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ONSET OF STREAMING FLOW IN THE SIPLE COAST REGION, WEST ANTARCTICA

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Onsets of West Antarctic ice streams feeding the Ross Ice Shelf are reviewed. Inland flow and streaming flow are defined to clarify what is meant by the onset. Various means to locate the onset are discussed including: crevasses, flow-stripes, surface elevations, bed elevations and driving stress, with the last being the most reliable. A new map of driving stress in West Antarctica is presented which clearly shows the location of maximum driving stresses. Recent work is summarized and used to draw conclusions that onsets appear to share the properties of temperate basal conditions, location in a channel, and presence of a sedimentary bed. Consideration of previous analysis of shear margins suggests that the controlling process in the onset of streaming is basal dynamics rather than internal weaknesses such as strain softening or lateral shear margins. Migration of onsets is also concluded to be an inevitable consequence of their kinematics and may be an episodic process.

INTRODUCTION

Ice streams are responsible for the unique dynamic character of the West Antarctic ice sheet. These fast-moving conduits discharge ice rapidly from the slower inland ice and feed the large floating ice shelves. They also determine the shape of this marine-based ice sheet. Ice streams have lower surface slopes which, in turn, reduce the elevation of the ice-sheet interior and permit inland penetration of storms which deliver greater amounts of precipitation farther inland than in East Antarctica [*Lettau*, 1969]. Thus, the cross-sectional profile of the West Antarctic ice sheet differs significantly from the classic near-parabolic profiles of either the Greenland or the East Antarctic ice sheets.

Ice-stream flow is radically different from inland flow. The West Antarctic ice sheet lies in a deep, sediment-laden marine basin warmed from beneath by a relatively high geothermal heat flux [*Drewry*, 1983]. These factors combine to create a well-lubricated, water-saturated bed on which ice streams can maintain fast motion through the generation of sufficient quantities of water by basal friction and shear heating. The deformable aspects of the marine, subglacial till are important if not directly in the motion of the ice, then at least in their malleability or erosion into a smooth subglacial surface devoid of frequent protuberances which might hinder ice flow [*Alley*, 1990; *Cuffey and Alley*, 1996]. The rapid flow is also responsible for faster response times of West Antarctica as perturbations of flow are transmitted rapidly upstream into the inland ice and downstream onto the ice shelf along ice streams [*Bindschadler*, 1997].

Ice streams develop from the slower moving ice in the central regions of West Antarctica. Understanding what conditions must be present, or must develop, for the ice to make this major transition in its mode of flow is fundamental to understanding where and why ice streams form. Far less work has been completed on the study of streaming onsets than has been undertaken on the streaming process itself. Most of the work has been surface observations and inferences from them rather than direct measurements of subglacial processes in the vicinity of an onset. Nevertheless, it is worthwhile consolidating the work to date to assess what is known, what is speculated and whether earlier ideas about onset dynamics remain viable in the light of more recent work. This paper addresses these topics, along with clarifying the terminology associated with the onset region.

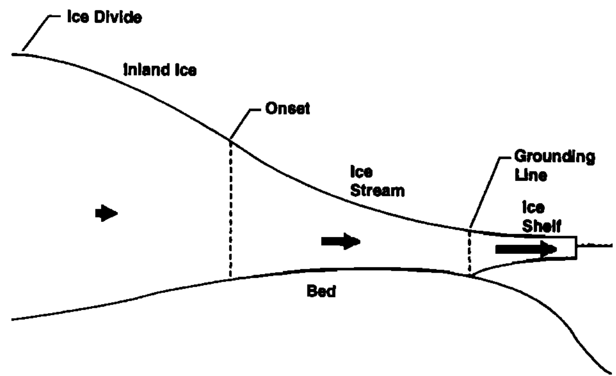

Fig. 1. Schematic cross-section of West Antarctica showing slow, inland ice, fast ice stream, and flat ice shelf. The transitions between these areas occur at the onset and at the grounding line. Length of arrows indicates flow speed. Surface and bed elevations illustrate progressively thinner ice from ice divide to shelf edge.

DEFINITION OF ONSET

The "ice-stream onset" or "streaming onset" (shortened hereafter as "onset") is defined here as the location of the transition between inland flow and streaming flow (Figure 1). This definition itself requires definitions of "inland flow" and "streaming flow". Inland flow is usually taken to describe flow resulting from internal deformation within the ice mass but, in West Antarctica, basal temperatures at the pressure melting point are likely so widespread that we expand the definition of inland flow here to allow some amount of basal sliding. The key characteristics of inland flow are that the gravitational driving stress (defined below) is balanced largely by basal shear stress and that the flow speed increases with increases in basal shear stress. The convex-up shape of ice sheets is a result of this relationship. A more detailed discussion of how much and what type of basal sliding occurs in inland flow is reserved for a later section.

Streaming flow, on the other hand, appears to be a distinctly different basal sliding process. Ice streams are the defining example, although any reference to their sliding mechanism was not included in the original definition of ice streams [*Swithinbank*, 1954]. Streaming flow appears to require both subglacial water pressurized to near the ice overburden pressure and subglacial till with a small shear strength. What fraction of the motion occurs by deformation within the till and how much is sliding of the ice over the till is still debated and is probably spatially variable [*Kamb*, 1991; *Kamb and Engelhardt*, 1998; *Alley and others*, 1989]. What typifies this type of flow is that speed increases with decreases in driving stress and that lateral shear, in addition to basal shear, provides an important resistance to flow [*Hughes*, 1977; *McIntyre*, 1985; *Whillans*, 1987; *Echelmeyer and others*, 1994; *Jackson and Kamb*, 1997; *Whillans and van der Veen*, 1997; *Harrison and others*, 1998]. The surface shape of an ice stream that results is concave-up as the slope decreases downstream from a steeper slope imposed by the inland flow regime, to a much shallower slope as the ice stream flows into the much flatter ice shelf (Figure 1).

These definitions of inland flow and streaming flow are shown in the next section to lead directly to means to locate the onset by examining either surface shape or the relationship between driving stress and flow speed.

LOCATION OF THE ONSET

Identifying fully developed ice streams is not difficult. They are many tens of kilometers wide, flow at speeds of many hundreds of meters per year, have heavily crevassed shear margins and have a more undulated surface topography than the adjacent, slower interstream ridges [*Bindschadler and Vornberger*, 1990]. Ice streams are fed through coalescence of an arborescent pattern of tributaries corresponding to a pattern of basal valleys [*Shabtaie and Bentley*, 1988; *Joughin and others*, 1999]. Each tributary has its own onset where it enters an ice stream. Multiple tributaries often feed a single ice stream forming larger ice streams [*Whillans and van der Veen*, 1993; *Scambos and Bindschadler*, 1991]. Downstream, the ice streams accelerate, they become thinner, and their beds become smoother on the multi-kilometer scale [*Retzlaff and Bentley*, 1993]. Most of these characteristics develop gradually making it difficult to trace an ice stream upstream and pinpoint the onset of each tributary.

Increasing the difficulty of onset location is the fact that, by our definition, the onset manifests a transition in flow mechanism at the base of the ice sheet, but measurements of basal conditions are sparse and often absent in potential onset areas. Yet, identification of the incipient ice stream is necessary to directing field studies of the physical processes critical to this flow transition. Thus, proxy indicators of the contrast in basal conditions have been sought to locate the onset. There may be no single best means to identify an onset region. We review various methods which have been employed, beginning with methods that rely on qualitative examination of the surface topography. There are two types of distinctive features that are numerous on ice streams: crevasses and flowstripes.

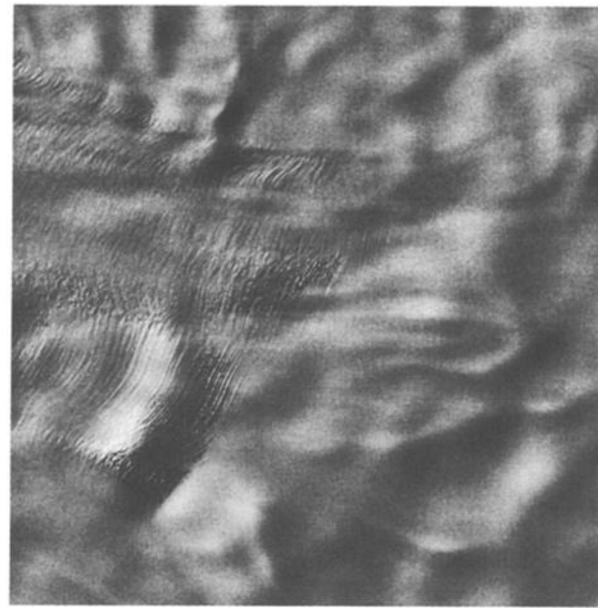

Fig. 2. Landsat image of an upstream area of ice stream E. Flow direction is toward bottom of image. Extensive transverse crevasses and flowstripe initiation both have been suggested as onset locators, however in this region they occur in different places. Image is 34.2 km on each side. Plate 1 shows approximate location.

Crevasses

Crevasses are formed when surface horizontal stresses exceed the tensile strength of ice. They are common on ice streams, especially at the margins, due to large velocity gradients. Some ice streams, particularly tributaries of B and E, have extensive sets of very long crevasses as their most upstream surface expression of fast flow (Figure 2). Individual crevasses are tens of kilometers in length, suggesting a regional and rapid increase in velocity [*Vornberger and Whillans*, 1986; *Stephenson and Bindschadler*, 1990]. This pattern is also seen on many Antarctic outlet glaciers, such as Byrd and Pine Island Glaciers which are not ice streams [*Bentley*, 1987; *McIntyre*, 1985], so this characteristic does not always represent an ice-stream onset. Nevertheless, when long transverse crevasses exist upflow of an ice stream, the inference that the crevasses locate the onset is reasonable. *McIntyre* [1985] noted that outside West Antarctica, this type of onset is correlated with a subglacial escarpment. *Bell and others* [1998] have made a similar correlation for one location in West Antarctica, but generally the onsets of the Siple Coast ice streams do not correlate with basal escarpments.

Flowstripes

Another distinctive type of ice-stream surface feature is "flowstriping"—development of narrow ridges or troughs a few hundred meters across, tens to hundreds of kilometers long and with amplitudes of a few meters. Flowstripes also have been referred to as "flowbands", "flowlines" and "stripes" by various authors. Different theories have been proposed to explain their formation [*Gudmundsson and others*, 1998; *Casassa and Brecher*, 1993] but all agree that their considerable length is the result of rapid ice motion which displaces them great distances over a time period much less than their relaxation time.

The formation of flowstripes has been modeled by ice sliding over perturbations in either basal topography or basal friction [*Gudmundsson and others*, 1998]. Flowstripes also have been observed emanating from surface bumps and margins of ice streams [*Merry and Whillans*, 1993]. Where surface velocities are available, flowstripes correspond roughly with regions moving at speeds in excess of 100 m/a [*Scambos and Bindschadler*, 1993; *Chen and others*, 1998].

Implicit in the association of flowstripes with ice streams is the assumption that the boundary between flowstripes and no flowstripes represents the onset of the ice stream. In other words, the subglacial conditions necessary for flowstripe generation are the same conditions required for streaming flow. *Dowdeswell and McIntyre*

[1987] showed that flowstripes affected the power spectrum of the surface elevation field enough that a broad discrimination was possible between the ice-stream region which contained flowstripes (their "type 3" surface) and the inland region ("type 2") which did not contain flowstripes. They quantified type 2 topography as having horizontal wavelengths generally greater than 20 km and vertical amplitudes of up to 16 meters with a root-mean-square of 4-5 meters. By contrast, type 3 topography had a much broader power spectrum often dominated by horizontal wavelengths less than 10 kilometers. The boundary of these two topography types afforded an estimate of the onset position, but the construction of the power spectrum required a large spatial sample, so the boundary, and therefore the onset location, is not particularly sharp.

Flowstripes are resilient features that can persist for centuries [*Gudmundsson and others*, 1998]. However, the flow field can change on shorter time scales [*Stephenson and Bindschadler*, 1988; *Bindschadler and Vornberger*, 1998]. There are many examples of flowstripes misaligned with a measured flow field, indicating that conditions are out of equilibrium. This condition can lead to misinterpretation of present onset conditions and flow direction if the most upstream flowstripes are equated to a present onset of streaming flow. An excellent example of this has been documented in the upstream reaches of ice stream D, where a band of flowstripes was interpreted as an onset region [*Hodge and Doppelhammer*, 1996], yet later measurements of velocity indicated flow is not presently parallel to the flowstripes [*Bindschadler and others*, in press]. The analysis of the velocity field (presented later) suggests this region is not presently an onset.

The formation process of flowstripes may not be unique, and therefore their use for locating onsets may be questionable. Figure 2 shows flowstripes of the type in which the upstream end of each flowstripe is associated with a surface topographic bump. This situation is common at the margins of narrow ice streams in the catchment areas of ice streams D and E, where complete coverage by Landsat imagery permits detailed examination [*Scambos and Bindschadler*, 1991]. Ice streams often become wider by the incorporation of adjacent ice, but the flowstripes that accompany this process do not locate an onset because the ice stream already exists upstream. The example of Figure 2 also casts crevasse formation in doubt as a reliable means of onset determination because the flowlines just described occur upstream of the crevasses.

A second situation in which flowstripes often occur is shown in Figure 3. Here, the upstream ends of a set of parallel flowstripes are not associated with any obvious surface feature. Flowstripes are extremely subtle, indicating subdued surface relief (or, in some cases, a sun-parallel orientation), making their initiation difficult to locate. In most cases, flowstripes continue downstream and become more clearly defined as the ice stream accelerates [*Stephenson and Bindschadler*, 1990]. The flowstripes of Figure 3 are unique because the flowstripes shown do not connect with an ice stream. This suggests the occurrence of failed onsets, an important possibility of onset dynamics.

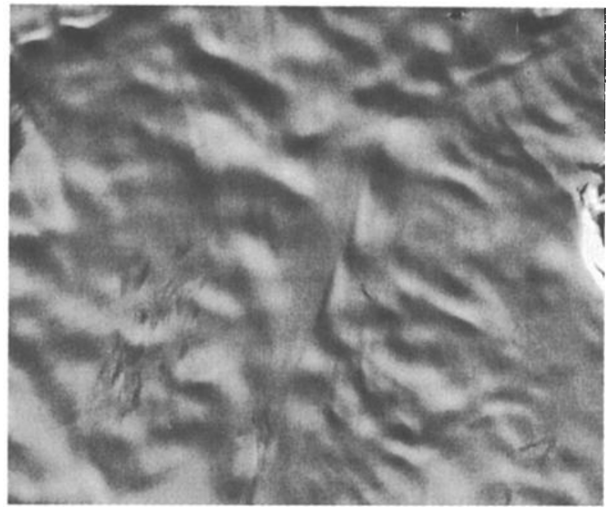

Fig. 3. Possible failed onset. Landsat image show a region of flowstripes (center of image) which fails to continue downstream (toward bottom of image). Image is 68.4 km x 57 km. Plate 1 shows approximate location.

Surface Elevations

Surface elevations furnish an additional method of locating onsets. The general profile along an ice-stream flowline (Figure 1) illustrates that the boundary between inland ice and the ice stream corresponds to the maximum surface slope and an inflection point in the along-flow elevation profile. This is a result of the convex-up elevation profile of inland ice and the concave-up profile of ice streams mentioned earlier. In practice, topographic variations on smaller scales make it difficult to apply this method without smoothing the surface profile, which partially defeats the purpose of finding the precise location of maximum surface slope. Because this method is closely tied to the use of maximum driving stress for onset location, it is developed more fully in the "Driving Stresses" section.

Bed Elevations

All the above methods rely on surface topography alone. Introduction of information on ice thickness can usually improve the identification. *McIntyre* [1985] has noted that the West Antarctic ice streams do not follow the general correlation observed elsewhere in Antarctica that fast flow initiates at a step in the bed elevation. *Retzlaff and Bentley* [1993] also commented that their extensive bed-elevation mapping of ice streams A, B, and C failed to show any large gradients in basal topography in the regions where they believed onsets must be located. However, *Bell and others* [1998] found one deep bed channel whose distinct headwall coincided with the initiation of flowstripes. They inferred this location was also the onset of streaming flow.

Driving Stress

Ice moves in response to the gravitational force exerted on it. This driving stress, τ_d, is defined as

$$\tau_d = \rho \; g \; H \sin\alpha \qquad (1)$$

where ρ is the average column density, g is the gravitational acceleration, α is surface slope and H is ice thickness. Slope increases downstream within the inland-ice regime because driving stress must increase to transport an increasing flux of ice [*Paterson*, 1994]. The specific relationship for a non-linear flow law produces the near-parabolic elevation profile [*Vialov*, 1970; *Paterson*, 1994]. By contrast, ice streams exhibit a decreasing slope downstream while their velocity is increasing. The diminishing slope produces a smooth transition to the nearly flat ice shelves. Decreasing ice stream thickness downstream is not only a result of downstream-dipping surface slope, but is also accentuated by the generally lower basal elevations in the interior caused by increased crustal depression underneath the thicker inland ice. This slope pattern creates an inflection point in the elevation profile at the onset.

Generally, surface slope varies spatially far more along an ice-sheet flowline than ice thickness (and certainly more than ice density) [*Whillans*, 1987], so the pattern of driving stress mimics the pattern of surface slope. Thus, just as there is a maximum in surface slope at the onset, the driving stress maximum is also nearly coincident with the onset.

Because driving stress depends on ice thickness, it cannot be determined from surface measurements alone. Data from regional surveys of ice thickness and surface elevation were used to calculate the pattern of driving stress [*Cooper and others*, 1982]. These data were collected from airborne surveys in a roughly orthogonal grid pattern with a spacing of approximately 50 km. As expected, where ice flowed into ice shelves, a zone of maximum driving stress occurred between the flat ice divides and the flat ice shelves. This was particularly apparent for the heads of ice streams A, B and C.

The largest source of error in those driving stress calculations was identified as the surface-elevation measurement. For this reason, we present a recalculation of West Antarctic driving stress (Plate 1) using a more accurate elevation data set based on satellite altimetry and a regridding of the same ice thickness measurements used by Cooper and others [*Bamber and Huybrechts*, 1996]. The spatial averaging for surface slope used in Plate 1 is 20 times the local ice thickness to average out the effect of longitudinal stress gradients on the effective shear stress at the bed [*Budd*, 1969].

The ice divides are evident by their very low driving stresses (red in Plate 1), caused by their very low surface slopes. The zone of maximum driving stress upstream of ice streams A-E is evident, ranging from 80 kPa (blue) to 150 kPa (brown) with the highest maxima and greatest widths on the north and south ends of the zone. This pattern expresses the overall bowl-shaped configuration of the Ross embayment section of the West Antarctic—the north end (Executive Committee Range) and south end (Whitmore and Transantarctic Mountains) of the zone maintain higher elevations, while the central elevations of West Antarctica are somewhat lower. These features contrast with outlet glaciers in East Antarctica, and with Pine Island and Thwaites Glaciers in West Antarctica as well, where driving stresses achieve maxima exceeding 150 kPa and where these maxima occur much closer to the coast [*Bentley*, 1987; *Cooper and others*, 1982]. The location of a recently discovered subglacial volcano [*Blankenship and others*, 1993] lies well away from the zone of maximum driving stress, suggesting it plays no role in ice-stream initiation.

Figure 4 shows 500-km-long longitudinal profiles of driving stress along center flowlines of ice streams B-E. These all show the characteristic feature of low driving stress at the divide, rising to a maximum, and decreasing along the ice stream to a value of about 20 kPa at the entrance to the ice shelf. The magnitude of the maximum ranges from 80 kPa for D to 150 kPa for B. The maximum is distinct and single for these two cases. The profile for ice stream C exhibits a modest secondary maximum upstream of the largest peak. Thickening and a topographic bulge have been identified at the head of the

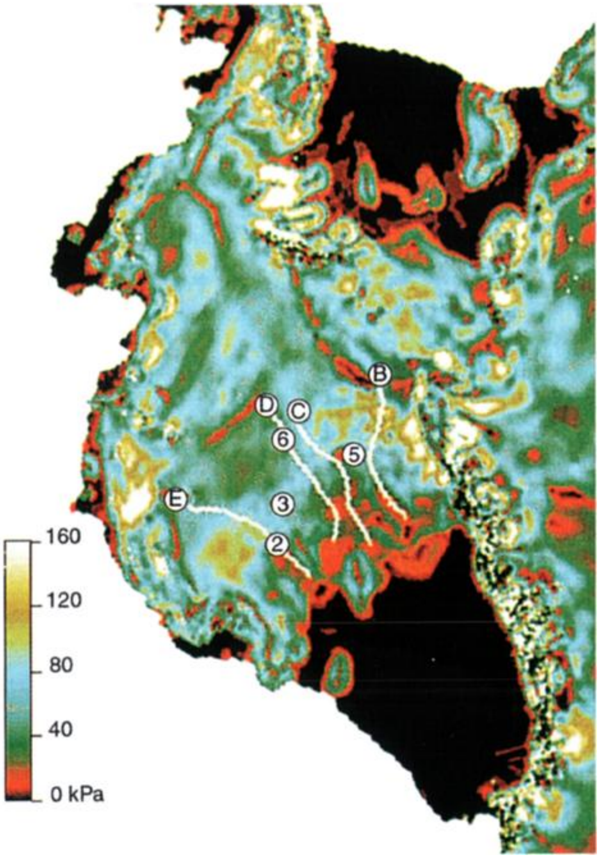

Plate 1. Distribution of driving stress in West Antarctica calculated from Equation 1 as described in the text. Large black areas correspond to Ross and Ronne/Filchner Ice Shelves. Solid white lines indicate positions of 500-km-long centerline profiles of Siple Coast ice streams B–E in Figure 4. Numbers indicate locations of Figures 2, 3,6 and 7. Ice divides appear as narrow red regions in central West Antarctica. Band of maximum driving stress across Siple Coast ice streams appears as brown areas on B profile and north of E profile and connecting blue band across C and D.

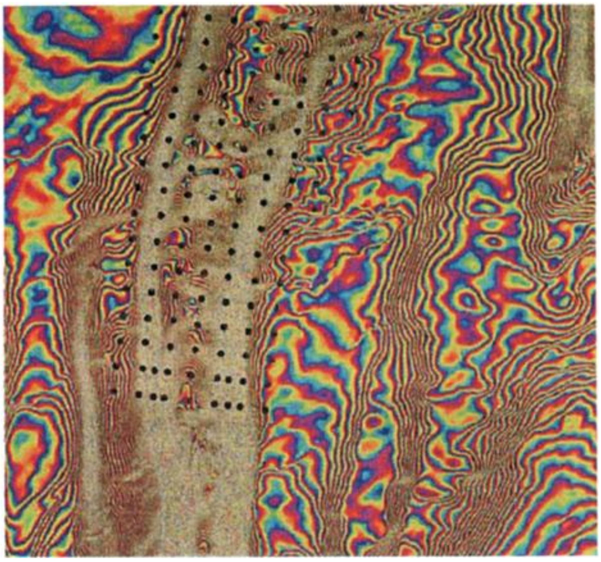

Plate 2. Synthetic aperture radar (SAR) interferogram from Radarsat data covering downstream one-third of ice stream D onset grid. Fringes correspond predominantly to relative velocity in direction toward satellite (toward top of figure). Points on figure correspond to downstream portion of gridpoints in Figure 6.

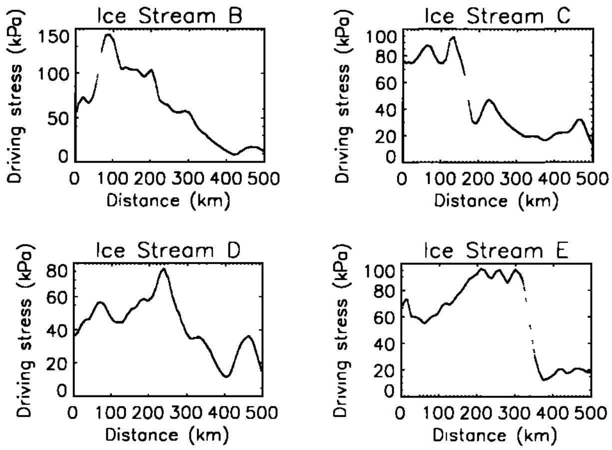

Fig. 4. Profiles of driving stress along ice streams B, C, D, and E. Profile locations are shown in Plate 1.

stagnant portion of this ice stream [*Joughin and others*, 1999]. The observed maxima in Figure 4 express this evolving condition and are probably not a reliable locator of the onset of this inactive ice stream. Ice stream E's profile displays a triple maximum suggesting a case where streaming flow begins, but is not sustained (c.f., Figure 3) until the position of the third maximum. This interpretation is supported by recent measurements of the flow pattern feeding the Siple Coast ice streams [*Joughin and others*, 1999].

DETAILED STUDIES OF ONSET AREAS

Recent fieldwork has focused on particular regions believed to be onset areas. One such study was based entirely on data collected remotely in the catchment area of ice stream C [*Bell and others*, 1998]. From the combined evidence of surface features, ice thickness, as well as magnetic and gravimetric signatures, it was concluded that initial streaming flow coincided with a 1000-m deep, 20-km wide bed channel in the basal topography. This correlation is reminiscent of the earlier correlations suggested by *McIntyre* [1985]. The onset's location was inferred from the appearance of flowstripes in 1-km resolution satellite imagery, but a new measurement of the velocity field shows that while ice does speed up at this location, the ice eventually slows as it meets the stagnant ice stream C [*Joughin and others*, 1999]. It is uncertain whether the ice does, or ever did, attain the streaming condition at this location. Nevertheless, the modeling of the gravity and magnetic signatures of this region demonstrated that the channel where speed increases contains a low density, non-magnetic sedimentary deposit at least one-kilometer thick.

The correlation of sedimentary units with fast flow is further advanced by surface field studies farther downstream along this channel. Reflection and refraction seismic surveys, in combination with measurements of surface ice velocity, demonstrate that a sharp contrast in ice speed coincides with the edge of a modeled subglacial sedimentary layer many hundreds of meters thick [*Anandakrishnan and others*, 1998]. Figure 5 shows the spatial correlation deduced by these authors, who also suggest that the thinner, narrower sedimentary unit (at km 40 of Figure 5) does not produce streaming flow because the width of this unit is narrow enough that side shear of adjacent slower ice effectively prevents streaming. This point is discussed more fully later.

A spatially more extensive field program designed specifically to locate and measure the surface deformation field associated with an onset was undertaken near Byrd Station on ice feeding into ice stream D [*Chen and others*, 1998; *Bindschadler and others*, in press]. Repeated surveys of surface markers placed in a regular grid (5-km spacing] quantified the surface velocity field and, from it, the surface-horizontal strain field [*Chen and others*, 1998]. A convergent, accelerating flow into the ice stream was apparent (Figure 6). As with the previous onset study, the developing stream coincided with a bed channel, although measurements to confirm the presence of a sedimentary layer were not made as part of the survey [*Bamber and Bindschadler*, 1997].

Analysis of the measured velocity field and associated ice-sheet geometry was conducted to locate the onset. The approach deemed most successful drew on the previously discussed difference of the driving stress versus velocity relationship for inland flow and for streaming flow. We now develop this subject more fully.

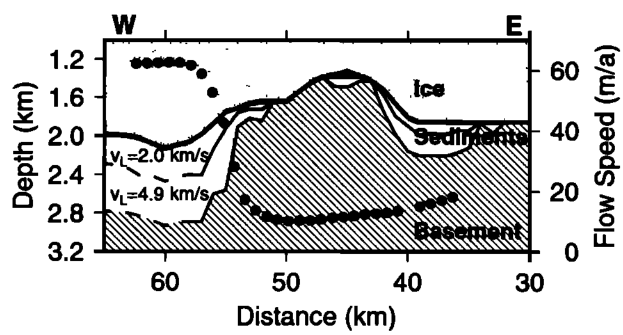

Fig. 5. Model of across-flow section near an onset calculated from inversion of seismic data and the GPS-measured surface velocity profile (from *Anandakrishnan and others*, 1998, Figure 4). Plate 1 shows approximate location.

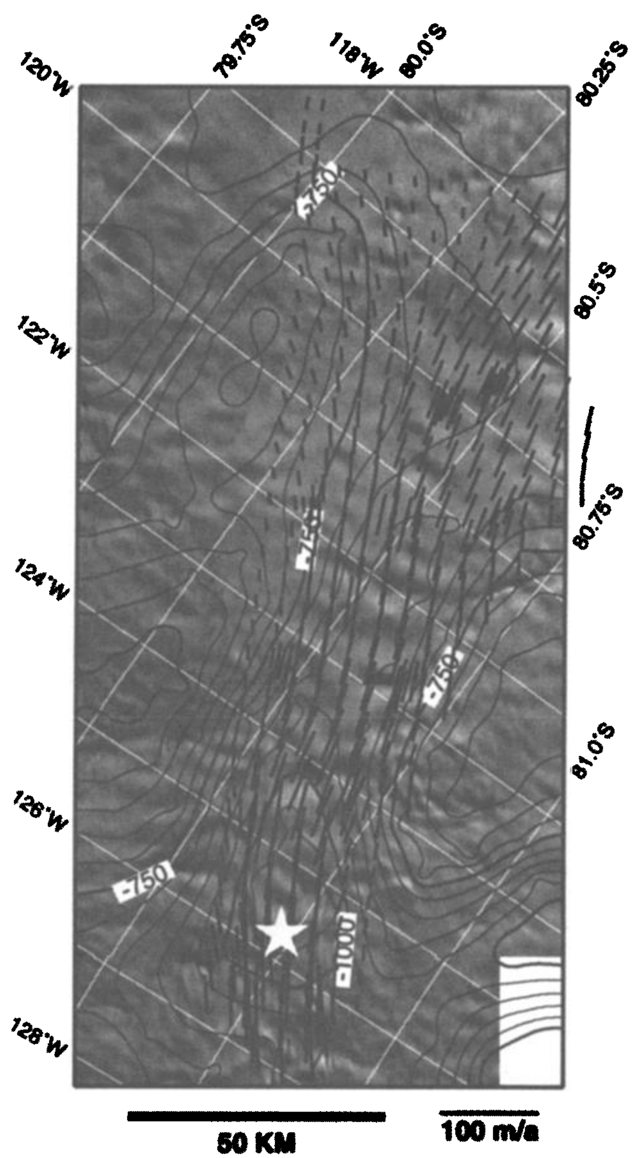

Fig. 6. GPS-measured surface velocities, Landsat imagery and bed elevation contours in the vicinity of the onset of ice stream D (from *Bindschadler and others*, in press). Data sources are discussed in text. Velocities are spaced 5 km apart. Image size is 152 km x 195.5 km. Plate 1 shows approximate location. Star is location of maximum driving stress.

The rheological properties of ice often are treated as a non-linear flow law of the form,

$$\varepsilon = A\tau^n \qquad (2)$$

where ε is the octahedral strain rate, τ is the octahedral stress, A is the temperature-dependent flow-law coefficient, and n is the flow-law exponent [*Glen*, 1955]. In the simplified case of parallel-sided, laminar flow, where bed-parallel shear is the only non-zero component of the stress tensor, the surface velocity, U_s, can be expressed as

$$U_s = U_b + 2\,A\,(\rho\, g\, \sin\alpha\, H)^n\, H\,/\,(n+1) \qquad (3)$$

where U_b is the basal sliding velocity [*Paterson*, 1994]. Substituting driving stress (Equation 1), this can be rewritten as

$$(U_s - U_b)/H = 2\,A\,\tau_d^{\,n}\,/\,(n+1) \qquad (4)$$

The ratio $(U_s - U_b)/H$ expresses the mean shear strain rate through the column [*Budd and Smith*, 1981]. Because the sliding component in the surveyed data is unknown, Figure 7 plots U_s vs. τ_d for the surveyed grid. Surface slope at each gridpoint was determined from a bilinear fit to surface elevations over a 40 km x 40 km area [*Bindschadler and others*, in press]. Open circles in Figure 7 follow the kinematic centerline, where the laminar flow approximation is most likely valid. These points show the expected pattern of increasing velocity versus increasing driving stress.

From the definition of streaming flow, the opposite relationship between driving stress and velocity is expected on the ice stream. Figure 8 includes points taken from data farther downstream on ice stream D and, as expected, an inverse relationship is evident [*Bindschadler*

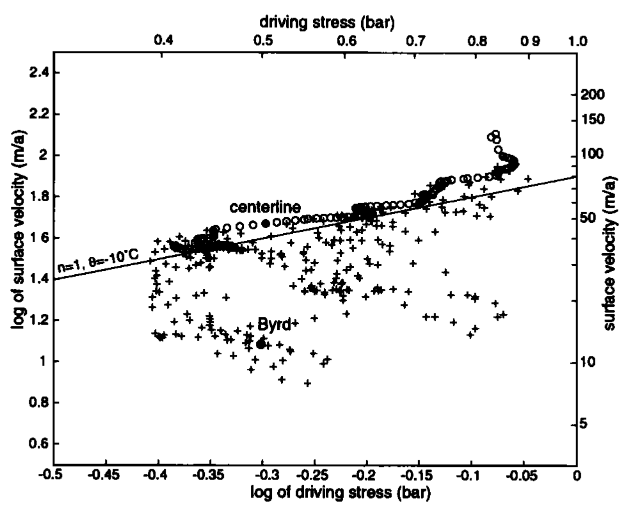

Fig. 7. Surface velocity versus driving stress. Data points are from ice stream D onset grid where bold circles are gridpoints along the kinematic centerline [*Bindschadler and others*, in press]. Solid line corresponds to n=1 at temperature of −10°C in Equation 4.

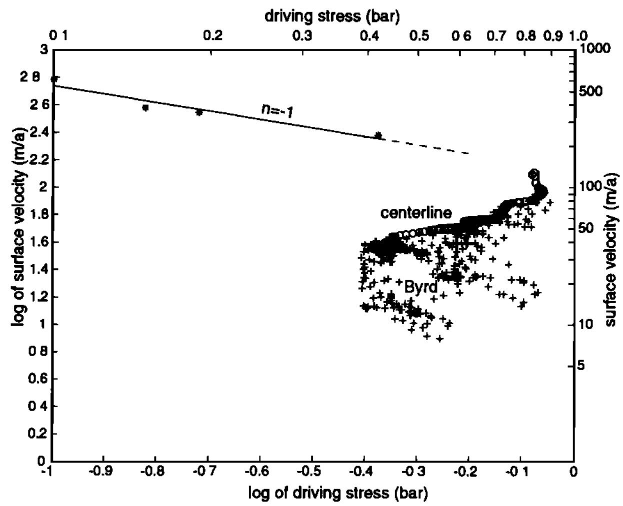

Fig. 8. Surface velocity versus driving stress. Crosses and circles are same data as in Figure 7. Additional data (4 asterisks) are taken from centerline positions distributed along ice stream D where velocities are given in *Bindschadler and others* [1996]. Solid line is n—1 linear fit to ice stream data.

and others, in press]. Thus, along the centerline of the developing ice stream, the driving stress rises almost steadily, and reaches a maximum value of 87 kPa before reversing to decreasing stresses as the velocity continues to increase. *Bindschadler and others* [in press] used this result to characterize the position where the slope of the relationship between τ_d and U_s/H changes sign as the onset of streaming flow.

A more detailed measurement of surface motion in the vicinity of ice stream D's onset was produced from synthetic aperture radar data from Canada's Radarsat satellite [*Gray and others*, 1998]. Two images were combined into an interferogram giving a very sensitive measurement of differences in path length from the satellite to the surface. Plate 2 represents the path-length difference as a series of color fringes. Each full fringe represents approximately 6 cm of motion along the path from the satellite to the surface (up from the bottom of the figure). Displacements are caused by both surface topography (due to parallax) and ice motion, but because the parallax component is rather small, most of the densely spaced fringing represents velocity gradients in the surface motion field. The fringe pattern confirms the general pattern of the grid survey measurements and includes more detail within the survey area. Most importantly, this study shows the complexity of the motion field within, and beyond, the survey grid. In particular, there is a region to the north (left in the figure) of the main tributary that illustrates a secondary flow parallel to the main tributary. Future interferometric results undoubtedly will prove extremely valuable in assessing the location of and patterns in the onset of streaming flow.

PHYSICAL PROCESSES AT THE ONSET

There have been no direct observations of the basal processes within an onset area. Nevertheless, the surface measurements discussed above, combined with known processes extant under active ice streams and at Byrd Station, in the inland ice regime, provide some basis for speculation of what may occur within an onset area.

Basal Temperatures

A necessary condition for streaming to begin is that the basal ice be at the pressure-melting temperature. The high probability of temperate basal conditions throughout much of West Antarctica has been mentioned earlier. A melting condition was observed at the base of the Byrd Station borehole more than 100 km upstream of ice stream D's onset, but the physical setting of Byrd—positioned on the upstream side of a large bed high—can be viewed as atypical [*Whillans*, 1979].

High accumulation rates will tend to lower the average temperature of the ice, but large geothermal heat flow will help offset this cooling [*Paterson*, 1994]. Thermal models of the West Antarctic ice sheet based on the steady state solution of *Robin* [1955] require improbably low values of geothermal heat flow to avoid a temperate bed [*Rose*, 1979; *Anandakrishnan and others*, 1998]. Steady state thermal conditions may not be reached on fast moving ice, but the journey of inland ice from the ice divide toward the onset area is measured in many thousands of years. Thus, through the combined processes of geothermal heating and strain heating in deforming basal layers, there is ample time for basal ice to be warmed to the pressure melting point.

Vertical stretching of ice will enhance the warming of basal layers contributing to the observed preference of streaming initiation in bed channels. This effect has been observed on Jakobhavns Isbræ, Greenland, where a thick temperate-ice layer created by vertical stretching as ice enters a deep bed channel is believed responsible for rapid shear deformation rates [*Iken and others*, 1993; *Funk and others*, 1994]. Vertical stretching experienced by ice flowing into bed channels also acts to decrease the vertical temperature gradients in the ice which, in turn, decreases the vertical conduction of heat upward through the column. The result would be a tendency to retain geothermal heat in the basal layers and deformation rates.

This effect is transient and would only last as long as it takes the temperature profile to re-equilibrate, but increased shear heating would delay equilibration time and maintain a thicker layer of temperate ice. If the basal ice is already temperate before entering the bed channel, the trapped heat would then be used to melt ice, producing water that could accentuate basal sliding.

Basal Sliding

The basal sliding process has been examined most extensively on temperate glaciers. In general, the sliding velocity, U_b, is related to the basal shear stress, τ_b, through a relationship of the form,

$$U_b = k\, \tau_b{}^m \tag{5}$$

The coefficient, k, is not constant in many proposed sliding laws and is specified to depend on factors such as water pressure and, in some cases, the sliding velocity itself [*Paterson*, 1994]. Early theoretical treatments of basal sliding suggested m=2, but subsequent approaches produced values of m=1 [*Paterson*, 1994]. The relevant point here is that as long as the exponent, m, is positive, the form of this sliding relationship fits with the earlier definition of inland flow—i.e., increasing velocity with increasing driving stress. It is only when the exponent, either of deformation (Equation 4) or sliding (Equation 5) is negative, that ice velocity increases with decreasing driving stress—the defining characteristic of streaming ice.

Evidence already presented argues strongly for a temperate bed and possibly a thick layer of temperate ice as the streaming onset is approached. In Figure 7, ice upstream of the onset is best fit by a flow-law exponent of unity (ice thickness in this region is nearly constant at 2300 $\pm$ 100 m; Bindschadler and others, in press]. Without more observational data of temperature and basal conditions, the issues of how the measured surface velocity is distributed between internal deformation and basal sliding and what conditions are responsible for this distribution cannot be pursued productively.

Subglacial Water

Subglacial water is known to play a critical role in streaming flow. Thus, the hydrologic evolution of the subglacial environment is likely key to ice-stream initiation. Although the driving stress is a maximum at the onset, the rate of water production depends on many factors. Geothermal heat flux is spatially variable, yet even a subglacial volcano, believed by some to be active, does not initiate an ice stream [*Blankenship and others*, 1993]. Strain heating also produces water in the lower layers of ice, but at rates much less than produced by basal friction. Frictional heating depends on the stress at the interface. Deforming basal tills could lower the efficiency of water production.

Surface slopes converge laterally toward ice streams and tributaries leading toward onsets occupy basal channels. These two conditions will tend to concentrate whatever water is produced toward the onset region. By contrast, the bulge now forming at the head of the stagnant ice stream C is diverting water from this site [*Anandakrishnan and Alley*, 1997; *Price*, personal communication]. How, and if, this ice stream reinitiates remains to be seen.

Lateral Margins

Theoretical analysis of shear-margin dynamics supports the idea that ice-stream initiation is controlled by the development of basal slip rather than by the development of lateral shearing. *Raymond* [1996] analyzed the delocalization of shear stress within an ice-stream margin, and showed that flow resistance extended a variable distance into the ice stream depending primarily on the sliding speed (expressed as a ratio of the deformation speed, and the flow law exponent). Three regimes were identified: low relative sliding speed, where the effect of the margin was local to the margin and the speed in the center of the stream was controlled by basal conditions; very high sliding speeds where the margin resistance controlled the speed across the entire ice stream; and an intermediate case where the ice stream speed was controlled by a combination of margin and basal resistance. The sliding speeds on active ice streams were found to place them in the category where both margin and basal processes are important. Applying his results to onset areas, lower relative sliding speeds are certain, placing the dynamics of the onset region in the category where margin resistance remains local.

In order for the dynamics of an onset area to be controlled by lateral drag from the margins, the onset would have to be no wider than a few ice thicknesses. The general observation from Landsat imagery is that tributaries form with characteristic widths of 20 km. Narrower cases do occur, but the widths increase shortly downstream by incorporation of ice across a margin. Figure 3 illustrates an example of a narrow flowstripe area believed to represent the early stages of streaming, but it does not widen and, by the disappearance of flowstripes, flow speed is inferred to decrease. Narrow tributaries feeding ice stream E from the south also illustrate a repeated pattern of acceleration and deceleration prior to becoming a fully developed ice stream [*Joughin and others*, 1999]. Thus, it is improbable that ice-stream initiation is controlled by the formation of "soft spots" of strain-softened ice creating a weak margin as has been hypothesized [*Merry and Whillans*, 1993; *Whillans and Bolzan*, 1987].

MIGRATION OF THE ONSET

The subject of onsets should not be left without discussing their possible migration, because this process has major implications for the evolution of the ice sheet. It is known that the West Antarctic ice sheet was much larger during the Last Glacial Maximum (LGM)—*Hughes* [1998] has modeled an ice sheet fully three times its present volume during this earlier epoch, with ancient ice streams extending to the edge of the larger ice sheet. A stratified ridge-trough system in the Ross Sea and under the Ross Ice Shelf also suggest that a number of ice streams extended to the ice-sheet's margin [*Anderson and others*, 1992].

What cannot be determined is whether those LGM ice streams originated at the same locations as the present ice streams. If so, they would have been many times longer than present and their thicknesses across the Ross Ice Shelf would have had to have been larger to be grounded. Higher inland elevations of a few hundered meters at most during the LGM [*Whillans*, 1979; *Borns*, 1995; *Hughes*, 1998] require that the average driving stresses for these ancient ice streams would have been considerably smaller. Yet, a characteristic of ice streams is that they move rapidly at low driving stresses. It is not known whether the fact that all the present ice streams feeding the Ross Ice Shelf are roughly the same length also is a significant characteristic.

A simple argument has been proposed to support the idea that the heads of ice streams are out of equilibrium [*Bindschadler*, 1997]. The lateral convergence of inland ice into ice streams feeding the Ross Ice Shelf is roughly a factor of two. In addition, ice thickness decreases by about a factor of two as ice streams form. Thus, based on mass continuity, ice speed should accelerate by a combined factor of four. In contrast, ice typically accelerates by a larger factor—ten in the case of the ice stream D grid. This non-equilibrium situation can result in a number of responses: the region of acceleration can thin; the upstream ice can accelerate; the downstream ice can decelerate; or the downstream width can narrow. (Note that the upstream reservoir cannot widen). Neither of the last two possibilities is observed. Either of the remaining first two possibilities lead to the upstream migration of the onset.

If ancient ice streams migrated, they did so at roughly the same rate as the ice sheet retreated. *Bindschadler* [1998] has calculated that for ice stream B, the retreat rate averaged 100 m/a for approximately the last 12,000 years. This rate compares favorably with two other attempts to calculate the inland migration rate of ice stream B. From the thinning rate at the onset [*Shabtaie and others*, 1988] and the regional surface slope, *Bindschadler* [1997] calculated a migration rate of 488 m/a. In a more local study, *Price* [1998] modeled the evolution of a peculiar crevasse pattern near the onset of one branch of the ice stream and calculated a mean upstream migration rate 230 m/a.

It is conceivable, even likely, that conflicts might arise between the requirement to migrate, from mass continuity, and the basal conditions necessary to support an onset. If only a limited number of sites capable of supporting an onset existed, as the onset region evolved, the suitability of the current onset might diminish as the suitability of nearby sites increased. Eventually, the onset might jump to a new position or, if no suitable sites were near, streaming flow might cease. Episodic migration was first suggested by *Whillans and Bolzan* [1987] and later supported by the analysis of *Price* [1998]. This effect would depend on the spatial heterogeneity of the basal conditions capable of supporting an onset—a parameter that cannot be quantified with our present understanding of onsets and basal conditions. Such a phenomenon might generate the large "rafts" of relatively undisturbed ice observed on ice stream B [*Whillans and Bolzan*, 1987; *Bindschadler and others*, 1988].

SUMMARY

Necessary conditions for streaming flow are subglacial water, a thawed ice/bed interface, and probably sediment with a low shear strength. Observations of onset locations are limited, but are consistent with the presence of sediment, either inferred from seismic or gravity and magnetic measurements and/or by the correlation of the incipient ice stream with a bed channel. They also are consistent with the likely presence of subglacial water because the shape of the ice sheet surface and the subglacial bed combine to concentrate subglacial water toward the onset region. The predominance of these factors and their correlation with known or strongly suspected onsets is further supported by modeling of the effects of lateral shear. This leads to the conclusion that the onset of an ice stream represents a radical transition in basal motion that is not controlled by internal weaknesses in the ice such as strain softening in lateral shear margins.

Proxy indicators based on surface features alone are the least reliable, yet the most convenient, means to identify the onset location of an ice stream. The absence of either flowstripes or crevasses is not reliable confirmation of inland flow. Thus, these are better used to confirm the existence of streaming flow than to pinpoint its initiation. Also, the longevity of flowstripes can lead to false interpretations if flow patterns have changed. Maximum driving stress is arguably the most reliable means to locate the onset of an ice stream in lieu of direct subglacial measurements, but is further enhanced if velocity and thickness measurements are available to examine more carefully the spatial relationship between driving stress, mean shear strain rate and inferred basal motion.

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