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STATE OF BALANCE OF THE ICE SHEET IN THE ANTARCTIC PENINSULA

by

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ABSTRACT

Data from ice rises on the east coast of the Antarctic Peninsula can be interpreted as showing that the ice is thinning at rates of up 0.5 m a^{-1}. However, a level line between two nunataks in Palmer Land showed no change in surface elevation over a period of 5 a. Melt rates on George VI Ice Shelf vary with position and may indicate that parts of the ice shelf are thickening at the rate of several m a^{-1}, presumably in response to a higher accumulation rate over the peninsula a few hundred years ago. A small valley glacier, Spartan Glacier, is wasting away at about 0.27 m a^{-1}. Ice fronts on both east and west coasts of the peninsula have been retreating for the last 30 a. It seems that there is general glacier recession in response to a warmer climate and decreased snowfall for at least the last 30 a, while parts of the peninsula are still thickening in response to a high accumulation rate several hundred years ago.

1. INTRODUCTION

Glaciological research is often justified because of its links with climate and climatic change. The Antarctic Peninsula lies in an area sensitive to environmental change and so is likely to be one of the first places to give an indication of trends which may affect the rest of the Antarctic ice sheet. The British Antarctic Survey (BAS) has carried out a number of studies in different ice-flow regimes and climatic environments over the past few years (Fig.1). Results from these studies are examined in this paper to see if they can be interpreted as showing a consistent picture for the state of balance of the ice sheet in the peninsula.

Many of the experiments were undertaken in support of the objectives of the Glaciology of the Antarctic Peninsula (GAP) project (Swithinbank 1974), which is an international attempt to investigate the climatic history of the Antarctic Peninsula in order to bridge the gap between the main Antarctic continent and South America.

2. SURFACE-LEVEL AND VELOCITY PROFILES

Martin and Sanderson (1980) have reported work on several small ice rises on the Larsen Ice Shelf on the east side of the peninsula. From measurements of velocity and from surface elevation profiles, they deduced values for the parameters n and B in the flow law of ice expressed as

$$\dot{\varepsilon} = (\tau/B)^n,$$

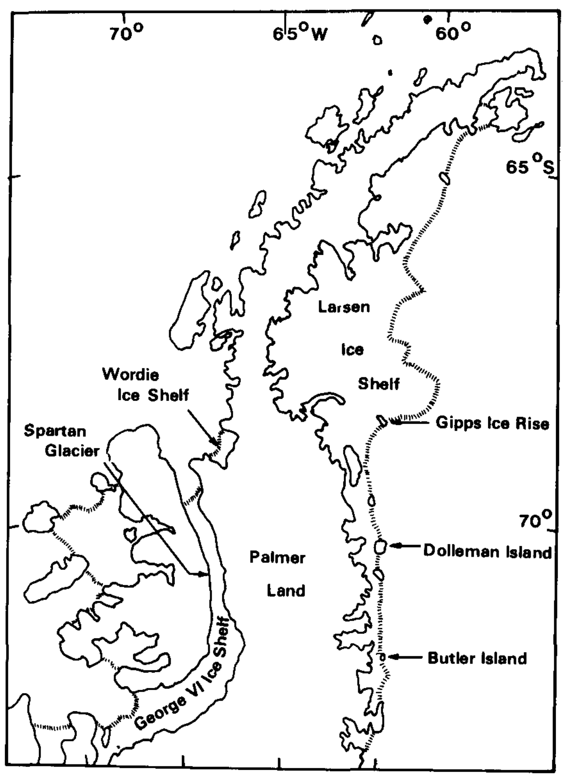

Fig.1. Sites in the Antarctic Peninsula considered in this paper where the state of balance can be inferred.

where $2\dot{\varepsilon}^2 = \dot{\varepsilon}_{ij}\,\dot{\varepsilon}_{ij}$ and $2\tau^2 = \sigma'_{ij}{}^1\,\sigma'_{ij}{}^1$ are the second invariants of the strain-rate and stress-deviator tensors. Surface velocities were measured on the west axis of Butler Island. In their analysis, the measured value of n was found to correspond within experimental error with the theoretical value of 3. Assuming n to be exactly 3 gave a value of B' (a weighted average over depth of B) of (1.01±0.11) x 10^8 N m^{-2} $s^{1/3}$ for a temperature of about -13.5°C. By contrast, surface elevation profiles of the east axis of Butler Island gave a value of B'' (another weighted average of B and typically of the order of

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4% less than B') of (1.49±0.26) x 10^8 N m^{-2} $s^{1/3}$. Martin and Sanderson (1980) described the discrepancy between the values for B' and B'' as being not very significant, saying that it may be due to non-steady state behaviour, and pointing out that Thomas and others (1980) suggested that the effects of recrystallization and fabric softening may account for the discrepancy. However, while bearing this in mind as a possibility, it is interesting to attempt an explanation for the discrepancy by considering non-steady state behaviour.

Martin and Sanderson showed that

$$B'' \propto a^{-1/3} \qquad (1)$$

where a is the accumulation rate. For Butler Island the accumulation rate was deduced from annual oxygen isotope variations in snow cores drilled to 10 m depth. Thus, the accumulation rate is the average value over the last 10 to 15 a, and was taken to be 0.23 m a^{-1} of ice. Let us assume that the value of B' is the "correct" one and postulate that the difference between it and the value of B'' is due to the measured accumulation rate a_m being less than that needed for steady state. Then the "balance" accumulation rate which would give a value of B'' equal to that of B' (or, more precisely, 4% less than B') is 0.83 m a^{-1}. This implies that, at present, the surface elevation of Butler Island is lowering at the rate of 0.60 m a^{-1}. This is a crude calculation and can only be used to give an indication that the ice rise may not be in a steady state.

The "balance" accumulation rate can be found by considering the vertically integrated equation of continuity (Mellor 1968). This can be written as

$$a_m - \frac{\partial h}{\partial t} = h(\dot{\varepsilon}_x + \dot{\varepsilon}_y) + \bar{u}_x \frac{\partial h}{\partial x} = a_b, \qquad (2)$$

where h is the ice thickness (assumed resting on a horizontal bed), x and y are horizontal axes in the direction of ice flow and perpendicular to the direction of ice flow, respectively, $\bar{u}_x$ is the average horizontal velocity, and a_b is the "balance" accumulation rate, when $\partial h/\partial t = 0$. Equation (2) can be applied to the western axis of Butler Island, using data quoted by Martin and Sanderson (1980). Direct measurements of velocities, and hence strain-rates, were made in the x direction only. In order to estimate $\dot{\varepsilon}_y$, we can use the parameter ξ used by Martin and Sanderson to correct their surface-level profiles for the fact that the ice rises were elliptical in plan; for two-dimensional ice rises $\xi = 1$ and $\varepsilon_y = 0$, while, for a circular plan, $\xi = 2$ and $\dot{\varepsilon}_y = \dot{\varepsilon}_x$. Martin and Sanderson took $\xi = 1.5$, implying in our case that $\dot{\varepsilon}_y = 0.5\dot{\varepsilon}_x$. Table I shows how the value of a_b varies along the profile. Although there are large fluctuations, especially near the summit, which may be caused by difficulties in finding a stationary point, the average value for a_b of 0.65 m a^{-1} is nearly three times that of the measured accumulation rate.

It is difficult to estimate errors in the values of a_b. Although Equation (2) should be applicable over the whole ice rise, it is uncertain how $\bar{u}_x$ is related to the measured surface velocity. Near the centre of the ice rise, longitudinal strain-rates dominate shear strain-rates, while over the greater part of the ice rise shear strain-rates predominate (Martin and Sanderson 1980). In the calculations, $\bar{u}_x$ has been taken as 0.8 times the surface velocity; this factor is unlikely to change significantly and in any case the term containing $\bar{u}_x$ is always smaller than the strain-rate term (Table I). An error as large as 20% in $\bar{u}_x$ will give a maximum error of 7% in a_b; the standard deviation of 0.19 m a^{-1} is a realistic error figure. We can confidently assume that the measured accumulation rate is considerably less than the "balance" accumulation rate. Taking a value of a_b as 0.65±0.19 m a^{-1} and using Equation (1) gives a value of B'' of (1.05±0.11) x 10^8 N m^{-2} $s^{1/3}$.

A strain network was set out on the summit of Gipps Ice Rise in 1974. When it was remeasured in 1975, only one leg of the network remained, aligned along the minor axis. We can use this measurement in Equation (2) to give a lower bound for the balance accumulation a_b. Near the summit of an ice rise (or an ice divide), the term $\bar{u}_x \frac{\partial h}{\partial x}$ tends to zero.

Taking values of $\dot{\varepsilon} = 1.87 \times 10^{-3}$ a^{-1} and h = 320 m gives a_b = 0.6 m a^{-1} compared with the measured value of 0.38 m a^{-1} (Martin and Sanderson 1980). Using the value for the balance accumulation, we can calculate, with the aid of Equation (1), that the value of B'' quoted by Martin and Sanderson for the Gipps Ice Rise is reduced from (1.35±0.24) x 10^8 N m^{-2} $s^{1/3}$ to 1.16 x 10^8 N m^{-2} $s^{1/3}$. By taking a value of ξ of 1.5, by analogy with the situation on Butler Island, a_b is increased to 0.9 m a^{-1} and B'' reduced to 1.01 x 10^8 N m^{-2} $s^{1/3}$, giving very good agreement

TABLE I. BALANCE ACCUMULATION RATE a_b ALONG WESTERN AXIS OF BUTLER ISLAND

leg	ice thickness (h) (m)	longitudinal strain rate ($\dot{\varepsilon}_x$) (a^{-1}) x 10^{-3}	longitudinal velocity (u_x) (m a^{-1})	surface slope ($\frac{\partial h}{\partial x}$) (radians)	(ξ = 1.5) $\xi h \dot{\varepsilon}_x$ m a^{-1}	$\bar{u}_x \frac{\partial h}{\partial x}$ m a^{-1}	a_b m a^{-1}
A-B	273	2.58	.77	0.018	1.05	.01	1.04
B-C	266	1.45	1.21	0.030	.59	.03	.56
C-D	257	2.06	1.83	0.032	.80	.05	.75
D-E	245	1.44	2.26	0.038	.53	.07	.46
E-F	234	1.86	2.81	0.042	.66	.09	.57
F-G	220	1.86	3.37	0.048	.62	.13	.49
G-H	205	2.48	4.11	0.052	.77	.17	.60
H-I	188	3.40	5.13	0.064	.96	.26	.70
						av:	.65
							±.19

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TABLE II. ACCUMULATION RATES AND THE CORRESPONDING VALUES OF B''

	Accumulation rate (m a^{-1})	Equivalent value of B'' (x 10^8 N m^{-2} $s^{1/3}$)
Butler Island		
measured	0.23	1.49 ± 0.26*
average using Equation (2)	0.65 ±0.19	1.05 ± 0.11
to agree with B'	0.83	0.97
Gipps Ice Rise		
measured	0.38	1.35 ± 0.24*
using Equation (2)	0.90	1.01

* Martin and Sanderson 1980

with the Butler Island values for B' and B'' calculated by assuming a_b to be 0.65 m a^{-1} (see Table II). Again, it is difficult to estimate errors but we can use the results to indicate that the assumption of non-steady state behaviour is consistent with the available data.

Dolleman Island, an ice rise between Butler Island and Gipps Ice Rise, had a three-leg strain rosette set out across its summit in 1975-76 which was remeasured in 1976-77. From these measurements, principal horizontal strain-rates $\dot{\varepsilon}_1$ and $\dot{\varepsilon}_2$ were calculated as (1.50±0.15) x 10^{-3} a^{-1} and (0.66±0.06) x 10^{-3} a^{-1}, respectively. Writing Equation (2) in the form

$$a_b = h(\dot{\varepsilon}_1 + \dot{\varepsilon}_2) = a_m - \frac{\partial h}{\partial t}$$

gives, for h = 462 m, a value for a_b of 1.00±0.1 m a^{-1}, compared with the measured value for a_m of 0.41 m a^{-1} (Martin and Sanderson 1980).

A comparison between measured and balance accumulation rates for the three ice rises is given in Table III. It can be seen that there is a remarkable uniformity between them, with a surface lowering of about 0.5 m a^{-1} and the balance accumulation rate being between two and three times the measured accumulation rate.

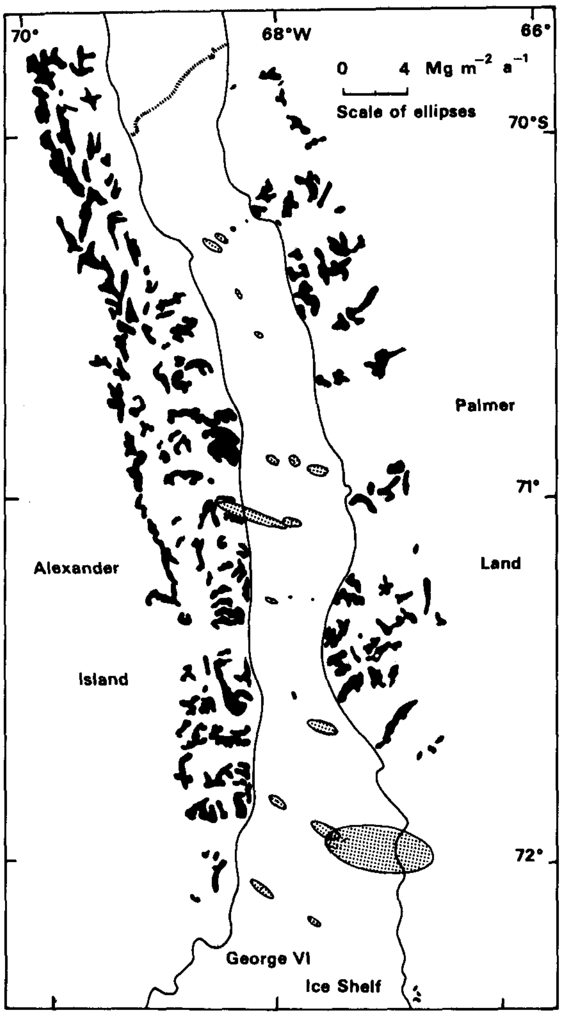

Fig.2. Bottom melting rates beneath George VI Ice Shelf calculated by assuming the ice shelf to be in a steady state. (Major axis of ellipse gives melt rate, minor axis the error.)

3. SURFACE-LEVEL LINES

In January 1975, a 2 km line across a snow-field in Palmer Land (70°35'S, 64°16'W) was optically levelled between two bench marks sited on rock (Bishop unpublished) in order to study glacier fluctuations. Swithinbank (personal communication) remeasured the line in February 1980 with a misclosure of 8 mm. No difference was noticed between the two profiles apart from the 200 to 300 mm random scatter which can be expected from transient sastrugi. This result implies that there has been no change in surface level over that time period and that the ice sheet in central Palmer Land is approximately in a steady-state condition.

TABLE III. RATES OF ICE THINNING

Ice rise	Measured accumulation rate (m a^{-1}) (a_m)	Balance accumulation rate (m a^{-1}) (a_b)	Rate of surface lowering (m a^{-1}) $\frac{\partial h}{\partial t}$	Ratio $\frac{a_b}{a_m}$
Butler Island	0.23	0.65	0.42	2.8
Dolleman Island	0.41	1.00	0.59	2.4
Gipps Ice Rise	0.38	0.90	0.52	2.4

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4. GEORGE VI ICE SHELF

Bishop and Walton (1981) have analysed the kinematic behaviour of George VI Ice Shelf and calculated the bottom melting rate m_b at many different points (Fig.2), using Equation (2) in the form

$$\left(\frac{\partial h}{\partial t} + m\right) = a_m - h\,\dot{\varepsilon}_z - u_x\,\frac{\partial h}{\partial x} = m_b, \quad (3)$$

where m is the actual bottom melt rate (positive for melting), m_b the "balance" bottom melt rate assuming $\partial h/\partial t$ to be zero, and $\dot{\varepsilon}_z$ the vertical strain-rate (positive extending). The pattern of melt rates m_b in Figure 2 has no simple relationship with parameters such as the time during which the ice has been afloat, or with distance from an ice-shelf boundary. It has been shown (Doake 1976) that the amount of heat available from the sea is the major factor in determining melting rates; the amount of heat conducted through the ice away from the ice-sea boundary is relatively small compared with possible fluctuations in heat supply from the sea (Bishop and Walton 1981). Because it is not known how various parameters such as current velocity, temperatures, salinity, etc., influence melting rates, and because there is no information about these parameters beneath George VI Ice Shelf (Lennon and others 1982), it is impossible to form an estimate for m. However, there is no reason to suppose that oceanographic conditions are sufficiently variable to account for the apparent melt-rate pattern: currents at the ice front are generally low, and temperature-salinity profiles with depth are very uniform across the northern ice front (Lennon and others 1982). This suggests that parts, at least, of the ice shelf may not be in a steady state. But, because of our lack of data, it is impossible to separate unambiguously the m and $\partial h/\partial t$ contributions to m_b. One possibility is that there is a uniform melt rate m of the order 1 m a^{-1} for the whole ice shelf, with parts of it thickening fairly rapidly. It is unlikely, though, that the ice shelf is thinning appreciably anywhere. For this to be the case, actual melt rates would have to be greater than the calculated "balance" melt rates, which are already high compared with melt rates calculated for other ice shelves (Thomas and Coslett 1970). The rather tenuous conclusion is that, unless there are diverse oceanographic environments beneath George VI Ice Shelf, parts of the ice shelf may be thickening at rates of up to several m a^{-1}.

5. RETREAT OF WORDIE ICE SHELF

Landsat satellite pictures of the Wordie Ice Shelf taken in 1974 and 1979 show the loss of 250 km^2 of ice shelf in 5 a with the ice front retreating by up to 7 km (Fig.3). Because of the crevassed nature of the grounding line and the relatively poor quality of the earlier picture, it is not possible to infer any change in position of the grounding line. Results from airborne radio echo-sounding of Wordie Ice Shelf in 1967 (Smith 1972) show that in the south-west corner the ice front extended 25 km further from the land than it did in 1974, and that, over a large area of the ice shelf which has subsequently disappeared, there was a shallow echo, interpreted as being caused by brine infiltration. Much of the ice shelf is severely crevassed, and Smith drew a distinction between the area east of about 67°50'W and the area to the west of that meridian. The eastern part is crevassed, while the western part contains rifts which extend to the bottom of the ice shelf, and which allow brine to percolate into the ice at sea-level. There has been a continual, though spasmodic, retreat of the ice front for at least 13 a. Recession appears to be halted temporarily when small ice rises "pin" the ice front. This behaviour conforms to experience and theory (Swithinbank 1957, Sanderson 1979). However, we cannot say whether there has been any similar change in grounding-line position. Without knowing this, it is impossible to decide whether the retreat has been caused by a reduction in mass flow from surrounding glaciers or by an increase in melt rate and net ablation of the ice shelf.

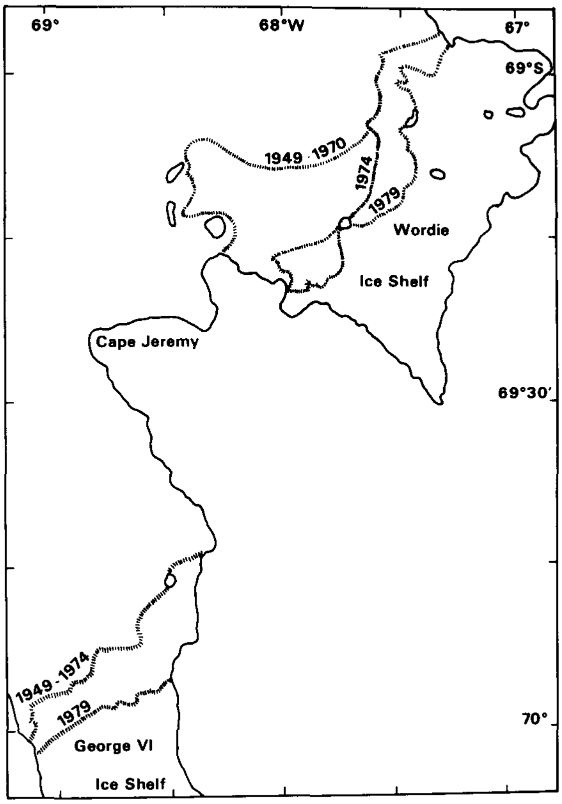

Fig.3. Recent recession of the ice fronts of George VI and Wordie ice shelves.

6. OTHER OBSERVATIONS

Ice fronts of George VI and Larsen ice shelves have receded in the last 30 a. Satellite pictures show a small recession at the northern ice front of George VI Ice Shelf between 1974 and 1979 (Fig.3). Reports from the British Graham Land Expedition 1934-37 suggest that in 1936 George VI Sound, north of the present ice front and as far as Cape Jeremy, was filled with rifted ice shelf floored by sea ice (Fleming and others 1938). Subsequent break-out of this rifted zone has been interpreted as a large (45 km) retreat of the ice front (Mercer 1968). It appears that disintegration of an ice front can proceed in several stages, from the initial formation of large crevasses, followed by rifts completely through the ice shelf but with fast ice preventing the escape of icebergs, to the final discharge of icebergs from the vicinity of the ice front. It is not always obvious whether a large area of floating ice is still joined to the main ice shelf throughout most of its depth or whether it has broken off but is prevented from drifting away. For example, Swithinbank (personal communication), when radio echo-sounding the Wordie Ice Shelf in 1967, described in his logbook the western rifted area (which had broken away by 1974) as "snowed-under icebergs". In an enclosed channel like George VI Sound there may be a long time-lag between the rifting and break-up of an ice front and its final disappearance. The practical difficulty of defining and locating the position of an ice front in a badly crevassed or rifted area should be borne in mind when considering evidence of retreat of ice fronts.

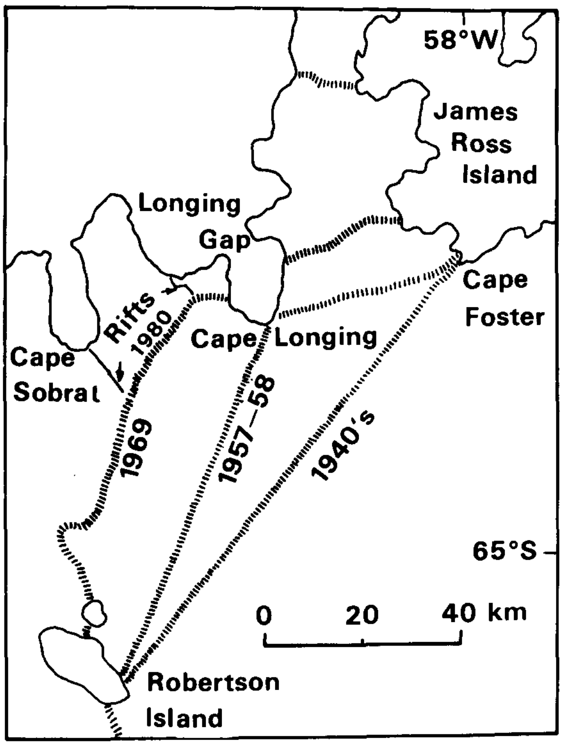

Fig.4. Recession of Larsen Ice Shelf since the late 1940s.

There is less ambiguity about recession at the northern end of Larsen Ice Shelf (Fig.4). Until the late 1940s, the ice front stretched from Cape Foster on James Ross Island to Robertson Island (Koerner 1964). Sometime between 1957 and 1958 it broke back to Cape Longing, and by 1969 had retreated another 15 km in some places. Evidence that retreat is continuing comes from BAS geological field parties over the last two years: in 1979-80 there was an unbroken surface between Cape Sobral and Longing Gap, while in 1980-81, the same geologist travelling the same route was halted by large rifts extending across the ice shelf to the ice front (Fig.4) (G Farquharson personal communication). It appears that recession at rates of up to 1 km a^{-1} is still continuing.

Mass- and energy-balance measurements undertaken between 1969 and 1974 on Spartan Glacier, a small valley glacier (6.3 km^2) draining into the western side of George VI Ice Shelf, show that the surface is lowering at an average rate of 0.27 m a^{-1} (Wager and Jamieson in press). Studies on a similar glacier, Hodges Glacier in South Georgia, show a pattern of advance and retreat of the snout over the last 100 a, but with a continual mass wastage for about the past 40 a (R Timmis personal communication).

7. DISCUSSION

Results from ice rises on the east coast of the Antarctic Peninsula suggest that thinning rates there may be of the order of 0.5 m a^{-1}. The central plateau of the peninsula shows no change in ice thickness over the last five years, while parts of George VI Ice Shelf on the west side of the peninsula may be thickening by several m a^{-1}. However, this rather confused picture shows certain similarities with results from the rest of Antarctica and from Greenland. Zwally and others (1981) have collected the few available estimates of rates of change of ice-sheet elevations and plotted them against their corresponding positions on a typical ice-surface profile. Near an ice divide or the summit of an ice sheet, thickness changes are low and negative (thinning). Rates of change of ice thickness increase towards the margin and become positive (thickening). There are too few results to know whether or not this is a general pattern for the whole of the Antarctic ice sheet. Because of the very different time scales associated with changes in average accumulation rates, and with the response of an ice sheet to these changes, it is possible that large areas of an ice sheet are never completely in an equilibrium state with their external environment. The smaller size of glaciers and ice shelves in the Antarctic Peninsula compared with the rest of Antarctica implies that their response times should be shorter. Therefore the effect of long-term changes in climate, which may ultimately affect the whole of Antarctica, should be seen first in the peninsula.

The apparent sensitivity of ice shelves to the climatic limit represented by the 0°C January isotherm (Mercer 1968) coupled with the 1.5°C rise in mean annual temperature for the peninsula over the last 30 a (Limbert 1980) could explain the observed contemporary disintegration of ice shelves around the peninsula. Although we might expect higher precipitation with increased temperatures because of the rise in water-vapour pressure, this does not appear to have happened on the east coast of the peninsula. This could be because other factors play a more important role in determining net accumulation rates, or it may be because of a shift in position of climatic regimes (Martin and Peel 1978). In any event, the break-up of an ice shelf may not necessarily be caused by a reduction in the ice flux which feeds it from the land. Theoretical models of ice shelves (Sanderson 1979) show that their equilibrium profiles are sensitive to melt rate, to the geometry of the side-wall restraint, and to pinning by ice rises and locally grounded areas. For example, a small increase in melt rate for an ice shelf in a bay with divergent sides could reduce the thickness of the ice shelf sufficiently to float once-grounded areas. Reduction of the up-stream restraining force would then allow the ice shelf to thin even more rapidly. A detailed knowledge of bay geometry and sea-bed topography is needed in order to calculate the response of an ice shelf to changes in melt rate. These changes can occur while conditions at the grounding line remain constant.

If there is a change in the net mass balance inland from an ice shelf, then these changes should ultimately affect the flow at the grounding line and cause it to migrate. The generally crevassed nature of surfaces near grounding lines make it difficult to locate the grounding line position accurately. Where sites are accessible, simple tiltmeters detecting the tital movement of an ice shelf near its grounding line can be used to find the position of the line to within a fraction of the ice thickness (Stephenson and others 1980). The results from George VI Ice Shelf, which suggest that the ice may be thickening in some areas, could be explained if the grounding line has advanced recently. Unfortunately we have no measurements which give information on past or present behaviour of the grounding line in this area.

Some semblance of order can be brought to the various, seemingly contradictory, values of thickening and thinning rates by considering the time scales which are appropriate to each measurement (Table IV). The form and flow of ice rises will respond to long-term averages of temperature and accumulation over periods of hundreds or thousands of years (Nye 1963). Accumulation rates deduced from oxygen isotope profiles show that the average accumulation for the last 10 to 15 a is smaller by a factor of 2 or 3 than that needed to keep the ice rises in their present form. The surface-level line shows no change over the five-year period 1975-80, while mass-balance measurements on Spartan Glacier only refer to the five-year measurement period 1969-74. All these results are fairly specific comparisons between recent accumu-

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TABLE IV. TIME SCALES FOR CHANGES IN ICE THICKNESS

Locality	Time scale (a)	$\frac{\partial h}{\partial t}$ (m a^{-1})	Principle of measurement
ice rise	10	-0.5	continuity of mass
level line	5	0	ice thickness
George VI Ice Shelf	~100's	+3	continuity of mass
ice fronts	~30(?)	(?)	continuity of mass
Spartan Glacier	5	-0.25	mass and energy balance

lation rates and the longer term average accumulation rates needed to maintain existing ice regimes. In contrast, the suggested ice thickening on George VI Ice Shelf is probably a delayed response to different climatic conditions several hundred years ago. This would be a typical time scale for changes in mass budget to affect the flow regime (Nye 1963) and to cause migration of the grounding line. George VI Ice Shelf is unusual in being so constrained that it experiences mainly compressive horizontal strain-rates. Over much of the ice shelf the vertical strain-rate term in Equation (3) plays a dominant role in determining melt rates (Bishop and Walton 1981). Thus changes in flow regime of glaciers feeding the ice shelf are likely to have a substantial effect on strain-rates. The great variability in "balance" melt rates could be explained if the large variety of glaciers flowing into George VI Ice Shelf each responded to changes in accumulation on different time scales (of the order of hundreds of years). It is more difficult to assign a time scale to the cause of ice-front retreats. Because of the sensitivity of ice shelves to their bay geometry and to sea-bed topography, as well as to their environment, a broad spectrum of response times may be real.

To summarize Table IV, ice thinning is associated with the recent climatic amelioration, while thickening may be happening in response to higher accumulation rates several hundred years ago.

CONCLUSION

Several experiments carried out in various ice-flow regimes in the Antarctic Peninsula using different techniques have yielded results which can be interpreted to give values for the rate of ice thinning or thickening. Ice rises on the east coast of the peninsula appear to be undernourished compared with the annual snow-fall they require to retain their present shape and flow rates. A level line between two nunataks shows that the ice surface has not changed over a period of 5 a. Melt rates under George VI Ice Shelf, calculated assuming steady state, are high and very variable. It is likely that actual melt rates are more uniform and smaller, with the ice shelf thickening in some places in response to a higher accumulation rate over the peninsula a few hundred years ago. Mass-balance measurements on a small valley glacier show that it is wasting away. Ice fronts have retreated for the last 30 a in some places. The general picture is one of ice-sheet decay in response to climatic amelioration over at least the last 30 a, when mean temperatures of both the air and presumably the sea have increased while net accumulation has decreased. However it appears that there was a higher snow-fall before then and some parts of the Antarctic Peninsula are still thickening as a response to this.

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