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Ice streams as the arteries of an ice sheet: their mechanics, stability and significance

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Abstract

Ice streams are corridors of fast ice flow (ca. 0.8 km/year) within an ice sheet and are responsible for discharging the majority of the ice and sediment within them. Consequently, like the arteries in our body, their behaviour and stability is essential to the well being of an ice sheet. Ice streams may either be constrained by topography (topographic ice streams) or by areas of slow moving ice (pure ice streams). The latter show spatial and temporal patterns of variability that may indicate a potential for instability and are therefore of particular interest. Today, pure ice streams are largely restricted to the Siple Coast of Antarctica and these ice streams have been extensively investigated over the last 20 years. This paper provides an introduction to this substantial body of research and describes the morphology, dynamics, and temporal behaviour of these contemporary ice streams, before exploring the basal conditions that exist beneath them and the mechanisms that drive the fast flow within them. The paper concludes by reviewing the potential of ice streams as unstable elements within ice sheets that may impact on the Earth's dynamic system.
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1. Introduction

In the last couple of decades of the 20th century, there was an increasing concentration in glacial geology on the mechanisms and consequences of fast glacier flow. In particular, attention focused on the significance of ice streams—corridors of fast ice flow within an ice sheet—and their impact on the stability of ice sheets. In many, although not all ice sheets, ice streams discharge the majority of the ice and sediment. For example, over 90% of all the ice and sediment discharged by the Antarctic Ice Sheet today

flows within its ice streams (Bentley and Giovinetto, 1991; Bamber et al., 2000). As a consequence, their occurrence and stability both in space and time is central to the dynamic behaviour of past, present and future ice sheets and is critical in interpreting their geological imprint. Ice streams have the potential to discharge large quantities of ice into an ocean basin, thereby impacting on the thermal and saline circulation therein. Ice streams provide a link between the ocean and cryosphere and episodes of ice streaming have been invoked as a potential forcing for high frequency millennial or sub-millennial climate change (Broecker, 1994b). Understanding the mechanics of fast ice flow and the intrinsic and extrinsic environmental controls which turn such flow on and off is an
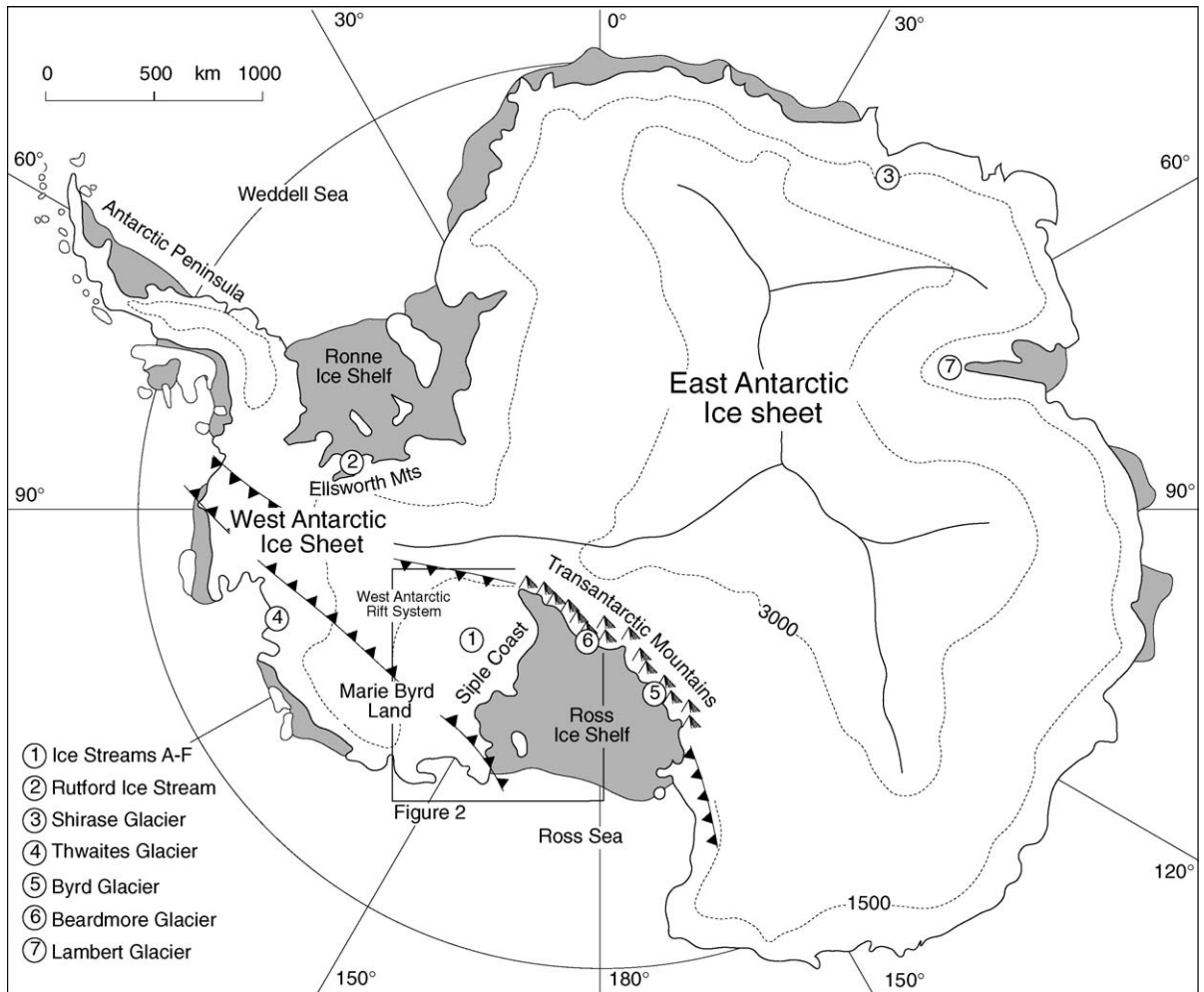
Fig. 1. Map of Antarctica showing the main locations mentioned in the text and the extent of the West Antarctic Rift System.

important research frontier and central to our ability not only to decipher the ancient glacial record but predict the future stability of today's ice sheets.

Ice streams can be investigated in two ways: by either looking at the geological imprint of ancient ice streams within Pleistocene (Ice Age) ice sheets (Ice Age; Stokes and Clark, 1999, 2001); or by investigating those ice streams which exist within today's ice sheets. Most research has concentrated on contemporary ice streams, which are best developed along the Siple Coast of Antarctica (Figs. 1 and 2), and scientists have investigated these ice streams intensively since the early 1980s. These ice streams named some-

what prosaically Ice Streams A to F[1] discharge 40% of the ice from the West Antarctic Ice Sheet (Price et al., 2001; Figs. 1 and 2). They have been extensively investigated due to their potential importance to the stability of the West Antarctic Ice Sheet (Alley and Bindschadler, 2001). The maritime nature of the West Antarctic Ice Sheet—much of its bed lies below sea level—has led many to question its stability, since a

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[1] Ice Stream B was renamed Whillans Ice Stream in 2001 in honour of Professor Ian M. Whillans who was a major figure in the study of West Antarctic ice streams and died in May 2001 (Anon., 2001).
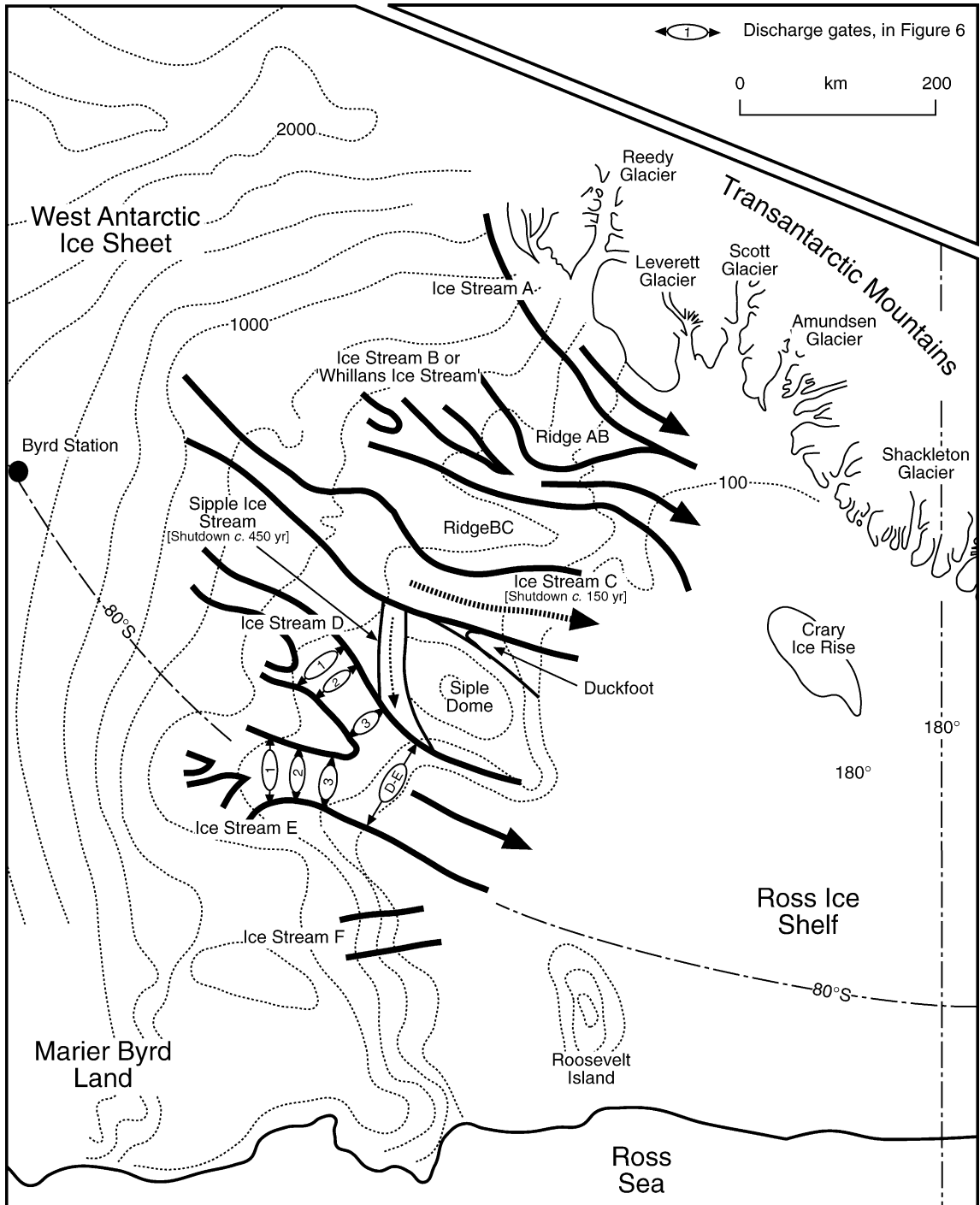
Fig. 2. Map of the Siple Ice Streams. See Fig. 1 for location.

collapse of the ice sheet would raise global sea level by the order of 5 or 6 m. As a result of the intense scientific interest in the Siple Ice Streams, they have stimulated several hundred research papers in the last 20 years. The aim of this paper is to provide an accessible overview of this substantial body of literature in order to provide some insight into the role of ice streams within the glacial system. Specifically, this review will focus on: the morphological and dynamic characteristics of the Siple Ice Streams; their spatial and temporal distribution; the mechanics of fast ice flow within them; and ultimately on the role of ice streams in general in coupling ice sheets, oceans and climate. I will start with a general definition of an ice stream, before moving on to examine the characteristics of the Siple Ice Streams in Antarctica.

2. Definition: what is an ice stream?

Swithinbank (1954) defined ice streams as 'part of an inland ice sheet in which the ice flows more rapidly than, and not necessarily in the same direction as, the surrounding ice'. Bentley (1987) argued from this definition, that an ice stream must be bounded laterally by ice rather than rock and that the ice must be grounded rather than floating as in an ice shelf (Fig. 3A). He distinguished between topographically controlled outlet glaciers, which may be associated with enhanced flow velocities, and ice streams, which lack significant topographic control. This distinctions has been continued since through the use of such qualifiers as 'pure' and 'topographic' ice streams (Stokes and Clark, 1999), largely because the term outlet
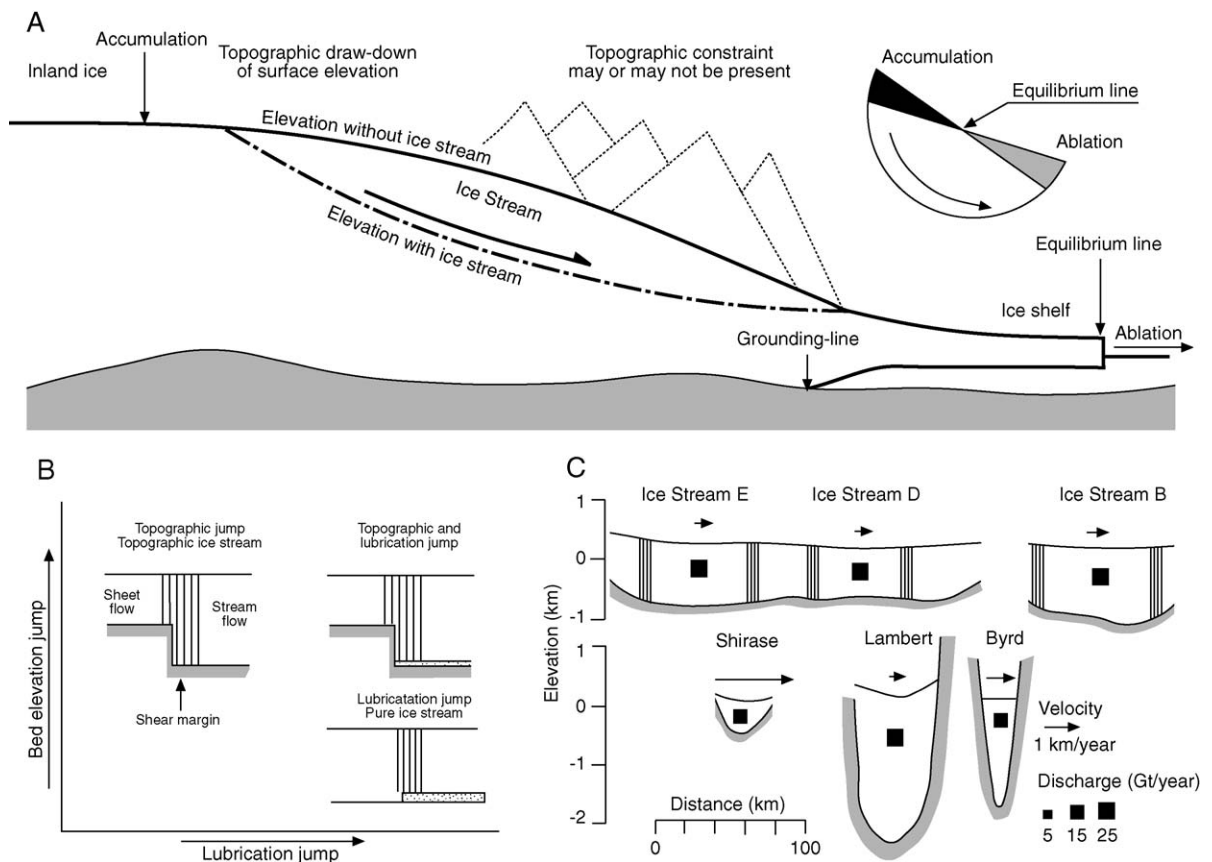

Fig. 3. (A) Schematic diagram of an ice stream. (B) Main types of ice-stream margin (after: Raymond et al., 2001). (C) Cross-sections through several pure and topographic ice streams (after: Bentley, 1987).

glacier need not necessarily be associated with en-hanced rates of flow. Topographic lows tend to ex-perience faster ice flow for several reasons. Firstly, the thicker ice, within topographic lows, results in a greater driving stress at the bed, and therefore, flow velocity, since the internal deformation of ice is strongly controlled by basal shear stress as defined by Glen's flow law (Paterson, 1994). Secondly, thicker ice means greater insulation and therefore increased basal temperatures, which not only enhan-ces the rate at which ice may deform, but may also enhance bed slip via increased basal melting and lubrication. Finally, meltwater is more likely to exist in topographic lows beneath an ice sheet, since the direction of subglacial water flow is driven both by ice surface slope and bed topography. Moreover, melt-water is more likely to exist beneath thick ice where the melt rate is greater. Clearly, all these variable may operate within feedback systems in which enhanced flow increases temperature and basal lubrication and in turn accelerates flow velocity (e.g. Clarke et al., 1977). The net result is a tendency for ice flow to accelerate within topographically constrained corri-dors. Where such corridors are absent, ice streams are associated either with corridors of ice which are rheologically weaker than the surrounding ice, or alternatively with some form of lubricated bed which facilitates rapid basal motion (Fig. 3B). As Fig. 3B acknowledges, it is possible for an ice stream to be both topographically constrained and have a lubri-cated beds and these distinctions are not easy to apply in practice (Bentley, 1987). The ice streams of the Siple Coast (Figs. 1 and 2) provide the only contem-porary examples of pure ice streams, with perhaps the exception of a rapid flow corridor identified in north-east Greenland (Fahnestock et al., 1993; Joughin et al., 2000). Glaciers such as the Amundsen, Beard-more, Shackleton or Byrd glaciers, which drain through the Transantarctic Mountains to discharge into the southern and western flank of the Ross Ice Shelf are all topographically constrained and therefore classed as outlet glaciers or topographic ice streams (Figs. 1 and 2). The Rutford Ice Stream provides a complication, in that it is flanked on one side by the Ellsworth Mountains and on the other by ice (Fig. 1; Bentley, 1987). Detailed observations have also shown that it is guided by both by basal topography and by a jump in lubrication (Fig. 3B; Smith, 1997).

Similarly, the Slessor Glacier shows little topographic constraint on the surface, but is constrained subgla-cially by a well-defined valley. Finally, the Lambert Glacier starts as a pure ice stream and then becomes topographically constrained (Fig. 1; Hambrey and Dowdeswell, 1994). It is important to emphasise that both pure and topographic ice streams may be asso-ciated with a significant flux of ice and debris, and are therefore both of importance to the geometry of an ice sheet and the patterns of erosion and sedimentation within it (Fig. 3C).

The physics of topographic and pure ice streams are however very different. As illustrated in Fig. 3A, ice streams lower the surface topography of an ice sheet; the degree to which an ice sheet is drawn-down is significantly greater for a pure ice stream due to the greater ice flow within it (Fig. 4A). Glacier flow is a result of a gravitational driving force, the down-slope weight of the ice, which is a function of the ice surface slope and ice thickness, and ice normally deforms as a result of basal shear stress in the range of 50–100 kPa (Paterson, 1994). Within an ice sheet flowing via sheet flow, this gravitational driving force is supported on the glacier bed. The resultant basal shear stress drives glacier flow and reaches a peak just below the equilibrium line, the point on a glacier surface where up-stream accumulation balances down-stream mass loss or ablation (Fig. 3A). For most Antarctic glaciers, this point lies at sea level, and the glacier may drain directly into floating ice shelves. Topographic ice steams in Antarctica show a peak in basal shear stress between 50 and 100 km from the margin or ground-ing-line as illustrated by the Shirase, Lambert or Thwaites glaciers in Fig. 4B. In contrast, pure ice streams (e.g. Whillans Ice Stream) show a peak in gravitational driving stress at the head of the stream followed by a down-flow decline to extremely low values—well below that necessary to deform ice significantly—near the grounding-line (e.g. 20 to 30 kPa; Fig. 4B). The peak in ice flow velocity within these streams is not coincident with the peak in glacial driving stress as is the case in topographic ice streams (Fig. 4C) and this dynamic difference helps distin-guish pure ice streams. In addition, pure ice streams appear to experience patterns of spatial and temporal instability, shifting location and undergo cycles of varying flow activity. It is this potential for instability which attracts the attention of glaciologists and the
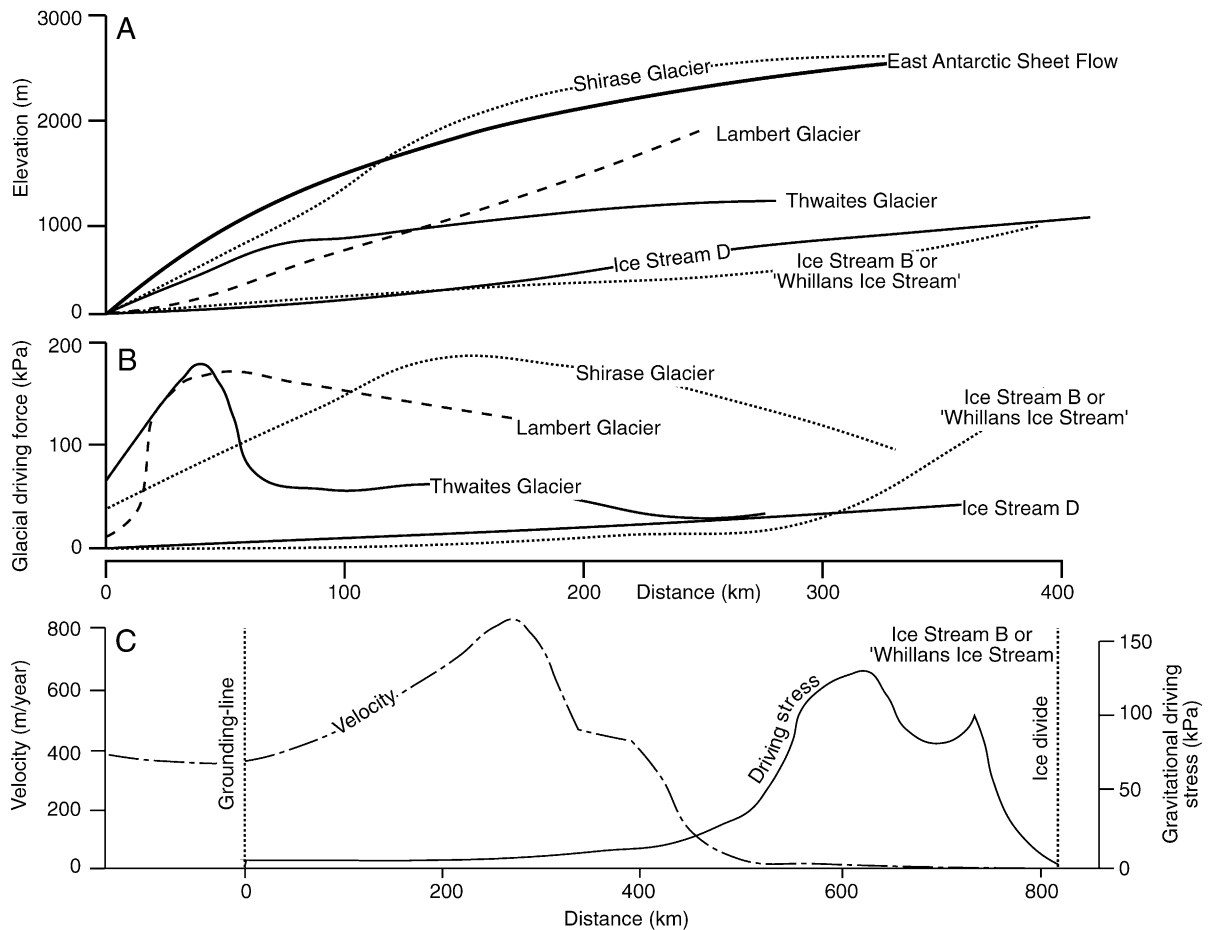
Fig. 4. (A and B) Longitudinal profiles of surface elevation and glacial driving stress for various ice streams. A typical profile for part of the East Antarctic Ice Sheet, moving via sheet flow, is provided for comparison (after: Bentley, 1987). (C) Longitudinal profiles of flow velocity and glacial driving stress for the Whillans Ice Stream (after: Alley and Whillans, 1991).

pure ice streams of the Siple Coast in Antarctica have, as the best examples available today, been extensively researched. As a result, the Siple Ice Streams form the main focus of this paper.

3. Morphology and dynamics of the Siple Ice Streams

The Siple Ice Streams, of the West Antarctic Ice Sheet, drain into the Ross Ice Shelf (Fig. 1). Ice Streams A to F are typically 50 km wide, 300 to 500 km long and have an average ice thickness of at least 1 km. Flow velocity within these streams

exceeds 0.1 km/year and often rises to over 0.8 km/year (Fig. 2; Shabtaie et al., 1987; Whillans et al., 1987, 2001). Each stream is defined by two lateral shear zones, which mark the transition from the slow (5 m/year) ice flow outside the stream, to the fast flow within the stream (100 to 800 m/year). Intense deformation within this shear zone results in a chaotic zone of surface crevasses 3 to 5 km wide (Echelmeyer et al., 1994), and to intense frictional heating as indicated by the temperature profiles in Fig. 5A (Harrison et al., 1998). The intervening ridges, between streams, form interfluves of ice frozen to the bed (cold-based), in contrast to the ice streams which are at pressure melting point; the junction between frozen and unfro-
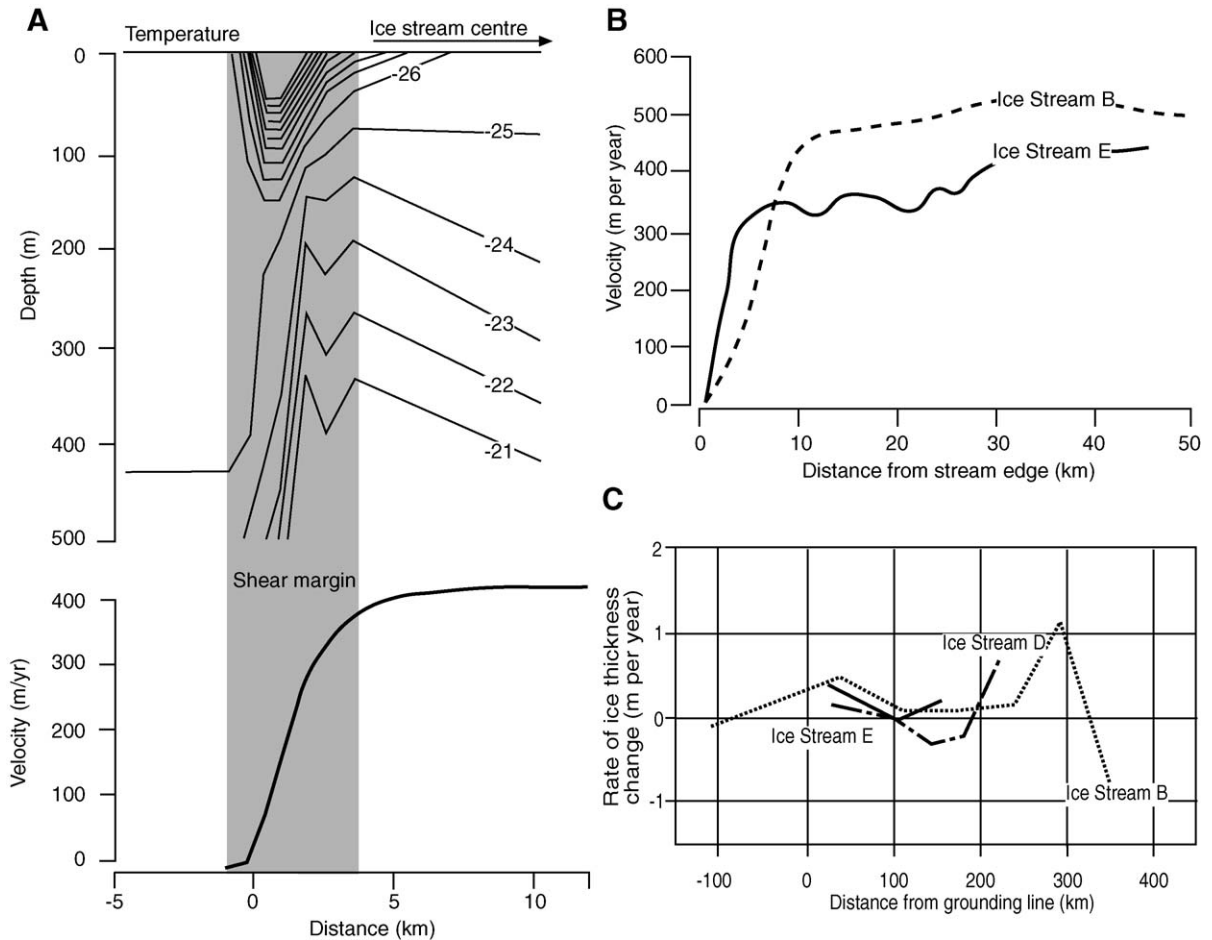
Fig. 5. (A) Change in englacial temperature across a part of the shear margin of Ice Stream (after: Echelmeyer et al., 1994; Harrison et al., 1998). (B) Comparison of the rate of change in ice velocity across the shear margins of the Whillans Ice Stream and Ice Stream E (after: Bindschadler and Scambos, 1991). (C) Changes in ice thickness along the length of several ice streams (after: Bindschadler, 1997).

zen basal ice occurs beneath the shear margins of the ice streams (Bentley et al., 1998). Each stream forms part of a dendritic drainage system, in which the trunk of an ice stream is fed by a network of potentially competing tributaries—the overlapping branches of several trees—which in some cases may drain a common accumulation area (Fig. 2; Joughin et al., 1999). Although the trunks of each ice stream are not particularly well constrained by subglacial topography (Shabtaie et al., 1987; Fig. 3B), the tributaries do appear to be focused along subglacial topographic lows, where concentrations of thicker and therefore warmer ice, subglacial sediment or water are likely (Joughin et al., 1999). These tributaries are not always

associated with surface crevasses, which tend to develop when ice velocity within a stream exceeds a threshold of about 100 m/year, while the typical rates of flow within a tributary may only be 25 m/year (Joughin et al., 1999).

The spatial and temporal structure of flow velocity within ice streams has been subject to widespread investigation using a range of survey techniques from direct surveys using various global positioning systems (e.g. Whillans et al., 1987) to earth observation systems using serial satellite images (e.g. Bindschadler et al., 1996), or more recently interferometric radar images (Joughin et al., 1999). Detailed observations of flow velocity were first made for the

Whillans Ice Stream (Ice Stream B) and its tributaries. The data showed a pattern of spatially varying flow velocity with rates from as much as 827 m/year to more typical values of around 400 m/year, contrasting with flow in the ice interfluves of as little as 5 m/year (Whillans et al., 1987, 2001). Fig. 6 shows several cross-sections, transverse to the direction of flow, through Ice Streams D and E determined from serial satellite images (Bindschadler et al., 1996). These flow profiles illustrate the complex nature of the flow regime within an ice stream and its tributaries and also emphasise the absence of a well-defined subglacial topography beneath a stream. Velocity declines rapidly at the stream margins and the rate of this change varies from ice stream to ice stream (Figs. 5B and 6). This variation between ice streams may be a function of a stream's spatial stability, with narrow shear zones developing where streams have greater stability and consequently allow the shear zone to become concentrated in an ever decreasing band of intense deformation (Bindschadler and Scambos, 1991).

The spatial pattern of ice velocity across the surface of an ice stream is complex and results in variations in the rate of strain causing local thinning and thickening of the ice (Bindschadler et al., 1987, 1993; Bindschadler and Scambos, 1991). If ice slows down-stream locally, then the ice must thicken to form a bulge, given uniform bed topography. This is of particular importance to our understanding of the basal processes responsible for fast ice flow. If the ice within a stream had a uniform rheology and moved as a single mass over a basal surface with a uniform level of lubrication, such local variation in flow should not occur. The fact that it does provide a clue to the spatial variability of basal conditions beneath an ice stream.

This spatial variation in strain results in a range of topographic features of the surface of the ice stream which have been described using a plethora of poorly defined terms, including 'mottles', 'warps', 'horsetails', 'chromosomes', 'flow stripes', and 'rafts'. For example, the surface of an ice stream contains an irregular topography of longitudinal ridges and troughs ('flow stripes') a few hundred metres across, tens to hundreds of kilometres long, and with amplitudes of a few metres (Whillans et al., 2001). These flow stripes are orientated parallel or sub-parallel to the direction of flow, and are not always in phase with local velocity fields (Shabtaie et al., 1987; Merry and Whillans, 1993). A range of possible explanations have been advanced to explain 'flow stripes'. For example, they have been modelled as the product of perturbation in ice motion over basal topography (e.g. Gudmundsson et al., 1998). Alternatively they have been interpreted as the result of the attenuation of ice surface topography at shear margins (Merry and Whillans, 1993). In addition to 'flow stripes', there are also a wide range of more irregular topographic highs and lows. 'Lumps' are topographic highs, typically 1.5 km long, 300 m and about 600 m apart. They form distinct trains within shear margins and are orientated transverse to the principal compressive stress. In contrast, 'warps' are larger compressional buckles on the surface of an ice stream (4 km long, 1.5 km wide and 8 km apart; Merry and Whillans, 1993). 'Mottles' are more subdued and longer wavelength forms that may occur both within ice streams and within areas undergoing sheet flow and have been explained in relation to flow around subglacial obstacles (Whillans and Johnsen, 1983). Some of these topographic forms appear to be stable, forming standing waves within the flow, while others may show evidence of up-stream migration (Hulbe and Whillans, 1997). These flow structures—'lumps', 'warps' and 'mottles'—have been variously explained as: (1) a consequence of transverse compression within a stream guided by variation in ice strength (Whillans and van der Veen, 1993; Hulbe and Whillans, 1997); (2) a result of the upwards propagation of basal topography which provides an obstacle to rapid basal motion (Bindschadler and Scambos, 1991; Gudmundsson et al., 1998); or (3) a result of spatial variations in basal lubrication (sticky spots; Alley, 1993; Hindmarsh, 1998c). A clear relationship between basal topography and surface topography is not always present (Whillans and van der Veen, 1993; Bindschadler et al., 1996), which suggests that spatial variation in the ease of basal movement beneath an ice stream may be of greater importance.

One feature of particular interest on the surface of ice streams is 'rafts' of smooth ice. These 'rafts' appear to move down-stream over time, and have been attributed to the assimilation of marginal slabs of cold ice during the headward or lateral erosion of a stream's margins. The entrained slabs are then transported down-stream (Shabtaie et al., 1987).
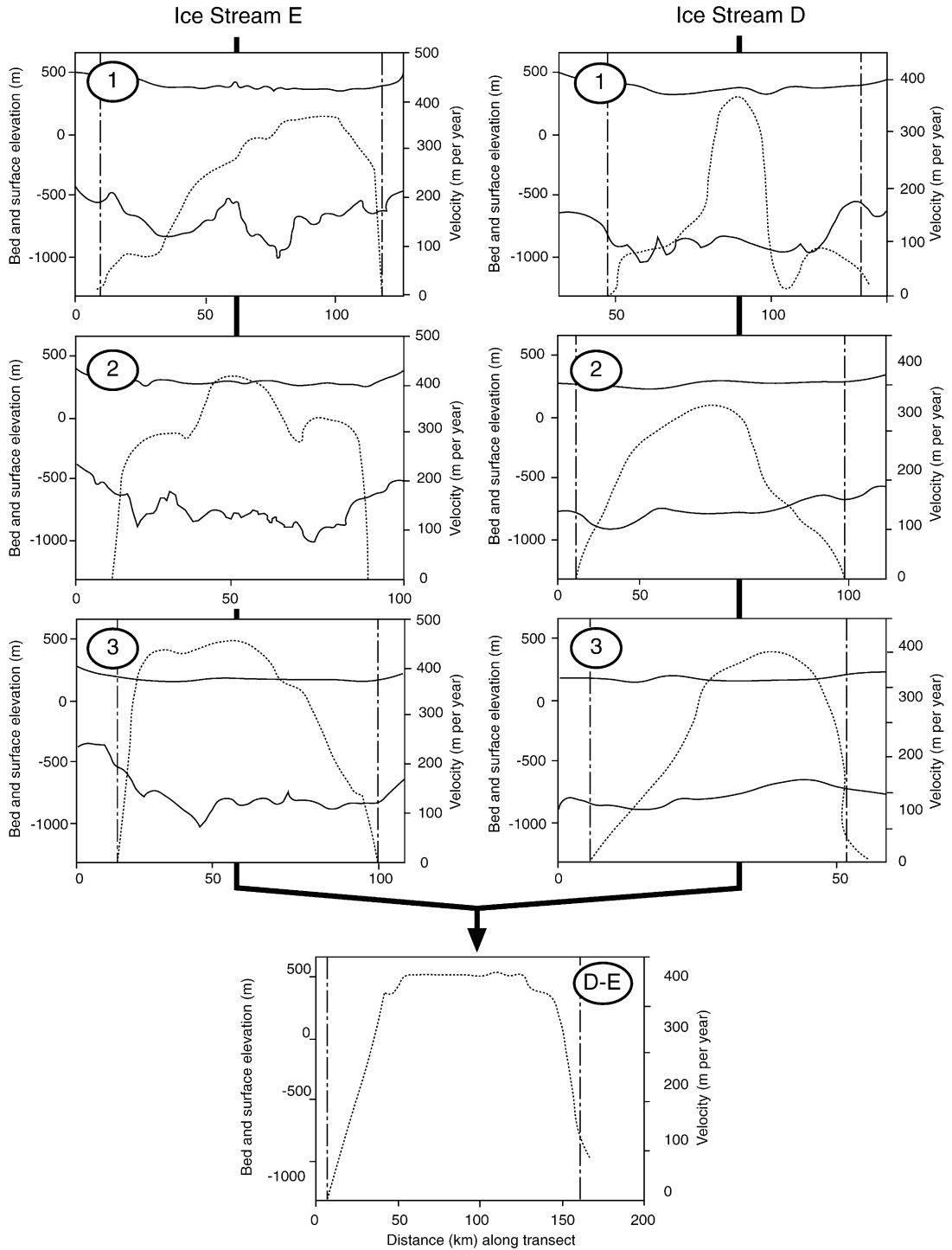
Fig. 6. Velocity cross-sections for Ice Streams E and D. Dotted line is velocity. The location of each cross-section is shown in Fig. 2 (after: Bindschadler et al., 1996).

4. Temporal behaviour of the Siple Ice Streams

In general, glaciers show fluctuations in ice velocity at an hourly, diurnal or seasonal scale, associated with changes in the availability of basal meltwater, as rainfall, heat and melt budgets vary through a day or season (Willis, 1995). High basal water pressures are essential for rapid basal motion since they counter the vertical weight of the ice above the bed (normal pressure), in much the same way as a hydraulic jack, which then reduces the friction at the bed facilitating basal motion. McDonald and Whillans (1992) showed that the Whillans Ice Stream has less than 0.4% variability in velocity on sub-annual scale, a fact that probably reflects a climate that gives minimal surface melt.

Velocity fluctuations do, however, occur on scales longer than a year (>1 to 10 s). For example, the mouth of the Whillans Ice Stream, close to the point at which it enters the Ross Ice Shelf, has been slowing, widening and consequently thickening over the last few decades (Stephenson and Bindschadler, 1988; Bindschadler and Vornberger, 1998). Since 1963, the mouth of the Whillans Ice Stream has widened at rate of 137 ± 34 m/year, while the movement of prominent ice rafts within the ice stream suggests that its velocity is slowing at a rate of 2.4% per year. In 1963, the velocity was 967 m/year compared with just 471 m/year in the mid-1980s. This rate of slowing is supported by other observations at the ice stream mouth and has resulted in a thickening (0.13 m/year) of the ice surface (Stephenson and Bindschadler, 1988). The deceleration of the Whillans Ice Stream seems to be continuing, and Joughin and Tulaczyk (2002) suggest that velocity decreased by 23% between 1974 and 1997 (Joughin and Tulaczyk, 2002). A similar fall of flow velocity and widening has been observed in the ice stream's upper tributaries, although the rate of widening is much less (10 m/year; Hulbe and Whillans, 1997; Harrison et al., 1998). Bindschadler (1997) has documented changes in ice thickness for some of the Siple Ice Streams (Fig. 5C).

The Siple Ice Streams not only show variability in width and thickness, but also undergo longitudinal extension. Theoretical arguments (Alley and Whillans, 1991), and inferred results (Bindschadler, 1997), predict that the point of onset of an ice stream should migrate inland at a rate of the order of 500 m/year.

Using crevasse patterns, Price and Whillans (2001) have recently measured the rate of up-stream migration for a tributary of the Whillans Ice Stream and obtained a rate of 230 m/year. If this migration continued, the ice stream head would reach the ice divide of the West Antarctic Ice Sheet in approximately 1400 years.

On much longer time-scales (>100 years), there is marked variability in the flow regime of an ice stream with the potential for an ice stream to switch on an off, as illustrated by Ice Stream C. Currently, Ice Stream C has little or no surface velocity with a flow rate of only 4 to 5 m/year (Whillans et al., 1987, 2001; Fig. 7). The surface of the ice stream is not visibly crevassed and the flow traces are subdued. However, radar-sounding shows structures typical of other ice streams at depth suggesting that it was once as active as the adjacent streams (Rose, 1979). Since the ice stream ceased to flow, snow has accumulated on the surface and therefore the depth of accumulation provides an indication of the age at which rapid flow ceased, given that the accumulation rate is known. Using this approach, Retzlaff and Bentley (1993) made five estimates of the date when Ice Stream C shutdown at different points along its length. The three most down-stream estimates gave ages of 130 ± 25 years, while further up-stream, the age fell to 100 ± 30 years, and active flow may only have ceased 30 years ago in the ice streams uppermost reaches. This suggests that a wave of stagnation propagated at a diminishing speed up-stream from the ice stream mouth (Retzlaff and Bentley, 1993). Despite being inactive, Ice Stream C still appears to be underlain by an unfrozen, sediment-rich bed (Bentley et al., 1998), and to have hydraulic properties similar to those of the Whillans Ice Stream (Kamb, 2001). Recent work by Jacobel et al. (2000) has suggested that the shutdown of Ice Stream C was a more complex event due to the presence of a folded shear margin on the northern side of the ice stream, to the south of the Siple Dome (Duckfoot, Fig. 2). The northern margin of the ice stream appears to have undergone a step-wise migration to the south just prior to its stagnation, perhaps associated with a decrease in flow velocity and a reduction in the width of the stream. During this migration, a region of ice-stream trunk and shear margin was folded between the recent (inner) and older (outer)
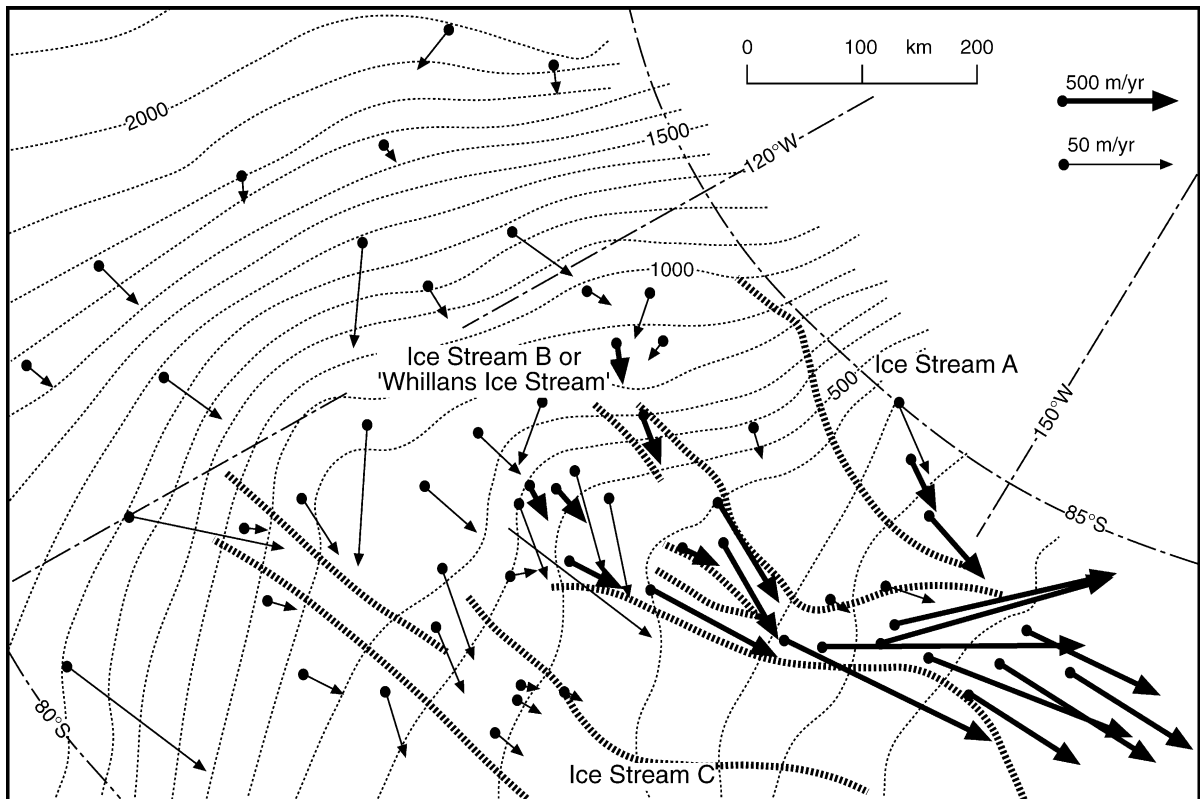
Fig. 7. Velocity data for the Whillans Ice Stream and Ice Stream C (after: Whillans et al., 2001).

shear margins of the ice stream (Duckfoot, Fig. 2; Jacobel et al., 2000).

There is also evidence for earlier and more substantial reorganisation of the Siple Ice Streams. Jacobel et al. (1996) reported the presence of a relict ice-stream trunk on the northeastern flank of the Siple Dome, referred to as the Siple Ice Stream (Fig. 2). This relict stream is cross-cut by several contemporary streams, indicating a major reorganisation of the drainage system in the recent past; estimates place the shutdown of the Siple Ice Stream at about 420 years (Gades et al., 2000; Jacobel et al., 2000).

Having reviewed the morphology, dynamics and temporal behaviour of the Siple Ice Streams we need now to consider the mechanisms that give these characteristics. Central to this problem is our understanding of conditions at the bed of an ice stream, since the rapid ice motion is concentrated at the bed.

Changes in the drainage organisation of the Siple Ice Stream, decadal variations in their velocity, and ultimately their potential to stagnate all have major implications for the mass balance of this sector of the West Antarctic Ice Sheet. Early estimates of mass balance suggested a net deficit of -20.9 ± 13.7 Gt/year with ice discharge, through the Siple Ice Streams, exceeding the rate of accumulation by the order of 25% (Shabtaie and Bentley, 1987). Most of this deficit was attributed to the hyper-activity of the Whillans Ice Stream (Whillans et al., 2001). More recent observations using more precise ice flow measurements based on synthetic aperture radar suggest a positive balance ($+26.8$ Gt/year), with rapid ice thickening over the stagnant Ice Stream C (Joughin and Tulaczyk, 2002). Significantly, this new work suggests that, contrary to previous estimates, the Whillans Ice Stream may have a balanced mass budget as a consequence of its continued slowdown. Whether this slowdown is part of a decadal fluctuation in velocity or part of a terminal decline is not clear, and consequently, the long-term mass balance implications are uncertain.

5. Basal boundary conditions and flow mechanisms for the Siple Ice Streams

5.1. Observations and mechanisms

The Siple Ice Streams represent a major problem in glacial mechanics: how do ice streams flow at rates of up to 900 m/year, despite their low surface slopes and correspondingly low (≤ 20 kPa) driving stresses? Initial ideas focused on either enhanced basal- and/ or marginal-ice shear due to stress-induced recrystallisation (Hughes, 1977), or enhanced basal sliding associated with elevated basal water pressures (Rose, 1979). In the mid-1980s, however, the discovery of 5- to 6-m-thick layer of soft sediment beneath the Whillans Ice Stream shifted the focus towards the concept of subglacial deformation. The idea is that saturated sediment beneath a glacier can deform at a lower shear stress than ice, and that its deformation may contribute anything up to 90% of a glacier's forward motion (Boulton and Hindmarsh, 1987). This discovery has not only had a significant impact on our understanding of ice stream flow but on glaciology as a whole.

The sediment layer was discovered using seismic reflection studies (Blankenship et al., 1986, 1987), which provided evidence for a highly porous and water saturated sediment layer beneath the Whillans Ice Stream. This sediment layer was interpreted as a subglacial till by Alley et al. (1986, 1987a), and the high levels of porosity were considered to be consistent with a dilated, and therefore actively deforming sediment. Further seismic data from the Whillans Ice Stream (Rooney et al., 1987) suggests that the layer is continuous over a distance of at least 8.3 km and that it is also present at a series of widely spaced (300 km) points along the ice stream. These data suggest that the till layer is more or less continuous beneath the ice stream. However, Atre and Bentley (1993) presented evidence to suggest that there may be some lateral variation in the properties and thickness of this subglacial till layer.

The presence of this subglacial till layer was confirmed by samples recovered by Engelhardt et al. (1990) using boreholes drilled through the Whillans Ice Stream. Sedimentary analysis of these, and subsequent samples, by Tulaczyk et al. (1998) has shown that the sediment consists of a clay-rich diamict containing marine diatoms, probably derived by the deformation of underlying marine and glacimarine sediments, of Tertiary age, similarly to those found beneath the Ross Sea. The degree of deformational mixing within these diamicts is, however, uneven, a fact which has been used to argue against it deepseated deformation (Tulaczyk et al., 2001).

The boreholes used to extract these samples have also provided information about the subglacial hydrology of the Siple Ice Streams, and hydrological observations have now been made beneath the Whillans Ice Stream and Ice Streams C and D (Engelhardt and Kamb, 1997; Kamb, 2001). This work suggests that the basal water flow system beneath all three ice streams examined is able to supply a substantial quantity of water at a pressure approximately equal to the weight of the overlying ice, thereby reducing basal friction. The borehole pressure records provide a glimpse of a complex water system that undergoes rapid local variations. For example, the pressure records in boreholes 500 m apart, beneath the Whillans Ice Stream, could not be correlated, whereas those just 25 m apart could (Engelhardt and Kamb, 1997). None of the fluctuations in basal water pressure observed seem to correlate with changes in ice velocity, as one might expect on more conventional glaciers (cf. Willis, 1995). This suggests that ice flow is insensitive to the high frequency—spatial and temporal—variation in water pressure present within the system and is probably controlled by some form of space and time integrated water pressure value (Engelhardt and Kamb, 1997). The detailed structure of the drainage system beneath the ice streams is not clear. However, modelling of borehole records suggests that the basal water system may consist of a continuous or discontinuous sheet of water forming a gap (ca. 2.5 mm thick) between the bottom of the ice and the top of the till (Kamb, 2001). Whether or not this distributed drainage system is associated with larger channels, such as the broad, shallow 'canals' which have been widely hypothesised for deforming beds (Walder and Fowler, 1994), is uncertain (Engelhardt and Kamb, 1997; Kamb, 2001). Surprisingly, the data obtained so far suggest that the hydraulic systems of the active and inactive Siple Ice Streams are very similar (Kamb, 2001).

The recognition of deformable sediment beneath the Siple Ice Streams linked the idea of subglacial

deformation firmly with rapid ice flow. This association has, however, been questioned in recent years by the direct observation of deformation beneath the Whillans Ice Stream and Ice Stream D (Engelhardt and Kamb, 1998; Kamb, 2001). A stake tethered to an instrument within a borehole was inserted into the subglacial sediments. As the stake was dragged, down-flow information was obtained about the vertical distribution of motion beneath the ice base. These observations provide conflicting data. Subglacial deformation accounts for 80% to 90% of the motion in Ice Stream D (30 to 60 cm below the ice base). In contrast, however, only 25% of the motion in the Whillans Ice Stream appeared to be due to subglacial deformation and basal sliding, or at least near surface deformation (within 3 cm of the ice base), accounted for most of the forward motion (Kamb, 2001). These data give an enigmatic glimpse into the spatial complexity of subglacial processes which appear to operate beneath an ice stream and have re-kindled interest in basal sliding as a mechanism of fast ice flow.

Interpreting these complex and somewhat contradictory observations is difficult. Part of the problem lies in the level to which point-specific observations, extracted at great logistical cost, can be extrapolated to infer the basal boundary conditions for an ice stream as a whole. For example, are the results of Engelhardt and Kamb (1998) indicative of widespread basal sliding beneath the Whillans Ice Stream, or is it simply a localised phenomena? The sensitivity of ice streams to different length scales of variation, in both space and time, is also unclear, but knowledge of this is essential to ensure that sampling scales are appropriate and that the extrapolations are valid. Answering such questions is challenging, and is currently a matter of opinion, and consequently, researchers have naturally tended to rely on the evidence which best fits their personal conceptualisation of the processes operating beneath an ice stream.

The key to all these processes, however, is our understanding of basal hydrology. Several researchers have drawn attention to its importance in determining the partitioning of flow between basal sliding and subglacial deformation and its temporal/spatial variation (e.g. Alley, 1989a,b; Iverson et al., 1998; Boulton et al., 2001b). Using this as a starting point, it is possible to develop a conceptual, and somewhat speculative, framework in which rapid spatial and temporal variation in basal water pressure are used to produce a model of basal motion that embraces both basal sliding and sediment deformation at various depths. Subglacial water pressure not only controls sediment strength, since enhanced pore-water pressures reduce the internal friction and cohesion of a sediment, but along with bed roughness, it controls the degree of coupling between the ice base and the bed. If we ignore bed roughness and simply consider water pressure, then it is possible to envisage a slip-stick cycle of behaviour associated with variation in water pressure (Fig. 8; Boulton et al., 2001b). As basal water pressure builds, it may float the ice from the bed leading to rapid sliding, with minimal bed deformation. However, as water pressure falls, the glacier sole will begin to couple to its bed and sliding will be concentrated in the sediment horizon of the bed with the greatest pore-water pressure relative to the sediment's strength. This is likely to be near the ice–sediment interface since sediment strength increases as effective pressure (overburden) increases with depth. As a result, deformation will tend to occur in near surface horizons, but the location of maximum deformation may vary through time as water drains vertically through the sediment pile. If water pressure falls below that necessary for deformation, then the sediment will start to consolidate and the glacier sole will become tightly coupled to the bed. During this 'stick mode', ice flow may only occur via creep, causing basal shear stress to accumulate until it is released by the next increase in basal water pressure.

If water pressure fluctuates over a large range, then all stages of the slip-stick cycle may operate. Alternatively, if the water pressure is maintained at any of these levels, or over a limited range, then the cycle may pause or be restricted to one mode, or a limited range of modes, within the cycle (Fig. 8). Equally, if one is dealing with a subglacial hydraulic system which is highly variable in space, as appears to be the case beneath the Siple Ice Streams (Engelhardt and Kamb, 1997), then one area of the bed may be sliding while an adjacent one is in a stick mode. When such variable behaviour is integrated over a large area of an ice stream's bed, a more uniform picture of ice flow may result and the local variation in basal process (and rate) becomes unrecognisable. The number of points or areas of an ice stream in each mode may control the average flow rate; for example, if a
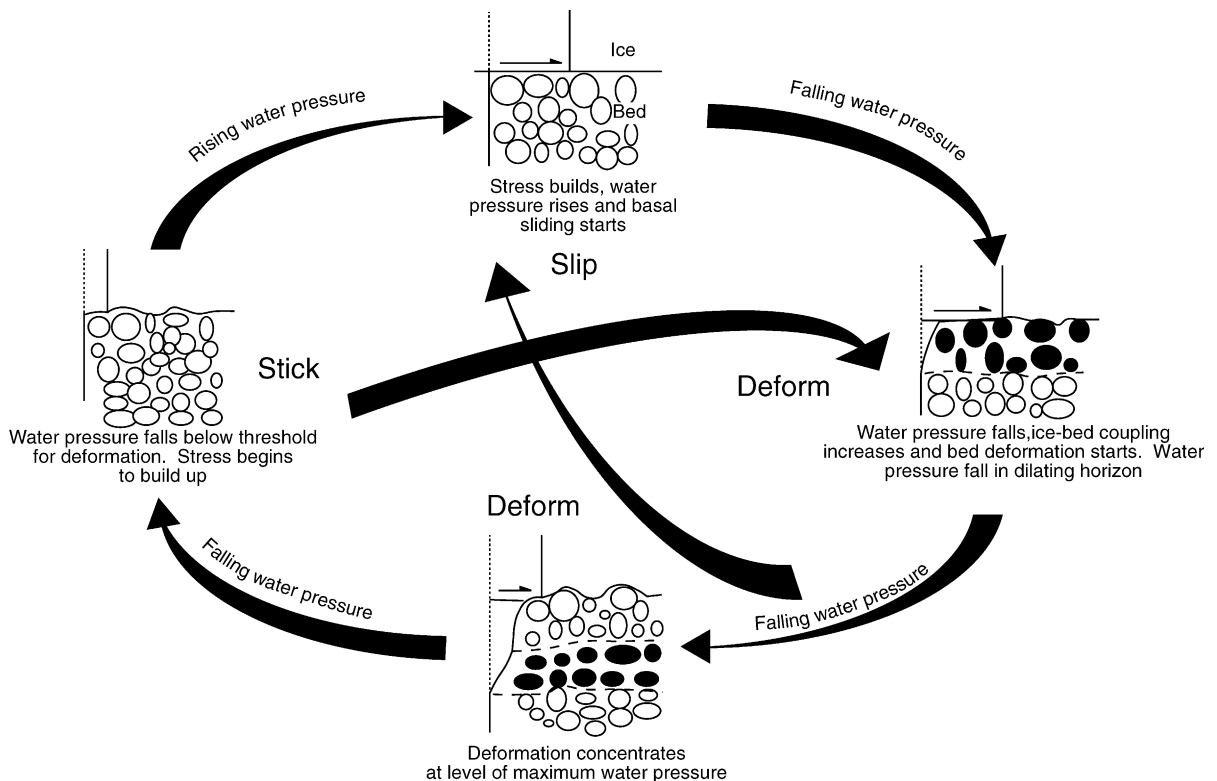
Fig. 8. Stick-slip cycle of basal motion, showing the importance of subglacial water pressure (after: Boulton et al., 2001b).

high proportion of the bed is in stick mode, then the rate of flow may be negligible. According to Anandakrishnan and Alley (1997b), the glacial driving stress of Ice Stream C is supported by numerous sticky spots, the failure of which accounts for the greater abundance of micro-earthquakes when compared to the active Whillans Ice Stream. Beneath Ice Stream C, the basal water pressure may be similar to that under the Whillans Ice Stream at a specific site, but on average, there may be a higher percentage of the bed in a stick mode, when compared to the Whillans Ice Stream. Consequently, ice-stream flow may be controlled by some form of statistical statement about the proportion of the bed in any given flow mode over a period of time. Such a statistical statement is not going to be immediately apparent from the point-specific observations of basal hydrology made to date. It is possible that, given the limited nature of the borehole observations to date, much of the observed complexity within the system may simply be accounted for by interpreting the results

as snapshot samples of a complex system. The visible results of this system—ice stream flow and character—are the time and space integrated products of this complex system. It is perhaps worth noting that this type of conceptual framework may work for the fine-grained clay-rich sediments beneath the Siple Ice Streams since they provide few clast-roughness elements with which to couple the bed to the basal ice. However, a glacier resting on a coarse-grained till may couple to its bed more effectively reducing the potential for de-coupled sliding and increasing the importance of bed deformation (Alley, 2000; Boulton et al., 2001b).

5.2. The theory of subglacial deformation as a key to understanding ice streams?

Ever since the discovery of a deformable till layer beneath the Whillans Ice Stream, our understanding of the mechanics of ice flow within an ice stream has become linked with the concept of subglacial defor-

mation, irrespective of recent suggestions that basal sliding may also be important. Our ability to model ice streams is linked to the development of flow laws that adequately describe the process of subglacial deformation. This has proven controversial in recent years with rival formulations being developed and applied to ice streams. Our understanding of ice streams has, therefore, become caught-up in this wider debate about the nature and mechanics of subglacial deformation. As a consequence, we need to explore this debate as part of our quest to understand ice streams.

Subaerial sediments such as those found on a hillside or in a glacier forefield tend to fail in a plastic fashion and their behaviour can be explained by classic soil mechanics. These sediments have a finite strength known as the yield strength above which they fail, and the rate of deformation (strain rate) is independent of the stress applied (Fig. 9). Sediments that behave in a plastic fashion cannot support a stress greater than this yield stress and failure tends to occur along a discrete horizon or shear plane. Early models of subglacial deformation presented by Boulton (1979) and Fisher et al. (1985) used the idea of plastic behaviour. These models, however, do not contain statements about the time-dependent behaviour of sediment when coupled to a glacier. In effect, these models do not consider the variables that controls the rate of deformation, something which is essential if

Fig. 9. Schematic graph showing the relationship between strain rate and stress, and the difference between a viscous and a plastic rheology.

one wants to model the patterns of erosion and deposition that result from subglacial deformation or model an ice stream, for example.

This problem can be overcome if one models till as a viscous fluid, in which the rate of deformation increases with the applied stress (Fig. 9; Boulton and Hindmarsh, 1987; Alley et al., 1987b, 1989; Hindmarsh, 1997). This type of model has been used effectively to predict the large-scale patterns of erosion and sedimentation induced by subglacial deformation within ice sheets and to explain a range of geological observations (e.g. Hart et al., 1990; Hart and Boulton, 1991; Boulton, 1996a,b). In the context of ice streams, Alley et al. (1987b,1989) developed a model of the Whillans Ice Stream based on a viscous formulation for the basal sediment. In this model, the gravitational shear stress of the ice stream is balanced by the strength of the till integrated over a wide area in which variations in rheology, including more rigid sticky spots, occur. In this model, the contribution of the lateral shear margins of the ice stream in supporting the down-slope weight of the ice (gravitational shear stress) was considered to be minimal. Consequently, ice-stream velocity is modelled as a function of the gravitational shear stress, and the thickness and viscosity of the till.

This model of the Whillans Ice Stream has been challenged by laboratory observations on material recovered from beneath the ice stream which indicate a behaviour more consistent with a plastic rather than a viscous flow law. The yield strength of the sediment is only 1 or 2 kPa, compared to the 20 to 30 kPa imposed by the down-slope weight of the ice (Kamb, 1991; Tulaczyk et al., 2000a). This implies that the bed is unable to support the gravitational shear stress of the ice stream. This challenge to the viscous till model is supported by observations made using subglacial sediment from other glaciers which also indicate a plastic rather than a viscous rheology when examined in the laboratory (e.g. Iverson et al., 1995, 1998; Hooke et al., 1997; Murray, 1997). As a result, two broad schools of thought have emerged in recent years. One school fully embraces a plastic description of subglacial sediments (Tulaczyk et al., 2000a,b), while the other maintains that, although tills may deform in a plastic fashion at a small-scale, their gross behaviour at larger-scales is still approximated by some form of viscous flow law (Hindmarsh, 1997;
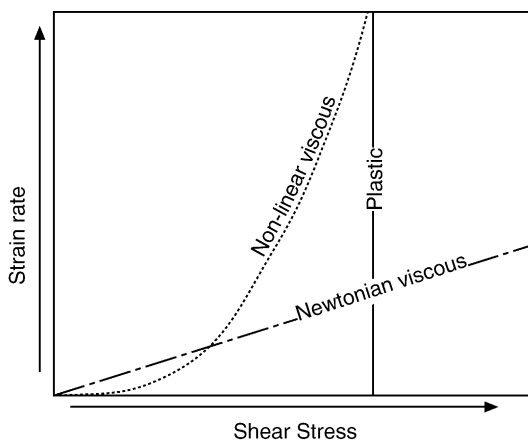
Alley, 2000). We shall consider each of these schools in turn.

Tulaczyk and others have developed a plastic model of till deformation and used it to explore the flow characteristics and stability of ice streams (Tulaczyk, 1999; Tulaczyk et al., 2000a,b, 2001). Conceptually, this approach places ice streams somewhere between two-end members. At one end of the spectrum, we have grounded ice masses which are firmly coupled to their beds such that the gravitational driving stress is equal to the basal shear stress. At the other, we have floating ice shelves where basal stresses are zero and the gravitational driving stress is supported by marginal and longitudinal stresses. According to Tulaczyk (1999), ice streams fall between these two-end members since they do not float in water but they do move over subglacial sediment so weak that it cannot support the gravitational driving stress. The weaker the subglacial sediment, the faster the ice-stream motion will be, because a greater fraction of the driving stress must be supported by ice deformation in the shear margins (Tulaczyk et al., 2001). Tulaczyk et al. (2000a,b) developed a model of plastic subglacial deformation. In this model, the strength of the sediment and its void ratio (porosity) are determined by the effective normal stress at the base of the ice stream (ice weight minus basal water pressure). Changes in effective normal stress result in changes in the void ratio of the till and therefore its strength. The effective normal stress is controlled not only by ice thickness, but also by basal water pressure. This is dependent on the rate of basal melting and is a function of both the geothermal heat flux, and the frictional heating caused by sediment deformation. Tulaczyk et al. (2000b) argued that till strength beneath an ice stream is therefore thermally controlled and they identified two feedback loops that lead to two possible stable states of motion. The first feedback occurs where basal melting due to shear heating within the sediment exceeds the rate of water drainage through the sediment. In this scenario, there is a build-up of stored water in the sediment, and consequently a fall in it strength, which in turn reduces the rate of shear heating within the sediment because there is less resistance to deformation. Over time, the strength of the sediment, and therefore the shear heating, will be reduced such that no change in water storage occurs in the sediment, that is the melting rate is equal to the rate of drainage. This feedback could maintain a fast flow mode (Tulaczyk et al., 2000b). The alternative feedback occurs, when the rate of melting is either less than the rate of drainage or is negative due to the freezing-on of water to the ice base. This will cause an increase in till strength, which in turn will increase the shear heating within the sediment, leading to an increased melt rate. If the melt rate is, however, strongly negative, then till strengthening, via water loss, may ultimately prevent subglacial deformation before the melt rate can be increased sufficiently to compensate for the deficit. This describes a second stable flow state in which ice can only move slowly via the internal deformation of the ice (sheet flow). Tulaczyk et al. (2000a) argued that switching between fast and slow modes may therefore be sensitive to a thermal trigger. As an ice stream thins due to its own flow efficiency, colder ice will be moved closer to the bed reducing the melt rate, thereby strengthening the subglacial till layer and perhaps ultimately triggering a shutdown of the ice stream.

Before we turn to the rival school of thought, it is worth drawing attention to one of the key implications of plastic flow models, namely that shear is restricted to a very shallow layer at the top of the till column. This is consistent with the observations made by Engelhardt and Kamb (1998) beneath the Whillans Ice Stream, where the majority of deformation occurred within the top 3 cm of the subglacial sediment layer.

Despite the field observations that suggest that laboratory samples of subglacial sediment fail in a plastic fashion, a number of attempts have been made to reconcile this with a large-scale viscous flow model. Perhaps the most elegant of these ideas is that proposed by Hindmarsh (1997). He acknowledges, as most researchers now do, that at a small-scale till fails by plastic failure, and that during such events, the rate of deformation is independent of the stress regime. However, he suggests that the net integration of multiple small-scale plastic failures is best approximated by a viscous flow law. Hindmarsh (1997) drew an analogy with ice, which on an atomic-scale behaves in a manner that is similar to plastic deformation, crystals deform by the migration of dislocations through the lattice (i.e. individual plastic failures). However, the net effect of multiple dislocations at a larger-scale is

the nonlinear viscous-type behaviour, which is an accepted characteristic of ice flow as defined by Glen's Flow Law. Hindmarsh (1997) argued that the same is true of subglacial till in which point-specific subglacial experiments and laboratory tests describe individual dislocations, the sum of which produces a viscous-type behaviour. Essentially he reconciles the two alternative models by placing them at different ends of a scale spectrum (Hindmarsh, 1997). Hindmarsh (1998a,b,c) also argued that the key to any model of subglacial deformation is its ability to explain the observed geological imprint produced when a glacier flows over soft deformable sediment. For example, in the wider context of glacial geology, such a model must be able to explain landforms produced by subglacial deformation such as drumlins (Boulton and Hindmarsh, 1987; Hindmarsh, 1998a,b), and the large-scale patterns of erosion and sedimentation within an ice sheets (Boulton, 1996a,b). Viscous-type flow models can explain such phenomena. More specifically, in the context of ice streams, Hindmarsh (1998c) argues that a model of subglacial deformation should be able to explain the meso-scale surface characteristics of an ice stream, such as the 'mottles', 'warps' and 'flow stripes' described from their surface (e.g. Merry and Whillans, 1993). He demonstrates how a viscous flow law linked to a model of subglacial hydrology can begin to explain such surface texture. As he points out the fact that a viscous flow approximation can produce these phenomena does not mean that it does, but his models do provide the potential to predict the behaviour of surface forms on ice streams which can be tested by satellite observations.

It is worth emphasising that viscous-type flow models predict the deformation of subglacial sediment over a much greater depth range than the plastic model which is consistent with geological observations (Hart et al., 1990), although not with the observations of Engelhardt and Kamb (1998) beneath the Whillans Ice Stream.

What is clear from the above discussion is that there is currently no consensus as to the type of flow law which best approximates subglacial sediment flow. This limits our ability to model ice streams and the two-end members remain one that assumes a plastic flow law and one based on a viscous approximation.

5.3. The search for a switch

The temporal behaviour of the ice streams, in particular, the shutdown of Ice Stream C some 150 years ago has interesting implications for the mechanisms of ice-stream motion, not least since any mechanical model should be able to predict such instability (e.g. Tulaczyk et al., 2000a). Ice Stream C is underlain by a layer of unfrozen soft sediment (Kamb, 2001) and its driving stress is similar to that of the Whillans Ice Stream, consequently, the contrasting behaviour of these two neighbouring ice streams must be due to some mechanical difference at the bed. Models of ice-stream shutdown fall into two broad categories: (1) those that view it as a product of the rapid ice discharge within the stream and therefore part of an ice stream's life cycle; and (2) those that see it as a consequence of the interaction between two adjacent streams. Within this framework, a range of hypothesises have been proposed to explain the shutdown of Ice Stream C as recently reviewed by Anandakrishnan et al. (2001), and include:

1. *Surging*: a small proportion of valley glaciers experience cyclic flow, in which episodes of fast flow are punctuated by periods of quiescence. It has been suggested that ice streams exhibit such behaviour (Hughes, 1975; Rose, 1979). One possible mechanism for this is a basal water feedback, in which ice flow accelerates as basal water pressure rises within a distributed drainage network at the glacier bed. However, as water pressure continues to rise in the distributed drainage system, it becomes unstable and collapses to a more localised, channel-based system, thereby reducing basal water pressure and rapid basal motion (Retzlaff and Bentley, 1993). Insufficient is currently known about the nature of the drainage system beneath the ice streams to evaluate this model and there is now some debate about the nature of this instability mechanism (Anandakrishnan et al., 2001).
2. *Loss of lubricating till*: the loss of a lubricating basal sediment layer, via basal erosion within the ice stream, could lead to a slow-down (Retzlaff and Bentley, 1993). However, the evidence points to a widespread soft-sediment layer beneath Ice Stream C (Atre and Bentley, 1993; Anandak-

rishnan and Bentley, 1993; Anandakrishnan and Alley, 1997a).

3. *Ice-shelf back-stress*: one of the consequences of subglacial deformation is the movement of subglacial till which is deposited at the down-stream end of an ice stream close to its grounding-line (Alley, 1989a,b). Thomas et al. (1988) suggested that deposition at the mouth of Ice Stream C may have increased ice grounding and back-stress, thereby stopping stream flow. However, Anandakrishnan and Alley (1997b) have showed that the grounding-line of Ice Stream C provides little restraint to ice flow making this idea unlikely.

4. *Ice piracy*: in this model, the Whillans Ice Stream captured the accumulation area of Ice Stream C thereby draining the ice which previously sustained flow in Ice Stream C. This type of mechanism has been observed in numerical models, in which ice streams self-organise as a result of mutual competition for ice (Payne and Dongelmans, 1997). There is, however, no ice flow evidence to support this idea and in fact, Joughin et al. (1999) have suggested that the reverse may in fact be true.

5. *Water piracy*: in this model, the Whillans Ice Stream captures the subglacial water that previously lubricated the bed of Ice Stream C, causing the till to consolidate and cease to deform in a ductile fashion (Anandakrishnan and Bentley, 1993; Anandakrishnan and Alley, 1997a,b). Subglacial water flow is driven primary by the surface gradient of the ice, which is approximately 10 times more effective than basal topography in driving water flow (Paterson, 1994). The idea is that the topography up-stream from the trunk of Ice Stream C was drawn-down and flattened in response to the inland migration of the ice stream. This reduction in surface slope caused a reorganisation of subglacial water flow; instead of being directed by ice slope down the trunk of Ice Stream C, basal topography increased in importance as the ice thinned, driving the water towards the adjacent Whillans Ice Stream. This reduced the amount of available basal water under Ice Stream C leading to an increase in the number of sticky spots and a loss of rapid flow, something which would be first felt at the grounding-line and would then migrate upstream (Anandakrishnan and Alley, 1997a,b).

6. *Thermal processes*: several investigations have illustrated the sensitivity of basal motion within an ice stream to thermal conditions (MacAyeal, 1993b; Payne, 1995; Tulaczyk et al., 2000b). Rapid ice flow within an ice stream causes viscous heating, and may lead to a warm bed, which in turn accelerates ice flow. However, rapid ice flow in an ice stream will ultimately cause ice thinning which may reduce the gravitational shear stress and bring cold ice closer to the bed, causing it to cool. This may in turn cause a shutdown in rapid basal motion. Tulaczyk et al. (2000a) illustrated the potential sensitivity of till strength to basal melt rates. The efficiency of ice discharge within an ice stream may therefore ultimately limit the active part of its life cycle.

Of these explanations, the water piracy model has gained greatest attention and is considered by some to offer the most likely explanation for the shutdown of Ice Stream C (Anandakrishnan et al., 2001). In essence, it combines two ideas, that an ice stream's flow efficiency may lead to its own end, and secondly that the lack of topographic control of the Siple Ice Streams may lead to their instability. The dominance of the water piracy model has been challenged by recent observations of Price et al. (2001) who suggest that the present day ice surface topography is a product and consequence of stagnation in the trunk of Ice Stream C, rather than its cause. Using velocity data for the up-stream regions of both the Whillans Ice Stream and Ice Streams C, they conclude that only the trunk of Ice Stream C stagnated, while its tributaries have remained active. The consequence of this is a build-up of ice at the confluence between the active tributaries and the stagnant ice-stream trunk, leading to a bulge in the ice surface and modification of the surface contours in the critical region between the two ice streams. According to Price et al. (2001), it is these contours that are used by Anandakrishnan and Alley (1997a) to justify the water piracy model. The evolution of the ice surface topography in the upper reaches of Ice Stream C is therefore crucial and the subject of a 'chicken and egg' type debate at present (Price et al., 2001; Anandakrishnan et al., 2001).

In truth, the cause of the shutdown of Ice Stream C is not known. Models tend to polarise between those which see it as a consequence of the interaction

between two adjacent and competing ice streams, and those who see it as a consequence of some internal instability within an ice stream itself, for example, as a result of its own flow efficiency. If the latter is true, then the other ice streams may show the same potential to shutdown as is perhaps indicated by the slowdown in the Whillans Ice Stream (Joughin and Tulaczyk, 2002).

5.4. Geology as an ice-stream template

There is evidence to suggest that some combination of saturated mobile sediment and a thin lubricating water layer is necessary to the initiation of rapid basal motion and consequently the formation of ice streams. Whether this occurs via subglacial deformation, via basal sliding over a thin water/sediment layer, or by some combination is as we have seen a matter of debate (Alley, 1989a,b; Engelhardt and Kamb, 1998). In both cases, however, there is strong reason to suspect a geological or substrate control on the initiation of rapid basal motion (Blankenship et al., 2001). If one favours a model of subglacial deformation, then the availability of deformable sediment becomes critical since erosion of a hard-rock substratum is unlikely to generate a sustainable sediment layer (Alley, 2000). One would expect therefore a strong correlation between the onset of rapid basal motion and the tectonically controlled boundary of a sedimentary basin, or the extent of a sediment drape.

The availability of basal water is also important for sediment deformation and critical if one favours a model of subglacial sliding. Consequently, spatial variation in the geothermal heat flux may be of relevance (Blankenship et al., 2001) and one might expect some correlation between areas of enhanced heat flow and the onset of rapid basal motion within ice streams (Blankenship et al., 1993). Enhanced heat flow will tend to be focussed in areas of subglacial volcanic activity or in regions composed of a mosaic of crustal blocks that vary in thickness and thermal history.

These ideas have been explored using aerogeophysical data designed to image the geological structure of the ice bed along the Siple Coast (Blankenship et al., 2001). The Ross Embayment is located above the West Antarctic Rift System. This rift system is bordered to the south by the Transantarctic Mountains

and to the north by Marie Byrd Land and consists of a succession of fault-bound basins which trend parallel to the ice streams of the Siple Coast (Fig. 1). The origin of this rift system has recently been attributed to the Cretaceous fragmentation of Gondwana (Dalziel and Lawver, 2001). The fault-bound basins within this rift system appear to have been infilled and draped by marine and glaciomarine sediments of mid-Cenozoic age and interbedded in some locations with a range of volcanic deposits (Blankenship et al., 2001). Bell et al. (1998) used gravity and magnetic data to define a sediment-filled basin incised by a steep-sided valley that appears to mark the onset of ice streaming in the upper reaches of the Whillans Ice Stream. Similarly, Anandakrishnan et al. (1998) presented seismic evidence to show that one of the lateral margins of the same ice-stream tributary lies above the boundary of a sediment-filled basin (Fig. 10). This work was amplified by Studinger et al. (2001) and Blankenship et al. (2001), both of whom present more regional evidence for the coincidence of ice streaming with the outcrop pattern of marine sediments and sediment-filled rift basins. These data suggest a strong coincidence in some, but not all, cases between the onset of rapid basal motion and the outcrop of soft-sediment. A word of caution is required when viewing these data, however, since the coincidence between basinal troughs, sediment and fast flow does not necessarily imply a causal link. Fast ice flow would tend to occur within basinal troughs irrespective of the presence of a layer of deformable sediment as discussed earlier. There are also several examples of the availability of sediment at the bed but the absence of ice streaming (Blankenship et al., 2001), which suggests the presence of additional controls.

Blankenship et al. (2001) also noted the coincidence between the initiation of several ice-stream tributaries and crustal zones with an enhanced geothermal heat flux, perhaps twice the continental average. These data clearly suggest that geology provides a spatial template for ice streaming, and that the sedimentological history of a region prior to glaciation is critical in determining the likelihood of streaming (Alley, 2000).

The idea that a suitable geological template is essential for ice streaming has been challenged by work on numerical models that have succeeded in simulating ice-stream flow characteristics over a uni-
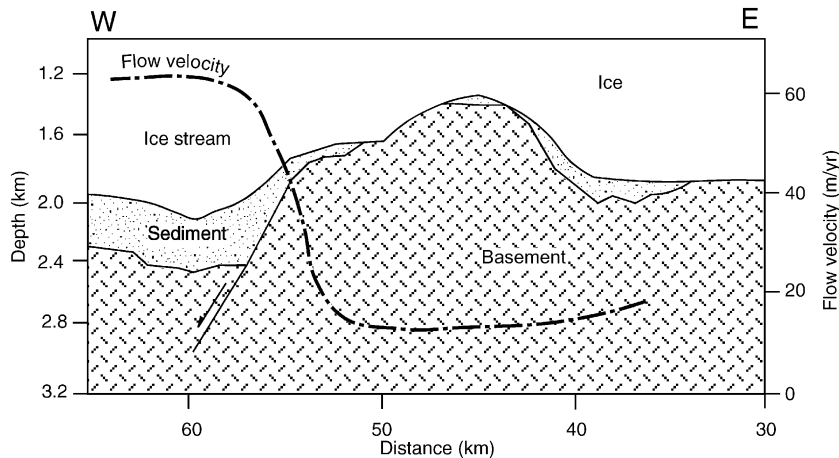
Fig. 10. Cross-section through a lateral margin of an ice-stream tributary. Note the coincidence of the ice-stream margin with the edge of the sedimentary basin (after: Anandakrishnan et al., 1998).

form substrate (e.g. Payne, 1995; Payne and Dongelmans, 1997; Payne and Baldwin, 1999; Hulbe and Payne, 2001). Numerical models of this sort emphasise the thermomechanics of ice streams and focus on the relationship between heat generation and ice flow. The viscosity of ice decreases with temperature, increasing the rate of ice deformation that in turn generates more heat leading to faster ice flow; a feedback known as creep instability (Clarke et al., 1977). In addition, if one introduces sliding, then the initiation of sliding starts a similar feedback loop via the friction generated at the bed. It is feedback loops such as these that give rise to topographically controlled ice streams, since thick ice is more likely to be warm and wet at the bed.

Payne (1995) used a simple two-dimensional ice sheet model to explore this coupling between flow and temperature. Within this model, he successfully simulated episodes of fast flow, with a self-limiting cycle of activity. Initially, the build-up of heat, due to ice flow, establishes a zone of warm marginal ice that flows faster than the inland ice. This causes a drawdown of the ice surface topography, with a maximum slope at the junction between cold and warm ice which helps accelerate flow at this point causing the thermal boundary to migrate inland forming a corridor of warm, fast ice (an ice stream). Flow within the ice stream accelerates since faster flow generates more heat, that in turn encourages faster flow, and the ice begins to thin. As the ice thins, the surface gradient

declines causing a decrease in basal shear stress and therefore glacier velocity. In addition, cold ice in the upper portions of the ice sheet is drawn down towards the bed within the ice stream. As a result, heat is more effectively conducted away from the bed causing it to cool and freeze, thereby facilitating the shutdown of the ice stream. The ice stream is therefore self-limiting and a period of ice accumulation is required before streaming can re-start.

Payne and Dongelmans (1997) have shown that this type of model can generate ice streams within an ice sheet resting on a flat bed, small flow perturbations are amplified and the ice streams self-organises by drainage capture until a set of equally spaced streams results. Payne and Baldwin (1999) applied this type of model to the Fennoscandinavian ice sheet of the last glacial cycle. On the basis of geological evidence, this ice sheet appears to have been characterised by a number of ice streams that traversed the hard-rocks of the Baltic Shield and consequently shows little relationship to the outcrop of deformable sediment (Punkari, 1995; Kleman et al., 1997; Boulton et al., 2001a). The ice streams could be explained by thermomechanical feedbacks triggered by the basal topography (Payne and Baldwin, 1999). The real issue, however, is whether this type of model can be applied to the Siple Ice Streams, where the level of topographic control is not pronounced. In order to answer this question, Payne (1998, 1999) has attempted to model the Siple Ice Streams. His model was able to

generate a series of ice streams in this area, triggered but not constrained, by the bedrock topography, which resemble the ice streams present today, although the flow velocities are substantially less. The modelled ice streams appear to go through internally generated cycles of growth and stagnation and the variability is concentrated on the Whillans Ice Streams and on Ice Stream C due to their mutual competition for ice. Clearly, this type of model suggests that ice streams are not simply restricted to areas of suitable substrate but are generated as natural modes of ice flow within an ice sheet. This is not to say that the rate of flow, or the detailed location of an ice stream may not be constrained by the substrate, but it does imply that the substrate alone may not be sufficient to generate an ice stream. This might help explain why in some cases, the correspondence between the outcrop geology and the pattern of streaming is not always clear (Blankenship et al., 2001). It is also worth noting that although these models can generate ice streams on a flat bed, basal topography is essentially the key trigger for the thermal feedbacks on which these models depend (Hulton and Mineter, 2000).

In summary, there is some observational evidence to suggest that geology might act as a template for ice streaming, but equally there is evidence to suggest that the coupling between flow and heat within an ice sheet is of equal importance. Caution is therefore required in asserting that ice streams can only occur where there is a suitable geological template.

6. The significance of ice streams

So far in this review, we have documented the morphology, dynamics and temporal behaviour of the Siple Ice Streams and reviewed the mechanisms which may account for these characteristics. We now need to apply this knowledge of contemporary ice streams in order to explore their role as a dynamic element in past, present and future glacial systems. In essence, we need to address a simple question: what is the significance and potential impact of ice streams on the Earth's dynamic system? We can view this question in several ways. Firstly, ice streams have an important role in the delivery of sediment to marine basins, since they concentrate the discharge of ice, the debris within it, and the subglacial sediment beneath

it. Elverhøi et al. (1998) have suggested that the sediment flux from ice streams is comparable to that of a large fluvial system, such as the Mississippi Delta. The location of ice streams within Pleistocene ice sheets may therefore be important to our understanding of the sedimentary architecture of large depositional basins. Since sediment has a higher potential for preservation in such basins than elsewhere, much of the geological record of ice on Earth is dominated by marine sedimentary basins (Eyles, 1993). Understanding ice streams may therefore have relevance to the interpretation of such basins and consequently our understanding of the Earth's early (pre-Cenozoic) glacial and climate history. In addition, the accumulation of sediment at the mouth of an ice stream, in what is termed a trough mouth fan, may pose a significant geohazard. At the Last Glacial Maximum, ice sheets advanced to the edge of the continental shelf on both sides of the North Atlantic discharging sediment via ice streams into large trough mouth fans at the edge of the continental shelf. If these relict fans were to collapse today, then a tsunami, or tidal wave, would result with the potential to impact on the densely populated regions which now surround the North Atlantic.

Secondly, ice streams clearly influence the geometry of the ice sheets in which they are located and consequently any reconstruction of a palaeo-ice sheet and its dynamics needs to be able to predict and model the location of any component ice streams (Stokes and Clark, 2001). Consequently, understanding the likely occurrence and the controls on their dynamics is important in the reconstruction of ancient ice sheets and the interpretation of the landform record associated with them.

Perhaps the greatest significance, however, of ice streams is their potential to impact on the stability of an ice sheet and in turn on the Earth's dynamic system as a whole. The data from the Siple Ice Streams reviewed above give a picture of an ice stream as a dynamic flow element. Ice streams not only discharge large volumes of ice, but shows a potential for instability through variations in flow velocity, migration and reorganisation of drainage pathways over time, and a potential to switch on and off. Like the arteries in our body, they are vital to the wellbeing of some ice sheets, and small changes may have a major impact on their vitality. They represent a potential for

internally driven change, independent of external forcing. This has been explored in relation to two issues, the interpretation of Heinrich layers in the marine sediments of the North Atlantic and the potential instability of the West Antarctic Ice Sheet. Each of these issues is examined briefly.

Heinrich layers were first observed in the marine sediments of the North Atlantic and consist of concentrated ice-rafted debris deposited during short (<100 year) events that occurred at irregular intervals between 60 and 10 thousand years ago (Heinrich, 1988). Each Heinrich layer was deposited during a period when the surface waters of the North Atlantic were exceptionally cold and fresh (Bond et al., 1992) and involved the discharge of large armadas of icebergs (Dowdeswell et al., 1995). The production of such a large volume of icebergs must have lead to at least the partial collapse of the ice sheet from which

they where derived. Provenance studies have suggested that much of this ice-rafted material was derived from Hudson Bay (Andrews and Tedesco, 1992; Andrews, 1998). Comparison of Heinrich events with the temperature record obtained from oxygen isotopes in ice cores taken from the Greenland Ice Sheet suggests that they are associated with episodes of pronounced cooling (Alley, 1998). The temperature records are dominated by two related oscillations, a low-frequency Bond Cycle and a higher-frequency Dansgaard–Oeschger Cycle (Fig. 11). Dansgaard–Oeschger Cycles show rapid jumps of temperature, with cold, dry and windy atmospheric conditions occurring around the North Atlantic when the ocean's surface water were fresh and cold (Bond et al., 1993). Progressive cooling through several Dansgaard–Oeschger Cycles defines a Bond Cycle. Bond Cycles are associated with a period of ice sheet
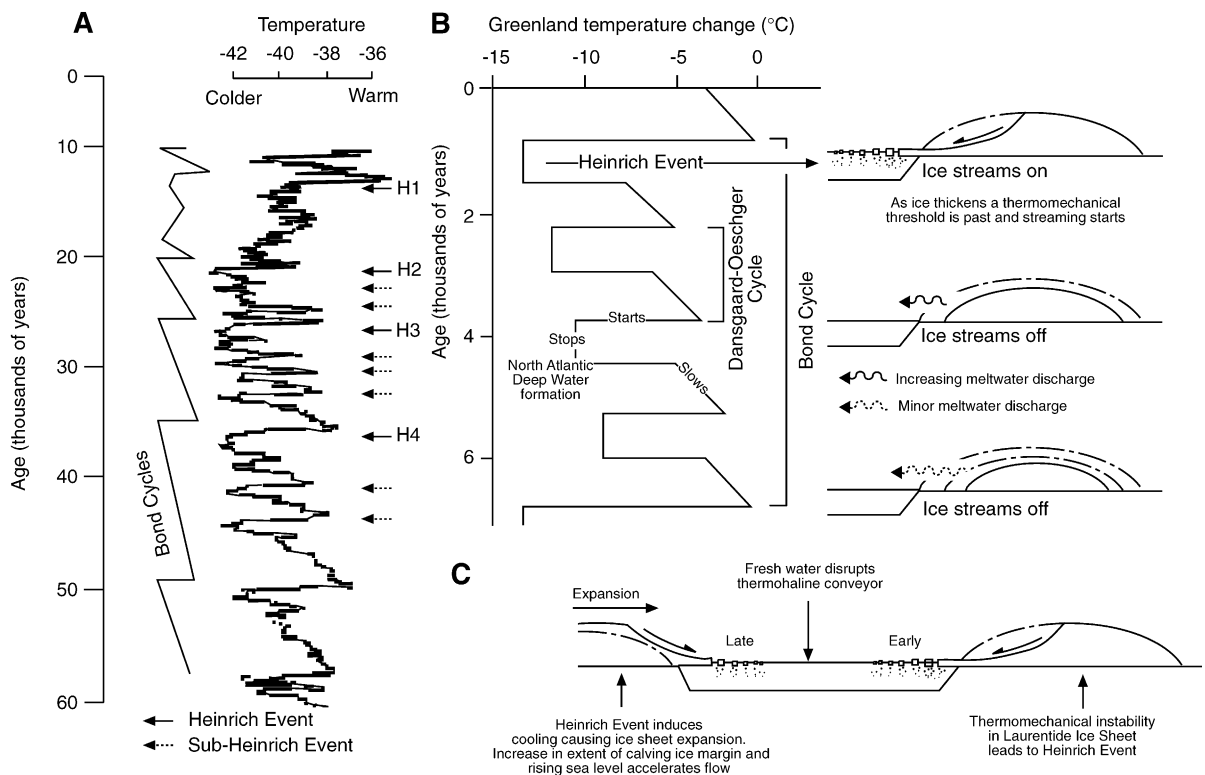

Fig. 11. (A) Temperature cycles (Dansgaard–Oeschger and Bond Cycles) within an ice core from the Greenland Ice Sheet (after: Bond et al., 1993). (B) Simple model of the relationship of Dansgaard–Oeschger Cycles, Bond Cycles, the formation of North Atlantic Deep Water, Heinrich events and possible oscillations in the activity of ice streams (after: Alley, 1998; Andrews, 1998). (C) Schematic illustration of how a collapse of the Laurentide Ice Sheet could precipitate increased ice rafting from other ice sheets (after: Andrews, 1998).

growth that is terminated by a Heinrich event and a period of intense cold. Smaller sub-Heinrich events also occur at the termination of some of the Dansgaard–Oeschger Cycles. Each Bond Cycle is terminated by a rapid rise in air temperature, and ice sheet retreat.

Dansgaard–Oeschger Cycles have been linked to formation of North Atlantic Deep Water (Broecker, 1994a,b). In today's interglacial, cold and salty water of the North Atlantic is dense enough to sink in the Nordic Sea towards the ocean floor, from where it flows south. This downward flux of water is balanced by a surface flow of warm fresh water from the tropics—the North Atlantic Drift—which keeps the climate of the North Atlantic equable. This thermohaline circulation is part of a larger conveyor that moves water and heat around both the Atlantic and Pacific oceans (Broecker et al., 1985, 1990; Broecker and Denton, 1990a,b). In the North Atlantic, the production of deep water is sensitive to the salinity of surface water and consequently inputs of fresh water may hinder its formation causing the thermohaline circulation, and therefore the North Atlantic Drift, to slow or stall, which will cool the North Atlantic region. Dansgaard–Oeschger Cycles have been explained in terms of oscillations in the efficiency of the North Atlantic Drift caused by variation in meltwater run-off from adjacent land areas (Broecker et al., 1990).

The larger Bond Cycles in the Greenland ice cores have been explained in terms of down-wind cooling caused by the expansion of the Laurentide Ice Sheet in the vicinity of Hudson Bay (Alley, 1998). Each of these expansions of the ice sheet during a Bond Cycle was terminated by an episode of ice sheet collapse during which large armadas of icebergs were released into the North Atlantic. The meltout of these icebergs deposited a Heinrich layer and provided freshwater which caused the thermohaline circulation, and the associated North Atlantic Drift, to stop initiating an intense cold spell. Having collapsed, the ice sheet would have little maritime margin left and therefore the discharge of icebergs and meltwater into the Atlantic would be reduced causing its salinity to rise, thereby restarting the thermohaline conveyor and the North Atlantic Drift (Broecker, 1994a; Broecker et al., 1992). The expansion and collapse of the Laurentide Ice Sheet in the vicinity of Hudson Bay may be due either to an external climate forcing or alternatively

due to an internal instability. With an external forcing, the ice sheet gradually grows and expands over the continental shelf and as the proportion of maritime margin increases, it becomes susceptible to rapid calving and collapse. In this scenario, the oscillation is driven by ice sheet expansion into a finite space, limited by the deep water at the edge of the continental shelf. The alternative is that the ice sheet's oscillation is driven by an internal trigger that causes it to collapse rather than by its continued expansion.

The search for an internal trigger to explain the growth and collapse of the Laurentide Ice Sheet in the vicinity of Hudson Bay has focussed on the role of ice streams switching on and off (e.g. MacAyeal, 1993a,b; Alley and MacAyeal, 1994; Marshall and Clarke, 1997). The Hudson Bay sector of the Laurentide Ice Sheet would have been a marine-based ice sheet like the West Antarctic Ice Sheet today. MacAyeal (1993a,b) first put forward a binge-purge model in which slow accumulation of ice within Hudson Bay, over a frozen bed of marine sediment, builds up during periods of quiescence and ice sheet expansion. As ice thickness increases, heat would be trapped at the bed, which begins to thaw leading to the formation of ice streams. Rapid ice streaming in the Bay, whose marine sediments would act as a geological template for the ice streams, leads to the rapid discharge of ice and the down draw of the ice sheet in this region. As the ice thins, the bed would freeze and streaming. In essence, it is the internal instability of the ice sheet, as manifest by ice streams, which drives the climate oscillation associated with a Heinrich event (Bond Cycle).

This model has been questioned on several grounds; firstly, evidence has emerged to suggest that the debris within some of the Heinrich layers is not just sourced in the Hudson Bay region, indicating that several ice sheets were involved (Groussett et al., 1993; Bond and Lotti, 1995; Rahman, 1995). Secondly, fluctuations in ice volume during and after a Heinrich event appear to have occurred not just around the North Atlantic, but also in the Southern Hemisphere (Lowell et al., 1995; Denton et al., 1999). This implies that either the events in the North Atlantic were sufficient to drive global change, or that ice bodies globally were responding to some form of external forcing. Modelling of the Hudson Bay ice streams by Marshall and Clarke (1997) has supported the idea of a thermomechanical trigger for ice streaming in this sector of the Laurentide Ice Sheet.

This work suggests, however, that the rate of ice discharge would not be sufficient to give the sedimentation rates inferred for Heinrich layers (Dowdeswell et al., 1995), and that the impact on sea level of this melting ice would have been much smaller than that observed. This clearly suggests that several ice sheets must have collapsed to give each Heinrich layer. Andrews (1998) demonstrates how the collapse of one ice sheet may cause other ice sheets with marine margins to collapse (Fig. 11C). McCabe and Clark (1998) show that ice sheets such as that in Britain may have expanded as a consequence of the cooling episode induced by a Heinrich event causing an increase in ice calving and debris supply to North Atlantic. This implies that Heinrich layers may be diachronous. Perturbation in the global thermohaline circulation in the North Atlantic could be sufficient to impact on the system as a whole and therefore explain the global pattern of events (Alley, 1998). The alternative model is that an external forcing caused global expansion of ice, and that as ice sheets advanced to the edge of their continental shelves, they began to calve rapidly drawing down inland ice, perhaps via ice streams, leading to ice retreat.

The jury is still out on whether the Heinrich events are the result of an internal instability within the ice sheet system, as manifest by the turning on and off of ice streams, or is driven by some external (global) control induced perhaps by an orbital forcing of solar radiation. This work, however, clearly illustrates the potential of internal instability within an ice sheet to switch streaming on and off, and to impact on the Earth's dynamic system as a whole.

One of the reasons that the Siple Ice Streams have been investigated so intensely is due to the fact that they are the most dynamic element within the West Antarctic Ice Sheet. This ice sheet is considered by many to be potentially unstable (e.g. Mercer, 1968; Weertman, 1974; MacAyeal, 1992; Oppenheimer, 1998). Much of the ice sheet's bed rests below sea level and consequently a retreat of the grounding-line could accelerate its flow leading to its collapse. In human terms, this would not be good news, leading to a rise in global sea level of between 5 and 6 m (Drewry, 1983). Critical, of course to the impact of such a sea-level rise is not only the magnitude but also the rate of the rise. Bentley (1998) suggests that a rapid rise in sea level would be perhaps five times (10

mm/year) that which occurred during the 20th century (2 mm/year). This is equivalent to an addition of 3000 km^3 of water to the oceans and would require an increase in flow velocity within the West Antarctic Ice Sheet of 10 to 20 times (e.g. 10 to 20 km/year in the Siple Ice Streams). It is perhaps interesting to note that this is about the ice volume required to deposit a Heinrich Layer (Bentley, 1998) and that the drainage area of the Ross Embayment is similar in size to the Hudson Bay sector of the Laurentide Ice Sheet. Despite strong spatial variations, the Ross Embayment as a whole has a mass balance which is close to zero, that is the accumulation is matched by the rate of ice flow (Bentley and Giovinetto, 1991). Recent estimates suggest that it has recently shifted into a positive balance (Joughin and Tulaczyk, 2002). For the system to collapse, the rate of flow would need to accelerate beyond the rate that snow could accumulate, which is currently not the case. It is difficult to see how the velocity fields of the Siple Coast could be increased sufficiently to cause the ice sheet to collapse (Bentley, 1998). In fact, despite a large level of internal instability, the system as a whole appears stable. The Ross Ice Shelf contains a record of ice-stream behaviour which is at least 1000 years old, since it takes that time for the ice discharged by the ice streams to reach the shelf edge (Fahnestock et al., 2000). Any major episode of instability should be recorded in its flow structure. Despite variation in the location and discharge from different ice streams, the system appears to have remained stable without any sign of a major reorganisation (Bentley, 1998). There is, therefore, no evidence for a level of unstable behaviour within the Siple Ice Streams that could cause the ice sheet to collapse dramatically in the near future, irrespective of anthropogenic global warming.

This does not mean, however, that the ice sheet will not ultimately disappear, or at least reduce in size, as it appears to have done at times during the Pleistocene (Scherer et al., 1998). Conway et al. (1999) have investigated the retreat of the grounding-line within the Ross Embayment and have concluded that it has been retreating at a rate of 120 m/year since the early Holocene (ca. 7500 years BP). If it continues to retreat at the present rate, then complete deglaciation would be achieved in a further 7000 years. Although this would cause a rise in global sea level, it would not occur any more rapidly than sea level has been rising

in recent years, unless the rate of ice flow accelerated dramatically. Retreat during the last 1000 years does not appear to have increased the rate of flow so why should it in the future? Recession of the grounding-line and decay of the ice sheet has been underway since the early Holocene and shows no acceleration due to anthropogenic warming or recent sea-level rise. In other words, the future of the ice sheet was pre-determined when retreat of the grounding-line was triggered in the early Holocene, probably by a rapid rise in sea level associated with the melting of mid-latitude ice sheets in the Northern Hemisphere. Reces-sion to the Siple Coast has continued in the absence of further forcing. Modelling studies have shown that the size of this response is out of proportion with the change in sea level (Le Meur and Hindmarsh, 2001; Hindmarsh and Le Meur, 2001). The grounding-line should have retreated as little as a few 10 s to a 100 km, as is the case for the East Antarctic Ice Sheet. The fact that retreat of the West Antarctic Ice Sheet is much greater is puzzling and must be due to some character-istic of the ice sheet and the system of ice flow within it. Recent modelling studies have tentatively suggested that the spatial and temporal instability of flow asso-ciated with ice streams may drive the retreat of the grounding-line (Hindmarsh and Le Meur, 2001). These conclusions are tentative, but might suggest that the instability of marine ice sheets is induced by ice streams, not through their flow efficiency, but more due to the spatial and temporal instability that they induce within the system.

This retreat of the grounding-line along the Siple Coast could be halted by the apparent shift towards a positive mass balance for the Siple Ice Streams, and Joughin and Tulaczyk (2002) suggest that this shift may signal the end of the Holocene deglaciation of the West Antarctic Ice Sheet. This may not necessarily be good news, however, since changes in the volume of ice discharged by the Siple Ice Streams may impact on the size and extent of the Ross Ice Shelf into which they drain. The positive mass balance of the Siple Coast reported by Joughin and Tulaczyk (2002) is a consequence of the shutdown of Ice Stream C and the continued slow-down of the Whillans Ice Stream. It is not known whether the slow-down of the Whillans Ice Stream is merely part of a decadal- or century-scale fluctuation in ice velocity, but if it is maintained, or even leads to the stream's stagnation, then the supply

of ice to the Ross Ice Shelf will fall. As the shelf continues to flow seaward, it will thin and perhaps begin to break up. This process may be facilitated by the effects of future climate warming which appears to have helped destabilise some of the smaller and thinner ice shelves along the Antarctic Peninsula (Vaughan and Doake, 1996). The collapse of the Ross Ice Shelf would expose over 400,000 km^2 of new shallow continental sea impacting on both the atmos-pheric and glacial systems of region. More impor-tantly, the growth of sea ice, and the associated brine exclusion, would turn the shelf into a producer of Antarctic Bottom Water. Denton (2000) has argued that an increase in the production of Antarctic Bottom Water could significantly change the thermohaline circulation. He argued that the thermohaline circula-tion has two stable states. Firstly, an interglacial mode in which North Atlantic Bottom Water sinks within the Nordic Sea, allowing warm surface water to penetrate much of the North Atlantic. Secondly, a glacial mode in which North Atlantic Bottom Water sinks south of Iceland, limiting the northwards pene-tration of the warm surface water. According to Denton (2000), one of the triggers for the onset of a glacial thermohaline circulation is an increase in the production of Antarctic Bottom Water. Loss of ice shelves around Antarctica and the associated increase in bottom water production could therefore plunge the North Atlantic region into the next glacial cycle. This event could, therefore, be linked to the behaviour of the Siple Ice Streams.

7. Conclusions

The aim of this review has been to provide an introduction and point of entry into the substantial body of literature that has built up about the Siple Ice Streams in Antarctica. I have examined their morphol-ogy, dynamics and temporal behaviour as well as considered the mechanics and basal conditions that might give rise to these corridors of fast ice flow. Ice streams are clearly exciting, dynamic elements within the glacial system, with the potential to impact on ice sheet stability and in turn on the Earth's ocean and climate system. They provide a challenging problem in glacial mechanics, but on the basis of the data reviewed here, one is left wondering whether they are

really a threat to ice sheet stability? The view of ice streams as unstable flow elements is based on the fact that ice velocity within them is an order of magnitude greater than conventional ice flow, and that streaming is an unusual or anomalous behaviour. This impression of instability has been reinforced by the apparent role of ice streams in the repeated collapse of the Laurentide Ice Sheet in the Late Pleistocene (Mac-Ayeal, 1993a,b). This view of instability is not necessarily consistent, however, with the data reviewed here from the Siple Coast, which gives an impression of a system which may be internally unstable, but which is collectively more robust. The Siple Ice Streams vary in width, length and flow velocity over time and have a complex dynamic history. At least one ice stream (Ice Stream C) has undergone a recent episode of stagnation, either as result of the mutual competition for water between two adjacent ice streams, or as result of some form of self-limiting cycle. There is also evidence for major reorganisation of the flow structure in the past, with at least one relict ice stream (Siple Ice Stream). The mass balance of individual ice streams is extremely variable, but collectively, they appear to be broadly in balance (Bentley, 1998). In fact, recent evidence suggest that accumulation may in fact exceed ice discharge due to the shutdown of Ice Stream C and the continued slowing of the Whillans Ice Stream (Joughin and Tulaczyk, 2002). The data therefore, present a picture of a system which is internally variable, and perhaps unstable, but which appears as a whole to exhibit stable characteristics. If the system as a whole was anything other than stable, then its instability should be manifested within the ice of the Ross Ice Shelf, which contains a thousand year record of ice-stream discharge (Fahnestock et al., 2000). In addition, recent work using interferometric satellite data to map ice flow velocity in Antarctica has shown that stream flow is more common than previously assumed, with a complex drainage system of tributaries feeding ice streams (Joughin et al., 1999; Bamber et al., 2000).

These observations give a picture of a flow system that is collectively stable, but individually variable. This does not preclude the possibility that the ice-stream activity will lead to the collapse of the West Antarctic Ice Sheet, but suggests that it is unlikely to occur as a consequence of the ice stream's own flow efficiency as previously thought. It may however occur as a result of the internal variability within the ice-stream system, which has been tentatively suggested as a possible cause of grounding-line retreat (Hindmarsh and Le Meur, 2001). Despite similarities in the size and geometry between Hudson Bay and the Ross Embayment, the Siple Ice Streams may not be good analogues for the unstable ice streams which caused the repeated collapse of the Laurentide Ice Sheet in the Late Pleistocene (Bentley, 1998; Alley and Bindschadler, 2001). For the Siple Ice Streams to behave in a similar way, one would need to increase flow velocity by a factor of 10 to 20 (Bentley, 1998). As a result, one is left wondering what environmental controls might operate, or what threshold must be past, for a system of ice streams to become sufficiently unstable to cause the collapse of an ice sheet?

Irrespective of the stability or instability of the Siple Ice Streams, ice streams remain as one of the most dynamic elements within the glacial system and challenge us to explain their formation and dynamic characteristics. In meeting this challenge, we may be able to predict or at least hypothesise where the thresholds to instability might lie, if they exist at all. Our understanding of the mechanisms of fast flow within ice streams has enhanced our knowledge of subglacial processes in general and ice streams have become caught-up in the wider debate about the nature and significance of subglacial deformation. I would suggest that one of the most fundamental research problems that needs to be addressed is to improve our understanding of the length scales of variability in basal condition to which ice streams respond as manifest not only in their macro-, but also meso-scale form. This is central to our ability to extrapolate and make sense of the point-specific results obtained from the bed of ice streams. Understanding the Siple Ice Streams does not simply require more subglacial observations, although this would help, but we also need to develop a greater understanding of the statistical variability in conditions both in space and time beneath an ice stream and of the sensitivities to them. One step in this direction is to improve our ability to model subglacial processes beneath an ice stream and to use these models to examine the sensitivity of a modelled ice stream to different scales of perturbation in basal conditions. Hindmarsh (1998c) has suggested that we need to develop critical tests, which can be answered easily from satellite observation, to evaluate

the alternative models of subglacial processes beneath an ice stream. For example, we should be able to predict from models of subglacial deformation, or sliding, the formation and behaviour of meso-scale topography on an ice-stream's surface (e.g. 'flow stripes', 'lumps' and 'warps') and to test these predictions against reality using satellite observations. If we can refine our ability to model ice streams, then we can formulate sampling strategies with which to obtain further basal data from the bed of an ice stream. It is also worth noting that the bed of Pleistocene ice streams (Canals et al., 2000; Stokes and Clark, 2001) may provide a complimentary, and more accessible, source of systematic information about the statistical variability in sedimentary properties over the bed of an ice stream.

In conclusion, if we are to understand the mechanics of ice streams, we need to improve our ability to model subglacial processes and in turn improve our data sets about not only basal conditions, but also about the surface topography and temporal behaviour of ice streams. The challenge is to map out the parameter space in which ice streams are stable and to identify the thresholds that might lead to instability of the system as a whole. It is a challenge that is fit for the 21st century.

Acknowledgements

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