

RESPONSE OF THE ANTARCTIC ICE SHEET TO A CLIMATIC WARMING: A MODEL STUDY

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ABSTRACT

It is generally believed that the increasing CO₂ content of the atmosphere will lead to a substantial climatic warming in the polar regions. In this study the effect of consequent changes in the ice accumulation rate over the Antarctic Ice Sheet is investigated by means of a numerical ice flow model. In the model runs, temperature increases linearly with time during 100 years, and is kept constant afterwards.

The results indicate that a climatic warming will probably lead to a sea-level lowering of some tens of centimetres in the next centuries. This is because for Antarctic conditions the increase in snow accumulation exceeds the increase in melting. This estimate does *not* take into account the effects of possible surging of parts of the Antarctic Ice Sheet and the response of the Greenland Ice Sheet (which may be quite different).

KEY WORDS Antarctic Ice Sheet, CO₂ effect on cryosphere

INTRODUCTION

Recently, the response of the earth's major ice sheets to a rapid climatic warming has been discussed in several papers. This subject is of wide interest because mankind is probably forcing such a climatic warming by increasing the CO₂ content of the atmosphere. Experiments with general circulation models of the atmosphere suggest that this warming will be largest in the polar regions (e.g. Manabe and Wetherald, 1980). In view of the high vulnerability of human activities to changes in global sea level, the response of the polar ice masses forms a potential threat.

About 90 per cent of the ice mass is stored at Antarctica, and the remainder forms the Greenland Ice Sheet. As far as sea level is concerned, the ice volume of mountain glaciers and the small ice caps in Northern Canada can be disregarded. It is thus not surprising that the Antarctic Ice Sheet has received most attention, although possible changes in the mass balance of the Greenland Ice Sheet have been studied by Ambach (1980) and Stauffer (1981). These studies show that for Greenland the loss of ice due to melting will increase substantially if a climatic warming of a few degrees K occurs.

So far, studies on the response of the Antarctic ice mass to a rapid warming have focused on the question whether inherent instabilities in parts of the Antarctic Ice Sheet exist, and whether they could be released by a rapid climatic warming. These instabilities involve ice surges, and proxy data on sea level provide some evidence that such events indeed occurred in the past. Quasi-periodic surging of parts of the Antarctic Ice Sheets was first suggested by Wilson (1964), and used to construct a theory for the quaternary glacial cycles. The subject has been discussed further by Hollin (1965, 1972) and Hughes (1973, 1975), among others. Budd and McInnes (1975) have shown that large-scale ice surges are physically plausible. Possible implications of ice sheet instability with regard to the CO₂ problem have recently been discussed by Mercer (1978).

A step forward in the quantitative assessment of the possibility that a CO₂ warming may trigger a

major ice surge on West Antarctica has recently been made by Thomas *et al.* (1979). They conclude that increased melting rates of the ice shelves that buttress the West Antarctic Ice Sheet may ultimately lead to the collapse of this sheet, but that this collapse will be rather slow (hundreds of years). Young (1979) and Fastook (1981), dealing with this problem by employing more detailed models, do not provide a conclusive answer about the possibility of surging in the near future. In fact their studies show how difficult it is to model the grounded ice-ice shelf transition in a proper way. It seems that much more detailed modelling of ice streams and ice shelves is necessary to obtain an answer that is somehow reliable.

Much less attention has been paid to the more regular effect of a change in the mass balance of the Antarctic Ice Sheet, probably because a climatic warming of a few degrees K will not yet cause substantial melting of ice. What has to be expected, however, is a significant increase in precipitation rate. In high latitudes, precipitation is essentially restricted by the low air temperature, not by the lack of atmospheric disturbances that force vertical motions. Consequently, a climatic warming will certainly increase the precipitation rate in the polar regions. In the case of Antarctica this means that snow accumulation will be larger over much of the continent, which will cause the Antarctic ice volume to increase. On the other hand, an increase in air temperature may lead to melting in the lower parts of the ice sheet. It is the aim of this study to investigate these effects by carrying out integrations with a numerical model of the Antarctic Ice Sheet.

The model to be employed is a global model in the sense that it does not deal with 'small-scale' phenomena like ice streams, but intends to calculate the overall distribution of ice thickness. In constructing this model the problem of 'steadiness' will show up. We will find that, according to the model, the present Antarctic Ice Sheet is not in a steady state (though not too far off). To estimate the effect of a climatic change, results have to be presented with respect to a run with constant climatic conditions.

The potential effects of a climatic warming will be investigated by first studying the sensitivity of the ice sheet to changes in the ice accumulation rate. After such a general survey we will discuss experiments more specifically directed to assess the effects of a climatic warming. This will be done on the basis of observational data on melting rates at the edge of the Greenland Ice Sheet (Ambach, 1972) and results from the numerical experiments of Manabe and Stouffer (1980) on the effect of increasing CO₂ content of the atmosphere.

THE ICE SHEET MODEL

The model of the Antarctic Ice Sheet to be used in this study is based on the assumption that the ice flows under vertical shear stresses and that a Glen type of flow law holds for the vertically-averaged ice velocity. This approach was used by the author in some paleoclimatic studies (e.g. Oerlemans, 1980), and some characteristics of this type of model are discussed in Oerlemans (1981). Although some differences exist, the structure of the model is basically the same as that used by Mahaffy (1976).

We will not make distinction between internal deformation and basal sliding. As proposed by Nye (1959), we write

$$u = k\tau_b^m \quad (1)$$

where u is the vertical mean horizontal ice velocity, τ_b the shear stress at the bottom of the ice sheet, and k and m are flow constants. Current values of m are in the 2 to 3.5 range (Paterson, 1969); we adopt a value of 2.5. The constant k in fact depends on ice temperature, but will be kept constant in this study. The value of k will be chosen such that the model produces a good simulation of the observed distribution of ice thickness in Antarctica.

Generalizing equation (1) to the two-dimensional case, and setting the basal shear stress proportional to $H\nabla H^*$, where H^* is the surface elevation and H the ice thickness, the vertical mean mass flow vector becomes

$$\mathbf{M} = KH^{m+1}[\nabla H^* \cdot \nabla H^*]^{(m-1)/2} \nabla H^* \quad (2)$$

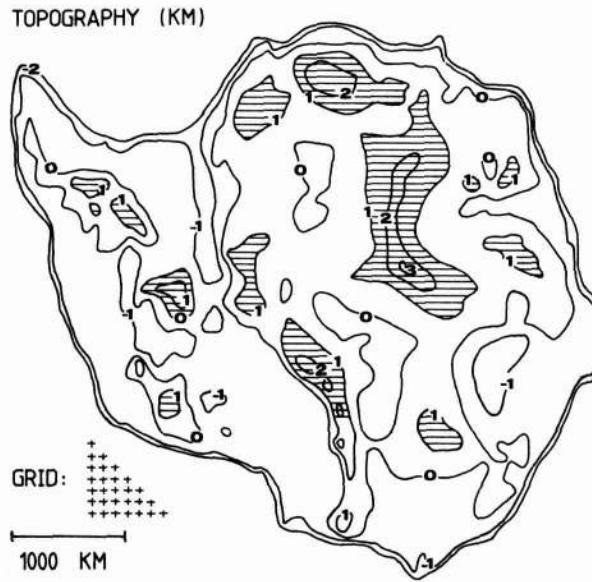


Figure 1. Bedrock elevation of the Antarctic continent, relative to sea level. The lowest contour is -2 km. Regions with an elevation over 1 km are shaded

If we denote the mass balance (or annual ice accumulation rate) by G , the conservation of ice mass is given by

$$\partial H/\partial t = \nabla \cdot \mathbf{M} + G \quad (3)$$

Equations (2) and (3), together with a suitable boundary condition, completely formulate the model. The equations are solved on a 100×100 km grid using central differencing in space and forward differencing in time. Bedrock topography is taken from Atlas Antarktiki (1966); it is shown in Figure 1. In all computations a time step of 10 yr is used, which renders the scheme stable for all situations of practical interest.

When a continental ice sheet reaches the steep edge of the continental shelf, a suitable boundary condition for a model simulating this sheet is to set $H = 0$ at this edge. This means that at that location the ocean acts as an infinite sink for ice, or, in other words, that the production of icebergs precisely equals the ice mass discharge from the continent. However, if the continental shelf is considerably below sea level (500 m, say), or if the continent-ocean transition is more gradual, such a boundary condition puts too heavy constraints to the behaviour of the model ice sheet (its response to a change in climatic conditions would be restricted to a change in ice thickness; expansion or shrinkage would not be possible). Thus, because large parts of the West Antarctic Ice Sheet are bounded by continental shelf *not* covered by grounded ice, it is obvious that a dynamic boundary condition is needed. The simplest way in which this problem can be treated is to introduce a floating ice condition, i.e. to remove the ice as soon as it starts to float. Unfortunately, because the ice-sheet edge is not resolved very well on the 100 km grid this still does not allow the sheet to expand. This can be understood as follows.

In Figure 2 an ice-sheet edge is sketched. On the model grid, the edge is represented by the ice thickness on the grid points: H_1 , H_2 and $H_3 (= 0)$; the index refers to the grid point. If $\partial H/\partial t$ at point 3 is positive, H_3 will also be positive. However, with a typical value of $\partial H/\partial t$ of 1 m/yr, H_3 can reach only 10 m within one time step. In most cases, this will not be enough to prevent floating and therewith loss of the ice. In reality, an ice sheet expands by pushing its snout forward, not by putting a thin layer of ice in front of the edge (as the model in fact does). So in the case of a bedrock below sea level, the model formulation fails to describe the movement of the edge in a proper way.

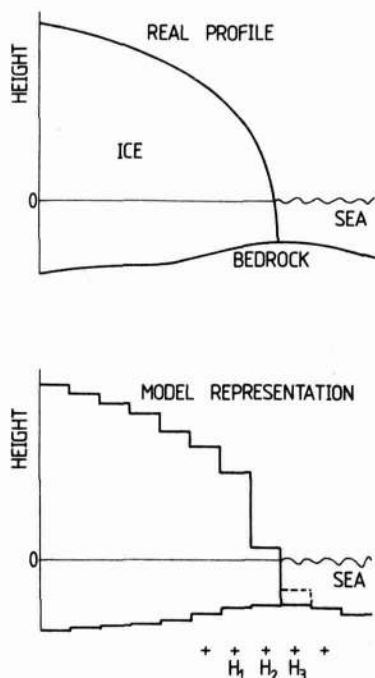


Figure 2. Model representation of the ice edge. In order to allow a smooth movement of the edge, arrangements in the numerical scheme have to be made (see text)

In order to solve this problem, the grid points in the vicinity of the grounding line are treated in a special way. First, we define an 'edge point' as a grid point adjacent to which both grid points with and without grounded ice occur. When such an edge point is encountered in the program, a check is made whether the ice thickness at this point (H_e) is larger than some prescribed factor times the average ice thickness (\bar{H}) of the surrounding points (having grounded ice). So

$$H_e > \alpha(\sum H)/L \quad (4)$$

where \sum denotes summing over the eight surrounding grid points and L is the number of these points with grounded ice. If condition (4) is not satisfied, a redistribution of ice over the area represented by the grid point is carried out. The ice mass in the grid point (being H_e times the area of the grid point) is redistributed in such a way that the thickness of the snout equals $\alpha\bar{H}$, where α measures the (prescribed) steepness of the edge. The floating ice condition is now applied to $\alpha\bar{H}$ instead of H_e , and the movement of the grounding line is traced 'within the grid point'. This procedure leads to a much better model representation of the ice-sheet edge.

To find a value for α , we consider a perfectly plastic ice sheet. Its profile is given by $H = \sigma(x)^{1/3}$ (e.g. Paterson, 1969), where σ is a constant and the edge is at $x = 0$. If we use two grid points to represent the edge, we should choose $x_1 = 50$ km and $x_2 = 150$ km. This immediately yields $H_1/H_2 = (1/3)^{1/3} = 0.577$. This value can be considered as a reasonable estimate for α . It is noteworthy that α does not depend on the flow constant σ . It should be viewed upon as a model constant determined by the resolution of the grid.

We finally note that the fraction of a grid-point area (10^4 km²) F_e covered by grounded ice is given by

$$F_e = H_e/(\alpha\bar{H}). \quad (5)$$

THE DISTRIBUTION OF SNOW ACCUMULATION

Snow accumulation varies widely over the Antarctic continent. It ranges from about 0.05 mie (metres of ice equivalent) per year in the central part of East Antarctica to over 0.6 mie/yr in some coastal regions. The present-day distribution is shown in Figure 3(a) (from Atlas Antarktiki, 1966).

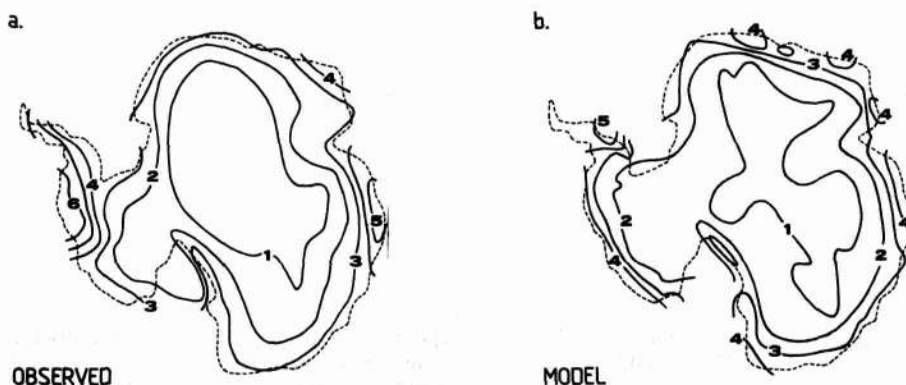


Figure 3. A comparison between observed (a) and modelled (b) precipitation rates. Unit: 0.1 m of ice equivalent/yr

The differences in accumulation rate over Antarctica are strongly related to surface conditions. In particular it depends on the surface slope (upslope precipitation), surface elevation (low temperatures of the air, implying a very small amount of precipitable water), and distance to the moisture source (the ocean). If one allows the model ice sheet to shrink or expand, it would be inconsistent to keep the accumulation distribution fixed. The model should have the freedom to shift the zones of high accumulation rates with the ice sheet edge. To meet this requirement, the precipitation was modelled in terms of the three environmental parameters mentioned above.

The best fit to the observed distribution of precipitation was obtained by setting

$$P = \max \{ (0.3 + 14S - 4 * 10^{-5} H^*) / C; 0.05 \} \text{ mie/yr.} \quad (6)$$

Here, P is the precipitation, S the surface slope, H^* the surface elevation and C a continentality factor. The latter is defined as

$$C = (0.5 + 0.5 N_R)^{-1}, \quad (7)$$

where N_R is the fraction of grid points with $H = 0$ within a circle of radius R around the point for which the precipitation rate is computed. Defined in this way, the continentality factor would be 1 for an island of one grid point, about 1.5 at the edge of the Antarctic continent and 2 in the interior of the continent. In all computations discussed in this paper $R = 500$ km. Since in the integrations the distribution of grounded ice varies slowly, C is computed only every tenth time step.

The precipitation field as computed for the present distribution of ice thickness is shown in Figure 3(b). In general, good agreement exists between the observed and calculated fields. The largest differences are found over the lower parts of the West Antarctic Ice Sheet and the Antarctic Peninsula. This is due to local meteorological conditions and is difficult to model. However, for the present purpose the results of the precipitation model are quite acceptable. No attempt was made to introduce a correction field because the results of the climatic change experiments have to be compared to a control run anyway (no matter whether the observed or computed accumulation field is used, according to the model the present ice sheet is not in a steady state).

THE CONTROL EXPERIMENT

Before any sensitivity studies can be done, a control experiment has to be defined. This is necessary because the flow constant K in equation (2) cannot be chosen such that the model produces a steady state ice sheet within 10,000 yr of integration (although convergence to steady state is evident). The reason for this may be shortcomings in the description of ice mechanics used here, or a real transient state of the Antarctic Ice Sheet—it probably is a combination of both.

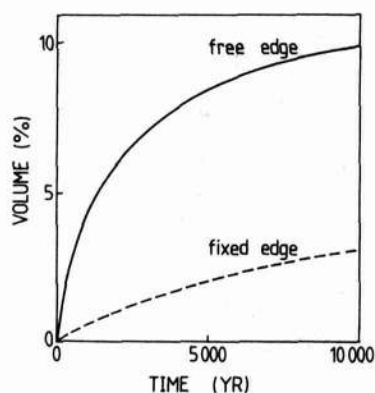


Figure 4. The control experiment. Starting with the present distribution of ice as initial condition, integration in time yields increasing ice volume (expressed as percentage of the present-day Antarctic ice volume, $2.3 \times 10^{16} \text{ m}^3$)

There are several indications that the West Antarctic Ice Sheet is more sensitive to environmental changes than the East Antarctic Ice Sheet (Hughes, 1975; Mercer, 1978). It is also well-accepted that during the last glacial maximum the ice volume of the West Antarctic Ice Sheet was considerably larger than it is now (e.g. Thomas and Bentley, 1978). In view of this, K was chosen such that the distribution of ice thickness over the eastern Antarctic continent is well-reproduced by the model. The corresponding value of K is $0.5 \text{ m}^{-3} \text{ yr}^{-1}$.

Starting with the present distribution of ice, the volume of the model ice sheet increases (Figure 4). After 10,000 yr the ice sheet is not yet in a steady state, although by this time the growth rate has slowed down considerably. For comparison, a run was made in which the edge of the ice sheet was fixed to its present location. In this case the volume also increases, but at a lower rate. Apparently, the increase in ice mass is due to an increase of the mean ice thickness *and* an expansion of the ice sheet. Figure 5 shows this more clearly. The distribution of ice thickness after 2000 yr of integration differs from the initial state in a notable way.

First, the ice cover over the Antarctic Peninsula expands. This may be partly due to the poor resolution of the Peninsula's geometry on the $100 \times 100 \text{ km}$ grid, but some evidence exists (Sugden and Clapperton, 1980) that this part of the Antarctic Ice Sheet is indeed growing. Second, there is a general tendency for the ice sheet to become steeper. This is at least partly due to systematic errors made in transforming a distribution of ice thickness from a map to a grid with poor resolution (as far as ice-sheet edges are concerned). The transition from grounded ice to ice shelf is not always clear: a grid square partly covered by grounded ice and partly covered by floating ice is in most cases classified as belonging to the ice sheet and assigned a small representative ice thickness. This obviously tends to make the edges of the 'grid ice sheet' less steep. In view of this, the ice sheet edge produced by the model after some time is probably more sound in a physical sense than the initial sheet.

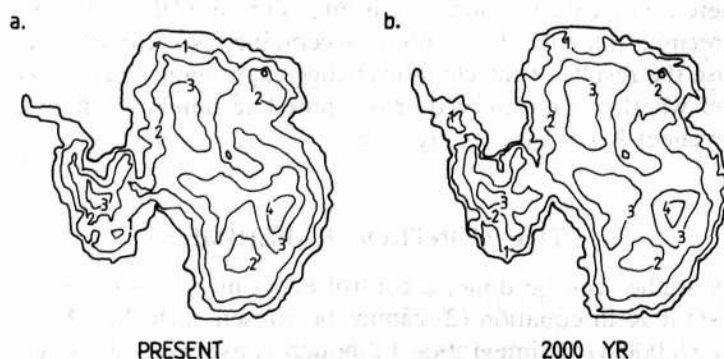


Figure 5. Distribution of ice thickness (km) as observed (a) and produced by the model (b) after 2000 yr real time simulation. Note the expansion of the ice sheet over the Antarctic Peninsula

In the climatic change experiments discussed later, most integrations are started with the ice distribution shown in Figure 5(b). This was done because the model has created an ice sheet in which unrealistically large gradients in ice velocity do not appear. Such gradients appear in the first years of the integration because the natural compensation between surface slope and ice thickness is distorted by observational errors. This may give rise to very large values of M in equation (2). Also, after 2000 yr the ice sheet does not expand anymore. This in fact demonstrates that the present model is capable of simulating the overall ice distribution of the Antarctic Ice Sheet, though it converges to a steady state in which the ice volume is larger than observed. The implication of this cannot be judged, however, because it is unknown whether the present ice sheet is close to a steady state.

In the following sections, the results of model runs will be presented with respect to the control run, defined here as the run starting with the ice distribution of Figure 5(b) without any external change in environmental conditions.

GLOBAL STABILITY TO CHANGES IN PRECIPITATION

Without having in mind any particular climatic change experiment, one can vary the accumulation rate G over the Antarctic continent to study the global stability of the ice sheet. Of course, one expects an increasing ice volume for increasing G (or P , because melting is not explicitly considered at this stage) and vice versa, but it is very likely that 'thresholds' exist in the response of the ice sheet. Such thresholds should, in principle, be due to the irregular topography of the bedrock (at least with the ice mechanics included in the present model). For example, it could be that a critical increase of G exists for which the shallow Ross Sea would be captured by the grounded ice.

In order to see whether such events show up in the model, five integrations were carried out in which G was changed uniformly by +100, +60, +20, -20 and -60 per cent. The ice-volume curves (with respect to the control run) of these integrations are shown in Figure 6. These curves reveal that the model ice sheet responds in a very regular way to the changes in G , at least during the first 1000 yr. Apparently, no thresholds are crossed.

The response of the ice mass to a climatic warming may be more complicated than Figure 6 would suggest because the corresponding change in G will vary from place to place. We turn to this in the next section.

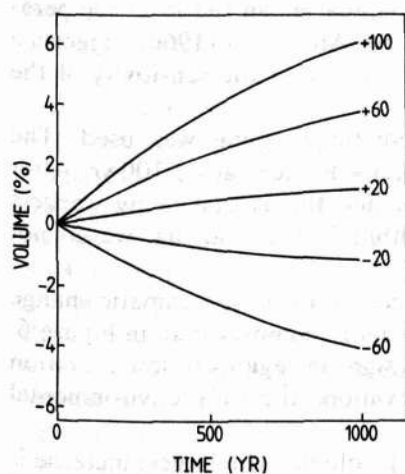


Figure 6. Response of the Antarctic ice volume to changes in the ice accumulation rate. The labels of the curves denote the (uniform) change in P in per cent

THE EFFECT OF A CLIMATIC WARMING

As argued in the Introduction, a temperature increase will cause an increase in precipitation rate because in the polar regions the maximum possible water content of the air determines, to a large

extent, the amount of precipitation. The best guidance for performing a climatic warming experiment, related to the CO₂ problem, is obtained from experiments with general circulation models of the atmosphere and ocean. Manabe and Wetherald (1980), using the GFDL model combined with a simple ocean model, found that for a doubling of the CO₂ content in the atmosphere surface temperatures in the polar regions could be raised by 8 K and precipitation rates by 30 per cent. In a more recent experiment with realistic geometry (the former study used a schematic representation of the ocean-land distribution) Manabe and Stouffer (1980) found that for the Antarctic continent the temperature increase would be about 3 K and the precipitation increase about 12 per cent (annual mean values). We will use these values as a guide in the climatic warming experiment.

If temperatures go up, melting of ice and snow may occur in the lower regions of the Antarctic Ice Sheet (at present melting is negligible, see Schwerdtfeger, 1970). To estimate the importance of this effect, measurements of ice and snow melt carried out at the edge of the Greenland Ice Sheet by Ambach (1972) were used. Conditions at that site are probably comparable to the conditions to be expected at the edge of the Antarctic Ice Sheet in the case of a substantial warming. Ambach's data suggest that melting rates can be computed by summing up positive daily mean temperatures. Assuming that the average surface albedo during the melting period is halfway between that of fresh snow and pure ice, it appears that 5 day-degrees melt approximately 2 cm water equivalent of ice. It is thus possible to estimate the annual melt if the annual mean temperature and the seasonal temperature variation (assumed to be sinusoidal) are known.

According to Atlas Antarktiki (1966), the annual temperature range in the coastal regions of the Antarctic continent is about 20 K, and in the lowest 1500 m (which is of interest here) fairly constant. So annual snow/ice melt can be expressed in annual mean surface temperature. The result is (for temperatures that are not too high):

$$AM = \begin{cases} (12 + T_s)^3 & \text{if } -12 < T_s < -5^\circ\text{C} \\ 0 & \text{if } T_s \leq -12^\circ\text{C} \end{cases} \quad (8)$$

Here, the annual melt is expressed in cm water equivalent of ice per year. In the model, the computation of the melting rate starts from the prescribed value of the annual sea-level temperature around the Antarctic coast. Surface temperatures are then computed from the surface elevation by using a lapse rate of 9 K/km. This value is suggested by the temperature maps in Atlas Antarktiki (1966).

The procedure described above requires knowledge of the present annual mean sea-level temperature around Antarctica (T_0). This quantity is not accurately known. Atlas Antarktiki (1966) suggests a value of -13°C , but it could also be -12°C or -14°C . At the end of this paper the sensitivity of the results to the choice of T_0 will be discussed.

In all forthcoming experiments the following scenario for a climatic warming was used. The temperature and precipitation rate increase linearly from zero to prescribed values at $t = 100$ yr. After 100 yr, temperature and precipitation are kept constant. The idea behind this is that many workers believe that a doubling of CO₂ in the atmosphere will take place within 100 yr, and that we do not know what will happen afterwards.

Figure 7 shows the results of four experiments differing only in the magnitude of the climatic change imposed. It is immediately obvious that the response of the ice sheet is more complex than in Figure 6. This is because the change in G varies substantially and even changes sign: in regions of low elevation melting exceeds the increase in precipitation whereas at higher elevations the only environmental change is the increase in snow accumulation.

In all runs shown the imposed climatic warming leads to an increasing volume. The largest increase is found for a small rise in T_s and a large increase in P , of course. But even with $\Delta T_s = +6$ K and $\Delta P = +12$ per cent the ice sheet ultimately becomes larger. Extended runs showed that it takes 10,000–20,000 yr before the change in volume settles down to a constant value.

In the experiments with a large temperature increase (dashed lines in Figure 7) an interesting feature occurs: after the initial increase in ice volume, and the subsequent decrease when the temperature rise

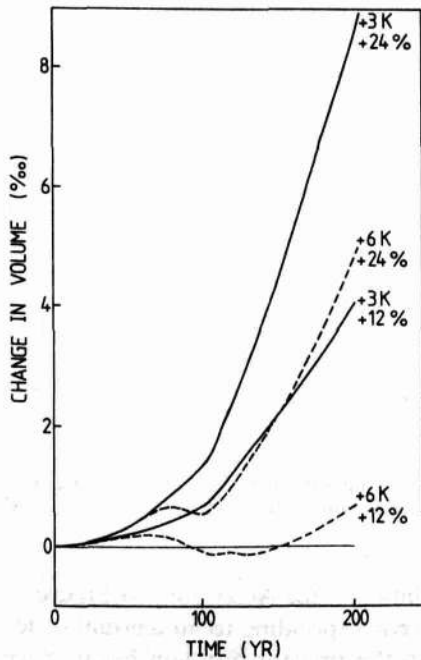


Figure 7. The climatic warming experiment. The labels denote the temperature change and the increase in precipitation rate. The (+3 K, +12%) experiment mimics the effect of a CO_2 doubling according to Manabe and Stouffer (1980)

becomes large, the volume starts to increase again. This phenomenon is due to the dynamic response of the ice sheet. If melting at the edge becomes large while at some distance from the ice-sheet edge the mass balance increases, the surface slope is no longer in equilibrium with the mass-balance field. As a consequence, the discharge of ice mass towards the edge increases, which compensates the lowering of the ice surface at the edge due to melting. So the temporary decrease in ice volume is due to the fact that the ice sheet needs some time to react to the increased gradient in the mass balance.

SENSITIVITY OF THE RESULTS

The experiments shown in Figure 7 suggest that the Antarctic ice volume will grow if the climate becomes warmer, but it is very important to know how sensitive these results are to changes in the model parameters. The most critical parameter is T_0 , the present-day annual sea-level temperature. To see the dependence on T_0 one of the preceding experiments ($\Delta T_s = +3$ K, $\Delta P = 12$ per cent) was repeated with $T_0 = -14^\circ\text{C}$ and $T_0 = -12^\circ\text{C}$. The resulting ice volume curves are shown in Figure 8. Apparently, the differences are not very large.

Another aspect is the dependence of the results on the initial state. This point was investigated by repeating the climatic change experiments with other initial states. Five states were used for this purpose, namely, states from the standard run (Figure 4, solid line) at $t = 0, 2000, 4000, 6000$ and 8000 yr (note that this also implies the use of different control runs in computing the changes in ice volume). The results of this test are shown in Figure 8 by vertical bars, indicating the range over which the change in ice volume varies. The dependence on the initial state increases with increasing T_0 . This is not surprising, because for higher values of T_0 melting is more important. Melting only occurs at low elevation, and the initial states differ mainly in the area of regions with low elevation. Nevertheless, the dependence of the model results on the initial state is rather small. The uncertainty due to other factors (T_0 , magnitude of the change in environmental conditions) is much larger.

DISCUSSION

According to the numerical experiments presented in this paper, the volume of the Antarctic Ice Sheet would probably increase if a climatic warming of a few degrees takes place. Once again it should be

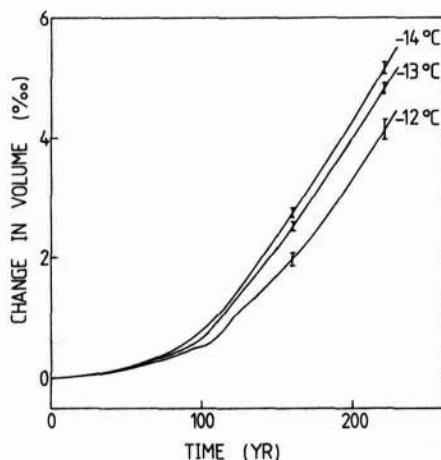


Figure 8. Sensitivity of the model results. The labels denote the present annual sea-level temperature on the Antarctic coast, used in the computation of melting. The vertical bars measure the dependence on initial conditions. All results refer to the (+3 K, +12%) experiment

stressed that this *only* concerns the effect of changes in the accumulation rate. According to Figure 8, the increase in ice volume has an order of magnitude of 0.5 per cent (corresponding to an amount of ice of 10^{14} m³) after 250 yr. Longer integrations are not very relevant to the present problem because we do not have any idea about the climatic conditions to be expected.

The importance of changes in the Antarctic ice volume is that they directly affect world sea level. It is not the intention to present an overview of how sea level would react to a climatic warming, but on the basis of the present results one aspect can be discussed. In the absence of any bedrock uplift or sinking in the Antarctic region, the change in world mean sea level (MSL) is given by

$$\Delta \text{MSL} = (-\rho_{\text{ice}} \Delta V_+ - \Delta V_-) / A. \quad (9)$$

In equation (9), ΔV_+ is the increase in ice mass above sea level, ΔV_- that below sea level, and A the area covered by the world ocean.

The change in ice volume (induced by a climatic warming) occurring in the present model is almost completely associated with the ΔV_+ term. The ΔV_- term is only a few per cent of the ΔV_+ term, indicating that expansion of the model ice sheet plays a minor rôle. So the drop in sea level associated with the climatic warming experiment can be inferred immediately from the total increase in ice volume (Figure 7). A simple calculation shows that a 1 per cent increase in ice volume corresponds to a drop in sea level of about 0.6 m. For the CO₂ scenario and associated climatic warming used in this study, the 'Antarctic mass balance effect' would lead to a sea-level drop of about 0.3 m in the next 200 yr.

This estimate is subject to a number of uncertainties. First of all, opinions about the effect of an increasing CO₂ content of the atmosphere differ widely. It is generally accepted that a warming has to be expected, but its magnitude is subject to much debate. Even if we did know the effect of CO₂ on climate, things would not be much clearer. The global carbon cycle is poorly understood and we are not yet able to predict (given the antropogenic CO₂ input) the CO₂ content of the atmosphere for the next centuries.

As stated earlier, this study only concerns the 'mass balance effect'. Inherent instabilities of the Antarctic Ice Sheet that could be triggered by a climatic warming are not considered. An obviously useful extension of the present model, allowing us to study such phenomena as ice surges, is to include thermodynamics of ice flow. In addition to this, for somewhat longer time integrations, movement of the bedrock in response to the varying ice load should be calculated. Such improvements are expected to be incorporated in future work.

It should be realized that the 'Antarctic mass balance effect' is only one out of many mechanisms that determine the response of sea level to a climatic warming. Expansion of water due to increasing

temperature, changes in atmospheric and oceanic circulation patterns, etc. are not negligible (see Lisitzin, 1974).

Finally, the steady-state problem remains. According to the present study the Antarctic Ice Sheet is not in equilibrium, but grows at a rate that is larger than the growth induced by the climatic warming. Present-day observations do not yet provide a reliable answer about the departure from equilibrium of the present Antarctic and Greenland Ice Sheets. It is obvious that more detailed modelling can help to solve this problem.

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