Antarctic ice streams: a review

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Abstract:

The stability of the West Antarctic Ice Sheet is influenced by the stability of the fast flowing ice streams that provide an outlet for much of the ice. These ice streams flow at speeds of up to 800m per year, two orders of magnitude faster than the surrounding ice sheet. Early research on ice streams developed detailed maps of their location, velocities and thickness. The cause of ice stream onset has been debated throughout the literature, and is still not properly understood, although current thought suggests that the change from bedrock to sedimentary basins coincides with the ice stream onset zone. Our understanding of ice stream mechanics changed dramatically in the late 1980’s when it was discovered that beneath Ice Stream B lay a saturated, deforming till layer. Subsequent research concentrated on the mechanisms for ice stream movement and the distribution of stresses within the stream. During the 1990’s research looked to find the major controls on ice stream movement, anaysling apparent ‘sticky spots’ and the role of margins. Evidence that Ice Steam C stagnated 150 years ago has sparked more recent work efforts to comprehend the overall stability of ice streams and their ability to start and stop, and to migrate at their boundaries. The methods used throughout ice stream research are considered and avenues for future research are presented.
**Introduction**

The stability of the West Antarctic Ice Sheet is strongly dependent on the stability of ice streams that are the main outlet of the ice sheet (Bentley and Clarke, 1987; Alley, 1990). The WAIS is traversed by about a dozen ice streams approximately 50km wide and 500km long, in which the ice is moving at speeds up to ~800m per year, in sharp contrast to the motions of the ~10m year in the general mass of the ice sheet outside the ice streams (Bentley and Clarke, 1987). To understand their stability, it is necessary to understand how forces are partitioned within the ice stream, their mechanisms of movement, and what controls their location and velocity. Glaciologists, physicists, mathematicians, and modellers, among others, have been investigating ice streams for over 50 years. This review brings together the knowledge that has been gained from their research, and reviews the way in which our thinking about ice streams has changed over time.

Firstly I define ice streams and describe why it is that they are important to study. I then provide an overview of the ice streams in Antarctica, their flows, surface features and what is known about their sub-glacial topography. I then present the knowledge of the onset of ice streams, followed by a review of the continually evolving debate about the mechanisms of ice stream movement. The hypotheses surrounding the stagnation of Ice Stream C are discussed critically, followed by a description of the changes and stability of ice streams more generally. The final sections review the methods that have been used in ice stream studies, and present the temporal changes in ice stream research.

**Definition and importance of ice streams**

Swithinbank (1954) first defined ice streams as a “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.” Therefore, an ice stream is bordered by ice and not rock, and is part of the inland ice sheet, therefore not floating. Bentley and Clarke (1987) state that the distinction between an ice stream and an outlet glacier is clear in principle, but muddy in practice, and that many Antarctic glaciers are intermediate in character. The Rutford Ice Stream, for example, is bordered with the Ellsworth Mountains on one side and the ice sheet on the other. Ice streams differ from the surrounding ice sheet in that the thick and slow-moving interior ice reservoir is generally fixed to the underlying bedrock, while the ice streams glide over lubricated beds at velocities of up to several hundred metres per year. One reason why it is important that we understand ice streams, is that the mass balance and stability of the West Antarctic Ice Sheet are, in large part, determined by the dynamics of the large ice streams that drain it. Some even go so far as to state that the evolution of the West Antarctic Ice Sheet drainage system almost certainly governs the process of ice sheet collapse (Bell et al., 1998).

**Location, flow, and surface features**

Radar sounding over West Antarctica in the late 1960s revealed the existence of the ‘Ross Ice Streams’ (Robin et al., 1970). Based on this data and on surface elevations
obtained during the International Geophysical Year West Antarctic traverses, Hughes (1973) prepared a sketch map of the West Antarctic ice streams (see Figure 1). Following airborne radar sounding in the 1970’s, Rose (1979) produced a much more detailed map of the area. The ice streams located along the Siple Coast, simply named ice streams A-F, have received the greatest amount of attention by researchers, as they are the major drainage outlets for West Antarctic ice. These ‘Ross Ice Streams’ have low surface elevation profiles, and their bed slopes are low and smooth. Generally, their low driving stresses increase continually inland to the heads of the streams.

Figure 1: Map of West Antarctica identifying the major West Antarctic ice streams presently active (labelled A-U). The letters are located at the ice dome - ice shelf junction. Dotted areas on the ice dome side indicate the concave surge basins of the ice streams. Dotted areas on the ice shelf side identify the bedrock troughs eroded by the surging ice streams when they retreated from the continental shelf margin (Hughes, 1977).
Figure 2: Map of the Siple Coast ice streams and their flow bands in the grid western part of the Ross Ice Shelf (Shabtaie and Bentley, 1987). Velocity vectors are shown, and a medium dark line indicates the ice shelf – ice sheet boundary (the grounding line).

It can be seen from Figure 2 that Ice Stream B has two tributary branches, named B1 and B2. The boundary between them exists as a suture zone for 250km downstream from the place where they merge (Shabtaie and Bentley, 1987). Ice Stream B reaches velocities of 827m per year (Whillans et al., 1987). Ice Stream C, in a nearby location, flows at only 5m per year (Whillans et al., 1987), however there is evidence that around 250 years ago, like Ice Stream B, it was fast flowing (Shabtaie and Bentley, 1987). The age of Ice Stream C’s halt is estimated from the depth of burial of crevasses, and is discussed later.

A similar contrasted pairing also occurs between the active Rutford Ice Stream (400m per year) and its slow moving companion occupying the Carlson Inlet drainage basin (Doake et al., 1987). Ice Stream B, according to the calculations of Shabtaie et al. (1987) and Whillans et al. (1987) is flowing at a rate that cannot be sustained. To stabilise itself, it must eventually reduce its flow rate by 50%, or if this is impossible, it much switch off for a while. Ice stream C is the opposite situation, in that it could maintain a flow of approximately 600m per year, but is nearly stagnant (Clarke, 1987).

Hindmarsh (1998) states that ice stream textures are important as they arise from the same mechanisms which are believed to control the large-scale flow of ice streams.
The surface of active ice streams A and B are heavily crevassed, and their boundaries are marked by chaotic, incoherent bands of surface crevasses, but there are no visible crevasses on the surface of Ice Stream C (Shabtaie and Bentley, 1987). Hulbe (1997) examined the surface texture of ice streams, identifying particular surface forms that move upstream, explaining this in terms of variations within the ice fabric. Troughs and ridges have also been identified on ice streams A and B (Shabtaie et al., 1987).

**Sub-glacial topography**

Ice stream dynamics and therefore the response of the WAIS to climate change, are strongly modulated by the underlying geology (Bell et al., 1998). All of the ice streams except for the Ross Ice Streams show deep sub-glacial valleys. Bentley and Clarke (1987) stated that the ice streams characterised by deeply incised beds are rather like true outlet glaciers despite their lack of visible confining walls. The Ross Ice Streams, however, are different. They are broader in comparison to their depth, show little transverse sub-glacial topography, and in some cases, such as the north boundary of Ice Stream D, do not even extend laterally to the edge of what little sub-glacial valley does exist. Drewry (1983) suggests that the valleys beneath ice streams result from erosion by the fast moving ice above and infers that the absence of deep sub-glacial troughs beneath the Ross Ice Streams is an indication of their youthfulness. Another possible explanation for the shallowness of the troughs is that erosion by a deforming sub-glacial till (as occurs at the Ross Ice Streams) may be much slower than erosion by sliding (Alley et al., 1987a).

Blankenship et al. (1986) reported through seismic reflection data the indication of longitudinal grooves or ridges in the substrate beneath the layer of till, causing variations in thickness of till across Ice Stream B. Also beneath the ice streams are water (Shabtaie et al., 1987), and under Ice Stream C there appears to be trapped water associated with the slopes of surface terraces. The presence of water and its importance for movement is discussed later.

**Ice stream onset**

A transition exists from slow inland flow where internal deformation and basal sliding are characteristic, to a mode of fast streaming flow in ice streams where basal processes dominate; this is termed the onset of an ice stream (Bentley and Clarke, 1987). The scale over which this transition in ice dynamics takes place remains unknown, although many studies have attempted to map the zone (Bindschadler et al., 2000). It is important to understand what causes the onset of an ice stream, as if the onset were to migrate inland, acceleration and thinning of reservoir ice would reduce ice-sheet volume, raising sea level (Bindschadler, 1998).

McIntyre (1985) found that in many cases, the ice stream head coincides with a bedrock step. Inland migration of the sheet-stream boundary from this pinning point was therefore deemed unlikely. Others state evidence to the contrary. Shabtaie et al. (1987) and Bentley and Clarke (1987) believe that the transition from ice sheet flow to ice stream flow need not be controlled by the bedrock topography and for this reason, ice stream boundaries could be mobile, rather than fixed. This apparent contradiction
is a result of variation between ice streams, in other words, while bedrock topography appears to be an important factor in controlling the onset of some ice streams, in others it is not important. Shabtaie and Bentley (1988) found that there is no step in either the bed topography or the surface topography at ice streams A, B or C.

More recent work by Bell et al. (1998) describes the onset region to Ice Stream B coinciding with a sediment filled basin incised by a steep sided valley. Near the head of adjacent Ice Stream C, seismic data and GPS surveying illustrate that ice stream flow occurs over a sedimentary basin with very close spatial coincidence of the ice stream margin and basin margin (Bindschadler, 1998). Further down stream, ice streams tend to occupy sub-glacial troughs, but the correlation is not universal. Anderson et al. (2001) also verify the results from previous over ice and aerial geophysical surveys, which indicate that ice stream onset coincides with the boundary between crystalline bedrock and sedimentary strata.

In support of these findings, it has been reasoned that a soft sedimentary bed is required to cause ice streaming because of its intrinsic low frictional resistance to flow, and owing to its high erodibility so as to generate till that can deform and lubricate ice motion (Anandakrishnan et al., 1998). Anandakrishnan et al. conclude that the configuration of ice streams and projections about the future of the West Antarctic Ice Sheet can be understood only in the context of sub-glacial geology and, in particular, the distribution, thickness, and composition of sedimentary basins, all of which are poorly known.

In terms of the ice flow at the ice stream onset, Bindschadler et al. (1987) and Shabtaie et al. (1987) present evidence that suggests that Ice Stream B contains within it large blocks of rheologically distinct ice that forms local topographic high or ‘rafts’ moving with the ice stream. They suggest that the sheet to stream transition is a ragged one involving the entrainment of entire blocks of inland ice by the active ice stream.

**Mechanisms of ice stream movement**

Initially it was thought that the mechanisms of ice movement were ice superplasticity (Hughes, 1977), and basal sliding (Rose, 1979). This early work was based on the premise that the sub-glacial bed is rigid and impermeable (Clarke, 1987). The entire thinking of the movement of ice streams changed dramatically when it was discovered that Ice Stream B is underlain by a layer of material that appears to be highly porous, water saturated, and weak enough to deform (Alley et al., 1987a; Alley et al., 1986b; Alley et al., 1987b; Blankenship et al., 1987a; Blankenship et al., 1986a; Blankenship et al., 1986b; Rooney et al., 1987). This finding was a result of the work by Alley et al. (1986b), Blankenship et al. (1987b), and Blankenship et al. (1986b), who used seismic sounding to hypothesise that the rapid motions are due to shear deformation of a layer of water-saturated till underlying the ice.

A series of boreholes drilled through Ice Stream B confirmed the presence that the physical conditions and basal till are suitable for deformation (Engelhardt et al., 1990a; Kamb and Anonymous, 1990). The cores indicated that the layer was at least 2m thick, compared with the seismically inferred thickness of 6.5 m. These findings
generated a great deal of attention from the glaciology community, and during the late 1980’s and early 1990’s a large amount of research was completed to further understand the mechanisms of movement.

The drilling beneath Ice Stream B not only confirmed the presence of a layer of till, but also the presence of water 1000m below an Antarctic ice stream. Engelhardt et al. (1990) described that the deformability of the sub-glacial till is made possible by high pore water pressure, which reduces its strength. At these ice streams the ice is below freezing, except at its base, where the source of the pore water is melting of basal ice, which is due to the frictional heat generated by till deformation, plus the geothermal heat flux conducted upward within the subfreezing ice mass (Alley et al., 1987a). (Kamb, 1991) highlights the basal water pressure feedback mechanism, whereby if the water pressure increases, the till pore pressure increases, the till deformation rate will increase, this will increase the melting rate, which will result in a further increase in basal water pressure, unless the increased pressure causes conduits to enlarge sufficiently. Kamb points out the immediate instability caused by this feedback.

Other research has attempted to determine the relative importance of each of the possible mechanisms of movement. Thorsteinsson and Raymond (2000) confirm that almost all of the velocity is produced by motion at the base. They report that deformation through the thickness of the ice (~1km) is expected to contribute <1% of the speed at the upper surface, and conclude that sliding at the ice-till interface must be the dominant mechanism of basal motion.

‘Sticky spots’ and other controls on ice stream movement

During the 1980’s and early 1990’s much work concentrated on the mechanisms of till deformation and the distribution of stresses in an ice stream. However, it was found that the shear strength of the till beneath Ice Stream B is too low to support the ice stream (Kamb, 1991). This led to a flurry of activity in the early to mid 1990’s on the controls or restrictions to ice stream movement. Alley (1993) first suggested that the velocity of an ice stream is controlled by ‘sticky spots’, where till is either stronger than elsewhere (or absent altogether), or where large bedrock bumps stick into the base of the ice stream, inducing drag. Field study of the surface velocity field of Ice Stream B however, failed to find these sticky spots (Whillans and Jackson, 1993; Whillans and Van der Veen, 1993).

On ice stream E, however, MacAyeal et al. (1995) found that basal friction is dominated by an irregular distribution of basal sticky spots. They found that they cluster in regions where Landsat imagery suggests structural features in the underlying bedrock. They also found that the regions without sticky spots were the same size as the region studied in the two 1993 studies, above, therefore hinting that perhaps the sticky spots of Ice Stream B had been missed. They concluded, however that while significant in the overall force budget, sticky spots are not the dominant resistive stress, and that side and grounding line drag are also important. Alley (1993), who first theorised that presence of sticky spots, also reported that they are probably not dominant in controlling Ice Stream B.
In ice streams, the width-thickness ratio is so much greater than in valley glaciers that it was generally considered that ice stream motion must be controlled mainly by resistive drag at the base, and that the margins were relatively unimportant. An alternative suggestion by Echelmeyer (1994) and Jackson and Kamb (1997) is that the shear margins play an important role in retarding the flow of the ice stream. Jackson and Kamb (1997) used hot water drill ice cores to look at the marginal shear stress of Ice Stream B. They found that 63-100% of the stream’s gravitational loading comes from the margins and much less comes from the base; therefore, margins play an important role in controlling ice stream motion.

The stagnation of Ice Stream C

The flow rates of Ice Stream C are two orders of magnitude slower than the surrounding ice streams B and D. It was suggested that Ice Stream C had previously been more active when Shabaiaa and Bentley (1987) reported that there are no crevasses on the surface of Ice Stream C, however numerous crevasses are buried at a depth of 35 m. They interpreted this to mean that Ice Stream C ceased to be active at that spot 250 years ago. More recent estimates date the stagnation of the main trunk of Ice Stream C to around 150 years ago (Anandakrishnan et al., 2001). The causes of this sudden decrease in flow remains to be a point of controversy in the literature. It is important that we understand the process of ice stream stagnation as this has implications for the overall stability the ice streams, and therefore of the Antarctic ice sheets also.

Alley et al. (1994) hypothesised that ongoing draw down of the ice sheet caused Ice Stream C to migrate into a region that caused a diversion of basal water from the lower parts of Ice Stream C. Anandakrishnan (1996) give evidence to support this hypothesis by measuring a change in seismicity at the basal water diversion zone. They explain that Ice Stream C stagnated in its lower part because of the loss of basal lubricating water. As a consequence, the ice has coupled to localised hard spots on the bed, causing local micro-earthquakes (measured during fieldwork) and slow movement due to friction. They go on to suggest that the loss of water from Ice Stream C has probably increased the water supply to Ice Stream B.

Jacobel et al. (2000) developed these ideas further to suggest that the shutdown of Ice Stream C was not a single event, but a sequence involving stagnation of ice and migration of the ice stream boundary (see Figure 3). Ground based studies confirm the inference from imagery that a series of former shear zones exist, decreasing in age towards the ice-stream centre. Jacobel et al. indicate that possible causes for the stepwise migration of the north margin of Ice Stream C include a gradual decrease in ice flux, a reduction in the available water or hydrostatic pressure in the basal till, or a freezing of the till layers on the northern side.

Anandakrishnan et al. (2001) review and evaluate various hypothesis for the stagnation of Ice Stream C and conclude that the most credible is the “water piracy” hypothesis (Alley et al., 1994). Price et al. (2001) review this finding by analysing the elevated surface topography between the stagnant ice stream and its active upstream tributaries. They conclude that the elevated topography formed because of compression and thickening of the confluence region that occurred subsequent to
trunk stagnation. This conclusion is contrary to the "water-piracy" hypothesis, which requires the observed topography to be in place prior to stagnation. Bentley et al. (1996) also disagree with the "water-piracy" theory, reporting that the cause of its stagnancy is not a lack of water at its bed as the bed is currently everywhere unfrozen and wet.

The causes of the stagnation of Ice Stream C are not yet fully understood, and it can be seen from the above paragraphs that contradicting evidence exists surrounding the "water-piracy" theory. This is an area of ongoing research that needs to be resolved in order to more fully understand the factors controlling the flow of ice streams.

![Figure 3](image.jpg) **Figure 3**: Time sequence showing likely evolution of the north margin of Ice Stream C, as shown in Jacobel et al., 2001.

The stability of ice streams

The ice streams of the Siple Coast have been the most extensively studied, and it has been found that they can undergo significant changes over relatively short time scales of 100 years or less. As is evident from the stagnation of Ice Stream C, discussed above, ice streams appear to be able to switch between slow and fast modes of flow (Clarke, 1987). Also of concern is the apparent ability for ice stream margins to migrate. This is evident in the past with Ice Stream C’s stepwise inward migration. Delineated tracks of ice-stream margins carried onto the Ross Ice Shelf show that large changes in ice-stream widths (and probably discharge) occurred during the last thousand years (Hughes, 1973). Bindschadler (1998) suggested that margins shift in response to changing thermal and hydrological conditions at the base, and that these changes in ice-stream width alter the balance of forces and the rate of ice discharge.

Feildwork has generated evidence that Ice Stream B’s margins are currently widening. Shabtaie and Bentley (1987) first reported this observation, more recently Echelmeyer
and Harrison (1999) quantified that the southern margin of Ice Stream B is moving outward into the inland ice, at a rate of at least 9.7m (± 1.1 m) per annum. This is coupled with a slowing of the ice stream by about 5% over the last 10 years. Furthermore, if this migration can be interpreted as a widening of the ice stream, then the lower speeds may indicate a temporal change in basal drag across the ice stream.

Ice Stream C also appears to be currently varying. Price et al. (2001) report that a tributary to Ice Stream C, C1, has changed flow since the main trunk of Ice Stream C stagnated. They suggest that this is as a result of stagnation induced, rapid thickening in the upstream regions of Ice Stream C, combined with a longer period of thinning in the upstream regions of Ice Stream B. As a result, approximately 50% of the ice from tributary C1 currently flows into Ice Stream B. If current trends in ice thickness change continue, Price suggests that the slope between the two will steepen, and more ice will be diverted from Ice Stream C into Ice Stream B in the future.

In contrast to this evidence for margin mobility, layering of radar-reflectors within the ice indicates that fast streaming motion was prevented at Siple Dome in spite of substantial basal melting and bed lubrication (Nereson, 2000). This suggests Siple Dome, which is bordered by ice streams C and D, is strongly resistant to streaming flow, has never been overridden by ice streams, and is maintained by stable boundaries associated with its sub-glacial geology.

Methods in ice stream research

The development of new technologies have enabled different investigation methods and advanced our understanding of ice streams. Initially many studies set out to map the location and thickness of ice streams (Robin et al., 1970), using techniques such as airborne radar sounding and seismic reflection. Rose (1979) reported that ice streams can be mapped by airborne radar sounding survey, as they are characterised by strong surface clutter. Shabtai and Bentley (1987) showed that these scattered returns on the radargrams can be attributed to crevasses, either shallowly buried or exposed. Furthermore, the lateral boundaries of the ice streams are sharply delineated by large and regular crevasses that denote zones of very strong shear.

Repeat aerial photography has also been used as it is effective in measuring velocities and surface slopes of glaciers in regions which are intensely crevassed (Whillans and Jackson, 1993). Such zones may be mechanically critical to glacier flow but are not safe for ground based work. Very closely spaced velocity and elevation data can be obtained at small unit cost. The difficulty is in obtaining quality photography, and that the photo measurement and analysis program is labour intensive.

The ability to drill through the ice and sample the sediments at the bottom has been tremendously beneficial to ice stream research. This has allowed ground truthing of what had been hypothesised though remotely sensed methods. Alongside the development of drilling techniques have been the improvements of precise measurement tools for such things as pressure and temperature.

Since the early 1990’s, increased satellite cover has meant that satellite imagery is an increasingly popular method for investigating the properties of ice streams
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(Bindschadler and Scambos, 1991; MacAyeal, 1992). Satellite imagery allows precision measurement of surface velocity over regions of the Antarctic ice sheet that are heavily crevassed and therefore inaccessible by ground or air transportation. More recently, satellites have been used with the development of Geographic Positioning Systems (GPS), which allows high precision survey of location and elevation.

Probably the largest development which has changed the way in which we investigate ice stream behaviour is the development of larger and more powerful computers. Since the late 1990’s, modelling to mimic current behaviours with the view to predict future behaviours has become more prevalent. It must be remembered that a model inherently simplifies the real world, and therefore encompasses a number of assumptions. Kamb (1991) quite rightly points out that valid modelling of the ice stream motions can only be done when the mechanism that controls the motion has been securely identified and quantitatively formulated.

One example of recent modelling work is that of Nereson (2000), who modelled the elevation of ice streams margins after stagnation. This work aids in interpreting the age of stagnation of past ice streams, and was successfully applied to the Siple Ice Stream, which was found to have stopped ~ 500 years ago. Another critical aspect of ice streams is their interaction with the ice shelf into which some of them flow, and this has been the subject of much modelling work. Many models (Hughes, 1977; Lingle et al., 1984) have suggested that without the ice shelves, the grounding line of the ice streams would retreat unstably, and possibly result in the complete elimination of inland ice. These models however unrealistically assume that ice movement at the grounding line is controlled solely by ice shelf spreading, and that ice movement of the ice stream immediately inland is controlled by basal shear stress. Van der Veen (1985) introduces a transition zone in the first 150km of the ice stream in which the longitudinal stress linearly decreases from its value on the ice shelf to zero, apparently producing more stable results than previous models.

Summary of ice stream research

The study of ice streams has been a relatively recent phenomenon, with most work being completed since 1970. The ice streams of the Siple coast (the Ross Ice Streams) have been the most thoroughly studied, perhaps at the expense of the other ice streams in Antarctica. A large majority of the work has been completed on Ice Stream B, the fastest moving of the Ross Ice Streams (Alley, 1990; Alley, 1993; Alley et al., 1986a; Blankenship et al., 1987b; Engelhardt et al., 1990b; Hulbe, 1997; Jackson and Kamb, 1997; Kamb, 1991). More recently, Ice Stream C has received a reasonable amount of attention (Anandakrishnan, 1996; Bentley et al., 1996; Jacobel et al., 2000; Nereson, 2000; Price et al., 2001; Shabtaie et al., 1987), and ice streams D and E have generated a moderate amount of work (Bindschadler et al., 2000; Bindschadler and Scambos, 1991; Jacobel et al., 1992; MacAyeal, 1992).

Initially most studies set out to map the location and thickness of ice streams, and describe their surface characteristics (Drewry, 1983; Robin et al., 1970; Rose, 1979). The 1980’s saw a large amount of work completed to investigate the sub-glacial sediments, with the revelation in 1987 that beneath Ice Stream B there lay an unconsolidated, saturated, deforming till layer. Subsequent coring also confirmed the
presence of water beneath the ice stream, and this led to a large research effort into the distribution of stresses at the base of an ice stream (Alley et al., 1986b; Alley et al., 1989; Alley et al., 1987c; Blankenship et al., 1987b; Engelhardt et al., 1990b; Engelhardt et al., 1990c; MacAyeal, 1989). When it was discovered that the base could not support the ice stream alone, research turned to investigate other mechanisms for controlling ice stream movement, such as sticky spots or margin restraint (Alley, 1993; Echelmeyer, 1994; Jacobel et al., 1990; Kamb, 1991).

In the late 1990’s research refocused to investigate bed properties and sub-glacial geology once more, using more refined radio echo sounding techniques coupled with satellite imagery. The stagnation of Ice Stream C has also been a recent research focus, and has led to more general work concerning the stability of ice streams, through interpretations of their past locations and flows (Anandakrishnan and Alley, 1997; Anandakrishnan et al., 1994; Anandakrishnan et al., 2001; Bentley et al., 1998).

**Conclusions**

An understanding of the factors that influence ice stream dynamics is important as they are the largest outlets for ice from the West Antarctic Ice Sheet. It is important that we can understand the stability of this mass of ice, as if it were to melt it would contribute up to 6 meters to global sea level rise (Patterson, 1994). Our understanding of ice streams have been influenced by the development of technologies that allow us to investigate such difficult and isolated terrain. Without remote sensing technologies such as airborne radar sounding, seismic reflection, and satellite imagery, our understanding of ice streams would be quite limited.

Ideas of basal deformation have changed our previous way of thinking that glacier mechanics were controlled by ice deformation and basal sliding. Future research however, needs to determine the causes of ice stream onset, and more thoroughly map sub-glacial geology in West Antarctica. In particular it appears that the distribution, thickness, and composition of sedimentary basins, may hold answers to where ice streams have previously flowed and the stability of their onset region. Recent work on the stagnation of Ice Stream C has aided in our understanding of ice stream stability, however this is also an area where further understanding is required. The concentration of work on the Ross Ice Streams, while interesting and important, has possibly been at the expense of an understanding of other ice streams, and therefore more effort should be given to research in other locations in the future. Finally, once our understanding of the controls and stability of ice streams is sufficient, the use of models to predict future changes will be a useful tool to anticipate the future of the West Antarctic Ice Sheet.
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