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The extent of basal melting in Antarctica

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Zusammenfassung: Die ersten Resultate einer umfassenden Studie der physikalischen Eigenschaften des antarktischen Inlandeises werden vorgelegt. Als Unterlagen dienten Karten der Oberflächen- und Felsbodentopographien, des Zuwachses und der Oberflächentemperatur (10 m), die dem Sovietatlas der Antarktis und dem Kartenband der amerikanischen geographischen Gesellschaft entnommen und in einigen Einzelheiten mit neueren Expeditionsergebnissen verbessert wurden. Mit der Annahme eines ausgeglichenen Massenhaushaltes machen diese Daten (nach Glättung über Entfernungen von etwa 100 km) es möglich, versuchsweise Karten und Profile der folgenden, bisher nicht dargestellten Merkmale zu entwerfen: Stromlinien des Eises, Eisgeschwindigkeiten und Dehnungsraten bei ausgeglichenem Massenhaushalt, Wärmezufuhr und -erzeugung an der Unterseite des Eises, Tiefenprofile der Temperatur, dielektrische Absorption von elektromagnetischen Signalen, vertikale Temperaturgradienten in der Oberflächenschicht des Eises, Temperaturen an der Felsgrenze unter dem Eis und Schmelzmengen.

Die Temperraturrechnungen stellen elektronisch-numerische Lösungen der Temperaturleitungsgleichung für Eissäulen dar und folgen deren Bewegung von der „Eisscheide" zum Rande des Inlandeises entlang den Stromlinien, die von den Mittelpunkten der westlichen und östlichen Antarktis ausgehen. Diese Rechnungen haben zu den hier vorgelegten Felsbodenisothermen und Schmelzgebieten geführt.

Die Resultate zeigen, daß unter dem größten Teil des ostantarktischen Eises die Temperaturen weit unter dem Druckschmelzpunkt liegen, dem sie sich jedoch schnell der Küste zu nähern. Andererseits wird der Druckschmelzpunkt im Inneren der Westantarktis erreicht, wo isolierte Senkungsgebiete im Felsuntergrunde oder hohe Eisgeschwindigkeiten vorkommen.

Abstract: The material presented summarizes the first results of a comprehensive study of the physical characteristics of the Antarctic ice sheet. The basic data used are maps of surface and base topography, accumulation rate, and surface (10 m) temperature, published in the Soviet Atlas of Antarctica and Map Folio 2 of the American Geographical Society and modified in some details by means of more recent expedition results. On the assumption of a balanced mass budget these data (smoothed over distances of the order of 100 km) have made it possible to construct tentative plan maps and profiles for the following features not previously mapped: ice cap flow lines, balance ice flow velocities and strain rates, basal heating rates, temperature-depth profiles, dielectric absorption of radar signals, vertical temperature gradients in the ice near the surface, basal temperatures, and melt rates.

The ice temperature calculations have involved following ice columns from the ice divides to the edge of the ice sheet along ice flowlines radiating from the centres of East and West Antarctica, continuously solving the heat conduction equation by digital computer. These calculations have provided the basal isotherms and outlines of melt regions presented in this paper. The results indicate that most of the East Antarctic ice sheet has base temperatures far below the pressure freezing point which is however rapidly approached near the coast. On the other hand the ice reaches the pressure melting point in pockets of depressed bed rock or regions of high velocity in the interior of West Antarctica.

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1. INTRODUCTION AND ASSUMPTIONS

The coverage of data on the physical characteristics of the Antarctic ice cap has reached a stage where it is possible to derive many other physical features of its dynamics and history on a broad regional view. Although in places no data exist for several hundreds of kilometers, these gaps are still small compared to the diameter of Antarctica, 4,000 km. Hence if as a first approximation we select points at about 200 km spacing and interpolate the available data for regional averages, we find that to cover the area of Antarctica $\sim$ 12,000,000 km^2 about 300 points are required.

The present paper gives the results of part of the first stage of a large project using such regionally smoothed values of ice surface elevation E, ice thickness Z, surface temperature Θ_s, und accumulation rate A, to calculate other physical feature. The techniques used have been developed with sufficient generality to ensure that as further data becomes available, new results will follow.

In order to calculate the history of the Antarctic ice cap, it is necessary to know its velocity distribution. To calculate the velocity the temperature distribution is required. Finally, to calculate the temperature distribution both the velocity and history need to be known. Because of the interaction of these parameters the present approach is to introduce a series of simplifying assumptions in calculating the first parameter, then using the distribution of this parameter to remove certain of the simplifications in calculating the other parameters. The more is known about any one parameter the more can be calculated for the others. The order chosen for this procedure is as follows: Temperature profiles are first calculated for an ice cap in steady-state balance i. e. it is assumed that Z, A, Θ_s have been constant with time at their present values.

This implies that the velocity distribution is also constant with time, and can be calculated as the "balance velocity" V from the present values of A and Z.

By comparing these calculated values of velocity and temperature profiles with measured values much can be learned about the present state of balance and the past history of the ice cap. New temperature and velocity distributions can then be calculated without the restrictive balance assumption.

As a first stage in this project balance velocities and steady-state temperatures have been calculated which allow a first estimate to be made of the basal temperatures over the Antarctic and of the extent of bottom melting and freezing.

2. MODELS FOR TEMPERATURE CALCULATIONS

2.1 *Stationary column*

Two basic models have been used for heat conduction in a moving medium to calculate the temperature distributions. The first, called the steady-state "Eulerian" column, has been discussed by *Radok, Jenssen and Budd (1970)*. In this model a column fixed in space over a given point in the bedrock is considered and the conduction of heat in the ice moving through it is calculated. For the simplest case with ice thickness Z, accumulation rate A, strain rate $\frac{A}{Z}$, horizontal advection and warming rate $\frac{D\Theta}{Dt}$, and thermal diffusivity of ice $\varkappa$ (all constant with depth) the differential equation for temperature Θ at level z above the bedrock may be written,

$$\varkappa \frac{d^2\Theta}{dz^2} + \frac{Az}{Z}\frac{d\Theta}{dz} = \frac{D\Theta}{Dt} \qquad (1)$$

The boundary conditions used are the basal temperature gradient γ_b and the surface temperature Θ_s. For a balanced state the surface warming rate following the motion is given by

$$\frac{D\Theta}{dt} = V\frac{\partial\Theta_s}{\partial x} = \alpha V\lambda \tag{2}$$

where α is the surface slope, V is the forward velocity (assumed constant with depth), and λ is the vertical gradient of annual mean temperature at the snow surface. For this simple column model with "*basal heating*" the basal gradient may be calculated from the geothermal heat flux gradient γ_G and the frictional heating, concentrated at the base, by

$$\gamma_b = \gamma_G + \frac{\tau_b V}{JK} \tag{3}$$

where τ_b is the basal shear stress, K is the ice conductivity and J the mechanical equivalent of heat.

Although analytical solutions for this equation are available, it is generally simpler, for large numbers of profiles, to use direct computer solutions by numerical finite difference methods. Generalisations to the above simple model are then readily incorporated. These include:

(i) Varying velocity and strain rate with depth. In this case the velocity may be prescribed as a function of height above bedrock (V_z) and the strain rate may be taken as,

$$\dot{\varepsilon}_z = \frac{A\,V_z}{Z\,V_s} \tag{4}$$

where V_s is the velocity at the surface.

Alternatively the shear strain rate $\dot{\varepsilon}_{xz}$ may be prescribed as a function of shear stress and temperatures Θ_z at height z above bedrock. This may then be integrated together with equation (1) to give both the temperature and velocity profiles as a function of a prescribed surface velocity.

(ii) *Internal heating.* As a result of the velocity varying with depth the heat production Q_z at height z above bedrock may be calculated from,

$$Q_z = \tau_{xz}\;\dot{\varepsilon}_{xz} \tag{5}$$

Here $\dot{\varepsilon}_{xz}$ is temperature dependent and so this equation must be solved simultaneously with the heat conduction equation.

(iii) *A variable diffusivity* may be prescribed as a function of depth or more appropriately as a function of temperature, (cf. Ratcliffe 1962) and this effect incorporated into the integration of equation (1).

These three modifications have been studied in detail and for typital conditions, have been found to cause only minor variations in the profile compared to typical variations in the major parameters Z, A, γ_b and $\frac{D\Theta}{Dt}$. Hence at this stage they may be considered as higher order refinements.

(iv) For an ice cap not in balance the warming rate may be taken as

$$\frac{D\Theta}{Dt} = (\alpha V + \frac{\partial Z}{\partial t})\lambda \qquad (6)$$

where $\frac{\partial Z}{\partial t}$ is the rate of surface lowering.

(v) When the base temperature reaches the melting point then the lower boundary condition becomes that the base temperature equals the pressure melting point and then the difference between the calculated base gradient, γ_c, and the geothermal gradient γ_b, determines the melt rate M from

$$M = (\gamma_b - \gamma_c)\frac{K}{L} \qquad (7)$$

where L is the latent heat of fusion for ice.

2.2 *Moving column model*

The second model, called the "Lagrangian column" is a development of the work of Jenssen and Radok (1963) whereby a column of ice is followed with moving co-ordinates from the centre of the ice cap along a flow line to the edge. In this case the warming rate is simply prescribed by the surface temperature as a function of distance along the flow line. The equation is then the same as (1) except that the co-ordinates are moving along the flow line with prescribed speed V and expanding or contracting according to the ice thickness; with melting the second term becomes $\frac{A}{Z}(z + M[1 - \frac{z}{Z}]) / \frac{\partial\Theta}{\partial z}$ where M is the local melt rate.

3. INFORMATION REQUIRED AND DATA AVAILABLE

For the Eulerian column the data required are for each point: the ice thickness (Z) (or surface and bedrock elevation E and b), accumulation rate (A), surface warming ($\alpha V\lambda$), base gradient γ_b, and the surface temperature Θ_s.

For the Lagrangian column Z, A, Θ_s and γ_b and V need to be specified as a function of distance along a flow line. In addition for both models the parameters K and γ_G have to be specified.

Data for Z, A and Θ_s for Antarctica are already available as compiled and mapped by Bentley et al. (1964) and Bakayev (Ed.) (1966). These have been updated by more recent traverse data from USARP SPQML7 traverses I and II (Cameron et al., 1968; Clough et al., 1968; Beitzel, 1969; Kane [personal communication]) and ANARE traverses inland of Wilkes and to the Amery Ice Shelf (Battye [unpublished]; Budd, 1966; Corry [personal communication]). A complete report including all these data maps with numerical values and all the derived maps and profiles is in preparation.

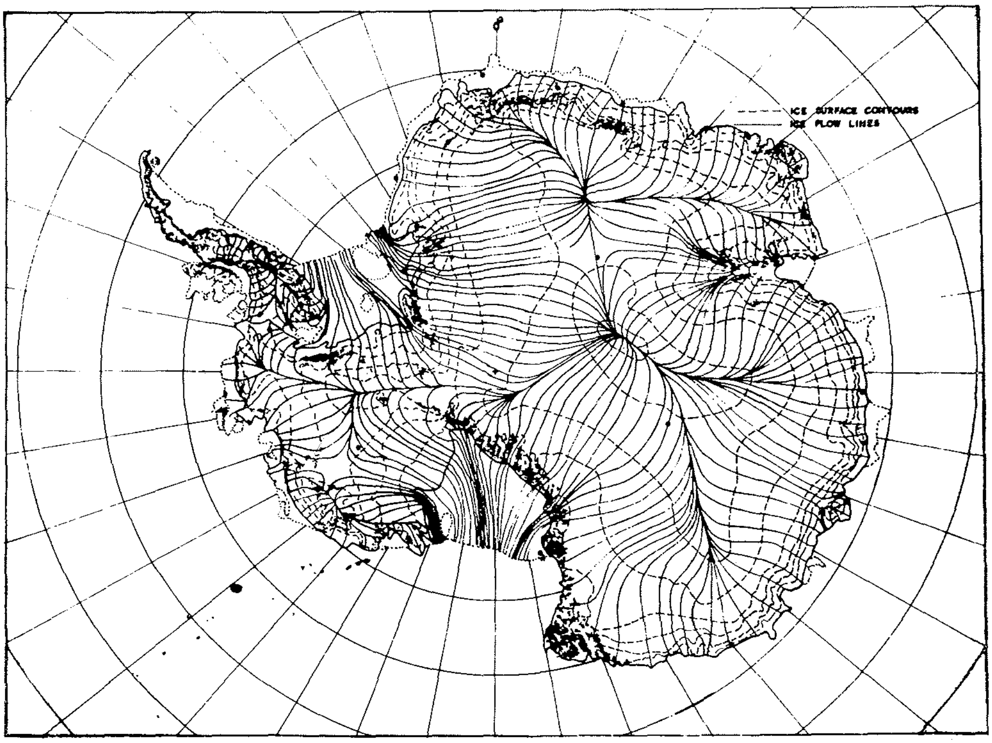
4. CALCULATIONS FROM DATA MAPS

From the elevation contours orthogonals have been constructed at approximately 100 km spacing. These have been used to represent the flow line patterns and are shown in Fig. 1. To select data points the 500 m elevation contours from 1,000 m to 4,000 m have been divided into 200 km segments. Sectors are thus defined by these segments and the flow lines through their end points. The data points have been chosen as the centres of these segments. This distribution rather than even spacing has been adopted for greater detail near the edges where the rates of change are greater. For the large ice shelves the flow lines and data points have been extended to the seaward edges. In other areas an extension to elevations less than 1,000 m has been avoided because of the high irregularity of conditions in those coastal regions.

Using these flow lines and the maps for ice thickness and accumulation rate the balance velocities have been calculated from

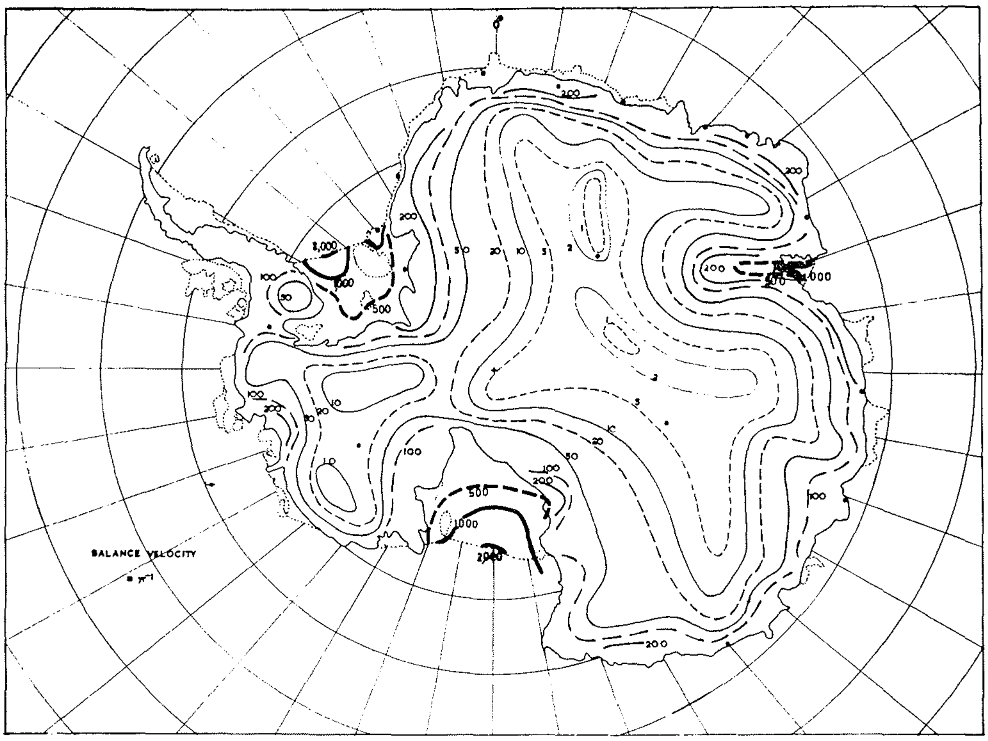
$$\frac{d(VYZ)}{dx} = AY \tag{8}$$

Y is the distance between the flow lines at distance x where the ice thickness is Z and the accumulation rate A.



1. Flowlines. The 500 m elevation contours for Antarctica from the 1:20,000,000 map of the Australian Division of National Mapping have been used to construct orthogonal trajectories with about 100 km spacing and smoothing to represent the general ice flow directions and drainage patterns for the Antarctic ice cap.

1. Stromlinien. Von den 500 m Höhenlinien in der von der australischen Division of National Mapping herausgegebenen Karte der Antarktis (Maßstab 1:20 Millionen) sind orthogonale Trajektorien mit etwa 100 km Abstand und Glättung entworfen worden, die allgemeine Strömungsrichtung und Abflußgebiete des antarktischen Eises wiedergeben.



2. Balance Velocity. From maps of ice thickness Z, accumulation rate A, and ice flow lines (distance apart Y at distance X) the average velocity V required for a balanced state has been calculated as $V = \frac{\iint AdYdx}{YZ}$ for sectors formed by the flowlines and 200 km segments of the 500 m spacing elevation contours

2. Geschwindigkeiten bei ausgeglichenem Massenhaushalt. Die Mittelgeschwindigkeit des Eises, die nötig ist, den Massenhaushalt auszugleichen, ist mit Hilfe von Karten der Eisdicke Z, des Zuwachses A, und der Stromlinien (mit Abstand Y bei der Entfernung X) aus der Gleichung $V = \frac{\iint AdYdx}{YZ}$ für Sektoren berechnet worden, die von den Stromlinien und 200 km Segmenten der Höhenlinien im Abstand von 500 m gebildet sind

Integration gives

$$V = \frac{\int AYdx}{YZ} \tag{9}$$

Hence the velocity may be obtained by integration (planimetric or numerical) of the accumulation rate over the sector and dividing by the ice thickness and sector length. The results of these calculations for balance velocities are shown in Fig. 2.

By obtaining gradients along the flow lines as

$$\alpha = \frac{dE}{dx} \text{ and } \lambda = \frac{d\Theta_s}{dx} \tag{10}$$

the surface slope and vertical temperature gradients are calculated for each of the data points. From these the balance steady-state warming rates are calculated as

$$\frac{D\Theta_s}{Dt} = \alpha V \lambda \tag{11}$$

and are shown in Fig. 3.

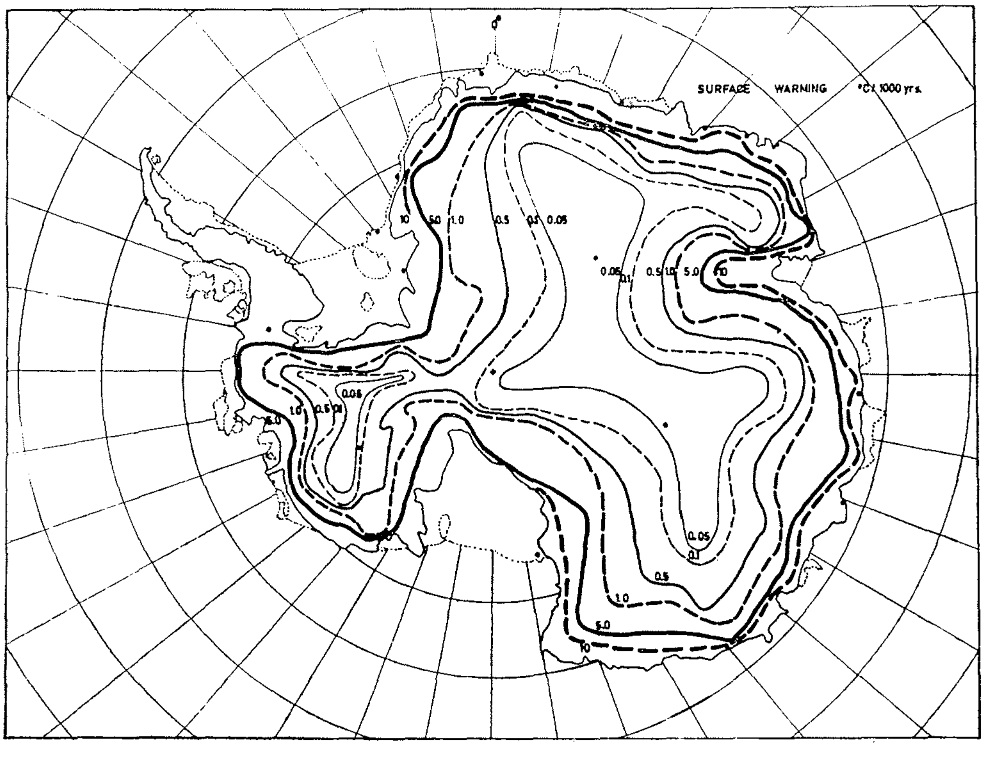
From the ice thickness and surface slope the basal stress is calculated as

$$\tau_b = \varrho g \alpha Z \tag{12}$$

and is shown in Fig. 4.

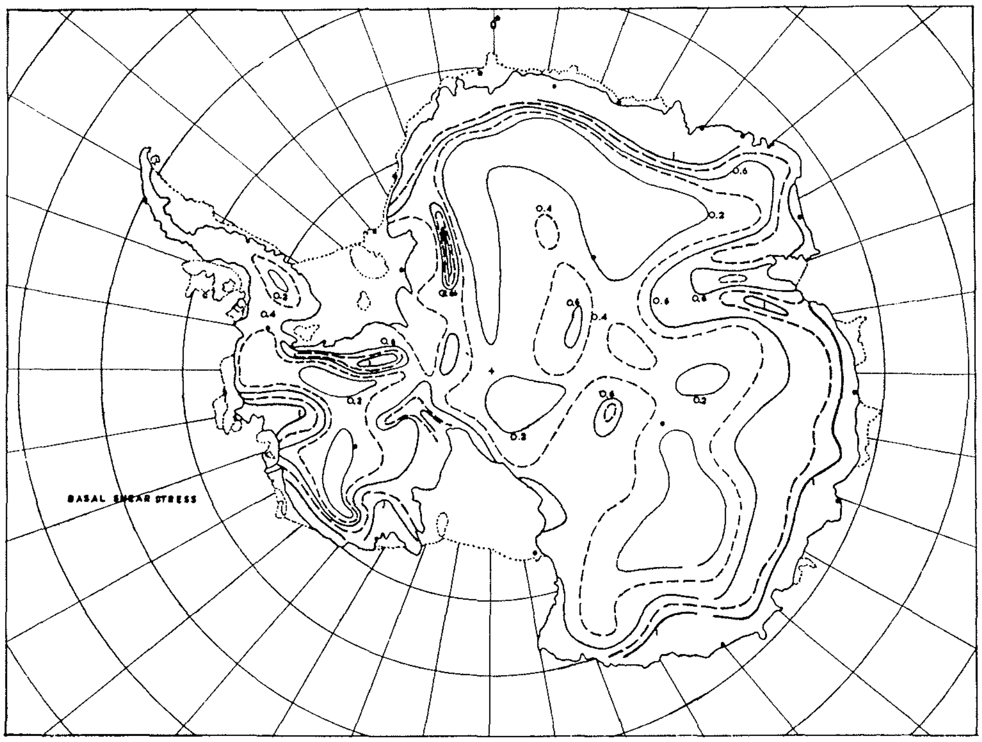
The base gradient (for basal heating in absence of melting) is then calculated from

$$\gamma_b = \gamma G + \frac{\tau_b V}{JK} \tag{13}$$



3. Surface Warming. From the maps of average velocity V and surface temperature Θ_s the rate at which the surface of the ice warms as it flows outwards along the flowlines (distance x) has been calculated as $\frac{D\Theta_s}{Dt} = V\alpha\lambda$, assuming that the ice cap shape and temperature distribution have remained constant with time

3. Oberflächenerwärmung. Die Erwärmung der Eisoberfläche während der Bewegung des Eises auf den Stromlinien (Entfernung x) nach außen ist mit Hilfe von Karten der Mittelgeschwindigkeit V und der Oberflächentemperatur Θ_s aus der Gleichung (11) mit der Annahme berechnet worden, daß die Form des Inlandeises und seine Temperaturverteilung sich mit der Zeit nicht geändert haben



4. Basal Stress. From the maps of ice thickness Z and surface slope α along the flowlines, smoothed over ~200 km, the basal stress τ_b (bars) has been calculated as $\tau_b = \varrho g \alpha Z$ for each 200 km segment of the 500 m elevation contours

4. Scherspannung am Felsboden. Die Scherspannung τ_b (Bars) ist mit Hilfe von Karten der Eisdicke Z und der Oberflächenneigung α (entlang den Stromlinien gemessen und über etwa 200 km geglättet) aus der Gleichung $\tau_b = \varrho g \alpha Z$ für alle 200 km Segmente der 500 m Höhenlinien berechnet worden

For East Antarctica in the present calculations the geothermal heat flux has been taken as 1 μ cal cm^{-2} sec^{-1}, typical of Precambrian Shields (cf. Lee and Uyeda, 1965). For the geologically younger West Antarctic the value chosen here is 1.2 μ cal cm^{-2} sec^{-1}. The variation in this base gradient over the Antarctic is illustrated in Fig. 5.

Finally from these values of Z, A, $\alpha V\lambda$, γ_b, and Θ_s temperature profiles have been calculated. For the moving "Lagrangian" column some 22 flow lines were selected to give a representative coverage and values read off at 100 km intervals. The computer output was for every 20 km but here we have used only the 100 km points. As an example of the basal temperature results from the moving column calculations Fig. 6 shows the base temperatures and melt regions as obtained using the data as described above.

The computer output for each position includes the following information:

1) the input data
2) depth profiles of: — temperature, temperature gradient, 2nd derivative, warming rate, velocity, integrated dielectric absorption

Table I

Example of computer output every 20 km along each flowline.

N.PTS	Z.INC	DEPTH	THRM.DIFF	BASE.TEMP	TAU	INST.MELT	TOTL.MELT	U.PICK
26	86.5602	2164.0060	42.7300	0.0000	0.4104	0.0000	0.0000	-2.0000

T.INC	TIME	T.PRINT	X.DIST	X.PRINT	T.EXIT	X.EXIT	PUNCH	SCALE
1.0000	7320.00	999999.00	100.0051	100.0000	999999.00	100.0000	1	1.0000

VARIABLE		CURRENT VALUE	RATE OF CHANGE PER UNIT TIME	VARIATION COEFFICIENTS ALONG X LINEAR	QUADRATIC
BEDROCK HEIGHT	(METRE)	-0.61374E 03	0.	-0.44300E-02	-0.19860E-06
ACCUMULATION	(M/YR)	0.21700E 00	0.	-0.55000E-07	-0.11000E-11
SURFACE TEMPERATURE	(DEG)	-0.28000E 02	0.	0.20000E-04	0.80000E-10
GEOTHERMAL HEAT FLUX	(DEG/M)	0.23364E-01	0.	-0.	0.
HEIGHT ABOVE SEA LEVEL	(METRE)	0.15500E 04	0.	0.16500E-02	0.50000E-08
DIVERGENCE ANGLE	(DEGREES)	-0.	0.	-0.	0.
HORIZONTAL VELOCITY	(M/YR)	0.15000E 02	0.	0.22000E-04	0.60000E-10

DEPTH	TEMPERATURE	TEMP.DIFF	TEMP.GRAD	2ND.DERIV.T	INTGD.ABSN	WARMING	TRAJ.NO	TRAJ.DEPTH
0.	-28.000	-0.000	********	********	0.000	0.420	1	2196.
87.	-28.194	-0.194	-0.229	-1.141	0.881	0.429	2	1825.
173.	-28.396	-0.397	-0.238	-0.911	1.754	0.437	3	1426.
260.	-28.606	-0.606	-0.245	-0.557	2.618	0.444	4	887.
346.	-28.819	-0.819	-0.248	-0.033	3.473	0.450		METRE
433.	-29.033	-1.033	-0.245	0.727	4.319	0.456		
519.	-29.240	-1.241	-0.234	1.805	5.157	0.463		
606.	-29.435	-1.435	-0.212	3.299	5.987	0.473		
692.	-29.604	-1.604	-0.175	5.323	6.810	0.486		
779.	-29.732	-1.733	-0.118	7.990	7.626	0.505		
866.	-29.801	-1.801	-0.035	11.401	8.440	0.532		
952.	-29.783	-1.783	0.081	15.617	9.252	0.568		
1039.	-29.648	-1.648	0.238	20.637	10.067	0.613		
1125.	-29.358	-1.358	0.441	26.361	10.889	0.667		
1212.	-28.870	-0.870	0.696	32.570	11.727	0.727		
1298.	-28.138	-0.138	1.005	38.910	12.590	0.790		
1385.	-27.115	0.885	1.368	44.898	13.491	0.849		
1472.	-25.756	2.244	1.780	49.940	14.449	0.898		
1558.	-24.023	3.976	2.228	53.347	15.488	0.925		
1645.	-21.893	6.107	2.696	54.248	16.647	0.914		
1731.	-19.359	8.641	3.157	51.261	17.978	0.821		
1818.	-16.444	11.556	3.566	42.024	19.556	0.559		
1904.	-13.220	14.780	3.856	23.528	21.485	0.003		
1991.	-9.826	18.173	3.941	-5.309	23.894	-0.910		
2077.	-6.476	21.524	3.750	-39.680	26.922	-2.022		
2164.	-3.423	24.577	********	********	30.683	-2.929		
METRE	DEG.CENT	DEG.CENT	DEG/100M	DEG/KM/KM	DECIBEL	DEG/1000YR		

TIME ELAPSED	=	7320.000	YEARS
DISTANCE MOVED	=	400.005	KILOMETERS
LATERAL DIVERGENCE	=	255.479	PER MILLION YR
STREAMLINES CONVERGE AT		-58.714	KILOMETERS DOWNSTREAM
CURRENT MELT	=	0.000	MICRON/YR
INTEGRATED CURRENT MELT	=	4241.943	MILLIMETRE
COLUMN SURFACE WARMING	=	0.420	DEG/1000YR
COLUMN ELEVATION CHANGE	=	-32.251	METRE/1000YR
IMPLIED SURF. LAPSE RATE	=	1.302	DEG/100M
DIELECTRIC ABSORPTION	=	30.683	DECIBEL (ONE WAY)
ABSORPTION TEMPERATURE	=	-22.000	DEGREES
SURF GRADIENT OF COMPUTED TEMPERATURES	=	-0.219	DEGREES PER 100 METRES
Base " " " "		3.244	

3) trajectory paths
4) time of travel
5) melt rates and total melt (or freezing).

An example of such a printout is shown in Table 1. Only the basal temperatures and melt rates will be examined here.

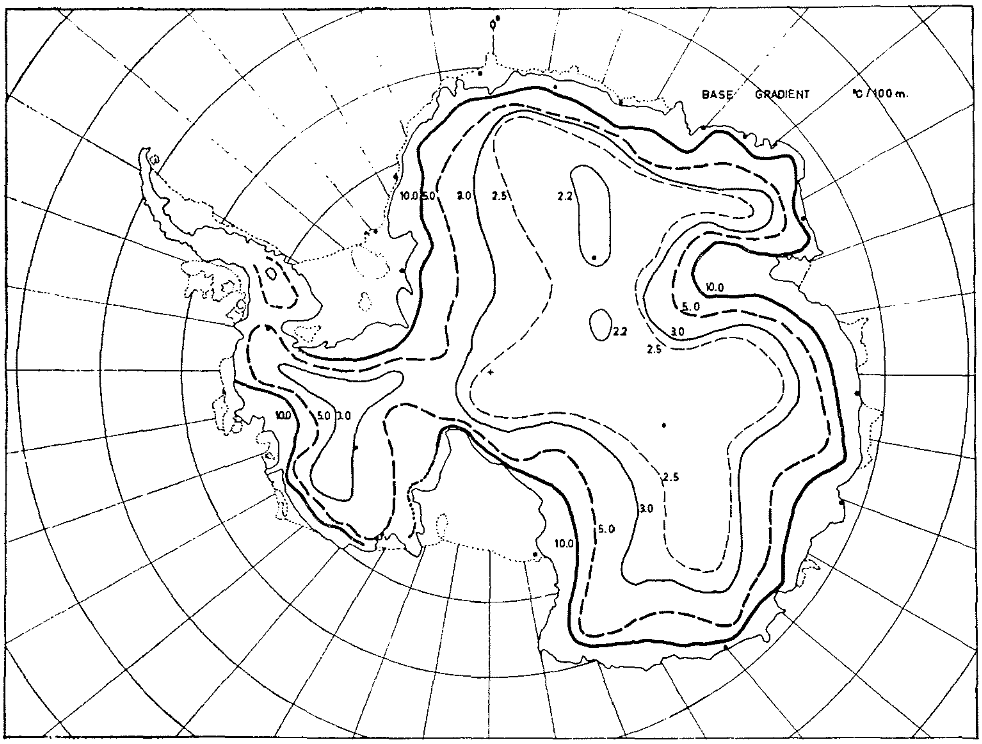
5. DISCUSSION OF RESULTS

5.1 *Base temperatures*

We first consider the map of basal temperature contours, Fig. 6. This shows that on the broad scale most of the base of the ice cap in inland Antarctica is cold with temperatures beween — 30° and — 10° C. In general as the ice flows towards the coast or into the large ice shelf basins the surface decreases in elevation and warms up and the base temperatures approach melting. In most cases melting does not set in until quite near the coast or near the ice shelf edge — generally not until the surface is below the 2,000 m elevation contour.

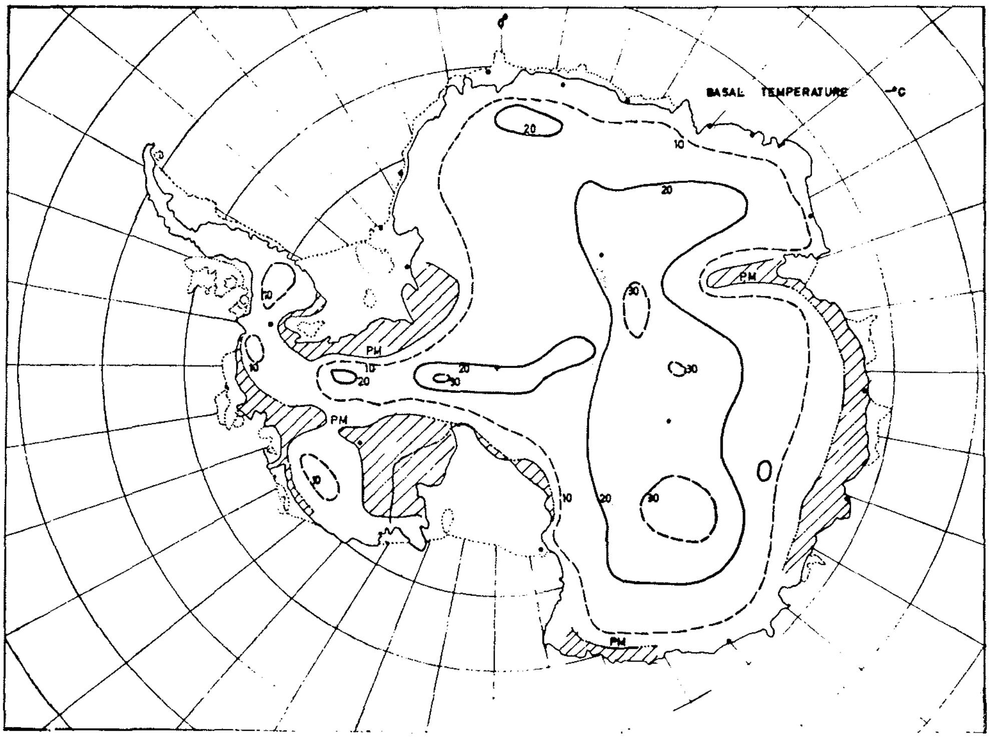
A similar pattern of basal temperature variation from the centre to the edge of East Antarctica was found by Shumsky (1970). However the inclusion of the term for surface warming in the present treatment results in the temperatures here not reaching melting until much nearer the edge than indicated by Shumsky's results.

In West Antarctic the low areas around the ice shelves are at pressure melting, but the most extensive area with the base temperature at pressure melting is the region of deep ice with comparatively low surface elevation in Byrd Land, and extending to the



5. Base Gradient. The temperature gradient near the base of the ice γ_b (in absence of melting) has been calculated as the sum of the geothermal heat flux gradient γ_G (taken as 2.2 and 2,6 °C/100 m in East and West Antarctic respectively) and the heat generated by the motion, assumed as being produced in a thin basal layer, $\gamma_b = \gamma_G + \frac{\tau_b V}{JK}$, where τ_b is the basal stress, V the average velocity, K the thermal conductivity, and J the mechanical equivalent

5. Bodengradient. Der Temperaturgradient γ_b am Boden des Eises (ohne Schmelzen) ist aus der Gleichung $\gamma_b = \gamma_G + \frac{\tau_b V}{JK}$ berechnet worden. γ_G entspricht dem geothermischen Wärmestrom und wurde als 2.6 °/100 m für die Ostantarktis und 2.2 °/100 m für die Westantarktis angenommen. Das zweite Glied ist die durch die Eisbewegung erzeugte Wärme, (als in einer dünnen Bodenschicht konzentriert angenommen) τ_b ist die Bodenscherspannung, V die Mittelgeschwindigkeit, K die Wärmeleitfähigkeit, und J das mechanische Wärmeequivalent



6. Basal Temperatures. From the maps of ice thickness, accumulation rate, surface temperature, velocity and basal stress, temperature depth profiles have been calculated following ice columns along 22 representative flowlines to the coast. The resulting basal temperatures have been contoured each —10 °C and for pressure melting point P. M. The shaded areas indicate bottom melting or freezing

6. Bodentemperaturen. Tiefenprofile der Eistemperatur sind mit Hilfe von Karten der Eisdicke, des Zuwachses, der Oberflächentemperatur, der Geschwindigkeit, und der Bodenscherspannung für Eissäulen entlang 22 Stromlinien berechnet worden, die ein zusammenhängendes Bild des Inlandeises ergeben. Die resultierenden Temperaturen am Boden des Eises sind durch Isothermen in Abständen von 10 °C und durch die Druckschmelzpunktisotherme dargestellt. Schraffierte Flächen bedeuten Schmelzen oder Gefrieren am Boden des Eises

Ross Ice Shelf. However, although the inner part of this region is melting, refreezing sets in as the ice shelf is approached; this suggests there is an inland melt lake under the ice of Byrd Land.

The magnitude of the melt rate is generally less than 10 mm/yr everywhere, except right near the coast where the rates at the 1,000 m elevation contour reach 60 mm/yr in the Lambert Glacier and about 30 mm/yr inland of Wilkes station, the Ninnis Glacier and the Amundsen Sea.

5.2 *Flow line profile*

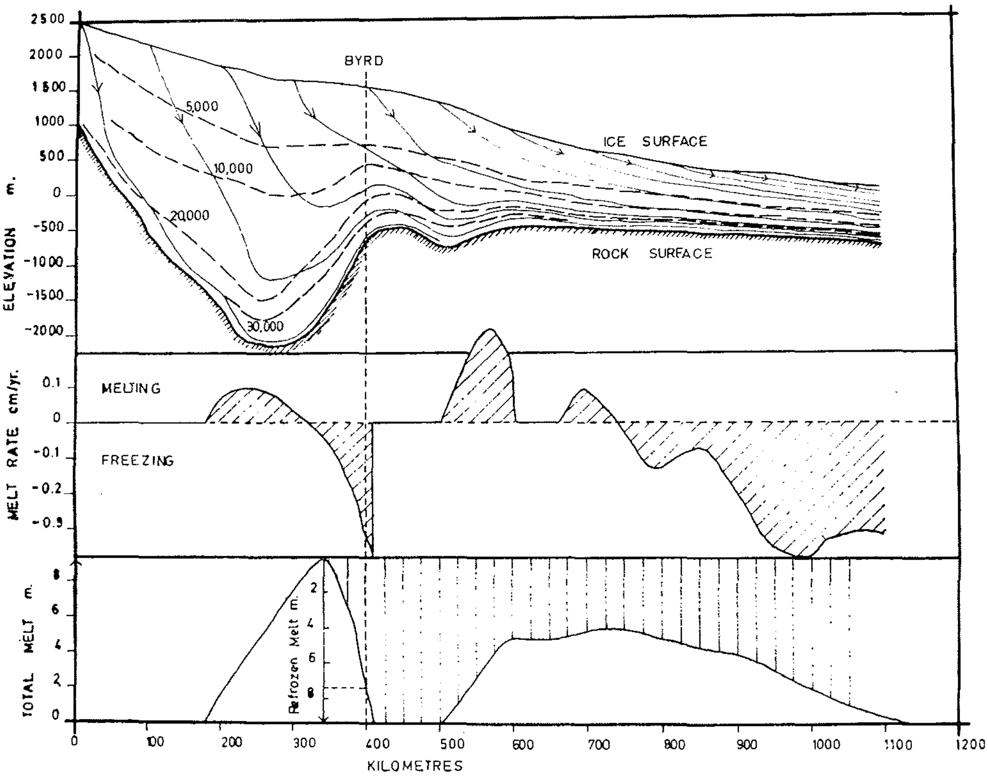
Of all the flow lines studied the one with the most interesting sequence of melt-freeze-melt alternation was the line in west Antarctica through Byrd station to the Ross Ice Shelf. The profile along that flow line (Fig. 7) shows that melting starts in the deep trough where the ice thickness is over 3,500 m. The meltwater refreezes on to the base of the ice as the ice flows up over a ridge, and this is followed by further melting and finally refreezing as the ice shelf is approached.

The amount of refrozen melt under Byrd station from Fig. 7 is about 7 m. This may be compared with about 5 m reported by Gow et al. (1968) from their examination of the cores of the deep drilling project at Byrd. Calculations with other values of γ_b indicate that a 20 % increase in geothermal flux increases the refrozen melt value to 13 m, whereas for a 20 % decrease the temperature drops below the pressure melting point at Byrd.

6 POSSIBLE VARIATIONS TO THE PRESENT OUTPUT

Finally to consider how closely the calculated values here may approach reality we look at the possible variations in the assumptions and input data.

Firstly the velocities for balance are most interesting in their own right. The calculated values for the velocity of the front of the Ross and Amery ice shelves are about double their measured values (cf. Dorrer et al 1969; Budd, 1966). This supports the calculations of many authors, e. g. Giovinetto (1964), Giovinetto et al. (1966), Giovinetto and Zumberge (1968), Budd et al. (1967), Loewe (1967), Bardin and Suyetova (1967), that



7. Flow Line Profile. For the flowline in west Antarctica through Byrd the profile of ice particle paths and the age of the ice (years) for steady-state are shown with the results of a typical calculation of basalt melt (or freezing) rates and the total amount of ice melted and refrozen, assuming that the melt water stays with the column

7. Ein Stromlinienprofil. Eistrajektorien und Alterslinien (in Jahren) bei ausgeglichenem Massenhaushalt werden für die Stromlinie durch die Byrdstation in der Westantarktis gezeigt, zusammen mit den Resultaten einer typischen Berechnung des Schmelzens oder Gefrierens am Boden des Eises und der geschmolzenen oder von neuem gefrorenen Gesamteismenge (unter der Annahme, daß das Schmelzwasser in der betrachteten Eissäule bleibt).

the Antarctic is not in steady-state but has a positive balance or excess in gain over loss by about a factor of 2. This means the actual velocities may be about half those of Fig. 2. Using lower velocities has the effect of lowering the base gradient and contributing to lower temperatures.

Secondly we consider the surface warming. If in fact the ice cap is out of balance and the surface is rising then the surface warming rates would be less than those used here. This effect is opposite to the previous one in that it would contribute to higher temperatures.

Higher geothermal fluxes would also contribute to warmer temperatures — but the values would have to be much higher to bring the central areas of Eastern Antarctica near to melting point.

Lastly the assumption of steady-state can be relaxed with the moving column model but it then becomes necessary to express the parameters A, Z, Θ_s, V as functions of time as well as position.

In the next phase of the programme systematic variations in V, γ_b and the other input parameters will be carried out to study their effect on the distribution of basal temperatures, melt rates and all the other calculated parameters in the output.

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22nd International Geographical Congress 1972

The 22nd Congress will open in Montreal, Canada on August 10, 1972. The program for the following week will include the presentation of technical papers, special planels and lectures, a series of workshops, films, an exhibition of maps, atlases, books and photographs, and local and regional excursions. Most of the Commissions plan to meet at host universities across Canada prior to the Montreal meetings. A broad program of symposia and field tours will take place before and after the main Congress, extending from coast to coast and from the United States border to the Canadian Arctic.

Throughout the Congress, the Canadian organizers will encourage active participation in discussion, as well as joint meetings to promote the dynamic interchange of ideas between scientists of diverse specialization. A compromise will be sought between the traditional formal offering of papers and novel methods of presentation. While the 13 Sections are broad enough to encompass the major areas of the study of geography, the proposed themes will set the focus on current problems and new research and concepts. The deadline for technical papers will be September 1, 1971.

Details of the Sections, Symposia, Commission meetings and Field Tours are listed in the *First Circular* which will appear in the May issue of the *IGU Bulletin*, 1970. Provisional registration for the Congress should be made by September 15, 1970. Separate copies of the *First Circular* and the application form may be obtained on request from the *Executive Secretary, 22nd International Geographical Congress, P. O. Box 1972, Ottawa, Canada.* Provisional registrants will be placed on the mailing list for the *Second Circular* to be issued in November, 1970.