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Response of the Antarctic ice sheet to future greenhouse warming

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Abstract. Possible future changes in land ice volume are mentioned frequently as an important aspect of the greenhouse problem. This paper deals with the response of the Antarctic ice sheet and presents a tentative projection of changes in global sea level for the next few hundred years, due to changes in its surface mass balance. We imposed a temperature scenario, in which surface air temperature rises to 4.2° C in the year 2100 AD and is kept constant afterwards. As GCM studies seem to indicate a higher temperature increase in polar latitudes, the response to a more extreme scenario (warming doubled) has also been investigated. The mass balance model, driven by these temperature perturbations, consists of two parts: the accumulation rate is derived from present observed values and is consequently perturbed in proportion to the saturated vapour pressure at the temperature above the inversion layer. The ablation model is based on the degree-day method. It accounts for the daily temperature cycle, uses a different degree-day factor for snow and ice melting and treats refreezing of melt water in a simple way. According to this mass balance model, the amount of accumulation over the entire ice sheet is presently 24.06×10^{11} m^3 of ice, and no runoff takes place. A 1° C uniform warming is then calculated to increase the overall mass balance by an amount of 1.43×10^{11} m^3 of ice, corresponding to a lowering of global sea level with 0.36 mm/yr. A temperature increase of 5.3° C is needed for the increase in ablation to become more important than the increase in accumulation and the temperature would have to rise by as much as 11.4° C to produce a zero *surface* mass balance. Imposing the Bellagio-scenario and accumulating changes in mass balance forward in time (static response) would then lower global sea level by 9 cm by 2100 AD. In a subsequent run with a high-resolution 3-D thermomechanic model of the ice sheet, it turns out that the *dynamic* response of the ice sheet (as compared to the direct effect of the changes in surface mass balance) becomes significant after 100 years or so. Ice-discharge across the grounding-line increases, and eventually leads to grounding-line retreat. This is particularly evident in the extreme case scenario and is important along the Antarctic Peninsula and the overdeepened outlet glaciers along the East Antarctic coast. Grounding-line retreat in the Ross and Ronne-Filchner ice shelves, on the other hand, is small or absent.

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Introduction

A potentially important effect of a climatic warming is sea-level change. Global mean sea level has risen by 10 to 25 cm over the last hundred years (Barnett 1983; Gornitz and Lebedeff 1987; Peltier and Tushingham 1989), and it is conceivable that larger changes will occur when global temperature increases by a few degrees C! By far the largest amount of ice is stored on the Antarctic continent. The area of the grounded ice sheet is 12 million km^2 and the average thickness about 2500 m. The largest thickness observed (by radar) is over 4700 m. The climate around the Antarctic continent is much colder than in Greenland (10 to 15° C temperature difference in the annual mean). As a result, the surface mass balance is positive everywhere. It varies from more than 0.5 m ice per year in the coastal areas to just a few cm in the continental interior. This is due to the very low air temperatures and the associated low moisture-content of the atmosphere.

The present state of balance is not well known, although there seems to be no evidence that the ice sheet is too far from equilibrium. This is corroborated in a study by Budd and Smith (1985), who reviewed all existing data on surface mass balance and ice velocity. They arrived at the conclusion that the total influx of ice is nearly balanced by the outflow, although on the basis of the present data, (still poor) an imbalance of up to 20% would be hard to detect. This would correspond to a change in sea level of 1.2 mm/yr. Recent modelling studies have indicated that the ice sheet must still be responding to the last glacial-interglacial transition, but that the associated rate of sea-level change is an order of magnitude smaller than the observed trend over the last hundred years (Huybrechts and Oerlemans 1988; Huybrechts 1990). Many questions still remain.

Even so, speculations about the possible contribution of the Antarctic ice sheet to future sea level have been made. In the late 1970s it has been stressed several times in the literature that the West Antarctic ice sheet, with its bed so far below sea level, could be so inherently unstable that a moderate warming may lead to a run-away situation in which the major part of the West Antarctic ice sheet disintegrates (e.g., Mercer 1978). This would raise sea level by up to 5 m in a few centuries time. Later modelling studies have suggested that this view is probably too dramatic (Van der Veen 1985) and that a sudden collapse in these short time scales is unlikely to happen. Nevertheless, the question of how Antarctica may contribute to changing sea levels is still frequently asked. So far, few studies have been conducted on the more regular effect of a change in surface mass balance.

From a general point of view, the relation between surface mass balance M and mean air temperature T is double-valued. In a very cold climate, where little melting occurs and precipitation amounts are restricted by low air temperature, the mass balance will generally *increase* with air temperature. In warmer climatic zones, where part of the precipitation falls as rain, the mass balance will generally *decrease* with temperature. While for most glaciers and the Greenland ice sheet $dM/dt<0$, the Antarctic ice sheet is mainly in the region where $dM/dT>0$, at least for the current climatic state. This has been noted by several authors (e.g., Robin 1977; Oerlemans 1982; Muszynski and Birchfield 1985) and it seems to be generally accepted that, in case of a climatic warming, the *surface* mass balance will increase by a small amount. This should lead to a fall in global sea level. When the warming is more than a few degrees C, however, melting in the coastal zone could also become important, in particular on the Antarctic peninsula. This point has received much less attention. The net effect on sea level would thus depend on which process is dominant for which type of temperature ice.

With regard to the response of the Antarctic (or any) ice sheet to climatic change, a distinction has to be made between between the *direct* or *static* effect of a changing surface mass balance and the secondary effect of a change in the dynamics of the ice sheet. On a short time scale (say, less than 100 years), it can be assumed that flow changes are too slow to significantly affect the amount of ice that is transported across the flotation line. Consequences for sea level can then be investigated by integrating mass-balance changes forward in time. So far, little attention has been paid to the static effect. When a somewhat longer time integration is envisaged or bottom melting on ice shelves becomes important, the stationary assumption is probably not justified anymore. In that case, grounding line retreat may occur and ice dynamics have to be considered too.

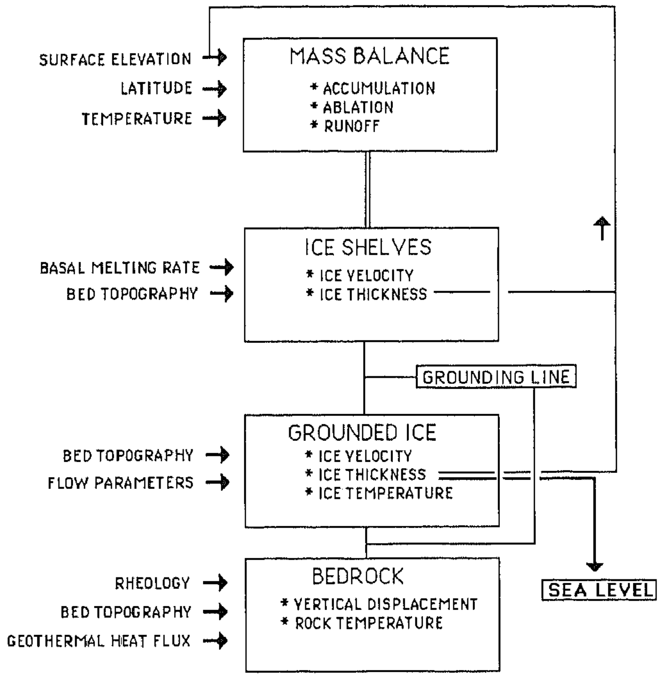

Fig. 1. Structure of the model used in this investigation. Input are given at the left-hand side. The mass balance drives the model, which has ice shelves, grounded ice, and bed adjustment as major components. The position of the grounding line is not prescribed, but internally generated. Ice thickness feeds back on surface elevation, an important parameter for the calculation of the mass balance

In the present study we have tried to perform a sensitivity test with a rather comprehensive model of the Antarctic ice sheet. The approach is quite similar to the one described in Oerlemans (1982), but with a much more refined treatment of the mass balance components, updated scenarios for the greenhouse warming and a much more detailed model for ice dynamics. In the model, both grounded ice and ice shelves are treated on a fine mesh (40 km grid size), by solving the full thermomechanical equations for ice flow in three dimensions. There is free interaction between the flow in grounded ice and floating ice, so the grounding line is *internally generated.* Figure 1 shows the structure of the model. While the formulation of the mass balance is discussed below in some detail, we refer to the accompanying paper (Huybrechts, this issue) for a description of the numerical ice-flow model.

In the interpretation of the results, we distinguish between the static effect (which can be studied with the mass balance model solely) and the dynamic effect.

Modelling the surface mass balance

In the present approach, the components of the mass balance and its perturbations are parameterized in terms of temperature, the principal forcing parameter. Although accumulation and runoff result from quite complex processes, involving the circulation pattern in the atmosphere and the energy balance at the ice sheet surface, such an approach is necessary to keep calculations times within reasonable bounds. Also, it seems that thermodynamic models for the melting/refreezing processes in the upper snow/firn/ice layers currently available are not sufficiently universal to justify application in the type of sensitivity tests we describe here.

Accumulation

The accumulation rate on the Antarctic ice sheet appears to be strongly related to air temperature, which controls the amount of vapour that can be advected inland. Such a relationship seems to be particularly strong over the inland plateau, where year-round temperatures are far below freezing, and snow deposition results from an almost continuous fallout of ice crystals and also some surface sublimation. Robin (1977) proposed that accumulation could be set proportional to the saturation-vapour pressure of the air circulating above the inversion layer. In Lorius et al. (1985) it was argued that the rate of change with temperature of the saturation-vapour pressure is the relevant quantity from a physical point of view. Indeed, the relation thus obtained between accumulation and temperature appeared to agree well with accumulation rates deduced from the ^{10}Be content in the Vostok ice core (Yiou et al. 1985).

To calculate accumulation rates, we have adopted the following procedure. As a first step, basic data sets for accumulation rate and surface temperature have been obtained by quadratically interpolating and smoothing raw data, as compiled at the Scott Polar Research Institute (Cambridge, UK), on the 40 km model grid. The resulting accumulation distribution is shown in Fig. 2. The mean annual surface temperature *change*

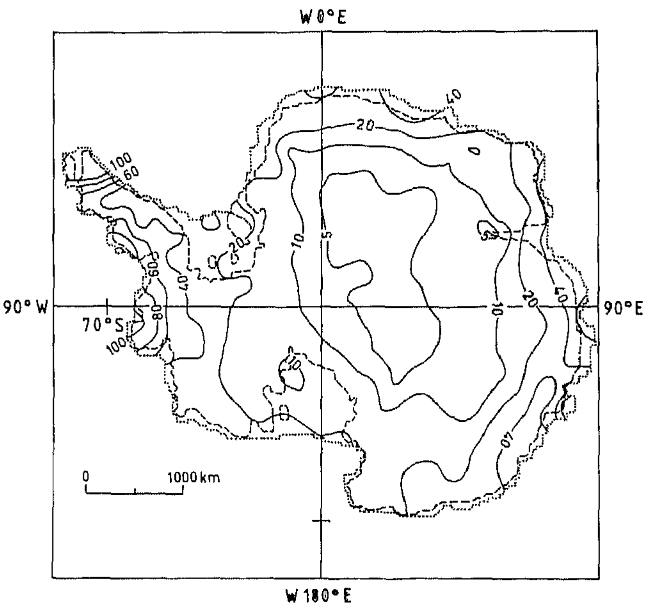

Fig. 2. Distribution of accumulation for present conditions as obtained from a computer interpolation on point measurements. *Labelled contour lines* are in cm/y

δT occurring at the grid points during the integration then consists of two parts: one associated with the imposed climatic warming $\Theta(t)$ and one with the change in surface elevation:

$\delta T = \gamma \delta H + \Theta(t)$, where
$\gamma = 5.1°$ C/km if $H \leq 1500$ m
$\gamma = 14.3°$ C/km if $H > 1500$ m

Lapse rates are suggested by a linear multiple regression study by Fortuin and Oerlemans (1990). To relate the temperature above the inversion layer (T_f) to the surface temperature (T_s) we used a relation proposed by Jouzel and Merlivat (1984):

$$T_f[\mathrm{K}] = 0.67\ T_s[\mathrm{K}] + 88.9$$

The accumulation rate for any perturbed climatic state is then obtained from the product of its reference (present) value, times the ratio of the derivatives of the saturation vapour pressure over a plane surface of ice for the reference and perturbed state:

$$M[T_f(t)] = M[T_f(\mathrm{ref})] \cdot \exp\left\{22.47\left[\frac{T_0}{T_f(\mathrm{ref})} - \frac{T_0}{T_f(t)}\right]\right\} \cdot \left\{\frac{T_f(\mathrm{ref})}{T_f(t)}\right\}^2$$

$T_0 = 273.16$ K is the triple point of water. In this approach, resulting accumulation rates typically differ by a factor of 2 for a 10° C temperature shift for conditions prevailing over central Antarctica. Although accumulation processes at the ice sheet margin and over the ice shelf are somewhat more complicated and also depend on cyclonic activity and depression paths, the above relation is nevertheless assumed to be valid over the entire ice sheet. This is a simplification and means that the pattern of low and high pressure areas is assumed to be unaffected by a climatic warming. This may be doubtful, but very little is known in this respect. For instance, available GCM simulations for a CO_2-doubling show little resemblance in their patterns of precipitation changes, not even in a qualitative sense (e.g. Schlesinger and Mitchell 1985). Furthermore, net accumulation also depends on such factors as snowdrift and evaporation, and the change in these with climatic change may be different from the change in precipitation alone. However, a more sophisticated treatment would be far beyond the scope of the present study and is probably not justified in the light of other approximations.

Runoff

At present, there is little or no surface melting in Antarctica. If temperatures go up, however, summer melting may occur in the lower reaches of the continent, and ultimately runoff towards the ocean may result. To estimate the importance of these effects, we have to rely on measurements taken in the ablation zone of the Greenland ice sheet. As shown by Braithwaite and Olesen (1989), there is a high correlation between positive degree-days and melt rates at the ice margin in West Greenland. The annual amount of positive de-

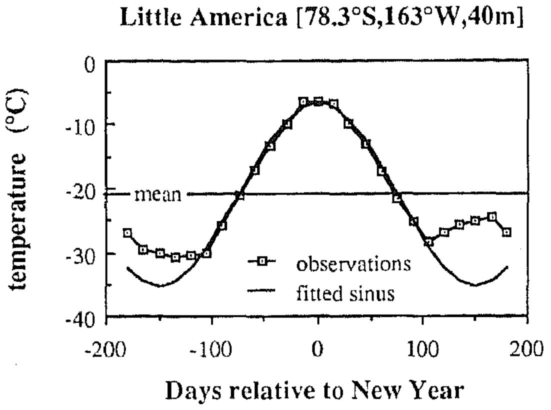

Fig. 3. Fitting of the temperature in the summer half-year with a sine function. This example is for station Little America

gree-days is parameterized in several steps. Data from 16 Antarctic coastal stations (World Survey of Climatology 1970) were used to relate summer temperature (mean of December and January) and annual temperature amplitude (both in °C) to geographical latitude ϕ (in °S):

$$T_{sum} = 25.11 - 0.39\phi + \gamma H \quad (r = 0.84)$$
$$A_{ann} = -30.74 + 0.59\phi \quad (r = 0.91)$$

The correlation coefficients appear to be sufficiently large to justify this approach. In the calculations the daily cycle was set to 3.02° C (mean value of 16 coastal stations with a standard deviation of 0.16° C) and temperature during the summer season was fitted by a sinusoidal function with a half-period of 5 months and top at January 1st. This is necessary as the yearly temperature march has a rather flat profile in winter, because insolation is zero. The procedure is illustrated in Fig. 3. By integrating the parameterized annual and daily temperature cycle through the year (with a 30 min time step) the number of positive degree-days (PDD) is now easily obtained. The result can be approximated quite accurately by a fourth-order polynomial:

$$\begin{aligned}\mathrm{PDD}\,[^{\circ}\mathrm{C\ day}] = & (58.259 - 2.201\,A_{ann} + 0.038\,A_{ann}^2) \\ & + (50.263 - 2.265\,A_{ann} + 0.045\,A_{ann}^2)\,T_{sum} \\ & + (12.326 - 0.788\,A_{ann} + 0.019\,A_{ann}^2)\,T_{sum}^2\end{aligned}$$

This is subject to the conditions that the summer temperature is between -3 and $+10°$ C, and that the annual amplitude is between 7 and 22° C

PDD represents the potential energy for melting. Figure 4 shows this potential for some selected stations. The potential energy is used for melting snow and/or ice according to the following two-step scheme:
Step 1. Snow is melted with a degree-day factor of 0.003 m ice-equivalent per unit of PDD, and the meltwater percolates into the snow cover and refreezes. Runoff occurs when enough latent heat is released by the refreezing process to raise the uppermost 2 m from the mean annual temperature to the melting point.
Step 2. Refrozen meltwater ('superimposed ice') and in a later stage glacier ice are melted with a higher degree-day factor of 0.008 m ice-equivalent per unit of PDD. (This higher value is essentially related to a lower albedo for bare ice as compared to snow.) Degree-day factors are from Braithwaite and Olesen (1989).

This process may stop at either of the stages 1 or 2, depending on the magnitude of PDD.

On the ice shelf, all melting is assumed to refreeze into superimposed ice and no runoff takes place. In view of the very small surface slope, this is to be expected for a long time to come. Also, at this stage melting beneath the ice shelves is not considered.

The treatment of the refreezing process sketched above is admittedly rather crude. However, an alternative technique seems difficult to construct with our present knowledge. Field studies of the energy transfer involved in the formation of ice lenses and superimposed ice have been very limited. Useful data have been obtained from field campaigns on the Laika ice cap in arctic Canada (Blatter et al. 1988) and on the Greenland ice sheet (Ambach 1963, 1968). Analysis of Ambach's results suggests that assuming that a 2 m thick (ice-equivalent) layer heats up annually by refreezing gives the correct order of magnitude.

Sensitivity of the surface mass balance

With the mass-balance model described above it is possible to study how the total ice budget gained (or lost) at the surface depends on temperature. Figure 5a dis-

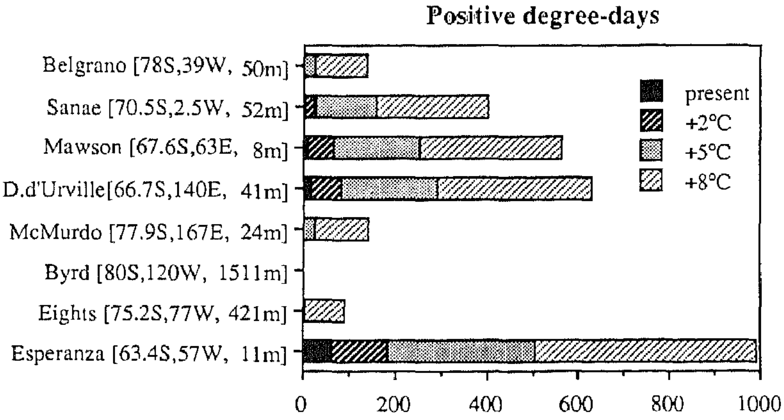

Fig. 4. Calculated amount of positive degree-days (PDD) for various Antarctic stations

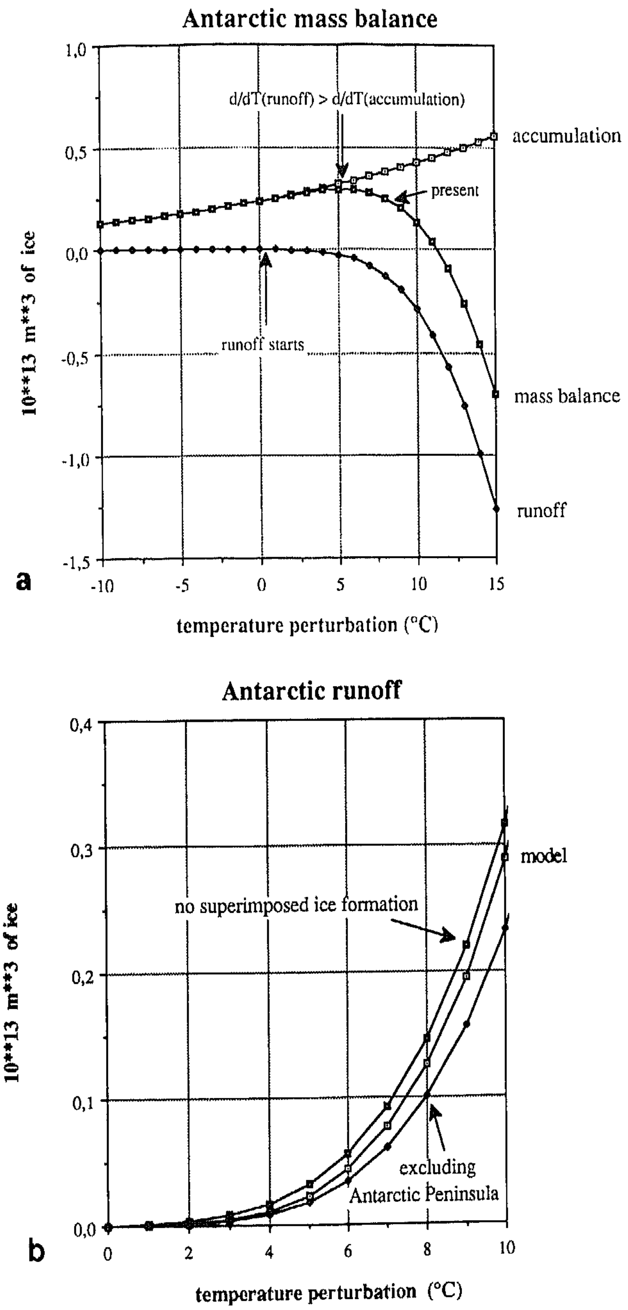

Fig. 5. a Dependence of Antarctic mass balance components on temperature relative to present. For a temperature rise below 5.3 K, the mass balance would be larger than today. **b** The effect of superimposed ice formation on the predicted runoff. The *lowest curve* indicates the relative importance of the Antarctic Peninsula

plays the surface mass-balance and its components integrated over the entire ice sheet. Only grounded ice has been considered as ice mass changes in the floating shelves do not effect global sea level. The surface elevation of the ice sheet, which is a necessary input parameter for the mass-balance model, was taken from the 'interglacial steady-state' as obtained by the numerical ice-flow model. For present conditions, we found that an amount of 24.06×10^{11} m^3 of ice equivalent is deposited on the surface every year. This amount of ice would lower global sea level by 6 mm/yr, were it not that the same amount is discharged across the grounding line to form an ice shelf and, ultimately, icebergs (assuming a stationary state).

For a 1° C temperature rise with respect to present conditions, the mass balance increases by an amount of $1.43\ 10^{11}$ m^3 of ice, i.e., by 5.9%. This corresponds to an additional mean accumulation-rate spread out over the entire grounded ice sheet of 10.3 mm/year water equivalent. Since the area of the oceans is about 29 times larger than the area of the Antarctic ice sheet, it would imply a rate of sea-level lowering of 0.36 mm/year.

Runoff appears to be negligible for the present climate, but we found it to increase progressively with temperature. This higher-order dependence results because a warming climate will not only increase melting rates, but also the length of the ablation season. In the calculation, the refreezing process does not make a large difference (Fig. 5b). As long as the temperature increase is below 5.3° C, the mass balance will increase, because the increase in accumulation can still outweigh the increase in runoff. At that stage, the total accumulation amounts to 32.75×10^{11} m^3, of which 8.7% runs off in summer. This would add an extra amount of 1.46 mm/year to the world's oceans, if ice sheet dynamics were to remain unchanged. This applies to the total surface balance, but the pattern of change depends on altitude, of course.

Temperature has to rise by 8.3° C to produce the present mass-balance again, and for a temperature increase of 11.4° C the total surface balance becomes negative. If such a situation were to last over a longer time (say thousands of years) ice shelves could not be sustained anymore and a large ice-free strip would develop on the Antarctic continent.

Figure 6 demonstrates how the surface mass-balance changes for various values of the imposed climatic warming. Near the coast, the mass balance will quickly decrease and even become negative, while the mass balance on the plateau increases. As a consequence, the balance gradient steepens in a warmer climate, which has significant consequences for the ice-flow regime. This point is taken up in the next section.

Modelling technique and strategy

Brief description of the ice-flow model

A full account of the mathematical equations in the model is described in the accompanying paper (Huybrechts, this issue). The model for the Antarctic ice sheet as used here is essentially a 3-dimensional extension of the thermo-mechanical flowline model described in Huybrechts and Oerlemans (1988). A basic assumption is that the ice flows in direct response to pressure gradients set up by gravity. Two processes contribute to the mass flow: internal deformation (plane shear) and sliding over the bed. The softness parameter of ice, which determines the rate of deformation, depends strongly on temperature. For this reason,

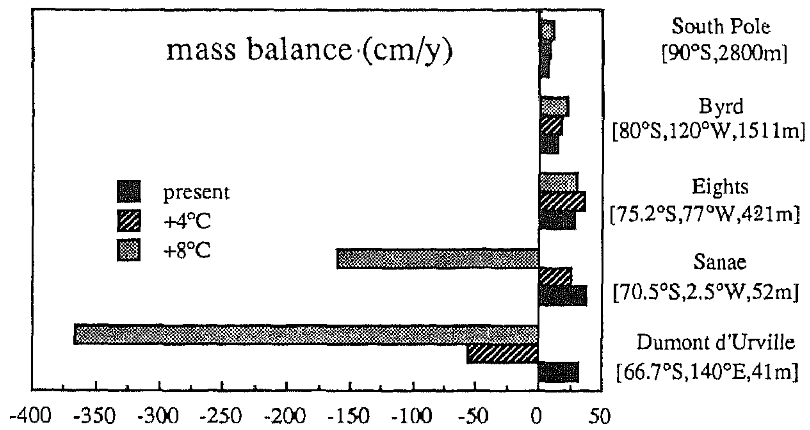

Fig. 6. Mass balance for present conditions and in a warmer climate for some selected Antarctic stations

the 3-dimensional temperature field is calculated. Sliding is restricted to regions that are at the pressure melting point.

Furthermore, a time-dependent numerical ice shelf model is included and the grounding line is treated in a dynamic fashion. Also, the response of the underlying bedrock to changing ice load is considered in some detail, including a temperature calculation.

All calculations (grounded ice and ice shelves) are done on the same grid. The grid point distance is 40 km. With 10 layers in the vertical, with closer spacing near the bedrock surface, and a few grid points in the bed for the temperature calculations, this leads to a domain of somewhat more than 200 000 grid points. All numerical integrations reported have been performed on a CRAY-2 computer.

Model input consists of bed topography, surface temperature, mass balance, thermal parameters and an initial state (which may be a thin slab of ice of equal thickness). The model then essentially outputs the time-dependent 3-dimensional ice sheet geometry, and the velocity and temperature fields. Bedrock elevation, surface elevation and ice thickness data for the *present ice sheet* originate from a data set provided by W. F. Budd (Budd et al. 1984), which was based on the map folio series published by the Scott Polar Research Institute (Drewry 1983). The undisturbed bedrock heights, needed in the bed adjustment calculations, have been reconstructed by assuming local isostatic equilibrium.

Interglacial 'steady state' reference run

Investigating changes in ice-sheet geometry related to greenhouse warming requires that a reference state is defined. Starting from the observed ice sheet and running the model forward in time with a prescribed temperature forcing is not a very meaningful approach, as it is unlikely that the input data are in full internal equilibrium with the model physics. The model ice sheet will certainly evolve from the present observed state, but it would be unclear why. Shortcomings in the description of ice mechanics, insufficient data coverage, or the fact that the present ice sheet is just not in steady state could all play a role. As a consequence, it would be impossible afterwards to distinguish between the 'natural model evolution' and the 'real' ice sheet response. One way to circumvent this ambiguity is to put forward a steady state and relax the model to equilibrium. The resulting steady-state interglacial reference run, although differing from the presently observed data in some detail, should then be used as input in a climatic change experiment.

To do this, the coupled velocity and temperature fields have first been integrated forward for 100 000 years in a quasi-diagnostic way (keeping ice thickness and the grounding-line position fixed). This significantly speeds up the whole procedure. Then the grounding line is allowed to migrate and the model relaxes to a stationary state for another 100 000 years. The result is shown in Fig. 7. Deviations from the present observed state include somewhat thicker ice in the Antarctic Peninsula (mainly due to poor bedrock data below the present ice shelf areas) and in the Pine Island and Thwaites Glacier catchment areas (West Antarctic ice sheet) and a slight recession of the grounding line at

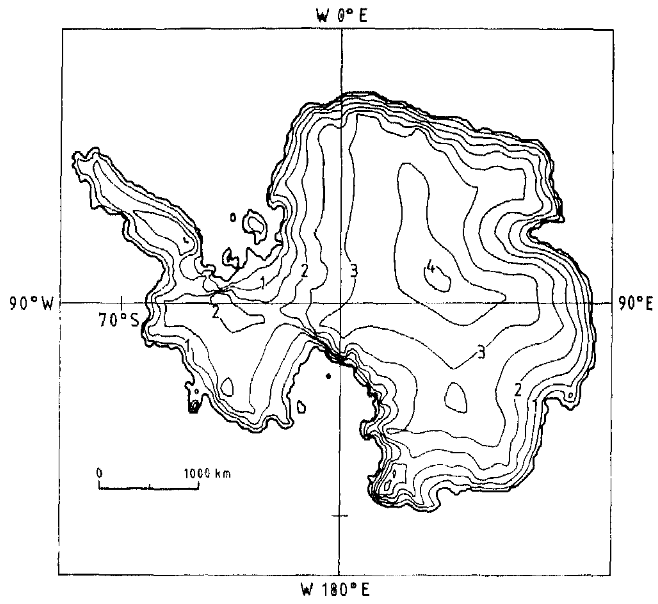

Fig. 7. The 'interglacial reference state' as produced by the model, showing surface elevation (in km)

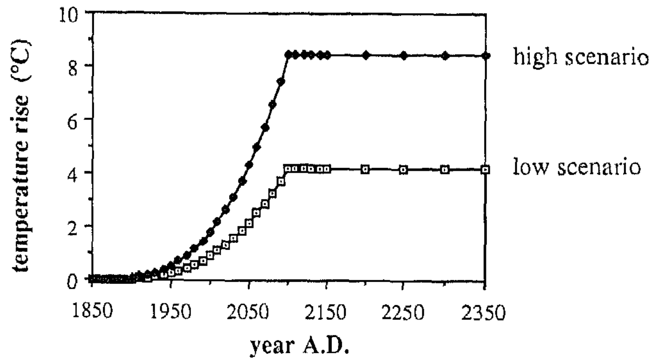

Fig. 8. The temperature scenarios ('high' and 'low') used to force the dynamic model. The lower curve is the 'Bellagio-scenario'

the seaward edge of the most overdeepended outlet glaciers of East Antarctica (Totten, Ninnis and Mertz Glaciers). However, considering the complexity of the model and the high number of degrees of freedom, the resulting configuration is certainly acceptable, and the model seems capable of reproducing the most important characteristics of the ice sheet.

Response of the ice sheet to climatic warming

The temperature scenario employed to force the mass balance model is taken from the Villach II meeting in the fall of 1987 (Climatic Change 1989 vol 15, nos. 1/2). It is also known as the Bellagio-scenario and represents a best guess on the basis of available model simulations. It is shown in Fig. 8 and predicts the temperature to rise by 4.2° C by the year 2100, starting from 1850, which is considered to be an undisturbed reference state. The probability that this figure is correct within 1.5° C is estimated to be 67%. This figure applies to global-mean surface-temperature, however. As sensitivity studies with GCMs seem to point to a larger response in polar latitudes (e.g. Manabe and Stouffer 1980), an alternative experiment was carried out with a temperature rise twice as much. Furthermore, temperature perturbations have been assumed to be uniform through the year and independent of latitude.

Static response

In view of the long response time scales of large ice sheets, in particular in the interior of the East Antarctic ice sheet, we first consider the 'static response', i.e., the present ice-sheet geometry is assumed to be in equilibrium with the velocity field, and changes with respect to the present mass balance are accumulated forward in time. So the velocity field remains unchanged, in effect keeping the amount of ice transported across the grounding line constant. In this calculation, changing surface elevations have been taken into account as a contribution to local temperature changes (although this contribution is very small) and the maximum amount of ice that can melt away is equal to the present ice thickness. The results in terms of equivalent sea-level rise are shown in Fig. 9. Since the imposed temperature perturbations are lower or equal to the 8.3 K mark (implying that the mass balance is always larger than at present), the expected increase in ice volume shows up. This is particularly apparent for the low scenario, where the volume increase is found to lower global sea level by 1.2 cm but the year 2000, 9.1 cm by the year 2100 and up to 41 cm by the year 2350. However, a somewhat more complicated response shows up for the high scenario, where the ice volume decreases somewhat around 2100, because the 8.4° C temperature perturbation just exceeds the 8.3° C mark slightly. This effect is restored later because at the edge, where melting rates have gone up to a few meters per year, the total ice thickness has melted away. Obviously, at this stage the assumptions made in this static experiment break down.

Dynamic response

Considering these important changes in ice thickness, in particular for the high scenario, the ice sheet is ex-

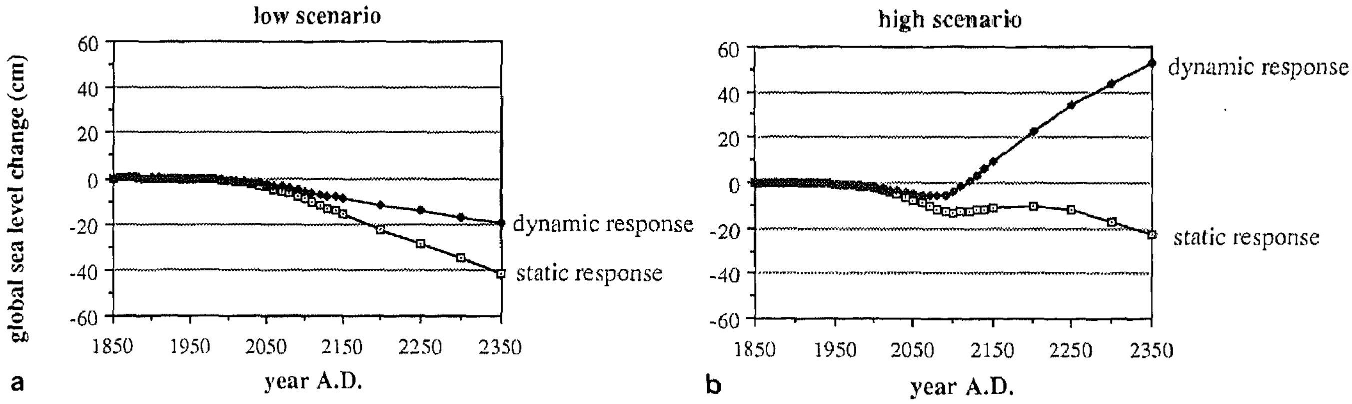

Fig. 9 a, b. Response of the Antarctic ice sheet expressed in global sea level changes for the 'low scenario' (a), and for the 'high scenario' (b) pected to show a dynamic response too. On the plateau this is not the case, and the increased accumulation rates lead to increased ice thicknesses of a few tens of meters at most, without a significant change in the velocity field at all. On the other hand, nearer to the coast and the grounding line the situation is quite different, because response times are much shorter here. Several dynamic processes can be distinguished that result in lower grounded ice-volumes as compared to the corresponding static run. For the small warming case, the mass balance increases in first instance everywhere and ice thickening is largest at low elevations. After a while, however, velocities in the lower reaches start to react to this and increase, thereby enhancing ice transport towards the ocean. This leads to thinning and partly counteracts the thickening due to increased accumulation. This sucking effect appears to be mainly due to increased ice-mass discharge on the ice shelves and is apparent at a few gridpoints inland. For a somewhat larger warming, the mass balance starts to decrease at lower elevations and eventually surface melting and runoff occur in the most northerly locations. At this stage, however, the resulting change in ice thickness near the grounding line is not large enough to produce significant grounding-line retreat. In first instance there is some seaward migration (if there is no melting), followed by a slight recession in some places, but as an overall picture the grounded domain has not changed its dimensions much.

This is essentially the picture that applies for the low scenario. By the year 2350 some seaward migration of the grounding line is produced in the most poleward locations (the Ronne-Filchner and Ross ice shelf) and a slight retreat is apparent in the Antarctic Peninsula and parts of the East Antarctic coast, in particular between 110 and 150° E. Basically, the net static effect is thus partly counteracted by an increased ice-mass discharge across the grounding line, and this leads to reduced rates of global sea-level lowering. Compared to the static effect, this reduction amounts to 18% by 2000, 36% by 2100 and 53% by 2350 (see Fig. 9a). This incease reflects the response time-scale near and beyond the grounding line. So, from this low greenhouse warming experiment, it can be concluded, that as long as the warming is below about 5° C, the Antarctic ice sheet will probably reduce global sea level by some tens of centimeters a few centuries later.

The results shown in Fig. 9b show that, the situation becomes quite different for a larger warming. In particular, when this warming exceeds 5–6° C, i.e. after 2060–2070 in the high scenario, thinning at the grounding line increases and significant grounding-line retreat sets in. Viscous spreading of the newly-formed ice shelf parts then greatly enhances ice-mass discharge from the grounded ice sheet. This has immediate consequences for global sea level and the dynamic effect becomes dominant now, leading to a global sea level *rise* by as much as 0.5 m after 500 years of simulated time. Nevertheless, grounding-line recession is still insignificant along the major ice shelves (Ronne-Filchner and Ross), because, even with a warming of 8.4° C, the number of positive degree-days is at these locations still below 50 and no runoff takes place. The areas where grounding line retreat has taken place at the year 2500 (when calculations end) are shown in black in Fig. 10.

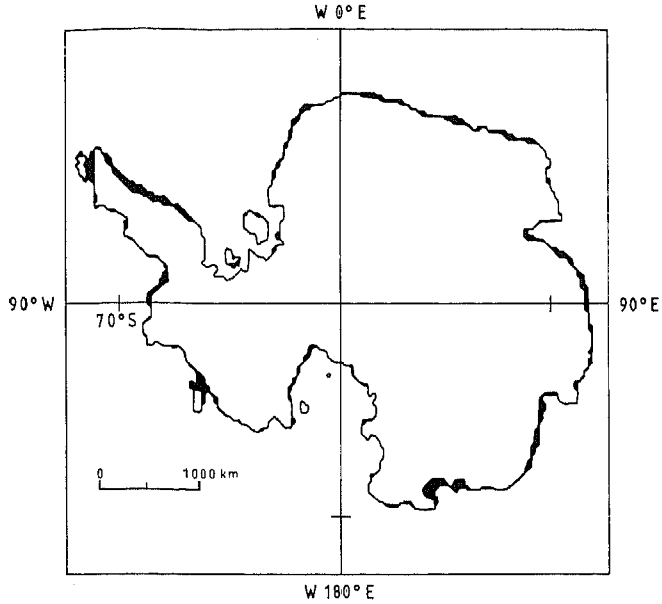

Fig. 10. Grounding-line retreat for the 'high scenario' experiment. The *black areas* indicate where grounded ice has been replaced by ice shelf by the year 2500 (when the integration stops)

Effect of melting beneath ice shelves

In the experiments discussed, basal melting beneath ice shelves has not been considered. This is simply because it is not known how the oceanic circulation and heat exchange beneath shelves will react to a climatic warming and higher ocean temperatures. In the parameterization of the surface mass balance, no surface runoff shows up in the catchment areas feeding the Filchner-Ronne and Ross ice shelves, not even in the rather extreme +8.4° C warming. Thus, it is clear that we did not find any sign of a significant grounding line retreat here, let alone a catastrophic collapse of the West Antarctic ice sheet (WAIS). Nevertheless, it has been suggested by several authors that increased melting could weaken the ice shelves and reduce their buttressing effect, ultimately inducing a collapse of the WAIS. To investigate this point somewhat further, we finally performed a rather drastic experiment, in which the mass balance on the ice shelf was reduced stepwise by 1 m/y at time zero, while keeping the surface climate fixed. In this experiment we did not consider the possibility of the ice shelf completely breaking up, so that calving could take place at the grounding line, but were only interested in the effect of ice shelf thinning on the grounded ice sheet.

Results for 500 years of simulated time are summarized in Fig. 11, which shows the number of gridpoints with grounded ice and sea-level change. As can be

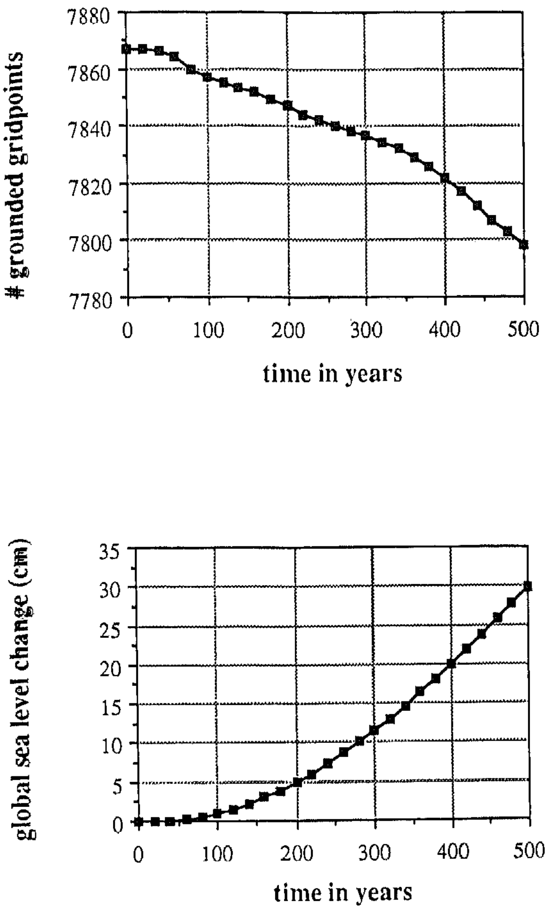

Fig. 11. Response of the Antarctic ice sheet to an instantaneously imposed melting rate beneath ice shelves of 1 m/yr. According to the model, the area of grounded ice would decrease by less than 1% in 500 years

judged from this figure, ice shelf thinning does lead to grounding-line retreat, but the total effect is still smaller than the direct surface mass-balance response in the high scenario. On the basis of these model results, we tend to conclude that a West Antarctic ice sheet collapse on a time scale of half a millenium or so is unlikely to happen. It is also realized, however, that the fast ice streams presently flowing through the ice sheet are not explicitly modelled here, and that they could possibly speed up the response under particular circumstances.

Conclusions

Numerical modelling of the Antarctic ice sheet and its surroundings certainly helps to make estimates of future sea-level change associated with greenhouse warming. In our view, the model used in this study is a well-balanced representation of current knowledge of the dynamics of ice sheets and ice shelves, supplemented with a mass-balance parameterization which is reasonable. The results of the numerical experiments can be summarized as follows:

1. For present conditions, an amount of 21.89 10^{11} m^3 water equivalent is deposited on the ice sheet surface every year and no runoff takes place.
2. For a rise in annual mean temperature of less than 5.3 K, the increase in accumulation outweighs the increase in runoff and a larger-than-present surface budget results. For a temperature increase of more than 8.3 K, the mass balance drops below the present value. To make the surface balance negative, a 11.4 K temperature rise would be needed.
3. Imposing the Bellagio-scenario, the model predicts a 6 cm sea-level lowering by 2100 AD due to changes on the Antarctic ice sheet and up to 20 cm 500 years after the onset of the warming (assuming that increased melting beneath ice shelves is not important).
4. Grounding line retreat sets in along the East Antarctic coast and in the Antarctic Peninsula if the warming exceeds 5–6° C. Imposing a temperature rise twice as much as the Bellagio-scenario would then result in a positive contribution to global sea level after the year 2100.
5. A stepwise 1 m/yr increase in the melting rate beneath ice shelves, all other things being equal, would cause a sea-level rise of only 5 cm in 200 years, and 30 cm in 500 years.

However, a number of uncertainties still remain. The weakest points in our approach probably are the treatment of the ice streams and the formulation of the melting/run-off process in the case of significant warming. Extensive field studies are currently being carried out to study the dynamics of ice streams, and one may hope that improved understanding and modelling will result. The treatment of the melting can also be improved, and here much can be taken from energy balance/mass balance studies that have been undertaken in other parts of the world. Also model predictions on the magnitude and the geographical distribution of the warming still differ widely and the effects of this warming on the oceanic circulation and melting rates beneath ice shelves are not well known. Nevertheless, in spite of the fact that our knowledge appears still inadequate in many respects, it seems very likely that in a warmer climate the Antarctic ice sheet will at first grow. Also the hypothesis of a catastrophic collapse of the West Antarctic ice sheet does not seem to be supported by the model results presented here.

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