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The Response of Large Ice Sheets to Climatic Change [and Discussion]

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The response of large ice sheets to climatic change

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SUMMARY

The prediction of short-term (100 year) changes in the mass balance of ice sheets and longer-term (1000 years) variations in their ice volumes is important for a range of climatic and environmental models.

The Antarctic ice sheet contains between 24 M km^3 and 29 M km^3 of ice, equivalent to a eustatic sea level change of between 60 m and 72 m. The annual surface accumulation is estimated to be of the order of 2200 Gtonnes, equivalent to a sea level change of 6 mm a^{-1}. Analysis of the present-day accumulation régime of Antarctica indicates that about 25% (*ca.* 500 Gt a^{-1}) of snowfall occurs in the Antarctic Peninsula region with an area of only 6.8% of the continent. To date most models have focused upon solving predictive algorithms for the climate-sensitivity of the ice sheet, and assume: (i) surface mass balance is equivalent to accumulation (i.e. no melting, evaporation or deflation); (ii) percentage change in accumulation is proportional to change in saturation mixing ratio above the surface inversion layer; and (iii) there is a linear relation between mean annual surface air temperature and saturation mixing ratio. For the Antarctic Peninsula with mountainous terrain containing ice caps, outlet glaciers, valley glaciers and ice shelves, where there can be significant ablation at low levels and distinct climatic régimes, models of the climate response must be more complex. In addition, owing to the high accumulation and flow rates, even short- to medium-term predictions must take account of ice dynamics. Relationships are derived for the mass balance sensitivity and, using a model developed by Hindmarsh, the transient effects of ice dynamics are estimated. It is suggested that for a 2°C rise in mean annual surface temperature over 40 years, ablation in the Antarctic Peninsula region would contribute at least 1.0 mm to sea level rise, offsetting the fall of 0.5 mm contributed by increased accumulation.

1. INTRODUCTION

Small imbalances in the mass budgets of the large ice sheets in Antarctica and Greenland have widely recognized implications. A one-sixth variation in the present annual mass gain of Antarctica, for instance, is equivalent to a 1 mm a^{-1} sea level change. The response of the Antarctic and Greenland ice sheets to climate forcing (principally temperature, and precipitation as derivative) has been addressed in several recent studies (Huybrechts & Oerlemans 1990; Huybrechts *et al.* 1991; Drewry 1991; Giovinetto *et al.* 1990). Most of these investigate the climate sensitivity of the mass balance of the ice sheet using relatively simple models of surface mass balance at a point. The ice sheet mass balance is the sum of the surface mass balance over the upper surface of the ice sheet, losses at the lower boundary, and discharge at the margin. Because the margin can be chosen as either the grounding line of the ice sheet or the calving front of the ice shelves it is important to establish a clear definition of terms to aid interpretation of historical studies of mass balance. The surface mass balance at a point is the difference between the accumulation rate from snowfall, ice crystal deposition and riming, and the ablation rate from melting, evaporation, sublimation and deflation.

Figure 1 shows a simple model of how surface mass balance varies with temperature. Warrick *et al.* (1990) find that small glaciers and the Greenland Ice Sheet lie mostly within domain 'b' whereas the Antarctic Ice Sheet with a much colder climate lies in domain 'a'. Huybrechts & Oerlemans (1990) used this model in a sensitivity study of the mass balance of the Antarctic ice sheet and concluded that with a warming of up to about 5°C there is a steady increase in accumulation resulting in an maximum mass of 640 Gt a^{-1} added to the ice sheet. This is equivalent to a sea level fall of (−)1.8 mm a^{-1}.

Most authors calculate a 'static' response, i.e. the ice sheet mass balance and sea level signals for a given climate scenario with no compensating changes in ice dynamics. Typically they consider that ice flow is relevant only on time scales of 10^2–10^3 years (e.g. Young 1981). Huybrechts & Oerlemans (1990), however, calculate a dynamic response and find that it becomes important at the margins of the ice sheet after 100 years or more when applying IPCC temperature scenarios.

(a) *Prediction of surface mass balance*

The algorithms for surface mass balance used in these models relate the accumulation rate (usually

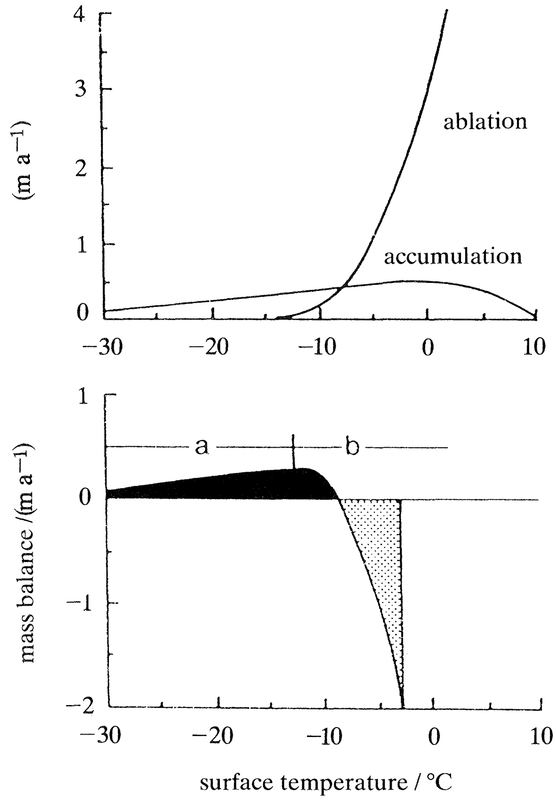

Figure 1. Schematic relationship between surface mass balance and mean annual surface temperature (from Warrick & Oerlemans, 1990). The upper diagram depicts ablation and accumulation and the lower diagram the surface mass balance as a function of temperature.

presumed equivalent to precipitation rate) to surface air temperature. Robin (1977) has suggested that precipitation over the Antarctic continent is controlled by advection of moisture in the troposphere above the surface inversion layer. Hence the rate of snowfall is proportional to the water available in the atmosphere (the saturation mixing ratio). This ratio is a function of the temperature above the inversion layer which, it is supposed, is related to the surface air temperature. Huybrechts & Oerlemans (1990) have used a simple linear relationship which Fortuin & Oerlemans (1990) found applicable over the interior region of the ice sheet. The Robin or 'thermodynamic' method is supported by deep ice-core data linking isotopically derived temperatures and accumulation rates through the last glacial cycle (e.g. Jouzel & Merlivat 1984; Lorius 1989).

Most models for Antarctica have ignored ablation. Huybrechts & Oerlemans (1990) adopted a degree-day technique for the Antarctic ice sheet margin and found only low surface run-off ($\leqslant 8\%$ of the accumulation over the whole ice sheet) for temperature increases of 5°C or less.

2. PRESENT DAY ICE SHEET MASS BALANCE

The interpretation of sensitivity or full modelling studies of the climate response of ice sheets requires knowledge of not only the present day mass balance but also the probable pattern of change such as the identification of important regional differences or the identification of distinctive and rate-controlled processes.

Table 1 is modified from Jacobs *et al.* (1992) which is the most up-to-date compilation of estimates of the Antarctic ice sheet mass balance. At present loss of mass has been less accurately estimated than input, giving rise to low levels of confidence for net balances. Jacobs *et al.* (1992) have ascribed accuracies of $\pm 20\%$ to total accumulation, $\pm 33\%$ to iceberg calving and

Table 1. *Antarctic mass balance estimates (Gt(10^{12}kg) a^{-1}) (modified from Jacobs et al. (1992))*

accumulation			loss					
grounded ice	ice shelf	total	calving icebergs	ice shelf melting	runoff	balance	date	reference
		1850	−1700	−550		−400	1963	Losev[a]
1490	356	1885	−1053	−293			1971	Barkov[a]
		2080	−1450	−200	−10	+420	1971	Bull[a]
		2000	−2400	−320	−60	−720	1978	Kotlyakov[a]
		1749	−855	−251		+487	1983	Meier[a]
		2000	−1800			+200	1985	Budd and Smith[a]
			−2300				1985	Orheim[a]
1468	495	1963					1985	Giovinetto & Bentley[a]
2158	475	2633					1986	Radok *et al.*
		2141					1989	Giovinetto & Bentley[a]
		2200	−2200			0	1990	Warrick & Oerlemans
		2406			~0		1990	Huybrechts & Oerlemans
1817	287	2104					1990	Fortuin & Oerlemans
						+40–400	1991	Bentley & Giovinetto
1528	616[b]	2144	−2016	−543	−53	−468	1992	Jacobs *et al.*

[a] References to these papers are given in Jacobs *et al.* (1992).
[b] The significantly higher ice shelf values are based upon a refinement of, and data additions to, the analysis of Giovinetto & Bentley (1985, 1989).

$\pm 50\%$ to ice shelf melting and surface runoff. It is unlikely that grounding line flux is better known.

The total accumulation is determined by integration of measured point values over the upper surface of the ice sheet. However, there are considerable problems with the quality, spatial coverage and time-representativeness of these measurements (Bull 1971; Giovinetto & Bentley 1985; Drewry 1991).

The mass balance of the grounded ice sheet has been estimated as the difference between accumulation (over the grounded area alone) and the ice flux at the grounding line (Bentley & Giovinetto 1991). This method can only be used for prediction of mass balance change if there is no feedback from processes on and beneath ice shelves to the response of the ice sheet to climate forcing. An alternative approach is to calculate the mass balance of the entire ice sheet, including ice shelves. It is then necessary to estimate output from iceberg discharge (Orheim 1985) and melting beneath ice shelves. Distinguishing the separate role of ice shelves allowed Jacobs *et al.* (1992) to offer alternative interpretations of the overall negative mass balance of the ice sheet (-468 Gt a^{-1}; table 1) as: (i) not significantly different from zero; (ii) the result of thinning of ice shelves or the slow retreat of calving fronts; or (iii) the result of excess flow across the grounding line (to maintain equilibrium of the ice shelves) resulting in a reduction of the grounded ice sheet volume. It is worth noting that the mass balance of the grounded ice sheet may be affected by rapid thinning of ice shelves because of changes in oceanographic conditions. This thinning may reduce the restraining back-stress on ice flowing from inland and lead to grounding line retreat (Jenkins & Doake 1991).

3. CLIMATIC ZONATION OF THE ICE SHEET

Most previous mass balance studies have assumed that the Antarctic Ice Sheet can be modelled as a single unit. This is a reasonable first approach, but on inspection of the geographical pattern of accumulation and of the factors influencing both precipitation and ice discharge significant regional variations become evident. Figure 2 shows the rate of accumulation over the whole continental area. Three principal zones are apparent and are summarized in table 2: (i) low accumulation over the main body of the ice sheet; (ii) higher accumulation around the maritime margin of the continent; and (iii) the distinctive contribution of the Antarctic Peninsula region.

The first two of these elements have been recognized by Fortuin & Oerlemans (1990) using a different criterion. In a statistical analysis based upon linear multiple regression of annual surface temperature and surface mass balance they identified three topographic regions: ice shelves (0–200 m ASL); an escarpment region (200–1500 m ASL, where there is high natural climate variability and which included most of the Antarctic Peninsula); and an interior region (more than 1500 m ASL). Only in the last region could there be reliable parameterization of surface mass balance with temperature, a result confirmed by Giovinetto *et al.* (1990).

It is apparent from table 2 and figure 2 that in terms of surface mass balance alone the Antarctic Peninsula plays a distinct, substantial, and to date, little investigated role in the overall response of the ice sheet. With its mountainous terrain, strongly differentiated climate regimes (with significant ablation and run-off at low levels) and mix of ice features (ice caps, outlet glaciers, tidewater glaciers, and ice shelves) modelling is more complex than for the interior ice sheet case.

4. RESPONSE OF ICE MASSES IN THE ANTARCTIC PENINSULA

(*a*) *The surface mass balance of the Antarctic Peninsula*

Doake (1985) has drawn attention to the contribution of the Antarctic Peninsula and its related drainage basins. The region comprises a mountainous, S-shaped belt stretching over 1500 km northwards from Ellsworth Land (latitude 75°S), with peaks rising to over 2500 m ASL. The Peninsula exhibits the warmest climate in Antarctica and acts as a major physical barrier to tropospheric circulation. Reynolds (1981) and Doake (1985) describe two distinct climate regimes on the east and west sides of the Peninsula. On the east side, dominated by the presence of extensive and persistent sea ice in the Weddell Sea, temperatures are 7°C lower than at the same latitude on the west side. Ice shelves extend 350 km further north on the eastern flank of the Peninsula, where almost all outlet glaciers feed into the Larsen Ice Shelf.

A large area of the Antarctic Peninsula has a mean annual surface temperature below -11°C. Peel (1992) has found that the relation between accumulation rate and temperature determined from ice cores taken in this area is consistent with the Robin (1977) relationship. He suggests that accumulation can be predicted from mean annual surface temperature with an accuracy of 17% for the central regions, 26% for east coast regions and 31% for Alexander Island. However, using the sea-level isotherms and lapse rates determined by Reynolds (1981), an area of some 20000 km^2 (2%) of the Antarctic Peninsula may be estimated to have a mean annual surface temperature warmer than -11°C. In this region the surface mass balance is the difference between accumulation and a non-zero ablation rate which can also be parameterized in terms of mean annual temperature. It is instructive to make a simple order of magnitude calculation of the contribution to sea-level rise which might be expected from this ablation. For warm glaciers it has been suggested that ablation increases with mean annual surface temperature by some 0.5 m a^{-1} K^{-1} (IPCC 1990). (Note that in this paper components of the surface mass balance are given in metres of ice per year.) This value may be appropriate for the northernmost parts of the Peninsula where

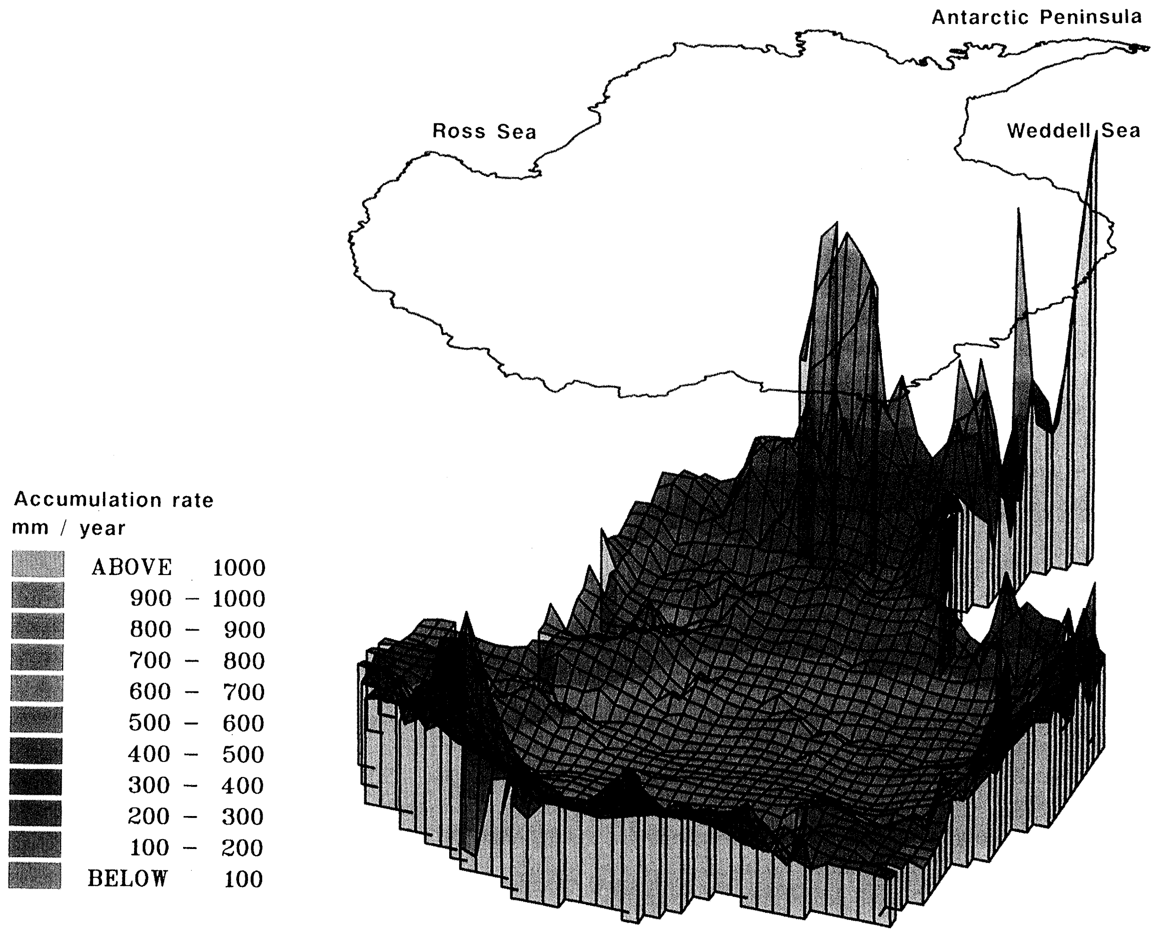

Figure 2. Pattern of accumulation rate over Antarctica. This is based upon a dataset compiled by one of the authors (D.J.D.) and originally intended as a separate sheet in *Antarctica: glaciological and geophysical folio* (ed. D. J. Drewry), Scott Polar Research Institute, Cambridge (1983)), updated with more recent measurements of accumulation especially in the Antarctic Peninsula and over the Ronne-Filchner Ice Shelf. Note the very low rate in the central regions, the increase towards the margins of the ice sheet and the considerable accumulation rates in the Antarctic Peninsula region (courtesy D. Vaughan).

mean annual temperatures are as high as -3°C but must be too large for the colder areas. A conservative value of $0.25\ \mathrm{m\ a^{-1}\ K^{-1}}$ applied to the whole area warmer than -11°C gives a total ablation of 4.6 Gt $\mathrm{a^{-1}\ K^{-1}}$ or 0.012 mm $\mathrm{a^{-1}\ K^{-1}}$ of sea level rise.

(*b*) *Dynamic response of the Antarctic Peninsula ice sheet*

Sea level rise from ice melt, dD, is related to the change in the volume of grounded ice, dV, by the equation

Table 2. *Regional contributions to Antarctic total accumulation*

	area (%)[a]		accumulation (Gt $\mathrm{a^{-1}}$)		accumulation (%)	
region	this paper	F & O[f]	this paper[b]	F & O[c]	this paper	F & O
interior ice sheet	65	62	858	880	40	42
ice sheet margins						
1. excl. ice shelves, excl. Peninsula	16	—	170	—	8	—
2. excl. ice shelves, incl. Peninsula	—	26	—	937	—	44
3. ice shelves	12	12	616[d]	287	29	14
Antarctic Peninsula	7[e]	—	500[e]	—	23	—

[a] Total conterminus area of Antarctica taken as 13 918 000 km^2 (including 2.4% ice-free terrain) (Drewry *et al.* 1982).
[b] Total accumulation taken as 2144 Gt $\mathrm{a^{-1}}$ (Jacobs *et al.* 1992).
[c] Total accumulation 2104 Gt $\mathrm{a^{-1}}$.
[d] From Jacobs *et al.* (1992).
[e] From Frolich (1992).
[f] From Fortuin & Oerlemans (1990).

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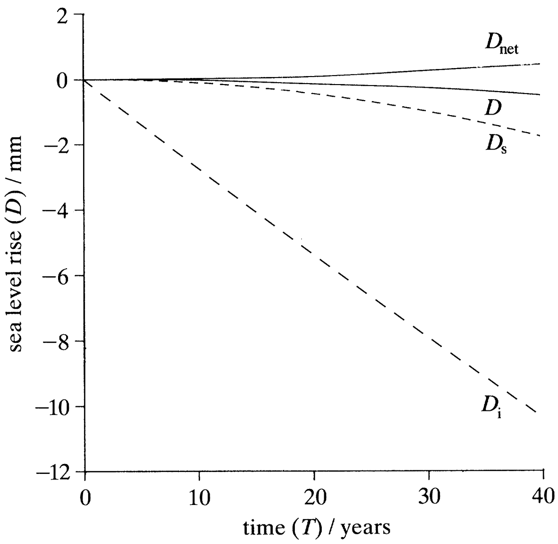

Figure 3. Sea level change for a 2°C warming of the Antarctic Peninsula over 40 years. (i) Initial equilibrium response D_i; (ii) transient response D; (iii) quasi-static response D_s; and (iv) net rise in sea level, D_{net}.

$$\mathrm{d}D = -(0.0025\ \mathrm{mm\ km^{-3}})\mathrm{d}V. \tag{1}$$

In areas of the Antarctic continent where the grounding line may be regarded as fixed and the thickness of the ice at this margin is very small compared to the average thickness at the ice divide, H, we may write

$$\frac{\mathrm{d}V}{V} = \frac{\mathrm{d}H}{H}. \tag{2}$$

Hence

$$\mathrm{d}D = -(0.0025\ \mathrm{mm\ km^{-3}})V\frac{\mathrm{d}H}{H}, \tag{3}$$

and the problem of predicting sea level rise reduces to the task of determining how average thickness at the ice divide responds to a specified climate change. We consider first the steady-state configuration of the ice sheet for different climate conditions and then the question of how the transition from one state to another might occur.

(i) *Equilibrium response*

Vialov (1958) has shown that for an ice sheet with spatially uniform accumulation rate, a, no ablation, a flat bed, and a fixed margin with zero thickness, the equilibrium value of the divide thickness is given by the equation

$$H_e = (2)^{\frac{n}{(2n+2)}}\left(\frac{a}{A}\right)^{\frac{1}{(2n+2)}}L^{\frac{1}{2}}, \tag{4}$$

where L is the distance from the divide to the margin, n is the index in Glen's flow law relating stress and strain rate tensors for ice and A is the vertically-averaged value of the rate factor. For $n = 3$,

$$\frac{\mathrm{d}H_e}{H_e} = \left(\frac{1}{8}\right)\left(\frac{\mathrm{d}a}{a} - \frac{\mathrm{d}A}{A}\right). \tag{5}$$

Hindmarsh (1992) has analysed a more complex model which is representative of conditions in the Antarctic Peninsula. He postulates a central ice reservoir along the spine of the Peninsula which feeds tributary glaciers terminating at calving fronts along the coast. The accumulation rate and bed-slope can vary with distance from the divide and the ice thickness at the margin need not be zero. Hindmarsh concludes that the equilibrium thickness at the divide is still proportional to the one-eighth power of the accumulation rate with a similar weak dependence on the rate factor A. Where the bed slopes down from the central region to the margin, the steady-state thickness H_e becomes less sensitive to changes in the accumulation rate and rate factor in the margin.

Suppose that climate change over the ice sheet is defined in terms of the mean annual surface temperature, T_s. The change in sea level resulting from an increase in T_s can be calculated from equations (1) to (5) provided that changes in accumulation and rate-factor can be determined for a given change in T_s.

The relation between T_s and accumulation has been discussed above. We have

$$\frac{\mathrm{d}a}{a} = f(T_s)\mathrm{d}T_s, \tag{6}$$

where, for temperatures below about −11°C, $f(T_s)$ is approximately constant ($f(T_s) = 0.1\ \mathrm{K^{-1}}$ at $T_s = -20$°C). Ablation in marginal regions of the Peninsula is, for the moment, ignored.

For ice sheets such as those in the Antarctic Peninsula where advection is the dominant mode of heat transport, the mean annual surface temperature at equilibrium is related to the basal temperature T_b at the divide by the equation (Paterson 1981).

$$T_b = T_s + (0.2\ \mathrm{K}\ a^{-\frac{1}{2}})\left(\frac{H_e}{a}\right)^{\frac{1}{2}}. \tag{7}$$

(The value of the parameter in this equation is calculated using geothermal heat flux $5.6\ 10^{-2}\ \mathrm{Wm^{-2}}$, thermal conductivity $2.1\ \mathrm{Wm^{-2}}$ and thermal diffusivity $1.15\ 10^{-6}\ \mathrm{m^2\ s^{-1}}$).

Because the vertical averaging of the rate factor is strongly weighted towards the base an estimate of A can be obtained by using the basal temperature in the Arrhenius equation for the rate factor. Thus

$$A = A_0 \exp\left(-\frac{E}{kT_b}\right), \tag{8}$$

where k is Boltzmann's constant ($1.381\ 10^{-23}\ \mathrm{J\ K^{-1}}$), E the activation energy for creep (we choose to use a value of $1.089\ 10^{-19}$ J between −50°C and −10°C) and T_b is measured on the absolute scale.

$$\frac{\mathrm{d}A}{A} = (7884\ \mathrm{K^{-1}})\frac{\mathrm{d}T_b}{T_b^2}. \tag{9}$$

As an example, consider an increase, $\mathrm{d}T_s$, of 2°C in mean annual surface temperature over the Antarctic Peninsula plateau from a present day value of −20°C. From equation (6) the corresponding change in accumulation rate will be $\mathrm{d}a = 0.2\ a$. If the present day accumulation rate is $0.5\ \mathrm{m\ a^{-1}}$, $\mathrm{d}a = 0.1\ \mathrm{m\ a^{-1}}$. Assuming that the equilibrium thickness at the divide for present day conditions, H_c, is between 500 m and 1000 m, then equation (7) gives a present day basal temperature T_b between -13.7°C and -11.1°C. Equations (5), (6), (7) and (9) are solved simultaneously for dT_b, dA and dH_c as a function of dT_s. For this example, the future equilibrium basal temperature lies between -12.3°C and -10.0°C, $dT_b = (1.4\text{–}1.1)$°C, $dA/A = (0.16\text{–}0.13)$ and $dH_c/H_c = (0.005\text{–}0.0088)$. Using Doake's (1985) estimate for the volume of grounded ice on the Antarctic Peninsula, $V = 1.8\ 10^5\ \text{km}^3$ in equation (3), the equilibrium response of sea level to the 2°C temperature rise is estimated to be a fall of $(-)2.3$ to $(-)3.9$ mm.

It is clear from the above example that at surface temperatures around -20°C the effect of an increase in accumulation is in the long term partly offset by the effect of warmer temperatures on the viscosity of ice. On the one hand more snow is deposited on the ice sheet, on the other ice discharges more rapidly to the sea so that the total volume of grounded ice when it reaches equilibrium under the new conditions will not be much changed. However, the situation alters dramatically if surface temperatures in the region of -5°C are considered. The highly non-linear variation of the rate factor with temperature means that, for a given dT_s, dA/A increases strongly with T_s. In addition da/a will be very small or even negative as ablation processes become significant at the ice sheet surface. Thus dH_c/H_c will be negative and sea level will rise. If all the ice in the Peninsula were to melt the sea level rise would be some 540 mm, which is roughly equivalent to the effect of melting all the world's small glaciers.

(ii) *Transient response*

If changes in surface temperature proceeded sufficiently slowly for the ice sheet to be always close to equilibrium the steady state analysis in the last section would be sufficient. However, the predicted climatic effect of greenhouse gas emmissions over the next 100 years, and the climate change that has been observed over the last 40 years in the Antarctic region (whether greenhouse gas induced or not), are sufficiently rapid to make it neccessary to consider the transient response of the ice sheet.

The basic timescale for ice sheet response is H/a, where H is the ice thickness at the divide and a is the accumulation rate. A full discussion of how this timescale is obtained is given in Hindmarsh (1990). In the Peninsula a is high and H is relatively low, so the response time is likely to be more rapid than for the continental ice sheet as a whole.

The ice sheet responds to an increase in accumulation by increasing vertical velocity and decreasing the basal temperature (equation (7)). The increase in vertical velocity is immediate but the adjustment of the temperature profile takes some time. Paterson (1981, p. 202) shows a series of dimensionless steady-state temperature profiles for various values of an advection parameter ($Ha/36.3\ \text{m}^2\ \text{a}^{-1}$) which may be used to assess the timescale of this process. For $H = 500$ m, for example, an increase in accumulation from 0.5 m a^{-1} to 0.6 m a^{-1} increases the advection parameter from 6.9 to 8.3. The temperature profiles for these two values are separated by about 0.2 H. Assuming that the advection velocity is equal to a, the timescale for the decrease in basal temperature to be achieved is about 0.2 H/a that is, 200 years.

If the increase in accumulation is associated with an increase in surface temperature (as in equation (6)) after a lag time of the order of H/a the warmer temperature will be advected to the base of the ice sheet and the basal temperature will increase. Thus it will take some time before the rate factor will increase to offset the effect of increased accumulation. A full discussion of these processes is given in Whillans (1978).

We have not attempted a complete analysis of the transient response of the Antarctic Peninsula ice sheet. However, it is possible to present a first estimate of sea level response on the 10 to 100 year timescale by considering the response to accumulation changes without taking rate factor changes into account.

The transient response is examined for an initial period during which $dA \approx 0$. The equilibrium thickness at the divide for the instantaneous conditions at time $t \ll H/a$ is then defined as

$$H_i(t) = (2)^{\frac{n}{(2n+2)}} \left(\frac{a(t)}{A(0)}\right)^{\frac{1}{(2n+2)}} L^{\frac{1}{2}}, \tag{10}$$

where $A(0)$ is the value of the rate factor at time $t = 0$.

Hindmarsh (1992) has suggested that in the initial period the height of the divide may be expressed as

$$H(t) = H_i(t) + h(t), \tag{11}$$

where h is a small perturbation to H_i. He then derives an approximate solution for the response of h to changes in accumulation with the rate factor held constant

$$h(t) = -KH_i\left(1 - \exp\left(\frac{-t}{T}\right)\right),$$

where

$$K = \frac{H_i}{(2n+2)^2 a^2}\frac{da}{dt}, \tag{12}$$

and

$$T = \frac{H_i}{(2n+2)a}.$$

For $h/H_i \ll 1$ as required by the perturbation analysis we must have $K \ll 1$ which, for $n = 3$, $H_i = 500$ m to 1000 m and $a = (0.5\text{–}1.0)$ m a^{-1} implies $(da/dt)/a \ll (0.032\text{–}0.13)\ \text{a}^{-1}$. For temperatures below about -11°C this implies that temperature change must be much less than (0.32–1.3)°C a^{-1} for equation (12) to hold. The time constant T for the exponential relaxation towards H_i is in the range 65 to 250 years. Sea level approaches the equilibrium value D_i for conditions at time $t \ll H/a$ according to the equation

$$D(t) - D_i(t) = K'H_i\left(1 - \exp\left(\frac{-t}{T}\right)\right), \tag{13}$$

where $K' = (0.0025\ \text{mm km}^{-3})VK/H$ from equation (3).

Consider the example of a 2°C rise in temperature from a present day value of -20°C which takes place over a period sufficiently long for K to be less than 0.1. For $H_i = 500$ m and $a = 0.5$ m a^{-1} this means at least 30 years. Since 1947, there has been a general warming trend amounting to 2.2°C for the Antarctic Peninsula (Jones & Limbert 1987), an average increase of 0.05°C a^{-1}. We therefore choose 40 years as the period over which the 2°C rise occurs and note that this is very much less than $H/a = 1000$ years. Figure 3 shows D_i and D calculated using the appropriate time constant, $T = 125$ years. The axes have been chosen so that at time $t = 0$, $D = D_i = 0$. Also shown is the quasi-static response, D_s, i.e. the fall in sea level which would occur if there were no extra dynamic adjustment to the increasing accumulation and the height at the divide increased according to the equation

$$\frac{\mathrm{d}H_s}{\mathrm{d}t} = \left(\frac{\mathrm{d}a}{\mathrm{d}t}\right)t. \tag{14}$$

The time at which the initial equilibrium response is equal to the quasi-static response is

$$t^* = \left(\frac{H}{4a}\right), \tag{15}$$

which in this example is $t^* = 125$ years. After 40 years $D_i = -10.4$ mm, $D_s = -1.8$ mm and $D = -0.5$ mm. The (longterm) equilibrium response D_e to a 2°C rise was calculated to be -2.3 mm in the previous section.

Returning to the estimate of the contribution of ablation to sea level rise we may calculate that at a rate of 0.012 mm a^{-1} K^{-1} ablation would contribute $+1.0$ mm over 40 years in this scenario, offsetting the fall in sea level of -0.5 mm predicted by the perturbation analysis. This is a conservative estimate since it takes no account of the increase in the ablation area over the period. Figure 3 shows the net rise in sea level, D_{net}, over the 40 year period.

How far can the perturbation analysis be used to estimate the contribution of Antarctic ice to sea level rise over the next 100 years? The IPCC Business-as-usual scenario (IPCC 1990) predicts an increase in mean global surface temperature of 0.03°C a^{-1}. The temperature increase in the peninsula region itself may be greater but as yet GCM predictions (which in any case suffer from inadequate spatial resolution in this area) do not agree on this point. For the IPCC rate $K \approx 0.05$ and the perturbation analysis is valid. However, the scaling approach we have used is best regarded as a way of clarifying the time scales and relative importance of the various processes contributing to sea level rise. The quantitative estimates should be regarded with extreme caution. We can say with confidence that the general pattern of response to increased accumulation will be an increase in the volume of grounded ice moderated by dynamic adjustments to the ice sheet flow. As far as we can judge, given the caveat expressed above, ablation from the marginal regions will offset the effects of accumulation in the central regions of the Peninsula and the net effect is likely to be one of sea level rise.

5. CONCLUSIONS

The contribution of the Antarctic ice sheet to sea level rise cannot be estimated by treating the continent as a single unit. Because of variations in ice thickness and accumulation rate, there will be different dynamic responses to climate change. Furthermore, some marginal areas experience significant ablation. Different areas require different modelling strategies.

As an example, the response of the Antarctic Peninsula ice sheet, which experiences some 25% of present day annual accumulation over Antarctica and significant ablation, has been examined. A first estimate of sea level response on the 10 to 100 year timescale suggests that a 2°C rise in mean annual temperature over 40 years will produce a net rise in sea level. Further development of the Hindmarsh model for the Antarctic Peninsula ice sheet, to include the effects of changes in the rate factor, and an improved estimate for ablation, will allow this estimate to be refined.

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Discussion

G. de Q. Robin (*Scott Polar Research Institute, Cambridge, U.K.*). In spite of his reservations about the overall quantitative significance of his estimates of mass balance changes over the Antarctic Peninsula, could Dr Drewry relate his estimates to those over the whole of Antarctica and to global sea level for a predicted climatic change. How would the total change of the Antarctic mass balance then affect global sea level.

D. J. Drewry. The IPCC estimate for the contribution of the whole Antarctic continent to global sea level change is (-0.3 ± 0.3) mm a^{-1} °C^{-1}. That is, for a warming of 2°C, sea level would fall by (24 ± 24) mm over 40 years. The contribution from the Antarctic Peninsula, which has 6.8% of the surface area of the continent would be a fall of 1.6 mm. Our estimates, however, suggest a rise of 0.5 mm for this scenario.

G. Weller (*Geophysical Institute, University of Alaska, U.S.A.*). In light of the recent results from third generation (coupled, transient) global circulation models (GCMs) which predict very small temperature increases for Antarctica, how will the currently projected ice sheet balance changes and consequent effects on sea level have to be modified?

D. J. Drewry. The recent results show smaller temperature increases for Antarctica because warming of the ocean in the region is now taken into account. However, the treatment of sea-ice dynamics in GCMs is still in need of improvement and the influx of water from sub-ice shelf melting is not included at all. Thus the predictions of future temperature from GCMs must still be taken with caution. We have estimated the net sea level rise that would occur if the warming trend over the Peninsula seen over the last 40 years continues at the same rate for the next 40 years as 0.5 mm. Recent GCMs suggest a slower warming of 2°C over 100 years rather than 40 years. For this case there would be a net sea level fall of 7 mm.

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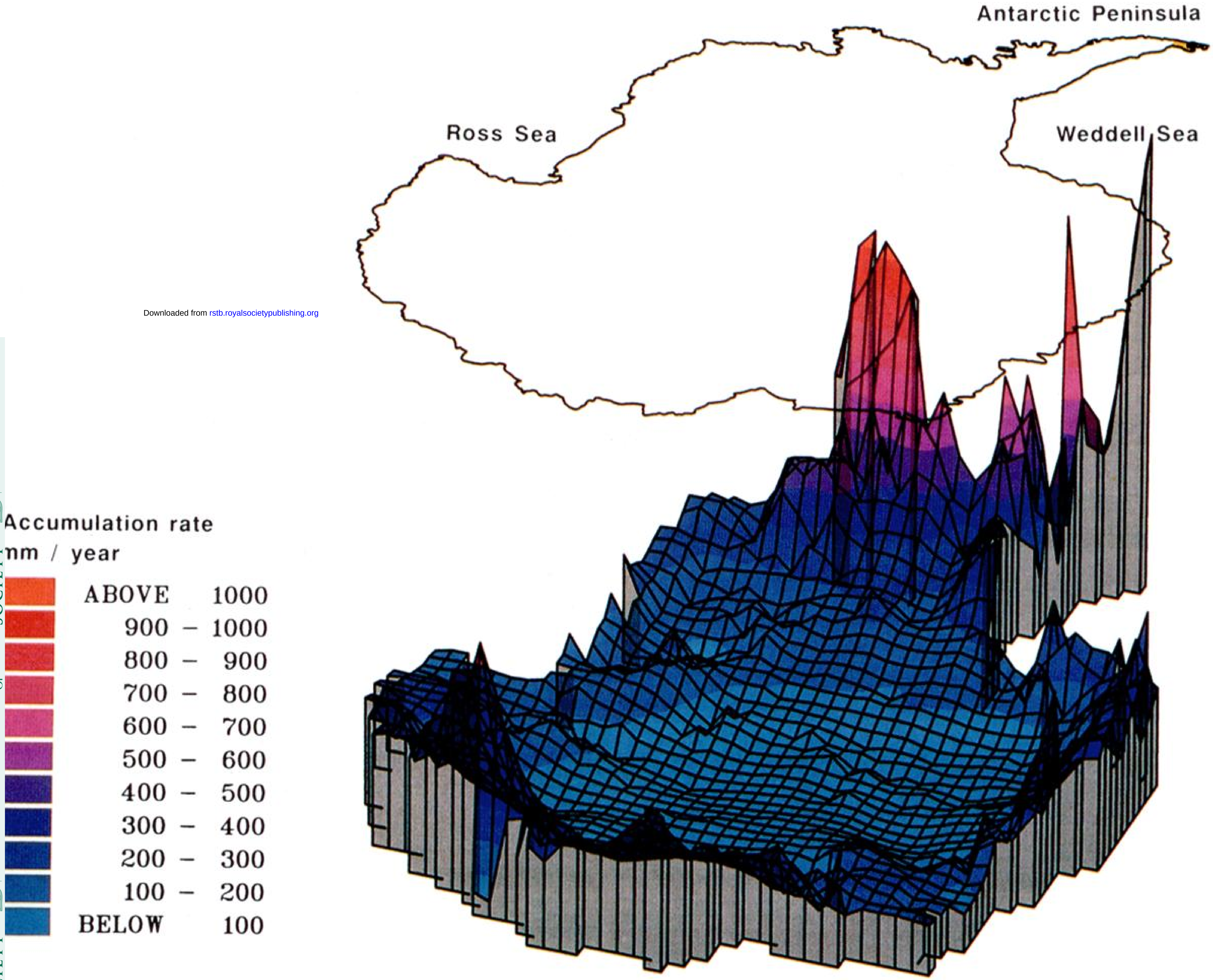

gure 2. Pattern of accumulation rate over Antarctica. This is based upon a dataset compiled by one of the authors).J.D.) and originally intended as a separate sheet in *Antarctica: glaciological and geophysical folio* (ed. D. J. Drewry), ott Polar Research Institute, Cambridge (1983)), updated with more recent measurements of accumulation pecially in the Antarctic Peninsula and over the Ronne-Filchner Ice Shelf. Note the very low rate in the central gions, the increase towards the margins of the ice sheet and the considerable accumulation rates in the Antarctic ninsula region (courtesy D. Vaughan).

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