

## THE DYNAMICS OF ICE-SHEET OUTLETS

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**ABSTRACT.** A comparison of data from aircraft altimetry, Landsat imagery, and radio echo-sounding has shown characteristic surface topographies associated with sheet and stream flow. The transition between the two is abrupt and occurs at a step in the subglacial topography. This marks the onset of basal sliding and high velocities caused by subglacial water; it results in crevassed amphitheatre-like basins round the head of outlet glaciers. It is also the zone of maximum driving stress beyond which values decline rapidly as velocities increase. This abrupt transition appears to be topographically controlled since basal temperatures are at the pressure-melting point well inland of the change in regime. The Marie Byrd Land ice streams exhibit qualitative differences from other ice-sheet outlets, however; the change to lower driving stresses is much more gradual and occurs several hundred kilometres inland. Such ice streams have particularly low surface slopes and appear in form and flow regime to resemble confined ice shelves rather than grounded ice. The repeated association of the transition to rapid sliding with a distinct subglacial feature implies a stabilizing effect on discharge through outlet glaciers. Acceleration of the ice is pinned to a subglacial step and propagation of high velocities inland of this feature seems improbable. Rapid ice flow through subglacial trenches may also ensure a relatively permanent trough through accentuation of the feature by erosion. This is concentrated towards the heads of outlet glaciers up-stream of the region where significant basal decoupling occurs. This may be a mechanism for the overdeepening of fjords at their inland ends and the development of very steep fjord headwalls.

**RÉSUMÉ.** La dynamique des émissaires de calotte polaire. Une comparaison des données d'altimétrie aérienne, des images de Landsat et de sondages radar a montré des topographies caractéristiques de surface associées à la calotte et l'écoulement. La transition entre les deux est abrupte et intervient à l'occasion d'un décalage de la topographie sous-glaciaire. Ceci marque le début du glissement basal et des vitesses élevées dues à l'eau sous glaciaire et a pour conséquence un bassin crevassé en amphithéâtre au début de l'émissaire. C'est aussi la zone du maximum des contraintes, au delà elles diminuent rapidement en même temps que les vitesses croissent. Cette transition abrupte semble être contrôlée par la topographie puisque la température à la base est un point de fusion bien plus loin à l'intérieur que le changement de régime. Néanmoins, les courants de glace de la Marie Byrd Land révèlent des différences qualitatives avec les autres émissaires de la calotte. Le passage à des états de contraintes inférieures est beaucoup plus progressif et intervient plusieurs centaines de kilomètres à l'intérieur. Tels courants de glace ont en particulier de faibles pentes de surface et ressemble en forme et écoulement à des shelves

guidés plutôt qu'à de la glace reposant sur le sol. L'association répétée d'un passage à un glissement rapide avec des caractéristiques distinctes à la base du glacier implique un effet stabilisateur en débit par des émissaires. L'accélération de la glace est soumise à une marche dans le lit et la propagation de ces vitesses élevées à l'intérieur est improbable. La possibilité de surges pour la calotte ne semble pas vraisemblable pour de tels émissaires dans les conditions actuelles. De rapides écoulement de glace dans des fossés sous-glaciaires peuvent aussi assurer une association assez permanente de ces fossés par l'érosion. Ceci est concentré au début des émissaires dans des régions où un débitage conséquent à la base intervient. Ceci peut être le mécanisme d'approfondissement des fiords à leur bord continental et le développement de leurs fronts très abrupts. L'association des régimes d'écoulement et de la topographie peut se révéler utile à l'explication des dépôts des anciennes calottes de glace.

**ZUSAMMENFASSUNG.** Die Dynamik von Auslassströmen aus Eisschilden. Ein Vergleich zwischen Daten aus Flugzeug-Profilmessungen sowie aus Landsat- und Radar-Echoaufnahmen hat Bildern charakteristische Oberflächenformen, verbunden mit dem Eisfluss in Schilden und Strömen, gezeigt. Der Übergang zwischen beiden ist abrupt und tritt an einer Stufe im Untergrund ein. Er markiert das Einsetzen des Gleitens am Untergrund und hoher Geschwindigkeiten infolge subglazialen Wassers und führt zu spaltendurchsetzten, amphitheatralischen Becken rund um die Ursprünge von Auslassgletschern. Dort ist auch die Zone grösster Treibkraft, unterhalb derer die Werte bei zunehmenden Fließgeschwindigkeiten schnell abfallen. Dieser abrupte Übergang scheint topographisch bedingt zu sein, da die Temperaturen am Untergrund schon weit landein vor dem Verhaltenswechsel sich auf dem Druckschmelzpunkt befinden. Jedoch zeigen die Eisströme in Marie Byrd Land qualitative Unterschiede zu anderen Auslassströmen des Eisschildes; der Wechsel zu geringeren Treibkräften ist viel stärker abgestuft und tritt schon einige hundert Kilometer landeinwärts ein. Sie haben ausnehmend geringe Oberflächenneigungen und scheinen in Form und Fluss mehr begrenzten Schelfeisen zu gleichen als aufstizendem Eis. Die immer wieder anzutreffende Verbindung des Übergangs zu schnellem Gleiten mit einer bestimmten subglazialen Erscheinung hat eine stabilisierende Wirkung auf den Abfluss durch Auslassgletscher zur Folge. Die Beschleunigung des Eises ist mit einer subglazialen Stufe verknüpft; die Ausbreitung hoher Geschwindigkeiten landein von dieser Erscheinung erscheint unwahrscheinlich. Schneller Eisfluss durch subglaziale Gräben dürfte ebenfalls eine relative dauerhafte Verbindung durch Verstärkung der Erscheinung infolge Erosion sicherstellen. Dies konzentriert sich gegen die Ursprünge von Auslassgletschern hin, stromaufwärts von jenem Gebiet, wo signifikante Ablösung vom Untergrund auftritt. Man kann darin einen Mechanismus für die Übertiefung von Fjorden an ihren rückwärtigen Enden und für die Entwicklung sehr steiler Fjordrückwände sehen. Die Verknüpfung des Eisflussverhaltens mit der Geländeform mag zur Aufklärung des Abflusses von früheren Eisschilden von Nutzen sein.

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INTRODUCTION

Although only 13% of the Antarctic coastline consists of outlet glaciers and ice streams (Drewry and others, 1982), their importance to the dynamic glaciology of the ice sheet is very much greater than this proportion suggests. It has been estimated that 90% of the accumulation falling inland of the coastal zone is discharged through such features (Morgan and others, 1982). Hughes (1977) identified them as the most "energetic" elements within an ice sheet and hence as exercising the major control on its overall stability. In this paper we shall draw together evidence on the form and flow of outlets in order to understand better their contribution to ice-sheet dynamics. The implications of these aspects of ice-sheet discharge will be considered with regard to their stability, to that of the ice mass as a whole, and in relation to the erosion of subglacial landforms.

ICE FLOW INTO OUTLET GLACIERS AND ICE STREAMS

Previous work

On the basis of physical processes, two major components to ice streams and outlet glaciers may be distinguished. They are typified by converging ice flow and by stream flow. An ice-sheet outlet may display elements of both, but their relative magnitude and importance will vary in response to topographic and glaciological controls.

Sheet flow is thought to be largely the result of motion by internal deformation of the ice mass, as has been shown by the relationship between driving stresses and mean shear strain-rates (Budd and Smith, 1981; Cooper and others, 1982). Similarly, the dynamics of the ice mass down the International Antarctic Glaciological Project flow-line in Wilkes Land in the region before the deep troughs of Vanderford and Totten Glaciers are explicable by internal deformation (Budd and Young, 1979); even where basal melting occurs, bottom sliding is

probably negligible. Flow within ice streams, however, is recognized as being due to basal sliding (Rose, 1979; Weertman and Birchfield, 1982). Drewry (1983[b]) found that data from Marie Byrd Land ice streams were satisfactorily explained by accounting for Budd's (1975) friction lubrication factor, thereby indicating major decoupling of the ice streams by water. However, Bindschadler (1983) demonstrated that this is only the case for near-zero water pressures. Where the subglacial water pressure increases, it is this and the resulting stress concentrations in areas where the ice is still in contact with the bed which control the sliding process.

Converging flow is indicated morphologically by basins and broad valleys in the ice-sheet surface which terminate in stream flow or the outlet itself. Contour maps of East Antarctica (for instance, Drewry, 1983[c]) show clear examples inland of Byrd, David, Ninnis, and Dibble Glaciers, the first extending over 200 km into the ice sheet.

Ice streams are clearly distinguished from adjacent sheet flow by their discrete boundaries which are often curvilinear for several hundred kilometres. Such margins are heavily crevassed shear zones separating the active streams moving at high velocities by basal sliding from the more slowly-flowing plateau ice (Rose, 1979). The ice streams become progressively isolated from adjacent ice masses as a result of shear decoupling at their sides and the development of glide fabrics (Hughes, 1977; Hughes and Fastook, 1981).

The presence of ice streams is often associated with subglacial terrain which causes flow to be channelled. Valleys in the ice-sheet surface have been noted to have corresponding subglacial features (Rose, 1979; Allison and others, 1982). Smooth bedrock echoes of uniform intensity up-stream of the grounding lines of ice streams in Enderby, Kemp, and Mac.Robertson Lands have also been taken to indicate that glacial erosion from high sliding velocities has substantially smoothed the bedrock and contributed to its lowering (Morgan and others, 1982). However, the ice streams of Marie Byrd Land flow in wide and relatively shallow, although smooth, depressions

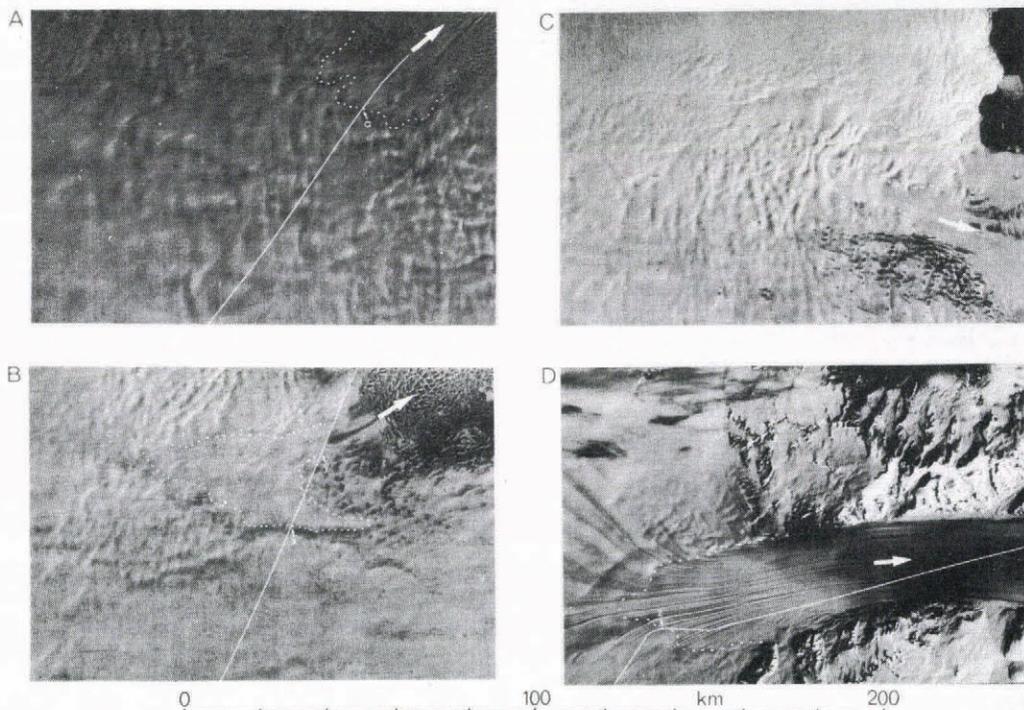


Fig. 1. Landsat images showing the transition in topography of the ice-sheet surface inland of outlet glaciers. Dotted lines separate irregular and smooth terrain types thought to indicate ice flow by internal deformation and basal sliding respectively. Large arrows give flow directions through the outlets. A: Thwaites Glacier (1205-14051); B: Frost Glacier (1446-23272); C: Dibble Glacier (1447-23214); D: Byrd Glacier (1526-18555). Letters, a, x, y, and z refer to points on surface profiles in Figure 2, which are shown as white lines across Thwaites, Frost, and Byrd Glaciers.

(Rose, 1979) and Thwaites Glacier shows no distinctive subglacial channel at all (Drewry, 1983[b]). These subtle landforms in association with such vigorous ice streams may suggest that the ice flow in these areas is either relatively youthful or transient (Drewry and Robin, 1983).

The transition of regime from converging flow to stream flow, a question of central importance in ice-sheet mechanics, is thought to occur abruptly in the region of outlet glaciers (Young, 1981). Budd and Young (1979) suggested that, as ice is channelled into deep subglacial troughs, an order-of-magnitude increase in velocity and an accompanying decrease in shear stress takes place due to the onset of basal sliding. Driving stresses fall from a peak of approximately 200 kPa to zero in the 100 km before the grounding line as the ice accelerates (McIntyre and Cooper, 1983). Budd and others (1979) found from laboratory experiments that this limiting shear stress is a function of roughness, normal stress, and velocity. The subsequent acceleration was thought to be dependent on the reduced effective normal stress (Budd and McInnes, 1978) as the subglacial water layer carries a greater proportion of the weight of the overlying ice.

#### The topography of the transition from converging to streaming flow

The transition in flow regime may be studied further by consideration of ice-sheet topography in the region of outlet glaciers. Landsat imagery shows an abrupt change in surface terrain inland of many outlet glaciers and ice streams, for instance, Thwaites, Frost, Slessor, and Byrd Glaciers (Fig. 1). Although irregular in plan form, this transition separates zones of undulating topography typical of the ice-sheet margin from the smoother, relatively featureless terrain of the outlet glaciers themselves. The boundary between these areas is, in places, heavily crevassed and often appears to be wind-scoured and free of snow, indicating a scarp-like step. Even when individual crevasses are not visible on Landsat imagery, radio echo-sounding shows that consistent cluttering of Z-scope records begins within 2 km of the transition in topography up-stream of Frost and Thwaites Glaciers. Similar crevassed basins have been reported inland of many outlet glaciers in Antarctica (Shumskiy, 1970) and in Greenland (Holtzschcher and Bauer, 1956).

Details of the form of the ice sheet are available in a few instances along flight lines which intersect basins inland of outlet glaciers (Fig. 2). Surface profiles down flow-lines to Frost, Thwaites, and Byrd Glaciers derived from aircraft altimetry with a predicted accuracy of 2 m (McIntyre, unpublished) show distinct features which enable them to be accurately matched with Landsat imagery. In all three cases, the transition between undulating and smooth terrain identified on the imagery corresponds to within 1 km with an abrupt increase in surface gradient. The gradients over these sections are 7.9, 7.7, and 7.2% for Frost, Thwaites, and Byrd Glaciers respectively and hence are between six and eight times the regional gradients of approximately 1.1, 1.2, and 0.9%. These steps are over twice as steep as slopes typically associated with surface undulations in these regions. They are clearly the surface expression of the subglacial topography, which is also shown in Figure 2. Radio echo-soundings show bedrock steps of between 400 and 2200 m directly beneath the steep slopes where these are intersected by flight lines, although the nature of the Byrd Glacier profile is not fully known because of intermittent data. On the basis of the correlation found between radio echo-soundings and Landsat imagery, we can assume that these marked subglacial escarpments occur around much of the heads of subglacial valleys, as can be seen in the fjord terrain of Norway, Baffin Island, and elsewhere.

#### Ice dynamics during the transition from converging to streaming flow

Direct measurements of velocity of Antarctic outlet glaciers are generally scarce, particularly inland of the coastal zone in which control points and recognizable features are most common. However, the virtually

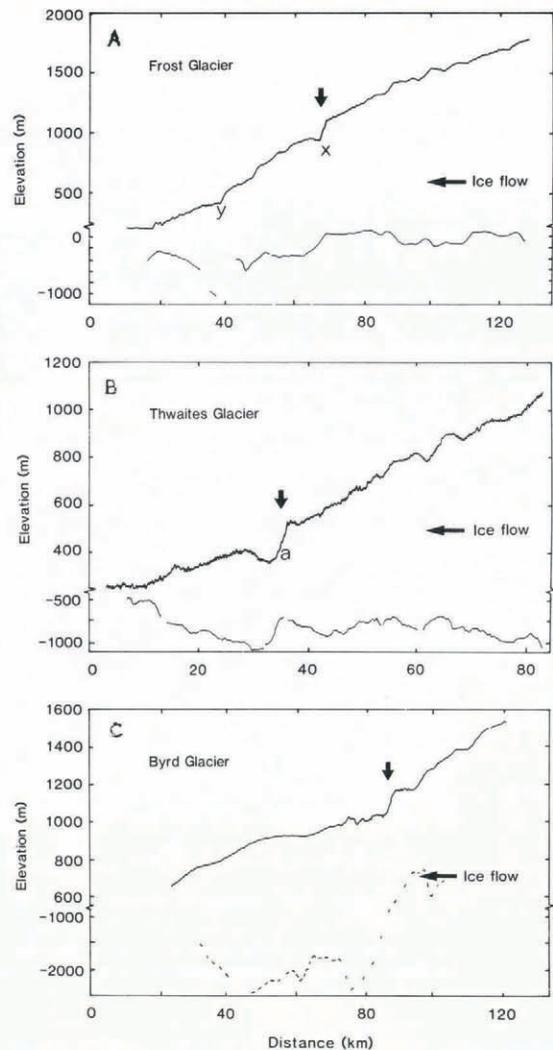


Fig. 2. Derived surface profiles from aircraft altimetry and bedrock profiles from radio echo-sounding down flow-lines to Frost (A), Thwaites (B), and Byrd (C) Glaciers showing the repetition of steep slopes in association with basal peaks which occur at the transition from sheet to stream flow. Bedrock profile for Byrd Glacier is dashed because of sparse data. Note that vertical scales for surface and bedrock profiles are different. Points a, x, y, and z refer to locations on Landsat images in Figure 1.

unlimited number of natural features on the surface of Byrd Glacier enabled Brecher (1982) to make detailed determinations of velocity variations within the fjord walls. He calculated surface velocities at 601 points photogrammetrically and contours of these in the upper fjord region are shown in Figure 3 superimposed on a Landsat image. With an expected error of less than 5%, they show good agreement with previous, independently determined velocities at lower parts of the glacier (Swithinbank, 1963; Hughes and Fastook, 1981). The absence of sufficient ground control in the region of converging flow made the calculation of velocities difficult. However, three velocity measurements were made by ground survey (written personal communication from H.H. Brecher) and, although of poorer relative accuracy because of their low values, provide velocities inland of the observed step in surface topography. Balance velocities have also been calculated for the Byrd Glacier drainage basin (Fig. 3) using data from Drewry (1983[a]). The good agreement between the measured and calculated values inland of the fjord walls suggests that the latter are accurate to a first approximation.

Figure 3 shows a significant change in motion

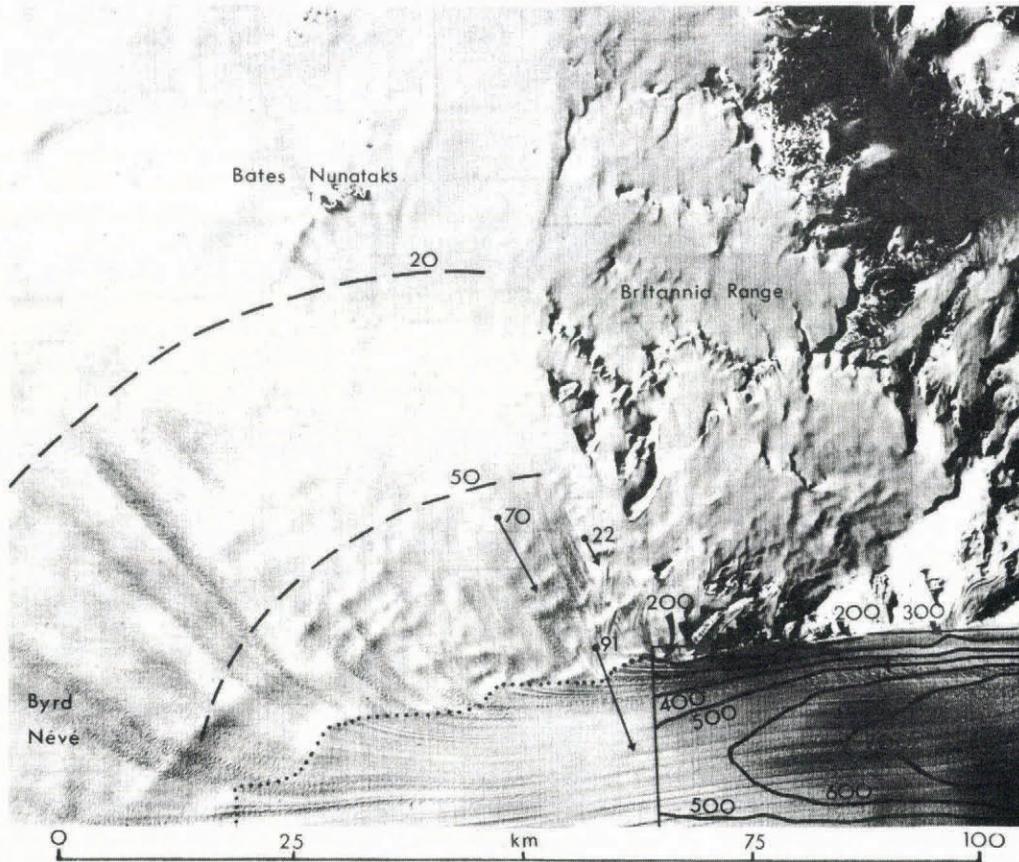


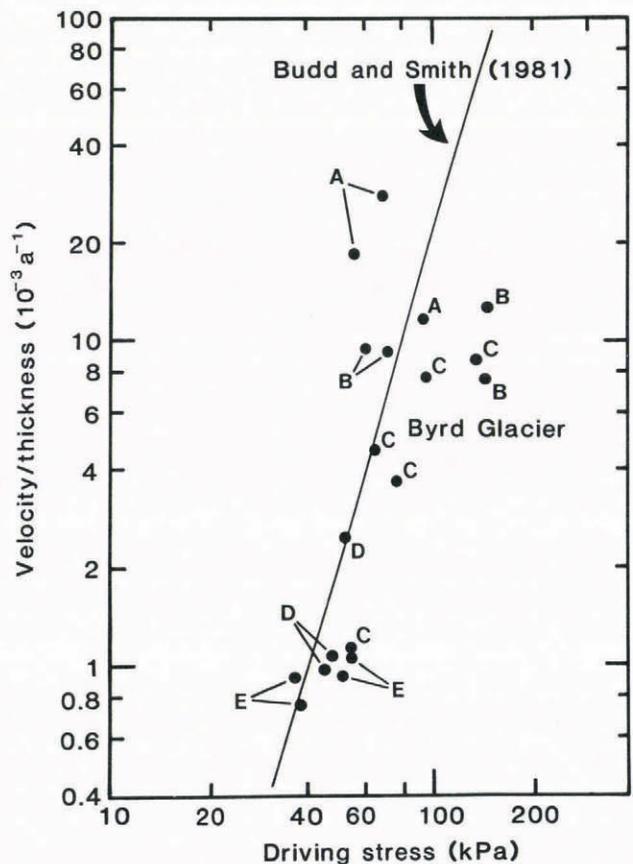
Fig. 3. Balance and measured velocities in the region of Byrd Glacier. Balance velocities (dashed lines) have been calculated from data in Drewry (1983[a]). Measured velocities (heavy solid contours) have been calculated photogrammetrically (Brecher, 1983); the three velocity vectors are from ground survey (written personal communication from H.H. Brecher).

associated with the onset of topography characteristic of streaming flow. Balance velocities increase gradually across the drainage basin and are approximately  $50 \text{ m a}^{-1}$  within 10 to 20 km of the topographic step. The measured velocities show that the ice is still moving at less than  $100 \text{ m a}^{-1}$  within 2 km of this point. Only 8 km down-stream, however, the ice has accelerated to  $500 \text{ m a}^{-1}$  and reaches a maximum of  $875 \text{ m a}^{-1}$  within another 55 km.

That motion inland of the step is largely due to internal deformation can be seen by comparing values of driving stress with the ratio of velocity to thickness (Fig. 4), which is proportional to the mean shear strain-rate through the ice column in the absence of sliding. Ice flow dominated by basal lubrication undergoes a transition to low driving stresses while sliding increases greatly (Young, 1981; McIntyre, unpublished). This is not seen for the data in Figure 4, which are taken from within 10 km of the topographic step in Byrd Glacier. The pattern is similar to that of East Antarctica which is thought to move by internal deformation (Budd and Smith, 1981). Although the scatter of points is quite high, this may be attributed to the small window (20 km) used to calculate driving stresses which is unlikely to have averaged out longitudinal stress gradients.

Direct evidence of subglacial water, and hence the

Fig. 4. Variations in mean shear strain-rate (velocity/thickness) with driving stress up-stream of the step in the Byrd Glacier profile. Regression line is for polar ice thought to be moving largely by internal deformation (Budd and Smith, 1981). Letters refer to distances up-stream of the step in the surface profile: A 10-19 km; B 20-39 km; C 40-59 km; D 60-79 km; E 80-100 km.



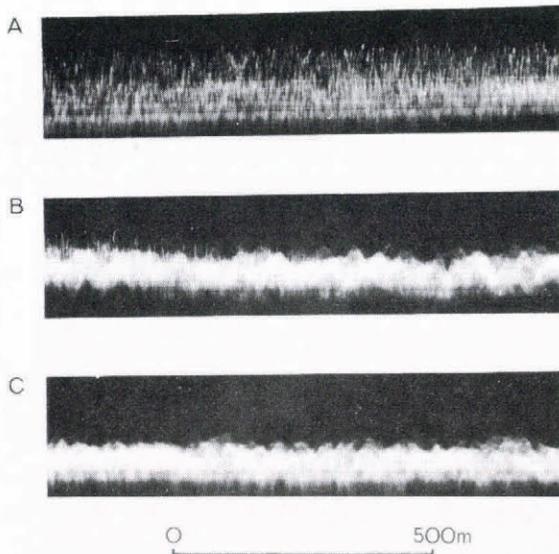


Fig. 5. ESM records of bedrock echoes down a flow-line to Byrd Glacier. A, B and C are 4 km up-stream, at, and 3.5 km down-stream (C) of the abrupt increase in surface slope.

possibility of basal sliding, down-stream of the maximum surface gradient comes from consideration of the strength of returned radio echoes. Figure 5 shows bedrock echoes recorded in the echo strength measurement (ESM) mode (Neal, 1976) down a flow-line to Byrd Glacier at points 4.0 km up-stream (A), at the mid-point (B), and 3.5 km down-stream (C) of the abrupt increase in surface slope. Unfortunately, because this mode of recording was used in only three Antarctic radio echo-sounding seasons in the joint programme involving the National Science Foundation, the Scott Polar Research Institute, and the Technical University of Denmark and because of the changing of film and faults in the recording camera, this is the only outlet glacier for which a complete profile of ESM data is available. Qualitative support for the patterns discussed below may, however, be found in the changing character of bedrock echoes in Z-scope profiles where they are not obscured by heavy crevassing.

Figure 5A shows the great variability in returned peak power which is characteristic of a rough ice-rock interface. The echo consists of many smaller reflections from individual facets; their amplitudes and auto-correlation lengths are related to the distribution of slopes and the auto-correlation length of the reflecting surface (Neal, 1976). Figure 5C, however, shows a very much more uniform level of returned power and is more typical of an interface between ice and ponded water. An intermediate reflecting surface is shown in Figure 5B, in which it is possible to see the transition from rock to water reflectors over a distance of about 500 m. Contrary to expectations, ESM amplitudes in these examples are generally less where the ice is underlain by water (B and C) than by rock (A). This is the result of the rapid increase in ice thickness associated with the step in subglacial topography and hence the greater absorption of the returned power.

Details of the transition from slow- to fast-flowing ice are further confirmed by patterns of driving stress, the mean stresses in balance with variable basal shear stresses and longitudinal stress gradients. It has been noted that down flow-lines to outlet glaciers there is typically a gradual increase to a peak of about 200 kPa within 100 km of the grounding line followed by a rapid decline to zero at this point (Budd and Young, 1979; Young, 1981; Cooper and others, 1982; McIntyre and Cooper, 1983). This pattern is also seen down flow-lines to Thwaites and Lennox King Glaciers. Using mean values over 10 km sections, the maximum driving stresses of 220 and 280 kPa down flow-lines to Thwaites

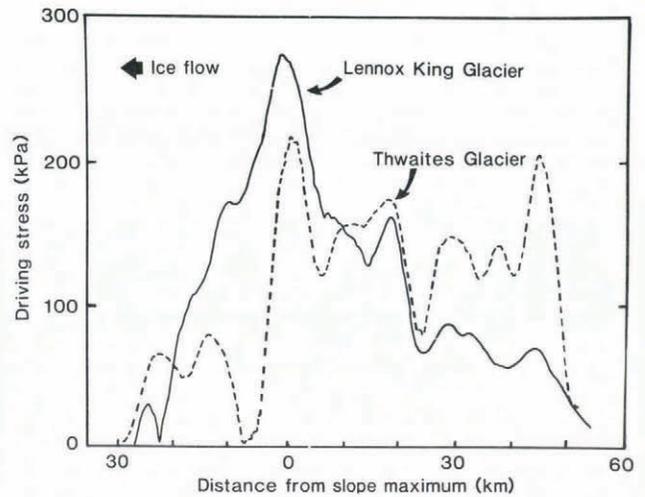


Fig. 6. Driving stress profiles down flow-lines to Thwaites and Lennox King Glaciers showing the change to low values after the peak value which occurs at the step in the surface profile. Data have been averaged over short distances (10 km) and so the pattern is distorted by longitudinal stress gradients.

and Lennox King Glaciers respectively occur, as expected, at the mid-point of the maximum surface gradient (Fig. 6). The observed rapid decline and subsequent low values of driving stresses may be attributed to a changed flow regime induced by lubrication by basal water, as was indicated by ESM records.

The three-dimensional nature of this transition in flow was shown by the arcs of high driving stresses round the heads of ice streams and outlet glaciers (Cooper and others, 1982; McIntyre and Cooper, 1983). They can now be seen to be associated with the amphitheatre-like basins which mark the change in regime of flow. The subsequent fall to a driving stress of zero at the grounding line, as was found in the two-dimensional analysis, was not reproduced in three dimensions in all cases because of the grid size which averaged data over 111 km squares.

Mechanisms for the presence of subglacial water in basins at the head of outlet glaciers, and hence for the onset of basal sliding, are not known with any certainty but two processes have previously been suggested. First, the zone of high convergence of flow, steep slopes, and high driving stresses may provide lubrication through the dissipation of frictional energy, strain softening, and increased water production (Budd and others, 1979). Second, Hughes (1977) suggested that the extensive zones of crevassing in the transition zone may provide suitable conduits for surface melt water to reach the bed if they remain water-filled, as discussed by Weertman (1973). Differentiation between these mechanisms is not possible given present knowledge of ice-sheet dynamics but we may consider the closely related role of basal temperatures in the change in regime.

Weertman (1963), in an analysis of ice flow through a mountain barrier, predicted the profile of the ice sheet and the heat balance at bedrock in the region of outflow. Although he did not allow for the observed concavity in surface form where basal sliding sets in, he recognized that basal ice in immature and very rapidly moving outlet glaciers, and possibly that of mature outlets too, would be at pressure-melting point. The coincidence of the zone of maximum stress and the apparent onset of sliding (Budd and Young, 1979) has been taken to indicate that the acceleration of the ice is induced by frictional heating (Budd and others, 1979), which will also reach a peak at this point. Using the model of Budd (1969), we can estimate basal temperatures inland of this zone in Byrd Glacier which occupies a substantial breach

in the Transantarctic Mountains. The model's assumptions, for instance the requirement of a uniform horizontal temperature gradient up-stream, were found to be valid since the parameters to which the calculation is most sensitive (that is, surface velocity and temperature and basal temperature gradient) vary only slow up-slope. Even assuming a basal temperature gradient of  $0.03 \text{ K m}^{-1}$  and a balance velocity of  $35 \text{ m a}^{-1}$ , the basal ice is at pressure-melting point 20 km inland of the transition to fast flow in this mature outlet. More realistic values are 0.03 to  $0.07 \text{ K m}^{-1}$  (Smith and Drewry, 1984) and  $50 \text{ m a}^{-1}$  (Fig. 3). We may therefore assume that the zone of warm basal ice due to frictional heating extends a considerable distance into the ice sheet from the subglacial step. On this basis, the transition to high-velocity flow is unlikely to be due to temperature and the importance of a topographic trigger may be assumed.

The preceding evidence linking high-velocity flow in outlet glaciers to a characteristic subglacial topographic feature has been largely confined to a few instances for which the necessary sources of data have been available. That the resulting conclusions are more widely applicable, and hence that the association is factual rather than coincidental, can be shown by supporting the discussion with data from other sources. Table I presents statistics relating to the transition in flow for nine glaciers. Sources used include topographic maps at a scale of 1 : 250 000 published by the U.S. Geological Survey and radio echo-sounding profiles; they are of poorer accuracy than the Landsat imagery and aircraft altimetry but confirm the previously noted details where overlap occurs. The results are consistent and the repetition of certain features should be noted. These are the marked step in the surface profile at the point of transition, the

occurrence of a bedrock sill at this point, the lower mean slopes down-stream of the step compared with those above, the sharp fall in driving stresses calculated over 10 km sections, and the onset of crevassing. All these points support the concept of a topographically initiated transition in flow regime shown by velocity measurements in Byrd Glacier (Fig. 3). Their repetition indicates that it is a consistent feature which controls discharge from the ice sheet.

*Types of ice-sheet outlet*

The flow of West Antarctic ice streams has often been regarded as being different from the steady-state regime of outlet glaciers (Hughes, 1973, 1977; Young, 1981). This is partly based on their gradual transition to low driving stresses with increasing shear strain-rates between 400 and 600 km up-stream of the grounding line rather than the abrupt decrease in the last 100 km of outlet glaciers (McIntyre and Cooper, 1983). Such behaviour does not necessarily indicate instability; the measured velocities of the ice streams have been closely approximated by steady-state balance velocities (Weertman and Birchfield, 1982).

The development of high velocities in outlet glaciers may be considered through the model of Budd (1975) which recognizes three classes of glacier (ordinary, fast, and surging) and two modes of flow. An empirical distinction of these flow regimes has been made on the basis of centre-line thickness as a function of surface slope and velocity. Data from Drewry (1983[a]) show that Byrd, Ninnis, Thwaites, and Frost Glaciers are all classed as fast polar glaciers by this relationship. They support the concept of a lubrication factor enabling the development of high velocities after a critical limit of velocity and basal stress has been reached.

TABLE I. STATISTICS OF GRADIENTS, DRIVING STRESSES, AND SILL HEIGHTS IN THE REGION OF THE TRANSITION TO HIGH-VELOCITY FLOW WITHIN OUTLET GLACIERS

Glacier	Gradient			Driving stress		Absolute height of subglacial sill	Comments
	above peak	at peak	below peak	at peak	below peak		
	%	%	%	kPa	kPa	m	
Frost	1.4	7.9	0.9	160	20	400	
Thwaites	2.0	7.7	1.1	120	30	400	
Byrd	1.3	7.2	0.4	170	40	2200	
Lennox King	1.5	6.2	0.8	160	20	1250	
Mackay	1.7	6.5 8.2	1.6	-	-	-	Two steps in profile close together.
David*	2.0	8.0	0.9	-	-	-	Escarpment round only 3/4 of basin.
Scott	0.9	-	0.9	-	-	-	No marked step visible in surface profile.
Scott*	-	7.1	1.6	-	-	-	Escarpment round only 3/4 of basin.
Mertz†	-	3.1	1.1	190	40	450	
Ninnis†	1.2	-	0.4	160	30	600 600	Two steps in bedrock profile, 20 and 50 km up-stream from driving-stress peak. No clear step in surface profile.

\* U.S. Geological Survey maps, 1 : 250 000.

† S.P.R.I. radio echo-sounding profile.

The relationship does not hold, however, for the West Antarctic ice streams which drain into the Ross Ice Shelf. Using mean values of thickness and balance velocity for Ice Streams B, C, D, and E (Rose, 1979), we find significant inconsistencies with the expected value of surface slope. Observed gradients (typically less than 0.1%) are less than predicted by an order of magnitude or more. An error of this size cannot be explained by a difference between calculated and real velocities or by inaccuracies in thickness or slope measurements. That the model does not fit the data suggests the inapplicability of using a lubrication factor to explain their presumed high velocities. In support of this, Young (1981) suggested that there is a qualitative difference between these ice-sheet outlets and fast polar glaciers. Bindschadler (1983) concluded that the dynamics of Ice Stream B were better explained by pressurized subglacial water. High water pressures result in basal separation and stress concentration where the ice is still in contact with the bed. It is these concentrations that control the sliding velocity.

A better understanding of these ice-sheet outlets is gained by comparing them with confined ice shelves which, in view of the above discussion and the suggested importance of pressurized water, may be expected to behave in a similar manner. In both cases, it is likely that the walls rather than the bases of the ice streams provide most of the restriction to flow. Indeed on the basis of localized flat surfaces with ice thicknesses indicating hydrostatic equilibrium, Robin and others (1970) suggested the presence of a series of "pseudo" ice shelves in this region of Marie Byrd Land. Using data on ice thickness, surface velocity, and surface slope for ice-shelf segments (approximately 50 km in width) between ice rises, grounding points, and land-based ice (Swithinbank, 1955; Robin, 1958, 1975; Rose, 1979; Thomas and others, in press), we find a similar disagreement with the model of Budd (1975). Surface slopes of ice shelves from Quarisen and locations between Roosevelt Island and Shirase Coast and between Cray Ice Rise and Shackleton Glacier in the Transantarctic Mountains are lower than fast-moving outlet glaciers by one to two orders of magnitude. It is therefore reasonable to consider the Marie Byrd Land ice streams as more closely related to ice shelves than to terrestrial ice. They appear to represent landward extensions of the ice shelf which penetrate the West Antarctic ice sheet. Their behaviour is consistent with ice which is flowing under the influence of lateral spreading rather than the basal shearing which dominates terrestrial ice flow.

## DISCUSSION

### *Stability of ice-sheet outlets*

Of central importance to the stability of any part of the Antarctic ice sheet is the transition from internal deformation to fast flow with basal sliding. Although little is known of the mechanics of this transition, it was shown above that the valley head which typically underlies outlet glaciers initiates fast flow through the acceleration of converging drainage down a steep gradient. It thus controls the higher velocities observed down-stream. If this is the case, then the rapid transfer of the transition zone inland, as occurs during the surge of a temperate glacier, seems particularly improbable. With the topographic step controlling the onset of basal sliding, it is likely to induce a form of stability in discharge from the ice sheet since the transition zone will be unable to retreat beyond it. The acceleration of ice through outlet glaciers is therefore pinned to a zone at the head of subglacial valleys by its dependence on topographic initiation. Even with minor fluctuations of the location of the onset of basal sliding due to variations in boundary conditions (such as ice-sheet flux, sea-level, or climatic regime), the head of the subglacial valley will inhibit propagation of a surge by acting as a sill beyond which grounding-line retreat cannot occur (Hughes, 1977; Thomas, 1979). Thus, ice sheets draining through subglacial fjord topography or mountain terrain

appear unlikely to undergo instabilities in flow regime. This is an important suggestion with regard to the Greenland ice sheet which drains almost entirely through outlet glaciers. It is interesting to note that no evidence of surge activity has yet been presented for this ice mass.

### *Ice-sheet erosion*

This scheme of topographically initiated stream flow is inherent to concepts of the selectivity of glacial erosion within ice sheets. Clayton (1965) and Sugden (1968) concluded that erosion is concentrated in valleys while plateau areas may remain largely unaffected by the passage of ice. Our evidence supports this scheme which suggests that ice-sheet motion is dominated by internal deformation while the steep gradients of subglacial valley heads initiate the basal sliding and fast velocities seen in outlet glaciers.

Several characteristic topographic components of glacial troughs, among them the steep valley head and associated overdeepened profile, have long presented difficulties to glaciologists and geomorphologists attempting to explain their origin. Sugden and John (1976) explained overdeepening by the relative position of the equilibrium line along the trough, which usually determines the location of the zone of greatest ice discharge. Here, velocity and ice thickness will tend to be greater than elsewhere, thus favouring higher than normal rates of erosion. Cray (1966) observed that the thickness of floating ice in a fjord becomes greater inland. An increase in ice thickness or uplift of the land would bring the bottom material gradually into contact with the moving underside of the ice, allowing direct erosive action to take place. Bathymetry of the valley floor should thus reflect the ice-water boundary at the maximum extent of the ice.

The preceding discussion raises the possibility that the development of a steep headwall and subsequent overdeepening may be associated with the onset of basal sliding. Following the commencement of glacial erosion at the valley rim, the pre-existing topography has a concave form while thickness and velocity increase rapidly down-stream, all of which may result in accelerated erosion (Holstedahl, 1967). Erosion will then increase to a maximum which is determined by the progressive decoupling of the ice from its bed by water. This hydrostatic buoyancy is due to the presence of water from frictional melting and from the sea. The presence of subglacial water may also increase the potential abrasion rate which is dependent on ice velocity and effective normal pressure (Boulton, 1974). Subglacial hydrostatic pressure will reduce the overburden pressure of the ice within the fjord, thereby changing the basal regime from one of till lodgement to one of possible abrasive down-wearing. At the grounding line, the ice becomes freely floating and so will have lost its erosive ability until the mechanism suggested by Cray (1966) becomes operative. Thus, the fjord will develop, first, a steep headwall due to the rapid onset of basal sliding, second, an overdeepened inland end as enhanced erosion increases with ice velocity, thickness, and topographic concavity, and, third, progressively higher bedrock elevations as hydrostatic decoupling reduces the ice's erosive ability. With the maximum erosive power concentrated at the valley head near the headwall, this feature is likely to retreat up-stream as suggested by Souchez (1967).

Assuming that these processes of headwall retreat and inland overdeepening occur, we may return to Figure 2 and compare the sub-glacial topography of Byrd and Thwaites Glaciers. The former exhibits a headwall of some 2200 m which is approximately 100 km inland from the seaward flank of the Transantarctic Mountains. If Byrd Glacier has existed for 25 million years either as an outlet for the East Antarctic ice sheet or as a glacier fed by more local mountain ice masses (Smith and Drewry, 1984), the headwall retreat through the mountains represents a mean rate of headwall erosion of  $4 \text{ m a}^{-1}$ . This may be an underestimate because of the uplift of the mountains. However, given that it occupies a major zone of tectonic weakness, the rate of erosion is not

unrealistic. Thwaites Glacier, on the other hand, shows a vertical step of only 400 m which is displaced 10 to 15 km inland of its grounding line. Although it now has velocities of about  $3 \text{ km a}^{-1}$  (Thomas and others, 1979) compared to a maximum of  $875 \text{ m a}^{-1}$  for Byrd Glacier (Brecher, 1982) and hence presumably a high erosive power, this has been taken to indicate that it is a youthful, or perhaps transient, feature (Drewry and Robin, 1983).

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