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ANTARCTIC ICE STREAMS: A REVIEW

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Abstract. An ice stream is a part of an inland ice sheet that flows rapidly through the surrounding ice. The distinction between an ice sheet and an outlet glacier is clear in principle but muddy in practice -- many Antarctic glaciers are intermediate in character. The "Ross ice streams," which flow through the West Antarctic inland ice into the Ross Ice Shelf, are distinct in character, differing even from other ice streams in the marine ice sheet of West Antarctica. Their surface elevation profiles are low, their bed slopes are low and smooth, and their driving stresses diminish monotonically downglacier. In transverse profile they are broader in relation to ice thickness and exhibit shallower subglacial troughs, than other ice streams. Many models for the fast sliding of glaciers have been applied to the Ross ice streams; most have included in some form a reduction in basal drag resulting from a lesser effective than glaciostatic pressure at the bed. Most models are only semiphysical at best, so need to be "tuned" by fitting to assumed parameters at some place or places in the real ice sheet. The recent discovery of a very small effective pressure beneath one ice stream consequently has led to some gross errors in the velocities predicted by the models. The difficulty may be resolved if it is true, as recent experiments suggest, that ice stream B, and by extrapolation other Ross ice streams as well, slide on a deforming bed that absorbs most or all of the differential motion between the ice and the bedrock.

Physiographic Aspects

Any review of ice streams properly should start with a definition. I will use the definition from Swithinbank [1954]: an ice stream is "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." There are two aspects in particular that we may consider from this definition: that an ice stream is bordered by ice not by rock (if it is bordered by rock, then it is an outlet glacier) and that, as part of the inland ice sheet, it is not floating (if the ice is floating, it is shelf ice).

Although this is a fine definition in principle, the distinction between ice streams and outlet glaciers in particular becomes rather hazy in practice. We may find good examples of pure ice streams such as ice streams B, C, D, and E (Figure 1) that feed into the Ross Ice Shelf, and good examples of outlet glaciers such as Byrd Glacier (Figure 2) and others that feed the Ross Ice Shelf through the Transantarctic Mountains. However, there are many fast-moving parts of the inland ice sheet in Antarctica that are more difficult to classify. For example, Rutford Ice Stream (Figure 3) is bounded on one side by the Ellsworth Mountains but on the other side by a smooth-surfaced low-lying part of the ice sheet. Many "ice streams" show characteristics of ice streams along parts of their lengths and characteristics of outlet glaciers in others -- the Lambert Glacier (Figure 4), for example. The borders of still others such as Jutulstraumen (Figure 5) are characterized by scattered nunataks penetrating the surface, although it is clear that the glacier lies in a deep valley in the subglacial bedrock [Decleir and Van Autenboer, 1982]. Others that occur in generally mountainous areas are largely free of any exposed rock along the borders, e.g., Slessor Glacier (Figure 6).

The geographic distinction between an ice shelf and an ice stream also becomes fuzzy when one considers active ice streams entering very small ice shelves that they dominate. A particular example of this is Pine Island Glacier (Figure 7), which maintains its clear identity through the small ice shelf in Pine Island Bay right to its edge at the ocean front. But an ice shelf spreads due to longitudinal stresses rather than shear stresses, which its floating base cannot support, so there is a profound dynamic difference between an ice stream and an ice shelf that justifies defining ice streams to be portions only of the inland ice.

Consider next some measured characteristics of fast-moving glaciers in Antarctica. (Ice streams and outlet glaciers exist also in Greenland, where in fact, ice streams were first recognized [Rink, 1877]. Jacobshavn Glacier, which is both an outlet glacier and an ice stream, is currently under investigation [Echelmeyer and Harrison, 1986], but detailed data are not yet available.) Longitudinal profiles of surface elevation, bedrock elevation, and driving stress for several East Antarctic ice streams are shown in Figure 8. The driving stress, τ_d, is given by

$$\tau_d = \rho_i g h \sin \alpha \qquad (1)$$

where ρ_i is the density of ice, g the acceleration of gravity, h the ice thickness, and α the surface slope. Also shown for comparison is an East Antarctic sheet-flow surface profile inland from Syowa Station [Shimizu et al., 1978].

The surface profiles of the smaller glaciers (Hays, Robert, and Rayner glaciers) do not vary a great deal from the East Antarctic sheet-flow profile, which presumably is an equilibrium profile. Nevertheless, a close look shows that each of these glaciers has an inflection point in its surface profile coastward of which the profile is concave up rather than convex up. Although the bedrock profiles for these glaciers are rather different, the result is that each glacier shows

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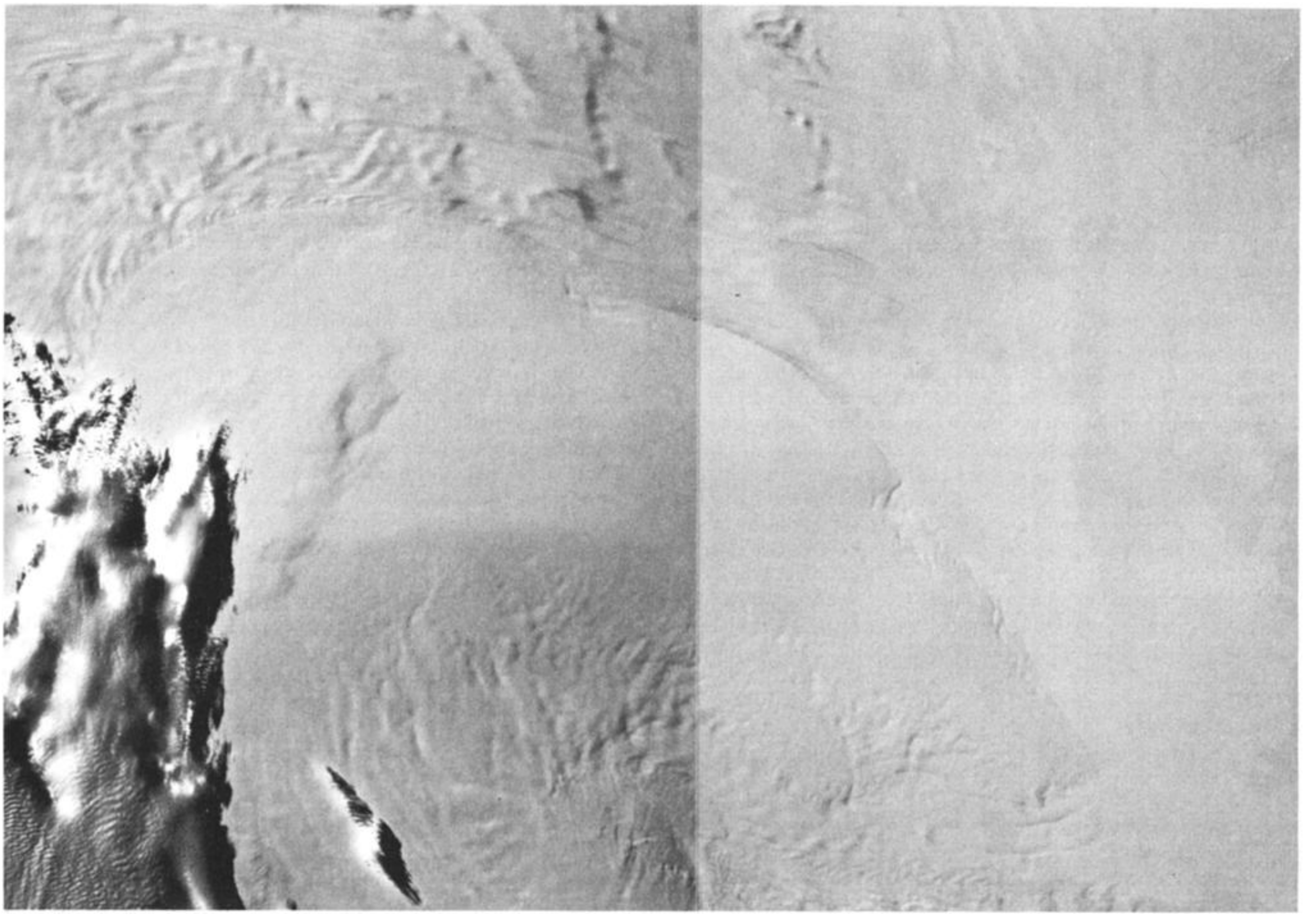

Fig. 1. Composite Landsat image of ice streams E and F. The ice streams are characterized by a "softened" surface appearance and distinct flow lines. Ice stream E is at the top left and center; ice stream F is bottom center; both flow into the Ross Ice Shelf at the right. Images courtesy of B. Lucchitta, U.S. Geological Survey, Flagstaff, Arizona.

a driving stress that is characterized by a peak somewhere within 100 km or so of the coast. The surface profile of Vanderford Glacier is much lower, presumably because of the very deep subglacial trough. Nevertheless, the profile of driving stress shows the same characteristic peak as the others; it is particularly similar to that of Shirase Glacier despite the strongly differing surface and elevation profiles.

The form of the driving-stress curve can be generalized as shown in Figure 9, in which the driving-stress profiles for the Syowa Station sheet-flow line and Shirase Glacier are compared. The characteristic relative maximum in driving stress for East Antarctic ice streams was pointed out by Cooper et al. [1982].

The Lambert Glacier profile is different because of the extremely low elevations for the first 180 km of its length. Comparison of the ice thickness and surface elevation profiles suggests that this whole section of the ice stream is afloat. W. F. Budd (personal communication, 1986) believes that the surface elevation is slightly above buoyancy, and he reports that there are irregularities in the surface that appear to reflect grounding. Nevertheless, it is clear that this section of the ice stream is being affected at most only very weakly by basal shear stress and that its dynamics must be ice shelf like rather than ice stream like. Thus the 180-km point appears to act as an effective grounding line.

In Figure 10 the Lambert Glacier profiles, redrawn to start from the 180-km point, are compared with the profiles for Byrd Glacier [Brecher, 1982; Hughes, 1975]; Shirase Glacier also is shown as a reference from Figure 8. The surface profiles for Byrd and Lambert glaciers look rather similar -- both show substantially lower profiles than Shirase Glacier, presumably because of their much deeper subglacial beds. Even with the lower surface slopes, the large ice thicknesses near the coast lead to driving stresses that peak within the first 50 km of the glaciers, whereas a more gradually rising driving stress characterizes the thinner Shirase Glacier.

A strongly contrasting picture is provided by the West Antarctic ice streams that feed the Ross Ice Shelf, which I will henceforth refer to as the "Ross ice streams." Smoothed profiles for three of them are shown in Figure 11. These ice streams are characterized by very low surface elevation profiles, low-slope beds, and very low driving stresses that increase continually inland to the heads of the ice streams. (Cooper et al. [1982] have shown that the driving stresses peak

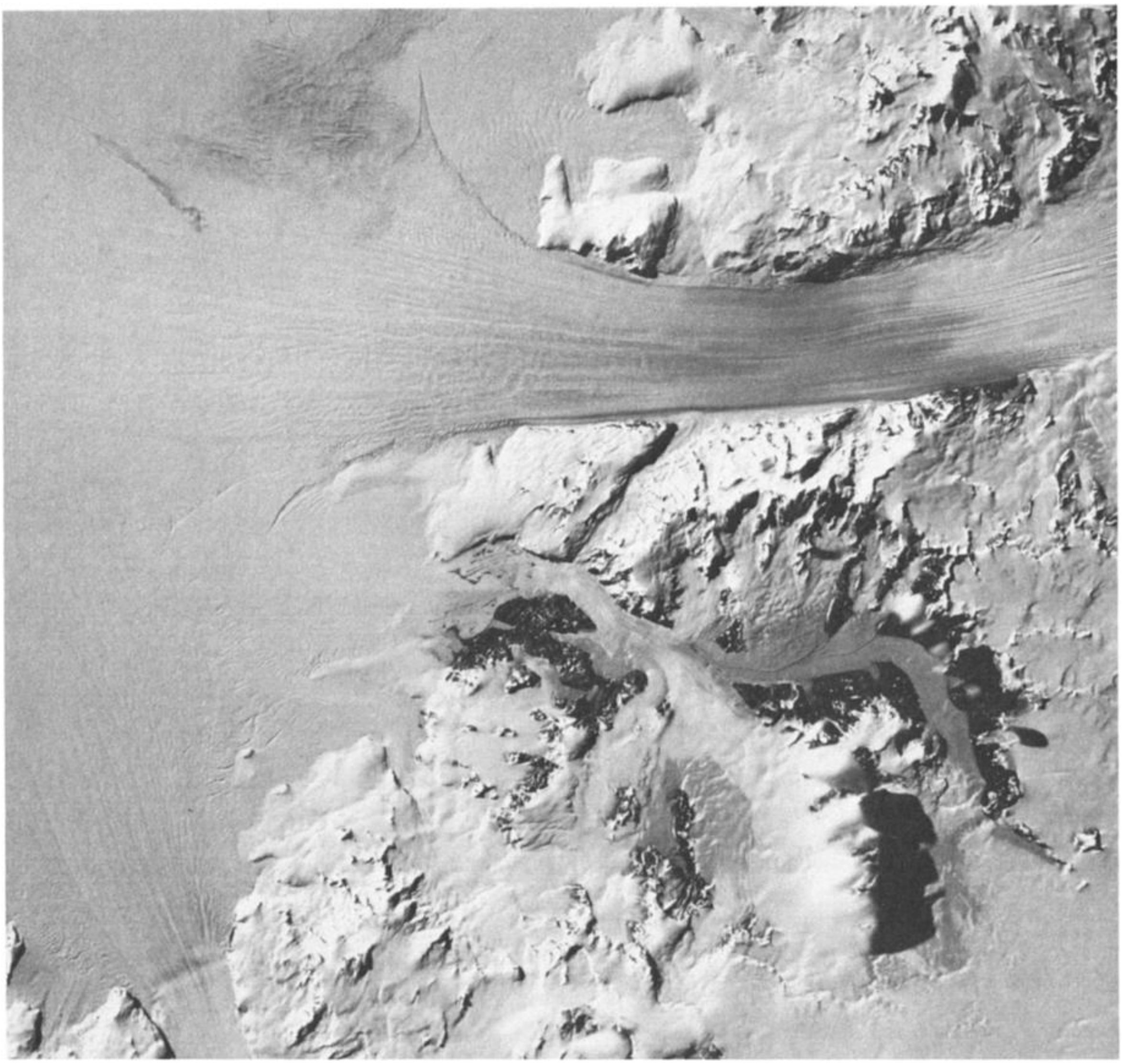

Fig. 2. Landsat image of Byrd Glacier, flowing from the East Antarctic plateau on the right through the Transantarctic Mountains to the Ross Ice shelf on the left. Also visible are Darwin Glacier (center) and Mulock Glacier (bottom left). Image courtesy of B. Lucchitta, U.S. Geological Survey, Flagstaff, Arizona.

there and diminish with closer approach to the ice divide.) Clearly, the Ross ice streams are strikingly different in character, and by implication different in dynamics, from the East Antarctic ice streams and outlet glaciers.

The Ross ice streams are different also from other West Antarctic ice streams. In Figure 12 profiles are shown for Thwaites Glacier and Pine Island Glacier, which flow into the Amundsen Sea, and for ice stream B and East Antarctic sheet flow. The Amundsen Sea ice streams are intermediate in surface profile, presumably reflecting the generally lower elevations of the West Antarctic ice sheet compared with East Antarctica. Nevertheless, they show the peak in driving stress near the coast that is typical of the East Antarctic glaciers but not the Ross ice streams.

Further differences appear in transverse profiles. Cross sections of most of the same glaciers together with velocities and fluxes are shown in Figure 13. All of the ice streams except the Ross group show deep subglacial valleys. The Ross ice streams are broader in comparison to their depth, show little transverse subglacial topography, and in some cases, such as the right-hand (grid north) boundary of ice stream D, do not even extend laterally to the edge of what little subglacial valley does exist.

Thus the Ross ice streams are significantly different in longitudinal and transverse profiles and in basal driving stress from other outlet glaciers and ice streams. On the other hand, the velocities and fluxes for these ice streams (Figure 13) are not remarkable. The velocities are moderate compared with the others, and the fluxes, although among the largest of any of the ice streams, are not exceptional. In fact, although ice stream B carries the largest flux, Lambert Glacier has a larger output than ice streams D and E [Allison, 1979; Shabtaie and Bentley, 1987].

In anticipation of a look at dynamic models

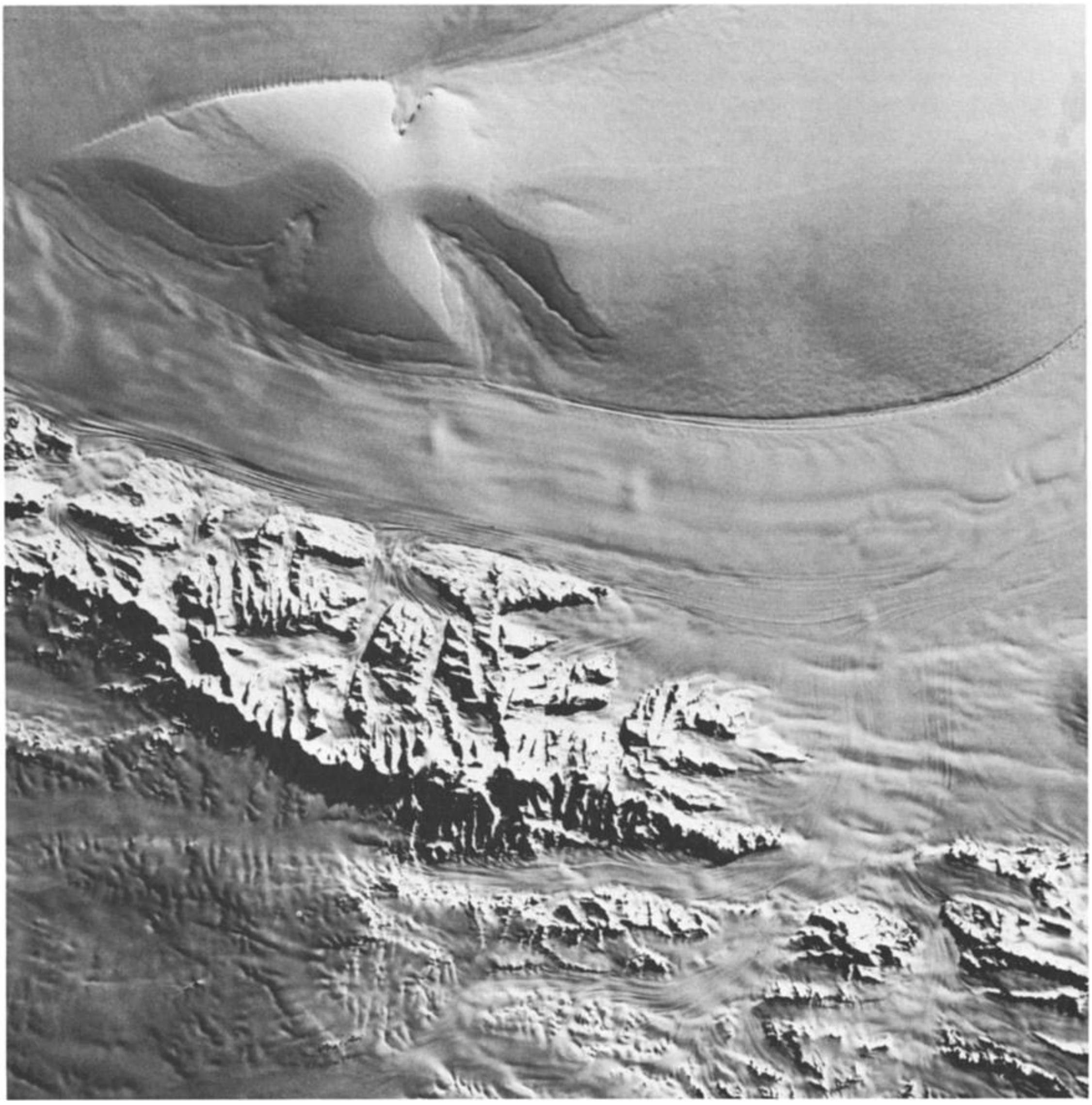

Fig. 3. Landsat image of Rutford Ice Stream, flowing from left to right between the solid rock wall of the Sentinel Range on one side, and low lying ice-covered Fletcher Promontory on the other. Is this an ice stream or an outlet glacier? Image courtesy of B. Lucchitta, U.S. Geological Survey, Flagstaff, Arizona.

of the Ross ice streams, it is worth considering a few of their other properties. From a detailed airborne radar-sounding survey, Rose [1979] showed that the ice streams are characterized, and indeed can be mapped, by their strong surface "clutter." From detailed work on the ground [Shabtaie and Bentley, 1987] it is known 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 crevasse systems that denote zones of very strong shear. These show visually in many places but in some, as along ice stream C, they also are buried. A remarkable feature of the Ross ice streams is that ice stream C has become inactive [Rose, 1979] -- apparently about 250 years ago, as estimated from the depth of burial of surface crevasses [Shabtaie and Bentley, 1987]. A successful dynamic model for the Ross ice streams must be able to explain this striking fact.

McIntyre [1985] has shown that one of the characteristics at the head of at least some fast moving glaciers is a short segment of steep surface slope that coincides with a bedrock step and with the transition from sheet flow to ice stream flow. This transition zone demarks a change in the general surface terrain from irregular on the upstream side to smooth and softened on the downstream side, and also a change in the strength and characteristic of the radar echo from weaker and more scattered upstream to stronger and more coherent downstream. If a similar characteristic exists at the heads of the Ross ice streams, it has yet to be discovered.

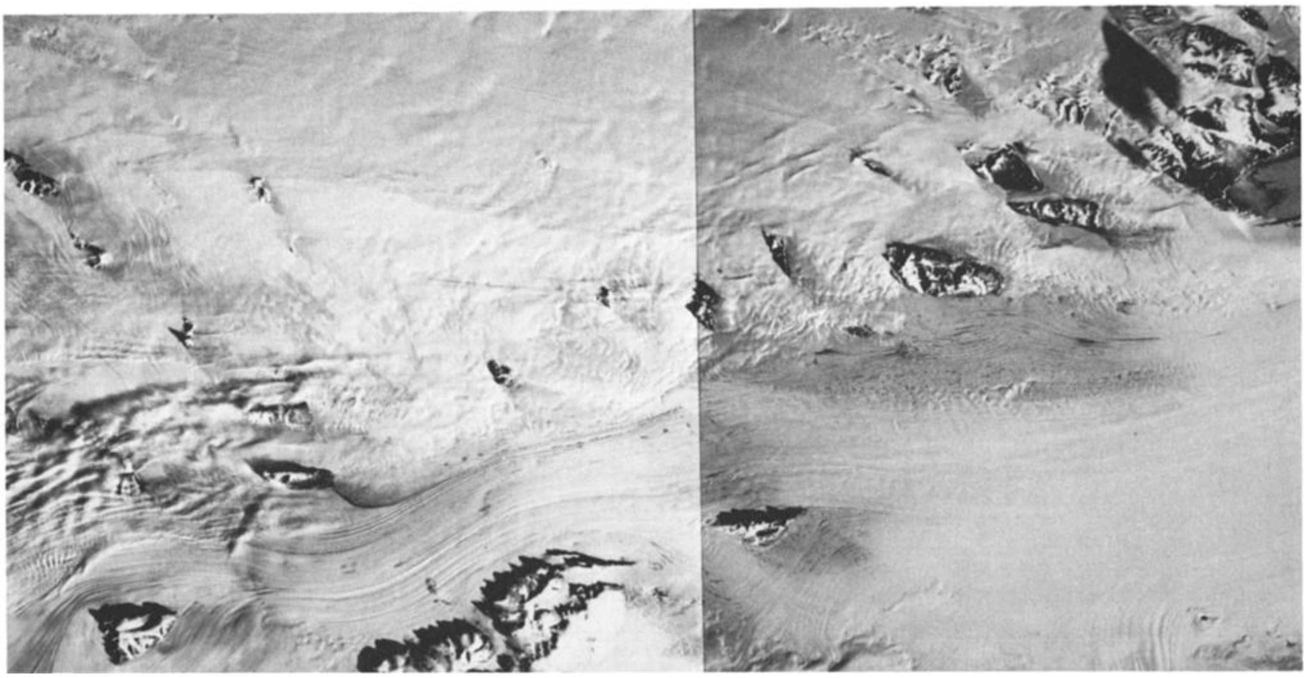

Fig. 4. Composite Landsat image of Lambert Glacier, flowing through the Prince Charles Mountains into the Amery Ice Shelf at the right. The grounding line is just at the right-hand edge of the photograph. Note the complex and varying nature of its boundaries. Images courtesy of B. Lucchitta, U.S. Geological Survey, Flagstaff, Arizona.

Dynamic Models

I will now undertake a brief review of the dynamic models that have been applied to the Ross ice streams, either individually or as part of the West Antarctic ice sheet. This will entail primarily a look at models for the fast sliding of glaciers. Throughout this discussion I implicitly assume that τ_b, the basal shear stress, is not significantly different from τ_d. Alley et al. [this issue (a)] and Whillans [1987] have both shown that this is a good approximation for ice stream B.

Weertman [1957], in his original paper on the physics of glacier sliding, pointed out that fast sliding could result from the presence of a subglacial water layer thicker than the controlling obstacle height. Presumably that would apply also to the more realistic beds that were analyzed by Nye [1970] and Kamb [1970]. In Weertman's basic model the sliding speed, u_b, is proportional to $\tau_b{}^2$ if the exponent in the Glen flow law is 3; however, if the subglacial water layer thickness exceeds the controlling obstacle height, then motion is by enhanced creep around large obstacles only, so $u_b \propto \tau_b{}^3$ [Weertman, 1964].

Lliboutry [1959, 1968, 1971] introduced into sliding theory the effect of cavities that form downstream from bedrock obstacles, particularly at fast sliding speeds. He found that these cavities, particularly when water filled, should lead to a double-valued relationship between τ_b and u_b. For fast flow, the effective pressure N_e (difference between glaciostatic overburden pressure and water pressure at the bed) is the strongly dominating term in the expression for τ_b; in fact, $\tau_b \simeq 2.75rN_e$, where r is the roughness [Lliboutry, 1968]. In this case, the sliding reduces to a case of simple Coulomb friction, in which the ratio τ_b/N_e gives the equivalent coefficient of friction and the speed u_b is independent of τ_b/N_e. Unfortunately, this very independence makes Lliboutry's theory difficult to use in modeling. For the average roughness used by Lliboutry [1971], the equivalent coefficient of friction is 0.3. Recent measurements of ice thickness and surface slope at camp "UpB" on ice stream B [Shabtaie and Bentley, 1987; Shabtaie et al., this issue] yield $\tau_d = 0.15 \times 10^5$ Pa, and Blankenship et al. [1986, this issue] imply $N_e = 0.5 \times 10^5$ Pa, so, remarkably enough, the measured ratio is also 0.3. However, the bed at UpB seems to be very different from the bed assumed by Lliboutry [1971], [Alley et al., 1986, this issue (a); Blankehship et al., 1986, this issue; Rooney et al., this issue], and furthermore Lliboutry's more recent, more complex analysis no longer appears to lead approximately to a Coulomb friction law [Lliboutry, 1979, equation 110], so the significance of this agreement is not clear.

Budd [1975] showed that glaciers can be classified fairly well according to the value of $u_s\tau_b$, which approximates the rate of basal heat generation; u_s is the velocity at the surface. Glaciers for which $u_s\tau_b \ll 50$ MPa m yr^{-1} are ordinary, nonsurging glaciers. A value of $u_s\tau_b \gg 50$ MPa m yr^{-1} characterizes fast Greenland glaciers, whereas $u_s\tau_b \simeq 50$ MPa m yr^{-1} is

Fig. 5. Landsat image of Jutulstraumen, flowing past scattered nunataks on Princess Martha Coast into the Fimbul Ice Shelf at the top. Image courtesy of B. Lucchitta, U.S. Geological Survey, Flagstaff, Arizona.

typical of surge-type glaciers (Figure 14). For UpB, however, the value of this product is only about 7 MPa m yr^{-1}. Thus there is the suggestion that something is different about ice stream B, and very likely other Ross ice streams as well.

Behind the concept of a critical value of heat generation is the idea that when the critical value is exceeded, the water melted by the heating increasingly lubricates the bed. Budd and McInnes [1974] and Budd [1975], following a suggestion of Budd and Radok [1971], used that idea for numerical modeling of surging glaciers. They assumed a reduced basal shear stress given by the form

$$\tau_b{}^* = \frac{\tau_d}{1 + \phi\, u_b\, \tau_d} \qquad (2)$$

where ϕ is a "frictional lubrication parameter" to be selected. The difference between the driving stress τ_d and the basal shear stress $\tau_b{}^*$ is supported by the longitudinal stress, which gives a contribution to u_b through longitudinal strain according to the normal flow law. Drewry [1983] found a good fit for West Antarctic ice streams to (2); to do so, he had to take a value of ϕ that was about 20 times Budd's [1975] preferred value, although still within the range of acceptable values given by Budd [1975].

Early attempts at modeling ice stream movement were aimed principally at the grounding line. Sliding laws were used for the ice stream that did not include explicitly the effect of basal water. Thomas and Bentley [1978] took a Weertman sliding law of the form

$$u_b = k(\tau_b)^m \qquad (3)$$

and carried out their modeling for both m = 2 and m = 3. In succeeding work of the same nature,

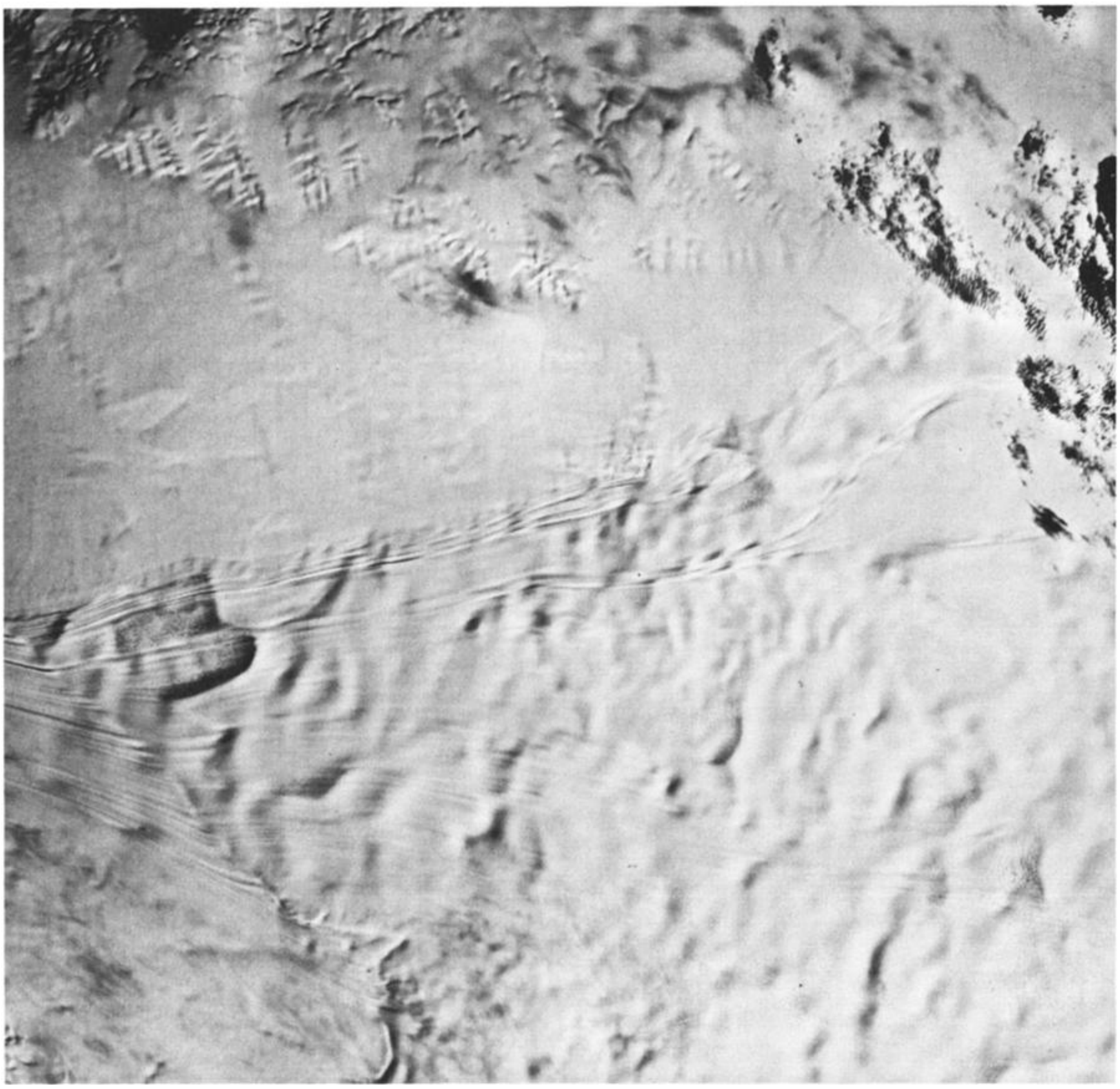

Fig. 6. Landsat image of the upper part of Slessor Glacier, flowing from right to left toward the Filchner Ice Shelf. Although substantial subglacial topography is evident, the ice stream borders are free of rock exposures. The edge of the Shackleton Range just appears in the lower left-hand corner of the photograph. For more information see Marsh [1985]. Image courtesy of B. Lucchitta, U.S. Geological Survey, Flagstaff, Arizona.

Thomas et al. [1979] used (3), but limited their consideration to m = 3. In both cases, k was determined from assuming $\tau_b = \tau_d$, using (1) and taking measured or estimated values of h, α, and u_s at the grounding line; the movement was assumed to be due entirely to basal sliding so that $u_b = u_s$. Hughes [1981] used (3) with m = 2 and adjusted k to fit West Antarctic flow-line profiles in a best average sense.

Hughes [1975] was the first to try to explain in particular the low, concave slopes that typify the Ross ice streams -- he suggested that the ice sheet was surging. He applied Robin and Weertman's [1973] theory, in which subglacial water is forced upstream by a strong gradient in τ_d, and in the scanty evidence available at the time he found a correlation between large gradients in slope and bright radar reflections that Robin et al. [1970] suggested might be subglacial lakes. However, a surging West Antarctic ice sheet would have to be grossly out of balance, whereas information available even then suggested that such an imbalance was not consistent with the input and output fluxes. Later, Hughes [1977] also suggested that fast movement near the grounding line probably came from progressive decoupling of the ice from its bed caused by a downstream thickening of a subglacial water layer.

Weertman and Birchfield [1982] pointed out that continuing fast flow in the Ross ice streams is consistent with mass balance. They then carried out an analysis into which they introduced the thickness of the water layer explicitly. By combining the effects of geothermal flux and heat of sliding, they calculated a basal water thickness of the order of 10 mm at the grounding line.

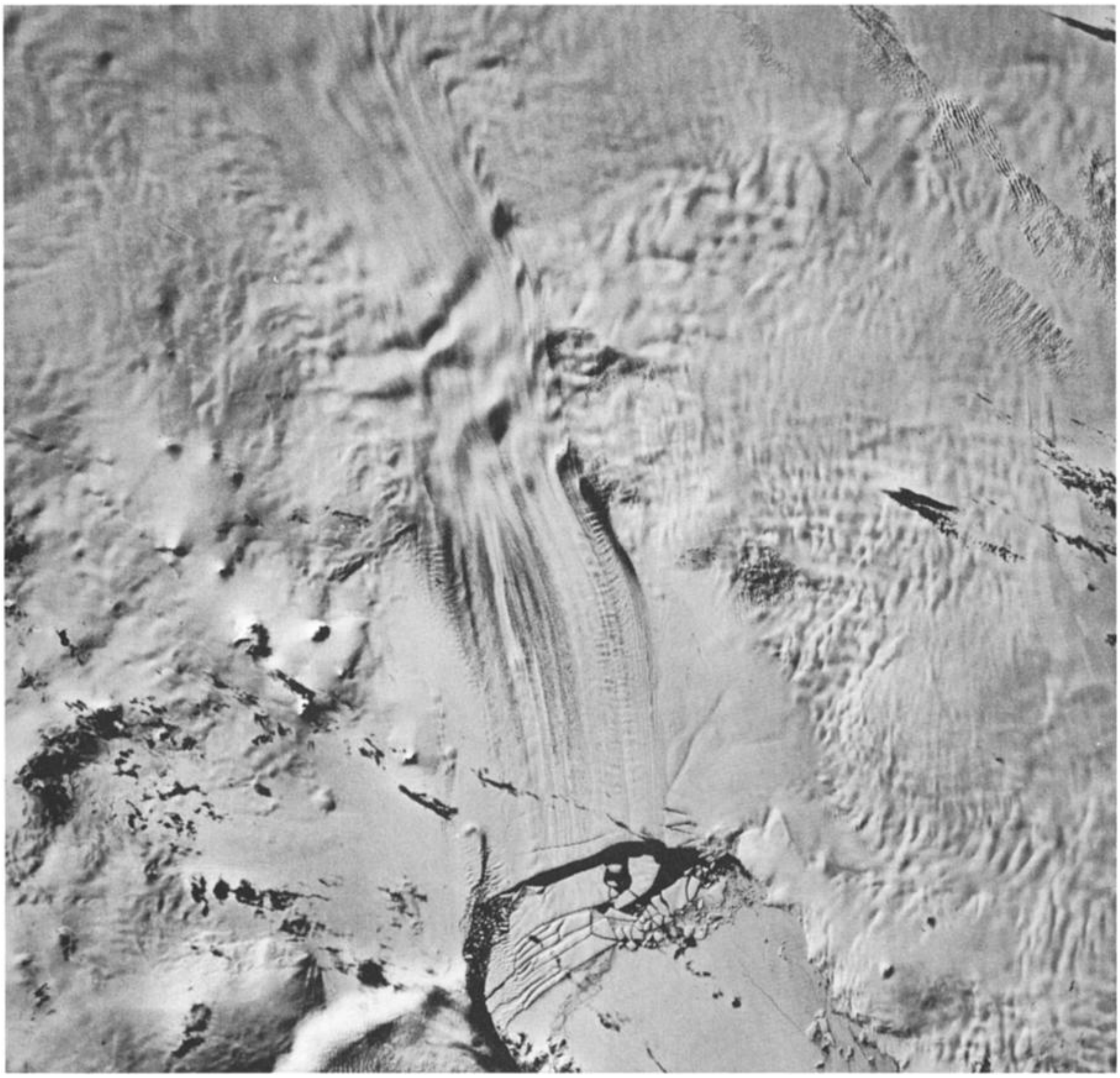

Fig. 7. Landsat image of Pine Island Glacier, flowing from top to bottom into a small ice shelf in Pine Island Bay. Flow lines carry clearly across the ice shelf. Image courtesy of B. Lucchitta, U.S. Geological Survey, Flagstaff, Arizona.

For a subglacial water thickness, d, greater than the controlling obstacle size, d_c, in Weertman's sliding theory, the sliding velocity relationship simplified to the form $u_b = (u_b)_o \, 10d/d_c$, where $(u_b)_o$ is the velocity that is obtained from the normal sliding relation. For d_c = 2-5 mm, d = 10 mm, and $(u_b)_o \simeq$ 20 m/yr, this yields u_b = 400-1000 m yr^{-1}, which is the right order of magnitude. For the inland end of ice streams, however, Budd and Jenssen [1987] calculate a typical meltwater-layer thickness of only a fraction of a millimeter. This would not lead to any significantly enhanced velocity, yet on ice stream B, the high velocity typical of the grounding line [Bindschadler et al., 1987] is found to diminish little for at least 200 km upstream [Whillans et al., this issue].

Weertman and Birchfield [1982] also calculated an equilibrium width (from ice divide to grounding line) for a fast sliding ice sheet of about 900 km, which is the right magnitude for West Antarctica. Equilibrium is based on a balance between the water production at the bed and the sliding speed, which are interdependent. However, the equilibrium width was unstable -- any increase in width would cause the ice sheet to expand, and any decrease would cause it to contract. Furthermore, even if their model ice sheet had the equilibrium width, it apparently would still be out of balance at all points upstream of the grounding line.

Another approach to quantifying the effect of basal water on sliding is to introduce the effective normal pressure, N_e, into the sliding law, as was originally done by Lliboutry [1959]. The recent impetus for this approach was provided by Budd et al. [1979], who found from laboratory experiments that sliding velocities could be expressed by an equation of the form

$$u_b = k \frac{\tau_b^{\,p}}{N_e^{\,q}} \tag{4}$$

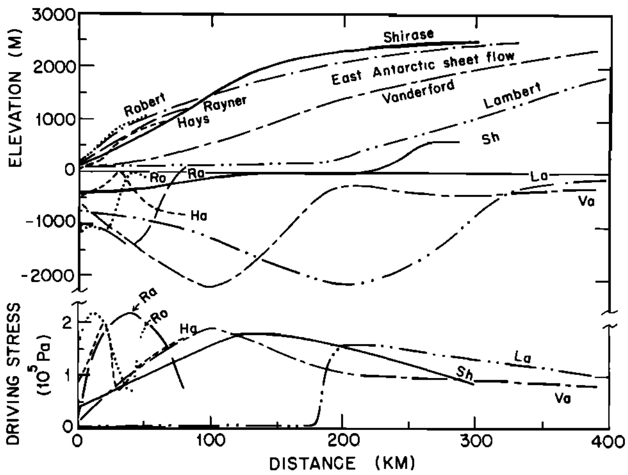

Fig. 8. Profiles of surface and bedrock elevations and driving stress for several fast moving East Antarctic glaciers. Surface profiles are labeled with the glacier name, bottom profiles and driving-stress curves with the corresponding two initial letters. Data are from Mae [1979] and Wada and Mae [1981] for Shirase Glacier; Meier [1983], Hays Glacier; Morgan et al. [1982], Robert and Rayner glaciers; Morgan and Budd [1975], Lambert Glacier; and Budd and Young [1979], Vanderford Glacier. Also shown for comparison is an East Antarctic sheet-flow surface profile inland from Syowa Station [Shimizu et al., 1978].

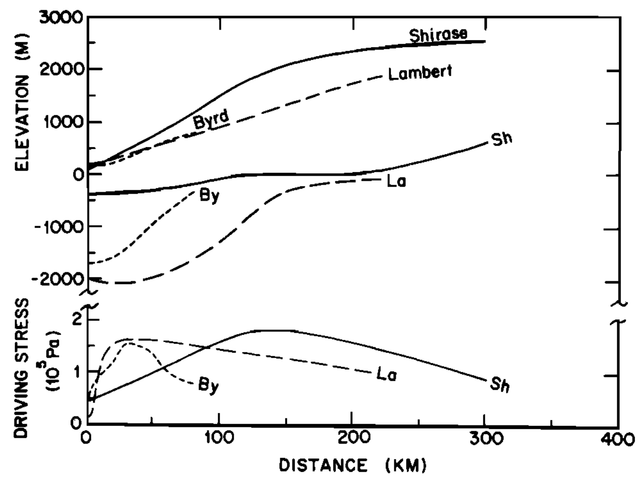

Fig. 10. Profiles of surface and bedrock elevation and driving stress for Byrd, Lambert, and Shirase glaciers. The Lambert Glacier profiles start at the effective grounding line seen at 180 km along the profiles in Figure 8. The Byrd Glacier data are from Hughes and Fastook [1981] and Brecher [1982].

in which the most appropriate values of p and q were p = 3, q = 1 at high normal stresses and p = 1, q = 2 at low normal stresses.

Bindschadler [1983] took N_e into account by introducing the bed separation index $I = \tau_b/N_e$. Since, according to sliding theory [Kamb, 1970], the mean deviation of the normal stress on the glacier bed from its average value is proportional to τ_b, I is a measure of the likelihood that locally the water pressure will actually exceed the overburden pressure and separation will occur. At relatively low sliding speeds, u_b can be expected to increase with I, but at high sliding speed, when cavitation is extensive, I, according to Lliboutry's [1968] theory (as mentioned above), is simply the effective coefficient of friction and u_b is nearly independent of I. (I also gives the minimum stoss slope on subglacial bumps needed to keep the glacier from simply sliding up and over those bumps under the push of the basal shear stress [Iken, 1981], so with increasing I, an increasing fraction of bumps may become ineffective impediments to sliding.)

Bindschadler [1983] showed from measurements and calculations for Variegated Glacier and Columbia Glacier that the sliding velocity does increase as I increases. For ice stream B he calculated a downstream diminishment in N_e, using the theory of Röthlisberger [1972], which could help to maintain the fast sliding despite a downstream decrease in τ_b. For the position of "UpB" camp, N_e calculated in this way would be about 5 x 10^5 Pa (5 bars), about 10 times the seismically determined value cited above.

Bindschadler [1983] did not use I directly to calculate velocities. Instead, he used the good presurge data from Variegated Glacier to test four different sliding theories and found a best fit to that glacier for a relation of the form

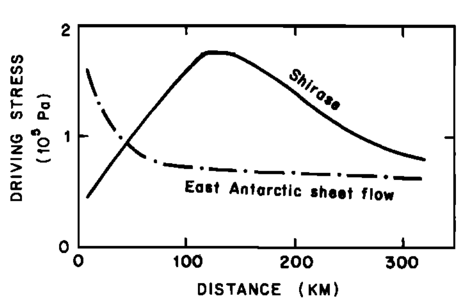

Fig. 9. Comparison of driving-stress curves for Shirase Glacier and the neighboring sheet-flow profile through Syowa Station. From Mae [1979].

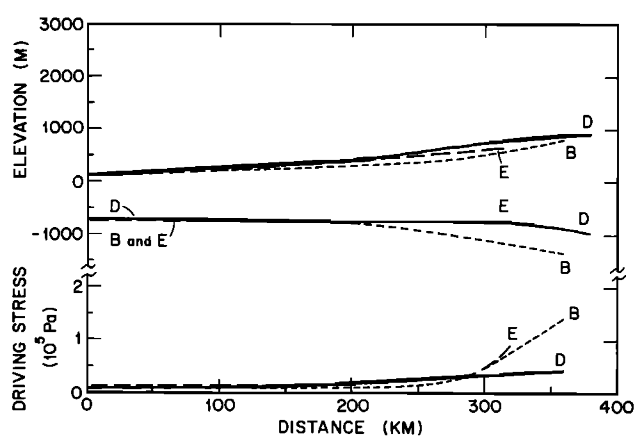

Fig. 11. Profiles of surface and bedrock elevations and driving stress for West Antarctic ice streams B, D, and E. Taken from Lingle [1986].

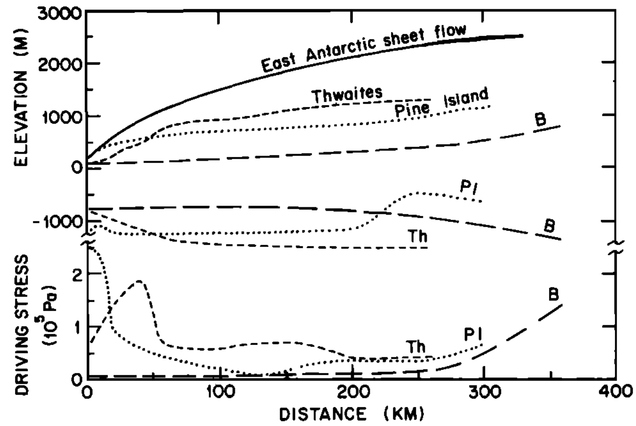

Fig. 12. Profiles of surface and bedrock elevations and driving stress for Pine Island and Thwaites glaciers [Lingle, 1986]. Profiles for ice stream B and the sheet-flow line through Syowa Station are included for comparison.

$$u_b = k \frac{\tau_b^3}{N_e} \tag{5}$$

i.e., (4) with p = 3, q = 1. Since velocity data were lacking, he did not attempt to fit a sliding law to ice stream B. (Equation (5) does not work for ice stream B; with the numerical values of τ and N_e cited above for UpB and k = 8.4×10^{-9} m yr^{-1} Pa^{-2} [Bindschadler, 1983], the calculated value of u_b is only 0.6 m yr^{-1}.)

Budd et al. [1984] carried out numerical modeling of the Ross Embayment section of the West Antarctic ice sheet. They took a sliding law of the form

$$u_b = k \frac{\tau_b}{N_e^2} \tag{6}$$

i.e., (4) with the low-stress values p = 1 and q = 2. Equation (6) was fit on the average to West Antarctic balance velocities corrected for a deformational component. For the effective pressure they took

$$N_e = \rho_i g h^* \tag{7}$$

where h^* is the height of the ice sheet surface above buoyancy; this is equivalent to assuming free communication between the subglacial water and the sea. Using this model they were able to reproduce the actual velocities along the grounding line fairly well, but the calculated velocities farther inland were much less than actually observed on ice stream B.

In two recent models, Budd and his collaborators have modified the denominator in (4) by the introduction of another adjustable parameter to provide a closer fit to the balance velocities. McInnes and Budd [1984] use an expression of the form

$$u_b = k_1 \frac{\tau_b}{N_e + k_2 N_e^2} \tag{8}$$

The adjustable constants k_1 and k_2 were calculated by fitting a flow line across West Antarctica incorporating Pine Island Glacier and ice stream B. A two-dimensional instead of a three-dimensional model was used to provide a much closer grid point spacing than was computationally practicable for a grid covering a large area. N_e again was calculated from the height above buoyancy (equation (7)). The modeling results

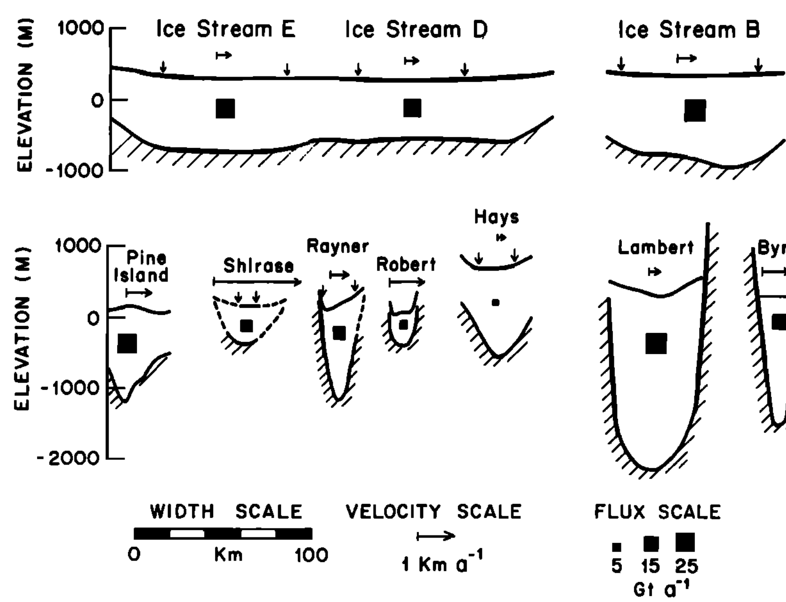

Fig. 13. Transverse sections, showing also glacier speeds (arrows) and fluxes (solid squares). Speed and flux data are from Allison [1979], Lambert Glacier; Brecher [1982], Byrd Glacier; Fujii [1981], Shirase Glacier; Meier [1980, 1983], Hays Glacier; Morgan et al. [1982], Rayner and Robert glaciers; Crabtree and Doake [1982], Pine Island Glacier; and Shabtaie and Bentley [1987], ice streams B, D, and E.

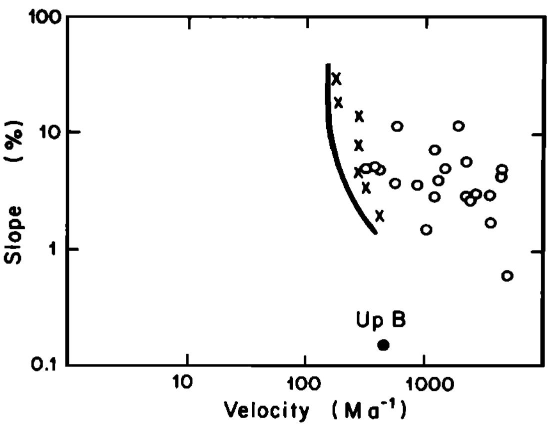

Fig. 14. Plot of surface slope versus velocity showing the grouping into fast moving Greenland glaciers (open circles), surging glaciers (crosses), and the approximate boundary for fast flow (solid line). Ordinary glaciers (not shown) lie to the left of the line. A measurement for ice stream B (solid dot) is also shown. Extended from Budd [1975].

showed a good, stable reproduction of the actual surface elevation profile of the ice sheet along that flow line. Jenssen et al. [1985] take as a sliding law

$$u_b = k_1 \frac{\tau_b}{(N_e + k_2 N_e^2)^2} \tag{9}$$

and adjust the parameters to fit the balance velocities in the Ross Embayment again in an areal sense; they find a small improvement in the fit over that obtained by Budd et al. [1984] using (6).

Rather than fitting a sliding law like (4) with a fixed value of k, several models allow it to vary along a flow line.

Lingle [1984] used a generalized form of (4) in which the effect of basal water was carried implicitly by a function, f(x), that could be tuned to fit the balance velocities on a particular ice stream. His equation was

$$u_b = \frac{k}{h} f(x)\tau_b^3 \tag{10}$$

the exponent 3 was chosen in accordance with the field observations of Kamb [1970] and, as mentioned above, is appropriate to movement by enhanced creep around large obstacles on the bed. (Beneath a fast moving ice stream the small obstacles may be either abraded off or covered by water.)

Lingle [1984] applied (10) to ice stream E in an analysis of how that ice stream would change through time. He found that to produce a fit to the current balance velocities for ice stream E itself, f(x) need vary only by a factor of 2 or 3, but to fit balance velocities all the way to the ice divide required a decrease in f(x) of 4 orders of magnitude. Obviously, some important physical process is hidden in f(x). In a similar analysis, Morland et al. [1984] found that for sheet flow in Greenland and Devon Island, f(x) need vary only within a factor of 2 or 3, but the applicability of their result to sliding is unclear, since the Devon Island ice cap, and probably also the relevant section of the Greenland ice sheet, is frozen to its bed (W. S. B. Paterson, personal communication, 1986).

Lingle and Brown [1987] use a sliding law of still another form:

$$u_b = k \frac{\tau_b}{N_e} \tag{11}$$

i.e., (4) with p = 1 and q = 1 or simply $u_b = kI$. They calculate u_b for ice stream B from balance velocities, just as Budd et al. [1984] did. The coefficient k in (11) is taken to be the minimum value needed to assure that the basal water beneath the ice stream is driven up the bed slope to the grounding line, i.e., does not simply accumulate under the ice stream. N_e is then calculated by Darcy's law assuming a permeable aquifer underlying ice stream B and continuing up to the ice divide. They attribute variations in N_e to the thickness of the assumed aquifer.

Finally, a new model based on an entirely different physical picture is presented by Alley et al. [this issue (b)]. This new picture involves a deforming subglacial till in which a linear viscosity is assumed, so that the sliding law becomes

$$u_b = \frac{h_b}{\eta} \tau_b \tag{12}$$

Here η and h_b are the viscosity and thickness of the subglacial layer, respectively. In order to fit the observed velocities, η has to decrease severalfold from the head to the foot of the ice stream.

Although it is the purpose of this review to concentrate on the ice streams themselves, some mention should be made of the junction between the ice streams and the ice shelves into which they flow. This junction, whether or not ice streams are specified for the inland ice, has received much attention because of the idea that the grounding line may be inherently unstable, or at least strongly susceptible to movement [Hughes, 1973, 1975, 1977; Weertman, 1974; Thomas and Bentley, 1978; Thomas et al., 1979; Lingle, 1984; van der Veen, 1985]. In relation to modeling ice stream dynamics, the crucial question is how to couple the ice stream, whose motion is controlled by sliding on its bed, hence by basal shear stress, to the ice shelf, whose motion is controlled by spreading, hence by longitudinal stresses. All models except van der Veen's [1985] have explicitly or implicitly assumed that ice movement at the grounding line is controlled solely by ice shelf spreading, but that movement of the ice stream immediately inland is controlled by basal shear stress. These models all yield an unstable retreat in the absence of any ice shelf, continuing to the complete elimination of the inland ice. Van der Veen [1985] introduces a transition zone in the first 150 km of the ice stream in which the longitudinal stress linearly decreases from its value on the ice shelf to zero. This apparently produces enough of a stabilizing influence, by increasing the response of the ice stream to changes at the grounding line, to prevent more than a moderate retreat of the grounding line in his models, even with no ice shelf.

Another stabilizing influence may stem from the occurrence of a deforming bed beneath an ice stream. Alley et al. [this issue (b)] point out the possibility of a negative feedback mechanism: an increasing rate of ice discharge at the grounding line, which might follow a hypothetical decrease in back stress from the ice shelf, would lead to faster transport of subglacial till, hence a thinning of the till, and therefore would decrease the ice velocity for a given basal shear stress.

Discussion

From this review it is clear that in neither the geographical nor the physical sense can an ice stream yet be defined precisely. It seems likely that ice streams characterized by deeply incised beds are rather like true outlet glaciers despite their lack of visible confining walls. Small East Antarctic ice streams have surface profiles that are not largely different from the standard equilibrium profile for sheet flow ice, although they do exhibit the characteristic concave-up shape near the coast. Driving stresses peak near the coast in both East Antarctic ice streams and outlet glaciers. A large ice stream like Lambert Glacier has a lower profile, presumably because its large flux lowers the surrounding ice level. In configuration, flux, and driving stress, there is little difference between Lambert Glacier and Byrd Glacier, the quintessential outlet glacier.

West Antarctic outlet glaciers are less characterized by deep subglacial troughs than their East Antarctic counterparts, and their surface elevations are lower, probably because of the deep-lying glacial floor in West Antarctica rather than any inherently different character of the ice streams themselves. Pine Island Glacier and Thwaites Glacier are similar to East Antarctic ice streams in ice thickness (~2000 m) and in having driving-stress maxima near the coast.

The Ross ice streams are different in several ways, as has already been noted above. The boundary that separates the West Antarctic inland ice from the Ross Ice Shelf is the only broad coastal stretch that is characterized by a complete absence of mountains and a bed everywhere below sea level. The ice streams, probably as a result, are broad compared with their thickness and have only slight subglacial depressions in many places, and in some instances the ice stream boundary does not even coincide with the edge of the subglacial depression. 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 subglacial troughs beneath the Ross ice streams is an indication of their youthfulness. Perhaps the continual movement of the grounding line over the recent geologic past has given the ice streams little time to erode their beds in one location -- it is noteworthy, in this regard, that the subglacial terrain of the Ross Embayment shows no physiographic feature corresponding to the grounding line [Bentley and Jezek, 1981]. Another possible explanation for the shallowness of the troughs is that erosion by a deforming subglacial till may be much slower than erosion by sliding [Alley et al., this issue (a)].

The Ross ice streams are further notable for their monotonically decreasing driving stresses from their heads to their grounding lines. In this regard they are unique among ice streams and outlet glaciers for which measurements exist.

All the numerical models that have been applied to the West Antarctic ice sheet and its ice streams, except for that of Alley et al. [this issue (b)], suffer from a dual, crucial, problem -- lack of measurements of the physical conditions at the bed and lack of a valid physical sliding law to use even if the conditions were known. (Alley et al. [this issue (b)] do know at least some of the physical conditions at one point on the bed, but their proposed sliding law is completely untested.) The consequences of that fact can be demonstrated dramatically by inserting the recently measured values of τ_b, u_b, N_e, and h^* for UpB on ice stream B into the various equations that have been used for modeling (Table 1). Two calculations have been made for each sliding law that includes N_e, one with the definition of N_e used by the original author(s) (where N_e was defined by (7), the measured value $h^* = 250$ m for UpB was used) and the other with the seismically determined value of N_e [Blankenship et al., 1986, this issue]. The measured value of u_s is 443 m yr^{-1} [Whillans et al., this issue].

The sliding law from Bindschadler [1983] was never intended nor claimed to be applicable to ice stream B. It is included in the table to show the very low sliding speeds that are implied by a dependence of u_b on $\tau_b{}^3$ together with the low driving stress on ice stream B, even for the small measured value of N_e, when the model is tuned to a nonsurging valley glacier. But the other models were all tuned specifically to the West Antarctic ice sheet, yet most fit poorly even with the authors' own definitions of N_e, and the low measured value of N_e yields absurdly high speeds. Of course, the fit would have been far better if the measured value of N_e had been used in the tuning process, but the point is that models that are not based on a sound, physically based sliding law can at best be used only very narrowly within the specific range of parameters to which they are tuned. They cannot be used with any confidence for extrapolation in time or space.

The use of height above buoyancy to determine the effective pressure is also not very satisfactory, except close to the grounding line. Whatever the form of the basal water connection to the sea, some hydraulic pressure gradient will exist; in general, it should increase the basal water pressure substantially above hydrostatic. Equation (7) therefore should yield an upper limit to N_e, but in most places N_e should be substantially less. For example, at UpB, (7) yields $N_e = 25 \times 10^5$ Pa, whereas N_e according to Bindschadler's [1983] calculation would be 5×10^5 Pa, and the measured value from the seismic measurements is 0.5×10^5 Pa [Blankenship et al., 1986, this issue].

TABLE 1. Calculated Values of Sliding Speed for Ice Stream B According to Various Models Using the Surface Slope and Ice Thickness Measured at UpB Camp

	Authors' Parameter Values			u_b for UpB, m yr^{-1}		
Equation	k or k_1	k_2, Pa^{-1}	N_e, 10^5 Pa	Using Authors' Value of N_e	Using Measured Value of N_e	Reference
$u_b = k\tau_b^2$	1.25×10^{-6} m yr^{-1} Pa^{-2}			280		Thomas and Bentley [1978]
$u_b = k\tau_b^3$	8×10^{-11} m yr^{-1} Pa^{-3} 8×10^{-10} m yr^{-1} Pa^{-3}			270 2,700		Thomas and Bentley [1978] Thomas et al. [1979]
$u_b = k \frac{\tau_b^3}{N_e}$	8.4×10^{-9} m yr^{-1} Pa^{-2}		5	0.06	0.6	Bindschadler [1983]
$u_b = k \frac{\tau_b}{N_e^2}$	5×10^{9} m yr^{-1} Pa		25	12	32,000	Budd et al. [1984], also van der Veen [1985]
$u_b = k \frac{\tau_b}{N_e}$	4,670 m yr^{-1}		3.5	200	1,400	Lingle and Brown [1987]
$u_b = \frac{k_1 \tau_b}{N_e + k_2 N_e^2}$	1.5×10^{4} m yr^{-1}	2.5×10^{-7}	25	55	4,400	McInnes and Budd [1984]
$u_b = \frac{k_1 \tau_b}{(N_e + k_2 N_e^2)^2}$	1.5×10^{11} m yr^{-1} Pa	3.5×10^{-7}	25	100	870,000	Jenssen et al. [1985]

Two calculations are presented where relevant, one using the definition of N_e preferred by the author(s) of the model, and one with the seismically determined value $N_e = 0.5 \times 10^5$ Pa. Other numerical values used are $\tau_b = 1.5 \times 10^4$ Pa and $h^* = 250$ m (cf. (7)). The measured value of u_b is 443 m yr^{-1}.

Measurements on ice stream B show little difference in speed between stations UpB [Whillans et al., this issue] and DnB 200 km downstream [Bindschadler et al., 1987] despite a factor-of-two downstream decrease in the driving stress. This strongly implies a low power dependence of u_b on τ_b. At the same time, the small value of N_e at UpB implies a low power dependence on N_e also. Thus taking p = 1 and q = 1 in (4), as Lingle and Brown [1987] did, might seem appropriate. The difficulty, however, is that there is neither theoretical nor experimental support for a sliding law of the form $u_b = kI$.

A subglacial deforming till may provide an escape from this dilemma. If the model of Alley et al. [this issue (b)] should prove to be conceptually correct, $u_b \propto \tau_b$ because of simple linear viscosity in the till. N_e does not appear explicitly, but presumably is a factor in determining the viscosity.

From the work presented in this conference and the important field projects that are under way in several places now, we may hope to approach a better understanding of ice streams and the dynamic principles that control them. Only then will it be possible to predict the future of the Antarctic and Greenland ice sheets, and understand the behavior of the great ice sheets of the past.

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