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CHARACTERISTICS OF ICE FLOW IN MARIE BYRD LAND, ANTARCTICA

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Abstract. Extensive radio echo-sounding has mapped the part of West Antarctica between Byrd Station, the Whitmore Mountains, the Transantarctic Mountains, and the Ross Ice Shelf. The ice sheet in this area is dominated by five major sub-parallel ice streams (A–E), which are up to 100 km wide and extend inland from the grounding line of the Ross Ice Shelf for about 400 km. Their positions have been determined by crevassing seen on radio echo-sounding records, trimetrogon photographs, and Landsat imagery. The ice streams are characterized by their flat transverse cross-sections, while the intervening ice sheet exhibits domes and ridges. Ice flow lines are defined from the ice-surface contour pattern and the trend of the ice streams. It is apparent from this work that the flow line passing through Byrd Station joins ice stream D.

The bedrock of the area is relatively smooth near the Ross Ice Shelf, becoming rougher near Byrd Station and especially so near the Whitmore Mountains. Bedrock troughs, which control the positions of the ice streams, are believed to have a tectonic origin.

In this paper the role of the ice streams in the glaciological regime of West Antarctica is investigated from radio-echo data and estimates of balance velocity, basal shear stress, and basal temperatures.

Résumé. *Caractéristiques de l'écoulement glaciaire dans le Marie Byrd Land, Antarctique.* Des sondages par écho radio ont cartographié une partie de l'Ouest Antarctique entre la Station Byrd, les Whitmore Mountains, les Transantarctic Mountains et le Ross Ice Shelf. La calotte glaciaire dans ce secteur est dominée par cinq principaux courants de glace subparallèles (de A à E), qui ont jusqu'à 100 km de large et s'étendent vers l'intérieur des terres depuis la ligne de décollement du Ross Ice Shelf sur environ 400 km. Leurs positions ont été déterminées par les systèmes de crevasses vus sur les enregistrements d'écho radio, par des photographies trimétriques et par les images Landsat. Les courants de glace sont caractérisés par leur section transversale plate tandis que le reste de la calotte dans l'interfluve montre des domes et des rides. Les lignes d'écoulement de la glace sont définies à partir de l'allure des lignes de niveau et l'orientation du courant de glace. Il apparaît que la ligne de courant passant par la station Byrd rejoint le courant de glace D.

Le fond rocheux dans ce secteur est à relief relativement doux près du Ross Ice Shelf devenant plus accidenté près de la station Byrd et tout spécialement près des Whitmore Mountains. Les dépressions du lit rocheux qui contrôlent la position des courants de glace ont pense-t-on, une origine tectonique.

Dans cet article le rôle des courants de glace dans le régime glaciologique de l'Ouest Antarctique est étudié à partir de sondages par écho-radio et d'estimations de bilan de vitesse, de cisaillement à la base et de température à la base.

Zusammenfassung. *Charakteristiken des Eisflusses in Marie Byrd Land, Antarktika.* Die Region der Westantarktis zwischen der Byrd-Station, den Whitmore Mountains, der Transantarctic Mountains und dem Ross Ice Shelf wurde durch umfangreiche Radar-Echomessungen aufgenommen. Der Eisschild wird in diesem Gebiet durch 5 grössere, annähernd parallele Eisströme (A–E) beherrscht, die bis zu 100 km breit sind und sich landeinwärts von der Aufsetzlinie des Ross Ice Shelf etwa 400 km erstrecken. Ihre Lage wurde aus Spaltenzonen, die aus den Radar-Aufzeichnungen zu erkennen sind, aus Trimetrogon-Aufnahmen und aus Landsat-Bildern bestimmt. Die Eisströme sind durch ihre flachen Querschnitte gekennzeichnet, während dazwischen Dome und Rücken auftreten. Die Stromlinien ergeben sich aus dem Muster der Höhenlinien auf der Eisoberfläche und der Zugrichtung der Eisströme. Die Untersuchung lässt erkennen, dass die Stromlinie durch die Byrd-Station sich mit dem Eisstrom D vereinigt.

Das Felsbett des Gebietes ist in der Nähe des Ross Ice Shelf relativ glatt, wird jedoch in der Nähe der Byrd-Station und besonders vor den Whitmore Mountains zusehends rauher. Tröge im Felsbett, welche die Lage der Eisströme bestimmen, haben vermutlich tektonischen Ursprung.

Der Beitrag untersucht die Rolle der Eisströme im glaziologischen Regime der Westantarktis, mit Hilfe von Radar-Daten und Abschätzungen der Bilanzgeschwindigkeit sowie der Scherspannung und der Temperatur am Untergrund.

Introduction

During the 1974–75 Antarctic field season the Scott Polar Research Institute (SPRI) combined with the Technical University of Denmark (TUD) to carry out airborne radio echo-sounding of the Antarctic ice sheet with the co-operation and support of the U.S. National Science Foundation (NSF). As part of this work, a 50 km grid (Fig. 1) was flown covering 0.5×10^6 km^2 of Marie Byrd Land, where ice streams had been reported on earlier reconnaissance radio echo-sounding flights (Robin and others, 1970[a], [b]). These ice streams are relevant to the recent debate concerning the stability and dynamics of the West Antarctic ice

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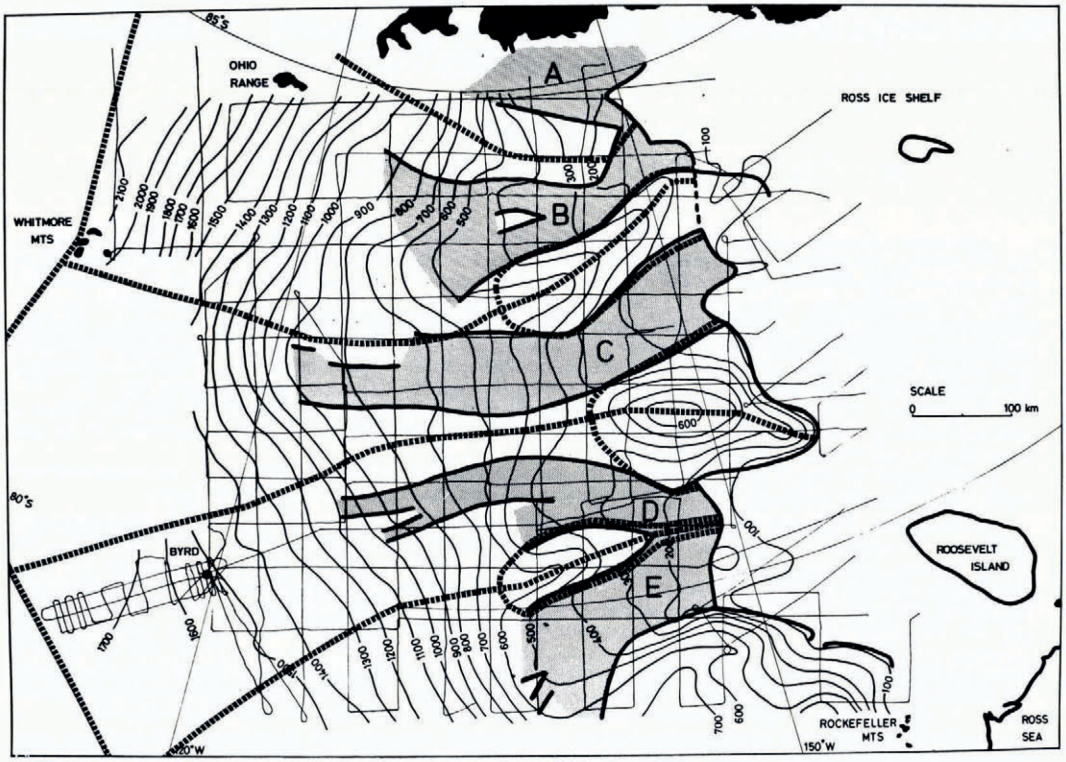

Fig. 1. Map of Marie Byrd Land showing 1974–75 radio echo-sounding flight grid, ice-surface contours (100 m interval), ice streams (stippled), and catchment-area boundaries (broken lines).

sheet–Ross Ice Shelf system (Hughes, 1973, 1977; Thomas, 1976[c]; Weertman, 1976; Whillans, 1976, 1977).

Details of data collection and reduction, together with a contour map and an analysis of the bedrock, have been presented elsewhere (Rose, in press).

Bedrock

Figure 2 shows a montage of 13 bedrock profiles with the positions of the ice streams marked. The smooth area adjacent to the Ross Ice Shelf, where ice streams are contained by broad channels (which continue westward under the Ross Ice Shelf (Robertson and others, in press) and the Ross Sea (Hayes and Davey, 1975)), is believed to be an extension of the Ross Sea sedimentary basin. The ridges are considered to be tectonically controlled (Rose, in press).

Further inland, ice-stream channels degenerate and the topography becomes progressively rougher. A major valley, often reaching 1 500 m below sea-level, connects with the trench beneath ice stream B and winds north towards Byrd Station to join the Bentley subglacial trench. Around the Whitmore and Thiel Mountains the topography is the roughest in the area; large blocks of heavily dissected mountains are divided by steep-sided valleys, typically 20 km wide and 2 km deep.

In general, the bedrock is well below sea-level. Removal of the ice sheet and isostatic uplift would produce only a few islands, principally in the Whitmore Mountains area, but also to the north and west of Byrd Station.

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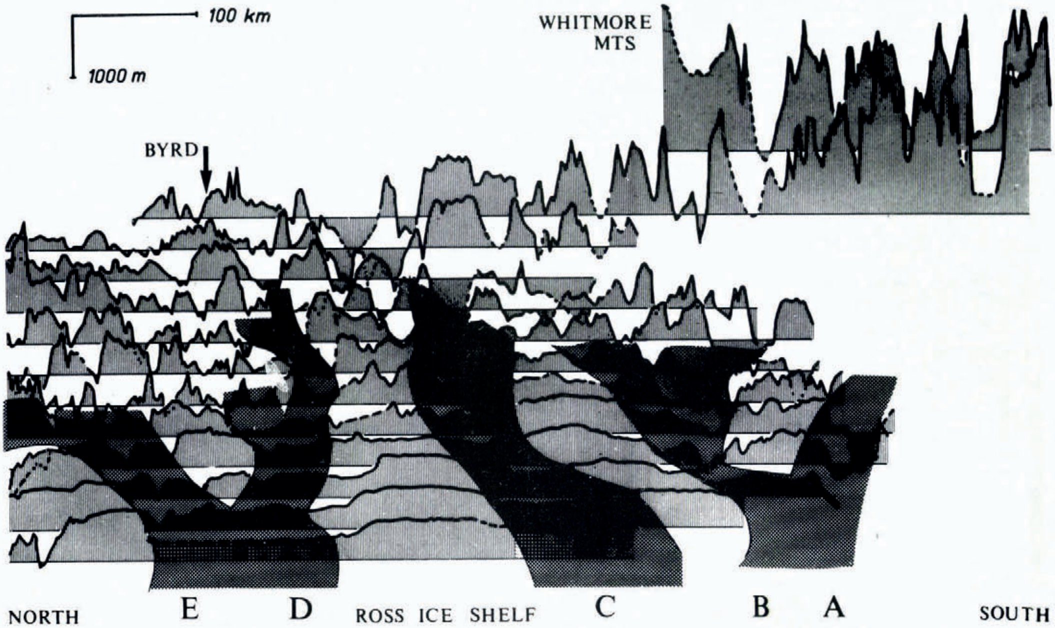

Fig. 2. Radio-echo profiles of bedrock at 50 km spacing traversing Marie Byrd Land parallel to long. 135° W. The common datum for profiles, the base of the shading, is at −1 000 m. The effective separation is 500 m. Positions of ice streams are shown by the heavier sections of profiles and the stippling. Interpolated data are represented by dashes. Dots show obscured sections.

Ice surface

Figure 1 shows ice-surface contours. The grounding line of the Ross Ice Shelf was defined by detecting the point at which the surface elevation (e) rose above that predicted for a freely floating ice sheet of the same thickness (t), using the relationship

$$e = 0.108t + 15.6, \tag{1}$$

taken from Crary and others (1962, fig. 14). These data were supplemented with trimetrogon photographs and Landsat imagery.

Ice streams

Ice streams were identified on the radio-echo films by cluttered echoes received from features, on or near the surface, that were associated with them (Fig. 3). Normally, sastrugi and other surface irregularities will reflect incident radio energy and produce a series of weak hyperbolae originating at or near the ice surface. In most cases this clutter is weak; it may not cover the whole width of the film nor will it obscure basal reflections or deep internal layering. Zones of intense surface crevassing will return strong echoes which can be detected at near-horizontal angles. The resulting distinct long-tailed bands of hyperbolae often totally obscure weaker bottom and internal reflections.

The radio-echo record (Fig. 3) is typical of all the ice streams. The results of mapping the ice streams from these zones of clutter are shown in Figure 1. At the inland extremities of the streams the radio-echo hyperbolae are still unmistakable.

Figure 3 illustrates some significant differences between ice streams B and C. The hyperbolae of ice stream B are more intense as shown by:

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(a) The bedrock reflection is obliterated more effectively and for a longer horizontal distance.

(b) Considering echoes from the boundary crevasse zones, the tails of the hyperbolae are longer, indicating stronger reflections for oblique rays in low-gain sectors of the antenna radiation pattern. These tails are continued at the top of the film record between the end of the suppression line and the ice-surface echo, and in line with this again at the bottom of the record. (This is due to the high pulse-repetition frequency of the transmitter.)

(c) The recovery time of the receiver, after saturation by the surface echo, is longer. This is shown by the depth of the black zone beneath the surface echo.

Trimetrogon photographs confirm this difference in crevasse intensity.

Ice stream E is sufficiently far north for it to appear on Landsat imagery. The severely crevassed northern boundary is seen separating the ice stream with its streamlines, rippled surface appearance, and patches of crevasses from the smooth undisturbed surface of the adjacent ice plateau. The ridge between ice streams D and E has a similar smooth appearance, while further inland the entire surface is covered by minor features, perhaps undulations, ≈ 5 km wavelength. The eastern extremities of ice stream E show as streams about 15 km wide with striated surfaces.

Glaciological studies

The radio-echo results and other pertinent data have been combined in order to evaluate some glaciological parameters relevant to the study of the ice streams. These are summarized below and discussed later with reference to particular ice streams.

(i) *Flow lines*. Flow lines have been drawn for the area assuming that these run normal to the ice-surface contours. Figure 1 shows the boundaries of the catchment areas of the various ice streams, domes, and ridges.

(ii) *Layers*. On the radio-echo records, internal layers are often seen within the ice sheet. The cause of these is uncertain but they are believed to be due to former surfaces which possess small dielectric variations (Robin and others, 1969; Paren and Robin, 1975; Whillans, 1976). These may be due to impurities or density changes between layers. Because of the low energy returned, layers cannot be seen near the top of the ice where they are swamped by the tail of the surface echo, nor near the bottom of warm ice where absorption attenuates the signal below the sensitivity of the system or where the continuity of layers has been destroyed by severe deformation of basal ice over rough terrain (Robin and others, 1977). Those areas in Marie Byrd Land where internal layers were recorded in 1974–75 have been mapped. Although detection of layers depends on receiver settings and variations in geometric losses caused by different aircraft terrain clearances, some observations can nevertheless be made. Layers are not seen in the ice streams. They are occasionally seen in the intervening domes and ridges. Near the ice divide, strong layers are seen which can be followed over long distances without gaps or major distortion (e.g. Whillans, 1976). The layers become weaker, discontinuous, and distorted as the ice flows down-stream, until they disappear altogether.

(iii) *Basal shear stress*. Basal shear stresses (Fig. 4) have been evaluated from surface slopes and ice thicknesses, averaging these values between adjacent 100 m ice-surface contours along flow lines, where these intersect flight lines. Near the grounding line on the ice streams, shear stresses decrease to less than 0.15 bar.

(iv) *Balance velocities*. Using the sparse snow-accumulation data (Bull, 1971; Whillans, 1975; Clausen and Dansgaard, 1977) combined with the flow-line map and ice-thickness data has enabled the steady-state balance flow velocities and volume flow-rates to be calculated

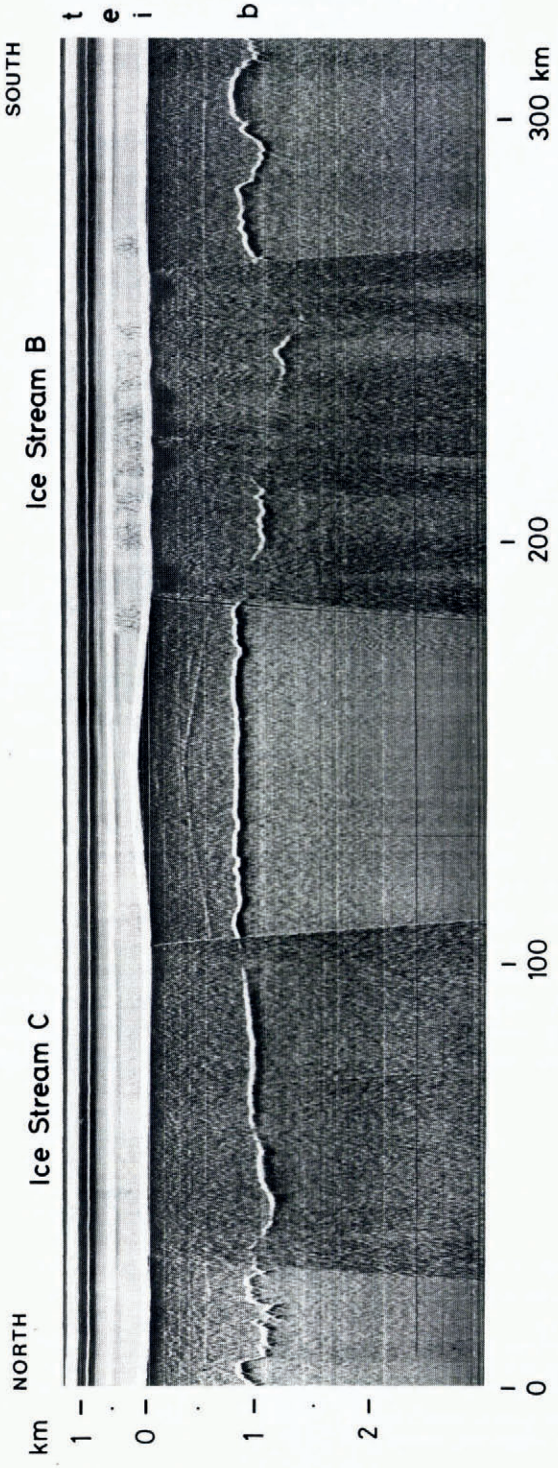

Fig. 3. Radio-echo film record of ice streams C and B (lat. 81.7° S., long. 139° W. to lat. 84.5° S., long. 141° W.). t = transmitter pulse, e = end of suppression, i = ice-surface echo, and b = bedrock echo.

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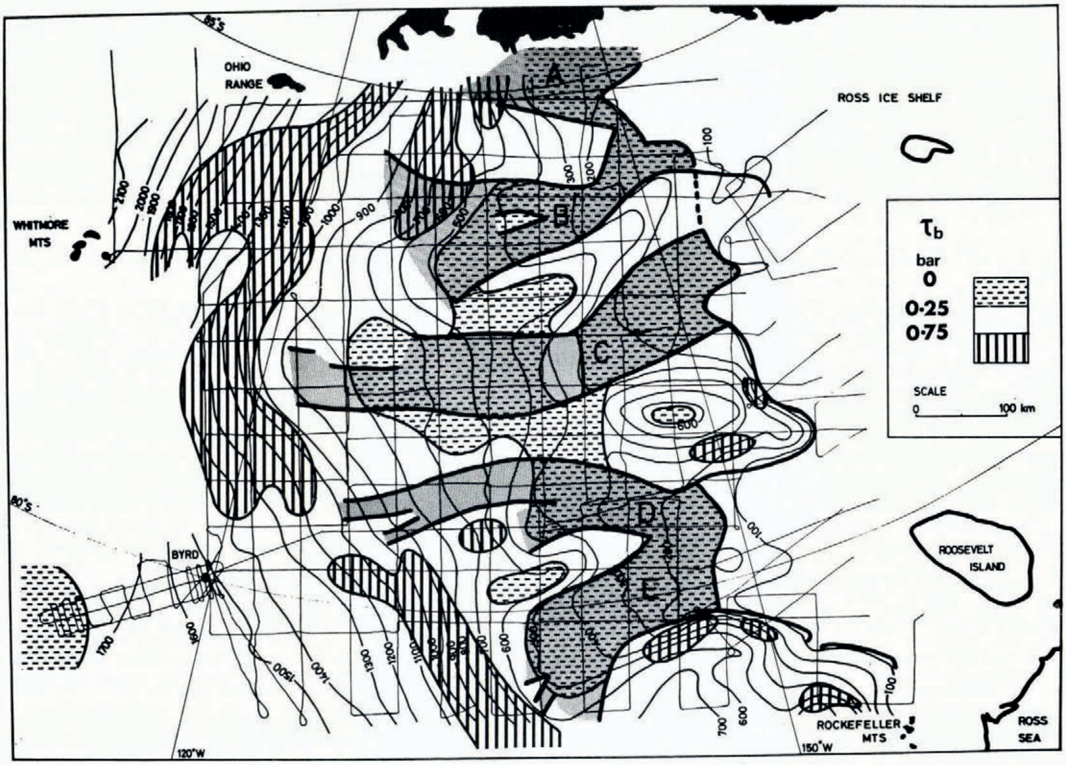

Fig. 4. Distribution of calculated basal shear stresses for grounded ice of Marie Byrd Land.

(Table I). Near the ice divide and on the various ridges and domes, velocities are expected to be less than 10 m a^{-1}. In the low shear-stress, low surface-slope areas of the ice streams, the ice accelerates to between 100 and 600 m a^{-1}.

(v) *Surface temperatures*. Bentley and others (1964) published a summary of temperature data. Since then, the Ross Ice Shelf Project (RISP) has collected more data (Thomas, 1976[a]). Detailed information from the ice streams themselves is still lacking.

Zwally and Gloersen (1977) published passive micro-wave pseudo-colour images of the Antarctic. The brightness temperature of firn is determined by the annual accumulation-rate and the mean annual temperature. Ice streams D and E, and the Siple dome are clearly

Table I. Comparison of Marie Byrd Land ice streams
(cf. Hughes, 1973)

	Ice stream			
	B	C	D	E
Catchment area (10^3 km^2)	163	122	104	131
Volume flow-rate (km^3 a^{-1})	18	13	11	24
Balance velocity				
at 500 m contour (m a^{-1})	100	100	50	200
at grounding line (m a^{-1})	450	200	300	450
at maximum (m a^{-1})	600	200	300	450
Lowest contour below which layers are not seen within ice stream (m)	600	900	1 000	600

seen. Zwally and Gloersen suggested that topographic channelling of inversion winds may affect firn structure, drifting, and surface temperatures, and they reported evidence of summer surface melting along the ice streams. Because of these problems, it has not been possible to use micro-wave images effectively in this study.

(vi) *Basal temperatures*. Estimates of the basal temperatures, or melt-rates where applicable, have been made. Because the necessary input data over the main part of this area are poorly known, sophisticated flow-line models were rejected in favour of a steady-state column determination of temperature.

Equation (4 : 58) of Budd and others (1971) was used in the temperature calculations:

$$\theta_b = \theta_s - Z\left[\gamma_b \frac{\operatorname{erf}(y)}{y} - \frac{V\alpha\lambda}{A} 2E(y)\right] \tag{2}$$

where

θ_b, θ_s = temperatures at base and surface,

Z = ice thickness,

γ_b = basal temperature gradient (geothermal+friction),

α = surface slope,

V = velocity,

λ = surface temperature–elevation gradient,

A = accumulation-rate,

$$y = \left(\frac{AZ}{2\kappa}\right)^{\frac{1}{2}},$$

κ = thermal diffusivity,

$$\operatorname{erf}(x) = \int_0^x \exp(-y^2)\,dy,$$

$$E(x) = \int_0^x \left[\exp(-y^2)\int_0^y \exp(t^2)\,dt\right] dy.$$

This is based on a model incorporating basal frictional heating and horizontal advection of cold ice from up-stream. The Byrd Station core hole to bedrock yielded an estimated basal gradient in the ice of 0.032 5 deg m^{-1} (Gow and others, 1968). With allowance for frictional heating due to a velocity of 12 m a^{-1}, 0.026 deg m^{-1} was taken as the geothermal component of this flux in the calculations, equivalent to 0.060 W m^{-2}.

Because of the improbability of this one spot value being applicable to the whole area, whose three principal geological regions are situated between the Cenozoic volcanic province of north Marie Byrd Land and the ancient blocks of the Horlick Mountains, the geothermal heat flux was varied by factors of 0.75 and 1.25 in repeat calculations. Velocities were also varied by these same factors. In this way, the sensitivity of the results to probable errors in accumulation-rates and size of catchment areas may be evaluated.

In general (Fig. 5), ice streams were found to exhibit bottom melting, while the intervening domes and ridges were below pressure-melting point. This reflects the important effect of frictional heating in the ice streams, where high velocities more than compensate for low basal shear stresses.

Changing the velocities in the calculations influences the melt-rates under the ice streams. The edge of the freezing zone is hardly moved, because, with low velocities, frictional heating is less important than geothermal heat.

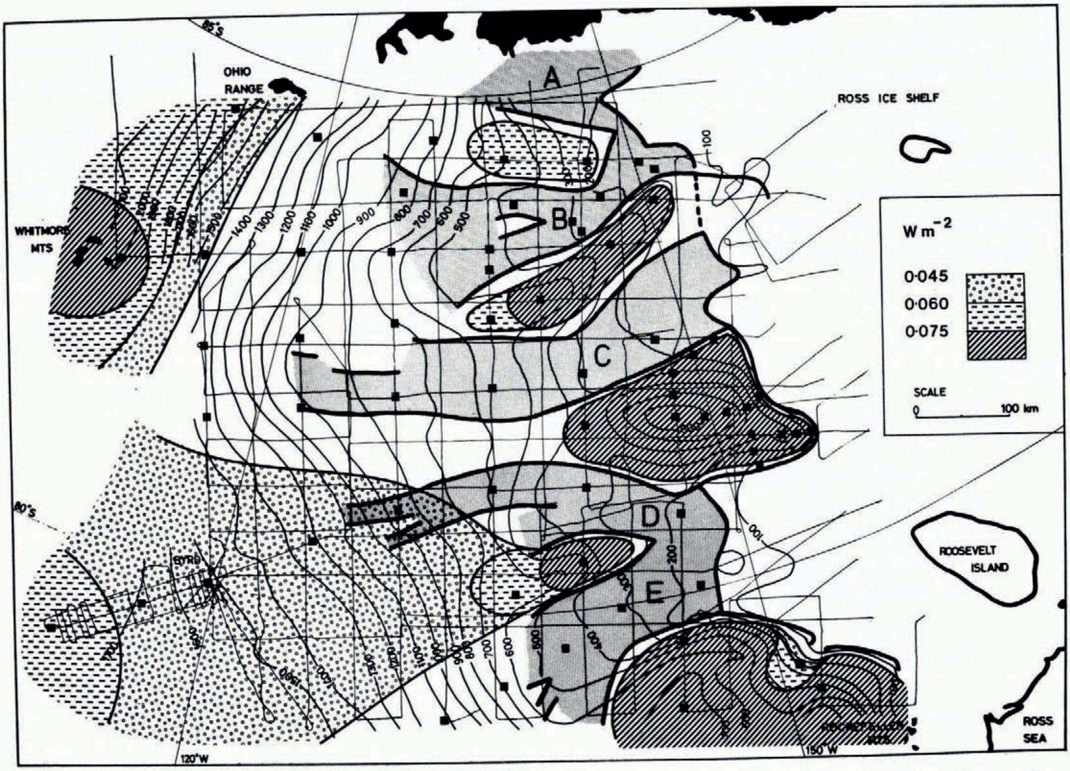

Fig. 5. Expected areas of basal freezing, for balance velocities and three different values of geothermal heat flux. Squares show points used for temperature calculations along flow lines.

Conversely, changing the heat flux (Fig. 5) influences the areas of basal freezing where ice is thick and temperatures are already near the pressure-melting point. The thinner ice of the various ridges and domes is colder, so that to cause basal melting the heat flux must be changed by a significantly greater amount.

Ice stream A and ridge A/B

Ice stream A, situated at the foot of the Transantarctic Mountains, drains ice from the high East Antarctic plateau via Reedy Glacier and the Shimizu Ice Stream. Flights over this area found very deep channels aligned with these two outlet glaciers. The remaining bedrock in this area is comparatively rough and forms the foothills to the Wisconsin Range. Balance velocities and temperatures have not been calculated for ice stream A owing to the difficulty in estimating catchment areas and volume flow-rates.

The extension of this area is the amorphous ridge separating ice streams A and B. The rock beneath (Fig. 1) rises steeply from the adjacent deep ice-stream channels and is much rougher than that beneath the other ridges. Temperature estimates (Fig. 5) show this ridge to be the least securely frozen to its bed. A basal flux of 0.065 W m^{-2} would cause melting over the whole feature (Fig. 5).

Ice stream B

This ice stream is known to be very active; measured velocities of over 500 m a^{-1} nearby on the Ross Ice Shelf (Thomas, 1976[b]) are compatible with balance velocities (Table I).

The ice stream is defined by prominent boundary crevasse zones and its surface exhibits numerous patches and bands of minor crevasses. The boundaries frequently cause the radio-echo bedrock returns to be obliterated, as shown in Figure 3 and by the interpolated sections of profiles in Figure 1. The northern boundary extends into the Ross Ice Shelf for 100 km as a 10 km wide band of well-developed parallel S-shaped crevasses, and inland for 250 km as a band of chaotically disturbed ice following the edge of the subglacial trough. This deepens towards its south side and to the east (Fig. 2). The north-east part of the ice stream has surface elevations 200 m lower than the south side but bedrock is 500 m higher.

100 km inland of the grounding line, frictional heating in the ice will melt about 12 mm of ice per year (for balance velocity and 0.060 W m^{-2}). From the present calculations, the edge of the area of basal freezing (Fig. 5) is estimated to be well inland. The ice leaving the high cold Whitmore Mountains area is warmed quite quickly as it thickens and moves into the deep trench. However, the steep slopes above the 1 000 m contour suggest that basal freezing may be present. The present calculations indicate that a 50% reduction in the geothermal heat flux to 0.030 W m^{-2} would be required before this took place.

Ridge B/C

This prominent ridge of ice has a small dome, at lat. 83° S., long. 139° W., elevation 595 m. The bedrock beneath is smooth. Because of transverse flow from the ridge into the ice streams, the small catchment area, and the low accumulation (0.1 m a^{-1}?), velocities will be very low, so geothermal heat is the dominant factor in temperature calculations. A high basal heat flux of 0.085 W m^{-2} would be required to initiate bottom melting.

Ice stream C

This stream is an enigma. Radio-echo records show areas of ice extending far inland which return clutter typical of other ice streams, yet the surface is almost featureless. Ross Ice Shelf velocities determined off the grounding line (Thomas, 1976[b]) suggest that this ice stream is stagnant, while calculated balance velocities reach 200 m a^{-1} (Table I). Robin (1975) estimated a balance velocity of 145 m a^{-1} near the grounding line by use of his ice-shelf flow lines and extrapolation of known ice-shelf velocities.

The channel beneath this ice stream is broad and smooth and its boundaries are poorly defined—exhibiting only shallow slopes (Fig. 2). Rock elevations are about −600 m, i.e. 100 m higher than the other ice streams. The head of the ice stream is located over a deep arm of the major north–south channel, where radio-echo records show discrete patches of very strong hyperbolae. Further down the stream, clutter is strong at the edges, and weaker but generally continuous across the ice stream (Fig. 3). Trimetrogon photographs show one feature running along a central flow line, and the distant north edge of the ice stream near the Siple dome. It is suspected that this latter feature is only seen because of the change in surface slope.

400 km inland from the grounding line along ice stream C is an area of low surface slopes situated over the west bank of the very deep channel. Calculated basal shear stresses are thus low here. Up-stream, contours are closely spaced and for a width of 250 km they curve around the flatter area. This suggests that the catchment area of ice stream C widens rapidly up-stream from this point and ice-stream lines should converge into this area. Interpretation of the radio-echo clutter, however, shows the ice stream being guided by the westward leading arm of the deep trench. This ice stream shows a greater area of low shear stress than the others (Fig. 4); note the small medium stress area within this zone.

Internal layers are not seen west of the 900 m surface contour, although they appear beside the ice stream beneath the 400 m surface contour. Behind ice streams B and E, layers are seen only behind the 600 m contour.

Basal temperatures are expected to be at the pressure-melting point over the whole length of this ice stream (for 0.060 W m^{-2}) unless velocities are reduced by 50%, when the very flat area near the grounding line will freeze to the bed. At 0.045 W m^{-2}, this will happen at 75% velocities.

Surface-contour configurations and radio-echo clutter are typical of ice streams, yet velocity measurements (Thomas, 1976[b]) off the grounding line, and the surface-appearance question present activity. Suppose that this ice stream exhibits surging behaviour and is now in a quiescent phase with low velocities. Old crevasses will be being slowly buried by accumulation at about 0.1 m a^{-1}. On radio-echo records, detail within the top 100 m of the ice is lost because of the beam width causing late returns, saturation of the receiver, and differentiation of the output. The origin of the hyperbolae is within this region, i.e. within the last 1 000 years of accumulation, an upper limit on the last date of activity.

Low velocities in the zone extending 150 km inland of the grounding line will generate insufficient frictional heat to prevent basal freezing with a steady-state model. For 50 km up-stream of this area, the increase in surface slopes and basal shear stresses may indicate a build-up of ice behind a stagnant patch.

During the active phase, ice would be removed from the present inland flat area, and the side boundaries of the catchment area would retreat. Southward spreading encroached on ice stream B, causing the present apparent conflict of flow lines and catchment areas here. Ice stream B is seen now as perhaps regaining this lost territory.

Measurements of velocity and strain-rates along ice stream C should be taken to test this hypothesis.

Siple dome

This major feature of the West Antarctic coast rises to at least 650 m elevation, over 300 m above the col to the east. The surface of the underlying rock is remarkably smooth and level, falling away steeply to the north (Fig. 2). Low accumulation and diverging flow result in low balance velocities and, although basal shear stresses are high under the steep slopes (Fig. 4), frictional heating is less important than the geothermal heat flux. All the basal ice is below −7°C, regardless of velocity, at 0.060 W m^{-2}. The need for fluxes of 0.086–0.120 W m^{-2} to initiate basal melting shows the stability of this dome.

Ice stream D

This ice stream drains from the area of Byrd Station. Near the Ross Ice Shelf it converges with ice stream E and vectors determined on the Ross Ice Shelf indicate that ice stream E is dominant, forcing the combined discharge to the south of Roosevelt Island. Near Siple dome, the channel beneath ice stream D is significantly wider than the ice stream. The latter can be defined clearly from the radio-echo records and the prominent "boomerang" crevasse field. This ice stream is formed by the coalescence of smaller ice streams as governed by the bedrock (Fig. 2). Velocities are expected to reach 100 m a^{-1} at the 400 m contour, where the basal shear stress drops below 0.25 bar (Table I). With 0.060 W m^{-2}, temperatures are found to be below the pressure-melting point only at the ice divide. 0.045 W m^{-2} will allow 400 km of the catchment area (Fig. 5) to freeze.

Ridge D/E

This small ridge has a corresponding bedrock high bounded by steep sides. As with ridge B/C and Siple dome, velocities are low and basal melting is unlikely with less than 0.075 W m^{-2}. Shear stresses reach 0.6 bar.

Ice stream E

The catchment area of this ice stream extends north, out of the area surveyed by 1974–75 radio echo-sounding, to the ice divide connecting the coastal ranges of Marie Byrd Land, where accumulation-rates exceed 0.2 m a^{-1}. Ice stream E therefore has a higher discharge than the other ice streams. High activity is shown by trimetrogon photographs and Landsat imagery of active crevasse fields and distinctive boundaries. Balance velocities reach 100 m a^{-1} at the 600 m contour and 450 m a^{-1} at the grounding line. Temperature calculations show sufficient activity for basal melting to occur over the whole area of this ice stream to the limit of the radio echo-sounding, even if the geothermal heat flux is zero.

Rockefeller Plateau

Bedrock here rises steeply from ice stream E to a flat surface 200 m below sea-level, falling away to the north (Fig. 2).

Ice accumulating here flows into ice stream E or joins the small ice stream draining this plateau (designated ice stream F; Hughes, 1977), which is clearly shown on Landsat imagery.

The small area of the plateau restricts velocities. 0.075 W m^{-2} are required to cause basal melting under ice stream F, and 0.110 W m^{-2} to melt the rest of the plateau.

Conclusion

Ice streams control the discharge of ice from Marie Byrd Land into the Ross Ice Shelf. The ice streams are situated above bedrock channels, and near the Ross Ice Shelf they have low surface slopes, low basal shear stresses, and high balance velocities capable of causing basal melting. The intervening domes and ridges are expected to be frozen to their beds.

Whereas evidence points to rough agreement between calculated balance velocities at the grounding line and velocities measured in the ice shelf off ice streams B, D, and E, ice stream C seems anomalous. There is little evidence of present activity. Measurements of velocities along this ice stream should be taken to test whether ice stream C is now in the quiescent stage of a surge cycle.

Acknowledgements

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DISCUSSION

R. H. THOMAS: Could the anomaly of ice stream C be explained by comparatively recent grounding of its lower reaches to form a large area of ice which would obstruct flow of the ice stream?

K. E. ROSE: This may well be possible.

C. R. BENTLEY: It appears that it would not take much modification of the surface elevations in the Byrd Station area to lead to the conclusion that the ice from Byrd Station flows down ice streams C or E rather than D. Do you think the errors in the measurements and contouring are small enough to rule out that possibility?

ROSE: Elevations at the grid intersections were tied together by random-walk and least-squares techniques, and were fixed to Byrd Station and points on the Ross Ice Shelf, assuming hydrostatic equilibrium. While the ice-surface contours allow a wide choice of possible

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catchment area boundaries for ice stream C, we consider that errors are insufficient to allow the ice from Byrd Station not to enter ice stream D. This may not always have been the case with different surface configurations in the past.

I. M. WHILLANS: It seems to me that the evidence points to the conclusion that ice stream C has stagnated.

ROSE: I agree that this ice stream is now stagnant but accumulation will surely build up surface elevations and slopes again until velocities increase, i.e. creating a surge.

T. J. HUGHES: How did you determine where the Marie Byrd Land–Ross Ice Shelf grounding line was, and how far off do you think the grounding line shown on your map might be?

ROSE: Along flight tracks the grounding line is shown where the surface elevation first exceeds the elevation expected for a freely floating ice shelf of the same thickness. This was determined to *c.* 2 km. The grounding line was interpolated between these positions. Navigation errors are estimated to be *c.* 5 km.