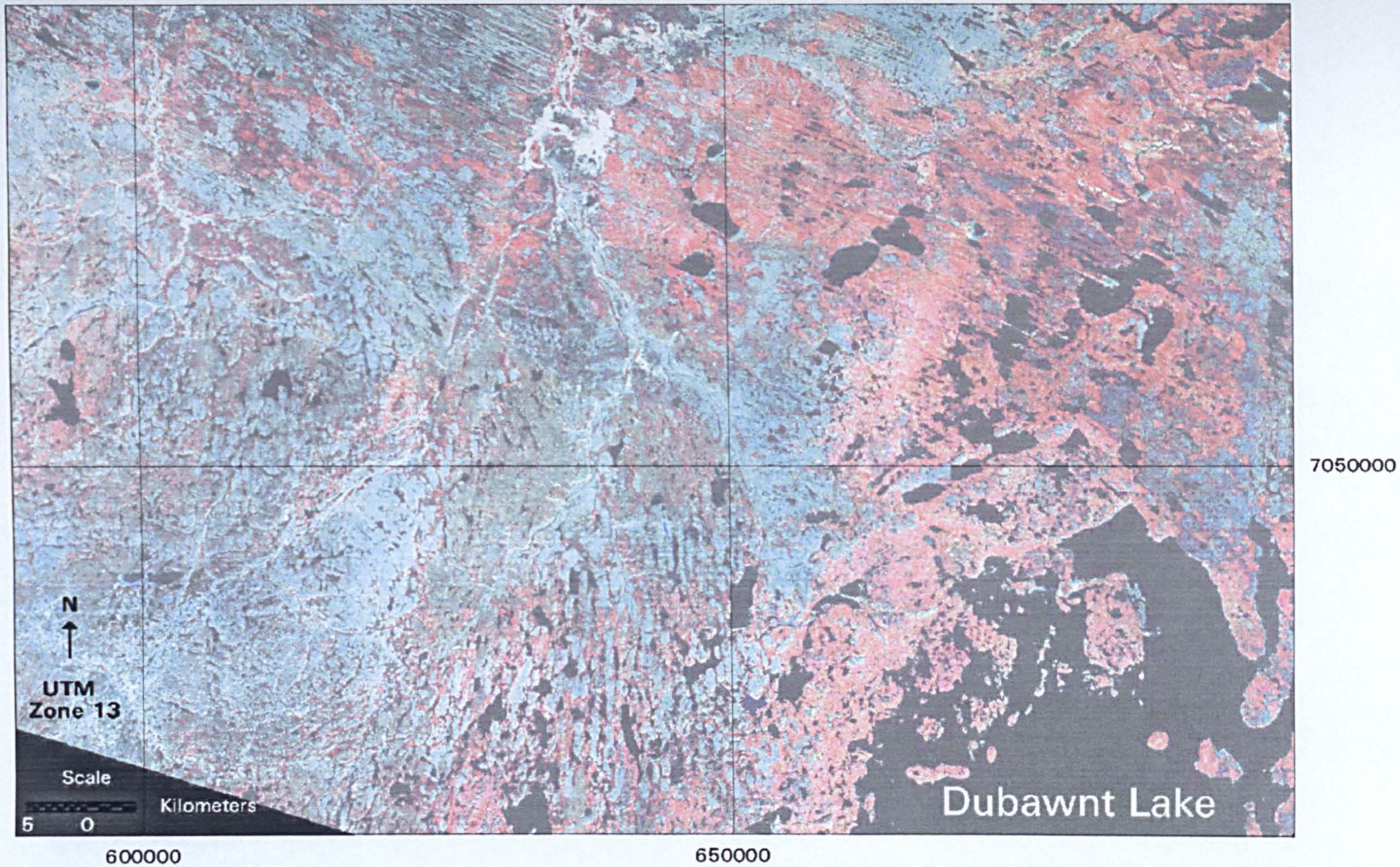

Figure 7.20. Landsat MSS image (bands 4,3,2) of ice stream tributary north west of Dubawnt Lake.

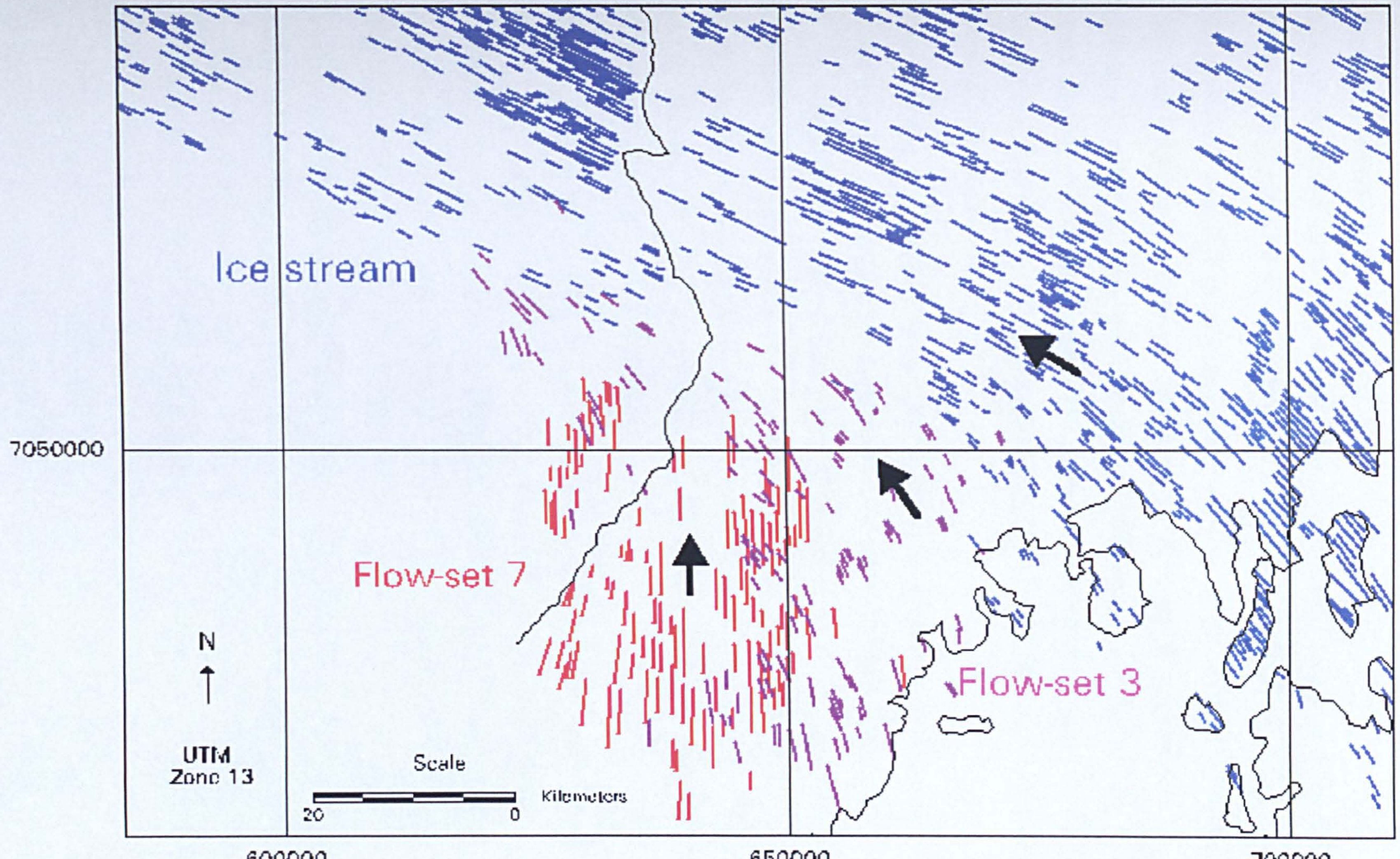

Figure 7.21. Lineation map showing inferred ice stream tributary lineations.

Closer inspection of the digital imagery reveals that the main ice stream trunk is superimposed on Flow-set 7. Therefore, Flow-set 7 is older and ceased to operate prior to the main ice stream shut-down. Furthermore, Flow-set 3 is superimposed upon the Flow-set 7 landforms but again becomes cross-cut by the ice stream. Although the bedforms are not notably long or attenuated, the coherent pattern feeds directly into the main ice stream and is very similar to Flow-set 7.

It is postulated that this pattern represents a switch of the tributary position as its catchment area became reduced. At first, the tributary was broad but narrowed substantially before shutting down altogether. This happened prior to the whole ice stream shutting down and provides further support that the ice stream was dependent on a large supply of ice. When this ice became exhausted, the tributary shut down, followed by the ice stream.

When viewing the regional topography in Figure 7.16 it can be seen that the ice stream tributary (unlike the main ice stream) may have been preferentially channelled by the topography. Typical elevations of around 280 m in the tributary feed downslope into the main ice stream channel (ca. 200 m). Furthermore, the tributary is bordered to the west by a plateau which reaches altitudes of up to 400 m. This can be seen on Figure 7.20 where the inferred tributary landforms border a curved feature which joins to the main ice stream margin.

Analysis of geological maps reveals no bedrock structures in the area and it is assumed that the plateau and the curved margin are of a similar lithology to the undifferentiated surficial materials abundant inside and outside of the ice stream. However, the curved feature is not an ice stream marginal moraine but rather, represents the eroded border of the higher ground to the west of the tributary.

The tributary flow-sets (3 and 7) identified in this study had not been previously recognised as a distinct flow pattern. Most of the research either failed to map these flow-sets, or grouped them together with the widespread and yet sporadic south-westerly ice flow patterns which occur further west (e.g. Bird, 1953; Aylsworth and Shilts, 1989b).

In this study, south-westerly flow patterns are mapped (Flow-sets 1, 9 and 10) but no evidence has been found to suggest that the tributary flow-sets should be grouped into the same flow patterns. Rather, the two tributary flow-sets (3 and 7) depict a distinct and coherent pattern of ice flow into the main ice stream channel. Furthermore, it is difficult to attribute the curved feature which borders the tributary (shown on Figure 7.20) to a southerly ice flow. This is because it is juxtaposed to both the ice stream and the tributary. Fieldwork investigations could fruitfully test this ice stream tributary hypothesis.

The identification of this tributary (if correct) has implications for contemporary ice stream research which suggests that some ice streams may be fed by subglacial valleys (tributaries) of faster moving ice in the onset zone (see Joughin *et al.* 1999) This evidence also supports the recent findings from West Antarctica which show that some ice stream margins are highly dynamic (Echelmeyer and Harrison, 1999) and in particular, that Ice Stream C shut-down was not a single event, but an ongoing process of margin migration (Jacobel *et al.*, 2000)

7.5.3. Ice Stream Timing and Significance.

It is suggested that the ice stream played a major role in the deglaciation of the ice sheet, which Dyke and Dredge (1989) hypothesised was rapid between 10,000 and 8,400 yr BP. Because of the configuration of the ice sheet (Figure 7.19), the ice stream probably operated after 9,000 yr BP and had shut-down by 8,500 yr BP. This approximate constraint is imposed by the ice sheet configuration at 9,000 yr BP and the fact that a major moraine system (the MacAlpine moraines) have been dated to mark a readvance prior to 8,500 yr BP. Figure 7.22 shows the location of the MacAlpine moraines and other major moraine systems on the Canadian Shield in relation to the Dubawnt Lake Ice Stream. Aylsworth and Shilts (1989a) suggested that these moraine systems reflect major ice recessional positions (stillstands) during deglaciation. It is clear from the position and proximity of the MacAlpine moraines to the Dubawnt Lake Ice Stream that the two are probably related and that the ice stream produced a lobate margin.

The extent of the ice stream implies that it drained a huge portion of the ice sheet, lowering the surface profile and probably resulting in a substantial thinning immediately prior to deglaciation. The inference that the ice stream was one of the last major flow events through the area is evidenced by the remarkable preservation of the ice stream bedforms and the numerous eskers systems superimposed on them.

Moreover, Aylsworth and Shilts (1989a) noted that esker tributaries commonly join trunk eskers at right angles. This suggests that ice flow was extremely slow or even non-existent after the ice stream shut down. It is tentatively suggested that the ice stream shut down when it had exhausted its supply of ice from upstream. Evidence for a shrinking catchment area is also provided by the inference that an ice stream tributary shut-down prior to the main ice stream.

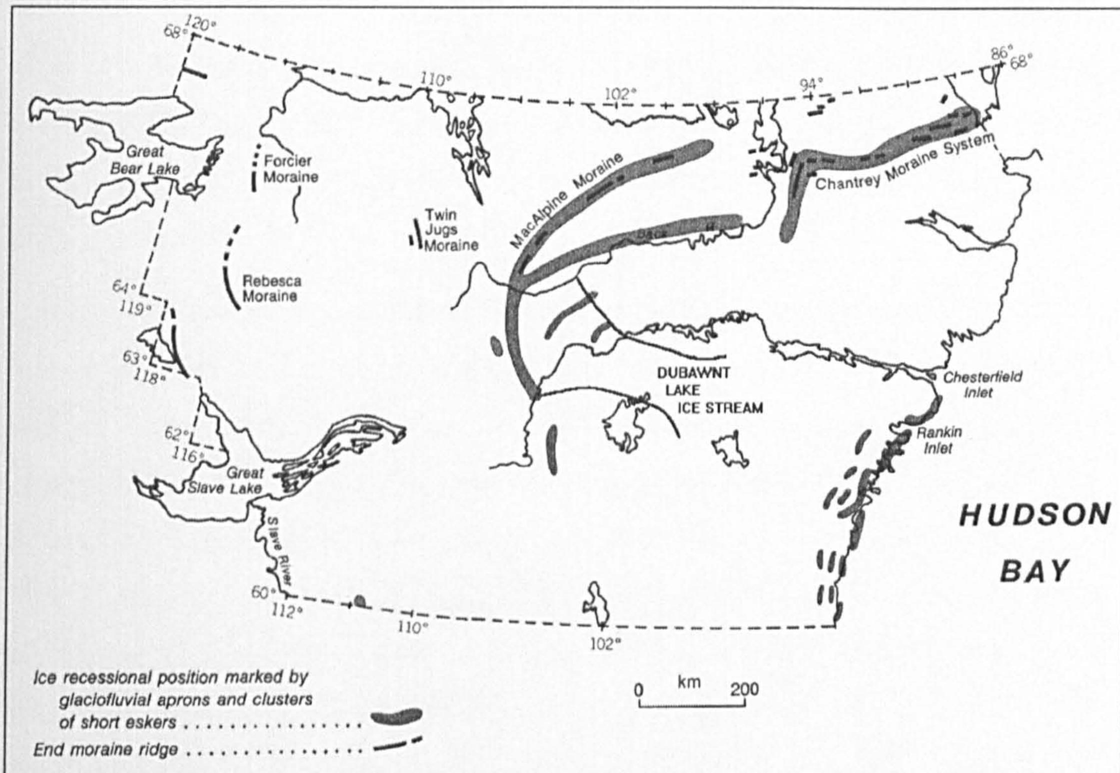

Figure 7.22. Moraine complexes on the Canadian Shield, representing major ice recessional positions. Note the proximity of the MacAlpine Moraine to the Dubawnt Lake Ice Stream, modified from Aylsworth and Shilts (1989a)

7.5.4. Ribbed Moraine: Implications for Ice Stream Shut-down.

Ribbed moraine are common throughout the area and cover many parts of the ice stream (see Aylsworth and Shilts, 1989a; b). Although no new mapping or analyses have been carried out in this study, their occurrence within the ice stream needs to be explained in the light of these findings. Ribbed moraine occur both within and outside of the margin and in some places, fields of ribbed moraine transgress the ice stream margin unaltered. Indeed, the ribbed moraine have been noted to lie superimposed upon lineations and occasionally appear to have partially broken up the drumlins. This makes the ribbed moraine unique from nearly all other documented cases (Clas Hättestrand, pers. com.) and provides strong evidence to suggest that they were formed after the drumlins had been laid down, and after ice stream activity.

There are two main explanations for the formation of ribbed moraine, the 'shear and stack theory' (e.g. Bouchard, 1989) and the 'fracturing theory' (e.g. Hattestrand and Kleman, 1999). The shear and stack theory explains ribbed moraine formation by the shearing and stacking of till slabs (including englacial material) during compressive ice flow. Ribbed moraine are subsequently formed by the basal melt-out of this material. In contrast, the fracturing theory suggests that ribbed moraine are formed at the transition from cold- to warm-based ice by fracturing of a pre-existing frozen till sheet. These hypotheses provide two possible scenarios for the ribbed moraine within the ice stream.

Scenario 1: Aylsworth and Shilts (1989a) stressed that any theory of ribbed moraine formation must be able to explain their proximity to the Keewatin Ice Divide, the close spatial relationship between drumlins and ribbed moraine, and the disparate compositional nature of these two types of landforms. They argued that their association with the Keewatin Ice Divide implies some glaciological control and that their morphology suggests they form under compressive ice flow. This is because the appearance of some ribbed moraine depict "inclined plates of sediment thrust one on top of the other" (see figure 16 of Aylsworth and Shilts, 1989a).

They suggested that this form results from the meltout of englacial debris thrust into the ice and results from a stagnating ice mass with little or no internal movement. In contrast, drumlins are thought to form under extensional ice flow. However, as Aylsworth and Shilts (1989a) noted, the close spatial relationship of the drumlins and ribbed moraine cannot be explained by this process alone. They favour an interpretation that the underlying sediments were a key control on bedform development; more deformable sediments produced drumlins whereas less deformable sediments produced ribbed moraine. If this is the case, ribbed moraine may reflect areas of high basal shear stress and possibly sticky spots beneath the ice stream. However, why do the ribbed moraine occur superimposed on drumlins? One explanation would be that sticky spots developed after the drumlins had been formed and may have been seed points of shut-down during the thinning of the ice sheet. The ribbed moraine may have been formed immediately after the ice stream shut down.

Scenario 2: It is difficult to fit the fracturing model of ribbed moraine formation to those found within the Dubawnt Lake Ice Stream. This theory suggests that ribbed moraines form when the warm-based margin of an ice sheet migrates inward during deglaciation. However, it can be assumed that the whole of the Dubawnt Lake Ice Stream was warm-based during operation and that it propagated inwards towards the cold based Keewatin Ice Divide, the onset zone of the ice stream. Given this configuration, we would expect ribbed moraine to only occur in the onset zone of the ice stream at the boundary between cold-based and warm-based ice. While it is true that the highest densities of ribbed moraine occur towards the onset zone, ribbed moraines are located up to 350 km downstream and possibly further. It is difficult to imagine that cold based islands were protected within the ice stream. Alternatively, ribbed moraine could have formed contemporaneously following a switch from cold- to warm-based conditions as the ice stream was activated. However, this can be discounted, because the ribbed moraine lie superimposed on the drumlins.

The following scenario is therefore proposed, which could account for the fracturing theory. After the ice stream shut down the ice sheet thinned and cold-based conditions resulted from a lack of friction at the bed and a reduction in pressure melting. However, the terminus would almost certainly be characterised by warm-based conditions as the ice sheet deglaciated and melted back. During deglaciation of the area, this warm-based zone migrated upstream and ribbed moraine were formed at the transitional boundary with the cold-based ice. However, why were the ribbed moraine only formed in certain places and why were the vast majority of ice stream lineations preserved without modification?

A universal theory of ribbed moraine formation remains elusive. Of the two main hypotheses presented here it would appear that neither can satisfactorily explain the widespread occurrence of ribbed moraine on a palaeo ice stream bed. It is tentatively suggested that the shear and stack theory is the most viable explanation. This theory has been modified by Dyke *et al.* (1992) who attributed sticking and sliding at the head of an ice stream on Prince of Wales Island to invoke their occurrence. If this is the case, some ribbed moraines may be the geomorphological signature of ice stream sticky spots.

7.6. Summary and Conclusions.

The Keewatin Ice Divide is thought to have represented one of the major domes of the Laurentide Ice Sheet. This area harboured some of the last remaining vestiges of the continental ice sheet and was one of the last areas to become ice free at the end of the last glaciation. This chapter presents evidence to suggest that a previously detected flow-set (or fan) (Kleman and Borgström, 1996; Bouton and Clark, 1990) was a significant ice stream, responsible for the rapid deglaciation of this portion of the ice sheet. The Dubawnt Lake Ice Stream fulfils at least four of the criteria for ice stream activity developed in this thesis (Chapter 4; Stokes and Clark, 1999). It fits the characteristic shape and dimensions, displays highly convergent flow patterns, contains highly attenuated bedforms and has very abrupt lateral margins.

Using satellite imagery, the ice stream is reconstructed at over 450 km in length, with widths varying from 305 km in the onset zone, 140 km in the main ice stream channel and diverging to 190 km at the terminus. The high parallel conformity of the bedforms, lack of internal cross-cutting relationships and extremely coherent flow pattern imply that the ice stream bedforms were generated isochronously and represent a snapshot view of the bed prior to shut-down.

The ice stream bedforms display some of the longest lengths in the literature, with maximum values of up to 12,743 m and elongation ratios as high as 48:1. Downstream variations in lineament length increase from the onset zone and reach peaks at around 220 km downstream. After this point, the lineation length decrease and drumlins become smaller as the ice stream diverges at the terminus. This pattern is mirrored by the downstream variation in elongation ratio. However, downstream variations in width peak much further upstream (between 140 an 180 km). This indicates that the longest bedforms are not the widest and that an abrupt increase in streamlining takes place between 180 and 220 km downstream. This coincides with a drop in bed elevation as the ice stream enters the Thelon sedimentary basin.

Although the bed topography and sedimentology did not control the overall location of the ice stream or its marginal positions, they did play an important role in encouraging fast flow velocities and may have increased downdraw in the onset zone. The influence of topography and geology on ice stream initiation were probably limited. It is more likely that the ice stream was driven by internal instabilities in the ice sheet during deglaciation. Climate amelioration at this time led to the formation of extensive proglacial lakes and along with significant marine inundation, these newly available moisture sources may have produced enhanced accumulation in the ice stream catchment area (the Keewatin Ice Divide). This would have created an oversteepened ice sheet profile which fed a proglacial lake and triggered ice streaming. The presence of glacial Lake Thelon would have provided an effective terminal environment for removing ice and accentuating drawdown. In addition, increased temperatures may have triggered a change in the thermal state of the bed, possibly causing a switch from cold-based to warm-based conditions, thus leading to rapid basal sliding.

It is suggested that the ice stream shut down when it ran out of ice causing widespread thinning of the ice sheet. The ice stream was the last major ice flow across the area and warm-based deglaciation is inferred from the numerous esker systems which lie across the ice stream bedforms at oblique angles. Prior to shut-down, the ice stream was fed by a small tributary which ceased before the main ice stream. This tributary gradually shrunk before shutting down all together, probably through a reduction in its catchment area.

Numerous ribbed moraine have been identified and mapped from large areas of the ice stream. It is difficult to reconcile current theories of ribbed moraine formation to their occurrence within an ice stream. Many of the Keewatin ribbed moraine fields are unique in that they lie superimposed upon drumlins. It is tentatively suggested that the shear and stack theory best supports their occurrence and that they were formed by the melt-out of englacial debris after the ice stream had shut down. If this is the case, the specific locations of ribbed moraine may be linked to the seed points of ice stream sticky spots.

The Dubawnt Lake Ice Stream led to the rapid retreat of the Keewatin Sector between 10,000 and 8,400 yr BP, probably after 9,000 yr BP. This is inferred from a number of dated ice marginal positions and a regional reconstruction provided by Dyke and Dredge (1989). The close proximity of the MacAlpine moraines to the ice stream terminus implies that the it had shut-down prior to 8,500 yr BP. The ice stream was responsible for the obliteration of one of the last major ice centres of the Laurentide

Ice Sheet and hints at the unpredictable and catastrophic importance of a major terrestrial ice stream forced by climate change.

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Chapter 8: Palaeo Ice Stream Geomorphology: Discussion Regarding Ice Sheet Stability and Ice Stream Functioning.

8.1. Introduction.

This chapter draws together several themes of the thesis and discusses some of the wider implications of the research. It begins with an appraisal of the geomorphological criteria and landsystems models developed in Chapter 4. In conjunction with the geomorphological criteria presented in this thesis, non-geomorphological criteria are briefly discussed and a subjective qualitative assessment of the palaeo-ice stream locations from Chapter 3 is presented.

Using the case studies in Chapters 6 and 7, the configurations of ice streams are explored, with particular attention paid to terrestrial ice streams, which have no modern analogues. Following this, the palaeo-glaciological implications of the M'Clintock Channel and Dubawnt Lake ice streams are discussed. What controlled the size of these ice streams? What was their likely ice flux and how did this effect the behaviour of the whole ice sheet? Was their activity linked? This also includes a discussion of the role of ice streams in ice sheets; do they play a regulatory role, or is their behaviour inherently unstable? Are some ice streams more stable than others? Are contemporary ice streams stable?

The third major section of this chapter is devoted to ice stream processes and functioning, coupling the fresh insights from this thesis with current theories in the literature. This section includes a discussion of the controls on ice stream location and initiation. A conceptual model is developed to summarise the main factors which affect the development of both pure and topographic ice streams.

The chapter concludes by highlighting several areas which future research could address.

8.2. Geomorphological Criteria: An Appraisal.

8.2.1. Significance of the Geomorphological Criteria.

In Chapter 3 of this thesis, a large number of the most pertinent palaeo-ice stream hypotheses in the literature were reviewed. This revealed that our understanding of palaeo-ice stream geomorphology is limited (Section 3.4). To address this problem, this thesis (Chapter 4; Stokes and Clark, 1999) developed several key criteria for identifying palaeo-ice streams. Notwithstanding the inherent problems of circularity (discussed in Section 4.5.1), it has been demonstrated that the criteria can be used to verify previously hypothesised palaeo-ice streams (Chapter 6), and aid the search for previously unidentified palaeo-ice streams (Chapter 7).

In the literature, the ice stream with the most compelling evidence for its activity was that described by Hodgson (1994) and referred to in this thesis as the M'Clintock Channel Ice Stream. Detailed investigation of this ice stream bed in Chapter 6 (Clark and Stokes, in press) revealed that it displayed at least five of the criteria and possibly seven (Section 6.2.1). Therefore, this ice stream potentially displays the almost complete landsystem. It is thus argued that it represents the best palaeo-ice stream track so far discovered and provides the securest validation of the ice stream landsystems models presented in Chapter 4. It is suggested that it portrays the 'text book' example of an isochronous record of ice stream activity.

Having found considerable support for the landsystems models in Chapter 6, Chapter 7 used them as an observational template to identify a previously undetected ice stream. This terrestrial ice stream (named the Dubawnt Lake Ice Stream), fulfilled four of the characteristic criteria. None of the other criteria have been identified. In some cases (i.e. ice stream marginal moraines), this may be a result of the imagery resolution, but the preferred explanation is that the other criteria are absent because of a lack of available 'soft' sediments. In contrast, the four criteria that are present (characteristic shape and dimensions, highly convergent onset zone, attenuated bedforms and abrupt lateral margins) are largely independent of specific geological controls and are more likely to be related to glacio-dynamic controls.

This raises an interesting question; are some criteria more valid than others? In short, the answer to this question is, yes. It is suggested that because some criteria are reliant on *both* fast ice flow and sedimentary conditions, that their occurrence will be less widespread. For example, deformable bed conditions and Boothia-type dispersal trains rely on the availability of a large supply of sediment in the ice stream onset zone, and for the latter, a distinctive lithology. Not all ice streams operated over such sedimentary basins. On the other hand, the characteristic shape and dimensions of an ice stream flow-set, its abrupt margin and its highly convergent flow patterns, require no specific sediment supply, although the development of subglacial bedforms is an essential pre-requisite.

In an appraisal of the criteria, it is possible to rank them in order of their significance. Of most importance are the characteristic shape and dimensions, the abrupt lateral margin and the highly convergent onset zone. These represent the fundamental characteristics of an ice stream. Almost as important are highly attenuated bedforms, but their occurrence may also be related to the sedimentary conditions of the ice stream bed. For example, some ice streams may 'struggle' to produce bedforms because of a lack of sediment. It is not presumed that fast ice flow has to produce subglacial bedforms, just that it may do.

Of next most importance are ice stream marginal moraines. Their occurrence is almost certainly a function of sediment supply but since so little is known about their formation, it is difficult to speculate on their distribution and significance. The other three criteria are all important, but their occurrence is equally governed by the underlying geology, as well as ice dynamics. Some marine-based ice streams may slide across their beds and this may not produce a substantial offshore sediment accumulation. In the same way, not all ice streams flow on a deforming bed and evidence of deformable sediments is in no way exclusive to ice streams, it has also been linked to surging glaciers for example. Finally, Boothia-type dispersal trains are extremely rare because they rely on specific ice dynamics over specific sediment types.

In summary, although none of the individual criteria are exclusive to ice stream activity, palaeo-ice stream investigations in Chapters 6 and 7 would appear to support their *collective* diagnostic use for identifying former ice streams. The M'Clintock Channel Ice Stream represents the best palaeo-ice stream record so far identified and provides our best means of validating the landsystems models outlined in Chapter 4.

It will remain difficult to unequivocally validate the individual criteria until we can gain an unimpeded view of the beds of contemporary ice streams. Although this appears unlikely in the foreseeable future, recent advances in marine sonar technology have begun to disclose geomorphology from the foregrounds of contemporary ice streams. Such research is potentially significant, because it allows us to make the crucial link between ice streams and their geomorphological products. Here we have evidence which can directly link the terrestrial record of palaeo-ice streams to their contemporary counterparts in West Antarctica.

8.2.2. Submarine Ice Stream Geomorphology.

Recently, it has been ascertained that both the East and West Antarctic Ice Sheets advanced a considerable distance across the continental shelf during the Last Glacial Maximum (LGM) (Bindschadler *et al.*, 1998). Their subsequent retreat (which began around 17,000 years ago in East Antarctica and 10,000 years ago in West Antarctica) has exposed the seafloor to scrutiny. In West Antarctica the ice sheet has retreated around 150 km from the continental shelf margin. This retreat was almost certainly facilitated by the Siple Coast ice streams whose tracks have recently been delineated by marine investigations beneath the Ross Ice Shelf.

These tracks display some of the criteria postulated in Chapter 4 (Stokes and Clark, 1999) as being indicative of palaeo-ice stream activity. For example, it has been argued that mega-lineations found in formerly glaciated parts of Canada can only form under extremely rapid ice flow such as in ice streams (Clark, 1993). For this reason, highly attenuated bedforms were predicted as being a diagnostic criterion for ice stream flow. Geophysical investigations in the Ross Sea of Antarctica permitted Shipp *et al.*, (1999) to identify features of appropriate characteristics and size revealing them to be the first marine discovery of Clark's (1993) mega-lineations. They discovered lineations in excess of 20 km in length along with attenuated drumlins (lengths 3-8 km, widths 0.5-0.75 km) in the foregrounds of the Siple Coast ice streams.

These findings considerably strengthen the hitherto untested diagnostic use of highly attenuated bedforms as indicators of fast ice flow. Moreover, several other criteria appear to have been detected. Shipp *et al.* (1999) identified sediment facies interpreted to reflect a deformation till which transported sediment downstream, much like a conveyor belt. This is inferred from facies architecture and "grounding zone wedges" which occur at the mouths of the troughs which the ice streams occupied. These troughs are the characteristic shape and dimensions of the ice streams and it is interesting to note that laterally accreting ridges separate the bathymetric lows from the higher inter-stream areas. It is suggested here, that these features may be analogous to ice stream marginal moraines. As such, the submarine geomorphology in the foregrounds of the Siple Coast ice streams potentially validates five of the criteria outlined in Chapter 4; characteristic shape and dimensions, highly attenuated bedforms, ice stream marginal moraines, evidence of deformable till and sediment accumulation fans.

A spectacular example of the potential of marine geophysics has been the recent discovery of a palaeo ice stream track off the northern tip of the Antarctic Peninsula (Canals *et al.*, 2000). Here, a cluster of mega-lineations referred to as a 'bundle structure' depicts a sinuous track of grooves and ridges covering an area 100 km long and up to 21 km wide, and shown in Figure 8.1. This is interpreted as the bed imprint of a palaeo ice stream that drained an expanded Antarctic Peninsula Ice Cap during the Last Glacial Maximum. While similar in scale and character to mega-lineation patterns observed terrestrially (Clark, 1993), what is exceptional is the longitudinal continuity of the features with ridge/groove structures extending uninterrupted for 100 km and displaying convergence and sinuosity over their length.

It may be that terrestrial records are more fragmented and less well preserved and we have yet to observe an uninterrupted ice stream imprint from convergent onset zone through the trunk of the ice stream. It is clear that this 'Boyd Strait Ice Stream' imprint is the most complete record so far discovered. This is attributable to the remarkable preservation potential of submarine geomorphology. If the M'Clintock Channel Ice Stream is the best terrestrial record of an ice stream, then the Boyd Strait ice stream is possibly the most complete ice stream signature. The bundle of mega-lineations represents the largest ice flow related landform ever identified.

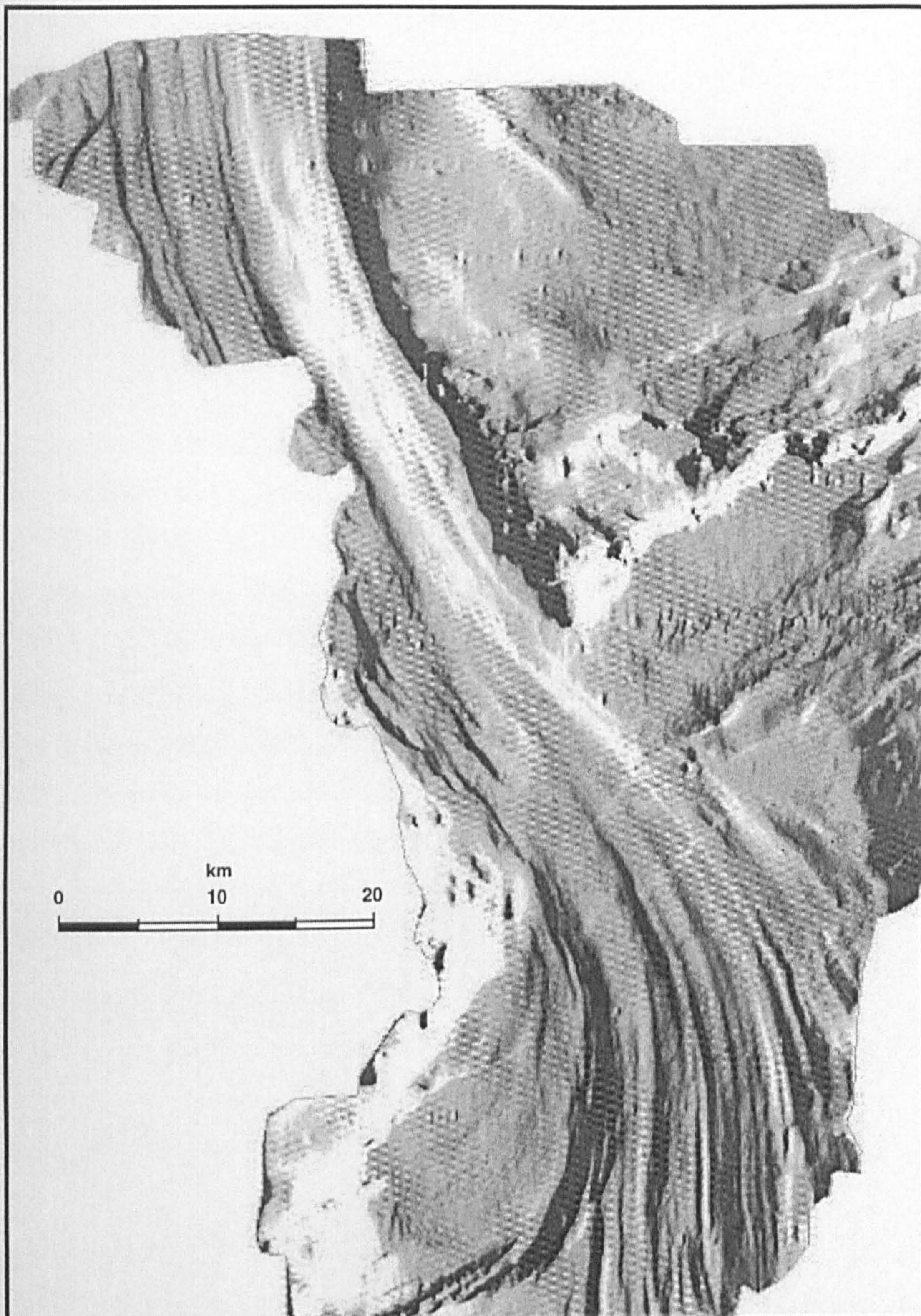

Figure 8.1. Submarine geomorphological imprint of the Boyd Strait palaeo-ice stream off the northern tip of the Antarctic Peninsula. Shaded relief image constructed from swath bathymetry data illuminated from the east (image kindly supplied by Miquel Canals, from Canals *et al.*, 2000).

In summary, submarine investigations in the straits and sounds which bordered former ice sheets and ice streams have the potential to provide a good test of the geomorphological criteria developed in this thesis. They also provide an invaluable record of ice stream activity, especially where terrestrial evidence is obscured or absent. Increased detection of submarine geomorphology will allow us to view hitherto inaccessible palaeo-ice stream locations and will thus overcome one of the main inhibitors to palaeo-ice stream research identified in Chapter 3 (Section 3.4.5).

8.2.3. Non-Geomorphological Criteria of Ice Stream Activity.

In addition to the geomorphological criteria developed in this thesis, ice stream hypotheses can also be augmented by non-geomorphological criteria. Bourgeois *et al.* (2000) suggested that reconstructed ice surface profiles and ice velocities can also be used in conjunction with the geological and geomorphological record.

Citing evidence from contemporary pure ice stream surface profiles (Bentley, 1987), they noted that they are characterised by a concave upper surface. This is in contrast to the generally convex profile of a slow flowing ice sheet. Therefore, it may be possible to invoke the presence of an ice stream from former ice surface profiles, perhaps reconstructed from trimlines or lateral moraines, or even from ice sheet models (e.g. Kaplan *et al.*, 1999).

In terms of former ice velocities, Bourgeois *et al.* (2000) suggested that estimated ice fluxes can be reconstructed from simple mass-balance calculations. Once ice stream dimensions and ice sheet thicknesses are known, accumulation can be assumed for the catchment area (e.g. from ice sheet and climate models) to produce an estimate of ice flux through the ice stream.

Bourgeois *et al.* (2000) used these two criteria in addition to the geomorphological criteria suggested in this thesis to identify the locations of several palaeo-ice streams in the Icelandic Ice Sheet at around 21,000 years ago (see Section 3.3.5). It is thus suggested that (where available) these 'glaciological' criteria for identifying palaeo-ice streams provide a useful addition to the geomorphological evidence of their activity.

8.2.4. Assessment of Hypothesised Palaeo-Ice Stream Locations.

Given the recent discoveries from submarine palaeo-ice stream tracks and using both the non-geomorphological criteria (outlined above) and the geomorphological criteria developed in this thesis, it is possible to assess the likelihood of a number of palaeo-ice stream locations. Although this process is entirely subjective and heavily biased by the reported evidence in the literature, it is a useful way of highlighting those ice streams which appear firm candidates for ice streaming and those which are less certain. These considerations are of critical importance to ice sheet modelling, where the prescribed locations of ice streams is crucial.

In Chapter 3 the evidence for a large number of palaeo-ice streams was presented in Tables 3.1 to 3.5. It is clear that some ice streams left behind several lines of evidence regarding their activity, whereas others left behind very little evidence (see Table 3.6). Indeed, the likelihood of a number of palaeo-ice streams should be questioned, whereas others can be assumed with far greater confidence. Tables 8.1 to 8.5 show all of the ice streams cited in the corresponding Tables 3.1 to 3.5 and include an assessment of the likelihood of their occurrence (1= unlikely, 5= likely).

The likelihood was not derived mechanistically, but rather, reflects a personal view (and often a 'gut feeling'). It is based on the evidence documented in the literature coupled with the insights provided through the research presented in this thesis. It is not intended to be a 'definitive guide' to palaeo-ice stream locations, but rather, it is hoped that it will open up the debate concerning their locations and increase the awareness of identifying them objectively.

It can be seen that some ice streams, such as the Hudson Strait Ice Stream, are far more likely candidates for ice streams than others. This is because they display a range of evidence for their activity. Evidence for the Hudson Strait Ice Stream has been found both terrestrially, and offshore and far reaching evidence in the form of iceberg rafted debris (Heinrich events) has even been used to infer the timing and velocity of this ice stream (see Dowdeswell *et al.*, 1995). In addition, ice sheet modelling of the Laurentide Ice Sheet has shown that a Hudson Strait Ice Stream is entirely consistent with the inferred ice sheet configuration (e.g. Fisher *et al.*, 1985).

Table 8.1. Hypothesised palaeo-ice streams of the Laurentide Ice Sheet and an assessment of their likelihood.

Reference	Location of Ice Stream(s)	Dimensions	Timing	Likelihood (5=likely; 1=unlikely)	Comments
1 and 5. Andrews *et al.* (1985); Laymon (1992)	Hudson Strait	not given explicitly	not given explicitly	5	A wealth of evidence from ice flow indicators showed that a large marine trough preferentially captured fast flowing ice from a large source area and over soft sediments. Evidence for this ice stream was also preserved in marine sediments from the North Atlantic whereby bands of iceberg rafted debris have been associated with its activity. Modelling experiments also supported the notion of this ice stream.
2. Dyke and Morris (1988)	Prince of Wales Island, Canadian Arctic	20 km wide and 100 km long	Last glacial maximum	5	Although relatively small compared to contemporary ice streams, the bedform pattern of this ice stream is a text-book example of fast ice flow. More importantly, the Boothia-type dispersal plume could only be produced by an ice stream.
3. Hicock (1988)	Albany Valley, James Bay Lowlands	ca. 50 km wide ca. 600 km long	Last glacial maximum	3	Although the presence of a dispersal plume implied fast ice flow, the bedform evidence for this ice stream is less obvious. This hypothesis was based on the properties of the underlying till which were thought to be of a deformation type. Difficult to ascertain if this ice flow was a spatially discrete channel of fast ice.
4. Boyce and Eyles (1991)	Simcoe Lobe (Ontario)	not given explicitly	less than 13,000 yr BP	3	This ice stream was based primarily on the longitudinal and lateral variations in the bedform morphometry coupled with evidence of subglacial sediment deformation. Like the Albany Valley ice stream (above), it is difficult to ascertain the spatial extent of fast ice flow, especially with little evidence to suggest abrupt lateral margins.
6. Kaufman *et al.* (1993)	perpendicular to the mouth of Hudson Strait	max. of 200 km wide and over 300 km long	9,900 to 9,600 yr BP	4	Evidence for fast ice flow of this ice stream was constrained by dating techniques. Although terrestrial evidence is scarce, ice flow indicators such as striae and erratics delimited the extent of the ice stream. Difficult to assess whether this was a readvance of the ice sheet margin or an ice stream *sensu-stricto*.
7. Hodgson (1994); this thesis (chapter 6) and Clark and Stokes (in press)	Victoria Island, Canadian Arctic Archipelago	at least 80 km wide and 200 km long	between 10,400 and 9,600 yr BP	5	Substantial evidence for an ice stream provided by a bedform pattern which displayed a remarkably abrupt margin juxtaposed to highly attenuated bedforms and ice stream marginal moraines. Evidence for fast ice flow was further constrained by dating techniques. Possibly the best terrestrial record of an ice stream ever identified.
8. Veilette (1997)	James Bay Ice Stream	ca. 200 km wide and 500 km long	ca. 9,000 yr BP	2	Evidence for ice stream activity based almost entirely on cross-cutting landforms thought to have resulted from an abrupt change in ice direction. Other than the distribution of streamlined landforms, little evidence to suggest the presence of an ice stream, more likely to be an advancing lobe.

Table 8.1. Continued.

Reference	Location of Ice Stream(s)	Dimensions	Timing	Likelihood (5=likely; 1=unlikely)	Comments
9. Patterson (1997; 1998)	Des Moines Lobe, Minnesota	max. 200 km wide and 900 km long	between 14,000 and 12,000 yr BP	4	This ice stream was postulated from a landform assemblage which appeared similar to that observed at the foregrounds of contemporary surging glaciers. Other evidence included convergent topography in the onset zone and the dating of several episodes of advance and retreat. Difficult to ascertain whether one ice stream produced a time-transgressive record or several separate advances and retreats of individual lobes produced the bedforms.
10. Kaplan *et al.* (1999)	Cumberland Sound Ice Stream	ca. 75 km wide and 150 km long.	not given explicitly	3	Terrestrial evidence for this ice stream is absent or obscured, but ice sheet modelling experiments indicated that an ice stream can develop given realistic boundary conditions and appropriate sliding parameters at the bed. Whether an ice sheet could support this ice stream and the Hudson Strait Ice Stream in close proximity is open to debate.
This thesis (chapter 7)	Dubawnt Lake, Keewatin.	450 km long and 140 km wide	probably between 10,000 and 8,400 yr BP.	5	The highly convergent bedform pattern with an abrupt margin and some of the longest bedforms ever identified fits the characteristics of an ice stream signature. Given the proximity of this ice stream to a large supply of ice (the Keewatin Ice divide) during deglaciation of an unstable ice sheet, it is likely that the flow was short-lived and not an advance of a slow-flowing lobe.

Table 8.2. Hypothesised palaeo-ice streams of the Cordilleran Ice Sheet and an assessment of their likelihood.

Reference	Location of Ice Stream(s)	Dimensions	Timing	Likelihood (5=likely; 1=unlikely)	Comments
11. Hicock and Fuller (1995)	Skeena Valley/continental shelf, British Columbia	ca. 12 km wide and ca. 100 km long	Late glacial	4	A wealth of data relating to till composition and properties suggested that deforming bed conditions were pervasive. Given the similarity in size of the Skeena Valley to those occupied by contemporary outlet glaciers and ice streams in Antarctica it is likely that ice flow was rapid. Whether or not this was an ice stream or an outlet glacier remains open to question.
12. Evans (1996)	Lillooet and Cheakamus Valleys, southern Coast Mountains, British Columbia.	not given explicitly	between 20,000 and 15,000 yr BP	2	Although topographic troughs would have preferentially channelled faster flowing deeper ice, it is difficult to substantiate evidence for ice stream activity. The distribution of abraded rock landforms were thought to demarcate ice streams, but these features could conceivably have been formed under deep slow-flowing ice.

Table 8.3. Hypothesised palaeo-ice streams of the British and Irish Ice Sheets and an assessment of their likelihood.

Reference	Location of Ice Stream(s)	Dimensions	Timing	Likelihood (5=likely; 1=unlikely)	Comments
1. Eyles *et al.* (1994)	North Sea Ice Stream	not given explicitly	not given explicitly	2	This ice stream was postulated from a landform assemblage similar to that produced by surging glaciers. There is little bedform evidence parallel to ice flow but the stratigraphy of tills are thought to reflect periodic surging of an ice lobe on shore. This ice stream hypothesis was also supported by modelling experiments but it is difficult to reconstruct its extent and activity.
2. Merrit *et al.* (1995)	Moray Firth Ice Stream (Scotland)	not explicitly given	ca. 13,000 yr BP	2	The scarcity of terrestrial evidence means that this ice stream was based almost entirely on marine deposits marking stillstands of a large tidewater glacier. Difficult to ascertain whether these deposits are related to a small outlet glacier or larger ice stream.
3. Knight and McCabe (1997); Knight *et al.* (1999); McCabe and Clark (1998)	Irish Ice Streams (north central Ireland) and Irish Sea Basin Ice Stream	up to 70 km long, width not given	between 15,000 and 14,000 yr BP	2	The reconstructed extent of these ice streams were remarkably small, given the large size of contemporary ice streams. Ice streams were invoked to have modified transverse ridges but the margins of the ice stream are not clear. More likely to be a period of discrete ice flow associated with warm-based ice, rather than ice streams *sensu stricto*.

Table 8.4. Hypothesised palaeo-ice streams of the Scandinavian Ice Sheet and an assessment of their likelihood.

Reference	Location of Ice Stream(s)	Dimensions	Timing	Likelihood (5=likely; 1=unlikely)	Comments
A – 0 (Figure 4). Punkari (1993; 1995; 1997); Kleman *et al.* (1997)	13 ice streams in the Scandinavian Ice Sheet including the Baltic Sea and Norwegian Channel Ice Streams	10's of kms wide, 100's of kms long	During deglaciation	5	Good evidence for several ice streams is revealed by bedform patterns which are densest in the centre of the lobes. Ice stream margins are demarcated by fluvial deposits known as interlobate complexes. The configuration of these lobes matches the time-transgressive retreat of the ice sheet margin during deglaciation.
B. Holmlund and Fastook (1993)	Baltic Sea Ice Stream	not given explicitly	ca. 11,000 yr BP	5	As above. The presence of the Baltic Sea Ice Stream is further supported by ice stream modelling experiments which depict a buoyant ice stream similar in size and configuration to ice streams in Antarctica.

Table 8.4. Continued.

Reference	Location of Ice Stream(s)	Dimensions	Timing	Likelihood (5=likely; 1=unlikely)	Comments
B - G. Dongelmans (1996)	Central and South Finland, Russian Karelia and northern Scandinavia	ca. 40 km wide 200-250 km long	During deglaciation	5	As above. Many of Punkari's (1993; 1995; 1997) ice streams are also evidenced by Dongelmans who invoked ice stream activity from distinct bedform patterns in between areas of little or no bedforms.
A. King *et al.* (1996); Sejrup *et al.* (1998)	Norwegian Channel Ice Stream	not given explicitly	multiple glaciations; last deglaciation up until 15,000 yr. BP	5	The location of this ice stream was inferred from both marine and terrestrial evidence. Terrestrial evidence in the form of elongated ridges are thought to mark the marginal area of the ice stream. Perhaps the most substantial evidence is found offshore, where the architecture of a large sediment fan has been attributed to periodic inputs from an ice stream in the channel.

Table 8.5. Hypothesised palaeo-ice streams of the Eurasian Arctic and Icelandic Ice Sheets and an assessment of their likelihood

Reference	Location of Ice Stream(s)	Dimensions	Timing	Likelihood (5=likely; 1=unlikely)	Comments
X and Z on Figure 4. Punkari (1995b)	Novaya Zemyla / Barents Ice Sheet	not given explicitly	not given explicitly	5	Evidence for these ice streams is similar to that thought to have been deposited by the Scandinavian ice streams, i.e. based largely on the bedform patterns produced by several terrestrial lobes.
Figure 5. Vorren and Laberg (1997); Dowdeswell *et al.* (1998)	Barents Sea-Svalbard Ice Sheet	not given explicitly	operated at glacial maxima throughout the Pleistocene	4	Terrestrial evidence of these ice streams is absent, but marine evidence in the form of several large sediment accumulations at the mouths of topographic troughs is thought to be indicative of ice streaming. Dating of these accumulations has shown that sediment delivery was periodic and facies architecture reveals that a deforming bed may have been important.
Fig 6. Bourgeois *et al.* (2000)	Icelandic Ice Sheet	20 km wide, 150-200 km long	LGM (21,000 yr BP)	5	The distribution of attenuated bedforms in topographic troughs coupled with inferred ice sheet profiles and mass balance calculations provides strong evidence for ice stream activity. Field evidence of a deformation till is also supported by sedimentary wedges offshore and some ice streams were located above geothermal heat anomalies which produced enhanced lubrication in the form of excess meltwater.

In contrast to the Hudson Strait Ice Stream, evidence for some ice streams remains inconclusive (Tables 8.1-8.5). For example, the Irish Ice Streams postulated by Knight and McCabe (1997) and Knight *et al.* (1999) appear to have left very little evidence of their activity. Their occurrence was invoked from areas of transverse ridges (Rogen moraine) which had been drumlinised following a change in the basal thermal regime. However, the margins of the ice streams are poorly defined and their size and shape is inconsistent with contemporary ice streams. The modification of the transverse ridges may well be a result of a change in the basal thermal regime, but this is not enough evidence to invoke ice stream activity.

In the same way, although the southern Laurentide Ice Sheet margin was almost certainly drained by terrestrial ice streams, finding substantial evidence of their activity remains problematic. Part of this stems from the fact that the margin positions fluctuated over short timescales and probably erased much of their evidence, but it also stems from a lack of understanding regarding the configuration of terrestrial ice streams. In addition, many palaeo-ice stream hypotheses are confused by the ambiguous use of the term 'ice stream'. In an attempt to address these issues, the following section explores the configuration of ice streams.

8.3. The Role of Ice Streams in Ice Sheets.

8.3.1. Ice Stream Configurations.

From the discussion above, it would appear that we can be reasonably confident of a number of palaeo-ice stream locations from former ice sheets. In this thesis, a wealth of evidence has been described to identify a marine-based ice stream and a terrestrially terminating ice stream. However, because we have no modern analogues of terrestrial ice streams, we are unsure what they look like. We have to rely on bedform evidence which has often been modified or is sometimes incomplete. This poses a number of important questions when trying to reconstruct the configuration of a terrestrial ice stream (Section 3.4.4). A fundamental question concerns the terminus; how does the ice stream evacuate ice rapidly?

It is suggested in Section 3.4.4 that because contemporary ice streams feed ice shelves or terminate in open water conditions, the removal of ice at the terminus is rapid and this maintains a high velocity. Terrestrial ice streams have a much less effective method for removing ice and presumably therefore, they have to advance, producing a large splayed ice lobe at its terminus (Section 3.4.4). Nevertheless, there are several problems when trying to envisage the configuration of a terrestrial ice stream.

Contemporary ice streams are bordered by slower moving ice all the way to the grounding line and they have a predictable lateral velocity gradient often described as 'plug flow' (Figure 4.2). This presents further questions with regards to the configuration of terrestrial ice streams. Are terrestrial ice streams characterised by plug flow all the way to the ice margin? Is fast flow restricted to a narrow zone along the central axis of the lobe, or does the divergence of flow producing the splayed lobe pattern result in a more uniform but slower regime of flow velocities? It is presumed to be the latter case, with a lobe of slower moving ice (i.e. below typical ice stream velocities) protruding beyond the overall ice margin.

In an attempt to conceptualise this, two possible configurations for terrestrial ice streams are shown in Figures 8.2a and 8.2b. Firstly, consider a terrestrial ice stream terminating with a large splayed flow pattern, producing a low angled lobe. The lobe may have a large surface area and low enough elevations to ensure that ablation matches the ice flux delivered by the ice stream. Figure 8.2a shows the surface profile and plan view of this configuration which has no modern analogue but closely resembles the inferred configuration of Dubawnt Lake Ice Stream described in chapter 7.

Alternatively, a terrestrial ice stream may terminate in a proglacial lake and evacuate ice rapidly via a calving front. This configuration is depicted in Figure 8.2b and again has no modern analogue. However, it is apparent that many terrestrial ice streams fed proglacial lakes, an example being the Simcoe Lobe of the southern Laurentide margin (Boyce and Eyles, 1991).

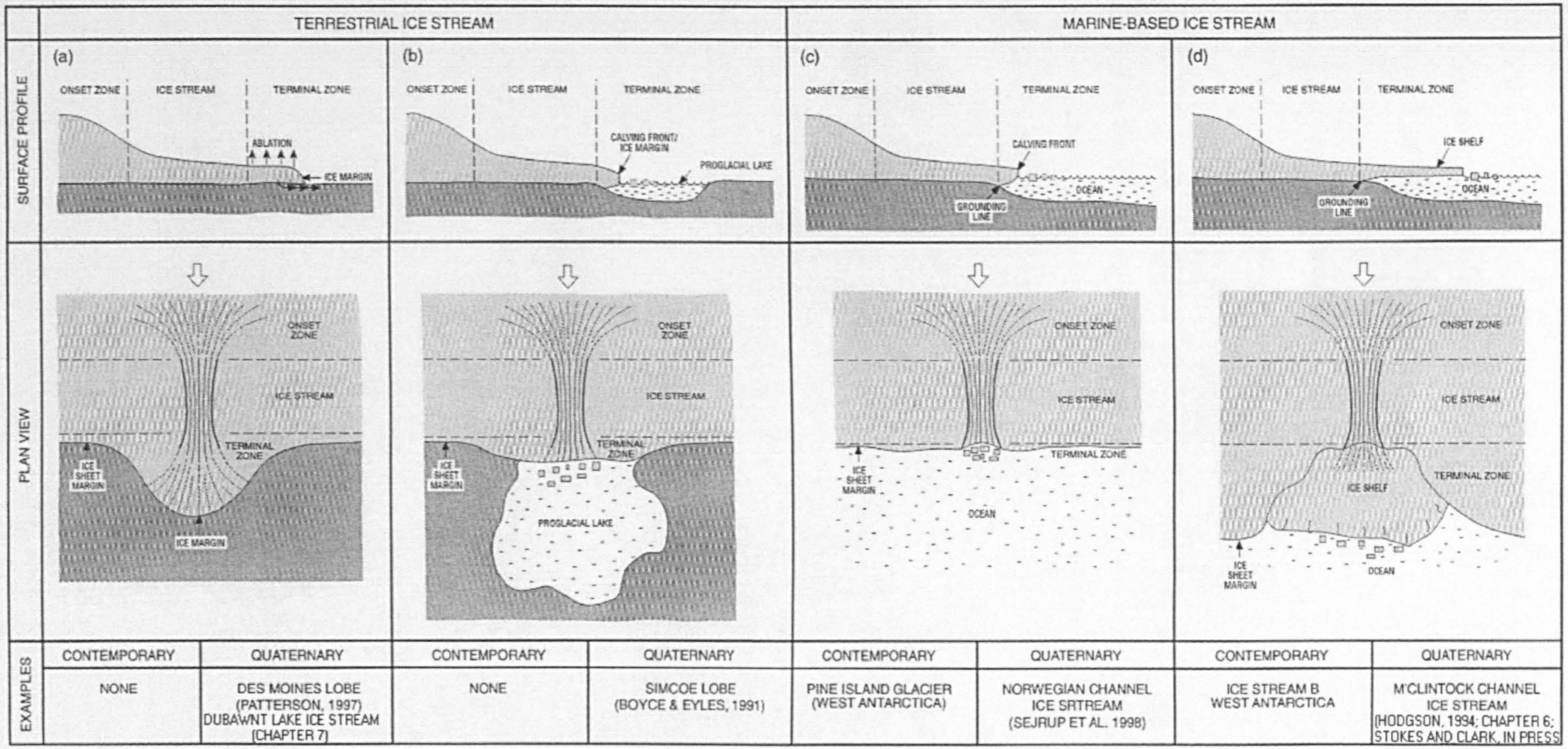

Figure 8.2. Conceptual configurations of terrestrial and marine-based ice streams and contemporary and palaeo examples. Some terrestrial ice streams terminate on land and have no way of rapidly removing ice. Margin advance produces splayed lobes which lowers the surface elevation of the ice sheet, enhancing ablation, see (a). Other terrestrial ice streams may drain into pro-glacial lakes and discharge their huge ice flux in the form of icebergs. This configuration is shown in (b). All contemporary ice streams are marine-based and can be broadly classified as those which drain into open water (c) and those which feed ice shelves (d).

It is plausible that the Dubawnt Lake Ice Stream may have fed a small proglacial lake which is well documented from fieldwork in the area (Section 7.5.1). However, the influence of this lake was not substantial (although it may have helped trigger ice streaming) because the terminus of the ice stream displays a divergent bedform pattern. This would not be expected if the ice stream maintained rapid flow along its length and entered a large proglacial lake. This mechanism can be seen in marine-based ice streams which terminate in open water conditions such as Pine Island Glacier in West Antarctica, which flows directly into the Amundsen Sea. Figure 8.2c shows this configuration and it is suggested that the Norwegian Channel Ice Stream may be an ancient example of such an ice stream (see Section 3.3.4). For comparison Figure 8.2d shows the configuration of marine-based ice streams which feed ice shelves. This is the situation observed on contemporary pure ice streams, such as those on the Siple Coast of West Antarctica, and the M'Clintock Channel Ice Stream is thought to be a suitable Quaternary example.

It is anticipated that the conceptual models of ice stream configurations (Figure 8.2) will help clarify the distinction between marine-based and terrestrial ice streams. Figure 8.2 also makes a distinction between the onset zone, trunk ice stream and terminal zone. This too is of paramount importance when attempting to reconstruct the configuration of an ice stream and identifying these zones will help to provide a more accurate interpretation of the bedform imprint.

In Chapter 3 (Section 3.4.4) a lack of modern analogues was cited a major inhibitor to terrestrial palaeo-ice stream research. It is hoped that these conceptual models (and Chapter 7) will go some way in overcoming the problems regarding their configuration and behaviour.

8.3.2. Glaciological Implications Arising from the M'Clintock Channel and Dubawnt Lake Ice Streams.

Investigations in this thesis (Chapters 6 and 7) have identified two ice streams which drained the Laurentide Ice Sheet between 11,000 and 8,000 years ago. Dating constraints from Hodgson's (1994) fieldwork and the extra timing constrained by the ice stream record on Prince of Wales Island indicate that the M'Clintock Channel Ice Stream operated between 10,400 and 10,000 years ago (see Section 6.4.6 for a discussion regarding the reliability of Hodgson's dating constraints). Although it is much more difficult to constrain the timing of the Dubawnt Lake Ice Stream, previous studies (most notably Dyke and Dredge, 1989) suggest that it probably operated between 10,000 and 8,500 years ago, possibly before 9,000 years ago. Key questions that have to be answered in the context of this chapter are;

1. How much ice did these two ice streams drain?

2. What affect did their ice flux have on ice sheet configuration?

3. Was their activity linked?

8.3.2.1. Size and Controls on Ice Stream Width.

The M'Clintock Channel Ice Stream was a large (720 x 140 km) marine-based ice stream, with a surface area of around 100,800 km^2. Based on the known extent of the ice sheet at around 10,000 yr BP (see Dyke and Prest, 1987a) its catchment area is estimated to be around 400,000 km^2, around 20% of the Keewatin Sector of the Ice Sheet (Section 6.5.6). Close to the ice stream terminus, the width of the ice stream (140 km) and the estimated minimum ice thickness (400 m; Hodgson, 1994) provide a cross sectional area of 56 km^2 (see Figure 6.20).

If the ice stream velocity is assumed to be the same as that inferred for the Hudson Strait Ice Stream (4 km a^{-1}, see Dowdeswell *et al.*, 1995), then the ice flux through the ice stream could have been as high as 224 km^{-3} a^{-1}. This is an order of magnitude higher than the calculated ice discharge of contemporary Ice Stream B (Bindschadler *et al.*, 1987) and emphasises its ice sheet wide significance during deglaciation. Indeed, its importance is comparable to the Hudson Strait Ice Stream, whose iceberg production during Heinrich events has been estimated at between 312 and 540 km^{-3} a^{-1} (see Dowdeswell *et al.*, 1995). Given the minimum and maximum estimates of the ice stream life cycle, then the M'Clintock Channel Ice Stream could have drained between 44,800 and 89,600 km^3 of ice in total.

The Dubawnt Lake Ice Stream was slightly smaller (450 x 140 km) with a catchment area estimated at around 190,000 km^2. Because the Keewatin Ice Sheet is thought to have retreated rapidly between 9,000 and 8,400 yr BP (see Dyke and Prest, 1987a;

Dyke and Dredge, 1989), this ice stream drained a significant proportion of the Keewatin sector and probably accounted for over 50% of its drainage.

It is difficult to estimate an ice flux for this ice stream because of a scarcity of ice thickness data from this region and the lack of any absolute dating constraints. However, its activity would have resulted in a considerable thinning of the ice sheet, leading to enhanced ablation and subsequent deglaciation. The distribution of numerous eskers overlying the ice stream imprint suggests that deglaciation occurred approximately parallel to the main ice stream axis, thus indicating that it was the last major ice flow through the area (Section 7.5.4).

If we refer back to Table 2.1 (Chapter 2), it can be seen that the drainage basin of the M'Clintock Channel Ice Stream (400,000 km^2) is far larger than the drainage basins of contemporary ice streams, the largest of which is Pine Island Glacier (220,000 km^2). This is because the dimensions of the M'Clintock Channel Ice Stream are far greater than contemporary ice streams, and this would be expected during deglaciation. In contrast, the catchment area of the Dubawnt Lake Ice Stream (190,000 km^2) is more comparable with contemporary ice streams, although it is larger than most. Again, this would be expected given the rapidity of deglaciation.

Of greatest note is the large width of the ice streams compared to their contemporary counterparts. Both ice streams display onset zones which typically converge into a main ice stream trunk around 140 km in width. This is the minimum width of both ice streams which is far wider than contemporary ice streams which usually range from 30 to 80 km (Table 2.1). Given the fact that the margins of both ice streams were thought to be controlled by internal glaciological mechanisms rather than geology or topography, it is important to consider what dictates the width of pure ice streams. Is their a maximum or minimum width for an ice stream? Why were the M'Clintock Channel and the Dubawnt Lake Ice Streams so wide?

Under glaciological regulation, the position of an ice stream margin is a delicate balance between advection of cold ice from outside the margin and the generation of shear heating from the fast flowing ice stream (Bindschadler and Vornberger, 1998). It follows then, that if advection is dominant, the ice stream will narrow, but if shear heating is dominant, perhaps due to an increase in velocity, then the ice stream will widen. This has important implications for the force balance of an ice stream and ultimately the amount of ice it can discharge.

Consider an ice stream in which side drag is the dominant resistive force. An increase in width would cause an increase in speed, because the driving force is increased with no increase in resistance. This would lead to a positive feedback mechanism whereby the ice stream can widen at a rate which is only limited by the amount of ice available upstream. This scenario may explain why the M'Clintock Channel and Dubawnt Lake ice streams were much bigger than their contemporary counterparts. They were dominated by side drag and increased in width until basal drag became relatively more important (i.e. the M'Clintock Channel Ice Stream) or ice supply from upstream became the limiting factor (i.e. the Dubawnt Lake Ice Stream).

This has implications for the Siple Coast ice streams of West Antarctica. They have a large supply of ice, so what is stopping them widening? Recent evidence from Ice Stream B (Bindschadler and Vornberger, 1998) suggested that it had widened by 137 m a^{-1} over the last 30 years (Section 2.3.5). However, this has been accompanied by a decrease in velocity and so basal drag must be relatively quite significant. For example, it could be possible that the ice stream widened onto sticky spots. This is particularly interesting, given the fact that the margins of Ice Stream D have been shown to overlie the boundary of a sedimentary basin. If Ice Stream D were to widen beyond this boundary perhaps it too would slow down.

In addition, it could be argued that ice stream margins are self-stabilising. This is because the shear zones alter the ice crystal orientation, thereby creating an easy-glide fabric (Kleman and Borgström, 1994). What is clear, however, is that large changes in ice stream behaviour can result from small changes in external forcing or internal dynamics (Bindschadler and Vornberger, 1998). This is what has also been demonstrated by the ice streams investigated in this thesis and it has important ramifications for abrupt changes in ice sheet configurations.

8.3.2.2. Effect on Ice Sheet Configuration.

The M'Clintock Channel Ice Stream would have had a profound effect on the configuration of the Laurentide Ice Sheet and yet most reconstructions of the Laurentide Ice Sheet place an ice divide in the channel. This ice divide is thought to have represented a stable feature of the Laurentide Ice Sheet and is a permanent feature of Dyke and Prest's (1987a) reconstruction from 18,000 yr BP, right up to deglaciation between 10,000 and 9,000 yr BP. Even accounting for soft sediments and deformable bed conditions, the model of Fisher *et al.* (1985) predicts an ice divide in the M'Clintock Channel.

Clearly, the evidence presented in this thesis (Chapter 6), and Hodgson (1994), suggests that an ice stream operated within the channel between 10,400 and 10,000 yr BP. It is suggested in Section 6.5.5 that although a major ice divide may have lay in the channel for a significant time following the Last Glacial Maximum, it cannot be accounted for at around 10,000 yr BP given the weight of evidence in favour of ice stream activity. Rather, it is suggested that sea-level rise led to a destabilisation within the channel which triggered an ice stream. Any ice divide that lay in the channel prior to this event would have provided a large catchment of ice, which was rapidly drained away at the ice stream terminus via an ice shelf or calving front in Viscount Melville Sound.

This is exactly the situation envisaged by Mayewski *et al.* (1981) who provided a glaciological reconstruction of the Laurentide Ice Sheet based on a review of the geological data from the ice sheet bed, coupled with insights from contemporary ice sheets in Antarctica. They suggested that climatic warming after 14,000 yr BP led to the disintegration of the Eurasian ice sheets which produced an increase in global sea levels. Their ice sheet reconstruction depicts a dome (or divide) in southern M'Clintock Channel which feeds an ice stream, initially in M'Clure Strait between 21,000 and 17,000 yr BP. This ice stream (and the divide) retreat southwards as the margin retreats, implying that the ice stream propagated inland as the margin retreated between 17,000 and 10,000 yr BP.

While their reconstruction supports the notion of an ice stream in M'Clintock Channel, it is suggested that the longevity of the ice stream may be overestimated. Rather, it is postulated that an ice divide did indeed lie over the M'Clintock Channel, possibly as late as 11,000 yr BP, and this scenario would account for the older westward flows on Victoria Island. If an ice stream lay in the channel throughout the late Wisconsinan, it is difficult to account for these flow patterns unless they are much older (i.e. pre-Glacial Maximum, ca. 18,000 yr BP). However, the dating constraints imposed by Hodgson (1994) would appear to suggest that the older westerly flows on Victoria Island occurred between 13,000 and 10,400 yr BP.

Alternatively, there could have been several episodes of ice streaming along the axis of the channel (northwards) which were punctuated by slower-flowing ice to the east and west when an ice divide built up. This situation is analogous to the binge/purge cycle postulated by Alley and MacAyeal (1994) to account for the cyclic behaviour of the Hudson Strait Ice Stream. If the binge/purge hypotheses proves viable, then perhaps the M'Clintock Channel Ice Stream displayed similar oscillatory behaviour. Unfortunately, the terrestrial evidence of its activity is solely related to its final cycle of operation. However, marine investigations in M'Clure Strait and the western Arctic Ocean, may provide fruitful answers regarding the temporal nature of its activity. Was this ice stream a one-off deglaciation event, or did it have a cyclical regulatory role in the ice sheet?

The final impact of this ice stream during deglaciation was to force the documented southward and eastward migration of the Keewatin Ice Divide (e.g. Cunningham and Shilts, 1979). This would have had a significant impact on the whole ice sheet, forcing the Keewatin dome south-eastwards towards Hudson Bay, a region prone to unstable ice sheet downdraw from the huge Hudson Strait Ice Stream.

The south-eastward migration of the Keewatin Ice divide probably occurred immediately prior to the Dubawnt Lake Ice Stream activity and may have been a pre-requisite for its rapid demise. Although the ice stream and therefore its catchment area were smaller than the M'Clintock Channel Ice Stream, its relative significance in a relatively smaller sector of the ice sheet should not be underestimated.

Retreat of the Keewatin Ice Sheet margin between 9,000 yr BP and 8,400 yr BP was rapid (cf. Dyke and Dredge, 1989) and it is suggested that the Dubawnt Lake Ice Stream was directly responsible for the final deglaciation of this sector of the ice sheet. If this ice stream had not developed, this sector of the ice sheet may have deglaciated far more slowly. In contrast, it rapidly shrank back towards Hudson Bay, where subsequent marine incursion led to the final collapse of the Laurentide Ice Sheet.

8.3.2.3. Synchronous or Asynchronous Response?

The timing of the M'Clintock Channel Ice Stream between 10,400 and 10,000 yr BP occurred during a period of ice sheet readjustment and was coeval with ice streaming in Hudson Strait (Heinrich event-0 occurring at 11-10 ka). Following this, other margin advances and stillstands have been documented, most notably the Gold Cove advance across Hudson Strait at 9,900 - 9,400 yr BP and the Cockburn Substage, which is bracketed between 9,000 and 8,000 yr BP (Miller and Kaufman, 1990; Kaufman *et al.*, 1993).

The Dubawnt Lake Ice Stream probably occurred between 9,000 and 8,500 yr BP and reflects an ice sheet wide stillstand prior to 8,500 ka. Indeed, it is suggested in Chapter 7 that the MacAlpine moraines (dated at 8,500 yr BP) reflect the maximum extent of the Dubawnt Lake Ice Stream during the Cockburn Substage, see Figure 7.22. Were these ice streams part of a pan-ice sheet destabilisation, or did they act independently?

The overall retreat of the Laurentide Ice Sheet between 19,000 and 15,000 yr BP has been attributed to increased hemispheric summer insolation and is thought to be a direct response to climatic forcing. However, as Clark P.U. (1994) noted, the rapid margin fluctuations after 15,000 yr BP cannot be directly linked to climatic forcing and he proposed that the rapid oscillations of the Laurentide Ice Sheet (and hence ice streaming), were associated with deforming sediment dynamics beneath the ice sheet.

In favour of this argument, Clark P.U. (1994) argued that areas of the Laurentide Ice Sheet which rested on deformable sediments displayed a highly lobate pattern, e.g. the southern and south-western Laurentide Ice Sheet. This is in contrast to areas of the ice sheet which rested on a rigid bed and displayed a more uniform pattern of retreat. Secondly, the documented ice sheet-wide advances at 21-19,000 yr BP and 15-14,000 yr BP (Heinrich events 2 and 1, respectively) only occurred over areas of deforming sediment, specifically in Hudson Strait and along the southern and south-western margin of the Laurentide Ice Sheet. Thirdly, the asynchronous response of ice sheet advances after 14,000 yr BP indicate a non-climatic response, but rather, scale specific responses of individual ice drainage basins. It is suggested that once the ice sheet margin retreated onto a 'hard' bed at around 10,000 yr BP, stable ice sheet behaviour prevailed, except in and around Hudson Strait.

Investigations on the M'Clintock Channel Ice Stream would appear to support the hypothesis of Clark P.U. (1994). This ice stream operated on an area of soft deformable sediments and it thus represented a margin of the ice sheet which was inherently unstable. Once the trigger for ice streaming was pulled (i.e. rise in sea level and an oversteepened ice sheet profile), ice streaming occurred. However, the possibility that the M'Clintock Channel Ice Steam operated more than once can not be ruled out. This situation happened to a number of ice sheet margins which rested on deformable beds between 15,000 and 10,000 yr BP and their response was not synchronous (cf. Patterson, 1997). The M'Clintock Channel Ice Stream represents the unstable response of an ice sheet resting on deformable sediments.

Evidence presented in this thesis (Chapter 7) suggests that ice streaming (and hence unstable ice sheet behaviour) not only occurred after 10,000 yr BP, but that it also occurred away from Hudson Strait and on an area *not* underlain by deformable sediments. It is suggested that this ice stream reflects an *ice sheet-wide* stillstand at around 8,500 yr BP (the Cockburn Substage), which is evidenced by end moraine systems from a number areas (see Dyke and Dredge, 1989).

These end moraines were mapped by Falconer *et al.* (1965) and dated to delimit the margins of the Laurentide Ice Sheet between 8,000 and 9,000 yr BP. Field investigations and aerial photographs revealed nearly 2240 km of end moraines which extended from eastern Keewatin (the MacAlpine moraines) across Melville Peninsula and on to Baffin Island. Figure 8.3 shows the locations of the major moraine systems and Falconer *et al.*'s (1965) postulated coverage of the Laurentide Ice Sheet.

Both the Cockburn moraines (on Baffin Island) and the Chantrey Inlet moraines (in northern Keewatin) are thought to represent periods of enhanced snowfall (Andrews, 1989; Dyke, 1984). The continuity of the moraine systems would thus suggest a pan-ice sheet response to a change in climate. It follows then, that the MacAlpine moraines and hence Dubawnt Lake Ice Stream activity, also reflect this climatically induced stillstand. This lends support to the hypothesis that the Dubawnt Lake Ice Stream was triggered by climate, see Section 7.5.1.

In conclusion, it would appear that the causes of the M'Clintock Channel and Dubawnt Lake Ice Stream activity were not linked. The M'Clintock Channel Ice Stream represents the non-steady response of an ice sheet resting on a layer of deformable sediments. Its timing coincided with a period of significant ice sheet readjustment at a time when ice streaming was common along margins which overlaid deformable sediments. In contrast, the Dubawnt Lake Ice Stream did not rest on deformable sediment, and its activity was almost certainly induced by climatic amelioration during ice sheet disintegration. It is unique in that it represents the only documented ice stream activity away from Hudson Strait after 10,000 yr BP. Whether other ice streams operated on the Canadian Shield during this time deserves particular attention.

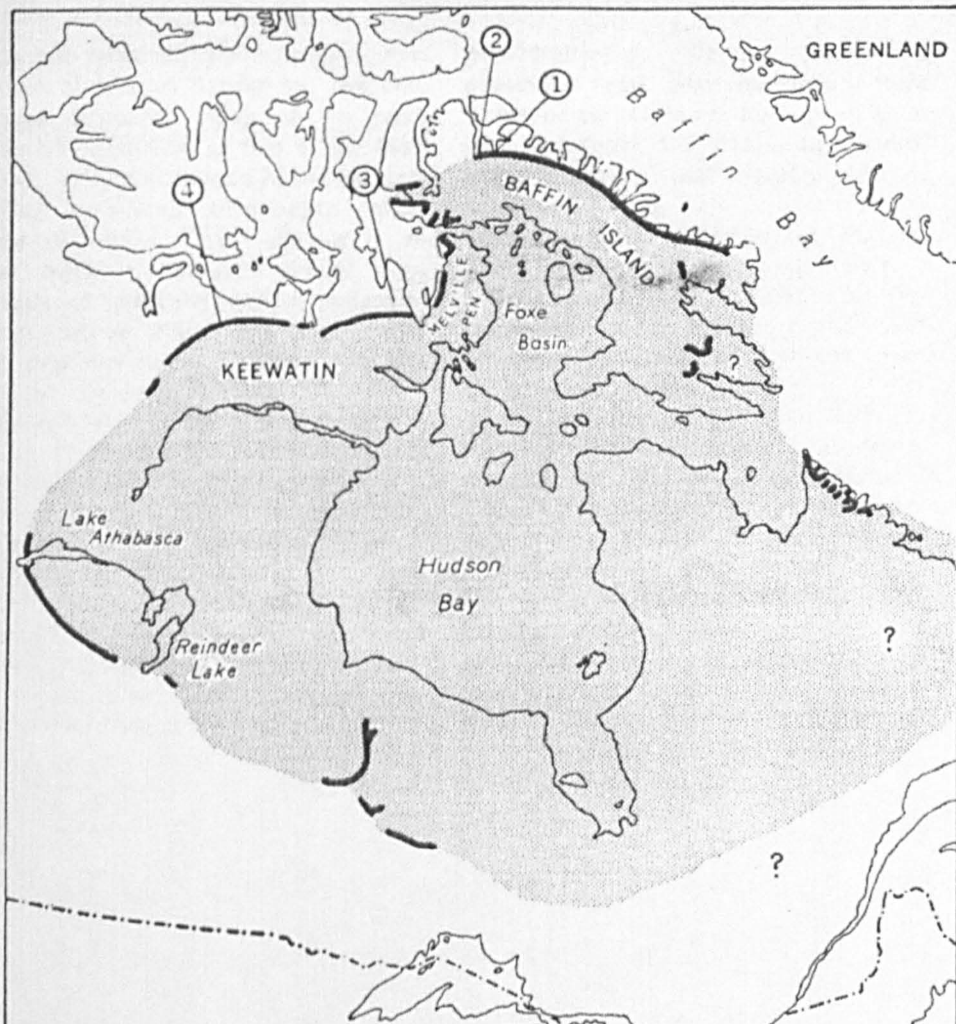

Figure 8.3. Generalised positions of the major moraine belts (thick black line) identified by Falconer *et al.* (1965): (1, 2), Cockburn age moraine system; (3) Chantrey Inlet moraines; (4) MacAlpine moraines, (taken from Falconer *et al.*, 1965). Compare with Figure 7.22 for location of Dubawnt Lake Ice Stream.

8.3.3. Ice Streams in Ice Sheets: Stable or Unstable Behaviour?

From the discussion above, it is clear that ice streams played a key role in the deglaciation of the Laurentide Ice Sheet. Of greater importance, is the recognition that some ice streams were switched on by geological conditions at the ice sheet bed, whereas the behaviour of others is related more to external climatic forcing. In addition, some ice streams may result from internal ice sheet instabilities, triggered by both geological and climatic factors. This raises a number of pertinent questions regarding the stability of ice stream behaviour. Are all ice streams potentially unstable? How many ice streams can an ice sheet support? Does an ice sheet need ice streams to deglaciate?

Oppenheimer (1998) noted that the potential instability of ice stream behaviour is a controversial subject. As outlined in Section 2.6, it is still not known whether ice stream activity is (a), a symptom of ice sheet instability, (b), a transient behaviour that may switch off prior to ice sheet collapse, or (c), a permanent process that links the inland ice to the ice shelves (Oppenheimer, 1998). He suggested that the answer to this questions lies in the extent of the basal lubrication beneath ice streams.

Basal lubrication was incorporated into MacAyeal's (1992) numerical model of the West Antarctic Ice Sheet, which appeared to suggest that rapid ice streaming, and hence ice sheet collapse may be sporadic and possibly chaotic. This is because the timescales of climate forcing are non-synchronous with the timescales of readjustment for an ice sheet which rests on a deformable bed. An interesting facet of this modelling, is that it implied that ice sheet collapse, and hence ice streaming, is not solely controlled by climate forcing. Rather, it is the distribution of deformable till that pre-disposes the ice sheet to ice streaming.

This supports the theory of Clark, P.U. (1994), that periods of Laurentide Ice Sheet instability were a direct result of an ice sheet whose margins lay on a deformable bed (see Section 8.3.2.3). This unstable behaviour is also evidenced by Heinrich events in the North Atlantic sedimentary record (Section 3.2.2). Bond *et al* (1992) found that the six most recent Heinrich events (at ~14,300 (H1); ~21,000 (H2); ~28,000 (H3); ~41,000 (H4); ~52,000 (H5); ~69,000 (H6) yr BP) were characterised by decreases in sea surface temperature, salinity and planktonic foraminifera, and most importantly, massive increases in iceberg discharge from the Laurentide Ice Sheet.

They also found that although the Heinrich events are characterised by low sea surface temperatures (and hence decreased foraminifera), the iceberg rafted debris only represented a narrow band within the sediment layers. This led them to tentatively postulate that shortly after a drop in sea surface temperatures, ice streams in eastern Canada began to advance rapidly, exporting huge amounts of icebergs. Indeed, evidence has been found which directly links the first two Heinrich events (H1 and H2) to the activity of an ice stream in Hudson Strait (Andrews and Tedesco, 1992).

The importance of these findings is that they illustrate considerable ice sheet instability and probable episodes of vigorous ice streaming. Of greatest importance, however, is the fact that these periods of ice sheet instability do not correspond to periods of climate change, occurring at shorter intervals (7,000 year cycles) than the orbital precession cycles (19,000 and 21,000 years) predict (cf. Broecker *et al.*, 1992).

In an attempt to model this complex behaviour, MacAyeal (1993) proposed a binge-purge oscillation for the Laurentide Ice Sheet which is controlled by the basal thermal regime of an ice stream in Hudson Strait. The 7,000 year cycle is thus determined by the time taken for an ice sheet to change its bed from a frozen to a thawed state. Once the bed of the Hudson Strait Ice Stream becomes thawed, enhanced lubrication of soft sediments causes rapid ice stream motion and profligate iceberg discharges into the North Atlantic. The purge phase thins the ice sheet very rapidly (perhaps in as little as 250 years) and leads to frozen bed conditions. This signifies the end of the purge phase and ice sheet growth begins its 7,000 year binge cycle.

The appeal of MacAyeal's (1993) model lies in its success in predicting the observed periodicity of Heinrich events detected from the marine record. This model (and the ocean sedimentary record) would thus support the notion that ice streaming is inherently unstable over deforming sediments. If this is the case, there are two major implications. Firstly, ice streams will tend to develop in those areas of an ice sheet underlain by soft sediments, and secondly, the amount of sediment available will dictate the life-cycle of the ice stream. However, this is further complicated by the basal mechanics of ice streams.

If an ice stream moves over a deeply deforming till, then it will transport that till in a downstream direction, depositing it at the terminus (e.g. Alley *et al.*, 1989). However, if, as has been suggested (Engelhardt and Kamb, 1998), that deformation is not as pervasive as previously thought, but rather, that the ice stream slides across its slippery surface, then ice stream activity can occur over longer timescales, without sediment exhaustion. This has implications for contemporary ice streams in West Antarctica. At least two ice streams have been shown to be underlain by deformable till and the margin and onset zone of Ice Stream D coincide with a sedimentary basin (see Anandakrishnan *et al.*, 1998; Bell *et al.*, 1998). Dictating (a) the extent and thickness of the sedimentary basin, and (b) the exact flow mechanism, will provide a useful estimate of their life-cycle and hence, potential for instability.

The disposition of the basal till is not the only control on ice stream activity because Ice Stream C has recently shut down despite being underlain by a layer of deformable till whose composition is very similar to that beneath Ice Stream B (cf. Anandakrishnan and Alley, 1997b). Instead, it has been suggested that Ice Stream C was eating its way inland and reached a point where the bed diverted water into Ice Stream B (Anandakrishnan and Alley, 1997a). The loss of water alone would not be able to shut down the ice stream, but it is inferred that a slight loss in lubrication exposed the ice stream bed to sticky spots in the downstream area (Anandakrishnan and Alley, 1997a). Thus, if Ice Stream C only shut down because of a chance coincidence of the bed geometry, other ice streams could propagate upstream without shutting down.

This process is not so simple, however, and some modelling experiments appear to show that ice streams are more ephemeral and that their activity plays more of a regulatory role. For example, if an ice sheet resting on a deformable bed conforms to the binge/purge cycle, ice stream activity would be preventing it from thickening, by rapidly draining away large quantities of ice from the interior. This would suggest that the ice streams are playing a regulatory role, preventing the whole ice sheet from building up and collapsing under its own weight.

The regulatory role of ice streams is supported by the modelling work of Payne and Dongelmans (1997) and Payne (1998; 1999) which suggested that neighbouring ice streams out-compete each other by capturing drainage basins. Payne has successfully modelled the location of the Siple Coast ice streams using a three-dimensional thermomechanical model (Section 2.5.3). This work suggested that ice streams arise when changes in bed topography caused localised accelerations in ice flow, increased friction, warmer ice and hence faster ice flow. This positive feedback mechanism allows ice streams to propagate inland because the lowering of the surface profile preferentially captures more ice from upstream and adjacent areas. However, when neighbouring ice streams compete for the same catchment area, the stronger ice stream will dominate. This leads to slower ice flow in the weaker ice stream and hence more friction, cooler ice and a further reduction in speed, eventually leading to ice stream stagnation.

Payne and Dongelmans (1997) and Payne (1998) cited field evidence from Ice Streams B and C to support this self-organisation, but acknowledged that the current paradigm for Ice Stream C shut-down favours a hydrological explanation. Nevertheless, Payne (1997) argued that the thermodynamics of ice streams and their basal meltwater are intimately linked and that long term changes in thermodynamics may well trigger short term hydrological changes.

The view that ice streams buffer the inland ice from collapse has been suggested by Blankenship *et al.* (1993). They used aerogeophysical techniques in West Antarctica and found that the onset zones of the ice streams coincided with an area of active volcanism beneath the ice sheet. This suggests that the heat supplied to the base of the ice sheet increases basal melting and provides a lubricating basal water layer which may trigger ice streaming. This situation has also been inferred to account for several palaeo ice streams draining the Icelandic Ice Sheet (cf. Bourgeois *et al.*, 2000).

Blankenship *et al.* (1993) speculated that the ice streams may not be able to propagate any further inland. Thus, if the ice stream termini retreat back to this geological boundary, the slow-moving interior of the ice sheet would be exposed to the ocean without the buffer of the ice stream system. This would represent a very unstable situation and implies that ice sheet collapse could take place irrespective of climate forcing (Blankenship *et al.*, 1993).

Some workers suggest that the inland migration of ice streams has already begun, and that ice sheet collapse may be inevitable. Conway *et al.* (1999) have dated the grounding line positions of the West Antarctic Ice Sheet in front of the Siple Coast ice streams throughout the Holocene. Their results indicated that the grounding line position of the ice sheet has receded by an average of 120 m a^{-1} during the last 7,500 years and this is supported by recent observations of grounding line retreat from Ice Streams B (Bindschadler and Vornberger, 1998) and C (Thomas *et al.* 1988).

Conway *et al.* (1999) argued that contemporary grounding line retreat is part of an ongoing recession which was triggered in the early Holocene. As such, the Siple Coast ice stream activity may be the fundamental mechanism through which the West Antarctic Ice Sheet will collapse. This collapse may not be catastrophic but complete disintegration within the present interglacial could be inevitable (Conway *et al.*, 1999).

In conclusion, although current research does not indicate that the ice sheet is retreating rapidly, areas of rapid change have been discovered on the Siple Coast ice streams (e.g. Bindschadler and Vornberger, 1998). It is important to note that deglaciation of Hudson Strait took some 6,000 years, although the ice stream was draining a significant fraction of the Laurentide Ice Sheet. Thus, ice stream behaviour is not always catastrophic. Indeed, some workers would argue that ice streams play a regulatory role. However, it is clear that when an ice sheet deglaciates, ice streams play a key role. More importantly, the Laurentide Ice Sheet continued to deglaciate once it had retreated back from the areas underlain by soft sediments and evidence in this thesis (Chapter 7) suggests that ice stream activity continued to operate over relatively hard bedrock. It becomes clear then, that whether or not ice stream activity is stable or unstable over deforming sediments, it is in no way restricted to such areas during deglaciation. Climatic forcing has a crucial role to play in unstable ice stream behaviour.

8.4. Factors Controlling Ice Stream Location, Initiation and Functioning.

It is clear from the preceding discussion that ice streams have a profound impact on ice sheet configuration and stability. It is also apparent that several factors, both external to and internal to the ice sheet, affect the location, initiation and functioning of ice streams. These can be grouped into a conceptual model outlining the primary controls on ice stream activity, shown in Figure 8.4.

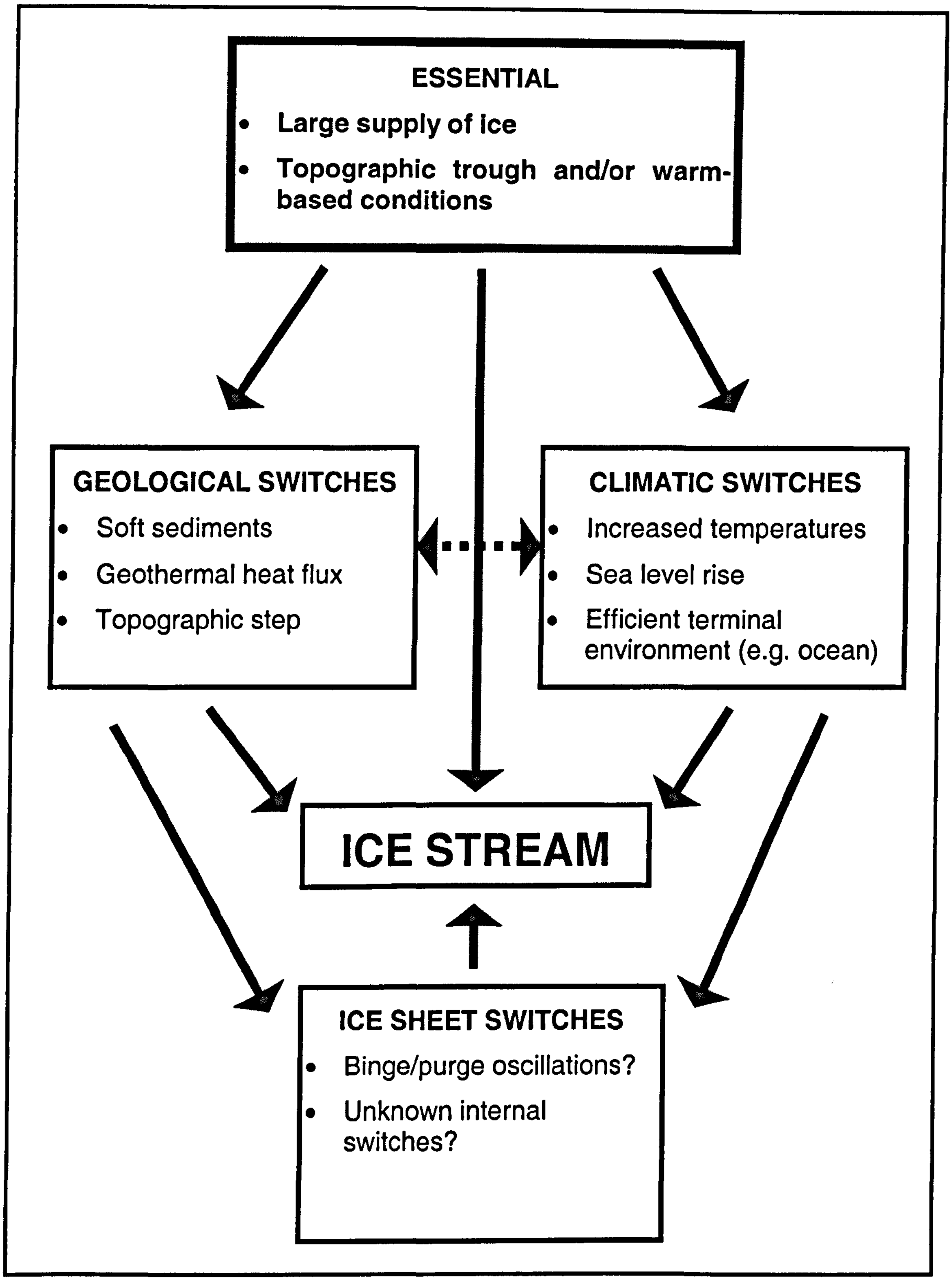

Figure 8.4. Conceptual diagram outlining the principal controls on ice streaming. Essential pre-requisites for ice streaming are a large supply of ice together with a topographic trough and/or warm-based conditions. In addition, several geological or climatic switches may encourage ice stream activity. Note that a combination of climatic or geological switches may be acting to trigger ice streaming and that they are not separate pathways (dashed line). These may also trigger internal ice sheet instability mechanisms, which may also lead to ice streaming.

It can be seen in Figure 8.4 that some factors are essential to ice stream activity, whereas others are conducive. Because of their rapid ice flux, ice streams will not develop unless they have a large supply of ice upstream. Once this pre-requisite is met, two types of ice stream may develop, a topographic ice stream or a pure ice stream. An essential factor for a topographic ice stream is a bedrock trough which preferentially captures ice from a wide source area. This convergent flow leads to a localised flow acceleration, greater friction, warmer ice and hence rapid ice flow. In addition, the deeper ice within a bedrock trough generates more pressure melting at the bed and increased amounts of lubricating water. Topographic troughs may also be preferential sites for sediment and meltwater accumulation. An essential pre-requisite for a pure ice stream is a melting bed, which encourages rapid basal sliding and/or pervasive deformation of underlying soft sediments.

In theory, these pre-requisites can lead to ice streaming with no other influences. However, glaciological regulation may inhibit ice stream activity and so it is more usual for several other factors to encourage or maintain fast ice flow. These factors are conducive to ice stream activity but are by no means essential. It is best to view them as a number of switches which are capable of turning an ice stream on. If enough of these switches are pressed at once, then ice streaming may be inevitable. These switches can be grouped into three broad categories, geological switches, climatic switches, and ice sheet switches, see Figure 8.4.

It is clear that certain geological factors are conducive to ice streaming. For example, the availability of soft sediments beneath an ice stream has a significant impact on the flow mechanisms and dynamics of ice stream behaviour. Highly saturated soft sediments may deform, thus providing little friction to the overlying ice. Alternatively, a layer of saturated soft sediments may provide a slippery surface over which the base of the ice stream can slide. Evidence from West Antarctic ice streams (Bell *et al*,. 1998) and from palaeo-ice streams (Chapter 6, and Clark and Stokes, in press) would appear to suggest that the availability of soft sediments is a key influence on ice stream initiation and functioning.

Another geological factor which may 'switch on' an ice stream is that of volcanic activity. Geothermal heat from volcanic activity beneath the ice sheet may warm the base of the ice, thus leading to increased meltwater production and an increase in basal sliding/till deformation. Volcanic activity has been associated with the onset zones of contemporary ice streams in West Antarctica (Blankenship *et al.*, 1993) and has also been inferred to have influenced ice stream locations in the former Icelandic Ice Sheet (Bourgeois *et al.*, 2000). As such, volcanic activity can be seen as another geological switch for ice stream activity.

The topography of the bed is an important geological control on the development of topographic ice streams, but pure ice streams may also be initiated by a step in topography in the inferred onset zone. Such steps produce a localised flow acceleration, warmer ice, more viscous ice and hence increased flow. This positive feedback mechanism may be responsible for the development of some ice streams and the locations of the Siple Coast ice streams have been successfully modelled solely on the basis of variations in bedrock topography (Payne and Dongelmans, 1997; Payne, 1998; 1999). Indeed, McIntyre (1985) noted that the onset zones of many Antarctic ice streams coincided with a topographic step and investigations in this thesis (Chapter 7) have shown that a topographic step influenced the fast flow of the Dubawnt Lake Ice Stream, see Section 7.5.1.

The geological switches, described above, have an immediate local effect on the flow of an ice sheet and the initiation/maintenance of ice streaming. Because of this, their influence can be easily observed and even quantified beneath contemporary ice sheets. They are described as 'digital switches' because they are either present or absent, their effect is either on or off. In contrast, climatic factors which influence ice streaming are far more complicated and difficult to observe. They are described as 'analogue switches', because they operate over longer timescales and their affect may be far more gradual and complicated by thresholds.

Climatic factors which may affect ice streaming include increased temperatures or increased accumulation. Increased temperatures will serve to increase the melting of an ice sheet, perhaps leading to an increased amount of lubricating water at the bed, or a change in the basal thermal regime of the ice sheet. Alternatively, increased accumulation may lead to an oversteepened ice sheet profile and increased pressure melting at the bed which may trigger ice streaming.

Changes in relative sea level are another important factor concerning ice sheet stability and ice streaming. An increase in sea level may destabilise a previously grounded ice sheet by reducing, or even removing the basal friction. In the same way, a marine terminus provides an efficient terminal environment for removing ice, thus accentuating marine downdraw.

These climatic factors are more difficult to observe and understanding and predicting their influence is very difficult. Although they are described as analogue switches which may operate over longer timescales, they are further complicated by feedback mechanisms and critical thresholds which are difficult to ascertain. For example, increased accumulation and ice sheet thickening may operate over a long timescale before it can affect the basal thermal regime of the ice sheet. Moreover, these climatic factors are strongly coupled to the behaviour of the ice sheet, and this introduces a third factor controlling ice streaming, that of internal ice sheet instabilities.

Internal ice sheet instabilities are the most complex control on ice streaming and result from the combined effects of the ice sheet with geological and/or climatic switches. An example of an ice sheet instability switch is the binge/purge hypothesis of ice sheet surging, described above in Section 8.3.3 (MacAyeal, 1993). This results from the complex interactions between the ice sheet thickness, the underlying sediments and the basal thermal regime of the ice sheet. Despite the fact that this hypothesis fits well with the available data, it is totally unproven as an actual mechanism. The problem is, that the geological, climatic and ice sheet switches operate on a number of different timescales. It is because of these temporal variations and feedback mechanisms that ice sheet collapse is sporadic and unpredictable (MacAyeal, 1992).

It is suggested here, that the Laurentide Ice Sheet collapsed as a result of both geological and climatic switches being turned on. Initially, the ice sheet was being drained by topographic ice streams and a number of pure ice streams which were switched on by the geology of the ice stream bed. The M'Clintock Channel Ice Stream is a good example of a geologically controlled ice stream at this time. These ice streams are fixed in time and space by the amount of available sediment. This means that unless an unlimited supply of sediment is available, they are not capable of deglaciating a whole ice sheet. In the case of the Laurentide Ice Sheet, geological ice streams were responsible for the shrinkage of the ice sheet but they ceased to operate once the ice sheet margin retreated onto hard bedrock. Deglaciation of the ice sheet also required an associated climatic forcing and the development of several ice streams which were switched on by climatic factors. The Dubawnt Lake Ice Stream is a good example of one such ice stream.

In terms of the stability of the West Antarctic Ice Sheet, it is suggested that collapse will only take place if further ice stream activity is forced by climatic switches or switches controlled by the internal dynamics of the ice sheet. Presently, it would appear that the ice sheet is being regulated by topographic ice streams and a number of geologically controlled ice streams, e.g. the Siple Coast ice streams. To deglaciate the West Antarctic Ice Sheet requires that several other ice streams are switched on by climatic factors and/or an ice sheet instability mechanism causes a major destabilisation. This may involve an increase in the number of ice streams within the ice sheet, or an increase in size of existing ice streams. A major climatic switch may be imminent or it could be gradually reaching an irreversible threshold. On the other hand, there may be a switch internal to the ice sheet which could be pressed irrespective of climatic factors but may be a simple response to the ice streams which are currently eating away at the inland ice.

8.5. Suggestions for Future Research.

From the discussion above, it is clear that our understanding of ice stream behaviour is incomplete. There are still too many unknowns regarding the behaviour of ice streams. Nevertheless, this thesis has demonstrated the beneficial use of palaeo-ice stream signatures to advance our understanding of these dynamic phenomena. Furthermore, this thesis has served to highlight several research areas which may provide fruitful answers to some of the more puzzling questions regarding ice stream behaviour.

This chapter has highlighted the importance of marine investigations in the foregrounds of contemporary ice streams. The aim of these investigations are to provide a refined reconstruction of the behaviour of the ice streams over longer timescales than contemporary observations permit. By revealing their submarine geomorphology, it is possible to elucidate when these ice streams reached their maximum extent and determine the precise nature of their subsequent retreat. However, this interpretation relies heavily on terrestrial analogues of ice stream geomorphology. This thesis has reconstructed the size and extent of two isochronously generated ice stream imprints, but further research on terrestrial ice stream imprints is necessary to improve our understanding of the geomorphology in the foregrounds of contemporary ice streams. In particular, more work is required on time-transgressive ice stream imprints. In doing so, it will be possible to unravel the dynamic history and temporal variability of contemporary ice streams.

In conjunction with the submarine research in the foregrounds on contemporary ice streams, it could also be worthwhile to look for ice stream geomorphology in the straits and sounds which bordered former ice sheets. Traditionally, terrestrial evidence has dominated the search for palaeo-ice streams. More recently, far travelled evidence of ice sheet discharges (Heinrich events) has been used to strengthen these hypotheses. With recent advances in marine geophysics, it is now possible to search for evidence of ice streams on the continental shelves which link the terrestrial and ocean sedimentary record. For example, submarine investigations in M'Clintock Channel and Viscount Melville Sound will help to test the hypotheses regarding the timing and behaviour of the M'Clintock Channel Ice Stream. Did the ice stream operate more than once? Have bedforms been generated within the channel? Does their pattern match the terrestrial record described in this thesis?

Since the discovery of a layer of highly saturated till beneath some contemporary ice streams, there has been much debate about its role in the basal flow mechanics. Does the layer of till pervasively deform, or does it provide a slippery surface across which the ice stream bed can slide? Elucidating the basal mechanics of ice streams is of paramount importance if we are to accurately model and predict their future behaviour. While borehole and seismic observations from contemporary ice streams provide invaluable information about the nature of the bed, palaeo ice stream beds provide an unimpeded view. Therefore, future research on palaeo-ice stream beds may help to ascertain the exact role of a deformable substrate. Samples can be taken easily and cheaply from both inside and outside the ice stream bed, and micromorphological till fabric analysis can provide information regarding the deformation history of the material.

Few workers doubt that the presence of a highly saturated till can facilitate fast ice stream motion. However, the role of basal drag in restraining ice streams has caused much greater conjecture. In particular, the role of sticky spots in retarding flow is not fully understood. It is difficult to assess what the relative importance is of areas of discontinuous till as opposed to bedrock bumps protruding into the base of the ice stream. Future research on palaeo-ice stream beds may be able to pin-point areas of the bed which acted as potential sticky spots. Moreover, it may be possible to ascertain the exact seed point of ice stream shut-down and how this propagated to the rest of the ice stream.

The more palaeo-ice streams we can investigate, the more we can learn about their behaviour. In particular, future research should concentrate on the role of ice streams during deglaciation. The disappearance of former ice sheets at the end of the last ice age provides our only case study of how ice sheets deglaciate. Only by studying their demise, will it be possible to assess the delicate state of contemporary ice sheets. The more ice streams we can find the more we can understand their role in deglaciation and abrupt climate change. Questions that need answering include; does an ice sheet need ice streams to deglaciate? Do ice streams increase in number during deglaciation or do they just get bigger? What inhibits ice stream activity?

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Chapter 9: Conclusions.

9.1. Identifying Palaeo-Ice Stream Geomorphology.

The importance of palaeo-ice streams can be summarised in four main points (Section 3.2);

- Because of their profound impact on ice sheet configurations we need to know where they were in order to accurately reconstruct former ice sheet histories.
- Their behaviour is inextricably linked to the ocean-climate systems where it has been demonstrated that not only do ice sheets respond to climate change, but that ice streams are responsible for forcing some of the most abrupt climate flips ever identified.
- Palaeo-ice streams provide an unprecedented opportunity to observe the bed of an ice stream, thus alleviating one of the main inhibitors to contemporary ice stream research.
- Some ice streams are powerful erosional agents. They are responsible for rapid landscape evolution and for producing some of the world's largest sedimentary fans.

A large number of workers have recognised the significance of finding palaeo-ice streams in formerly glaciated environments. Chapter 3 of this thesis represents the first ever review of palaeo-ice stream research in the literature. It highlighted several inhibitors to palaeo-ice stream research and revealed that our understanding of ice stream geomorphology is limited (Section 3.4). A result of this, has been that hypothesised locations have tended to outweigh meaningful evidence for their existence. In recognition of this problem, eight geomorphological criteria have been predicted to aid the identification of palaeo-ice streams (Chapter 4; Stokes and Clark, 1999);

1. Characteristic shape and dimensions (>20 km wide; >150 km long).
2. Highly convergent flow patterns.
3. Highly attenuated subglacial bedforms (length:width ratio >10:1).

4. Boothia-type erratic dispersal trains (Dyke and Morris, 1988).

5. Abrupt lateral margins (<2 km).

6. Ice stream marginal moraines.

7. Glaciotectonic and geotechnical evidence of pervasively deformed till.

8. Submarine till delta or sediment fan (only marine-based ice streams).

These criteria are developed from the known characteristics of contemporary ice streams, thereby introducing an objective basis for their identification. Although it is difficult to validate the individual criteria, recent submarine investigations from the foregrounds of contemporary ice streams would appear to support their diagnostic use (Section 8.2.2). Of greater significance is the notion that the individual criteria can be grouped together to comprise a characteristic landsystem of ice stream activity, shown in Figures 4.4 and 4.5 (Chapter 4). These conceptual models provide an observational template for future palaeo-ice stream research.

The application of the geomorphological criteria is two-fold. Firstly, the presence or absence of the criteria can be used to assess the credence of already hypothesised palaeo-ice stream locations. Secondly, the criteria can be used to search for evidence of palaeo-ice streams which have yet to be identified. Using satellite imagery and aerial photography to map glacial geomorphology, this thesis (Chapters 6 and 7, respectively) has successfully demonstrated both applications of the criteria to palaeo-ice stream research, see below.

9.2. The M'Clintock Channel Ice Stream.

The M'Clintock Channel Ice Stream was first postulated by Hodgson (1994) who inferred its activity from a spectacular drumlin field with an extremely abrupt margin on Victoria Island in the Canadian Arctic. Analysis of the bedform pattern using satellite imagery (Landsat TM and Synthetic Aperture Radar), aerial photography and Digital Elevation Models (DEM's) revealed that five of the geomorphological criteria were present. The flow-set revealed the characteristic shape and dimensions, displayed a highly convergent onset zone, was characterised by highly attenuated bedforms, and had an extremely abrupt margin which was demarcated by several ice stream marginal moraines. Detailed investigations of the ice stream bed and surrounding areas led to the following key discoveries;

- The extent of the ice stream is reconstructed at 720 km in length and its width converged from 330 km to 140 km close to the terminus. This is far larger than was originally envisaged by Hodgson (1994) and makes it comparable in size to the Hudson Strait Ice Stream, known to have had a profound effect on the configuration of the Laurentide Ice Sheet.
- The ice stream bedforms are substantially longer, wider and more elongated than any of the neighbouring non-ice stream flow-sets. Maximum values of lineament length (7,950 m), width (750 m) and elongation ratio (27:1) are represented by the ice stream flow-set.
- The margin of the ice stream flow-set is extremely abrupt and there is a change from north-south orientated lineaments to hummocky terrain over a distance of around 100 m.
- The ice stream flow-set depicts a remarkably coherent pattern, with the standard deviation of lineament orientation within 100 km^2 of grid squares yielding a value of only 4.5° along a flow-band. The exceptional parallel conformity and extremely abrupt margin of the flow-set are taken to reflect an isochronous (snapshot) view of the ice stream bed prior to shut-down.
- The morphometry of the ice stream bedforms display several systematic trends in the downstream direction. Close to the ice stream terminus, bedforms become shorter and more closely packed together and hard bedrock becomes progressively more exposed.
- Outside of the ice stream margin, till thicknesses have been estimated at 50 m (Hodgson, 1994). Within the ice stream, close to the inferred terminus, preliminary estimates of sediment thickness suggest a maximum surface lowering of around 45 m which decreases to 35 m at 60 km upstream. Extrapolating these estimates of surface lowering further upstream provides a maximum sediment flux of 73,000 m a^{-1} m^{-1} (i.e. per metre width of the terminus) for the ice stream. This represents a much higher figure than has been estimated for other ice streams (e.g.

contemporary Ice Stream B) but should be no surprise given the larger size and presumably higher velocities.

- The combination of a high sediment flux, fast flow velocity and the creation of bedforms over fine grained sediments is most easily explained by the operation of a deformable till layer beneath the ice stream.
- Using the estimates of surface lowering (and hence, sediment exhaustion) it is inferred that the ice stream experienced frictional shut-down as a result of cutting down through pre-existing unconsolidated sediments until it reached bedrock which possessed higher levels of basal friction. Somewhat surprisingly, it first reached bedrock close to the terminus, retarding flow there, which then propagated upstream.
- It is thus concluded that the location of this ice stream was controlled by the topographic trough, and the availability of soft unconsolidated sediments pre-determined its life-cycle. In this case, 50 m of sediment was required to permit ice stream activity for between 200 and 400 years. It is interesting to note that the onset zone of the ice stream coincides with the boundary between the hard (crystalline) rock and the soft (carbonate) rock.
- In some locations, the ice stream margin is juxtaposed to several ice stream marginal moraines. These ridges maintain a fairly constant width of between 500-800 m, reach lengths of up to 23 km and usually stand between 10 and 40 m above the surrounding terrain. No topographic steps or changes in geology coincide with the ice stream margin.
- It is thus suggested that the width of the ice stream was governed by internal glacio-dynamics. This implies that the margins were not fixed in time or space, and the identification of relict ice stream marginal moraines would appear to reflect subtle switches in margin positions.
- This ice stream would have had a profound effect on the dynamics of the ice sheet. Its activity is constrained to between 10,400 and 10,000 yr BP, and assuming a velocity typical of the Hudson Strait Ice Stream (4 km a^{-1}; Dowdeswell *et al.*, 1995), it could have drained up to 80,000 km^3 of ice from its catchment.

- The long-held assumption of an ice divide within M'Clintock Channel at this time needs re-evaluating. The preferred explanation is that an ice divide lay in M'Clintock Channel prior to ice stream activity but that a destabilisation of the ice sheet margin, perhaps as a result of sea-level rise, may have triggered ice stream activity. The possibility that the ice stream operated more than once can not be ruled out.

The ice stream was a major component of the Laurentide Ice Sheet during deglaciation, accelerating its demise, and probably having far reaching effects on sediment delivery to the continental shelf and on Arctic oceanography. It represents a good opportunity for further investigation of ice stream beds and for validation of numerical modelling of ice stream functioning. Given the inaccessibility of contemporary ice stream beds, this may constitute an effective way of advancing process knowledge.

These findings also have implications for the functioning of contemporary ice streams and the stability of the West Antarctic Ice Sheet. The behaviour of the M'Clintock Channel Ice Stream is further proof of the non-steady response of ice sheets and climate change, and of the enormous effects that ice streams can make to mass balance and stability of an ice sheet. If frictional shut-down of ice streams arising from sediment exhaustion is typical, then it implies a limited time-span of operation for those Antarctic ice streams which are currently operating on, and depleting their limited reserves of basal glaciomarine muds.

9.3. The Dubawnt Lake Ice Stream.

The geomorphological criteria developed in Chapter 4 were used to identify the location of a previously undetected ice stream from the Keewatin Sector of the Laurentide Ice Sheet. Several previous investigations in this area had noted the unique nature of the bedforms but ice stream activity had not been invoked. Geomorphological mapping using satellite imagery (Landsat MSS) detected four out of a possible seven criteria for ice stream activity. The ice stream flow-set fits the characteristic shape and dimensions, displays highly convergent flow patterns, contains highly attenuated bedforms and exhibits very abrupt lateral margins. Detailed investigation of the ice stream bed and surrounding areas led to the following key discoveries;

- This terrestrial ice stream is reconstructed at over 450 km in length and 140 km in width, but the terminus diverges to around 190 km in width.
- Maximum values of lineament length (12,743 m), width (990 m) and elongation ratio (48.3:1) are represented by the ice stream flow-set. These bedforms are more elongated than those associated with the M'Clintock Channel Ice Stream (Chapter 6) and portray some of the most attenuated subglacial bedforms ever identified.
- The southern margin of the ice stream flow-set is abrupt and is bordered by areas of little or no bedforms. Along the northern margin, older bedform patterns have been cross-cut by the ice stream indicating that this flow pattern is the youngest in the area.
- The ice stream flow-set depicts a remarkably coherent pattern despite the convergence and divergence upstream and downstream of the main trunk. For the central portion of the ice stream, over an area of 720 km^2, the standard deviation of lineament orientation does not exceed 3.8°. This exceptional parallel conformity and the abrupt margin of the flow-set are taken to reflect an isochronous (snapshot) view of the ice stream bed prior to shut-down.
- The morphometry of the bedforms depict several trends from the onset zone through to the terminus. The longest bedforms occur approximately half way down the ice stream (with lengths of up to 12.5 km and elongation ratios as high as 48:1) and the remarkable parallel conformity of the lineaments resembles a ridge/groove pattern (see Figure 7.9, Chapter 7).
- It is assumed that variations in lineament elongation ratio provide a useful proxy for ice velocity. The bedform pattern reveals exactly what we would expect from a terrestrial ice stream whose velocity increases in the onset zone, passes through a maximum in the main trunk of the ice stream and slows down as the ice diverges towards the terminus.
- Bedforms with the highest elongation ratio coincide with a drop in elevation where the ice stream entered a sedimentary basin. The topographic step and

availability of sediments encouraged fast ice flow, but were not the primary controls on ice stream location.

- The relatively flat Canadian Shield suggests that the location of the ice stream within the ice sheet lies irrespective of regional topography.
- Although the ice stream lay over a sedimentary basin, the margins of the basin extend considerably beyond the ice stream margins and no elongated bedforms are found there. It is inferred that although the less resistant material (compared to crystalline bedrock) would have encouraged fast ice flow, it did not have a significant influence on the location of the ice stream.
- It is thus suggested that the Dubawnt Lake Ice Stream was climatically induced and that increased accumulation in the onset zone (arising from newly available moisture sources, i.e. marine incursion/proglacial lakes) led to an oversteepened ice sheet profile during a time of climatic amelioration. Climatic warming would also have had a direct effect on the thermal state of the ice sheet bed, probably forcing a change from a cold to warm-based conditions. The formation of a proglacial lake may have triggered ice stream activity but it could not support a substantial calving margin and the ice stream produced a low angled divergent lobe (as evidenced by the bedform pattern).
- It is postulated that the ice stream moved by rapid basal sliding owing to increased amounts of subglacial water. Increased friction would have led to increased water at the bed and higher temperatures would have enhanced the production of meltwater.
- It is suggested that the main ice stream shut down when it ran out of ice. This thinned the ice stream and led to slower flow velocities and increased ablation. This is further evidenced by the shut-down of a tributary in the onset zone of the ice stream. This tributary gradually shrunk before ceasing altogether, probably through a reduction in its catchment area.
- The terminus of the ice stream and its close association with the MacAlpine moraine complex suggests that its activity occurred immediately prior to 8,500 yr BP. Given the reconstructed configuration of the ice sheet at this time, it is

postulated that its activity took place after 10,000 yr BP but probably occurred after 9,000 yr BP.

- The extent of the ice stream (surface area 72,000 km^2) implies that it drained a huge portion of the Keewatin Ice Sheet, lowering the surface profile and probably resulting in a substantial thinning which was followed by widespread deglaciation.

The ice stream was responsible for the obliteration of one of the last major ice centres of the Laurentide Ice Sheet and hints at the unpredictable and catastrophic importance of a climatically induced terrestrial ice stream. The identification of an ice stream tributary which shrunk prior to ice stream shut-down is further evidence for highly dynamic ice stream margins. More importantly, it supports the recent evidence from Ice Stream C (West Antarctica) that ice stream shut-down is not a single event, but involves the stepwise migration of ice stream margins (Jacobel *et al.*, 2000). This implies that subtle changes in margin positions may be the most obvious way of assessing the potential instability of contemporary ice streams in West Antarctica.

9.4. Glaciological Implications.

The documented cases of palaeo-ice stream activity described in this thesis provide several fresh insights regarding the behaviour and functioning of ice streams.

- The M'Clintock Channel and Dubawnt Lake ice streams were considerably wider than their contemporary counterparts. It is inferred that their dominant resistive force was side drag imposed by the slower flowing ice which bordered them. Therefore, the ice streams could increase their width until basal drag became relatively more important (i.e. the M'Clintock Channel Ice Stream), or, ice supply from upstream became the limiting factor (i.e. the Dubawnt Lake Ice Stream). Both mechanisms led to ice stream shut-down.

- The activity of the M'Clintock Channel Ice Stream was pre-disposed by the nature of the underlying geology. Its functioning, and ultimately its life-cycle, was determined by the amount of sediment available. Although an ice sheet resting on a layer of soft sediments is thought to be inherently unstable, unless the amount of sediment is unlimited, these ice streams have a limited time-span of operation. It is suggested that these 'geological' ice streams display a non-synchronous response to climate and may play a merely regulatory role in an ice sheet.

- The location of other ice streams, such as the Dubawnt Lake Ice Stream, is irrespective of the underlying geology and topography, and it is more likely that they are switched on by a change in climate which alters the ice sheet configuration and may result in a shift in the thermal state of the ice sheet bed. Their life-cycle, and therefore their potential for ice sheet instability is only governed by the amount of ice available upstream. As such, this type of ice stream behaviour is far more significant in terms of ice sheet collapse.

- Evidence would appear to suggest that contemporary ice streams in West Antarctica are geologically fixed (see Bell *et al.*, 1998; Anandakrishnan *et al.*, 1998). Their behaviour is thought to be unstable (Bindschadler, 1998) and has been shown to be ephemeral (see Jacobel *et al.*, 2000). It is suggested that their role in any potential ice sheet collapse may be overestimated, unless they have an unlimited supply of sediment available to them.

- Therefore, to collapse the West Antarctic Ice Sheet requires that either (a) existing ice streams increase in size or vigour, and/or (b), several other ice streams are 'switched on'. It is suggested that deglaciation will require climatic switches to turn on several other ice streams. Climatic switches may involve an increase in temperature and/or sea level rise. The problem here, of course, is that unlike the disposition of geology, the influence of climate is far more complicated. This is further compounded by ice sheet and climate coupling, and critical thresholds of climate change. Is the West Antarctic responding to short term or long term climate forcing? Has collapse already started?

To answer these questions, we need to refine our reconstructions of the deglaciation of former ice sheets. This demands a knowledge of those ice streams that deglaciated the northern hemisphere ice sheets at the end of the last ice age. As stated in Chapter 2, predicting the future of the West Antarctic Ice Sheet is reliant on a better understanding of palaeo-ice streams. It is hoped that this thesis will add one more piece to the jig-saw that will reveal the complete picture of ice stream functioning in ice sheets.

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