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Modelling the long-term response of the Antarctic ice sheet to global warming

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ABSTRACT. The primary effects of global warming on the Antarctic ice sheet can involve increases in surface melt for limited areas at lower elevations, increases in net accumulation, and increased basal melting under floating ice. For moderate global warming, resulting in ocean temperature increases of a few °C, the *large increase in basal melting can become the dominant factor in the long-term response of the ice sheet.* The results from ice-sheet modelling show that the increased basal melt rates lead to a reduction of the ice shelves, increased strain rates and flow at the grounding lines, then thinning and floating of the marine ice sheets, with consequential further basal melting.

The mass loss from basal melting is counteracted to some extent by the increased accumulation, but in the long term the area of ice cover decreases, particularly in West Antarctica, and the mass loss can dominate. The ice-sheet–ice-shelf model of Budd and others (1994) with 20 km resolution has been modified and used to carry out a number of sensitivity studies of the long-term response of the ice sheet to prescribed amounts of global warming. The changes in the ice sheet are computed out to near-equilibrium, but most of the changes take place within the first few thousand years. For a global mean temperature increase of 3°C with an ice-shelf basal melt rate of 5 m a^{-1} the ice shelves disappear within the first few hundred years, and the marine-based parts of the ice sheet thin and retreat. By 2000 years the West Antarctic region is reduced to a number of small, isolated ice caps based on the bedrock regions which are near or above sea level. This allows the warmer surface ocean water to circulate through the archipelago in summer, causing a large change to the local climate of the region.

INTRODUCTION AND BACKGROUND

There has been a continuing controversy over recent years regarding the degree of warming and minimum ice extent which may have prevailed in the Antarctic during the Pliocene (see, e.g., Webb and Harwood, 1991; Barrett and others, 1992; Sugden, 1992). The debate continued with a number of presentations at this symposium concerned with the interpretation of fossils in the Sirius Formation of the Transantarctic Mountains, and attention has also been drawn to evidence of a warmer Pliocene climate from other areas such as the Marine Plain of the Vestfold Hills (Quilty, 1996). An occasional theme in these discussions is whether numerical models of the ice sheet can illuminate the range of possible ice-sheet climate sensitivities and the response times for such changes.

A second application of interest for such models is the long-term response of the Antarctic ice sheet to the projected climate changes caused by the increase of atmospheric concentrations of CO_2 and other greenhouse gases.

When considering the stability of the Antarctic ice sheet it is important to recall the remarks of Oerlemans and Van der Veen (1984) and Van der Veen (1986) that substantial and rapid partial deglaciation over millennial time-scales would have a negligible impact on the global energy budget. This is because, over these time-scales, only a very small fraction of the energy transport from the atmosphere and ocean is required. The persistence of the Antarctic ice sheet is accordingly more dependent on the conditions of the atmosphere and ocean which restrict the flow of heat to the ice. An important example is the way in which at present the continual formation of sea ice around the periphery shields the floating ice shelves from warmer circumpolar waters.

The question of the equilibrium extent of Antarctic ice as a function of increasing magnitudes of climatic warming has been addressed using ice-sheet modelling by Huybrechts (1994). However, that study did not include grounding-line dynamics or ice-shelf flow, so that the extent of the ice, including ice grounded below sea level, was "entirely controlled by the surface mass balance". In effect the climatic forcings used by Huybrechts (1994) to represent the warmer climates were the changes in net surface accumulation resulting from increasing precipitation with temperature, offset against increasing ablation and surface melting which can predominate at the lower elevations. At present, and for small amounts of warming, net surface ablation in the Antarctic is very small and most of the ice loss occurs from the calving of icebergs and basal melting under the floating ice. Some ice shelves at present have basal growth of marine ice, but it is understood that such growth can occur only following cooling from melting from further under the ice (cf. Jacobs and others, 1992). The basal-ice net balance is very sensitive to the temperature of the ocean water circulating underneath the ice shelves, as shown by Williams and others (1998).

An important early response to climatic warming in the Antarctic is an increase in the mean temperature of the water in front of the ice shelves at depths below the ice-shelf base, which allows the circulation of warmer water underneath the ice shelves (O'Farrell and others, 1997). This results in increased basal melting and decreased basal freezing, and appreciable net total melt rates can prevail, even for ocean temperature just 1°C above the present conditions throughout the water column at the ice front (Williams and others, 1998). Recent inferences by Nicholls (1997) that basal melting could decrease with warmer conditions were based on observed changes through a seasonal cycle with an ocean-temperature-change amplitude of about 0.1°C. This is quite different from the future warming conditions considered here, where the warmer, more saline Circumpolar Deep Water increases in temperature and, because of the reduction in deep mixing under the diminished sea ice, becomes able to circulate under the ice shelves with a temperature of 2–3°C above freezing.

It is recognised that the mechanics of the transition between the grounded ice and floating ice is in general quite complex (e.g. Morland and Shoemaker, 1982; MacAyeal, 1987; Paterson, 1994; Hindmarsh, 1996). At present, the thick ice sections of the major ice streams flowing into the grounding zones of the existing large ice shelves have relatively low strain rates in that region due to the resistance of the embayed ice shelves (cf. Thomas and MacAyeal, 1982). By comparison those thick outlet glaciers which have unimpeded floating tongues have large strain rates, associated with their large ice thicknesses, following the Weertman (1957)-type floating-ice strain-rate relation (cf. Crabtree and Doake, 1982; Morgan and others, 1982; Budd and others, 1987).

As the ice shelves thin from increased basal melting, increased deformation and calving can also set in, and, as the thinner ice-shelf ice is removed, the thicker ice regions near the grounding lines of the ice streams become free to strain at the larger rates associated with their free-floating thickness in a manner comparable to that of the existing unimpeded thick outlet glaciers. This process, which is illustrated in the model results below, then allows the ice streams to also thin and flow more rapidly towards their grounding lines, which then retreat. Consequently, the processes of *increasing basal melt rates* of the ice shelves, followed by the *increasing strain rates of the thick floating ice*, near the grounding lines, as the ice shelves disappear, are the *dominant processes involved in the retreat of marine-based ice sheets* as a result of increasing warming. These processes were incorporated into the forcing for an ice-sheet model by Budd and others (1987) to study the response of the West Antarctic ice sheet to future warming. A high-resolution flow-band model with 20 km longitudinal resolution was used with prescribed increases in thinning rate, associated with increased melting rates, and strain rates dependent on the ice thickness, to simulate the transient and equilibrium responses to prescribed warming and melt rates. A coarser-resolution model with a 100 km grid was also used to obtain the similar transient and equilibrium responses for the whole Antarctic ice sheet attained after 10 000 years. These models did not explicitly include the ice shelves except in determining the thickness and strain-rate changes near the grounding lines from the increased melting, and the extent of the floating ice to the calving front.

Explicit ice shelves were included in the model of the whole Antarctic ice sheet at 20 km resolution by Budd and others (1994) to examine the similar response to future warming using prescribed basal melting rates increasing as a function of temperature. These model runs were restricted to the next 500 years because of the large amount of computer time required. The simulations also included the impact of increasing net accumulation associated with warming, taken from the increases in precipitation minus evaporation ($P - E$) derived for future warming scenarios by Budd and Simmonds (1991). The amount of warming considered was not enough for significant surface melting to occur, and corresponded to increases in global mean temperature of up to 3.2°C.

More recently, the same ice-sheet model was used with the output from a *fully coupled atmosphere–ocean–sea-ice model* simulating the transient response to gradually increasing atmospheric concentrations of greenhouse gases up to three times equivalent CO_2 and then keeping that level constant for a further 500 years (O'Farrell and others, 1997). In that study, the forcing for the *ice-sheet ice-shelf model* included increased basal melting as a function of increasing water temperature (at the depth below the ice shelf) using temperatures extrapolated from model gridpoints adjacent to the ice shelves, while the increased net accumulation was explicitly computed in the coupled model from changes in ($P - E$).

It was also found from the climate-model simulation that surface temperatures did not rise high enough for surface melting to become significant in the total mass balance. A number of sensitivity studies were also carried out with the ice-sheet model to examine separately the effects of the increase in net surface accumulation alone and the rapid onset of the prescribed water-temperature changes. Initially the increased accumulation offset the loss of grounded ice to the sea, but as time progressed and the ice shelves retreated, the loss from the grounded ice, due to increased flow rates, became dominant and the ice-sheet volume decreased. Both Budd and others (1994) and O'Farrell and others (1997) started with control simulations based on present conditions where there is no net basal melt under the large ice shelves, and melt for floating ice occurs only for the most northerly latitudes around Antarctica. The warming-anomaly simulations were then carried out with prescribed additional melt rates (for sensitivity studies) of 1, 3 and 10 m a^{-1} in Budd and others (1994) and for values corresponding to 1° and 2°C warming (giving 6.6 and 18.6 m a^{-1} from the Russell-Head (1980) relation) in O'Farrell and others (1997), as well as the time-varying forcing from the climate-model output. In the simulations of O'Farrell and others (1997) the ice-sheet model was not run out to equilibrium, mainly because of the extensive amount of computer time required.

Here we concentrate on the period after the ice shelves have largely disappeared, and assume that the warmer conditions are maintained with the warmer water circulating in the exposed embayments to allow the higher melt rates to continue at a constant prescribed level as the ice-sheet model runs out to a new equilibrium.

The main objective of this paper is to show that the relatively modest ocean warming imposed here over an extended period can change the character of Antarctic ice shelves, introducing a major disequilibrium near the grounding line of the marine parts of the ice sheet, which can alter the geometry of the ice sheet and the local climate considerably on a millennial time-scale. The much greater melting effect of small increases in ocean temperature not only far outweighs the upper surface ablative power of rising air temperature, but also imposes a strain regime at the grounding line which drains out the marine ice sheet by drastically altering the geometry of the ice in the ice-shelf basins.

The more specific aim is therefore to show the long-term response of the ice-sheet model and the changes in the ice configuration for the continued fixed degree of warming obtained from the previous climate-model study. We find that the bulk of the West Antarctic ice sheet, in particular that part grounded well below sea level, can disappear for a relatively small amount of global mean warming (3°C). The time required to reach a new steady-state equilibrium is of the order of 20 ka, but most of the changes take place within the first few thousand years, so we concentrate here on describing the changes taking place over the first 2 ka of the warming.

In any case, once the contiguous West Antarctic ice sheet is removed in a warmer climate, the warmer surface ocean water with reduced sea-ice cover is able to circulate around the residual archipelago. This results in dramatically altered circumstances for the regional climate, and calls for reconsideration of the longer-term climate forcing. As discussed later, one can expect dramatically increased mean summer temperatures for the region compared with those which prevail at present.

BASAL MELTING AND ICE-SHELF GEOMETRY

Whatever the details of force balances involved in the interaction of grounded ice sheets and the major embayed ice shelves of Antarctica, it is clear that the strain rates experienced by the thick ice at the grounding lines of these shelves are substantially lower than the free thinning rates that would be expected if they abutted the ocean as ice cliffs (Budd and others, 1987). Once basal melting from contact with a warmer ocean overtakes the net snow accumulation over the ice shelves , and the ice-shelf front begins to retreat, ice shelves in the major bays take on a different and simpler character. In a negative local net mass-balance regime, ice shelves are naturally shorter from grounding line to calving front, so that the Weertman-type strain thinning and the basal melting both work to taper the ice-thickness profile. For melt rates of several m a^{-1}, these short shelves are subject to essentially unrestrained "free" thinning rates reaching back to the vicinity of the grounding line. By contrast, under a positive local ice-shelf mass-balance regime, an ice shelf may become sufficiently thick to extend over a greater distance, with strain thinning acting in opposition to the thickness advection and net surface and basal accumulation, resulting in filling of bays, and contacts with distant islands or other pinning points, and producing the embayed situation evident today for the major Antarctic ice shelves, where the transverse shear stresses across the ice shelf play an important and complicating role (Budd and others, 1982; Thomas and MacAyeal, 1982).

FORCING FOR THE ICE-SHEET MODEL

The basal melt rates used by Budd and others (1994) were based on the free-floating ice-melt rates of Russell-Head (1980), supported by the iceberg dissolution rates derived by Budd and others (1980) and Hamley and Budd (1986). The results of Williams and others (1998) indicate that if there is ice growth taking place under an ice shelf at present, a certain amount of additional melting is required to eliminate the ice growth completely, and this reduces the amount of net melt that would occur without the freezing. While extensive extra melting was found for modest increases in ocean temperatures, the increases were less than those estimated on the simpler basis of the free-floating ice and iceberg studies. Therefore in this study (as we do not explicitly treat the high-resolution modelling of the water circulation under the ice shelves), we choose conservatively low melt rates for the water temperatures in comparison with those given in Budd and others (1994, table 3). Those values were based on the results from Russell-Head (1980) and can be represented by the relation

$$M = c(\Delta\theta)^{\frac{3}{2}}, \qquad (1)$$

where M is the basal melt rate (in m a^{-1}), $\Delta\theta$ is the difference from the pressure-melting point for the in situ ocean water, and c is a constant with the value of 6.57 m a^{-1} $K^{-3/2}$. From the coupled model results of O'Farrell and others (1997) the mean ocean temperature increase for the water column below the ice fronts is more than 1°C by the time of $2 \times CO_2$, and over 2°C by $3 \times CO_2$ with the warming slowly increasing further over time.

It should be noted that the climate-modelling runs were not coupled to the ice-sheet modelling and did not include the ocean circulation under the ice shelves or the change in geometry as the ice shelves disappeared. Here we have assumed the warmer water continues to circulate through the exposed embayments and under the residual floating ice, giving the same melt rates.

Observations of floating-ice melt rates in the Antarctic are still relatively limited. A summary has been given by Budd and others (1987, table 2). This includes reference to values up to 7 m a^{-1} with present water temperatures. For higher water temperatures, melt rates have been inferred from iceberg dissolution (Budd and others, 1980; Hamley and Budd, 1986) which appear to be compatible with the laboratory-scale results of Russell-Head (1980). Holdsworth (1982) derived ice-edge melt rates over 8 m a^{-1} in water at −1.3°C, about 0.5°C above freezing point. These values illustrate the uncertainty of melt rates as a function of water temperature, but suggest the values used here are not unreasonable for the warmer conditions considered.

For the ice-sheet model a series of simulations have been carried out with different melt rates to determine the sensitivity of the final configuration, and the timing of the retreat, and ice-volume changes, to the increased melting. It was found in the earlier studies that the main effect is that the ice shelves melt more slowly for the lower melt rates over the first 100 or 200 years. By 500 years the differences are relatively small, as the high strain rate becomes dominant, and by 2000–4000 years the ice-sheet-edge configurations, which are close to equilibrium, differ very little. We present here, as an example, the results of a conservative choice of melt rate, 5 m a^{-1}, to illustrate the ice-sheet response.

Sensitivity studies have also been carried out for changes in the net accumulation. By $3 \times CO_2$ the coupled climate model gives a mean accumulation-rate increase of about 50% over the present-day accumulation. Although the accumulation rate could be expected to increase even further as the ice sheet retreats and more ocean water is exposed, the sensitivity studies show that this is relatively unimportant for the final ice-sheet-edge configuration. The reasons for this are that the melt rates are still large compared with the surface net accumulation rate, and that the area of the ice sheet over which the snow can accumulate decreases.

Therefore, for the case chosen for our standard long simulation, the base-level net accumulation rate is set at 150% of the present distribution. Within the model it also increases where ice-sheet elevation lowers, in accordance with the elevation desert effect described by Budd and Jenssen (1989). Other aspects of the ice-sheet model are also similar to that described by Budd and Jenssen (1989) and Budd and others (1994), although it is worth emphasising the major features which control the boundary regions of the grounding zones. For deep bedrock the ice-sliding relation allows a smooth transition from grounding to floating, with increasing sliding velocity (V_s) in regions of decreasing basal shear stress (τ) due to the influence of decreasing effective normal stress, taken as proportional to the thickness of ice above buoyancy (Z^*), through the relation

$$V_s = k_1 \tau (Z^* + k_2 Z^{*2})^{-1}, \qquad (2)$$

where $k_1 = 1\,\mathrm{m^2\,a^{-1}\,Pa^{-1}}$, and $k_2 = 400\,\mathrm{m^{-1}}$, and the minimum value for Z^* is set at 20 m.

The simplified ice-shelf model used in the studies by Budd and others (1994) and O'Farrell and others (1997) was designed to treat the strong transverse shear strain rates present in embayed ice shelves. For the present study, which concentrates on the period after the major ice shelves have retreated, where the floating ice is primarily in the Weertman free-straining regime a different simple ice-shelf model is appropriate. The ice-shelf flow direction is based on the large-scale slope of the ice shelf, while the increments in flow rate are calculated from the vertical strain rate of the ice thickness appropriate to the free-floating strain-rate regime. Calving occurs when the ice thickness falls to 250 m. Although this is a crude but simple form of calving relation, it is found that the major results are not very sensitive to the prescribed calving thickness. The Weertman-type ice-shelf free-thinning rate ($\dot{\epsilon}_z$) can be expressed as a function of ice thickness (Z) as

$$\dot{\epsilon}_z = k(\theta) Z^n, \qquad (3)$$

where n is taken as 3, and $k(\theta)$ is directly related to the ice-deformation properties for the mean column temperature θ through the ice; for example, for $\theta \approx -20°$C, and isotropic ice rheology $k \approx 1.5 \times 10^{-10}\,\mathrm{m^{-3}\,a^{-1}}$. To give an appreciation of the large increases of vertical strain rates ($\dot{\epsilon}_z$) and annual strain thinning rates ($\mathrm{d}Z/\mathrm{d}t$) with increasing thickness, Table 1 provides a few values based on −20°C. For the warming run presented here, the value of k is taken as $k = 3 \times 10^{-10}\,\mathrm{m^{-3}\,a^{-1}}$, based on the laboratory studies of ice-flow properties and the observed thinning rates of ice fronts and outlet glaciers (e.g. Budd and others, 1987; Budd and Jacka, 1989). In addition, a conservative maximum limit on the vertical strain rate has been set at 10% a^{-1} to show that excessive values are not required for the large retreat. This simple model provides an appropriate ocean-margin boundary condition for the grounded ice sheet and represents reasonably well the ice shelves which are free to strain unimpeded by embayment restrictions.

The increasing basal melt rates (up to 5 m a^{-1}) are thus the primary forcing for the retreat, whereas the increase in accumulation at 150% of the present only contributes typically up to about 0.3 m a^{-1} to the rate of ice-shelf thickness change.

Table 1. Free-floating ice-strain rates ($\dot{\epsilon}$) and rates of ice-thickness change ($\mathrm{d}Z/\mathrm{d}t$) for $\theta = -20°C$

Ice thickness	Z(m)	250	500	750	1000
Strain rate	$\dot{\epsilon}$(% a^{-1})	0.24	1.92	6.48	15.3
Ice-thickness change	$\mathrm{d}Z/\mathrm{d}t$(m a^{-1})	0.60	9.60	48.6	153

RESULTS FROM THE SIMULATIONS

A set of perspective views of surface elevation are presented in Figure 1 to show the pattern of retreat of the ice sheet over time. It takes the first 100 years in the transient coupled climate model for the temperature changes to reach the values required for the higher melt rates (5 m a^{-1} in this case). After that the ice shelves thin, calve and retreat within 100–200 years, for quite a wide range of melt rates. The change in net surface accumulation has little effect on that timing. For the results presented here, the model was started with the present grounded-ice-sheet configuration and the climate-change forcing applied immediately.

After the major ice shelves disappear, the high floating-ice strain rates for the thick ice flowing from the ice streams develop and the ice flow from the grounded areas increases. The ice streams thin and the grounding lines retreat, particularly in those regions where the bedrock lies furthest below sea level. Eventually, the saddle region of West Antarctica (approximately 79° S, 115° W) becomes afloat to form a new transient ice shelf. Continued thinning, melting and calving progressively removes the grounded-ice connections between those isolated regions which remain because they have bedrock close enough to sea level at their new grounding lines for the grounded ice to survive. After 2000 years the area of ice cover is already greatly reduced, but the ice thickness is still quite large on these separate grounded areas. The ice thicknesses in these isolated regions decrease rapidly towards their equilibrium values, which are approached more slowly and asymptotically over the next few thousand years. The effect of the increase in accumulation rate is to reduce the rate of lowering of the residual ice caps in West Antarctica, and to increase the ice elevation in central East Antarctica, even though the ice-sheet area decreases. The slow responses to isostatic bedrock adjustment and equilibration to the new accumulation regime mean that a true steady state is not achieved for a much longer period, but these further changes are relatively small.

The changes in ice volume with time for the first 2000 years for the simulation of Figure 1 and also for the case without the additional 50% accumulation are shown in Figure 2. The figure shows separately the changes in total ice and floating ice, and the change in the volume of ice above floating, converted to the contributions to sea-level change. Note that the volume contribution to sea-level change is substantially less than the grounded-ice volume change because of the large fraction of the grounded ice which is below sea level, or rather below the level required for floating. The earlier papers addressed the warming for the period over which the major ice shelves retreated. Two hundred years of the forcing used here brings us to a comparable situation, and the subsequent changes are shown in Figure 2.

Contour maps comparing the surface elevation from the initial state to the configuration after 2000 years for the standard simulation are presented in Figure 3.

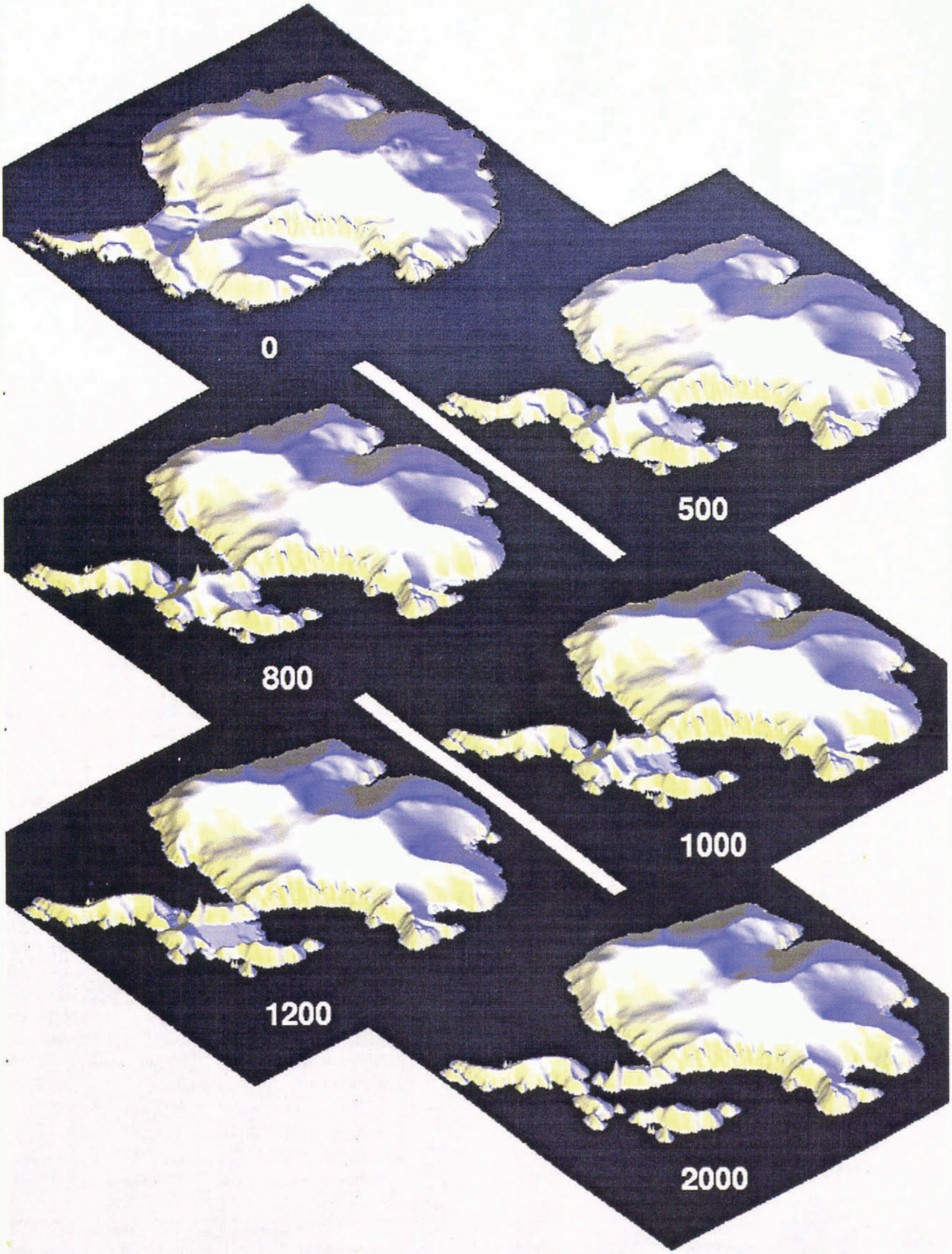

Fig. 1. A time sequence of perspective views of the changing Antarctic ice sheet from the model, from the present to 2000 years after the onset of warming, with 5 m a^{-1} basal melt rate and 50% increase in accumulation rate. The time from the start is shown in years for each frame.

The long-term changes in ice volume and sea level are similar to those presented by Budd and others (1987) except that in the early stages the increase in accumulation over Antarctica (which was not included in the earlier study) decreases the rate of sea-level rise from ice flow to some extent. This is similar in the first 500 years to the results of Budd and others (1994) and O'Farrell and others (1997), showing the dependence on the magnitude of the accumulation rate. The 50% increase in accumulation rate here is equivalent to a lowering of about 3 mm a^{-1} in sea level, which partially

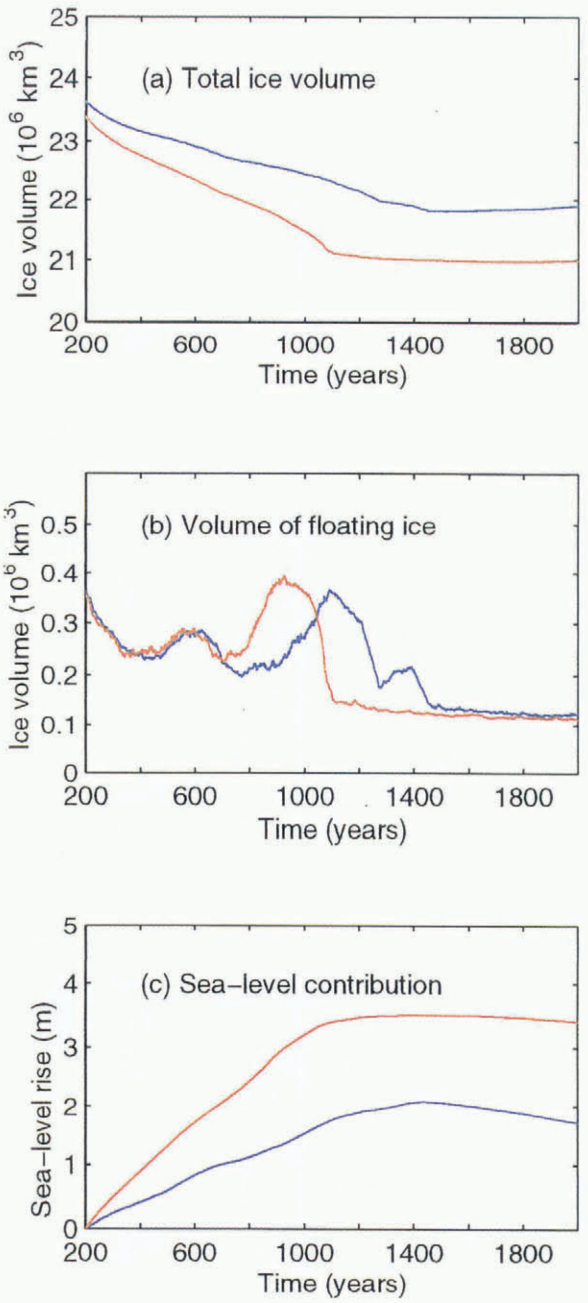

Fig. 2. The progress of changes for the simulation illustrated in Figure 1, shown by the plots of (a) total ice volume, (b) floating-ice volume (both 10^6 km^3) and (c) sea-level contributions from the changes (blue curves). Also shown are the results for the corresponding case without increased accumulation (red curves). The large increase in floating ice occurs when central West Antarctica becomes afloat.

offsets the rise from ice flow. The final configuration is primarily defined by the edges of the bedrock regions reaching more than about 250 m below sea level, where the ice can float, thin from strain and melting, and finally calve. Bedrock isostatic adjustment allows the bedrock to rise slowly and this adds to the longer time required to reach a complete equilibrium, but does not significantly alter the basic picture as shown in Figure 1. (It provides a slightly stabilising effect on the residual grounded ice.) The total contribution to sea-level rise after 2000 years amounts to about 3.4 m for the case without increased accumulation. For the 50% higher accumulation case a net contribution to sea-level rise of 1.8 m is obtained. This is consistent with the cumulative effect of the higher accumulation rate and the decreasing area of grounded ice. Other aspects which would need to be considered in refining these estimates would involve possible further increases in accumulation associated with the altered climate, and the state of dynamic balance in the ice-sheet model. At present, the ice sheet is believed to be close to balance with the Holocene accumulation regime, but as the present model has an ice-shelf treatment which is not designed to describe that period, we start from the present ice-sheet configuration rather than an exact equilibrium condition.

DISCUSSION AND CONCLUSIONS

The present study represents only a preliminary attempt to understand the possible extent of the Antarctic ice sheet under somewhat warmer conditions, such as a Pliocene warm period, an extreme Pleistocene interglacial, or in an environment following an extended period of future warming from increased greenhouse gases. The purpose has been to show the importance of the processes of basal melting and the increased strain rates of unimpeded thick ice in driving the retreat of marine ice sheets. It is apparent that about 3°C of global mean warming, corresponding to 2°C of warming for the water under the ice shelves and resulting in 5 m a^{-1} basal melt rates, is enough to cause the demise of the marine ice regions of West Antarctica and a retreat of coastal ice towards more firmly grounded regions elsewhere, over a time period of about 2000 years. From sensitivity studies it is also apparent that smaller amounts of warming and melt rates (e.g. 3 m a^{-1}) would cause a similar demise over comparable time-scales but with somewhat more floating ice at the early stages. These response times are relatively short in the context of possible Pliocene warm episodes. The minimum amount of warming or increased melt rate required to result in an extensive retreat of the West Antarctic ice sheet has not yet been determined. That would need a much more comprehensive treatment of the ocean warming from the coupled climate model, combined with high-resolution modelling of the water circulation under the ice shelves, coupled to basal melting and freezing, similar to that carried out by Grosfeld and Gerdes (1998) and Williams and others (1998) but extended to the complete Antarctic coastline currently fringed by ice shelves. It would also be necessary to include the feedback of the changing geometry of the ice topography and ocean basins into the coupled climate model and into the high-resolution models for the ocean circulation under the remaining floating ice shelves.

To determine the modified climatic conditions which would prevail for the final ice configuration, a new coupled model run would be required with the imposed warming and including the feedbacks from the changing geometry. We propose to continue research in that vein, but a preliminary indication of the resulting further alteration in regional climate can be made by relating the temperatures over present-day ice and ocean regions to the new ice-sheet configuration. At present, the mean January surface air temperatures in central West Antarctica and also along the Ross Ice Shelf next to the Transantarctic Mountains are in the range −15° to −20°C. With the reduction in sea ice which accompanies

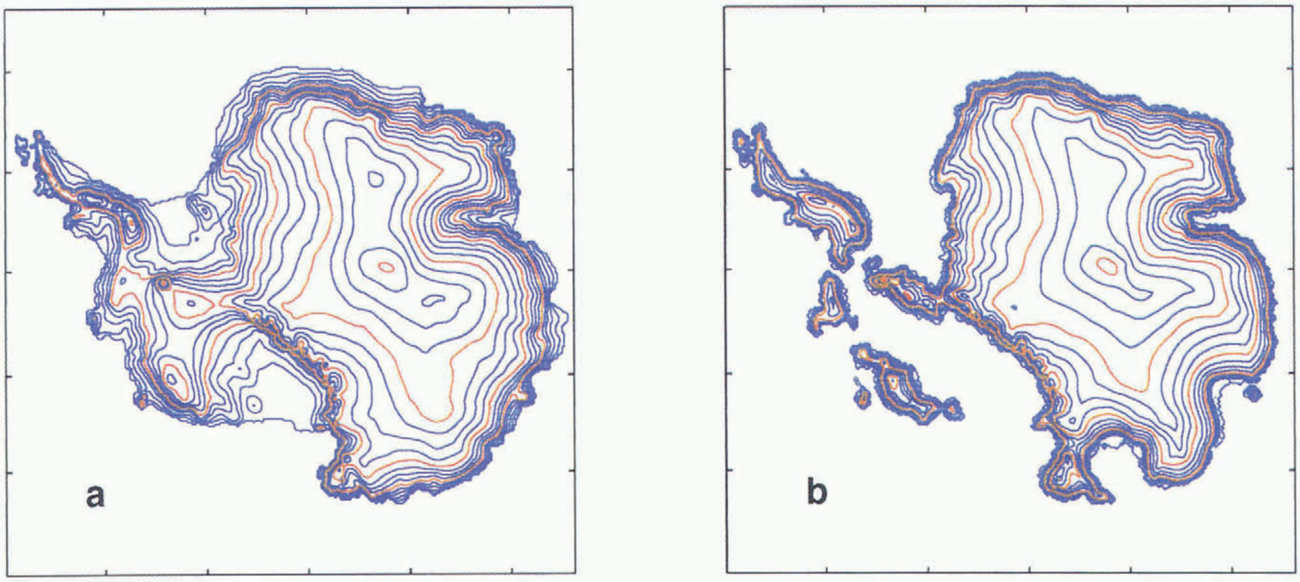

Fig. 3. Elevation contour maps (0, 0.1 then 0.25 km spacing with km contours highlighted), shown for (a) the initial configuration (present) and (b) after 2000 years, for the case of Figure 1.

the warmer scenario (+3°C), and the circulation of warmer water (~+2°C) from further north (e.g. the Ross Sea) through the newly exposed archipelago, it can be expected that surface ocean temperatures over this region would average about +1°C during the summer minimum sea-ice period. This is equivalent to a local regional warming of +16° to +21°C, relative to the present temperatures, resulting from an assumed initial global warming of only 3°C. The ice changes in the Antarctic could also have a small feedback to the global mean change (cf. Simmonds and Budd, 1991).

In summary, it is worth emphasising that the West Antarctic ice sheet is sensitive to relatively small amounts of global warming (of the magnitude expected to result from more than a doubling of CO_2 concentration), provided that the warming is imposed for long enough (of the order of 1000 years). The resultant contribution to sea-level change from the Antarctic ice sheet over that period could be expected to amount to about 2 m over 1000–2000 years. Once the ice shelves disappear, after the first few hundred years, it is likely that the continued retreat would be irreversible until near to the new equilibrium configuration. This is because the high grounding-line strain rates would continue to apply unless decreased by very much lower ice temperatures or restrained by regrowth of thick embayed ice shelves. The time required for these to return is comparable to the time taken for the continued retreat to move essentially to completion.

Consequently, we conclude that the demise of the West Antarctic ice sheet could occur irrevocably from a global warming of a few °C imposed for only a few hundred years. This represents a relatively short period within the context of the Pliocene or even a previous interglacial. For future scenarios of global warming, resulting from increasing greenhouse gases, the results presented here also illustrate the possible longer-term consequences if the warmer conditions continue for several centuries.

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