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Geomorphology under ice streams: moving from form to process

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ABSTRACT: Ice streams are integral components of an ice sheet's mass balance and directly impact on sea level. Their flow is governed by processes at the ice-bed interface which create landforms that, in turn, modulate ice stream dynamics through their influence on bed topography and basal shear stresses. Thus, ice stream geomorphology is critical to understanding and modelling ice streams and ice sheet dynamics. This paper reviews developments in our understanding of ice stream geomorphology from an historical perspective, with a focus on the extent to which studies of modern and palaeo-ice streams have converged to take us from a position of near-complete ignorance to a detailed understanding of their bed morphology. During the 1970s and 1980s, our knowledge was limited and largely gleaned from geophysical investigations of modern ice stream beds in Antarctica. Very few palaeo-ice streams had been identified with any confidence. During the 1990s, however, glacial geomorphologists began to recognise their distinctive geomorphology, which included distinct patterns of highly elongated mega-scale glacial lineations, ice stream shear margin moraines, and major sedimentary depocentres. However, studying relict features could say little about the time-scales over which this geomorphology evolved and under what glaciological conditions. This began to be

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addressed in the early 2000s, through continued efforts to scrutinise modern ice stream beds at higher resolution, but our current understanding of how landforms relate to processes remains subject to large uncertainties, particularly in relation to the mechanisms and time-scales of sediment erosion, transport and deposition, and how these lead to the growth and decay of subglacial bedforms. This represents the next key challenge and will require even closer cooperation between glaciology, glacial geomorphology, sedimentology, and numerical modelling, together with more sophisticated methods to quantify and analyse the anticipated growth of geomorphological data from beneath active ice streams.

Keywords: *ice stream, glacial geomorphology, West Antarctica, subglacial bedforms, subglacial sediments*

1. Introduction

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: p. 301). They are large features (1-10s of km wide and 10-100s of km long) and their significance arises from their high velocity (100s-1000s m a^{-1}) and their correspondingly large ice flux. In Greenland, for example, ice streams account for approximately 50% of the ice sheet's annual mass loss (van den Broeke *et al*., 2009), and this figure increases to around 90% in Antarctica (Bamber *et al*., 2000), where there is much less surface melt (the other major process through which ice sheets lose mass). Thus, the dynamics of ice streams holds important implications for ice sheet mass balance and associated impacts on global sea level, which has become an increasingly serious concern since the 1970s (Hughes, 1973; Mercer, 1978; Alley et al., 2005; Shepherd et al., 2012; Hanna et al., 2013; Joughin et al., 2014). Although there are uncertainties about the precise atmospheric and oceanic factors that force increased discharge via ice streams (Carr et al., 2013) and the extent to which subglacial topography modulates their response (Pfeffer et al., 2008; Nick et al., 2013; Morlighemi et al., 2014; Favier et al., 2014), their rapid flow is largely governed by processes at the ice-bed interface, which act to sustain, enhance or inhibit their motion (see reviews in Bentley, 1987; Bennett, 2003; Stokes et al., 2007). Explicit comparisons have rarely been made, but the ice-bed interface is analogous to the 'boundary layer' in fluvial and aeolian environments, where shear stresses (both basal and lateral in the case of ice streams) oppose the flow of the overlying medium. This analogy extends further because processes within the boundary layer create a distinctive geomorphology that is characterised by subglacial bedforms that resemble features created in fluvial and aeolian environments (Shaw 1983; Fisher and Shaw, 1992; Shaw et al., 2008; Clark, 2010; Ely et al., 2016; Barchyn et al., 2016).

The study of ice stream geomorphology is, therefore, critical to our understanding of ice stream behaviour. The creation of landforms beneath ice streams results from erosion, transport and deposition of sediment which, in turn, is intimately linked to the mechanisms through which they are able to flow so rapidly (e.g. Ó Cofaigh et al., 2002; Schoof, 2002; Alley et al., 2003; Bingham and Siegert, 2009; Graham et al., 2009). Over short time-scales (years to decades), the evolution of subglacial geomorphology influences the roughness of the bed and the routing and pressure of subglacial meltwater, which further impacts on bed 'stickiness' and modulates the flow of the ice stream (e.g. Schoof, 2002). Over longer time-scales (centuries to millennia), these processes may alter the subglacial morphology in a more gradual manner, such as over-deepening the bed (Alley et al., 2003; Cook and Swift, 2012) and creating major sediment depocentres at their terminus (Alley et al., 2007). Unlike fluvial and aeolian environments, however, the basal environment of ice streams is notoriously difficult to access and observe, and there is far more uncertainty in securely linking form to process. This, in turn, severely limits attempts to understand ice stream behaviour and parameterise their basal conditions in numerical modelling (Bentley, 1987; Bennett, 2003; Ritz et al., 2015).

Despite the difficulty in observing the subglacial environment, our understanding of ice stream geomorphology has grown rapidly in the last three decades, from almost complete ignorance to a detailed knowledge of their geomorphological imprint. This has been brought about by two main approaches: (i) borehole and geophysical investigation of modern (active) ice streams, and (ii) sedimentological, geomorphological and geophysical investigation of well-preserved palaeo-ice stream beds. The aim of this paper is to comprehensively review progress in these two main areas from an historical perspective; highlight the key questions that remain; and discuss the opportunities that are likely to arise that will enable them to be addressed. Section 2 reviews the progress made through the pioneering geophysical and borehole investigations on active ice streams, mostly in West Antarctica, from the 1970s to the early 2000s. Section 3 reviews the parallel progress made by those seeking to identify and characterise the beds of palaeo-ice sheets from the late 1980s to early 2000s. Section 4 returns to a more recent body of work aimed at imaging the beds of active ice streams (generally post-2000), and highlights the links between these geophysical investigations of modern ice streams and the large growth in data from palaeo-ice stream beds. This convergence of knowledge has led to some important advances in our understanding of the geomorphology of ice stream beds, but it remains a key challenge to securely link landforms to processes. Section 5 focusses on highlighting the key questions that remain and briefly highlights opportunities that might arise to address them. It is concluded (Section 6) that future surveys of modern-ice stream beds are likely to offer unprecedented opportunities to study ice stream geomorphology that will require the continued cross-pollination of ideas and concepts from each of the two main approaches, together with a third approach – numerical modelling.

2. Early Exploration of the Beds of Active Ice Streams

2.1. The Discovery and Definition of Ice Streams

The first use of the term ‘ice stream’ can be traced back to Rink (1877), who used ‘Is-Ström’ (i.e. ‘ice stream’ or ‘ice current’) in relation to the Greenland Ice Sheet (GrIS) to describe “those outer parts of the inland ice which are moving with greater rapidity towards the ice-fjords than the rest of its outer margin” (p. 369). Almost two decades on, Bell (1895) used ‘ice stream’ to describe flow in the vicinity of Hudson Strait, which is the first use of the term in relation to a palaeo-ice sheet (see Brookes, 2007). In a paper entitled ‘Antarctica and Some of its Problems’, Edgeworth (1914) appears to be the first to use the term in relation to that region, noting that ice streams “differ from ordinary glaciers in that they are not bounded by rock walls, but are simply slight depressions in the general surface of the inland ice, marking areas where the ice is in more rapid movement than in adjacent areas” (p. 611). Thus, whilst the definition of an ice stream is often attributed (e.g. in Bennett, 2003; Bentley, 1987) to a note by Charles Swithinbank (Swithinbank, 1954), he was simply stating a preference for the term used by Rink (1877) and Edgeworth (1914) over other possibilities "on account of the impression it conveys of movement" and because "it was the term first used in describing these features" (p. 186; see also the note by Roscoe (1954) in the same issue).

The early descriptions and discussions of ice streams (Rink, 1877; Edgeworth, 1914; Swithinbank, 1954) emphasised that they have no exposed rock to define their lateral margins, thereby distinguishing them from outlet glaciers. However, adhering to such a strict definition is often deemed impractical (cf. Bentley, 1987; Paterson, 1994), not least because the vast majority of ice streams *sensu stricto* become bordered by rock-walls in coastal regions, where they flow through deep fjords, or may lie within deep glacial valleys where rock walls lie just beneath the ice surface. This is not the case for a small number of present-day ice streams that flow over relatively flat terrain along the Siple Coast of West Antarctica, and for several ice streams that operated at the land-terminating margins of palaeo-ice sheets (Margold et al., 2015b). Thus, a more useful distinction from a process point of view is to define ice streams according to their topographic setting (Bentley 1987). Following Bentley (1987), Stokes and Clark (1999) used the term 'pure ice streams' (see also Paterson, 1994) for those not associated with large troughs, and 'topographic ice streams' for those that are constrained by topography. Truffer and Echelmeyer (2003) used the terms 'ice stream' (for 'pure') and 'isbræ' (for topographic) to describe these two types, and pointed out important differences in their geometry, flow mechanisms, force balance, hydrology and potential for instability. However, they also emphasised (as did Bentley, 1987) that these two types are end members along a spectrum, with many ice streams displaying characteristics of both types (see Table 1 in Truffer and Echelmeyer, 2003). For example, enhanced flow within the tributaries of an ice stream might be strongly steered by topography, even if the lower reaches of the main ice stream 'trunk' is not, as has been suggested for ice streams along the Siple Coast (Joughin et al., 1999). In this paper, I will refer to ice streams in the broadest sense and include fast flowing outlet glaciers that are topographically controlled (cf. McIntyre, 1985; Bentley, 1987; Paterson, 1994).

2.2. West Antarctic Ice Streams and 'Glaciology's Grand Unsolved Problem'

Given their early recognition (e.g. Rink, 1877) and now well-established significance (e.g. Bennett, 2003), there were very few studies of ice streams until the 1970s, see Fig. 1. Indeed, following Swithinbank's (1954) brief note, the first scientific paper with "ice stream" or "ice streams" in its title was Hughes' (1977) review of 'West Antarctic Ice Streams', but it is important to note that several major outlet glaciers (ice streams in the broadest sense) had been studied, e.g. Lambert Glacier in East Antarctica (e.g. Morgan and Budd, 1975; Allison, 1979). Ice streams had, of course, been recognised for some time, but mechanisms of fast glacier flow had been investigated almost exclusively in relation to surge-type glaciers (e.g. Meier and Post, 1969). Indeed, in a comment published at the end of Meier and Post's (1969) classic paper ('What are glacier surges?'), Johannes Weertman was concerned that they had ignored fast-moving outlet glaciers (i.e. ice streams *sensu lato*), suggesting that their flow mechanisms may actually be similar to surging glaciers, and imploring the community not to over-look the "outlet glacier problem" (Weertman, 1969: p. 817). With hindsight, it is clear that the observations, concepts and theories that had been applied to surging glaciers would turn out to be instrumental in shaping much of the early work on ice streams, not least because many of those who had studied surging glaciers began to turn their attention to ice streams.

It was concern over the stability of the marine-based West Antarctic Ice Sheet (WAIS) that instigated a concerted effort to understand its dynamics and stability. Hughes (1972, 1973) was one of the first to hypothesise that the junction of grounded ice sheet with a floating ice shelf (the grounding line) might be unstable if the bed deepens inland. He pointed out that this was the case for the WAIS (Hughes 1972, 1973) and went so far as to suggest that retreat was already underway. A further analytical treatment by Weertman (1974) confirmed the basic idea of marine-ice sheet instability, but he was more equivocal in stating whether or not the WAIS was already undergoing marine ice sheet instability (see Joughin et al., 2014 and Favier et al., 2014 for more recent analyses of this important issue). Weertman (1974) acknowledged the possibility that WAIS was disintegrating, but also noted that a complete treatment of this problem must take into account the fast moving ice streams that exist along the Siple Coast, where it flows into the Ross Ice Shelf. Later work by Hughes (1977) argued that the ice streams themselves were inherently unstable, but he acknowledged that the factors driving their motion were not well understood and highlighted the need for further observations, which he suggested should include bed profiling using radar and seismic surveys, and borehole investigations. The need for these investigations was made more urgent by a series of high-profile papers that debated the potential instability of the WAIS (e.g. Thomas, 1976; Weertman, 1976; Whillans, 1976; Mercer, 1978; Thomas et al., 1979), and which Weertman (1976) referred to as 'Glaciology's grand unsolved problem'.

Hughes' (1977) invocation for geophysical exploration of ice streams was, in fact, already underway as a result of collaboration between the UK's Scott Polar Research Institute (SPRI) and the Technical University of Denmark (TUD), with the cooperation and support of the National Science Foundation (NSF). Airborne radio echo-sounding of a 50 km grid in Marie Byrd Land was undertaken, where the Siple Coast ice streams (then named A, B, C, D and E)[1] had earlier been reported based on reconnaissance flights using the same technique (Robin et al., 1970). The results were reported in Rose (1979), which was one of the first studies to investigate the basal conditions beneath active ice streams, together with Morgan and Budd's (1975) radio-echo sounding of the Lambert Glacier basin. A key conclusion from Rose's (1979) study was that the lower reaches of the ice streams were contained within shallow, broad channels within a relatively smooth topography that becomes progressively rougher in their upstream reaches. Rose also calculated basal shear stresses using ice surface slopes and thicknesses and found that they were less than 15 kPa close to the grounding line. His first-order calculations of basal temperatures confirmed that the beds of the ice streams were warm-based, whereas the inter-stream ridges were below the pressure melting point. Similar conclusions were reached by Morgan and Budd (1975), and by Stephenson and Doake (1982), which was one of the first geophysical studies of Rutford Ice Stream, West Antarctica, and which highlighted the sensitivity of its grounding line position to changes in ice thickness. Crabtree and Doake (1982) also used radio-echo sounding to provide some of the first data from the bed of Pine Island Glacier and noted its potential vulnerability to rapid retreat from a bedrock sill, a topic which has attracted

[1] These ice streams have since been re-named Mercer (A), Whillans (B), Kamb (C), Bindschadler (D) and MacAyeal (E), but the original names are used here to aid cross-referencing to the older literature.

considerable attention in recent years (Favier et al., 2014; Smith et al., 2017). Rose (1979) also pointed out the enigmatic behaviour of Ice Stream C, which appeared to be inactive, but displayed many of the surficial and internal characteristics of the neighbouring active ice streams.

Another key conclusion from Rose's (1979) investigation was that the flow mechanism of pure ice streams was fundamentally different from that of normal ice sheet flow. Similar arguments were put forward by McIntyre (1985) who noted the difference in flow mechanisms between major ice streams/outlet glaciers that occupied deep topographic troughs (and sometimes coincide with topographic steps) and those on along the Siple Coast, which he suggested might have more in common with ice shelves. Notwithstanding the enigmatic Ice Stream C, therefore, a key question was how these particular ice streams achieved their high velocities, given the extremely low ice surface slopes and low driving stresses. This prompted a series of pioneering geophysical investigations aimed at characterising their basal environment and ascertaining the mechanism(s) of flow.

2.3. Ice Stream Geophysics: A Paradigm Shift in Glaciology?

In 1983, a major field programme was initiated to study the Siple Coast ice streams and geophysical techniques targeted one of the two major tributaries of Ice Stream B at the 'Upstream B camp' (UpB), where ice thicknesses approach 1,000 m and velocities exceed 400 m a^{-1} (Whillans *et al*., 1987). The field survey during the first season of the programme (1983-1984) deployed seismic reflection techniques to investigate the nature of the contact zone between the ice stream and the bed, and the results were presented in two papers (Blankenship *et al.,* 1986; Alley *et al*., 1986), that Clarke (1987) soon suggested would resonate "for many years" (p. 8840). These surveys revealed a layer of sediment averaging 5-6 m thick immediately beneath the ice with very low compressional (P) and shear (S) wave speeds, see Fig. 2 (Blankenship *et al*., 1986). Such low wave speeds were taken to imply that the sediment was highly porous and saturated with water, and was too weak to support the shear stress exerted by the overlying ice. Further arguments relating to till porosity, the ice stream force balance, and the water balance were presented in the accompanying paper by Alley *et al.* (1986), who argued that “deformation within the till is the primary mechanism by which the ice stream moves” (p. 57).

Boulton’s (1986) commentary that appeared alongside these papers hailed them as a ’paradigm shift’ in glaciology because most previous models of glacier flow had assumed that ice moved over a rigid surface. However, he also emphasised that a ‘rigid bed model’ would have seemed quite implausible to glacial geologists who were familiar with the sediments and landforms on the beds of palaeo-ice sheets, noting that its survival for so long “is an indication of how little note the glaciological community took of the views of geologists” (p. 18). Indeed, Boulton pointed out that the discovery of deformable sediments beneath these ice streams was entirely consistent with what had long been known from investigations on palaeo-ice sheet beds, such as glacially induced deformation structures (Slater, 1926) that are often associated with subglacial bedforms, such as drumlins (Stanford and Mickelson, 1985). Moreover, Boulton and Jones (1979) had earlier presented experimental evidence of till deforming beneath Breiðamerkurjökull in Iceland, the results of which received a more thorough treatment in the classic paper by Boulton and Hindmarsh (1987), which also described the geological consequences in terms of the development of subglacial bedforms. Boulton’s (1986) commentary ended with a plea for glaciologists and glacial geologists to work together on coupling the form and structure of ice sheet beds with ice dynamics “to the benefit of both disciplines, in a way which until now has been sadly lacking” (p. 18).

It is often overlooked that both Blankenship et al. (1986) and Alley et al. (1986) speculated about buried features within the till layer, albeit from observations with limited resolution (Fig. 2), and complicated by acquisition artefacts such as hyperbolae (see Rooney et al., 1987a). Blankenship et al. (1986) noted how the occurrence of more than one reflector suggested some stratification within the till layer, which was nearly flat in the along-flow direction. In contrast, there was much less continuity in the reflectors in the across-flow direction, and they noted a ‘hump’ that seemed to rise near, or even to, the base of the ice (Fig. 2), with another layer-bottom reflector indicating localised relief of 8 m. They described the emerging picture as “one of ridges and valleys in the basal surface of the sub-glacial layer, trending parallel to the axis of the ice stream” with their alignment suggesting “they are formed by erosion” (p. 56). Alley et al. (1986) went further and described the features as flutes (citing Gravenor and Meneley, 1958), suggesting they may be a characteristic erosional consequence of deforming till, and outlining further implications of their erosional-deformational model. They argued that variations in till thickness over the flutes should cause fluctuations in basal drag that affect ice stream flow, with higher drag over the ridge crests. A further corollary is that the estimated till flux at UpB is equivalent to steady state erosion of ~0.5 mm a^{-1} over the catchment, which would eventually be transported and delivered to the grounding line to form a 'till delta' (Alley et al., 1986).

An AGU Chapman Conference on Fast Glacier Flow in 1986 provided an opportunity for several workers to discuss the pioneering discoveries from under Ice Stream B (see introduction by Clarke, 1987), and a collection of papers were published in a special issue of the *Journal of Geophysical Research* in 1987 (Fig. 1). However, only a handful of these papers discussed the geomorphology of the ice-bed interface, and none in any detail. Indeed, Bentley's (1987) influential review in that special issue simply reiterated Hughes' (1977) earlier plea for more data on the bed conditions, emphasising that all of the existing numerical models of ice stream flow (except for Alley et al., 1987b, also in the same issue) suffer from a lack of measurements of the physical conditions at the bed and the lack of a valid sliding law. Shabtaie et al. (1987) described the morphology of Ice Streams A, B and C from airborne radar sounding and over-snow traverses, but their focus was primarily on the ice surface morphology. However, they were able to confirm many of the underlying topographic features reported by Rose (1979) and they also detected localised patches of strong reflectors at the bed of Ice Stream C, which they took to imply the existence of a water layer at least several centimetres to several metres thick, and which was probably ponded in some locations.

A series of four papers expanded on the earlier geophysical investigations at UpB Camp. Blankenship et al. (1987) presented more detailed results of one of the active seismic measurements, calculating that the porosity of the till was greater than 0.32 and probably closer to 0.4, and was saturated with water at a pore pressure only about 50 kPa less than the ice overburden pressure (9,000 kPa). The second paper by Rooney et al. (1987a) offered greater insight into the characteristics of the buried features within the till layer in this location. They reported data from two parallel seismic reflection lines that were 8.3 km long and oriented perpendicular to ice flow and 360 m apart. These lines revealed a near-continuous till layer averaging 6.5 ± 0.5 m and characterised by an undulating surface of ridges and troughs aligned parallel to ice flow at the bottom of the till layer. They considered whether the largest of these features may have been a drumlin, but suggested that the dimensions were not typical of drumlins. Rather, and following Alley et al. (1986), they argued that these buried features (200-1,000 m wide and 6-13 m deep) were similar to flutes observed on the bed of the Laurentide Ice Sheet (LIS) (citing Gravenor and Meneley, 1958, and Smith, 1948). Interestingly, Smith (1948) preferred the term 'giant glacial grooves' for the features he observed along the Mackenzie Valley in northwestern Canada, and which extend up to 13 km in length (see Section 3.3).

Rooney *et al.* (1987a) also drew attention to the base of the till layer, where a gently dipping sedimentary structure (interpreted to be sedimentary bedrock: see also Rooney et al., 1987b) was truncated by the till in an angular unconformity that suggested that the till was eroding, or had eroded, its bed. Indeed, they pointed to disturbed zones of anomalous dip direction immediately beneath the till that might represent the fracture and entrainment of large bedrock rafts into the till layer or deeper sub-sole deformation structures some 30 m below the ice-bed interface. In the third paper, Alley et al. (1987a) presented further evidence to argue that the till beneath Ice Stream B was deforming, eroding its bed, and being deposited at the grounding line as till deltas tens of kilometres long (see also Alley et al., 1989a). They pointed out that evidence for such deltas may already have been found in the grounding line mapping of Shabtaie and Bentley (1987), which showed lobate regions of grounded ice tens of kilometres long at the mouths of Ice Stream's A and B. Such deltas might be particularly important for ice dynamics because their deposition serves to stabilise the grounding line by locally increasing the water depth required for flotation (see later work by Alley et al., 2007). A further implication was that till deltas should mark recessional positions of marine-based ice sheets. The fourth paper (Alley et al., 1987b) developed a simple one-dimensional numerical model of ice stream flow on deforming till with a linear viscous rheology, although the possibility of partial decoupling across a water film was considered and, they argued, was likely to be proportionally more significant closer to the grounding line.

A important theme from the collection of papers on 'Fast Glacier Flow' (Clarke, 1987) was that the geometry and pressure of the basal water system was “very important in facilitating rapid basal velocities” (Alley, 1989a: p. 110). Irrespective of whether the ice slid over a till layer or deformed it, rapid motion would only occur if the water pressure was relatively close to the ice overburden pressure (Boulton and Hindmarsh, 1987; Clarke, 1987). In turn, the water pressure depends on the water supply and its drainage path, which is likely to be influenced by the bed geomorphology and may also play a role in shaping the form of the bed (e.g. through erosion and deposition). In relation to Ice Stream B, it had been argued that excess subglacial water was incapable of being conducted through subglacial aquifers (Lingle and Brown, 1987) or advected within a deforming bed (Alley et al., 1987a), and so it was assumed that it drained at the ice-bed interface. However, the form of that basal drainage system was largely unknown. Alley and co-workers developed a series of hypotheses regarding the water-drainage system that might be associated with a deforming till layer and which could be tested by direct borehole drilling to the ice-bed interface (Alley, 1989a, b). The major conclusions were that most of the ice stream motion should arise from bed deformation; that most of the subglacial water flowed in a distributed system approximating a thin film at the ice-bed interface (with local thickenings up to a few mm, but with no channels or linked cavities); and that effective pressures were small and should decrease down the ice stream (Alley 1989a). In terms of the geomorphology, and in the absence of channelized meltwater supplies (e.g. from supraglacial sources), they argued that it was unlikely that ice streams underlain by a metres-thick layer of saturated sediment would host any channelized drainage systems, such as those which might produce discrete channels eroded into the till, because sediment within the till layer would creep rapidly into incipient channels or cavities with lower pressures (Alley, 1989a; Boulton and Hindmarsh, 1987). It was generally assumed, therefore, that the subglacial drainage system would remain distributed and would erode and transport relatively small quantities (<0.1 m^3 m^{-1} a^{-1}) of fine-grained sediment (Alley *et al.,* 1989a), and effectively doing very little geomorphological ‘work’. Elsewhere, Doake et al., (1987) imaged the bed of Rutford Ice Stream in the vicinity of the grounding line and showed that the ice stream was intermittently grounded on raised areas of the bed that formed “esker-like” features (Doake et al., 1987: p. 8951).

2.4. Direct Borehole Access: Sliding versus Deformation or Ploughing

Despite the paradigm shift in understanding brought about by geophysical investigations, Alley *et al*. (1989b) acknowledged that the details of glacier motion on a deforming bed remained elusive and suggested that "bore-hole observations undoubtedly will be necessary to solve the problems of till deformation beneath thick ice" (p. 137). Indeed, the series of papers published by Alley in 1989 (Alley, 1989a; 1989b; Alley *et al*., 1989b) were partly motivated by the knowledge that Barclay Kamb and Herman Engelhardt (and co-workers) would soon be drilling to the base of Ice Stream B (Alley, 1989a). The initial results were reported in Engelhardt *et al*. (1990) who were the first to describe the physical conditions at the base of a fast moving Antarctic ice stream. Similar work was also being undertaken at Jakobshavns Isbræ in West Greenland, with the results published in Iken et al. (1993), also discussed below.

Using a hot water method, Engelhardt et al. (1990) drilled several boreholes to the base of Ice Stream B and confirmed that it was at the pressure melting point (as first calculated by Rose, 1979) and that basal water pressure was within about 160 kPa bars of the ice overburden pressure (broadly similar to the values calculated by Blankenship et al., 1987). Moreover, when the boreholes reached the bed, the water levels within them dropped very rapidly, revealing a connection to a basal water conduit system. Engelhardt et al. (1990) sampled some of the basal sediments driven into suspension and, through the use of a borehole penetrometer and piston corer, were able to sample a muddy, unsorted mixture with rock clasts as large as 5 cm, thus resembling a basal till. Weight loss on drying confirmed the relatively high porosity (~40%) calculated from earlier indirect methods (e.g. Blankenship et al., 1987). The high porosity, together with a mixed age of recovered microfossils and bedrock sources would be expected from a till that was deforming, but the discovery of a well-connected basal water system at high pressure led Engelhardt *et al.* (1990) to conclude that "the proportions of ice stream motion due to basal sliding, subglacial till deformation and ice superplasticity remain unknown" (p. 59).

Although it did not contradict the earlier work based on geophysics, the borehole data from Ice Stream B and other non-ice stream settings (e.g. Blake et al., 1992; Humphrey et al., 1993; Iverson *et al*., 1994; 1995) had begun to trigger a re-evaluation of the deforming bed mechanism and, in particular, question the linear viscous rheology of the saturated till (Murray, 1997). Crucially, it allowed workers to conduct experiments on the till recovered from beneath Ice Stream B and, building on concepts and experiments on the mechanics of clay-rich soils, Kamb (1991) tested freshly-cored till from Ice Stream B using a shear box method. Kamb found that the till sheared rapidly and underwent ‘catastrophic’ failure that resembled a plastic material with a yield stress of around 2 kPa. This suggested that the flow law for the till was highly non-linear and Kamb presented further arguments that the deforming bed mechanism would be unstable if a linear viscous rheology was assumed, due to the feedback from the generation of basal water by shear heating in basal till (i.e. the ice stream should rapidly increase in velocity). That such instability had not been observed, he argued, was evidence against a viscous till rheology.

In order to further constrain the basal mechanics, Engelhardt and Kamb (1998) returned to the UpB camp and deployed a tethered stake apparatus into the surface of the till layer via a borehole. Their aim was to determine what proportion of the ice stream motion might be accommodated by basal sliding, rather than till deformation. Displacement of the tether line indicated that basal sliding *sensu lato* (i.e. potentially including a contribution from shear deformation of the till very close to the ice base) predominated, and may account for as much as 83-100% of the ice stream’s motion, with any till deformation restricted to a shear zone 3-25 cm thick at the top of the 9 m thick basal till. If correct (and applicable across the entire ice stream), the implications of this discovery were profound, because the basal boundary conditions and flow law for ice streams would be radically different for basal sliding compared to pervasive till deformation. Engelhardt and Kamb (1998) also pointed out that the erosion, transport and deposition of till would be different under a sliding regime and, although they refrained from any discussion of the geomorphology, it follows that these different mechanisms could produce quite different landscapes beneath ice streams. To further complicate the picture, a similar study from Ice Stream D, also using a tethered stake instrument, showed that most of the ice stream motion (80%) was accommodated by till deformation (Kamb, 2001). Thus, it appeared that the ratio of sliding to till deformation varied from one ice stream to another and also through time, with Kamb (2001) noting how these contrasting results illustrated the “difficulty of making valid generalisations about the ice stream [flow] mechanism” (p. 174). It also became clear that the flow mechanisms of ice streams located in deep subglacial troughs could be quite different from the Siple Coast ice streams, as earlier work had postulated (e.g. McIntyre, 1985). Indeed, several boreholes drilled through Jakobshavns Isbræ, West Greenland, showed that the base of that ice stream is characterised by a layer of temperate ice of substantial thickness that deforms to support a large proportion of the movement of the ice stream. Later seismic investigations by Clarke and Echelmeyer (1996) indicated that much of the basal interface was probably underlain by compacted, non-deforming sediment, confirming the importance of ice deformation (enhanced creep) as the principal flow mechanism.

Suggestions of a near-plastic till rheology under the Siple Coast ice streams were further corroborated by two influential papers by Tulaczyk *et al*. (2000a, b). New triaxial ring-shear tests (and confined uniaxial tests) on the UpB till (Tulaczyk et al., 2000a) confirmed a nearly plastic till rheology, whereby the failure strength of the till was strongly dependent on effective stress, but practically independent of strain and strain rate. Moreover, Tulaczyk *et al.* (2000a) formulated a Compressible-Coulomb-Plastic (CCP) till model that was able to produce the viscous-like vertical distributions of strain in the upper few centimetres of the till layer, that had been suggested based on the borehole observations (Engelhardt and Kamb, 1998), and which was consistent with textural and compositional properties (Tulaczyk *et al*., 1998). In the second paper, Tulaczyk *et al*. (2000b) formulated a new analytical ice stream model, termed the 'undrained plastic bed' model, in which the storage of water within the till was an important component. Significantly, this modelling was able to produce two thermo-mechanically controlled states for ice streams, one with a strong bed and low ice velocities, and one with a weak bed and higher velocities. This suggested that ice streams might be susceptible to thermally triggered instabilities, during which small perturbations in the basal thermal energy balance could grow and led to the generation or elimination of the basal conditions required for ice streaming.

A challenge for the plastic bed model was to explain the geomorphology of ice stream beds in terms of the erosion, transport and deposition of sub-ice stream tills. Several workers had argued that a viscously deforming bed was entirely consistent with the production of flutes and drumlins at the ice-bed interface (e.g. Alley *et al*., 1986; Rooney e*t al.,* 1987a; Boulton and Hindmarsh, 1987; Hindmarsh, 1998a; b) and till deltas at the grounding line (Alley *et al*., 1989a). Indeed, Hindmarsh (1998a; b) had demonstrated that, under certain conditions, a viscously-modelled till sheet was unstable and that infinitesimal perturbations in amplitude were able to grow into drumlin-like features that were consistent with the sedimentological features assumed to reflect deforming beds (Hart, 1997). Hindmarsh (1998a; b) clearly acknowledged that, on the small scale, till behaved plastically (e.g. Iverson et al., 1997; 1998), but argued that a viscous law represented the simplest model compatible with large-scale observations on ice sheet beds. Hindmarsh argued that the net effect of numerous small-scale failure events (at scales of 0.1 to 1.0 m) is that of viscous behaviour at spatial scales >1 km (see also Hindmarsh, 1997; Fowler, 2003), but this has been questioned more recently based on the stick-slip motion of the lower reaches of Ice Stream B (Tulaczyk, 2006; see also Iverson, 2010).

In contrast, the geomorphological implications of a nearly plastic till rheology were unclear. In order to address this, Tulaczyk *et al*. (2001) proposed a qualitative ploughing model that was consistent with the experimentally-determined Coulomb-plastic rheology and which invoked keels in the base of the ice ploughing through a metres-thick layer of sediment and transporting it downstream. Interestingly, whilst they acknowledged that there was little direct evidence for the existence of ice keels beneath modern West Antarctic ice streams, they pointed out that earlier geophysical surveys were, at least, consistent with the notion of a 'grooved' till layer (Rooney et al., 1987a; Novick et al., 1994) and that bedforms left behind by palaeo-ice streams strongly supported their assertions (e.g. mega-scale glacial lineations: Clark, 1993; Canals et al., 2000). They presented a series of qualitative arguments to suggest that the keels can be generated by ice passing over a bed that is stronger than the ice (e.g. hard bedrock) and then survive as the ice ploughs through a weaker bed. Like the viscous till model, this ploughing mechanism produces a net till flux down the ice stream, albeit an order of magnitude lower than estimates for pervasive till deformation spanning several metres (see discussion in Section 5.1).

To summarise, although the detailed geomorphology of ice stream beds remained elusive, attempts had been made to reconcile the two quite different till rheology models with what was known about the conditions at the ice-bed interface under modern ice sheets, and influenced, to some extent, by observations of subglacial bedforms from palaeo-ice sheet beds.

2.5. In Search of Ice Stream Sticky Spots

Notwithstanding the debates about the rheology of sub-ice stream tills, it had become widely accepted that they offered very little resistance to flow in the areas that had been investigated. Indeed, early calculations of the force balance along ice streams indicated that a significant fraction of the driving stress must be supported by 'side drag' against the slower-moving ice at the lateral margins (Echelmeyer *et al.*, 1994), perhaps approaching as much as 100% on their lower-most ice 'plains' (Bindschadler *et al.,* 1987). However, these values were typically closer to around 50% further upstream (Raymond *et al.,* 2001; their Table 1), which suggested that basal shear stresses must contribute some component of the resistive stresses. Similar results were obtained from a study of the force balance of Rutford Ice Stream (Frolich and Doake, 1988) where it was found that there is a marginal zone in which lateral stresses are high, but that the centre of the ice stream is dominated by basal friction. Moreover, it is unlikely that bed resistance is entirely uniform and Alley (1993) highlighted several lines of evidence from the Siple Coast ice streams that were suggestive of localised areas of higher basal drag, which he referred to as 'sticky spots' (see Table 1 in Alley, 1993). Evidence included ice surface rumples and crevassing (Vornberger and Whillans, 1986), spatially variable till thicknesses (Rooney et al., 1987a), and inversions of basal drag based on ice surface data (MacAyeal, 1992), but their potential geomorphological expression was largely speculative.

Alley (1993) argued that large bedrock bumps penetrating a layer of till and protruding into the base of an ice stream were the most likely cause of sticky spots. He argued that for typical ice stream velocities (450 m a^{-1}), hemispherical bumps of 1 m radius covering <1% of the bed, or 10 m radius covering <2% of the bed, would theoretically be able to support a basal shear stress of 10 kPa. Empirical support for their existence is found from the correspondence between areas of minimum surface velocity and undulations in ice surface morphology that are thought to represent bedrock ridges, e.g. on Ice Stream E (Bindschadler and Scambos, 1991; MacAyeal, 1992; MacAyeal *et al*., 1995).

Discontinuity in the basal till layer was another possible cause of stickiness identified by Alley (1993), but he pointed out that such discontinuities were unlikely to support very high basal shear stresses because of their tendency to collect a layer of lubricating water from the surrounding till, unless they also possessed abundant large bumps. Nonetheless, there was evidence for their existence beneath modern ice streams. Based on acoustic impedance data, Atre and Bentley (1993) reported laterally varying basal conditions beneath Ice Stream B and C that were inferred to reflect changes in the nature of the sediments composing the bed and/or their physical state (e.g. porosity). Rooney et al. (1987a) also noted a 300 m till-free swath (or at least thinner than the seismic resolution of 2 m) from under Ice Stream B, which Alley (1993) suggested could support up to 13% of the basal shear stress in that region. Observations of varying till thicknesses had also been made on another West Antarctic ice stream. Smith (1997a, b) used seismic reflection data to argue that part of the bed of Rutford Ice Stream (running alongside the Ellsworth Mountains and into the Ronne-Filchner Ice Shelf) was characterised by soft, saturated sediments (that may be deforming), whereas other areas were highly consolidated or consisted of poorly lithified sedimentary rock that offered more resistance to ice flow. Later work by Smith (2006) used seismic techniques to detect micro-earthquakes under Rutford Ice Stream and showed that the stiffer areas of the bed that are likely to support basal sliding were much higher friction than areas where thicker regions of till were inferred to be deforming. Incidentally, more recent work by Smith et al. (2015) showed that clusters of microearthquakes most likely represent sticky spots characterised by stiffer/low porosity sediments, but that they accommodate only a small amount of the total basal motion.

Alley (1993) also suggested that raised regions of the ice stream surface could cause moderate increases in shear stress because ice thickness anomalies are around ten times more effective than bed topography in influencing subglacial water flow. Thus, ice surface perturbations can divert water around the underlying region of the bed and cause an increase in till stickiness due to a localised deficit of lubricating water. Alley (1993) suggested that Ice raft “a” (an elliptical feature 10 km by 5 km and a few metres high, near the mouth of Ice Stream B: cf. Bindschadler et al., 1987), might be an obvious candidate for this type of sticky spot. Elsewhere, under Ice Stream C, it was suggested that areas of well drained (‘stickier’) till of the order of 10^2 m^2 generated micro-earthquakes (Anandakrishnan and Bentley, 1993; Anandakrishnan and Alley, 1994). Indeed, the diversion of subglacial water from Ice Stream C and into Ice Stream B has been hypothesised to account for its shutdown

(Anandakrishnan and Alley, 1997), although others have invoked a mechanism involving the freeze-on of subglacial water (Price et al., 2001).

Although not considered in Alley's (1993) 'search for ice stream sticky spots', evidence has since emerged that some 'thermal' sticky spots might be caused by the freeze-on of subglacial water. Lliboutry (1987) had recognised that islands of cold-based ice could freeze to high spots in bedrock roughness at scales of 10^1 to 10^3 m, but he pointed out that this would be unlikely under ice streams where frictional heating would offset any tendency towards freezing. However, modelling showed that ice stream thinning and the advection of cold ice closer to the bed may help promote basal freezing (Payne and Dongelmans, 1997; Christoffersen and Tulaczyk, 2003a; b), and several boreholes from Ice Stream C have encountered a thin layer of frozen till between the basal ice and unfrozen till (Carsey et al., 2002). Evidence of 'freeze-on' may also have been detected under Ice Stream B (see review in Kamb, 2001). Due to thermodynamic feedbacks, it has been hypothesised that basal freeze-on may spread quite rapidly and act to shut-down ice streams (Tulaczyk et al., 2000b; Bougamont et al., 2003a; b).

Thus, soon after the discovery of weak tills beneath ice streams (Blankenship et al., 1986; Engelhardt et al., 1990), a body of evidence had emerged to suggest the presence of localised sticky spots that might explain spatial and temporal variations in ice flow, including their stoppage and reactivation (e.g. Anandakrishnan and Alley, 1997; Bougamont et al., 2003a, b). However, whilst inverse methods, geophysical, and borehole observations provided important evidence of their existence, they lacked the spatial coverage and resolution to confidently identify their cause and distribution and, aside from the inferred presence bedrock bumps and ridges, the geomorphological expression of ice stream sticky spots was largely unknown.

2.6. Summary

Although ice streams were identified in the late 1800s (Rink, 1877), their importance was only recognised as recently as the 1970s, in relation to the potential instability of the WAIS (e.g. Hughes, 1977). This triggered a major growth in their investigation (Fig. 1) that began with the pioneering geophysical investigations of the Siple Coast ice streams (e.g. Blankenship et al., 1986; Alley et al., 1986), soon followed by direct borehole observations (e.g. Englehardt et al., 1990) and a concerted effort to understand and model their rapid flow. By the end of the 20th century it had been shown that they flow over a bed of 'soft' sediments and that their motion was accommodated by a combination of deformation within the till layer and/or sliding across its surface, interrupted by the presence of localised sticky spots. Comparisons between different ice streams showed that their basal conditions were highly variable both within and between individual ice streams (e.g. Kamb, 2001), and that ice streams that flowed through deep subglacial troughs did not necessarily require a layer of soft deforming sediment (Iken et al., 1993; Clarke and Echelmeyer, 1996). However, our knowledge of the geomorphology of ice stream beds remained very limited. Geophysical investigations (e.g. Rooney et al., 1987a) appeared to show buried flutes (or grooves) within till, with several lines of evidence also pointing to the deposition of till deltas at ice stream grounding lines (Alley et al., 1987a; Shabtaie and Bentley, 1987; Alley et al., 1989a). Moreover, these indirect observations were limited in terms of their vertical and horizontal resolution, and were almost exclusively obtained from a very small number of 'pure' ice streams that flowed over relatively flat sedimentary beds and were underlain by several metres of till. Several authors appealed to the notion that observations from palaeo-ice sheet beds would help inform interpretations of subglacial processes beneath ice streams (e.g. Boulton, 1986; Murray, 1997), but the study of palaeo-ice stream geomorphology was very much in its infancy and few ice stream beds had been identified with any confidence.

3. Identification and Characterisation of the Beds of Palaeo Ice Streams

3.1. In Search of Palaeo-Ice Streams

The first attempt to incorporate the location of ice streams into a palaeo-ice sheet reconstruction appears to be from Denton and Hughes (1981), who depicted numerous ice streams in the former mid-latitude ice sheets of the northern hemisphere, based largely on topographic inference, see Fig. 3a. Their reconstruction clearly benefitted from Hughes' knowledge of West Antarctic ice streams (Hughes, 1977), but Andrews' (1982) was more sceptical and, in his review of ice sheet reconstructions, he pointed out that "it is not known whether or where ice streams existed in the Laurentide Ice Sheet" (p. 25). It is interesting to note, however, that although virtually nothing was known about the geomorphology of ice streams, Luchitta et al. (1981) had speculated that ice streams may have carved outflow channels on Mars, based largely on the surface expression of Antarctic ice streams (see Kite and Hindmarsh (2007) for a more recent rendition of this argument).

The importance of incorporating ice streams into palaeo-ice sheet reconstructions was becoming increasingly recognised and Dyke and Prest (1987) included them as convergent flow-lines in their Late Wisconsinan and Holocene reconstruction of the Laurentide Ice Sheet (LIS). There remained, however, a lingering scepticism that ice streams could be identified in the palaeo-record and very few had been identified. Mathews (1991), for example, questioned whether we knew enough about their basal conditions to say anything about the landforms they created, but he drew attention to some features on the continental shelf off Canada that might be considered tracks of former ice streams, including cross-shelf troughs between Vancouver Island and the Queen Charlotte Islands in western Canada (Luternauer and Murray, 1983) and another marine trough extending southeast from the St Lawrence Estuary in eastern Canada. He also noted that the large lobate moraines of the southern margin of the LIS (e.g. the Des Moines and Lake Michigan Lobes) may represent another possible type of ice stream track, given their similar size to Antarctic ice streams and their low surface slopes. Mathews (1991) concluded by suggesting that other possible candidates may exist (e.g. in the Canadian Arctic Archipelago) and they merit investigation "once the criteria for identifying ice stream tracks become better established" (p. 267).

Mathews' (1991) pessimism was, however, misplaced, because Art Dyke and colleagues had already recognised evidence for ice streaming in the central Canadian Arctic (Dyke, 1984; Dyke and Morris, 1988), largely on the basis of carbonate-rich tills dispersed through areas of igneous or metamorphic bedrock. Dyke and Morris (1988) called these 'Boothia-type' dispersal trains and described a classic example from Prince of Wales Island that is shown in Fig. 4. They noted how the lateral margins of the dispersal train were very abrupt and that a convergent pattern of drumlins feeds into the plume and become more elongated along its central axis, noting that the overall pattern resembled ice streams in Greenland and Antarctica. Dyke and Morris (1988) also noted a "peculiar 'lateral shear moraine', a feature unknown from elsewhere" (p. 86). This feature is at least 68 km long (with minor gaps) and runs parallel to another major drumlin field. Dyke and Morris (1988) speculated that it may have marked a shear zone at the lateral margin of an ice stream.

Thus, around the same time that researchers were striving to glimpse the bed of modern ice streams (e.g. Fig. 2) in West Antarctica (Blankenship et al., 1986; Rooney et al., 1987a; Engelhardt et al., 1990), Dyke and Morris (1988) were providing the first detailed description of their elusive bed geomorphology (Fig. 4) and this paper would go on to have a profound influence on the emerging study of palaeo-ice streams.

3.2. Ice Lobes, Surging and Terrestrial Ice Streams

Around the same time as Dyke and Morris' (1988) influential work at the northern margin of the LIS, many of the lobes of the southern margin were attracting attention as possible zones of ice streaming (Dredge and Cowan, 1989) because they had been calculated to possess low surface slopes (Mathews, 1974) that were similar to those in West Antarctica. Hicock and co-workers also reported Boothia-type dispersal trains (Fig. 4a) and deformation tills associated with drumlins around the Great Lakes region, that were interpreted to reflect ice streaming (Hicock, 1988; Hicock and Dreimanis, 1992). Indeed, Alley et al.'s (1986) discovery of a deforming till beneath Ice Stream B led him to conclude that the widespread uniform till sheets of the southern margin of the LIS were also consistent with deposition from deforming subglacial sediment layers beneath relatively short-lived ice streams (Alley, 1991), that had earlier been described as 'surges' resulting from the build-up of pressurised subglacial water (Wright, 1973; Clayton et al., 1985). In relation to the Des Moines Lobe, Patterson (1997) argued that the "persistence of the lobate margin, the distinctive style of glacial erosion and deposition along the lowland, all led to the interpretation that an ice stream, analogous in size to those in West Antarctica, may have drained a large portion of the central Laurentide Ice Sheet" (p. 260). Indeed, Patterson (1997) drew heavily on West Antarctic analogues, suggesting that an ice flux similar to that from Ice Stream B (30 km^3 a^{-1}: Bindschadler et al., 1987) could have created the Des Moines lobe in around 1000 years. A further paper by Patterson described specific glacial landform assemblages that characterised several ice streams at the southern Laurentide margin (Patterson, 1998). These included a lowland suite of level-to-streamlined fine-grained till, sometimes associated with highly elongate drumlins (Bluemle et al., 1993) that typically terminated towards the lobe margins, where thrusting of glacial sediment was evident in association with hummocky topography and major moraine systems.

3.3. Giant Glacial Grooves and Mega-Scale Glacial Lineations

The observation of exceptionally long narrow drumlins along some hypothesised ice stream tracks at the southern Laurentide margin (Bluemle et al., 1993) appeared to suggest a link between their formation and rapid ice flow. Giant glacial grooves and other highly elongate landforms had been reported several decades earlier (e.g. Smith, 1948; Dean, 1953), but not explicitly linked to ice streaming, which is not surprising given that ice streams had yet to be fully recognised in modern ice sheets (see Sections 2.1 and 2.2). In central North Dakota, however, Bluemle et al. (1993) reported drumlin ridges up to 27 km in length and with typical length to width ratios ranging from 30:1 to 50:1, with a maximum of 240:1. They suggested that the features were formed in basal ice cavities and that sediments were squeezed into the lower-pressure cavity from laterally adjacent regions of the ice stream bed during ice thrusting.

Similar, albeit much larger, patterns of streamlining were also reported by Clark (1993). Using Landsat satellite imagery, he identified a “hitherto undocumented and much larger form of ice moulded landscape” (p. 1) which comprised streamlined glacial lineations with typical lengths of between 8 and 70 km, widths between 200 and 1300 m and spacings between 300 and 5 km. Clark (1993) termed these features ‘mega-scale glacial lineations’ (MSGLs) and discussed a variety of possible origins, concluding that they probably formed as a result of subglacial deformation and attenuation of sediments around inhomogeneities in till, similar to that earlier proposed for drumlins (e.g. Boulton and Hindmarsh, 1987). If correct, he argued that their great length would be a product of the duration of flow and the basal ice velocity, i.e. a 50 km long MSGL could form in as little as 50 years under velocities of 1,000 m a^{-1}, or as much as 5,000 years under velocities of 10 m a^{-1}. He then proposed that because ice sheet flow-lines are unlikely to have remained stable for long time periods (Boulton and Clark, 1990), MSGLs are likely to form under conditions of extremely rapid flow such as ice streams or surges. This qualitative link between bedform elongation and ice velocity has been further corroborated by more detailed quantitative analysis of drumlins and MSGLs on several palaeo-ice stream beds (e.g. Hart, 1999; Stokes and Clark, 2002; Briner, 2007; Stokes et al., 2013a), despite the fact that the precise mechanism(s) of their formation and elongation remains enigmatic (see discussion in Section 5.2).

3.4. Geomorphological Criteria to Identify Palaeo-Ice Streams

Several workers also noted that the flow-sets containing elongate subglacial bedforms were consistent with the presence of palaeo-ice streaming. Hodgson (1994) drew attention to the abrupt lateral margin of a >100 km long field of highly elongate drumlins on Victoria Island, Canadian Arctic, thought to delimit the position of the lateral shear margin of a palaeo-ice stream (Fig. 5), and similar to that proposed by Dyke and Morris (1988) on nearby Prince of Wales Island (Fig. 4). Elsewhere, Punkari (1995a, b) postulated the existence of eleven ice streams within a broad marginal zone of the Fennoscandian Ice Sheet (FIS), based largely on satellite imagery that revealed elongate drumlins arranged in fan-shaped flow patterns that contrasted with glaciofluvial landform assemblages in the inter-lobate zones (Punkari, 1997). Similar fans were described by Kleman and Borgström (1996) who termed them ‘surge fans’ and noted their distinctive bottle-neck flow patterns (convergent and then divergent) of glacial lineations that probably form rapidly and during the decay stages of an ice sheet. Kleman and Borgström (1996) identified the Dubawnt Lake surge fan in central Canada as a ‘type landscape, which has subsequently been shown to represent a major, but short-lived, ice stream track during final deglaciation of the Keewatin sector of the LIS (Stokes and Clark, 2003).

In terms of the geomorphology of ice streams, therefore, significant progress was made during the 1990s in terms of identifying evidence of ice streaming amongst the glacial geomorphological record of palaeo-ice sheet beds. These studies suggested that ice streaming should leave behind sedimentological evidence of fast ice flow in the form of heavily deformed tills and distinctive erratic dispersal trains that often depicted convergent flow-patterns (e.g. Dyke and Morris, 1988; Hicock, 1988; Alley, 1991; Patterson, 1997; 1998). Many of these flow-sets, or fans (cf. Kleman and Borgström, 1996), also contained highly elongate glacial lineations (MSGLs), which were postulated to reflect rapid ice velocities (e.g. Clark, 1993); and some were characterised by abrupt lateral margins (e.g. Hodgson, 1994) and lateral shear margin moraines (Dyke and Morris, 1988; Stokes and Clark, 2002a). Taken together, these were argued to represent the key 'geomorphological criteria' for identifying palaeo-ice streams, which Stokes and Clark (1999) encapsulated in a series of landsystems models (shown in Fig. 6) depending on whether the ice stream terminated in water or on land, and whether the glacial lineations were formed rapidly and synchronously or time-transgressively during ice margin retreat. Stokes and Clark (1999) hoped that these observational templates would aid the identification of ice streams in palaeo-ice sheets. Indeed, in the first review of the evidence of palaeo-ice streams in 2001, it was clear that only a very small proportion of palaeo-ice streams had been identified with any confidence (Stokes and Clark, 2001). In part, this may have been because all modern ice streams are marine-terminating and unlike, for example, surge-type glaciers, it was difficult to observe the geomorphology in their recently deglaciated foregrounds and identify modern analogues (Evans et al., 1999). As a result, Stokes and Clark (2001) suggested that geophysical observations of the submarine glacial geomorphology recently exposed in front of modern-day ice streams (e.g. Shipp et al., 1999) held huge potential to link the terrestrial record of mid-latitude palaeo-ice streams with their contemporary counterparts in West Antarctica.

3.5. Offshore records of palaeo-ice streams based on marine geophysics

The early work identifying palaeo-ice stream footprints was based largely on terrestrial glacial geology and was considerably aided by advances in satellite remote sensing in the early 1990s (see examples in Fig's 4 & 5). In a similar manner, advances in marine geophysical techniques during the late 1990s opened up new research aimed at identifying the geomorphology of palaeo-ice streams offshore (Ó Cofaigh, 2012). Early work using seismic and side-scan sonar focussed on glacially influenced continental slopes in the Polar North Atlantic (e.g. under the auspices of the 'Polar North Atlantic Margins' (PONAM) programme). Key contributions related to ice stream geomorphology included the discovery of large sedimentary fans at the mouth of cross-shelf troughs (termed 'Trough Mouth Fans': Vorren and Laberg, 1997), whose volume and architecture were interpreted to reflect rapid sedimentation by ice streams (Hooke and Elverhøi, 1996; King et al., 1996; Dowdeswell et al., 1996; Elverhöi et al., 1997; Vorren and Laberg, 1997; Vorren et al., 1998; Batchelor and Dowdeswell, 2014). In many cases, rapid sedimentation was interpreted to reflect transport via a deformable bed that was subsequently deposited at the ice stream grounding line and then remobilised down-slope (see Fig. 7), consistent with the observations and modelling of modern ice streams (Alley et al., 1986; 1987a, b; 1989a, b; see Section 2.3). One of the key advantages of these marine records is that the architecture of the depocentres could be used to bracket the age of sediment deposition and, hence, rates of sediment flux (Dowdeswell et al., 2004; Nygård et al., 2007). These estimates typically span an order of magnitude from 100 m^3 a^{-1} (Alley et al., 1989a) up to 1000 m^3 a^{-1} per metre width of ice stream terminus (Dowdeswell et al., 2004), although far more extreme rates have also reported, e.g. 8,000 m^3 a^{-1}, which are comparable to some of the world's largest rivers (Nygård et al., 2007).

A further advantage is that marine-based seismic data can provide spatially extensive information of sub-surface sediment characteristics of the ice stream geomorphology, which can often be used in conjunction with geotechnical analysis of sediment cores. In this regard, several studies have shown that the uppermost layer of the sub-surface of the former ice stream track is typically composed of an acoustically transparent till that is generally under-consolidated compared to the material beneath, see Fig. 8 (Dowdeswell et al., 2004; Ó Cofaigh et al., 2005; 2007; Nygård et al., 2007). This is often interpreted as a low shear strength ('soft') deformation till that facilitated the rapid basal motion of the ice stream (Dowdeswell et al., 2004; Ó Cofaigh et al., 2005; 2007) and which is subsequently deposited at the terminus (Fig. 7). Marine geophysical surveys of palaeo-ice stream beds indicates that the thickness of this till layer typically averages ~5-10 m (Dowdeswell et al., 2004; Ó Cofaigh et al., 2005; 2007; Livingstone et al., 2016a), which is similar to the inferences from the early geophysical explorations of modern West Antarctic ice streams (see Section 2.3). Detailed analyses of these metres-thick ‘soft’ tills, however, has shown that they are typically characterised by a range of structures indicative of subglacial shear and that such shearing tends to be concentrated into thin (0.1 – 0.5 m thick) zones, indicating that deformation was unlikely to be pervasive throughout the entire thickness and was more consistent with localised near-plastic rheology (Ó Cofaigh et al., 2005). These datasets also reveal that MSGLs are typically formed in this till layer (Fig. 8), and recent work suggests that MSGLs are clustered more closely and reach greater lengths in areas of relatively thicker till (6-12 m), compared to areas where the soft till is thinner or where the till is stiffer (Livingstone et al., 2016a).

The increasing resolution of swath bathymetry data from the late 1990s also allowed workers to map the large-scale distribution of landform-sediment associations on continental shelves surrounding present-day ice sheets in Antarctica (e.g. Wellner et al., 2001) and Greenland (e.g. Evans et al., 2009) and those formerly occupied by mid-latitude palaeo-ice sheets in North America (e.g. Shaw et al., 2006) and NW Europe (e.g. Ottesen et al., 2005; 2008). This allowed numerous palaeo-ice streams to be identified based on the location of cross-shelf troughs separated by zones of slower-flow on the adjacent banks (e.g. see recent inventories by Ottesen et al., 2005; Livingstone et al., 2012; Batchelor and Dowdeswell, 2014; Margold et al., 2015a). Collectively, these studies also indicated that palaeo-ice sheets likely had very similar velocity patterns to modern ice sheets (Margold et al., 2015b). Moreover, these studies revealed that the geomorphology within the cross-shelf troughs was similar to that observed in terrestrial settings (see Section 3.4), with MSGLs typically observed on the mid- to outer-cross-shelf troughs and sometime in association with lateral shear margin moraines, see Fig. 9 (Wellner et al., 2001; Ó Cofaigh et al., 2002; Ottesen et al., 2005, 2008). This helped to strengthen the association between specific glacial landforms (e.g. MSGLs, ice stream shear margin moraines) and ice streaming, particularly where they are observed in troughs that lie immediately in front of modern-ice streams (Shipp et al., 1999; Wellner et al., 2001; Anderson et al., 2002; Mosola and Anderson, 2006).

Marine geophysical studies were also able to show a general down-flow evolution of landforms along palaeo-ice stream beds that were not fully captured by the landsystems models developed earlier from terrestrial ice stream beds (e.g.

Stokes and Clark, 1999). Ó Cofaigh et al. (2002), for example, showed that bedforms exhibited a progressive increase in elongation with distance along the Marguerite Trough palaeo-ice stream in West Antarctica. The inner shelf was characterised by short, irregular drumlins and crudely streamlined forms at the bedrock surface. Across the mid-shelf and to the outer shelf, these bedforms become progressively more elongate, with MSGLs exhibiting very high elongation ratios up to 90:1. This was interpreted to reflect increasing ice velocities as the ice flowed over the transition from crystalline bedrock on the inner shelf, to a softer sedimentary substrate on the outer shelf (see also Wellner et al., 2001, Ottesen et al., 2008). Indeed, Ó Cofaigh et al. (2002) also hypothesised that the substrate was important in determining the flow mechanism of the ice stream, with basal sliding dominating over the crystalline bedrock and subglacial sediment deformation and/or ploughing over the softer sediments further down-stream. More recent work by Graham et al. (2009) highlighted the importance of substrate control on ice flow, but also recognised that different substrates can preserve a different, and often time-transgressive, record of ice stream flow, which is illustrated in Fig. 10. Using data from the western Amundsen Sea Embayment, they argued that whilst the outer shelf areas characterised by MSGLs formed in a soft substrate are likely to record the final phase of ice stream flow, landforms on the hard bedrock of the inner shelf areas, such as meltwater channels and grooved or streamlined bedrock features (Fig. 10), are more likely to have formed over longer time-scales and perhaps over multiple glaciations (see also Larter et al., 2009). Graham et al. (2009) also recognised clear spatial variability in the landform signature, such that bedforms of quite different size and shape can occur in close proximity, which might be related to local changes in basal conditions (e.g. sticky spots: see Sections 2.5 and 3.6), particularly in the inner to mid-shelf areas.

Marine geophysical records of formerly marine-based palaeo-ice streams have also identified landforms that lie transverse to ice flow and which often occur superimposed on the flow-parallel bedforms. The largest and most prominent of these are grounding zone wedges (GZWs), which are asymmetric sediment depocentres (with steeper ice-distal slopes) typically several kilometres long (down-flow) and ranging in thickness from 10-100 m (Fig. 9: Dowdeswell and Fugelli, 2012; Batchelor and Dowdeswell, 2015). They are typically composed of subglacial till and often contain seaward dipping reflectors which indicate progradation of sediment as it is deposited at the grounding line, as predicted by Alley et al. (1986), who suggested ‘till deltas’ would be a consequence of their deforming bed model (Section 2.3, Fig. 7). The formation of GZWs is, therefore, thought to require high rates of sediment transport to a grounding line position that is relatively stable for 10s to 100s years (Graham et al., 2010; Dowdeswell and Fugelli, 2012; Batchelor and Dowdeswell, 2015; Livingstone et al., 2016a). For example, radiocarbon dates that bracket the age of GZW formation on the outer shelf of Marguerite Trough require sediment fluxes >1,000 $m^3 a^{-1}$ (per meter width of the grounding line) (Livingstone et al., 2016a). Because their formation also requires a stable grounding line position, at least for a few decades, it has been noted that they tend to form in locations where the geometry of the ice stream’s terminus would have been constricted by a narrower or shallower topographic setting (Jamieson et al., 2012; Graham et al., 2010; Batchelor and Dowdeswell, 2015). Although progradation of a sediment via a deforming layer is often invoked to account for GZW formation, Dowdeswell and Fugelli (2012) noted that channel-like incisions are associated with some of these features and may suggest that delivery of sediment via pressurised subglacial meltwater, perhaps associated with drainage of subglacial lakes (see Section 4.3).

Narrower recessional moraines have also been identified superimposed on marine-based palaeo-ice stream beds (Shipp et al., 2002; Graham et al., 2010). These are typically <15 m thick and 300 m wide and possess sediment volumes that are an order of magnitude less than GZWs (Batchelor and Dowdeswell, 2015). They occur in assemblages containing 10s to 100s of parallel to sub-parallel ridges and are thought to form by ice pushing of sediment, including folding, faulting and thrusting, during minor ice sheet re-advances within overall retreat (Shipp et al., 2002; Ottesen and Dowdeswell, 2006, 2009). The significance of both GZWs and transverse ridges is that they record the style of retreat of marine-based ice streams, see Fig. 11. The absence of these features is often taken to infer rapid retreat of the ice stream, with sporadic occurrences of GZWs taken to infer episodic retreat punctuated by still-stands, and sequences of minor recessional moraines taken to infer relatively slow retreat of a grounded ice margin (Fig. 11) (Dowdeswell et al., 2008; Ó Cofaigh et al., 2008).

More recently, a smaller-scale series of transverse ridges (termed 'corrugations': Graham et al., 2013) have been identified in high resolution swath bathymetry on some palaeo-ice stream beds (Jakobsson et al. 2011; Graham et al., 2013). These ridges are just a few metres high and their crests are separated by ~60 to 200 m, with a spacing observed to decrease seaward (Jakobsson et al., 2011). They are thought to be formed by the periodic and tidally-modulated grounding of large icebergs (cf. Lien et al., 1989) within an ice mélange, and have been hypothesised to result from rapid disintegration of an ice shelf (Jakobsson et al., 2011). However, Graham et al. (2013) proposed a more generic origin that involves the forward motion of a ploughing ice keel that might occur beneath an ice shelf, or in the vicinity of the grounding line, or behind newly calved icebergs at the grounding line.

3.6. In Search of Palaeo-Ice Stream Sticky Spots

As noted in Section 2.5, several studies had confirmed the existence of sticky spots beneath active ice streams (e.g. Anandakrishnan and Alley, 1994; MacAyeal et al., 1995; Price et al., 2002; Joughin et al., 2004), but their geomorphological expression was largely unknown. This point was emphasized by MacAyeal et al. (1995: p. 247) who stated that "if sticky spots exist and, if their influence on ice stream flow is important, *the characterisation of sticky spots* [emphasis added] may prove essential to understanding how ice sheets evolve with time". As a result, several workers recognized that the geomorphology on the beds of palaeo-ice streams offered huge potential to characterize their nature and distribution (see review in Stokes et al., 2007). Thus, it was argued that where subglacial bedforms (drumlins and MSGLs) were shorter and/or deviated around areas of exposed bedrock or topographic highs on the ice stream bed, that these patterns represented sticky spots (Hodgson, 1994, Clark and Stokes, 2001, Stokes et al., 2007). It was also noted that these areas were particularly common in the onset zones of marine-based ice streams (e.g. Fig's 9, 10) where ice flow was likely accelerating, but had not reached typical streaming velocities (Canals et al., 2000; Wellner et al., 2001; Ó Cofaigh et al., 2002; Shipp et al., 2002; Graham et al., 2009). However, Clark and Stokes (2001) noted a more unusual down-stream decrease in bedform elongation on the M'Clintock Channel Ice Stream bed in the Canadian Arctic Archipelago (Fig. 5), whereby MSGLs transitioned into shorter drumlins that were located amongst areas of exposed bedrock. These bedrock-dominated areas were interpreted as till-free sticky spots (cf. Stokes et al., 2007) and Clark and Stokes (2001) hypothesised that 'frictional shut-down' of the ice stream occurred when the soft lubricating sediments at the bed became non-existent. The small incorporation of erratic material in till samples from the ice stream bed (~4%: see Hodgson, 1993) also indicated that sediment supply from upstream was minimal, highlighting the importance of till continuity for the operation of some ice streams (cf. Alley, 2000).

Variations in the routing and pressure of subglacial meltwater were also becoming increasingly linked to variations in glacier bed stickiness and the concept of a 'mosaic' of deforming and stable spots was becoming widely accepted (e.g. Boulton et al., 2001, Piotrowski et al., 2004). Such variations had already been invoked to explain the shut-down of Ice Stream C, e.g. the re-routing of subglacial meltwater and/or its freeze-on to the base of the ice (Anandakrishnan and Alley, 1994; Bougamont et al., 2003a). The sedimentological evidence of the deforming bed mosaic was described in Piotrowski et al. (2004) and evidence of 'stick-slip' behavior had been described in tills from several palaeo-ice stream beds (Hicock and Dreimanis, 1992; Lian et al., 2003; Knight, 2002). The geomorphological expression of this type of sticky spot, however, was largely unknown, but several workers noted the rather unusual occurrence of transverse features, known as ribbed moraine, superimposed on highly elongate bedforms that were the classic signature of ice stream flow (e.g. Fig. 4: Dyke and Morris and Morris, 1988; Dyke et al., 2002) .

Ribbed moraines form transverse to ice flow and generally occur in fields of closely spaced ridges of the order of 10 m high, 100 m wide and 1,000 m long (Hättestrand and Kleman, 1999; Dunlop and Clark, 2006). Stokes et al. (2007) noted that they are associated with palaeo-ice stream beds in two situations: (i) exclusively in the onset zone, merging into drumlins and MSGLs that exist down-stream (Fig. 4: Dyke and Morris, 1988; Dyke et al., 1992; De Angelis and Kleman, 2008); and (ii), in discrete patches within the main ice stream trunk, clearly superimposed on top of MSGL and drumlins, see Fig. 12 (Stokes and Clark, 2003; Stokes et al., 2006). Their occurrence on ice stream beds is unusual because they are typically found in the interior regions of ice sheets (Aylsworth and Shilts, 1989; Hättestrand, 1997; Kleman and Hättestrand, 1999). Thus, Stokes et al. (2007) argued that their appearance likely indicates an abrupt change in ice dynamics and most likely a switch to 'slow flow' that is related to the development of sticky spots.

In ice stream onset zones, ribbed moraines have been reported to develop at the transition between the cold-based catchment and warm-based ice stream (Dyke and Morris, 1988; Dyke et al., 1992; Hättestrand and Kleman, 1999; De Angelis and Kleman, 2008). Dyke and Morris (1988) were the first to document ribbed moraines in the onset zone of a palaeo-ice stream on Prince of Wales Island in the Canadian Arctic Archipelago (Fig. 4) and they invoked oscillations of the basal thermal regime between cold and warm-based ice that led to acceleration and deceleration of ice velocity and attendant infolding and stacking of debris. De Angelis and Kleman (2008) also noted that the geomorphological imprint of four ice stream onset zones they studied in the NE Laurentide Ice Sheet (including the one studied by Dyke and Morris, 1988) matched the typical characteristics of partially frozen beds, e.g. palimpsest fluvial landscapes or older glacial landscapes juxtaposed with areas with fully developed glacial lineations and often associated with ribbed moraines (Kleman and Borgström, 1996).

In contrast, there are a few cases where ribbed moraines occur in the main trunk of the ice stream. In the case of the Dubawnt lake Ice Stream bed, central Canada (see Fig. 12), Stokes and Clark (2003) reported their patchy appearance superimposed on MSGLs, and suggested that they formed through glaciotectonic shearing and stacking of debris under compressional ice flow that resulted from a loss of water in the till layer, perhaps related to basal freeze-on (cf. Stokes et al., 2008). More recently, Trommelen et al. (2014a) have also linked ribbed moraines to sticky patches with the subglacial bed mosaic of northern Manitoba. They argued that 'pristine' ribbed moraines were preserved under cold-based ice, but that other patches of ribbed moraine were partially to fully streamlined during a transition to warm-based ice during deglaciation. Interestingly, their detailed study of till composition and dispersal under the Hayes Lobe (terrestrial ice stream) in NE Manitoba (Trommelen et al., 2014b) revealed the presence of sticky spots that were identifiable as islands of 0.2 to 4 km wide patches of till dominated by a local composition, but surrounded by a regional farther-travelled till sheet. The cause of these inferred sticky spots is intriguing because they occur in an area of abundant warm-based streamlined bedforms, but they are possibly related to variations in subglacial water pressure (Trommelen et al., 2014b).

Most recently, Winsborrow et al. (2016) noted the presence of glaciotectonic rafts and hill-hole pairs on an ice stream bed associated with the former Barents Sea Ice Sheet and argued that they represent the geomorphological expression of sticky spots. In this location on the continental shelf, however, they noted that the sticky spots coincided with subsurface shallow gas accumulations and that the formation of gas hydrates could have desiccated, stiffened, and thereby strengthened the subglacial sediments, which led to the development of a sticky spot. On the basis of these observations, they suggested that a further cause of sticky spots is likely related to pore-water piracy and sediment stiffening due to subglacial gas-hydrate accumulation, and that such a process could be more prevalent under extant ice streams than hitherto recognised.

3.7. Summary

Around the same time that ice streams were gaining prominence in glaciology in the 1980s, glacial geomorphologists were attempting to identify their 'footprints' on the beds of palaeo-ice sheets. Denton and Hughes (1981) were the first to fully recognise the importance of ice streams to understanding past ice sheet dynamics, but their reconstruction of ice stream locations in Northern Hemisphere ice sheets (Fig. 3a) was initially met with scepticism, which is perhaps understandable given that there were no criteria to objectively identify ice stream footprints in the palaeo record, and very few had been identified based on empirical evidence (see Section 2.3). During the late 1980s, however, pioneering work on erratic dispersal trains (e.g. Fig. 4: Dyke and Morris, 1988) was soon followed by the recognition that ice streams created a distinct geomorphology that included convergent bedform patterns, highly elongated glacial lineations (MSGLs), ice stream shear margin moraines, and major sedimentary depocentres (till deltas, grounding zone wedges, trough mouth fans). Collectively, these geomorphological criteria (cf. Stokes and Clark, 1999) formed a series of landsystem templates (e.g. Fig's 6, 10 and 11) that enabled their objective identification on palaeo-ice sheet beds. The increased use of satellite remote sensing and marine geophysical techniques led to the identification of hundreds of palaeo-ice streams, both onshore and offshore, such that some inventories are probably close to complete (Ottesen et al., 2005, 2008; Livingstone et al., 2012, Batchelor and Dowdeswell, 2014; Margold et al., 2015a, b; Dowdeswell et al., 2016).

By the early 2000s, we knew what a former ice stream bed looked like in terms of its geomorphology, including the key diagnostic landforms of rapid ice flow – mega-scale glacial lineations (Clark, 1993) – and new insights regarding how sticky spots might be manifest in the geomorphological record. These landforms have been used to reconstruct where and when ice streams operated (e.g. Ottesen et al., 2005; Stokes et al., 2016a) and as both input and constraint data for numerical modelling of individual ice streams (e.g. Jenson et al., 2005; Jamieson et al., 2012) and ice sheets (e.g. Stokes and Tarasov, 2010). These observations provided a clear geomorphological framework for those studying the beds of modern ice streams using geophysics (Section 2) and led to increased cross-pollination of ideas between the two disciplines. However, palaeo-ice stream beds are a time-integrated and often fragmentary record of ice stream operation that these data could say little about the time-scale over which processes were operating. Indeed, palaeo-ice stream research was in its infancy with respect to the time-scales over which the geomorphology was created and evolved and under what ice dynamical conditions (ice velocity, thickness, etc.). Smith et al. (2007: p.127) also noted that “the interpretation of former ice dynamics from the subglacial bedforms left behind following deglaciation is also limited by a lack of observed analogous features beneath modern glaciers”. This would require a renewed effort to observe the geomorphology and basal conditions on modern ice stream beds using geophysical techniques with much higher resolution.

4. Geophysical Investigation of the Geomorphology Beneath Modern Ice Streams

4.1. Discovery of Subglacial Bedforms Beneath Modern Ice Streams

Soon after the early work on the ‘pure’ Siple Coast ice streams in West Antarctica (Section 2.1), efforts were underway to investigate the basal conditions beneath Rutford Ice Stream, also in West Antarctica (King et al., 2004; Smith, 1997a, b; Smith et al., 2007). Unlike the Siple Coast ice streams, it flows through a deep asymmetric bedrock trough, bounded on one side by the Ellsworth Mountains, and with ice thicknesses ranging from ~2000-3200 m and ice velocities 300-400 m a^{-1} (Smith, 1997a, b; King et al., 2007). As such, its characteristics are more typical of the majority of extant ice streams in Greenland and Antarctica that are, to a large extent, controlled by underlying topography. Building on the earlier work on this ice stream (e.g. Stephenson and Doake, 1982; Doake et al., 1987), Smith (1997a, b) used seismic reflection profiles from transects perpendicular to the ice stream bed to show that the characteristics of the bed material varied both along and across the ice stream. The majority of the bed was interpreted as dilatant water-saturated sediments undergoing pervasive deformation, but other areas were interpreted to be non-deforming and perhaps accommodating basal sliding (Smith, 1997a). Smith (1997b) also noted that the proportion of the ice stream width over which bed deformation occurred appeared to increase down-stream over a distance of ~50 km (see also Doake et al., 2001). Localised areas of very high basal shear stress (sticky spots) were also inferred (see later work by Smith, 2006), although Smith (1997a) noted little correspondence between these areas of the ice stream bed and surface velocity data or satellite imagery of the ice stream surface. Of significance to ice stream geomorphology was that, although the profile data were only collected in 2-dimensions, Smith (1997a) tentatively interpreted a 'bump' approximately 400 m wide and 50 m high as a drumlin composed of soft deforming sediments sitting on a more rigid substrate. This appears to represent the first direct evidence of a subglacial bedform beneath an active ice stream.

Later work by Smith et al. (2007) collected seismic reflection data from the same location as those presented in Smith (1997a). In addition to the data collected from the original lines in 1991, Smith et al. (2007) presented data from 1997 and 2004, when the ice stream flow had not changed significantly. The repeat measurements enabled them to detect changes in the bed topography (geomorphology) and associated changes in subglacial hydrology (from acoustic impedance) over a 13 year period, shown in Fig. 13. Between 1991 and 1997, they measured erosion of sediment averaging 1 m a^{-1} across a 500 m wide area, which Smith et al. (2007) noted was much higher than previous estimates of subglacial erosion in other subglacial environments and under the Siple Coast ice streams (<0.1 to 100 mm a^{-1}: Alley et al., 1987, 1989; Alley et al., 2003; Hallet et al., 1996;

Humphrey and Raymond, 1994), although more comparable with recent work from Pine Island Glacier, which estimates similarly high rates of 0.6 ± 0.3 m a^{-1} (Smith et al., 2012). They ruled out the possibility that the erosion was due to subglacial meltwater because any free water at the ice-bed interface would be expected to occur elsewhere in the deeper parts of the ice stream's trough, but they were unable to pinpoint the exact mechanism through which the sediment was removed. The second comparison between the lines acquired in 1997 and 2004 indicated that the erosion across this area had ceased and that a mound of material 10 m high and 100 m wide appeared (Fig. 13). An additional flow-parallel line from 2004 also suggested that the mound extended at least 1 km upstream. Smith et al. (2007) noted that the dimensions and sedimentary characteristics of the mound were typical of a drumlin, thereby confirming the earlier supposition from Smith (1997a).

Ascertaining the mechanism of drumlin formation was more difficult, but Smith et al. (2007) considered two possibilities. The first invoked an advecting groove in the base of the ice that was infilled by soft deforming sediment, similar to that suggested for flutes (cf. Boulton, 1976). The second invoked a rheological instability in the bed material that had been modelled by Hindmarsh (1998a, b). The former mechanism was favoured on the basis that the feature was longitudinally continuous as well as relatively invariant with time (in both geometry and properties). Notwithstanding these uncertainties, the key conclusion was that an ice stream can reorganise its bed rapidly over just a few years or less, and the direct observation of the formation of a subglacial bedform had broken new ground in the study of ice stream geomorphology.

Around 160 km further upstream, in the onset zone of Rutford Ice Stream, King et al. (2007) presented both seismic and radar data to image the bed in a region where the ice flow velocity accelerated from 72 to >200 m a^{-1}. In the slower-moving part they observed a transverse moraine 2 km wide and 1.5 km long, which they interpreted to be composed of unconsolidated sediment undergoing active deformation. King et al. (2007) suggested that this feature might be analogous to a drumlinised ribbed moraine (Dunlop and Clark, 2006, Trommelen et al., 2014b). About 30 km downstream, where velocities exceeded 95 m a^{-1}, they observed several "drumlins of classical form" with elongation ratios between 1:1.5 and 1:4 (King et al., 2007: p. 665). Based on the composition of the drumlins and their conformity with the ice base, they interpreted the drumlins as active depositional features that were similar to the feature observed by Smith et al. (2007), and were consistent with erosion and deposition within a mobile deforming bed (Boulton and Hindmarsh, 1987). In addition to the confirmation that bedforms were being created under the ice stream, King et al.'s (2007) study was significant because it was the first to demonstrate a clear link between ice velocity and bedform elongation that had only been hypothesised from work on palaeo-ice stream beds (see Section 3.3), i.e. the drumlinised ribbed moraine were found under ice velocities of ~72 m a^{-1}, whereas the more elongate drumlins were observed beneath ice flowing at ~125 m a^{-1} (see also Smith and Murray, 2009). King et al. (2007) also noted that the drumlin further down-stream (observed by Smith et al., 2007) was lower (10 m vs 30–50 m) and narrower (100 m vs 200–500 m) than the drumlins they had imaged in the onset zone.

4.2. Formation of MSGLs beneath a West Antarctic Ice Stream

Interestingly, Smith et al. (2007) had ended their paper documenting drumlin formation under Rutford Ice Stream by suggesting that future work might ascertain "whether or not the drumlin evolves into a mega-flute" (p. 130). This point was, perhaps, a tacit acknowledgement that subglacial bedforms under the fast-flowing trunk of the ice stream would be expected to be much longer than drumlins and more similar to the MSGLs (cf. Clark, 1993) that had been observed on palaeo-ice stream beds (Section 3.3). Ascertaining the presence of MSGLs, however, would require data from a much larger area and Smith and Murray (2009) were able to do this by combining a number of previously-collected seismic lines (e.g. reported in Smith, 1997a, b) distributed over an area of 140 km^2. Interpolation of these datasets allowed them to construct schematic cross section of the ice stream bed, shown in Fig. 14, which indicated dilatant deforming till in two troughs that run either side of a central high. The central high was characterised by a mosaic of deforming and stiffer till (cf. Piotrowski et al., 2004; see Section 3.6), and they noted a local downstream transition from deforming to stiffer till, where they inferred basal sliding was more important. They confirmed the location of previous mounds and drumlins (see Smith, 1997a, b; Smith et al., 2007) and noted how some of the deforming material associated with these bedforms extended down-stream over the stiffer till.

Significantly, they also noted that the 'bump', previously interpreted as a drumlin (e.g. Smith 1997a, b), continued down-stream for at least 17 km and, as a result, they suggested that it should be referred to as a MSGL (Clark, 1993).

Subsequently, King et al. (2009) acquired new high resolution radar data from a large area of the ice stream bed, including the area where previous seismic lines had been acquired (Smith et al., 1997a, b; Smith and Murray, 2009). Combining the new radar data with the previously-collected seismic data clearly revealed an assemblage of MSGLs beneath ice flowing at around 375 m a^{-1}. This assemblage, shown in Fig. 15, was characterised by a pattern of ridges and troughs with wavelengths transverse to ice flow of 300 to 1,000 m and with peak-to-trough amplitudes ranging from 5 to 90 m, with a mean of 10 m. The longest ridges extended for >18 km and had elongation ratios ranging from 15:1 to >35:1. Indeed, the MSGLs observed under Rutford Ice Stream were largely indistinguishable from those observed on palaeo-ice stream beds (Fig. 15b) and this study provided the first conclusive evidence that MSGLs are diagnostic of ice stream flow, which had been postulated well over a decade earlier (Clark, 1993). The King et al. (2009) 'bed-map' has recently been extended in the downstream direction to cover an area 18 x 40 km and is now available as a dataset that comprises both ice surface, ice thickness and bed elevation data (see King et al., 2016)

As in previous work (Smith, 1997a, b, King et al., 2007; Smith and Murray, 2009), the MSGLs were inferred to be developed in areas of dilatant deforming till that formed part of a dynamic sedimentary system that underwent both erosion and deposition on decadal time scales. King et al. (2009) also confirmed that the basal sliding zones, where the till is stiffer and sliding is inferred to be taking place across the surface of the sediment, coincided with areas of the bed that had a more subdued topography and with fewer and poorly streamlined bedforms. The binary distribution of acoustic impedance data were also deemed to be compatible with observations from marine geophysical observations of MSGLs formed in a dilatant 'soft' till overlying a more consolidated and stiffer till (Fig. 8) (Dowdeswell et al., 2004; Ó Cofaigh et al., 2005; 2007). King et al. (2009) also noted that none of the existing theories of MSGL formation can readily explain the observations, but they favoured a till instability mechanism (cf. Hindmarsh, 1998a, b) based on the close association between MSGLs and the areas of dilatant deforming sediment.

Advances in geophysics were also allowing other ice stream beds to be imaged at higher resolution. Jezek et al. (2011) applied radar tomography methods to very high frequency airborne synthetic aperture radar data to measure ice thickness on Jakobshavn Isbræ in West Greenland and construct a 3-dimensional map of the bed topography over an area 5 x 20 km. They were able to detect, for the first time beneath the Greenland Ice Sheet, assemblages of elongate ridge-groove landforms oriented in the direction of the ice flow. These had wavelengths of 150 to 500 m, amplitudes of 10 to 30 m and extended along flow >10 km, often with a slightly sinuous nature. Although these features broadly resembled those identified on Rutford Ice Stream (King et al., 2009), Jezek et al. (2011) noted that the bed topography and its roughness and structure have the appearance of a bedrock surface with little to no unconsolidated sediment. Calculations of basal drag also indicated a much higher basal shear stress (around 120 kPa) under this ice stream than, for example, those typically found under the Siple Coast ice streams (see Section 2.2). Thus, the features were interpreted as 'bedrock mega-grooves' that may, in part, be influenced by the underlying bedrock structures, similar to those observed in hard-bed palaeo-ice stream landform assemblages (Smith, 1948, Bradwell, 2005; Bradwell et al., 2008, Eyles, 2012; Krabbendam et al., 2016). If correct, this implies that the features are largely erosional.

More recently, Schroeder et al. (2014) used data from radar scattering to image the bed of Thwaites Glacier, West Antarctica. In the upstream region of the ice stream, they were able to detect a corrugated bed with amplitudes (~20 m) and wavelengths (~500 m) that were similar in scale to MSGLs found under Rutford Ice Stream (King et al., 2009) and on palaeo-ice stream beds (Livingstone et al., 2012; Spagnolo et al., 2014). However, further downstream, they found topographies that were consistent with a much rougher bed, characterised by outcropping bedrock. This interpretation was also consistent with previous observations of the subglacial water system (Schroeder et al., 2013) and modelling of the basal shear stresses (Joughin et al., 2009), which found relatively high basal shear stress and channelized subglacial water in the lower trunk region, but with relatively lower basal shear stresses and a distributed water system in the upper trunk and tributaries of the ice stream (see Section 4.3). Elsewhere, on Thwaites Glacier, airborne ice-penetrating radar data aimed at ascertaining the controls on the location of the eastern margin of the ice stream may have inadvertently detected an ice stream shear margin moraine (see Fig. 4a in MacGregor et al., 2013), although this was not discussed in that study.

4.3. Direct Observations of Subglacial Hydrology and Till Deltas

The geomorphology associated with subglacial meltwater storage and drainage was also beginning to be uncovered using new remote sensing and geophysical processing techniques. The presence of a distributed system of shallow 'canals' cut into the underlying soft till had been predicted by theory (Alley, 1989a; Ng, 2000; Walder and Fowler, 1994), but evidence of their existence beneath ice streams was very limited on both modern and palaeo-ice stream beds. Carsey et al. (2002) had previously drilled into a 1.4 m deep water-filled gap close to the the shear margin of Ice Stream C, and Atre and Bentley (1994) interpreted strong reflections >50 km upstream of the grounding line under Ice Stream B to be ponded water . The pioneering borehole drilling by Engelhardt et al. (1990) was also inferred to have punctured an active basal drainage system, but direct evidence of the configuration of subglacial channels or canals remained elusive.

One of the first studies to document the geomorphology associated with meltwater activity under an active ice stream was by King et al. (2004), who deployed seismic techniques on Rutford Ice Stream to document the presence of a water-filled canal in deforming till, measuring at least 1 km by 0.2 km and with depths < 1 m. King et al. (2004) noted that the estimated water layer thickness of their canal was between 0.4 and 0.6 m and that its dimensions were long (> 1 km) and thin (<200 m) and not topographically constrained. Thus, they inferred that they had imaged part of a drainage system, rather than an isolated pond. Later work by Murray et al. (2008) used radar and seismic data from the same ice stream to infer several locations where water was interpreted to exist as shallow canals in locations where till was assumed to be deforming. These features were 50 m wide and comprised of water <0.2 m deep. They noted that the features were considerably wider than the 3–5 m predicted by theory (Ng, 2000) or the 1 m inferred by Engelhardt and Kamb (1999), but were narrower than the canal interpreted by King et al. (2004).

More recent work by Schroeder et al. (2013) on Thwaites Glacier, West Antarctica, also reported substantial volumes of water ponding in distributed canals and increasing in area where ice flow approaches a major bedrock ridge. They used the angular distribution of energy in radar bed echoes to characterise the configuration of subglacial water systems and found that a system of broad distributed channels transitioned into a network of fewer concentrated channels on and over the ridge. The transition between these systems occurred with increasing ice surface slope and basal water flux, which would be predicted by theory (Walder and Fowler, 1994). Schroeder et al. (2013) also noted that the transition to a channelized system coincided with an increase in basal shear stress (Joughin et al., 1999), which would also match theoretical predictions.

Elsewhere, the detection of subglacial lakes beneath modern ice masses was a rapidly evolving field (Siegert et al., 2005) and several studies had inferred water bodies (small lakes/ponds) beneath ice streams (e.g. Gray et al., 2005; Fricker et al., 2007; Peters et al., 2007; Fricker and Scambos, 2009; Smith et al., 2009). Gray et al. (2005), for example, detected ice surface uplift and subsidence on both Kamb Ice Stream (formerly Ice Stream C) and Bindschadler Ice Stream (formerly Ice Stream D), which they attributed to transient movements of subglacial water. The inferred volumetric transfers of water over a 24 day period (~20^7 and ~10^7 m^3 on Kamb and Bindschadler, respectively) constituted two major outburst floods of considerable magnitude. Peters et al. (2007) also inferred a major meltwater body beneath a tributary of Bindschadler Ice Stream that was trapped by a local reversal in the ice surface slope where the ice stream flowed over rough bed topography. The feature was 5-10 m thick (deep) and at least 1 km long, and existed in a region of the bed with alternating regions of soft and stiff till.

Around the same time, Bindschadler and Choi (2007) suggested that ice stream onset zones would likely be characterised by the presence of subglacial lakes due to the interdependence between rapid ice flow, surface topography and the spatial distribution of subglacial water. Using theories of how subglacial topography is transmitted to the ice surface topography, and how this affects subglacial hydraulic potential and water storage, they posited that the transition from tributary to full ice stream flow closely agrees with where subglacial water may first be stored. It is clear, however, that shallow (5-10 m) lakes can exist closer to the grounding line, such as Subglacial Lake Whillans (e.g. Horgan et al., 2012). A further key study by Fricker et al. (2007) documented the presence of several relatively large (~50-330 km^2) subglacial lakes under Mercer and Whillans Ice Streams (formerly Ice Stream's A and B, respectively) that were associated with ice surface movements of the order of metres over several months, thought to be related to lake volume changes. Later work by Smith et al. (2009) also drew attention to the observation that some subglacial lakes in Antarctica appear to be clustered beneath ice streams and other work noted their correspondence to subglacial topography, such as ridges, or areas where there are abrupt contrasts in basal conditions, such as sticky spots (Fricker et al., 2007, 2010; Fricker and Scambos, 2009; Sergienko and Hulbe, 2011). Indeed, Fricker et al. (2010) pointed out that the juxtaposition of subglacial lakes and sticky spots is likely due to the fact that the high-friction sticky spots help generate additional meltwater and also cause strong gradients in ice thickness, which feeds back to influence the pattern of hydraulic potential at the bed. This is supported by modelling undertaken by Sergienko and Hulbe (2011), who demonstrated that ice flow over a sticky spot causes favourable conditions for the development of subglacial lakes.

Thus, it was becoming increasingly clear that ponded meltwater, and channels and canals, were perhaps more abundant on ice stream beds than previously recognised (e.g. Sergienko and Hulbe, 2011), and there was clear evidence that it water could drain rapidly (Gray et al., 2005; Wingham et al., 2006; Stearns et al., 2008). However, the configuration of the hydrological system during such drainage events was largely unknown. If these drainage events were widely distributed across the bed (i.e. as a thin film or in shallow canals), it could considerably enhance basal lubrication, but flow through a single or small number of larger channels would have much less influence. To assess subglacial water flow to the grounding line of the Siple Coast ice streams, Carter and Fricker (2012) combined lake volume estimates derived from remote sensing with a model for subglacial water transport. They noted that subglacial meltwater outflow tends to concentrate in distinct locations (see also Le Brocq et al., 2013) and that whilst the mean outflow at the grounding line is around 60 m^3 s^{-1}, it can increase to 300 m^3 s^{-1} during synchronized flood events. Carter et al. (2017) have also modelled the drainage of lakes in which channels are mechanically eroded into the underlying deformable subglacial sediment. They noted that conventional models based on 'R-channels' cut into the overlying ice are unable to reproduce the timing and magnitude of lake drainage events under ice streams. Their modelling showed how water pressures change during a flood event, such that the initial drainage is associated with a high pressure distributed system, but that this system subsequently collapses to a channelized system. Their modelling was able to replicate both the inferred magnitudes and recurrence intervals of lake-volume changes derived from laser altimeter data on Mercer and Whillans Ice Stream. Release of large volumes of meltwater could also generate significant subglacial erosion and transport, but the extent to which these events do 'geomorphological work' under ice streams is largely unknown (see Section 5.1).

In relation to sediment transport, geophysical investigations close to the grounding line of active ice streams were now also able to confirm the presence of till deltas that had been hypothesised in the early work by Alley et al. (1987a; 1989a) and had subsequently been detected on the beds of numerous marine-based palaeo-ice streams (see Section 3.5). Anandakrishnan et al. (2007) conducted radar surveys in the grounding zone of Whillans Ice Stream, to image a wedge of subglacial sediment with a maximum thickness of 31 m and extending for >12 km, with the imaged volume on the order of 10^5 m^3 per unit width (they acknowledged that they were unable to detect the entire feature). Given that the grounding line has likely been in this location for 1,000 years (Conway et al., 1999), they calculated a sediment flux of around 10^2 m^3 m^{-1} a^{-1}, which is at the lower end originally predicted by Alley et al. (1989a) for the same ice stream (10^2 -10^3). This rate also implies long-term erosion rates of just over 0.1 mm a^{-1} over Whillans Ice Stream and its catchment, although Anandakrishnan et al. (2007) noted that erosion is likely to be concentrated in discrete areas where basal melting occurs over poorly consolidated sedimentary substrates. Elsewhere, Doake et al. (2001) used a similar approach to Alley et al. (1989a) to estimate a higher till flux of 4 x 10^3 m^3 m^{-1} a^{-1} at the grounding line of Rutford Ice Stream. As in most previous work, the mechanism of sediment transport was assumed to be via mobile layer of deforming sediment tens of centimetres or more, rather than via meltwater and/or melt-out of basal debris. Anandakrishnan et al. (2007) drew attention to the similarity between the wedge and grounding zone wedges identified on the sea-floor beyond the limits of the Ross Ice

Shelf (Shipp et al., 1999; Mosola and Anderson, 2006; Domack et al., 1999; see Section 3.5) and, in a companion paper, Alley et al. (2007) demonstrated how such a wedge serves to locally reduce water depths and thereby helps stabilise the ice stream's terminus. In locally stabilising the grounding line, the development of a wedge also allows for focussed deposition in that location, rather than the spreading of deposits more uniformly over a broader area (Anadakrishnan et al., 2007).

4.4. Sticky Spots and 'Traction Ribs'

Following the early work that discovered sticky spots beneath active ice streams (see Section 2.5), new geophysical techniques coupled with inversions of basal drag based on control methods began to uncover spatially organised patterns of high basal drag approximately perpendicular to ice stream flow (MacAyeal, 1992; MacAyeal et al., 1995; Price et al., 2002; Joughin et al., 2004). MacAyeal et al. (1995) had first noted bands of higher basal shear stress several tens of kilometres in length and several kilometres in width on Ice Stream E, which they referred to as 'sticky strings'. Price et al. (2002) also detected transverse bands of alternating low and high basal shear stress along a tributary of Bindschadler Ice Stream (formerly Ice Stream D), with the bands of high basal shear stress on the order of 10 times the mean values elsewhere. They attributed these bands to the possibility of bedrock ridges striking obliquely across the ice stream and also pointed out that areas of the low shear stress likely contained ponded water in between (see previous section). Joughin et al. (2004) subsequently utilised higher resolution velocity datasets to improve the calculations of basal shear stress beneath all of the Siple Coast ice streams. Their results were consistent with previous studies, which indicated a weak bed beneath ice streams, interrupted by localised sticky spots, and with clear patterns of transverse banding apparent, particularly in the tributaries of the ice streams. The cause of this 'banding' in basal shear stress was unknown, but Joughin et al. (2004) speculated that they might result from ice flowing across the topographic and tectonic fabric of the underlying terrain, with ice motion accomplished through a combination of soft-bed deformation over sedimentary basins and basal sliding over bedrock ridges. More recently, Peters et al. (2007) also noted three broad bands along a 12 km profile in the surface slope of Bindschadler

Ice Stream that were linked to alternating areas of high and low basal shear stress and with water ponding in the areas of low basal shear stress.

More recently, two papers have drawn attention to the apparent ubiquity of these regular patterns of basal shear stress on a number of Antarctic and Greenlandic ice streams, and including non-ice stream areas (Sergienko and Hindmarsh, 2013; Sergienko et al., 2014). These studies utilised higher resolution data on ice velocity, elevation and thickness to calculate basal shear stresses using standard inverse techniques from previous work (MacAyeal, 1992). The increased resolution of these inversions, compared with previous efforts, revealed the presence of regular ‘rib-like’ patterns of very high basal shear stress (typically ~200–300 kPa) embedded within much larger areas of near-zero basal shear stress (Fig. 16). They were found to be widespread throughout areas of slow and fast flow, but they were most pronounced in arcuate patterns within the onset zone of ice streams (Sergienko et al., 2014), as previously noted by Joughin et al. (2004). These patterns had a clear surface expression and were seen in the calculations of the driving dress, which are independent of the inversion technique, spatial resolution and its regularisation method. These enigmatic patterns were termed ‘traction ribs’ (Sergienko and Hindmarsh, 2013) and varied in size from several kilometres to tens of kilometres in length and a few kilometres wide, with the long axes aligned approximately transverse to ice flow, but often deviating by ~30°–60° from ice flow direction (Sergienko and Hindmarsh, 2013). It is not clear what causes these regular patterns in basal shear stress, but Sergienko and Hindmarsh (2013) noted the correspondence between the traction ribs and areas of high hydraulic gradient, and suggested that subglacial water may play a role in rib formation. They suggested that the ribs are likely to be regions of variable effective pressure that cause localised strengthening along the base.

In terms of the geomorphology, the resolution of the geophysical data was unable to reveal whether ‘traction ribs’ had a topographic expression at the bed and whether they were related to an underlying geological control, such as bedrock bumps or ridges (as proposed by Joughin and others, 2004). However, Sergienko and Hindmarsh (2013) noted that their pattern (Fig. 16), if not their dimensions, resembled subglacial bedforms observed on palaeo-ice sheet beds, such as the ‘mega-ribs’ reported by Greenwood and Kleman (2010), and the far more ubiquitous 'ribbed moraines' (Hättestrand, 1997; Hättestrand and Kleman, 1999; Dunlop and Clark, 2006). Despite their similarity in pattern, it was noted that the traction ribs were intermediate in scale between ribbed moraines and mega-ribs and that ribbed landforms at this scale had not been imaged on palaeo-ice streams. Recently, however, Stokes et al. (2016b) reported ribbed bedforms from palaeo-ice stream beds in western Canada that resembled both the pattern and scale of the traction ribs reported by Sergienko and Hindmarsh (2013) and Sergienko *et al.* (2014) (see Fig. 16e, f). Using Digital Elevation Models, they mapped >1,000 rib-like features on four previously-identified palaeo-ice streams from the Interior Plains of Western Canada (Evans et al., 1999, 2008, 2014; Ross et al., 2009; Ó Cofaigh et al., 2010a). Measurements of their length, width, spacing and amplitude showed that they were near-identical to traction ribs, which is consistent with the notion that similar bedforms exist beneath active ice streams and which help explain the transverse banding in basal shear stress. The formation of the ribbed bedforms remains conjectural, but the observations from palaeo-ice streams (Stokes et al., 2016b), coupled with those from modern ice masses and numerical modelling, suggest they might be related to wave-like instabilities occurring in the coupled flow of ice and till and modulated by subglacial meltwater drainage (Dunlop et al., 2008; Chapwanya et al., 2011; Sergienko and Hindmarsh, 2013, Fowler and Chapwanya, 2014). Once initiated, high basal shear stresses over the upstanding ridges might also induce glaciotectonism of subglacial sediments, as has been reported for some of the transverse ridges in western Canada (Evans et al., 1999, 2008, 2014).

4.4. *Summary*:

Until the turn of the century, the limited resolution of geophysical techniques had prevented the detailed imaging of the geomorphology of active ice stream beds and much of our understanding of their geomorphology was gleaned from investigations of palaeo-ice stream beds (Section 3). However, advances in the resolution and processing of remote sensing and surface and airborne geophysical techniques led to some important discoveries of subglacial bedforms that had previously evaded detection beneath modern ice masses. These studies were able to confirm that many of the landforms first identified on palaeo-ice stream beds, such as MSGLs, could be directly linked to ice stream flow (King et al., 2009; Smith and Murray, 2009). They also confirmed the link between ice velocity and bedform elongation (cf. King et al., 2007), which was initially proposed based only on observations from palaeo-ice stream beds (e.g. Clark, 1993). They reinforced the notion that basal conditions are spatially variable, especially in the onset zone of ice streams, and that their motion over a mosaic of sticky and slippery spots is likely accommodated by both till deformation and basal sliding (Smith, 1997a). Higher resolution datasets have also enabled more detailed inversions of basal drag, which revealed a hitherto unrecognised regularity in bed stickiness ('traction ribs') on many ice stream beds (Joughin et al., 2004, Sergienko and Hindmarsh, 2013; Sergienko et al., 2014), which may have a geomorphological expression as ribbed bedforms observed in the palaeo-record (Stokes et al., 2016b). These sticky spots and traction ribs have been shown to be intimately linked to subglacial hydrology (Peters et al., 2007), and the last decade or so has seen an increase in the detection of subglacial lakes (Gray et al., 2005; Peters et al., 2007; Smith et al., 2009) and meltwater drainage systems under active ice streams (King et al., 2004; Schroeder et al., 2013).

Repeat surveying of active ice stream beds (e.g. Smith et al., 2007; Smith and Murray, 2009) has also revealed an active sedimentary system where material is being eroded, transported, and deposited over very short, likely sub-annual time-scales, which is impossible to glean from palaeo-ice stream geomorphology. Moreover, these studies have also allowed estimates of the rates of subglacial erosion. Long-term erosion rates averaged over ice stream catchments have been estimated to be 0.1 - 0.5 mm a^{-1} (Alley et al., 1986; Anadakrishnan et al., 2007), but locally may be as high as 1 m a^{-1} in some locations (Smith et al., 2007; 2012). Once mobilised, this sediment is thought to be transported down-stream, with sediment fluxes at the grounding line typically of the order of 10^2 - 10^3 m^3 m^{-1} a^{-1} (Alley et al., 1989a; Anandakrishnan et al., 2007), but sometimes higher (Doake et al., 2001)

5. Future Challenges and Opportunities: Moving from Form to Process in Ice Stream Geomorphology

It is clear that our understanding of ice stream geomorphology has grown rapidly since the early pioneering investigations in West Antarctica in the 1980s (Section 2).

Initially, our knowledge of the morphology of ice stream beds was largely gleaned from geophysical investigations of modern ice streams in Antarctica. This was soon augmented by observations from palaeo-ice sheets beds and the recognition that ice streams created a distinctive geomorphology that is very different from slower ice sheet flow (e.g. Dyke and Morris, 1988). This led to a rapid convergence of ideas and collaboration between glaciology and glacial geomorphology, and went some way to addressing Boulton's (1986) earlier plea for greater interdisciplinarity between these sub-disciplines. Moreover, the last decade or so has has seen even greater collaboration between these two sub-disciplines (see also Bingham et al., 2010) and this is largely due to recent technological and methodological advances that have enabled the geomorphology of active ice stream beds to be imaged with unprecedented resolution (Bingham et al., 2010), revealing subglacial bedforms whose interpretation has been considerably aided by their palaeo-counterparts (e.g. Smith and Murray, 2009; King et al., 2009). As such, we now have a very detailed knowledge of the morphology of ice stream beds and an emerging understanding of how this relates to the distribution of sediments, water and basal shear-stresses. However, our understanding of how the morphology (form) relates to processes is much more limited, and I argue that this represents the next major challenge and will require even closer cooperation between glaciologists, glacial geomorphologists, sedimentologists, together with numerical modelling. For example, the precise mechanisms through which sediment is eroded, transported and deposited under ice streams are largely unknown and remain subject to debate (e.g. deep deforming layer versus shallow deforming layer, channelized meltwater versus canals and water films). These processes are intimately linked to both the growth and decay of subglacial bedforms, but our understanding of bedform creation and evolution is also poorly understood. The way in which ice stream geomorphology influences or even controls ice stream flow over a range of time-scales is also poorly understood, and yet these are a crucial boundary condition of numerical models that aim to predict future ice stream dynamics and their potential contribution to sea level. These issues are now discussed as a series of key questions, together with future opportunities that might enable them to be answered.

5.1. How is Sediment Eroded and Transported Beneath Ice Streams?

The convergence of knowledge from studies of both palaeo- and active ice stream geomorphology indicates that ice streams are responsible for considerable sediment fluxes to their grounding lines, see Table 1. As noted by Bougamont and Tulaczyk (2003), this is a key feature of ice streams that must be reproduced by models that attempt to simulate their flow. Moreover, and as noted in Section 4.3, these high fluxes are important because sediment deposition at ice stream grounding lines can stabilise retreat by serving to locally reduce water depths (Alley et al., 2007). The mechanisms through which sediment is eroded and transported by ice streams is also intimately linked to their basal shear stress, which represents a key boundary condition for numerical modelling of ice stream/ice sheet dynamics (Ritz et al., 2015).

The values in Table 1 are mostly derived from bracketing the timing of the deposition of large Trough Mouth Fans (e.g. Nygård et al., 2007) and grounding zone wedges (e.g. Livingstone et al., 2016a), or from modelling sediment transport based on assumed ice and sediment properties (e.g. Jenson et al., 1995; Bougamont and Tulaczyk, 2003). Typical sediment fluxes are of the order of 100 to 1,000 m^3 m^{-1} a^{-1} (i.e. per meter width of ice stream terminus), but more extreme values have been estimated from some very large palaeo-ice streams in the Laurentide and Eurasian Ice Sheets. Given these typical values, a key question is: how is the sediment eroded and transported?

Much of the early work invoked viscous deformation of a metres-thick layer of saturated sediment (Alley et al., 1986; Blankenship et al., 1986), which can readily explain the high sediment fluxes and appears to be consistent with observations of both palaeo- (Jenson et al., 1995; Alley, 1991) and modern ice stream sediments and geomorphology (Alley et al., 1987a; Hindmarsh, 1998a, b; Anderson et al., 2002; Dowdeswell et al., 2004; Ó Cofaigh et al., 2005; 2007; Mosola and Anderson, 2006; King et al., 2009; Smith and Murray, 2009). This traditional model has been viewed as the simplest explanation, but it has been repeatedly challenged (and to a large extent been replaced) by evidence that sub-ice stream tills exhibit a near-plastic rheology (Kamb, 1991; Tulaczyk et al., 2000a; Tulaczyk, 2006; Iverson, 2010) and that any deformation should be restricted to a thin zone (of the order of centimetres rather than metres) at the top the till layer and/or accompanied by basal sliding across the surface of the till (Engelhardt et al., 1990; Engelhardt and Kamb, 1998).

This is important because if sub-ice stream tills behave as Coulomb-plastic substrates with deformation restricted to narrow shear planes, it potentially limits the thickness of the mobile sediment layer to just a few centimetres and precludes large sediment fluxes via a 'viscously-deforming bed' mechanism (Tulaczyk et al., 2001).

In order to reconcile a Coulomb-plastic rheology with relatively high sediment fluxes (Table 1), it is necessary to invoke other mechanisms of sediment transport. One possibility is the ploughing of ice keels through the sediment (Tulaczyk et al., 2001). This mechanism could clearly increase the sediment flux, but generates values (i.e. <100 m^3 m^{-1} a^{-1}) that are at the lower end of those predicted by a viscously-deforming bed model (Tulaczyk et al., 2001). Thus, more recent work has sought to identify other possible sediment transport mechanisms beneath ice streams. Having observed metres-thick layers of frozen sediment on borehole camera imagery from Kamb Ice Stream (formerly Ice Stream C), Christoffersen et al. (2010) suggested that basal freeze-on is an important mechanism of sediment entrainment when an ice stream shuts-down, and that this sediment can be transported and subsequently released via basal melting when it reactivates. Based on various activation-reactivation scenarios, their estimated sediment fluxes via this mechanism (see Table 1) were more comparable to previous estimates that assumed a viscous-bed model (e.g. Alley et al., 1989a). Numerical modelling by Bougamont and Tulaczyk (2003) was also able to generate relatively high sediment fluxes (Table 1) based on the combined effects of ploughing and the transport of frozen till via plug-flow. Thus, englacial debris transport has potential to be an important component of the relatively high sediment fluxes generated by ice streams (Christoffersen et al., 2010).

High sediment fluxes beneath ice streams might also imply high rates of erosion. Typical steady-state estimates from ice stream catchments averaged over centennial time-scales are generally of the order of 1 mm a^{-1}, but there is one example of 1 m of sediment being removed over a just a few years under Rutford Ice Stream (see Table 2). Precisely how sediment is eroded is another major area of uncertainty. It has been noted that sub-till erosion by a viscous deforming bed might be too low to sustain high sediment fluxes (Cuffey and Alley, 1996). Indeed, in the absence of other mechanism of erosion to maintain a layer of lubricating sediment, this might make some ice streams susceptible to sediment exhaustion and a transition to a higher friction bed. Such a process has been invoked to explain the downstream decrease in bedform density and an increase in exposed bedrock on the M'Clintock palaeo-ice stream in Arctic Canada that led to its eventual shut-down (Clark and Stokes, 2001). A similar transition from a very thick (>10 m) to much thinner deforming layer (<10 m) has also been identified under Pine Island Glacier, West Antarctica (Smith et al., 2013). Given the high erosion rates observed under this ice stream (Smith et al., 2012), Smith et al. (2013) hypothesised that the sediment was being progressively eroded and that, if this process continued, the underlying basement rocks might soon outcrop, potentially increasing basal drag and reducing the ice flow.

Sediment freezing on to the base of the ice has been shown to be a particularly effective mechanism of eroding sediment from the bed and into the basal ice layer (Christoffersen et al., 2010). In addition, it has been proposed that basal ice keels might protrude into the sub-strata and cause erosion to replenish a layer of softer till (Tulaczyk et al., 2001; Clark et al., 2003). It is interesting to note, however, that very few studies have invoked erosion and transport via subglacial meltwater (Table 2). Meltwater has been shown to preferentially pond in the onset zone of active ice streams (e.g. Bindschadler and Choi, 2007; Peters et al., 2007; Schroeder et al., 2013) and there is a wealth of evidence from palaeo-ice stream beds that channelized meltwater flow has incised into bedrock in these locations (e.g. Ó Cofaigh et al., 2012; Graham et al., 2009). However, marine geophysical studies have also drawn attention to the fact that there is no evidence for channelized meltwater flow over the soft till that typically characterises fields of MSGLs further down-stream (e.g. Ó Cofaigh et al., 2002; Dowdeswell et al., 2004; Graham et al., 2009). This has often been interpreted to reflect the fact that subglacial meltwater was transported through the soft till matrix and does little geomorphological work, but Ó Cofaigh (2012) pointed out that this may simply reflect the resolution of geophysical techniques. Indeed, canals and ponded water in soft sediment have been detected beneath several West Antarctica ice streams, including small subglacial lakes (Horgan et al., 2012; see also Section 4.3) and recent modelling suggests that lake drainage events under ice streams might take place in sediment-floored canals (Carter et al., 2017).

Moreover, recent work from Greenland has suggested that suspended sediment in meltwater emerging from an outlet glacier implies erosion rates as high as 4.8 ± 2.6 mm a year (Cowton et al., 2012), which are much higher than had hitherto been assumed (Table 2). Given the emerging evidence of canals and channels beneath some West Antarctic ice streams (King et al., 2004; Schroeder et al., 2013) and that channelized meltwater flow is likely to reach the grounding line of ice streams (Le Brocq et al., 2009; 2013; Dowdeswell and Fugelli, 2012); it seems plausible that subglacial meltwater erosion and transport could play a role in influencing sedimentary processes beneath ice streams and shaping their geomorphology, but one which has been largely overlooked. Indeed, so little work has addressed how meltwater drainage might impact on the geomorphology of ice stream beds that a recent review of the progress in observing and modelling Antarctic subglacial water systems failed to mention the topic (Fricker et al., 2016). In this regard, observations and data from the beds of palaeo-ice streams are a potentially powerful constraint on the configuration of the basal drainage system during lake drainage events. For example, recent work by Livingstone et al. (2016b) has documented evidence of the mechanism and geometry of lake drainage events associated with relict subglacial lake locations on the bed of the Laurentide Ice Sheet. Consistent with work under modern ice streams (See Section 4.3), they showed that the palaeo-lakes were shallow (<10 m) lenses of water perched behind ridges orientated transverse to ice flow and that they periodically drained through canals incised into the substrate. The canals were typically 200-300 m wide, 4-10 m deep and between 700 and 1,700 m. They also reported that these canals sometimes transition into eskers (formed in R-channels) that likely represent the depositional imprint of the last high magnitude drainage event and supports the notion that subglacial meltwater is a potentially powerful agent of erosion under ice streams.

In summary, determining how sediment is eroded and transported beneath ice streams remains a key challenge. Several key mechanisms have been proposed and these hypotheses are summarised in Fig. 17. The next logical step is to attempt to understand which of the processes in Fig. 17 are most important and the extent to which they vary both within and between individual ice streams.These issue have attracted the attention of researchers for several decades, but there is little consensus over the primary mechanisms and this fundamentally limits our ability to model ice stream flow and introduces large uncertainties to predictions of their future behaviour. It is clear that sediment fluxes are high and it is likely that a number of mechanisms contribute to sediment erosion and that it is highly variable in space and time (Fig. 17). Whilst a large number of studies have invoked erosion and transport associated with a mobile deforming bed, it is possible that the accretion (via basal freeze-on) and subsequent release of sediment from the basal ice layer is also very important. It is also likely that meltwater plays an important role, but this has received very little attention and is a key area for future work to address. Resolving these issues will require major improvements in the resolution of geophysical techniques that can image the beds of active ice streams. Repeat surveys of the bed across larger areas than has hitherto been possible could also be prioritised. Ideally, it would be fruitful to compare surveys from various positions down ice streams (onset, trunk and grounding zone). In addition, *in-situ* monitoring of the processes at ice stream grounding lines might also be possible with advances in underwater observing systems. Such data would provide a powerful test for numerical modelling.

5.2. How are Bedforms Created Under Ice Streams?

The erosion, transport and deposition of sediments under ice streams (Fig. 17; Section 5.1) is intimately linked to the creation of subglacial bedforms. Their growth and evolution introduces basal boundary conditions (e.g. roughness) that are different from a featureless bed and which are likely to modulate the velocity of ice streams (Schoof, 2002). The development of such bedforms is also likely to be intimately linked to subglacial water routing and pressures. However, a further key challenge is that there remains little consensus as to how such bedforms are initiated, evolve and perhaps also decay under ice streams.

As noted above (Section 3.7), we now have a very good understanding of what landforms are produced by ice streams (see examples in Fig's 5, 9, 12, 15). The most ubiquitous features on ice stream beds are mega-scale glacial lineations (MSGLs) (King et al., 2009; Spagnolo et al., 2014). Although most workers have tended to focus on (and map) the ridges, this landscape is perhaps best described as a corrugated surface, and it may be misleading to only focus only the positive relief features (see also Section 5.3). Indeed, assemblages of MSGLs appear to resemble the scale and pattern of (albeit much rarer) features seen on the bed of ice streams flowing over a largely bedrock substrate and described as mega-grooves (Bradwell, 2005; Bradwell et al., 2008; Jezek et al., 2011; Krabbendam et al., 2016), although this needs testing quantitatively and there are far fewer studies of hard-bedded landform assemblages. Thus, ice streams flowing of a harder bedrock substrate appear to reorganise their bed in a similar fashion to those flowing over soft-sediments, which may point towards mechanisms that lead to net erosion of the subglacial sediments (see Eyles et al., 2016). It is also clear that ice streams that flow through deep troughs (e.g. Rutford Ice Stream: King et al., 2009) create MSGLs that are near-identical to those on the beds of pure ice streams (e.g. the Dubawnt Lake Ice Stream; Stokes et al., 2003). This suggests that although the geometry and force balance of pure and topographic ice streams might be quite different (e.g. Bentley, 1987; Truffer and Echelmeyer, 2003), their basal processes and geomorphology might have more in common than previously suggested. Indeed, the only obvious difference between the geomorphology of different types of ice streams is that MSGLs on marine-based ice streams are rarely associated with any major meltwater-related features, such as channels, or eskers (Livingstone et al., 2012). In contrast, eskers are commonly reported in association with terrestrially-terminating ice stream beds, but it is often assumed that they were formed during or after ice stream activity (cf. Stokes et al., 2003).

The burgeoning availability of satellite imagery and digital elevation models from both onshore and offshore has enabled workers to collate datasets of several thousand mapped MSGLs and quantitatively characterise their morphometry (e.g. Stokes et al., 2013a; Spagnolo et al., 2014). A recent study by Spagnolo et al. (2014) analysed around 4,000 MSGLs from eight ice stream settings: three offshore (Norway and Antarctica), four onshore (Canada), and one from under Rutford Ice Stream, West Antarctica. This revealed that the typical length of MSGLs is lower than previously suggested (mode 1,000–2,000 m; median 2892 m), and that previous work had perhaps tended to over-emphasise the more extreme values of length and elongation ratio. Indeed, whilst the elongation of MSGLs (mode 6–8; median 12.2) is typically higher than features described as drumlins, Spagnolo et al. (2014) found that these values, and those of their width (mode 100–200 m; median 268 m), overlap; which suggests the two landforms are part of a morphological continuum and may share a similar origin (see also Stokes et al., 2013a; Ely et al., 2016). They were also able to show that MSGLs can initiate and terminate at various locations on an ice stream bed and their exceptional parallel conformity is accompanied by a fairly regular lateral spacing (mode 200–300 m; median 330 m). This, they suggested, provides support to the idea that MSGLs are a spatially self-organized phenomenon (see also Spagnolo et al., in press). They also noted that the size, shape and spatial arrangement of MSGLs are consistent both within, and also between, different ice stream beds, and that this is likely to reflect a common mechanism of formation, which is largely insensitive to local factors.

The wealth of data on MSGL size and shape (e.g. Stokes et al., 2013a; Spagnolo et al., 2014; Ely et al., 2016) provides a useful test of hypotheses of MSGL formation, but there is no universally-accepted mechanism of their formation. There are four main ideas that seek to explain their creation: (i) subglacial deformation of till and attenuation downstream (Clark, 1993), (ii) catastrophic meltwater floods (Shaw et al., 2000, 2008), (iii) 'groove-ploughing' by roughness elements (keels) in the basal ice (Tulaczyk et al., 2001; Clark et al., 2003), and (iv) a rilling instability in the basal hydraulic system (Fowler, 2010). Schoof and Clarke (2008) also developed an idea that subglacial flutes may be formed through a transverse secondary flow in basal that can excavate sediment in a cork-screw like fashion and suggested that it might also be applicable to MSGLs. However, the generation of secondary flows in ice is not straightforward and they noted that the formation of much wider MSGLs would require around 1,000 years, which is inconsistent with rapid evolution of bedforms reported under Rutford Ice Stream (Smith et al., 2007; King et al., 2009). The sediment deformation (i) and meltwater flood (ii) hypotheses are extensions of ideas that have been proposed to explain drumlins and, as such, they appeal to the notion of a subglacial bedform continuum (cf. Stokes et al., 2013a; Ely et al., 2016). In contrast, the groove ploughing (iii) and rilling instability (iv) hypotheses appeal to processes that might act to carve a grooved till surface.

The sediment deformation theory is, arguably, seen as most consistent with observations from both modern and palaeo-ice stream beds (e.g. Hart, 1997; Hindmarsh, 1998b; Smith and Murray, 2009; King et al., 2009; Clark, 2010). Having formally recognised and named MSGLs, Clark (1993) suggested that the extensive literature on other ice-moulded bedforms provided a useful starting point and that the incremental action of ice flow in streamlining MSGL through subglacial deformation/erosion (cf. Boulton, 1987) seemed to be a likely explanation. He argued that if the development of other ice-moulded bedforms, such as drumlins, could initiate by subglacial deformation around inhomogeneities in till, then similar processes might form MSGL, with the difference in scale resulting from increased basal ice velocities. High strain rates, coupled with a plentiful supply of sediment, might lead to subglacial deformation and attenuation of drumlins into much more elongate MSGLs. In support of this hypothesis, MSGLs are commonly seen to develop downstream of drumlins (Ó Cofaigh et al., 2002; Graham et al., 2009; Stokes et al., 2013a), and sometimes formed side by side with drumlins (Stokes et al., 2013). They are also predominantly associated with and composed of a 'deformable' till layer (Dowdeswell et al., 2004; Ó Cofaigh et al., 2005, 2007; Smith and Murray, 2009; King et al., 2009). The preferred spacing of MSGLs (Stokes et al., 2013a; Spagnolo et al., 2014) is perhaps more difficult to reconcile with initiation from pre-existing inhomogeneities, which are more likely to be randomly located, but recent modelling of the coupled flow of ice, subglacial water and sediment suggests that a regular pattern of 'emergent' MSGL (cf. Clark, 2010) could arise from an instability in the deforming bed (Fowler and Chapwanya 2014). Encouragingly, such modelling is able to make tentative predictions of bedform dimensions, which appear to resemble observations of MSGLs, although these comparisons are not straightforward (see discussion in Fowler and Chapwanya, 2014). A potential challenge to the deforming bed hypothesis, however, is that some MSGLs are characterised by 'cores' that consist of crudely stratified glaciofluvial sediments, overlain by till (e.g. Ó Cofaigh et al., 2013). However, these observations may simply indicate that any subglacial deforming bed must have eroded down into pre-existing sediments (Boyce and Eyles, 1991; Stokes et al., 2013b), unless these sediments were laid down during MSGL formation. In addition, a comprehensive study of MSGLs on the bed of a palaeo-ice stream in Poland has shown that their formation likely involves (at least in that location) a process whereby sediment is continuously accreting via a shallow plastically-deforming till layer associated with an inefficient drainage system (Spagnolo et al., 2016).

The meltwater flood hypothesis (Shaw et al., 2008) is largely based on observations of paleo-ice stream geomorphology in Antarctic cross-shelf troughs and invokes catastrophic discharge of turbulent subglacial meltwater. This hypothesis has been applied to drumlin formation (e.g. Shaw, 1983) and large-scale terrestrial flutings (Shaw et al., 2000) and is based largely on form analogy between MSGLs and similar bedforms and patterns created by broad, turbulent flows in water and air. The form analogy is persuasive and Shaw et al. (2008) also point to the abundance of meltwater features and tunnel channels in ice stream onset zones (e.g. crescentric and hairpin scours around the stoss end of drumlins and MSGL), and numerous gullies and channels that often characterise the continental slope. However, there is a general absence of meltwater channel features either upstream or downstream of the MSGLs on most ice stream beds and there is the long-standing issue as to whether the magnitude of meltwater required to form such floods is plausible (e.g. Clarke et al., 2005; Ó Cofaigh et al., 2010b). It is also the case that there have been no observations of significant meltwater floods during the formation of bedforms under Rutford Ice Stream (Smith and Murray, 2009; King et al., 2009), which Shaw and Young (2010) acknowledged would "oblige us to take a long, hard look at the megaflood hypothesis" (p. 199). As noted, however, meltwater drainage events have been observed under some ice streams (see Section 4.3) and a key area for future work is to ascertain the nature and form of these events and the likely impact on sub-ice stream geomorphology (Livingstone et al., 2016b)

The groove-ploughing hypothesis invokes roughness elements in the basal ice (keels) that are able to plough through soft sediments and excavate grooves which leave MSGLs as erosional remnants (Clark et al., 2003). An important assumption with the groove-ploughing mechanism is that grooves can plough through sediment for a sufficient distance (several kms) without thermodynamic and mechanical degradation. Clark et al. (2003) argued that larger keels (e.g. 30 m wavelength, 5 m amplitude) are more likely to survive (e.g. Thorsteinsson and Raymond, 2000) and that survival distances of 10-100 km are plausible, depending on ice velocity (Thorsteinsson and Raymond, 2000; Tulaczyk et al., 2001; Clark et al., 2003). Clark et al. (2003) outlined several predictions of the groove-ploughing hypothesis. An obvious implication is that the transverse roughness of ice stream beds should greatly exceed explicitly measured or quantified roughness, which has been observed in numerous studies (Siegert et al., 2004; Bingham and Siegert, 2007; 2009). A further prediction is that MSGLs should occur in areas down-stream of regions where basal roughness is produced (e.g. downstream of zones of flow convergence and/or more resistant bedrock), and this has also received much observational support (Ó Cofaigh et al., 2002; Graham et al., 2009; Livingstone et al., 2012). Clark et al. (2003) also predicted that the transverse groove spacing should be related to the spatial frequency of roughness that is generated upstream, but MSGLs have been observed to initiate within existing ‘grooves’ (Stokes et al., 2013; Spagnolo et al., 2014); and the finding that MSGLs exhibit a preferred lateral spacing (Stokes et al., 2013a; Spagnolo et al., 2014) is difficult to reconcile with this prediction. Additionally, and where it has been explicitly measured (e.g. Stokes et al., 2013a; Spagnolo et al., 2014), there is little indication that groove width and depth decrease in the down-stream direction, which was also predicted by Clark et al. (2003). Thus, whilst there is some observational support for groove-ploughing and it is intuitively attractive, it does not appear to be the primary mechanism through which MSGLs are formed.

The rilling instability theory was put forward by Fowler (2010) and is based on the theory that a uniform water-film (~2-3 mm deep) flowing between ice and deformable subglacial till would be unstable and rilling would occur, similar to that seen in subaerial settings and which results in a number of narrow streams separated by intervening ridges. Fowler (2010) numerically modelled this process and, for a particular choice of parameters used, he found a preferred spacing (distance to nearest lateral neighbour: 394 m), length scale (52.9 km) and amplitude (12.3 m) which were broadly consistent with observations (Spagnolo et al., 2014). However, these results were strongly dependent on the choice of parameters selected. The appeal of this mechanism is that it can simulate the regularity and spatial organisation of MSGLs (Spagnolo et al., 2014; in press). Moreover, this process has since been incorporated into a more sophisticated treatment that combines both fluvial sediment transport and the coupled flow of ice and sediment to produce a continuum of ribbed moraine, drumlins and MSGLs (Fowler and Chapwanya, 2014).

In summary, there is no consensus as to how MSGLs are generated beneath ice streams and how the processes summarised in Fig. 17 might contribute to their formation. Given the available observations of both the geomorphology and composition of MSGLs, it seems likely that their formation primarily involves a mobile layer of deforming till (albeit of uncertain thickness) that is transported/attenuated downstream (cf. King et al., 2009) and which likely involves net erosion of the subglacial sediments over long (> centennial) time-scales. A further appeal of this as a primary mechanism is that it may also explain the broader continuum of landforms found on ice stream beds, including drumlins and ribbed moraines, and the recently inferred 'traction ribs' (Section 4.4). However, given that the pervasive and viscous deformation of several metres of till has largely fallen out of favour in recent years, it is likely that localised erosion from both ice keel ploughing and meltwater might also take place with grooves, in addition to basal freeze-on and subsequent melt-out. These mechanisms might also help to reconcile the high rates of sediment erosion and transport beneath ice streams (see Tables 1 and 2), but their relative roles remains largely unknown. In terms of other bedforms, ice stream shear margin moraines have received minimal attention in the literature (Stokes and Clark, 2002; Hindmarsh and Stokes, 2008; Batchelor and Dowdeswell, 2016), and their origin is even more enigmatic, despite their potential role in helping to stabilise lateral shear margins and the width of ice streams. They represent a key area for future work to address, in addition to a more quantitative comparison between soft-bedded MSGLs and their qualitatively similar bedrock mega-grooves.

5.3. How Can We Better Quantify Subglacial Roughness Beneath Ice Streams?

Studies of ice stream geomorphology (e.g. Stokes et al., 2013a) have perhaps been preoccupied with mapping landforms and quantifying the size and shape of individual features. This is probably because the early work on palaeo-ice streams was simply trying to identify their location and incorporate them into a palaeo-glaciological reconstruction (see Section 3.1). It soon became clear that extracting 'metrics' from these mapped features could lead to new insights with regard to the subglacial landscape and the formation of subglacial bedforms (see previous section), but there are other means of quantifying ice stream geomorphology that have perhaps been under-utilised, particularly with respect to subglacial roughness and how flow-sets of bedforms evolve through time.

Bingham et al. (2010) defined bed roughness as the vertical variation of an ice-sheet bed with horizontal distance and noted that it can be quantified in various ways. Radar-derived ice bed topography has been used most widely to investigate the roughness of ice stream beds and one approach has been to simply characterize roughness based on the standard deviations of individual bed elevation points from an interpolated bed surface (Rippin et al., 2006). More sophisticated techniques involve assessing bed roughness based on the power spectra derived from Fast-

Fourier-transforming bed-elevation profiles (Taylor et al., 2004), which has been used to quantify bed roughness across several West Antarctic ice streams (Siegert et al., 2004; Bingham and Siegert, 2007a; Bingham and Siegert, 2009). Taken together, these have clearly shown that the bed roughness beneath ice streams is much smoother than beneath non-streaming ice and that roughness also tends to decrease down ice streams.

Bingham and Siegert (2009) usefully outlined a framework for the geomorphological interpretation of bed roughness in Antarctica, suggesting that the smoother beds of ice streams likely result from pre-existing marine sediments, together with warm-based ice flowing at high velocities, which promotes subglacial erosion and deposition. They suggested that it would be worthwhile to apply parallel methods to former ice sheet beds in order to develop a methodological framework that captures the signature of roughness in formerly glacial landscapes, and this would seem an obvious priority for future work. That is, methods could be developed to quantify the roughness of various assemblages of landforms on palaeo-ice stream beds. If it can be shown that different landform assemblages have different roughness signatures, then this opens up huge potential to use measurements of roughness on modern ice streams to infer distinct geomorphologies that may lie at their bed. To that end, Fourier analysis was recently used to analyse MSGLs from a number of different ice stream beds, which is one of the first attempts to quantify and compare MSGL topographhies, but without the need for mapping individual bedforms (Spagnolo et al., in press). The results from that study concluded that assemblages of MSGLs were very similar across different settings, perhaps reflecting pattern evolution via downstream wavelength coarsening, and consistent with the instability theory for subglacial bedforms (e.g. Fowler et al., 2014).

It would also be useful to quantify how roughness changes along a flow-line from the slow-flowing ice (close to the ice divide), through ice stream onset zones, and into the main trunk, and see how this compares to ice velocities. Is there, for example, a threshold roughness that dictates the location of ice stream onset? Does the roughness on a soft-bedded ice stream differ from a hard-bedded ice stream? Can soft-bedded ice streams transition into hard-bedded ice streams through a process of sediment exhaustion? Another key area of investigation would be to measure and quantify how values of roughness change through time. A fundamental and yet largely unanswered question is: do ice stream beds get rougher or smoother and over what time-scales (or can they do both, depending on underlying geology and setting)? This could be measured on active ice streams over short (annual to decadal time-scales), but it might also be possible to extract measurements of roughness from palaeo-ice stream beds that are well-dated in terms of their duration. The quantification of ‘mature’ versus ‘younger’ ice stream geomorphology would provide a useful framework for interpreting the age of ice stream beds in both modern and palaeo-ice sheets. Moreover, if it can be shown that roughness evolves through time in a predictable manner, this would help parameterise the evolution of basal shear stress in numerical ice sheet models that aim to predict future ice sheet dynamics.

Related to the issue of roughness is the recent analysis of the size-frequency distribution of subglacial bedforms in flow-sets. Large datasets of bedform metrics have revealed that drumlins and MSGLs display smooth unimodal frequency distributions with a positive skew (Clark et al., 2009; Stokes et al., 2013a; Spagnolo et al., 2014). When their length, amplitude or spacing are converted to their natural logarithm, the resulting size-frequency distribution is well-defined as log-normal (Fowler, 2013; Hillier et al., 2013; Spagnolo et al., 2014). This distribution is extremely common in nature and is thought to emerge from a large number of independent events in which incremental growth (or decay) can occur (Fowler et al., 2013; Spagnolo et al., 2014). Thus, several studies have argued that the observed log-normal distributions of MSGLs (Spagnolo et al., 2014) implies an element of randomness in their development, such that growth phases occur randomly and for random duration, likely alongside episodes of decay, e.g. perhaps associated with episodes of bedform erosion and deposition (Fowler et al., 2013; Spagnolo et al., 2014; Hillier et al., 2013, 2016). As noted by Hillier et al. (2013), this is consistent with the geophysical studies that have revealed variable bed conditions in both space and time (Smith, 1997a, b; Murray et al., 2008) and subglacial landforms (King et al., 2007; Smith and Murray, 2009) that evolve rapidly on sub-decadal timescales (Smith et al., 2007; King et al., 2009). Hillier et al. (2013) also noted, however, that it is unclear whether this variability arises from the dynamics of ice–sediment–water interactions (e.g. basal stick–slip events) or from interactions between bedforms. Notwithstanding this uncertainty, this stochastic approach contrasts with a deterministic view whereby proto-bedforms of known size and shape always evolve similarly with time to a predictable final morphology. A key challenge, therefore, is to try and incorporate these elements of stochasticity into models of bedform genesis (see next section).

Similar to the analyses of roughness outlined above, Hillier et al. (2013) also noted that it would be useful to ascertain the stability or otherwise of bedform populations through time, i.e. are they in steady state? Using observations of flow-sets beneath modern ice streams, it might also be possible to create statistical models that link the physical processes to observable characteristics of bedform populations under ice streams (see Hillier et al., 2016).

5.4. Can We Numerically Model Ice Stream Geomorphology?

As noted in Section 5.2, there are several competing hypotheses that seek to explain ice stream geomorphology, almost exclusively focussing on the formation of MSGLs. Unfortunately, only very rarely have these ideas developed to the stage where they can be captured in physically-based numerical models of the ice-bed interface, and this is a further key challenge for future work to address. Of the various hypotheses relating to ice stream geomorphology, only the deforming bed/till instability theory has seen some success in being able to produce a recognisable geomorphology (Hindmarsh, 1998a, b; Schoof 2007; Chapwanya et al., 2011; Fowler and Chapwanya, 2014). The advantage of these models being able to predict ice stream geomorphology is that they can be tested against large datasets of landform metrics that now exist (e.g. Spagnolo et al., 2014), although this has proved difficult, except for ribbed moraine (Dunlop et al., 2008; Chapwanya et al., 2011). The ultimate aim would be for a model to be able to simulate the formation of landform assemblages that resemble those on ice stream beds, much like Dunlop et al. (2008) were able to simulate modelled ribbed moraines that matched observations.

Recently, Barchyn et al. (2016) took a reduced complexity approach in order to try and simulate recognisable bedforms using simple mechanisms based on shallow sediment dynamics that are hypothesised to exist subglacially (e.g. Fig. 17). Although their focus was on subglacial bedforms in general, rather than specifically on ice stream geomorphology, they found that bedforms readily emerge, see Fig. 18, and their interactions mirrored those in aeolian and fluvial geomorphology. Despite its reductionist approach, they were also able to replicate a bedform continuum, whereby transitions between ribbed moraines and elongate flow-parallel bed forms were associated with increasing ice speeds and decreasing sediment thickness. In their model, drumlins transition into MSGLs because the lower ice-to-bed pressure and high ice velocities force bedforms to elongate by extending cavities and low-pressure areas on the lee-side of bedforms. This model was also able to predict till fluxes (of the order of 10 m^3 m^{-1} a^{-1} for ice flowing at 150 m a-1), which they noted was similar but perhaps an underestimate compared with values elsewhere. As in most previous modelling efforts, the aim was to try and replicate the characteristic dimensions of subglacial bedforms (lengths, widths, heights) and sediment properties and processes were necessarily simplified. Clearly, as these modelling experiments evolve, they are likely to benefit from more sophisticated paramaterisations based on known sediment properties of the ice stream bedforms.

The value of these modelling approaches is that observable characteristics of bedforms (their geomorphology) can then be securely linked to processes and the properties of the overlying ice (thickness, velocity, etc.). Not only would this help inversions of ice sheet dynamics from palaeo-ice stream geomorphology, but it would also allow forward runs to observe the time-scales over which the geomorphology evolves and influences basal shear stresses. Such knowledge is required to reduce the uncertainties in numerical models of future ice sheet/ice stream dynamics, where the parameterisation and evolution of basal shear stresses is a key unknown (e.g. Ritz et al., 2015).

In developing the next generation of numerical models of sub-ice stream processes, it is useful to consider what key aspects of ice-bed interface (Fig. 17) should be targeted by a successful model. Based on this review of the literature, it seems that the following, at least in part, are salient features of the subglacial environment of ice streams:

- The efficient transport of sediment down-stream to the grounding line to account for high sediment fluxes and the building of grounding zone wedges (essential)
- The creation of highly elongate subglacial bedforms (MSGLs) that elongate with ice velocity (essential) and show transitions from drumlins to MSGLs (desirable)

- A mechanism (or mechanisms) to erode and replenish subglacial sediment (desirable)
- A mechanism (or mechanisms) that generates spatial and temporal variability in basal shear stress (desirable)

6. Conclusions:

Rapidly-flowing ice streams are an important component of ice sheet mass balance and associated impacts on sea level. However, it was only in the 1970s that their importance was fully-recognised and the study of ice streams began in earnest. It soon became clear that their flow was governed by processes at the ice bed interface that was characterised by a unique geomorphology. Elucidating how this geomorphology is created and evolves is critical to understanding and modelling ice stream flow, and our knowledge has grown rapidly over the last three decades, from almost complete ignorance to a detailed knowledge of the morphology of ice stream beds. This has been brought about through (i) geophysical investigations of active ice streams, mostly in the West Antarctica Ice Sheet, and through (ii) the investigation of sediments and landforms left behind by ice streams in areas formerly occupied by ice sheets. This paper has reviewed progress in these two main areas from an historical perspective in order to identify key areas of progress in both glaciology and glacial geomorphology and how they have been brought about. Emphasis has been placed on the extent to which these sub-disciplines have converged to help understand ice stream geomorphology, together with the key challenges that remain and how they might be overcome.

From this review, it is clear that we now have a very detailed knowledge of the morphology of ice stream beds and an emerging understanding of how this relates to the distribution of sediments, water and basal shear-stresses. However, our understanding of how the morphology (form) relates to processes is much more limited, and I argue that this represents the next major challenge and will require even closer cooperation between glaciologists, glacial geomorphologists, and sedimentologists, together with numerical modellers. Knowledge from both palaeo and modern ice stream studies has allowed for improved estimates of the rates of subglacial erosion and transport (Tables 1 and 2), but the importance of various mechanisms of sediment erosion and transport (Fig. 17) is largely unknown, and processes involving subglacial meltwater are potentially important, but have perhaps been overlooked in terms of shaping sub-ice stream geomorphology. Furthermore, sediment erosion, transport and deposition are likely to be intimately linked to the growth and decay of subglacial bedforms, which modulate the velocity of ice streams through their influence on basal roughness and subglacial hydrology. Unfortunately, only very rarely have formational hypotheses (e.g. of MSGLs) been developed to the stage where they can be captured in physically-based numerical models of the ice-bed interface, and this is a further key challenge for future work to address. Such modelling is likely to benefit from the integration of the known properties of subglacial sediments, which are readily accessible on palaeo-ice stream beds but have perhaps been under-utilised.

The way in which ice stream geomorphology influences or even controls ice stream flow over a range of time-scales is also poorly understood, and yet such knowledge is required for numerical modeling that aims to predict future ice stream dynamics and their potential contribution to sea level. Studies of ice stream geomorphology have perhaps been preoccupied with mapping landforms and quantifying the size and shape of individual features (object-based morphometry). Quantification of these ‘metrics’ has led to some important insights with regard to the formation of MSGLs, but other means of quantifying ice stream geomorphology have perhaps been under-utilised, particularly with respect to how topography (or roughness) can be quantified without the need to ‘map’ individual features (general geomorphometry). In addition, and with repeat surveys of ice stream beds now being undertaken (King et al., 2017), it should be possible to examine how fields (or flow-sets) of bedforms evolve through time, which could be explored by statistical analysis of size-frequency distributions. These methods of quantifying sub-ice stream geomorphology also offer a tangible means to help parameterise the evolution of basal shear stress in numerical ice sheet models and there is room for much closer collaboration with those modelling bedforms and basal shear stresses in aeolian and fluvial environments. Given the increasing resolution of digital elevation models from both palaeo and modern ice stream settings, there are several key questions that should be tractable within a few years: is there a threshold roughness that dictates the onset of ice stream flow? Do ice stream beds get rougher or smoother and over what time-scales? To what extent is the geomorphology on hard and soft-bedded ice streams different, and how might this influence their flow?

In summarising his thoughts on the Cordilleran Ice Sheet Symposium in 1990 and noting the “currently fashionable term ‘ice stream’”, Mathews (1991: p. 265) stated that “with so little known about the conditions and processes operating at the bed of contemporary ice streams, it seems doubtful than the site of an ancient ice stream can be identified solely from a track engraved on the substratum. It will remain a challenge until more is learned about the subglacial topography of the Greenland and Antarctic ice sheets and the erosional processes under both slow and fast-moving ice”. It is pleasing to see that three decades of collaboration between both glaciologists and glacial geomorphologists has overcome this challenge and we now have a detailed knowledge of the glacial geomorphology engraved by ice streams. However, there remains a challenge to learn more about the processes that create ice stream geomorphology and how they impact on ice dynamics. The next three decades are likely to see further growth in the imaging of geomorphology under active ice streams, which should help meet this challenge and provide the necessary constraints for numerical models that are urgently required in the context of ice sheet stability in a warming climate.

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Table 1: Examples of Published Estimates of Ice Stream Sediment Fluxes

Ice stream (ice sheet)	**Sediment flux per meter width of ice stream terminus ($m^3\ m^{-1}\ a^{-1}$)**	**Principal mechanism**	**Reference**
Modern ice streams:			
Ice Stream B (WAIS)	<0.1	**Meltwater** drainage system	Alley et al. (1989a)
Ice Stream B (WAIS)	< 88	**Ploughing**	Tulaczyk et al. (2001)
Ice Stream B (WAIS)	~100	**Deforming layer**	Anandakrishnan et al. (2007)
Ice Stream B (WAIS)	150	**Deforming layer**	Kamb (2001) and Engelhardt and Kamb (1998), cited in Anadakrishnan et al. (2007)
Ice Stream C (WAIS)	~230	**Ploughing**	Bougamont and Tulaczyk (2003)
Ice Stream C (WAIS)	525 - 875	**Freeze-on** and **melt out**	Christoffersen et al. (2010)
Ice Stream B (WAIS)	100-1000	**Deforming layer**	Alley et al. (1989a)
Rutford Ice Stream (WAIS)	4000	**Deforming layer**	Doake et al. (2001)
Palaeo-ice streams:			
Lake Michigan Lobe (LIS)	~100	**Deforming layer**	Jenson et al. (1995)
Marguerite Bay palaeo-ice stream (WAIS)	~100 - 800	**Deforming layer**	Dowdeswell et al. (2004)
Isfjorden (EIS)	560-980	**Deforming layer**	Hooke and Elverhøi (1996)
Marguerite Bay palaeo-ice stream (WAIS)	>1,000	**Deforming layer**, plus melwater and/or freeze-on and melt out	Livingstone et al. (2016a)
Norwegian Channel Ice Stream (EIS)	8,000	**Deforming layer**	Nygård et al. (2007)
Hudson Strait Ice Stream (LIS)	800 - 37,300	Sediment entrained in **basal ice layer** (related to Heinrich events)	Dowdeswell et al. (1995)
M'Clintock Channel Ice Stream (LIS)	73,000	**Deforming layer**	Clark and Stokes (2001)

Table 2: Examples of Published Estimates of Ice Stream Erosion Rates

Ice stream (ice sheet)	Erosion rate (mm a^{-1})	Primary mechanism and context	Reference
Ice Stream C (WAIS)	~0.04 – 0.1	**Basal freeze-on**: estimate of the accretion of sediment into a basal ice layer during cycles of basal freeze-on	Christoffersen et al., (2010)
Ice Stream B (WAIS)	0.1	**Deforming layer**: long term catchment average	Anandakrishnan et al. (2007)
Ice Stream B (WAIS)	0.1 - 0.4	**Deforming layer**: Steady state erosion over catchment	Alley et al. (1987b)
Ice Stream B (WAIS)	~0.5	**Deforming layer**: Steady state erosion over catchment	Alley et al. (1989a)
Ice Stream C (WAIS)	~0.2 – 06	**Basal freeze-on** and **ploughing**: numerical modelling of tributaries and trunk	Bougamont and Tulaczyk. (2003)
Ice Stream B (WAIS)	<0.73	**Ploughing**: average rate for a 120 km flow-line upstream of UpB Camp	Tulaczyk et al. (2001)
Leverett Glacier (Greenland)	4.8	**Meltwater erosion:** average rate of erosion from the catchment	Cowton et al. (2012)
Pine Island Glacier (WAIS)	600	**Deforming layer**: average rate of erosion estimated from ice surface/thickness measurements at one location, averaged over 49 years	Smith et al. (2012)
Rutford Ice Stream (WAIS)	1,000	**Deforming layer**: localised erosion across 500 m wide areas associated with drumlin formation	Smith et al. (2007)

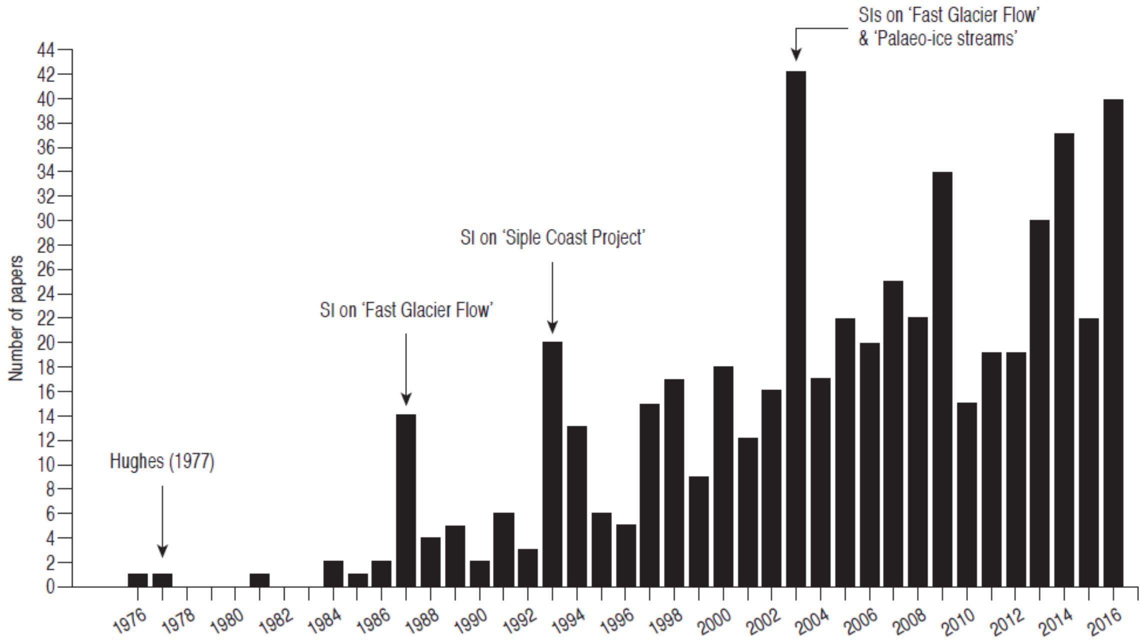


Figure 1: The number of papers with "ice stream" or "ice streams" in the title from the ISI Web of Science catalogue (n = 538; search date: 14 March 2017). The paper from 1976 is a conference abstract from the Midwestern AGU meeting (Bentley, 1976) and so the first major paper was by Hughes (1977). A slow growth in outputs covering ice streams only begins in the mid-1980s, following the pioneering geophysical studies aimed at characterising their subglacial environment (e.g. Blankenship *et al*., 1986; Alley *et al*., 1986). Indeed, the anomalous peak in 1987 is due to a special issue (SI) of the *Journal of Geophysical Research* that contained papers from the Chapman Conference on 'Fast Glacier Flow: Ice Streams, Surging and Tidewater Glaciers' (see Clarke, 1987). A further notable peak results from a collection of the 'Siple Coast Project' papers that were published together in *Journal of Glaciology* in 1993 (vol. 133). The most productive year remains 2003, which coincides with the production of two thematic issues related to ice streams: the 'International Symposium on Fast Glacier Flow' in the *Annals of Glaciology* and the 'Palaeo-Ice Stream International Symposium' in *Boreas* in 2003. Note that the focus here is on the emergence of the term "ice stream" in scientific papers and that some studies of ice streams/outlet glaciers will be missing (e.g. those on Pine Island and Thwaites Glacier, W. Antarctica; Lambert Glacier in East Antarctica; and Jakobshavns Isbræ, in W. Greenland), in addition to some edited book chapters.

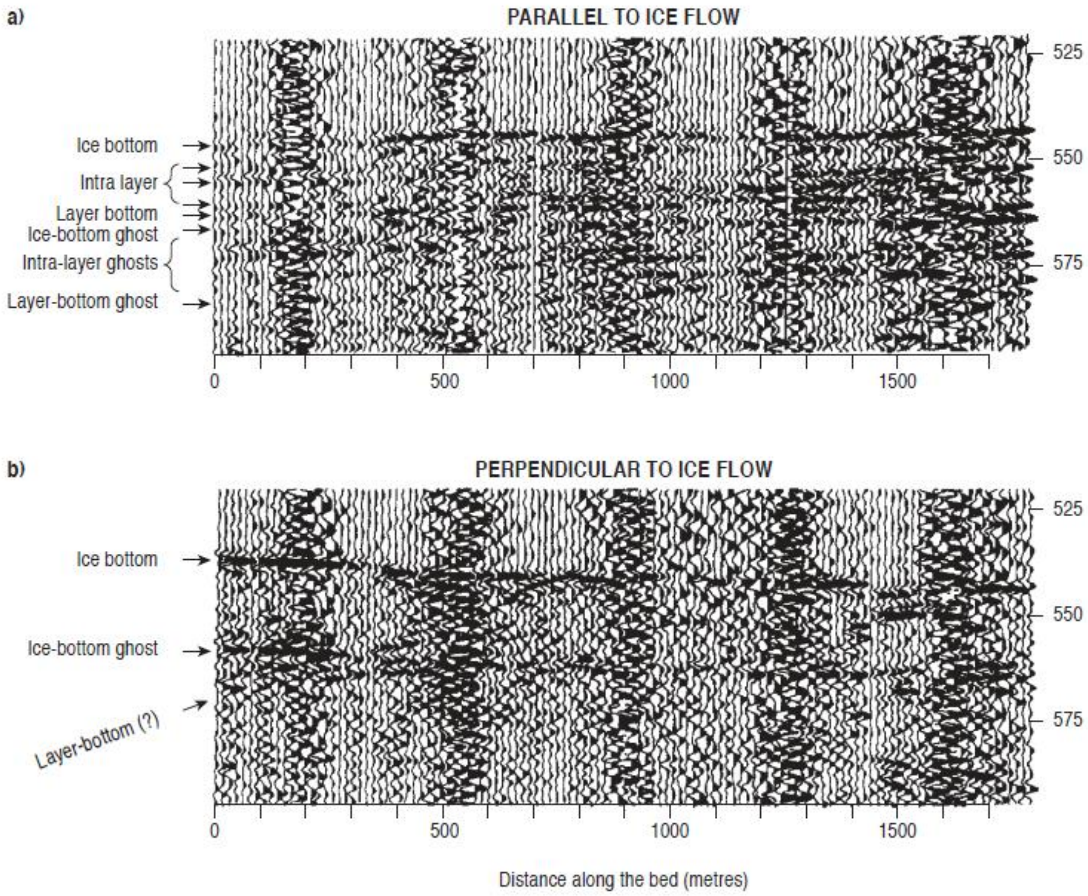


Figure 2: One of the first glimpses of the geomorphology of an active ice stream bed, from the UpB site on Ice Stream B, West Antarctica (from Blankenship et al., 1986). Blankenship et al. (1986) noted that flat, parallel reflectors in the along-flow direction (a) contrasted with those observed in the across-flow direction (b). Note the 'hump' in one of the reflector just beneath the ice bottom (i.e. buried within the till layer) around 800 m along the profile, labelled 'layer bottom (?)' on panel (b). Reproduced by permission of Nature Publishing Group.

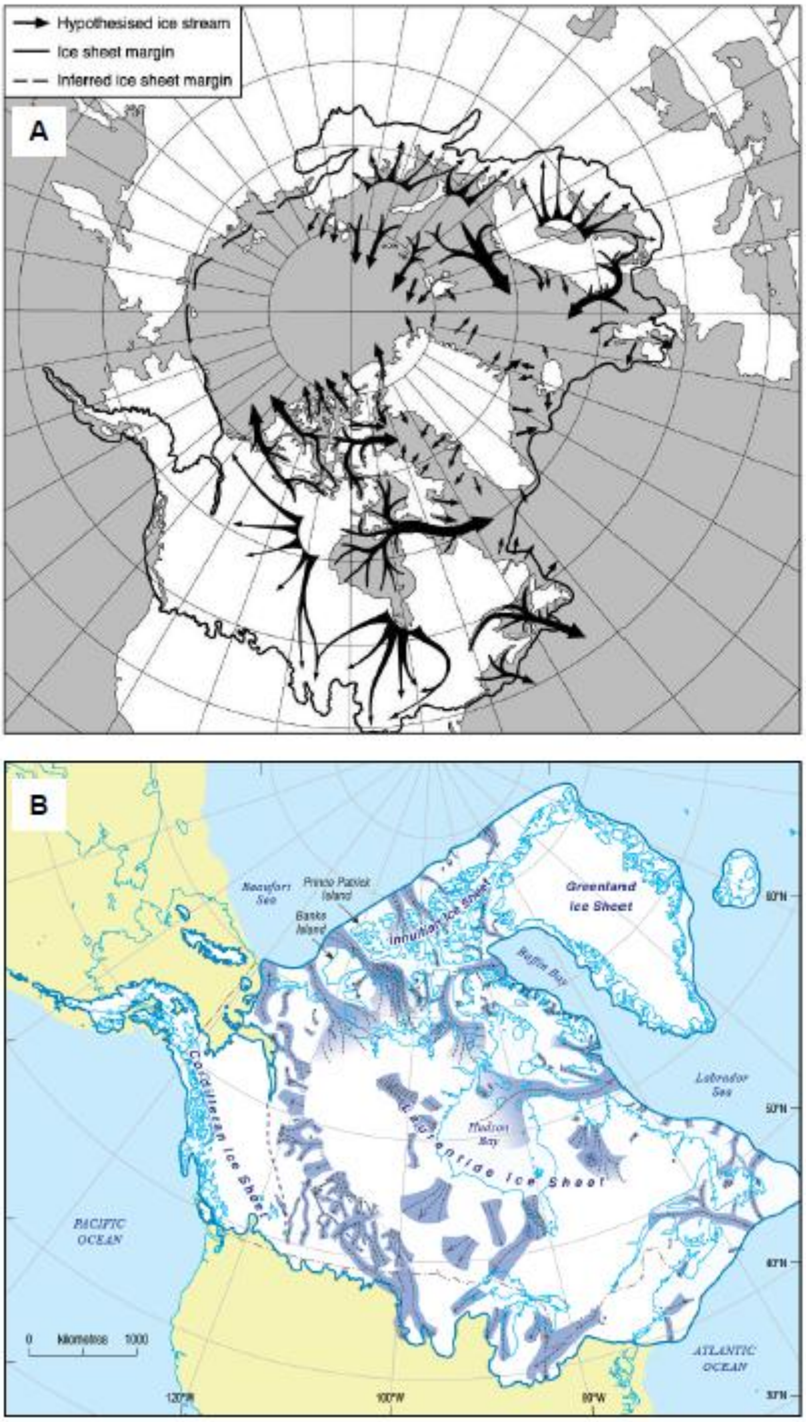


Figure 3: (a) Denton and Hughes (1981) hypothesised location of ice streams in the northern hemisphere ice sheets (from Stokes and Clark, 2001). Although largely based on topographic inference and some notion of spatial self-organisation, many of these ice stream locations have since been confirmed in recent inventories derived from detailed mapping of ice stream beds, such as the Laurentide Ice Sheet (e.g. Margold et al., 2015a), which is shown in (**b**), and taken from Stokes (2017).

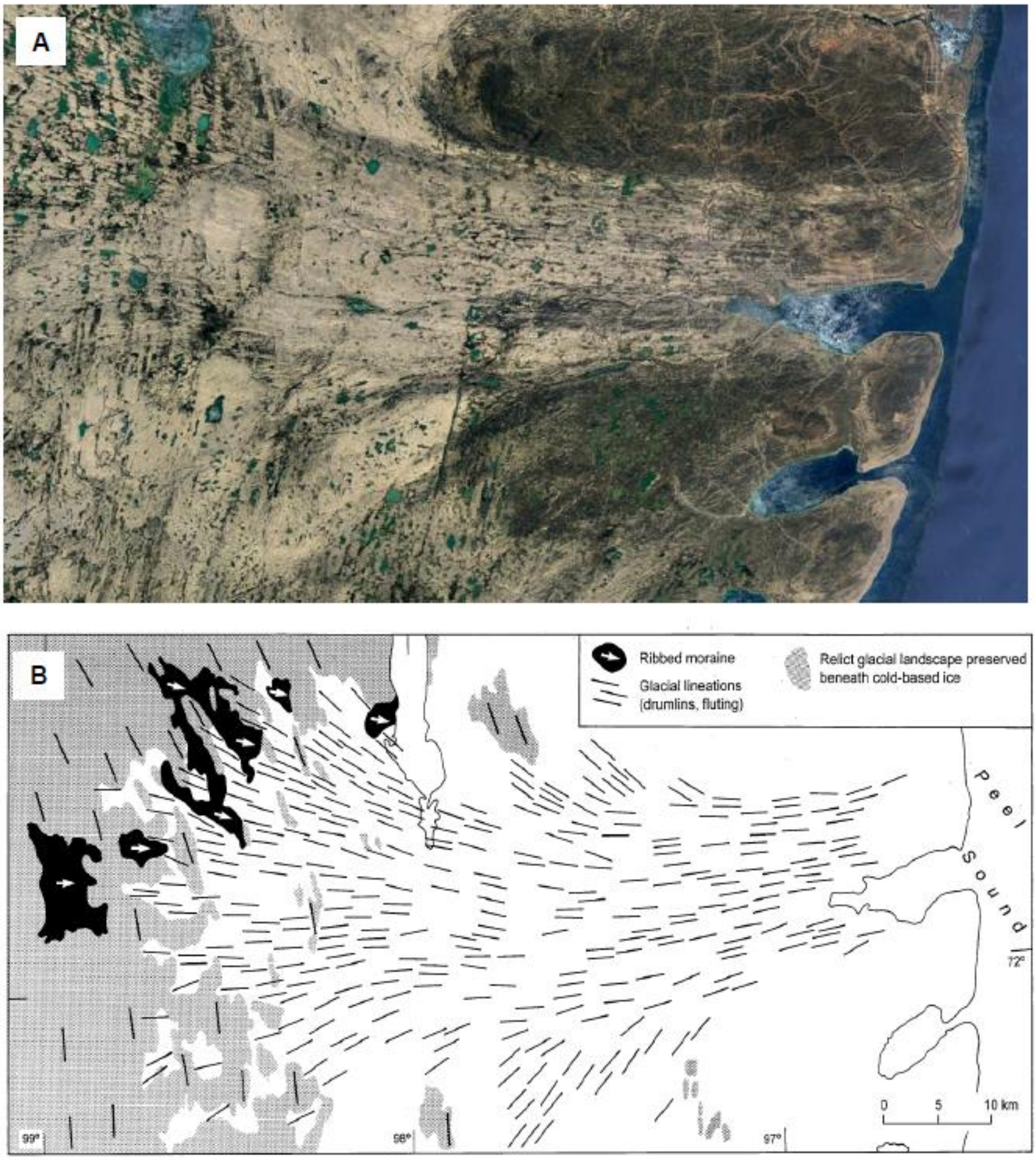


Figure 4: (a) Landsat image (data from Google Earth: Landsat / Copernicus) of a Boothia-type erratic dispersal train on Prince of Wales Island, Canadian Arctic. Dyke and Morris (1988) and Dyke et al. (1992) described this classic example and noted how the lateral margins of the dispersal train were very abrupt and coincided with a convergent pattern of drumlins feeding into the plume. (**b**) Simplified map of the glacial geomorphology from Hättestrand and Kleman (1999), which they redrew from Dyke et al. (1992). Reproduced by permission of Elsevier.

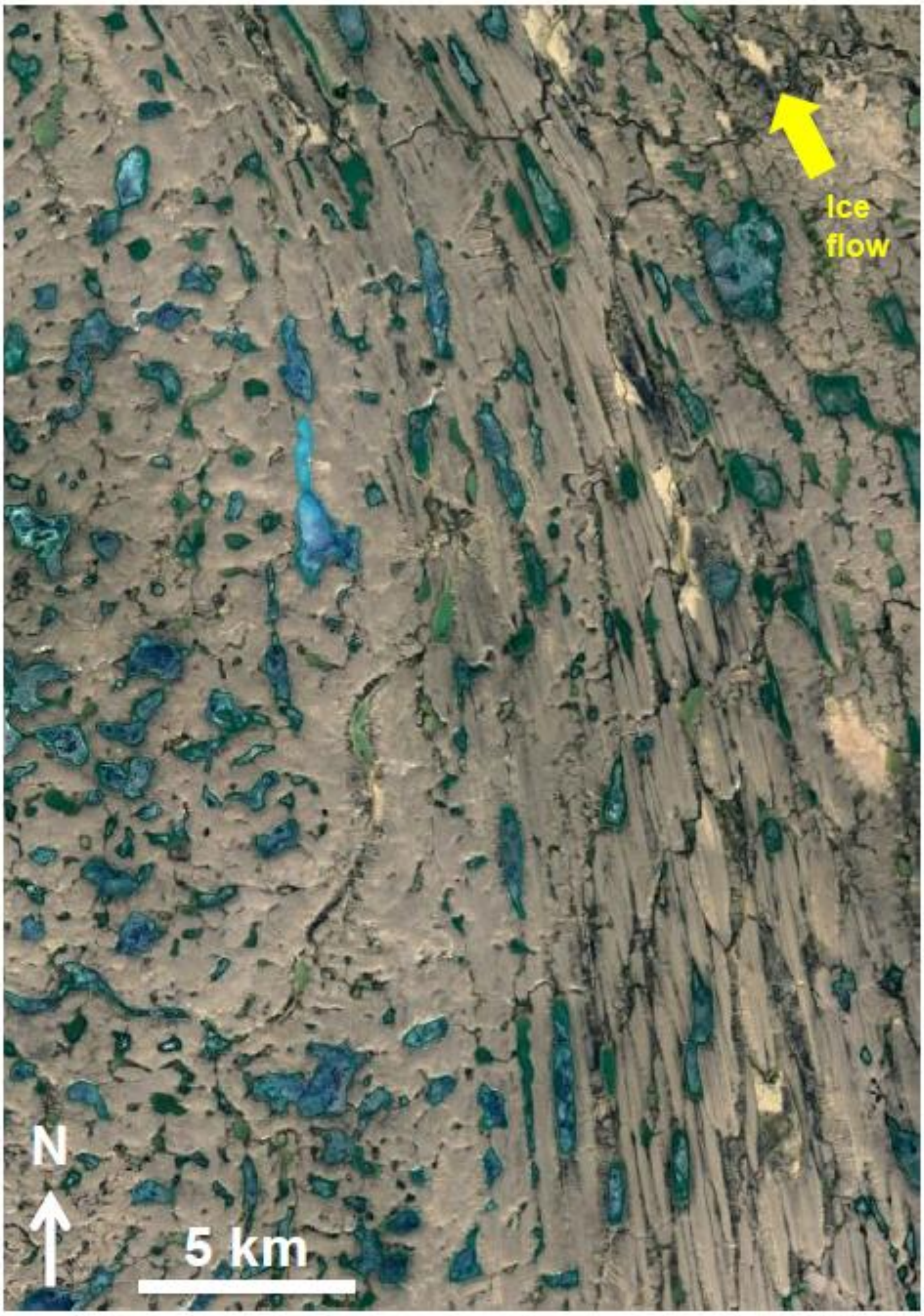


Figure 5: Landsat image (data from Google Earth: Landsat / Copernicus) of the abrupt lateral shear margin of the M'Clintock Channel Ice Stream on Victoria Island, Canadian Arctic, first identified by Hodgson (1994).

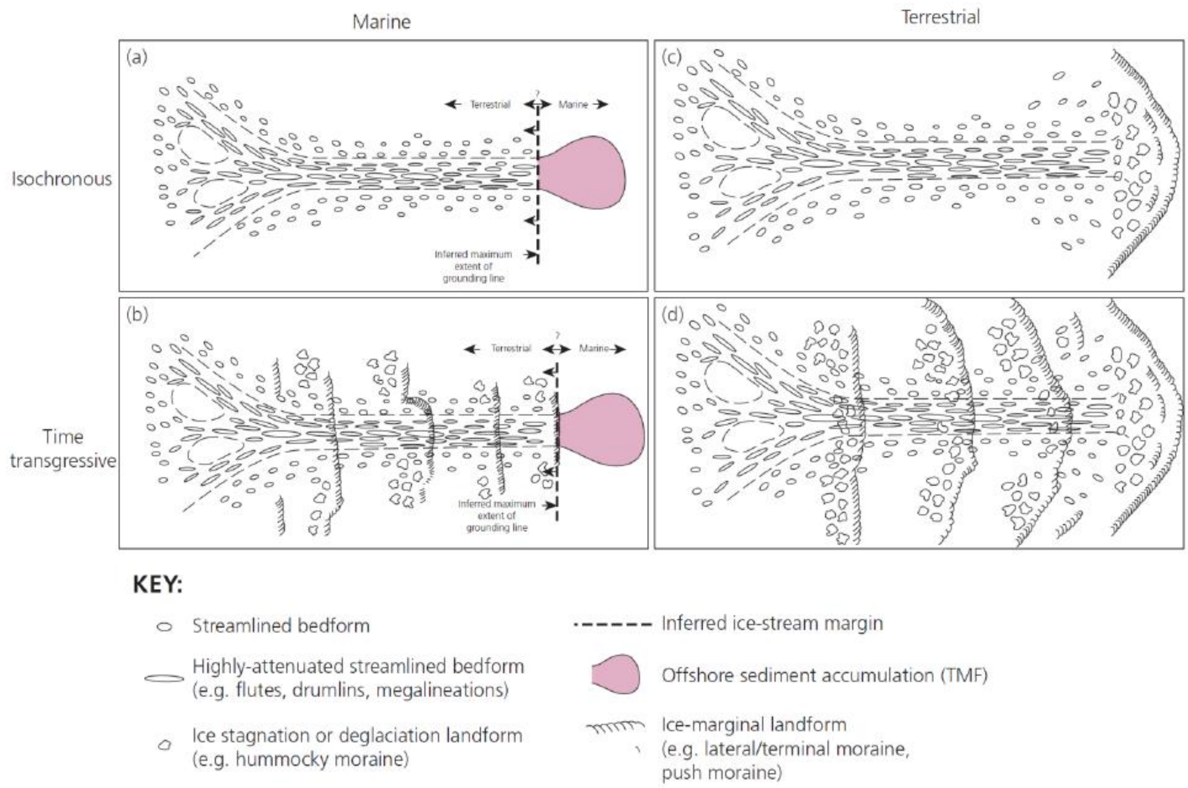


Figure 6: Idealised landsystems models of the geomorphology of palaeo-ice stream tracks from Stokes and Clark (1999) depending on whether the ice stream was marine-terminating (**a** and **b**) or terrestrially-terminating (**c** and **d**) and whether the imprint was formed near-synchronously prior to ice stream shut-down (**a** and **c**) or time-transgressively during ice stream retreat (**b** and **d**). These imprints were designed to help workers identify palaeo-ice stream tracks on the beds of palaeo-ice sheets. Reproduced from Benn and Evans (2010) with permission from Hodder Education, London, UK.

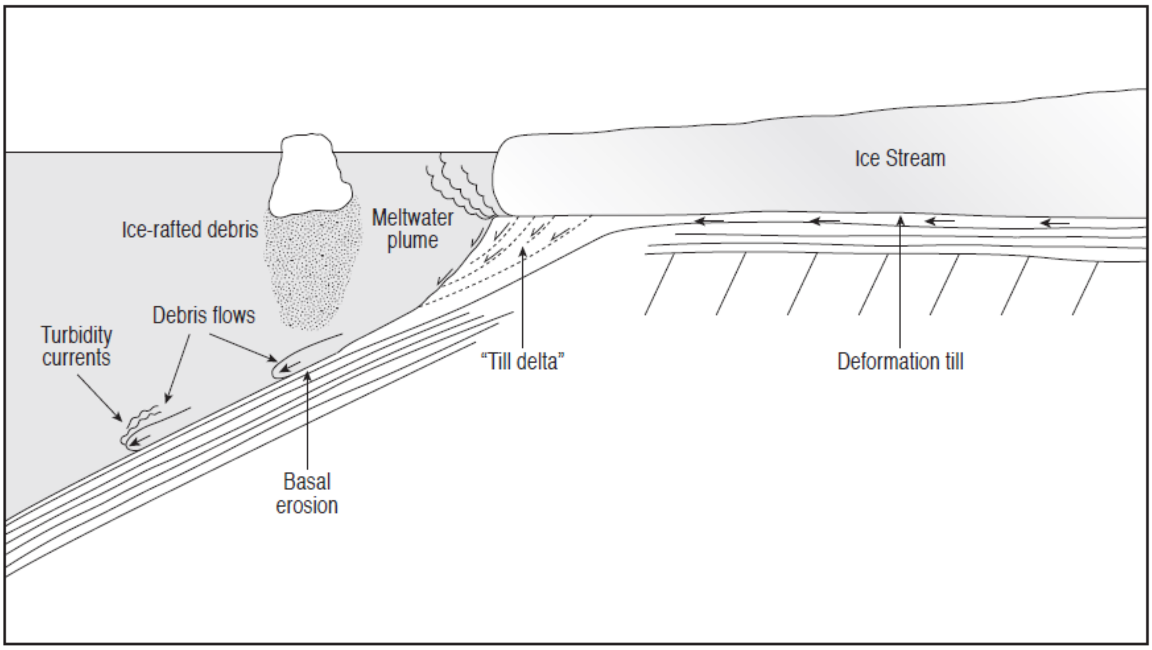


Figure 7: Schematic model showing sedimentary processes and the development of a till delta at grounding line of a marine-terminating ice stream at the edge of the continental shelf (e.g. during the Last Glacial Maximum) (redrawn from Laberg and Vorren, 1995)

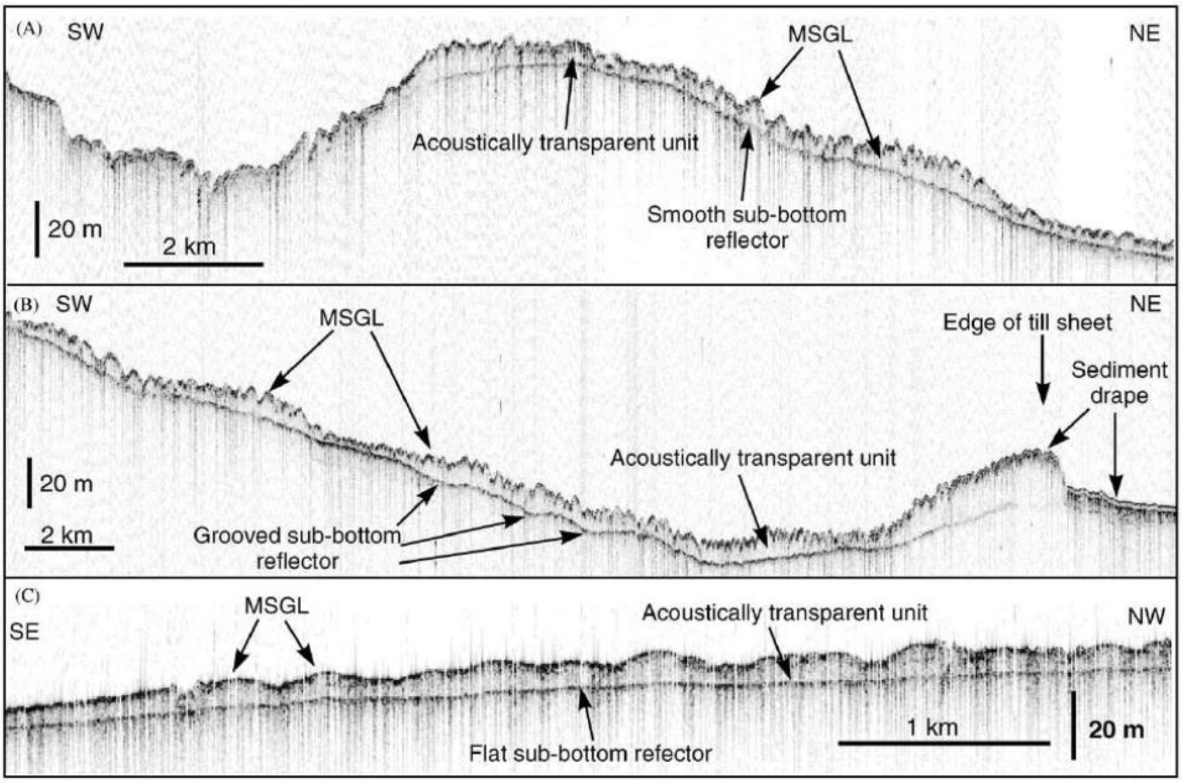


Figure 8: TOPAS sub-bottom profiler records from outer Marguerite Trough palaeo-ice stream, Antarctic Peninsula (from Ó Cofaigh et al., 2005) showing mega-scale glacial lineations formed in an acoustically-transparent, low shear strength till averaging around 5-10 m thick. Reproduced by permission of Elsevier.

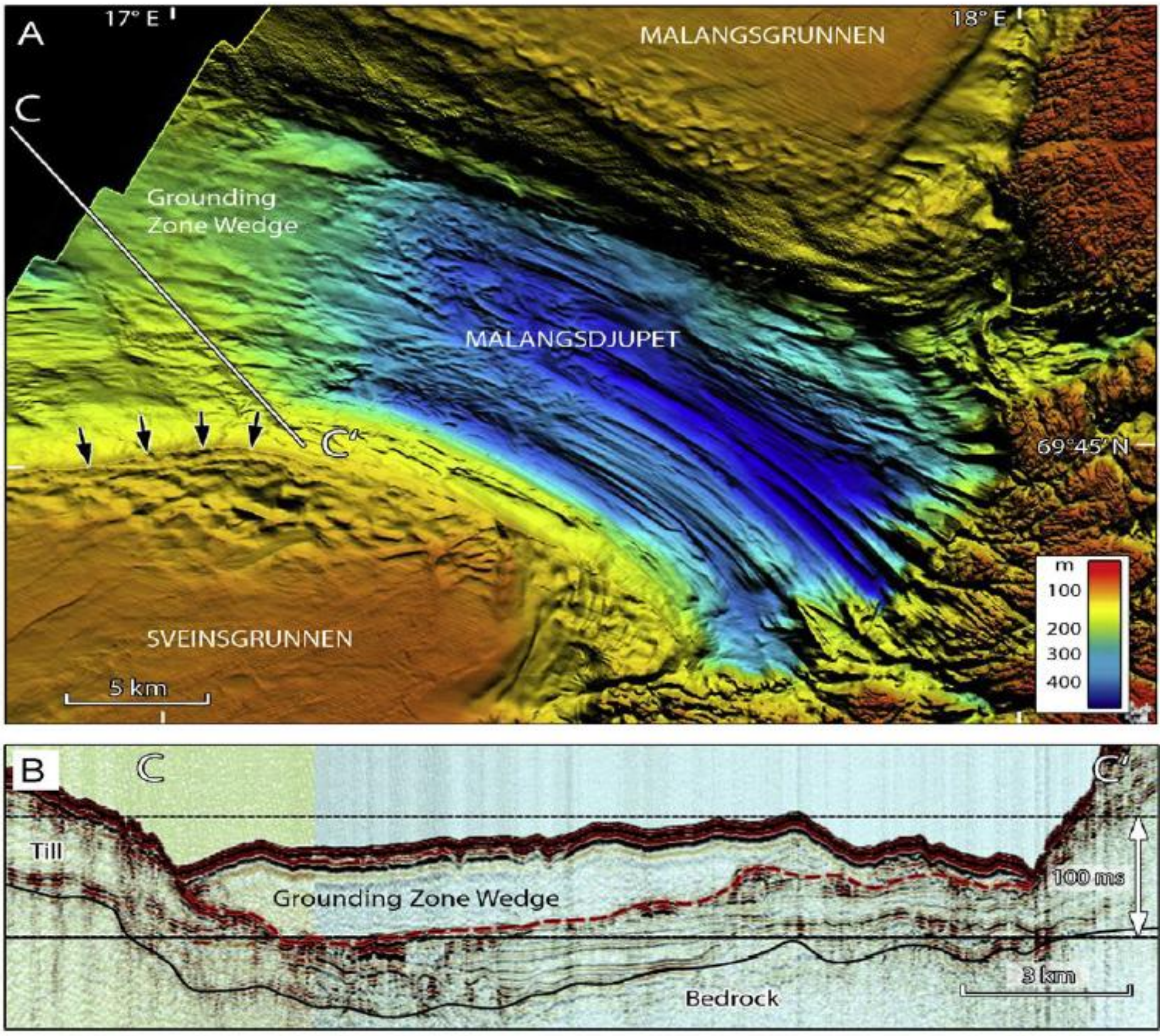


Figure 9: (**a**) Swath bathymetry of the Malangsdjupet palaeo-ice stream bed in northern Norway (from Ottesen et al., 2008) showing the typical geomorphology of a marine-terminating ice stream with rough bedrock in the onset zone merging into drumlins and mega-scale glacial lineations in the deepest part of the cross-shelf trough. Also not the location of a grounding zone wedge (cross section from seismic data shown in (**b**)) and lateral shear margin moraines (black arrows). Reproduced by permission of Elsevier.

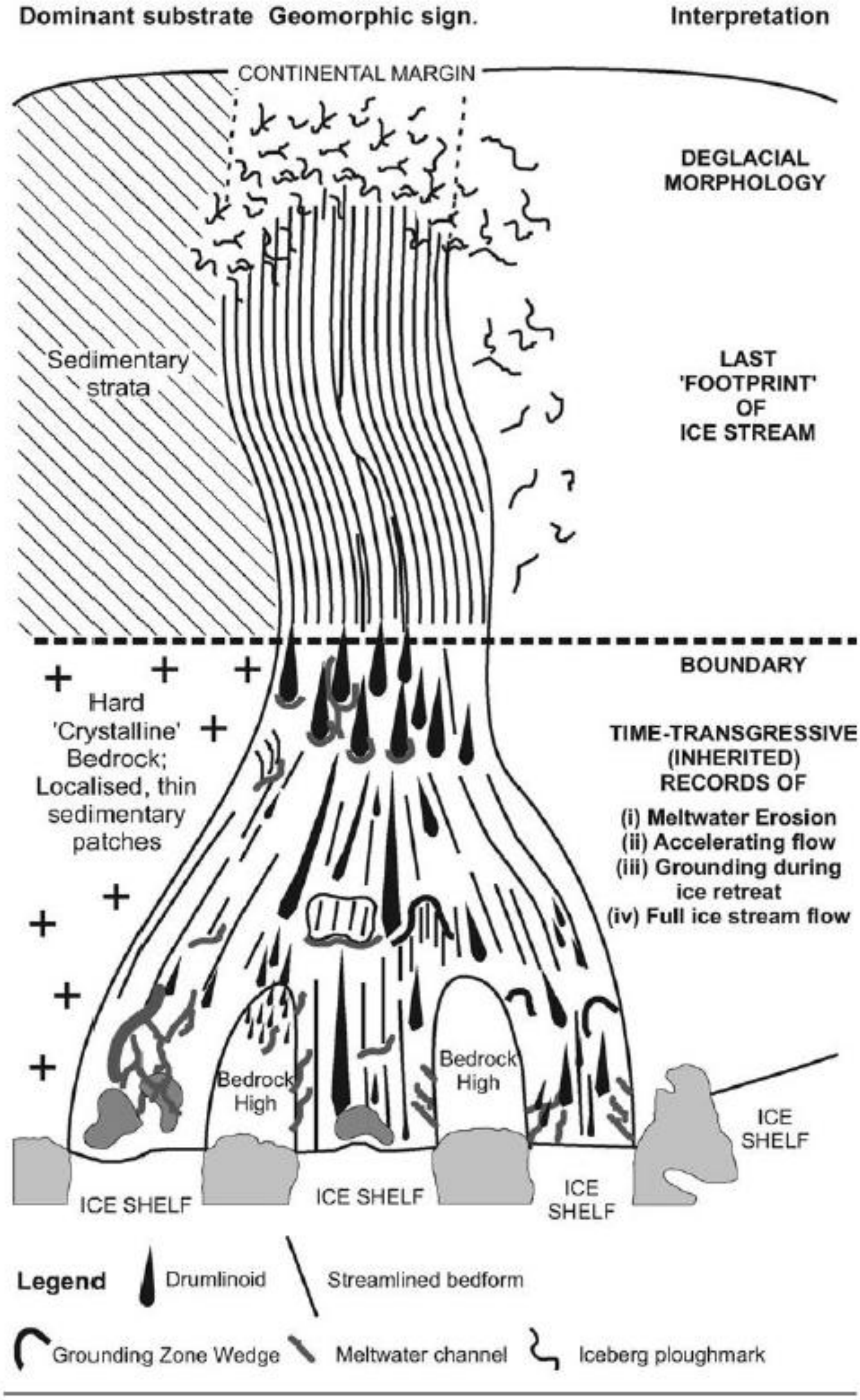


Figure 10: Conceptual model of palaeo-ice stream geomorphology on the west Antarctic continental shelf (from Graham et al., 2009) highlighting the complex time-

transgressive landform assemblage, particularly in ice stream onset zones. Reproduced by permission of Elsevier.

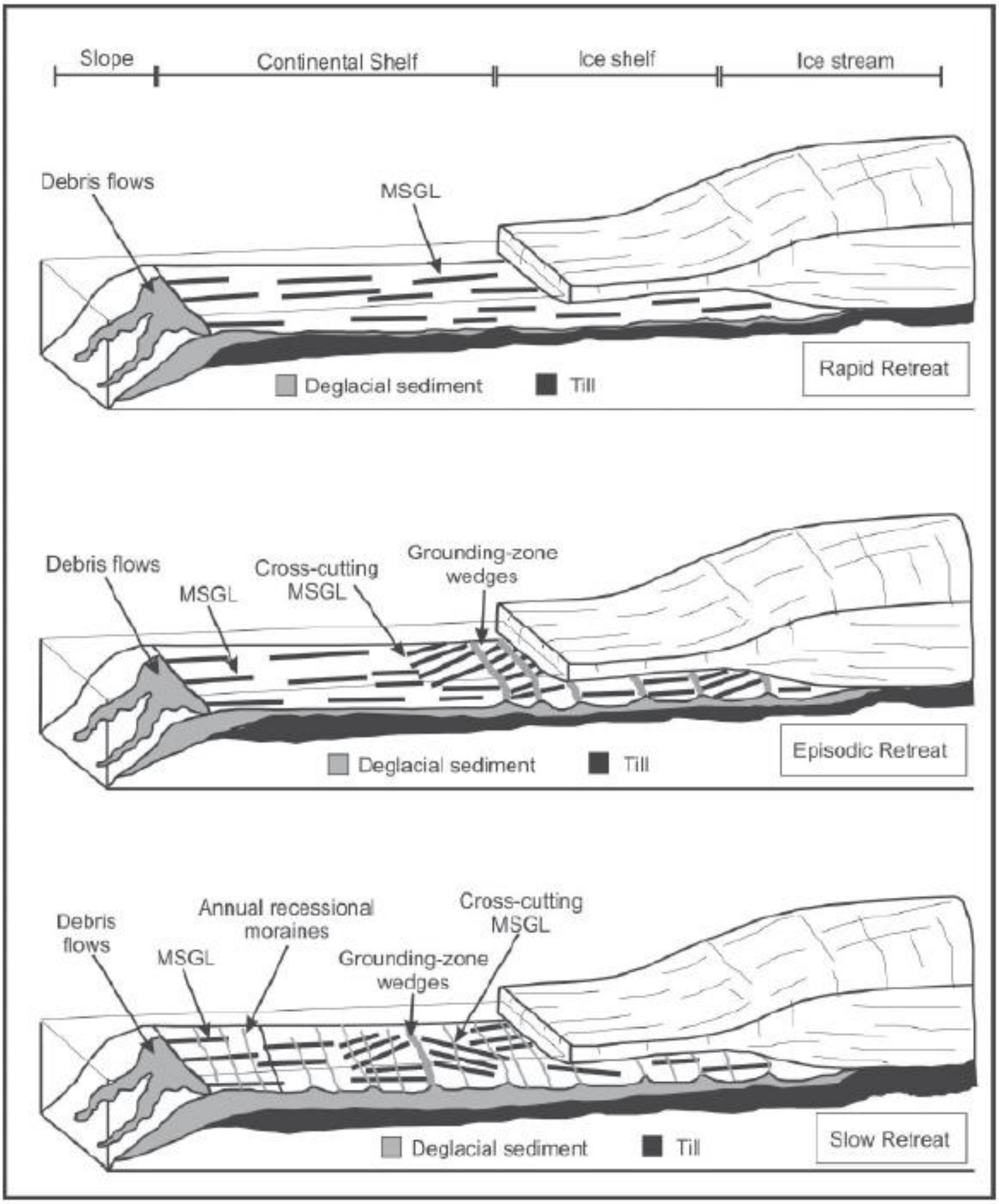


Figure 11: Idealised landsystems models of the geomorphology of palaeo-ice stream tracks that have undergone rapid retreat (top), episodic retreat (middle) and slow retreat (bottom) (from Ó Cofaigh et al., 2008). Reproduced by permission of John Wiley and Sons Ltd.

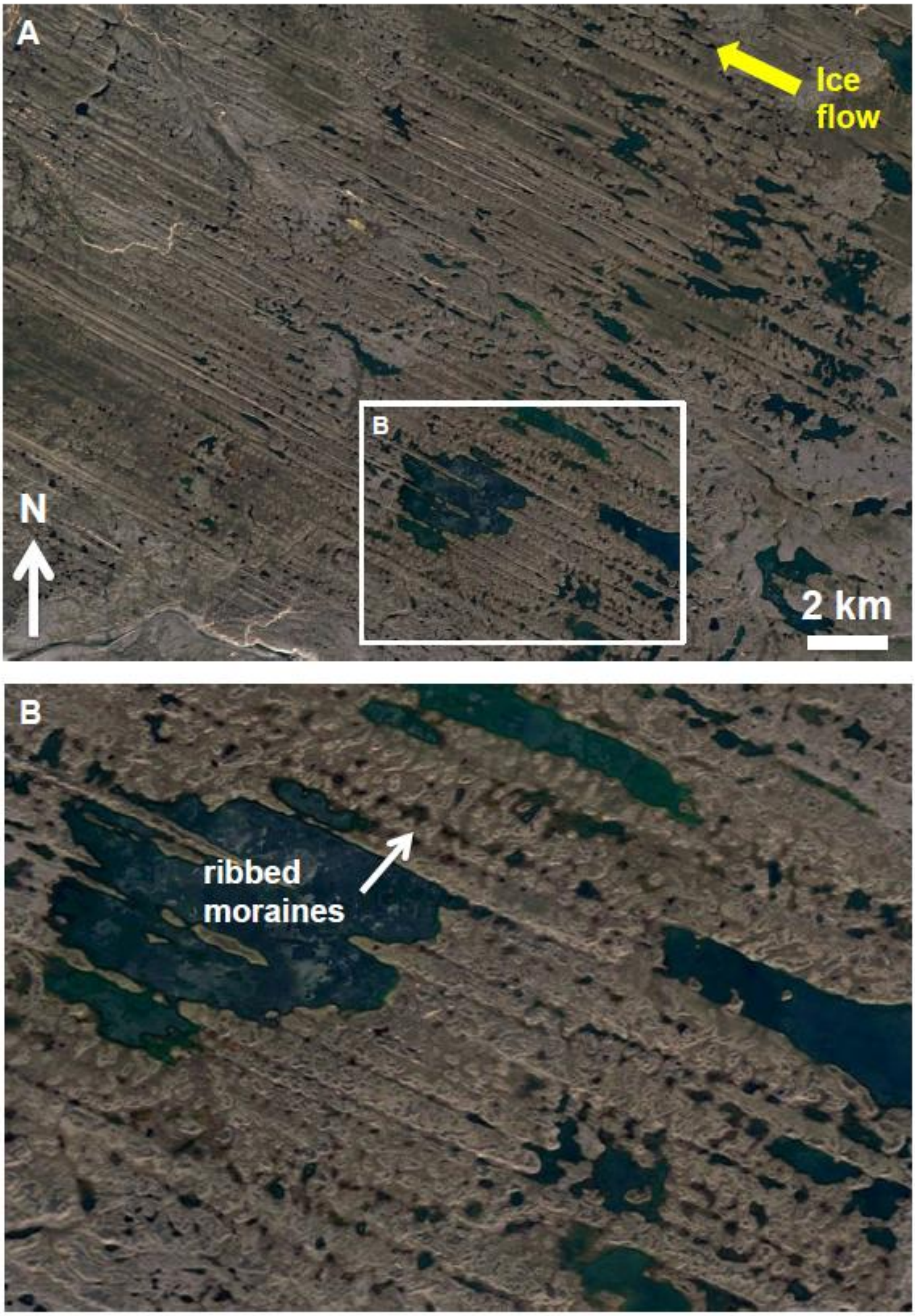


Figure 12: Landsat image (data from Google Earth: Landsat / Copernicus) of ribbed moraines superimposed on mega-scale glacial lineations in the main trunk of the Dubawnt Lake Ice Stream bed. It has been hypothesised that the ribbed moraines

are a manifestation of sticky spots during ice stream shut-down (Stokes and Clark, 2003).

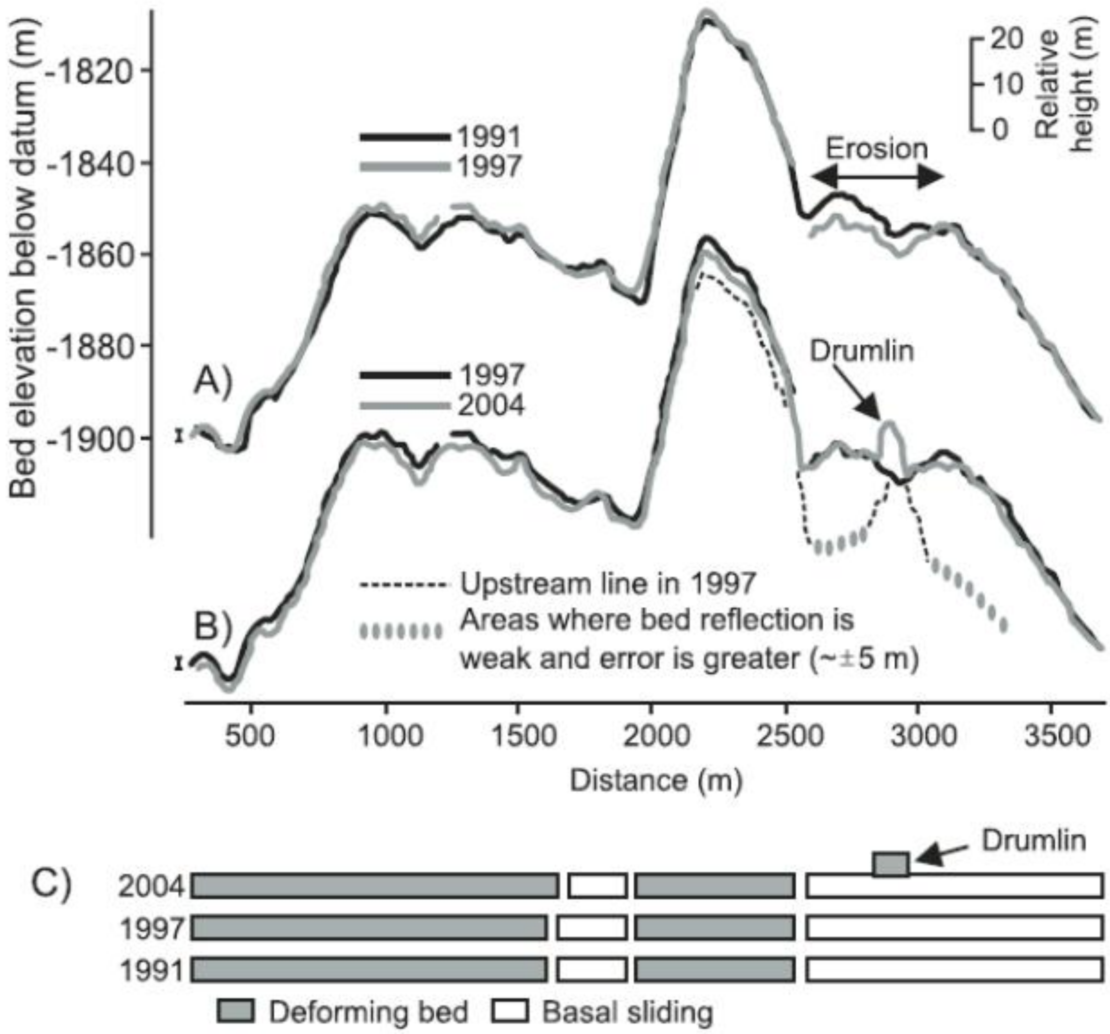


Figure 13: Bed topography and basal conditions at the base of Rutford Ice Stream, West Antarctica (reproduced from Smith et al., 2007, courtesy of the Geological Society of America). (**a**) Bed topography compared between 1991 and 1997. (**b**) bed topography between 1997 and 2004 and from am upstream line in 1997. (**c**) Inferred basal conditions from acoustic impedance data. Ice flow is into the page and error bars are given to the left of each of the lines in (a) and (b).

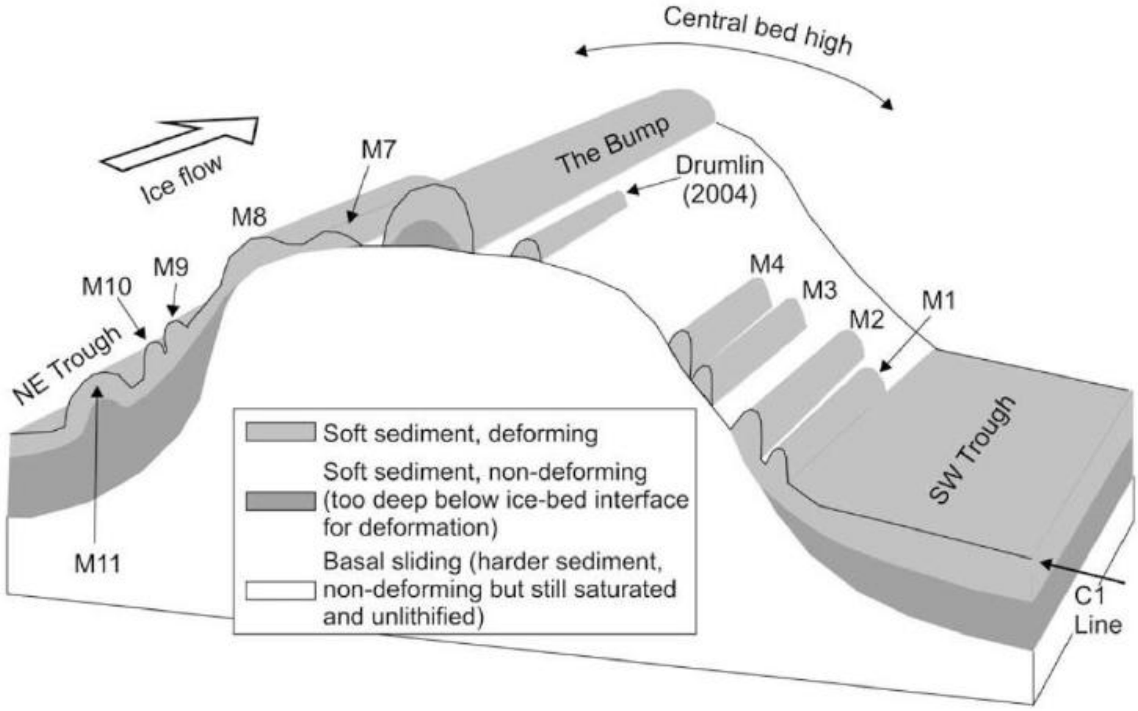


Figure 14: Schematic cross section and reconstruction of the geomorphology at the bed of Rutford Ice Stream, West Antarctica (from Smith and Murray, 2009). The line labelled C1 is the same profile as shown in Fig. 13. Reproduced by permission of Elsevier.

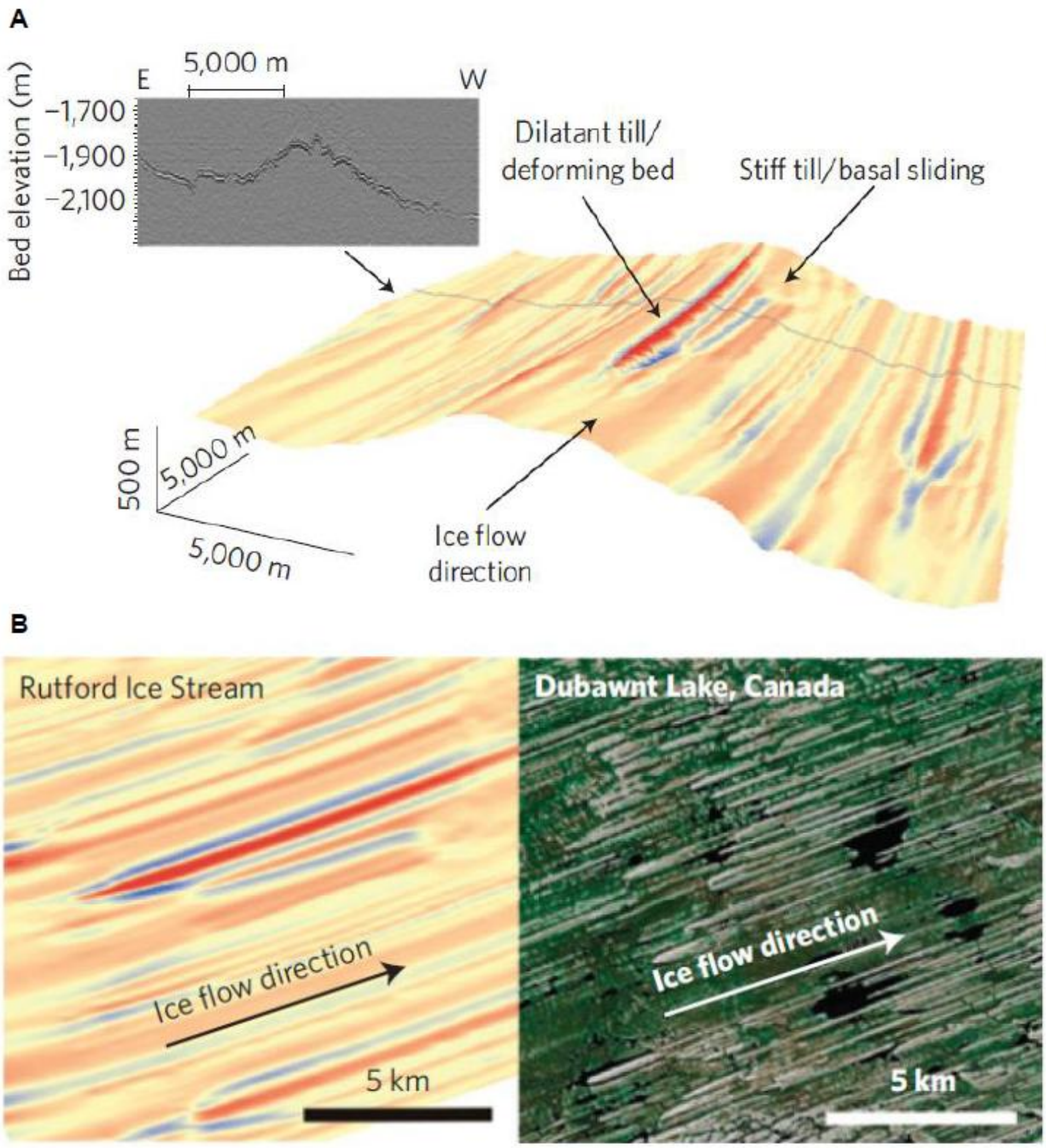


Figure 15: (a) Three dimensional image of the bed of Rutford Ice Stream (from King et al., 2009) looking downstream (see also Fig. 14), with the colour shading showing the difference between the short-wavelength topography and a long wave-length trend surface (example radar profile shown top left). (**b**) Comparison between the mega-scale glacial lineations on the Rutford Ice Stream bed and those on the Dubawnt Lake Ice Stream bed, Canada (Stokes et al., 2003). Figure merged from King *et al*. (2009).

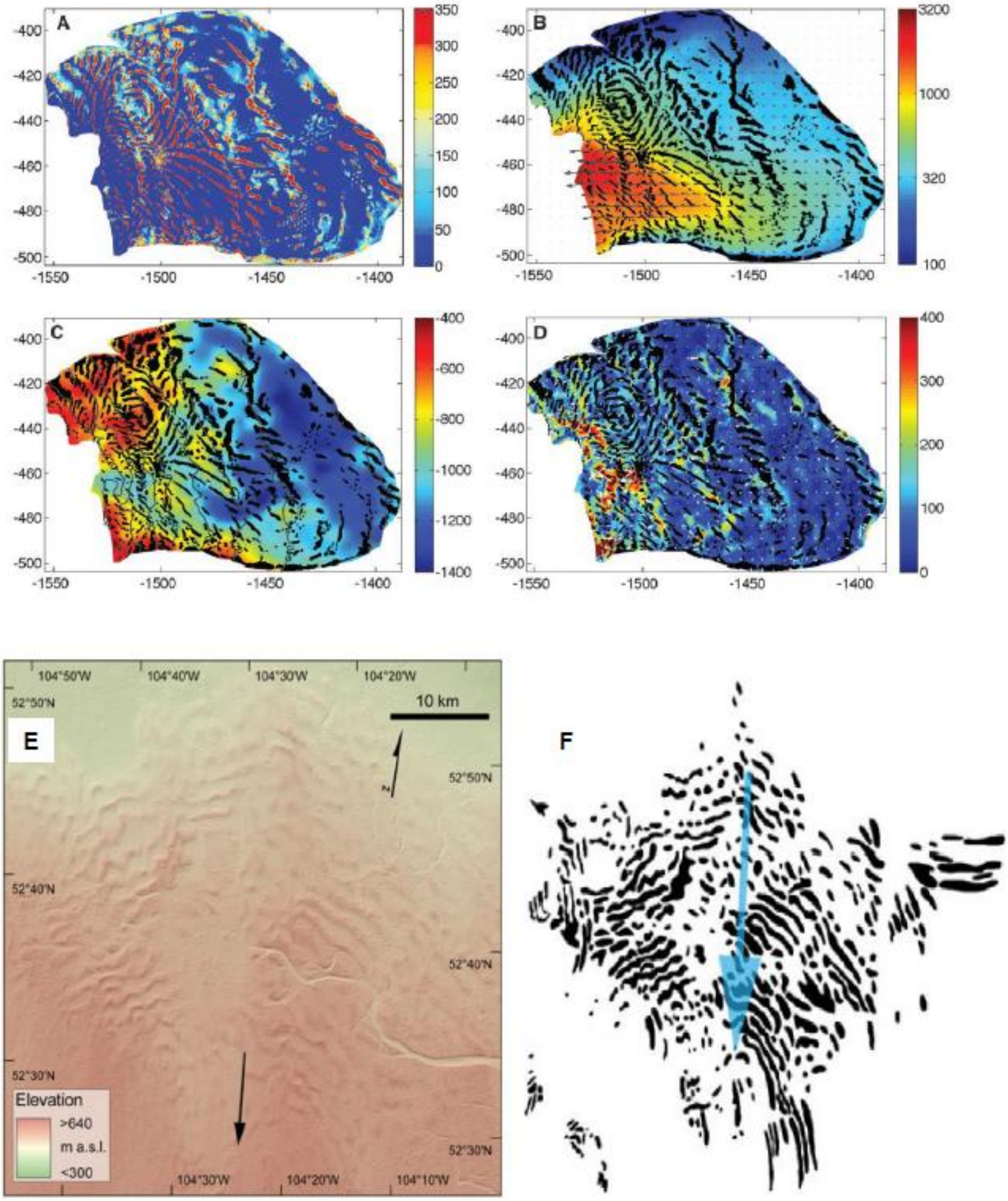


Figure 16: (a) Inverted basal shear stress for Thwaites Glacier, West Antarctica (in kPa); (**b**) Observed ice surface speed with black patches (‘traction ribs’) showing locations where basal shear stress (in a) is >100 kPa); (**c**) Bed elevation and traction ribs (in black); (**d**), Gradient of the hydraulic potential (in pascals per meter), with traction ribs (in black). (**e**) Example of a cluster of low-amplitude rib-like ridges aligned transverse to ice flow on a digital elevation model of the Buffalo Ice Stream corridor in Saskatchewan (see Ross et al., 2009 and Stokes et al., 2016b). (**f**) Black

patches show the location of the central area of rib-like features in (**e**). (**a-d**) taken from Sergienko and Hindmarsh (2013: reproduced by permission from The American Association for the Advancement of Science) and (**e-f**) taken from Stokes et al. (2016).

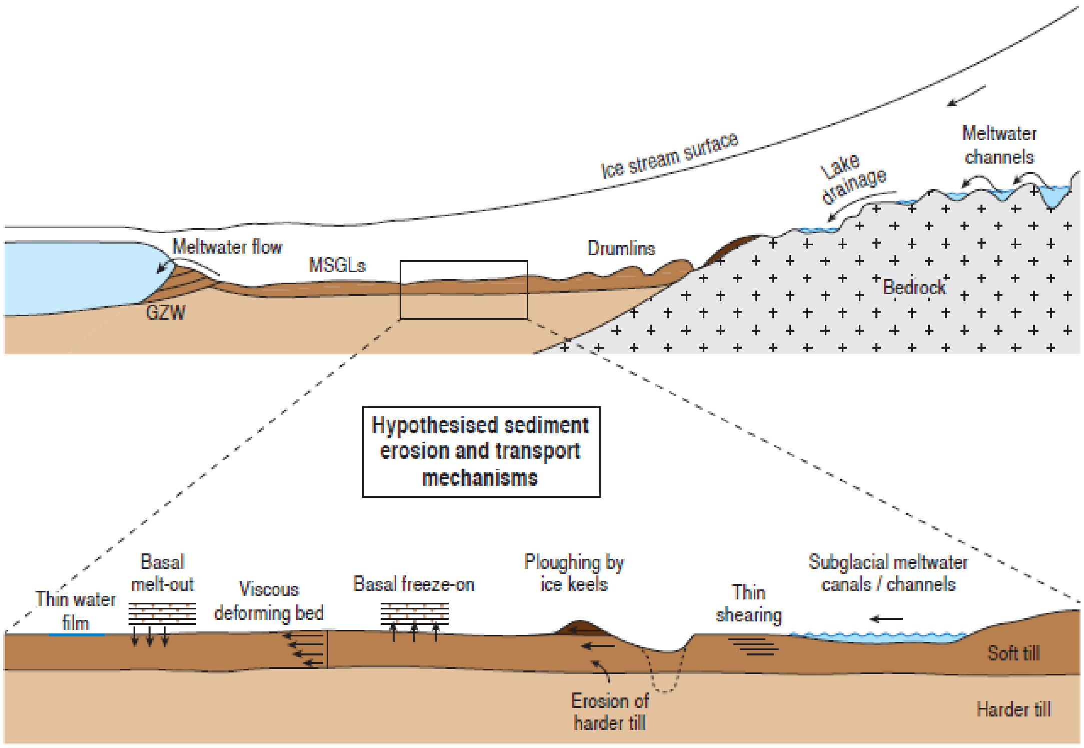


Figure 17: Simplified cartoon (not to scale, but resembling the example in Fig. 9) illustrating key mechanisms of sediment erosion and transport that have been hypothesised to take place under ice streams. Our understanding of these processes is poorly constrained, particularly with respect to the mechanisms and time-scales of sediment erosion, transport and deposition, and how these relate to the growth and decay of subglacial landforms shown in (**a**) and the evolution of basal shear stresses.

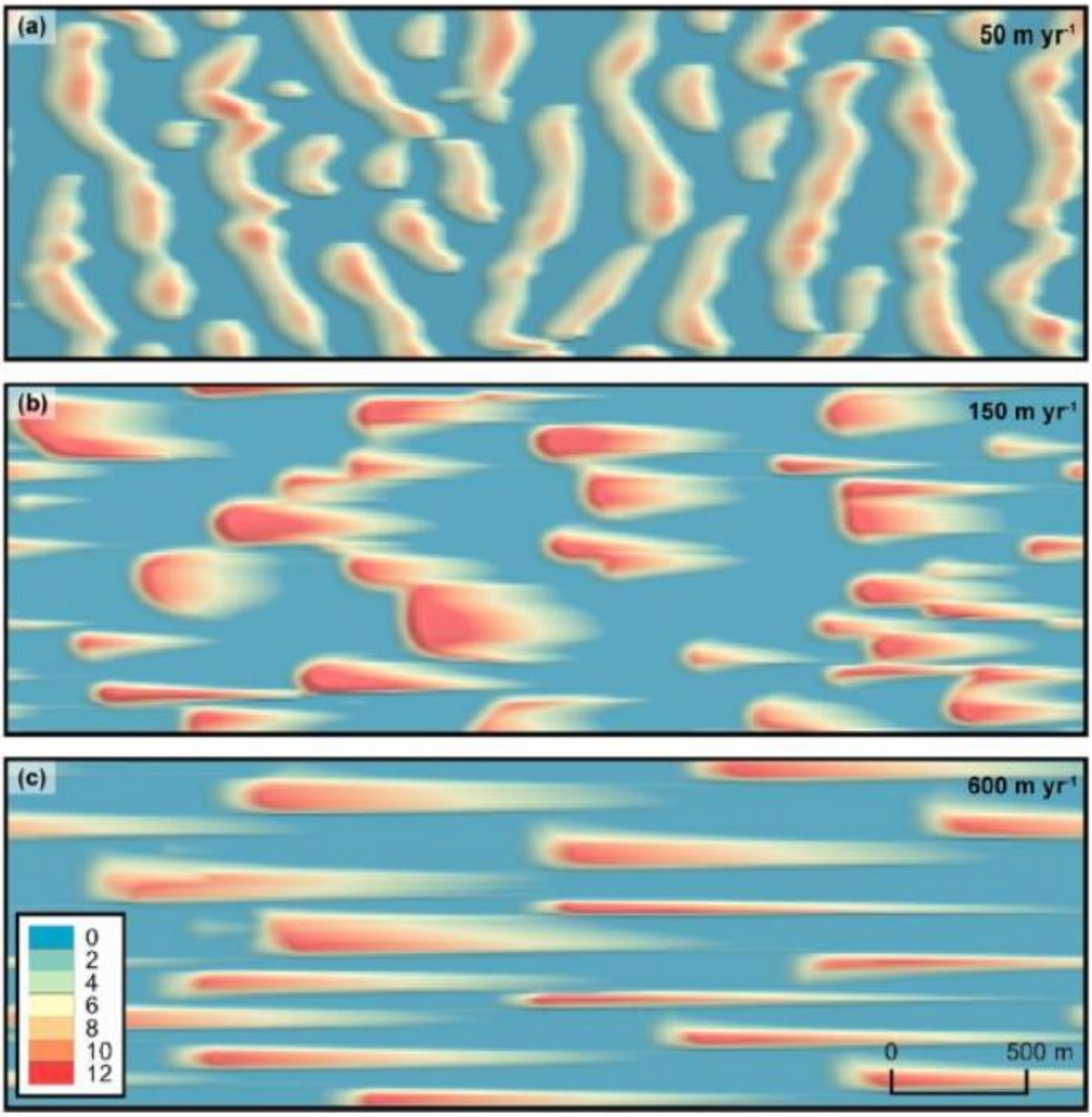
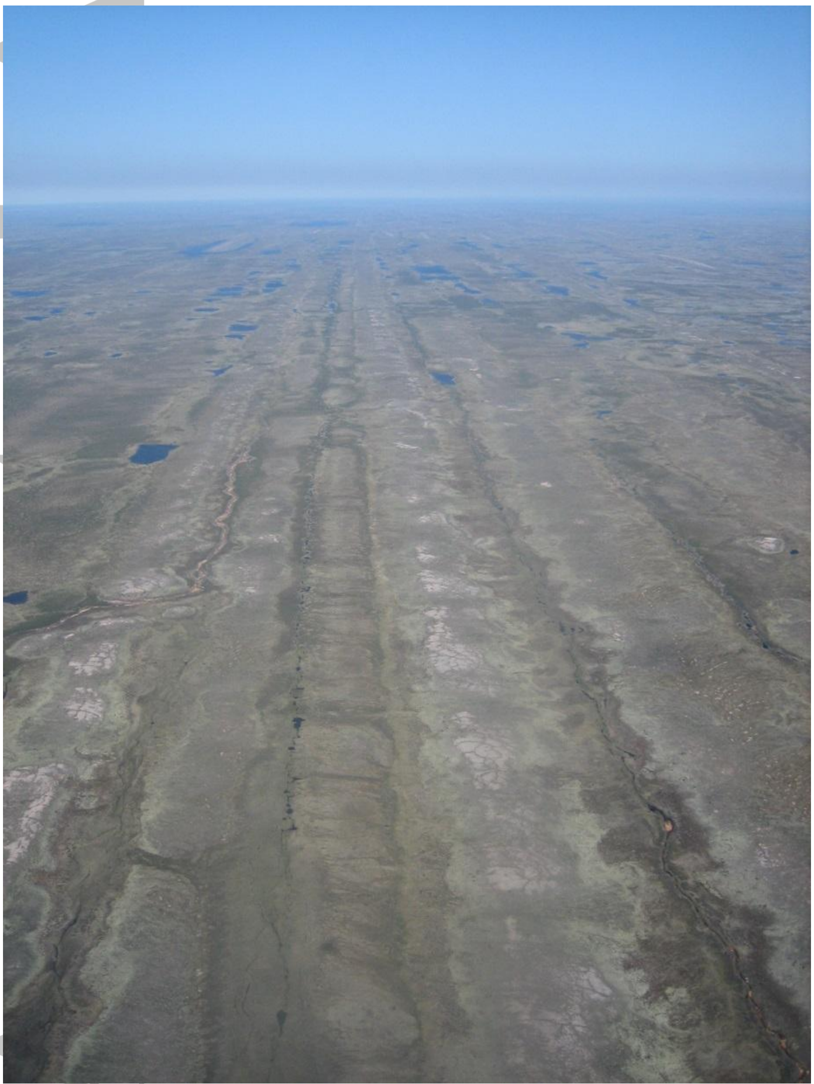


Figure 18: Samples of output from a reduced complexity model aimed at replicating the initiation and development of subglacial bedforms under ice streams (from Barchyn et al., 2016). The transition between ribbed moraines (**a**), drumlins (**b**) and mega-scale glacial lineations (**c**) is associated with increasing ice speeds and declining sediment thickness. Reproduced by permission of John Wiley and Sons Ltd.

Geomorphology under ice streams: moving from form to process

Chris R. Stokes

How much do we know about the geomorphology beneath ice streams? This paper reviews the state of science, emphasising how studies of both modern and palaeo-ice streams have converged to take us from a position of near-complete ignorance to a detailed understanding of their bed morphology. However, key uncertainties remain, particularly regarding the mechanisms and time-scales of sediment erosion, transport and deposition, and how these lead to the growth and decay of subglacial bedforms



Mega-scale glacial lineations on the Dubawnt Lake Ice Stream bed, Nunavut, Canada