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On the response of the Greenland ice sheet to greenhouse climate change

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Abstract

Numerical computations are performed with the three-dimensional polythermal ice-sheet model SICOPOLIS in order to investigate the possible impact of a greenhouse-gas-induced climate change on the Greenland ice sheet. The assumed increase of the mean annual air temperature above the ice covers a range from $\Delta T = 1^\circ\text{C}$ to 12°C , and several parameterizations for the snowfall and the surface melting are considered. The simulated shrinking of the ice sheet is a smooth function of the temperature rise, indications for the existence of critical thresholds of the climate input are not found. Within 1000 model years, the ice-volume decrease is limited to 10% of the present volume for $\Delta T \leq 3^\circ\text{C}$, whereas the most extreme scenario, $\Delta T = 12^\circ\text{C}$, leads to an almost entire disintegration, which corresponds to a sea-level equivalent of 7 m. The different snowfall and melting parameterizations yield an uncertainty range of up to 20% of the present ice volume after 1000 model years.

1 Introduction

A major component of Global Change is the greenhouse-gas-induced climate change owing to the accumulation of several trace gases in the atmosphere caused by mankind (“greenhouse gases”: water vapour, carbon dioxide, methane, CFC). The direct consequence is an increase of the earth’s global mean surface temperature, depending on the future emission of greenhouse gases and on the sensitivity of the climate system. The latter is usually described by the steady-state temperature increase under the conditions of a doubling of the atmospheric CO_2 concentration, $\Delta T_{2\times\text{CO}_2}$. According to the Intergovernmental Panel on Climate Change (IPCC, 1996), an estimate for this value is $1.9^\circ\text{C} \leq \Delta T_{2\times\text{CO}_2} \leq 5.4^\circ\text{C}$, the rather large uncertainty being due to the effect of multiple positive and negative feedbacks. The impact on other climate parameters such as the cloud cover and the precipitation is equally uncertain.

It is generally believed that this temperature increase will be even more pronounced in polar regions, mainly due to the positive feedback of reduced sea-ice extent and, for the northern hemisphere, reduced snow cover on the ice-free land masses, and consequently reduced albedo. This entails the risk of partial melting of the present large ice sheets on time scales of 10^2 - 10^4 years, accompanied by a sea-level rise to which Antarctica would contribute with approximately 65 m and Greenland with 7 m in case of entire disintegration (Schönwiese, 1992). In order to shed light on this problem, here the 3-d ice-sheet model SICOPOLIS (see below) is applied to the Greenland ice sheet, and simulations are presented with various air-temperature scenarios, snowfall scenarios and melting scenarios.

An important question is whether the response of ice sheets is a smooth function of climate change, or whether critical thresholds exist beyond which a qualitatively different behaviour of ice sheets can be expected. In this context three processes have been discussed, namely (a) large-scale surging triggered by a lubricating layer of basal sediment between ice and bedrock (MacAyeal, 1992), (b) instability of the grounding line (which separates grounded from floating ice) of marine ice sheets (Thomas, 1979), and (c) creep instability due to the reduced ice viscosity with increasing temperatures (Clarke et al., 1977). As far as it is known, the prerequisites for large-scale surging and grounding-line instability are only fulfilled at present by the West Antarctic ice sheet with a potential sea-level rise of 5 m; the Greenland ice sheet may, if any, undergo creep instability.

Dynamic simulations concerning the response of the Greenland ice sheet to greenhouse-gas-induced climate change have already been discussed by e. g. Huybrechts et al. (1991), Letréguilly et al. (1991), Fabré et al. (1995), de Wolde et al. (1997) and van de Wal and Oerlemans (1997). The problem is further assessed by the current EISMINT (“European Ice Sheet Modelling INitiative”) model intercomparison project [for a description of the simulations cf. Ritz (1997), publication of results in preparation], where several models are applied to one standard greenhouse scenario, and emphasis is put on model differences. Here it is investigated how the Greenland ice sheet reacts on a variety of possible greenhouse scenarios, and the influence of uncertainties associated with changes in snowfall and surface melting (the most relevant processes for short-term ice-sheet dynamics) is considered.

2 The ice-sheet model SICOPOLIS

SICOPOLIS (“SIMulation CODE for POLythermal Ice Sheets”) is a three-dimensional dynamic/thermodynamic ice-sheet model based on the continuum-mechanical theory of polythermal ice masses (Fowler and Larson, 1978; Hutter, 1982, 1993; Greve, 1997a). It simulates the time-dependent extent, thickness, velocity, temperature, water-content and age for any specified grounded ice sheet as a response to external forcing. External forcing is given by (i) surface temperature, (ii) surface mass balance (snowfall, surface melting), (iii) sea level surrounding the ice sheet and (iv) geothermal heat flux from below. The model is discussed in greater detail by Greve (1997b). If not mentioned otherwise below, the settings (physical parameters, Glen’s flow law, Arrhenius-type rate factor complemented by a linear dependence on the water content, no-slip conditions on a cold ice base, Weertman-type hard-bedrock basal sliding on a temperate ice base) are those of Greve et al. (1998).

In this study, SICOPOLIS is applied to the Greenland ice sheet, and a series of greenhouse-warming simulations from the present ($t = 0$) to 1000 years into the future

($t = 1000$ yr) are discussed. The simulations are driven by surface-temperature increases from $+1^\circ\text{C}$ to $+12^\circ\text{C}$, with the sea level kept constant. A variety of snowfall and melting scenarios are applied, and the geothermal heat flux is kept constant at $Q_{\text{geoth}}^\perp = 65 \text{ mW/m}^2$. This value is significantly larger than values used by other authors for Greenland simulations (Huybrechts et al., 1991; Letréguilly et al., 1991; Huybrechts, 1994; Fabré et al., 1995; Calov and Hutter, 1996), but provides optimum agreement for the present ice sheet resulting from a palaeoclimatic simulation (Greve, 1997c; see below). Horizontal grid spacing is 20 km. The vertical resolution is 51 grid points in the cold-ice region (temperature below pressure melting), 11 grid points in the temperate-ice region (temperature at pressure melting), if existing, and 11 grid points in the lithosphere.

A previously simulated present state of the Greenland ice sheet is applied as initial condition for the greenhouse simulations. It is the result of a palaeoclimatic simulation (Greve, 1997c) which was driven by a surface-temperature reconstruction of the GRIP core drawn in central Greenland (Johnsen et al., 1995), and which covers 250000 years of climate history, or two entire glacial-interglacial cycles. The agreement between this simulated ice sheet and the real ice sheet proved to be very good (total ice volume 6.4% or circa 40 cm sea-level equivalent too large), so that its use as an initial state for the greenhouse simulations is well justified.

3 Simulations on greenhouse warming

3.1 Control runs

The surface temperature (more precisely: mean annual air temperature above the ice), T_{ma} , is assumed to be decomposed into the present spatial pattern and a purely time-dependent deviation:

$$T_{\text{ma}}(x, y, z, t) = T_{\text{ma}}^{\text{today}}(x, y, z) + \Delta T_{\text{ma}}(t), \quad (1)$$

where x, y, z is a Cartesian coordinate system, in which x and y span the horizontal plane, and z is the direction anti-parallel to gravity. The present surface temperature, $T_{\text{ma}}^{\text{today}}$, was parameterized by Reeh (1991) based on data by Ohmura (1987). Greenhouse scenarios for the surface-temperature deviation, ΔT_{ma} , are prescribed by a delayed increase with the asymptotic value ΔT_{ma}^0 and the delay time τ_{gw} :

$$\Delta T_{\text{ma}}(t) = \Delta T_{\text{ma}}^0 \left(1 - e^{-t/\tau_{\text{gw}}}\right), \quad \tau_{\text{gw}} = 100 \text{ yr}. \quad (2)$$

The choice for τ_{gw} reflects the expected time scale for a greenhouse-gas-induced climate change; however, its exact value is of minor importance for the simulated ice-sheet response.

The snowfall rate, S , is kept at its present distribution, S^{today} , which was gridded by Greve et al. (1998) based on data by Ohmura and Reeh (1991), Bolzan and Strobel (1994), Wilhelms (1996):

$$S(x, y, t) = S^{\text{today}}(x, y). \quad (3)$$

This is referred to as scenario **const_S** (cf. Letréguilly et al., 1991; Fabré et al., 1995). The reason for the use of additional data to the classical compilation of Ohmura and Reeh (1991) is the larger density of measurements in the Summit region (Summit: highest point of the ice sheet at $72^\circ 34' \text{ N}$, $37^\circ 38' \text{ W}$) and the improved representation of the low-accumulation area northeast of Summit. It is not accounted for the fact that in warmer temperatures an increasing fraction of the total precipitation will fall as rain rather than snow. The error induced by this is small, however, because under such conditions surface melting (see below) outweighs snowfall by far.

For the melting rate at the surface, a degree-day model with different degree-day factors for the melting of snow, β_{snow} , and for the melting of ice, β_{ice} is applied (scenario **control_M**; cf. Braithwaite and Olesen, 1989; Reeh, 1991; Braithwaite, 1995):

$$\begin{aligned} \beta_{\text{snow}} &= 3 \frac{\text{mm WE}}{\text{d-K}}, & \beta_{\text{ice}}(\phi = 60^\circ \text{N}) &= 7 \frac{\text{mm WE}}{\text{d-K}}, \\ \beta_{\text{ice}}(\phi = 80^\circ \text{N}) &= 10 \frac{\text{mm WE}}{\text{d-K}}, \end{aligned} \quad (4)$$

with linear interpolation for other latitudes ϕ .

The latitude-dependent ice-melt factor was introduced following Braithwaite (1995). In this paper ablation on the Greenland ice sheet was simulated by more sophisticated energy-balance modelling and then related to the degree-day approach. It was found that a constant degree-day factor underestimates the runoff in the colder parts of the ice sheet, so that a northward increase appears more appropriate.

The response of the lithosphere on the varying ice load is treated by a local-lithosphere-relaxing-asthenosphere model (Le Meur and Huybrechts, 1996), in which the asthenospheric time lag τ_V is the only parameter. Following these authors, it is chosen as $\tau_V = 3000 \text{ yr}$.

With the above specifications, twelve reference simulations ggh01, ggh02, ..., ggh12 (“control runs”) are carried out, the temperature increase, ΔT_{ma}^0 , being varied from $+1^\circ \text{C}$ to $+12^\circ \text{C}$ in steps of 1°C .

Fig. 1 shows the time evolution of the mean annual air temperature, ΔT_{ma} , the total ice volume, V_{tot} , the sea-level equivalent, H_{sl} , the freshwater discharge due to melting and calving, Q_{fw} , the ice-covered area, A_{ib} , the maximum surface elevation, h_{max} , and the volume of temperate ice, V_{temp} , for runs ggh03, ggh06, ggh10. Obviously, the impact of

the $+3^{\circ}\text{C}$ temperature rise (ggh03) is rather small. After 1000 model years, the ice-volume decrease is 10.2%, corresponding to a sea-level equivalent of 72 cm. By contrast, the disintegration is distinctly more pronounced for runs ggh06 ($+6^{\circ}\text{C}$ temperature rise) and ggh10 ($+10^{\circ}\text{C}$ temperature rise): for these simulations, the molten ice volume is 34.4% ($+6^{\circ}\text{C}$) and as large as 83.5% ($+10^{\circ}\text{C}$), corresponding to sea-level equivalents of 2.4 m ($+6^{\circ}\text{C}$) and 5.8 m ($+10^{\circ}\text{C}$).

It is striking that these retreats are virtually not accompanied by changes in the maximum surface elevation. Even in run ggh10 ($+10^{\circ}\text{C}$), h_{max} changes merely from 3267 m at $t = 0$ to 2950 m at $t = 1000\text{yr}$. This is so because the eastern part of the ice sheet situated between approximately 70°N and 76°N is very stable as a consequence of the high bedrock elevation in this region [see Fig. 2, which depicts the simulated present topography of Greenland (Greve, 1997c), which is used as the initial condition for the greenhouse simulations, and the simulated topographies at $t = 1000\text{yr}$ for the three control runs ggh03, ggh06, ggh10].

The simulated present value of the freshwater discharge, $559\text{ km}^3/\text{yr}$, increases to the maxima of $916\text{ km}^3/\text{yr}$ ($+3^{\circ}\text{C}$), $1559\text{ km}^3/\text{yr}$ ($+6^{\circ}\text{C}$) and even $3484\text{ km}^3/\text{yr}$ ($+10^{\circ}\text{C}$) during the modelled period. This intensified source of cold freshwater may weaken the poleward transport of warm surface water in the north Atlantic, and therefore act as a negative feedback on the northern hemisphere climate change. Effects of that type can naturally not be included in an isolated ice-sheet-modelling approach as it is pursued in this study. This shortcoming indicates clearly that future work should aim at modelling the entire coupled system atmosphere/ocean/ice-sheets in order to deal more accurately with the dynamic effects which may result from the interaction of the several components of the climate system.

The tendency of the temporal evolution of the temperate ice volume, V_{temp} , is the same in all cases: an initial increase is followed by a more or less pronounced decrease. As it was already discussed by Greve (1997b), the initial increase is due to the elevated surface temperatures, which makes the ice sheet as a whole slightly warmer. This trend does not continue because the ice volume decreases, with the consequence that, first, the internal heat production due to strain heating is reduced, and second, the lithosphere raises due to the isostatic adjustment, both causing the near-basal ice temperatures to decrease.

Fig. 3 gives for the twelve control runs ($\Delta T_{\text{ma}}^0 = +1^{\circ}\text{C} \dots +12^{\circ}\text{C}$) the total ice volume, V_{tot} , at $t = 1000\text{yr}$, the sea-level equivalent, H_{sl} , at $t = 1000\text{yr}$ and the time required to produce a sea-level equivalent of 40 cm, $t_{\text{sl}40}$. 40 cm are a threshold value for the sea-level rise in the North Sea which, when exceeded, would entail a cost-intensive reconstruction of the north German harbours (Schellnhuber, personal communication 1997). However, note that the Greenland ice sheet is not the only component which affects global sea-level

changes, so that $t_{\text{sl}40}$ is not the time at which the global sea level increases by 40 cm.

Evidently, the response functions for $V_{\text{tot}}(t=1000 \text{ yr})$ and $H_{\text{sl}}(t=1000 \text{ yr})$ are smooth s-shaped functions of the asymptotic temperature increase ΔT_{ma}^0 , which vary from an unaffected ice sheet for $\Delta T_{\text{ma}}^0 = 0^\circ\text{C}$ to an almost disintegrated ice sheet for $\Delta T_{\text{ma}}^0 = 12^\circ\text{C}$. Catastrophic events which would manifest themselves in discontinuous responses of the ice sheet do not appear. 40 cm sea-level equivalent is reached within the modelled period (1000 yr) for $\Delta T_{\text{ma}}^0 \geq 3^\circ\text{C}$, and the required time, $t_{\text{sl}40}$, decreases down to 122 yr for the extreme scenario with $\Delta T_{\text{ma}}^0 = 12^\circ\text{C}$. This demonstrates clearly that even in the absence of sudden catastrophic collapses, as they are discussed for the West Antarctic ice sheet (MacAyeal, 1992), the melting of ice sheets as a consequence of the greenhouse climate change can affect considerably the life on earth in the near future.

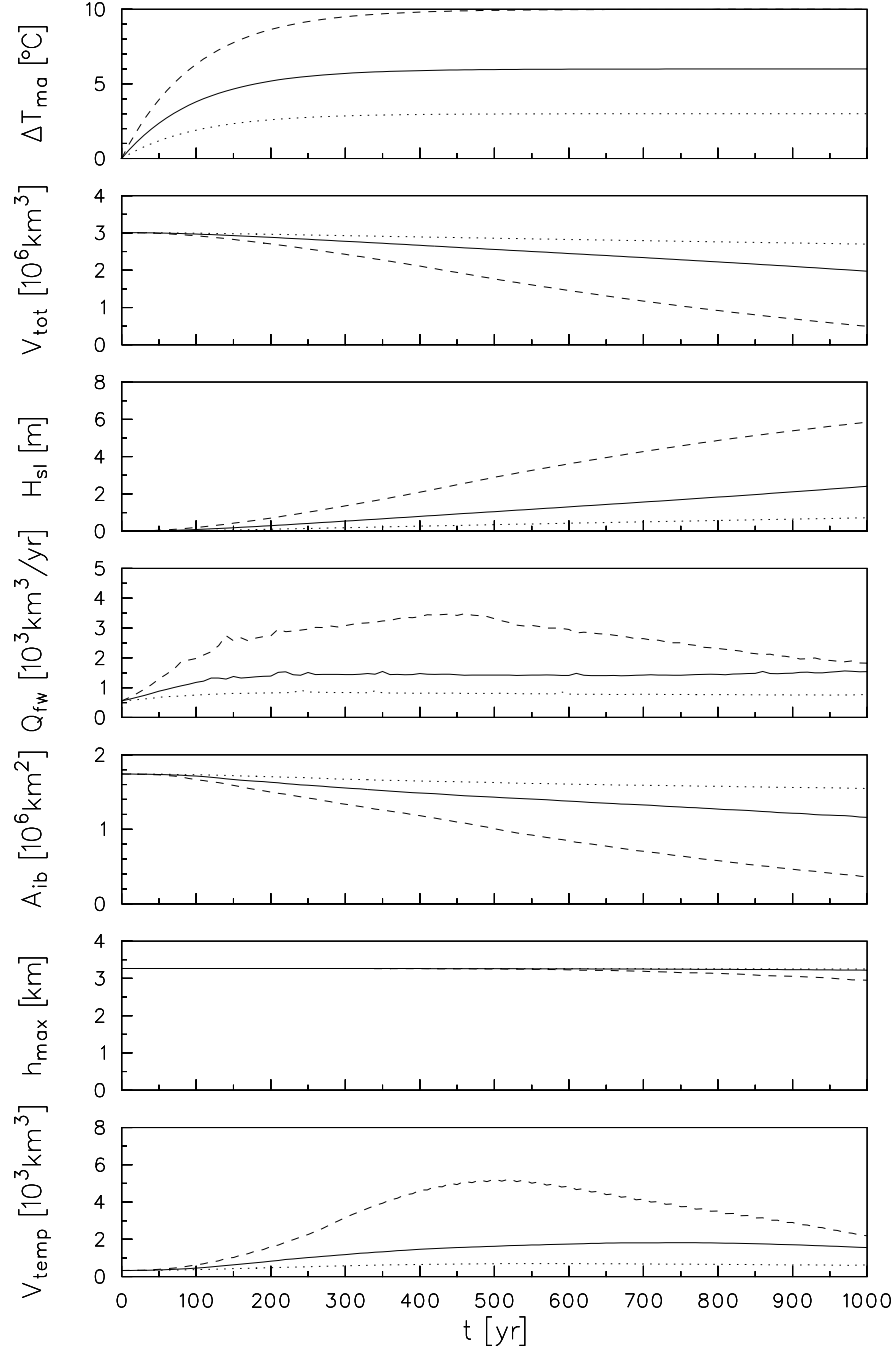


Figure 1: Temporal evolution of the mean annual air temperature (external forcing), ΔT_{ma} , the total ice volume, V_{tot} , the sea-level equivalent, H_{sl} , the freshwater discharge, Q_{fw} , the ice-covered area, A_{ib} , the maximum elevation, h_{max} , and the volume of temperate ice, V_{temp} , for the control runs ggh06 (solid), ggh10 (dashed) and ggh03 (dotted).

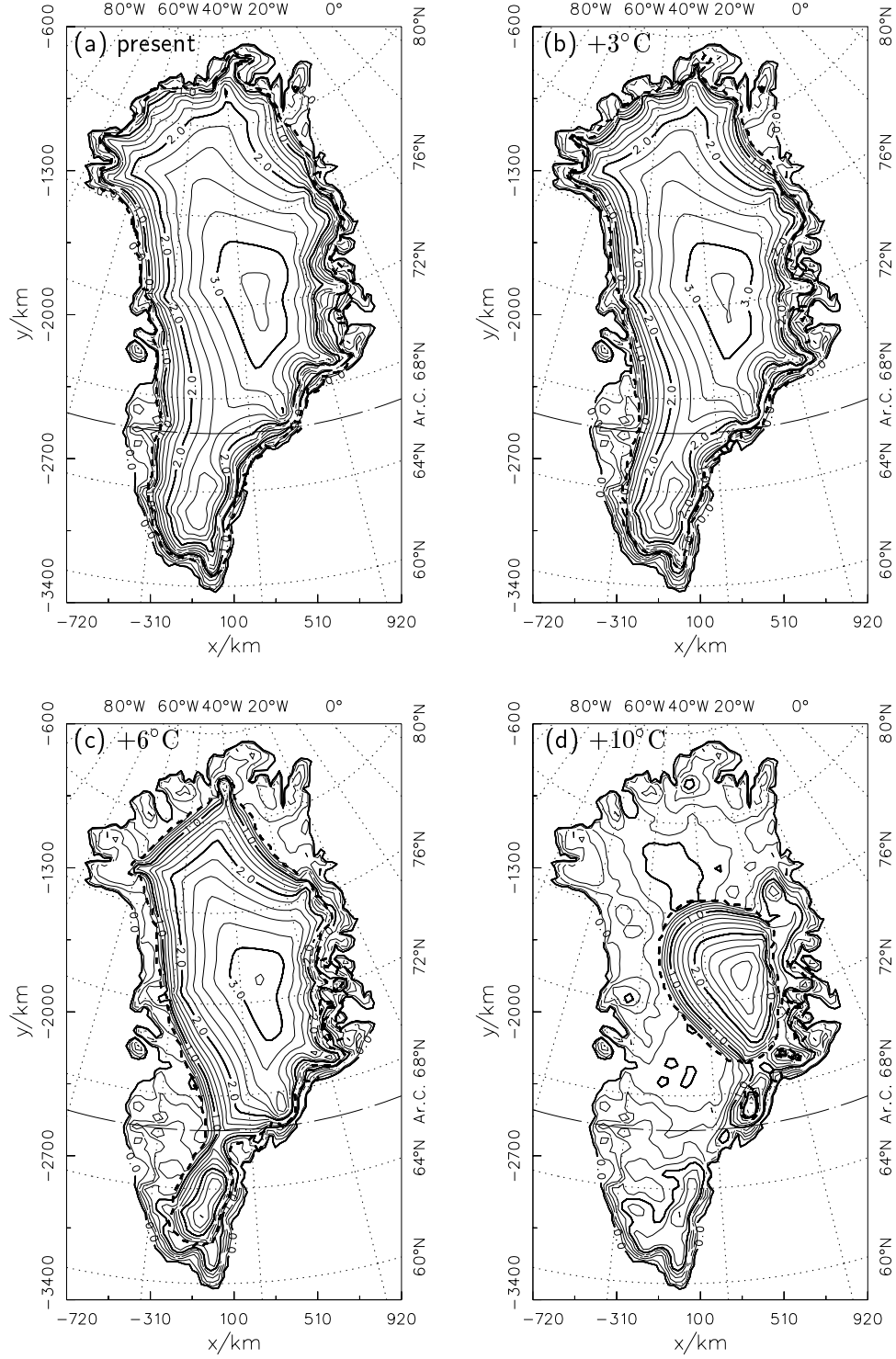


Figure 2: Simulated surface topography of Greenland, in km a.s.l. (spacing 200 m). (a) Simulated present state (Greve, 1997c), used as initial condition for the greenhouse simulations. (b) Control run ggh03 ($\Delta T_{\text{ma}}^0 = 3^\circ\text{C}$), at $t = 1000$ yr. (c) Control run ggh06 ($\Delta T_{\text{ma}}^0 = 6^\circ\text{C}$), at $t = 1000$ yr. (d) Control run ggh10 ($\Delta T_{\text{ma}}^0 = 10^\circ\text{C}$), at $t = 1000$ yr. The dashed heavy lines indicate the ice margin.

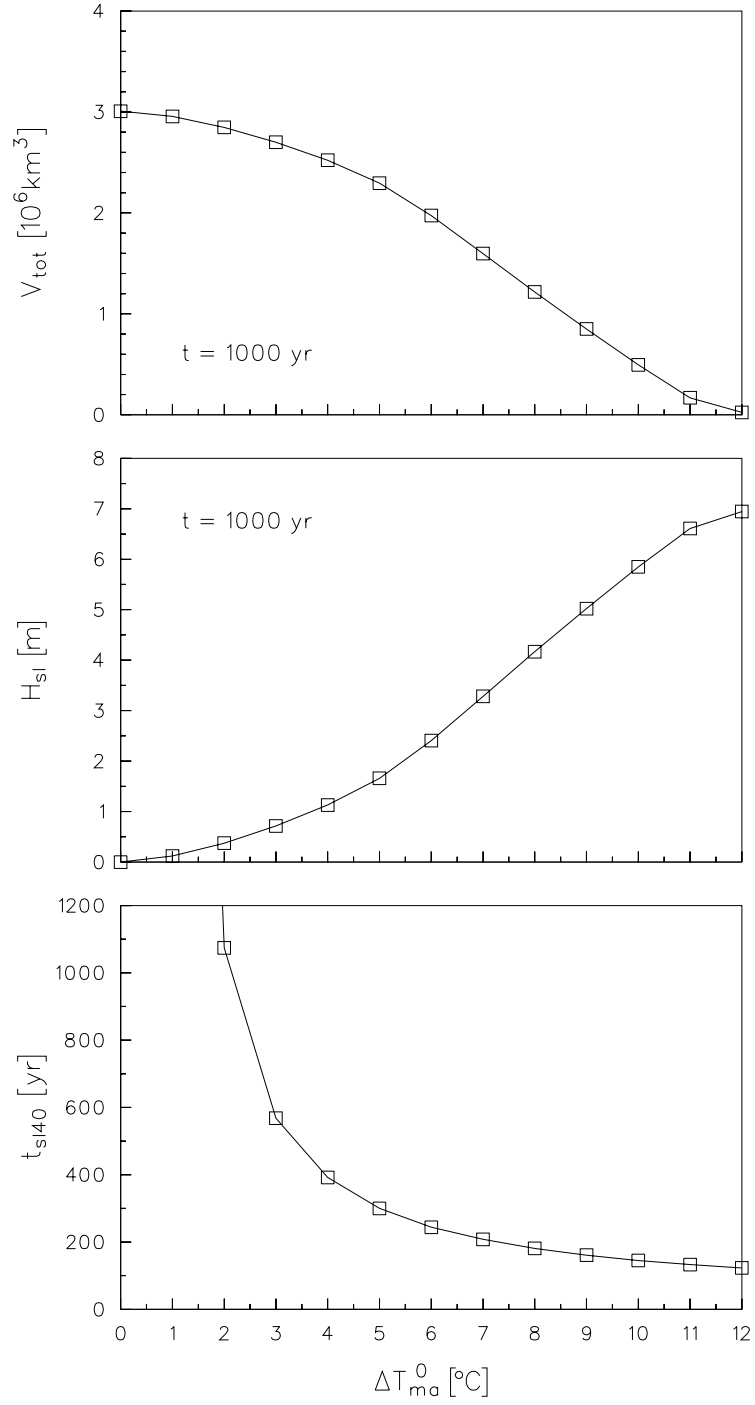


Figure 3: Simulated total ice volume, V_{tot} , sea-level equivalent, H_{sl} (both at $t = 1000 \text{ yr}$), and time at which the sea-level equivalent reaches 40 cm, $t_{\text{sl}40}$, as functions of the temperature rise, ΔT_{ma}^0 , for the control runs ggh01-ggh12.

3.2 Variation of the snowfall rate

The sensitivity of the disintegration process of the ice sheet to the assumed snowfall input is now investigated. To this end, three snowfall scenarios different from Eq. (3) are defined:

- Scenario `lin_S`:

$$S(x, y, t) = S^{\text{today}}(x, y) \cdot (1 + \gamma_s \Delta T_{\text{ma}}), \quad \text{with } \gamma_s = 0.03^\circ\text{C}^{-1}. \quad (5)$$

This relation stems from ice-core reconstructions of the past accumulation rate in central Greenland (GRIP, GISP2) which indicate a 75% reduction of S for glacial-maximum conditions (specified by $\Delta T_{\text{ma}} = -25^\circ\text{C}$) and the assumption of a linear relationship between S and ΔT_{ma} [Greve (1997c) and references therein]. It is extrapolated here for the greenhouse scenarios.

- Scenario `exp_S`:

$$S(x, y, t) = S^{\text{today}}(x, y) \cdot e^{\gamma_f \Delta T_{\text{ma}}}, \quad \text{with } \gamma_f = 0.078^\circ\text{C}^{-1}. \quad (6)$$

This is a similar reconstruction of the accumulation history derived from the GRIP core, but now based on an exponential correlation (Huybrechts et al., 1991; Fabré et al., 1995).

In this context it should be noted, however, that some of the past accumulation-rate changes are very likely of dynamic rather than of thermodynamic origin (Kapsner et al., 1995). As there is no evidence that future warming will further increase the dynamic contribution to snow accumulation, this scenario may represent a large overestimate.

- Scenario `decr_S`:

$$S(x, y, t) = S^{\text{today}}(x, y) \times \left\{ 1 - 0.05 \cdot (1 - e^{-t/\tau_{\text{gw}}}) \right\}. \quad (7)$$

The scenario leads to a decrease of the snowfall by 5%. This value was obtained by Ohmura et al. (1996) as the mean accumulation change for Greenland computed with the high-resolution ECHAM3 T106 GCM (General Circulation Model) under $2 \times \text{CO}_2$ conditions. The spatial variability simulated by these authors is not accounted for.

These parameterizations are used to drive the following simulations: Runs ggh13, ggh14, ggh15 with scenario `lin_S` and $\Delta T_{\text{ma}}^0 = 3^\circ\text{C}, 6^\circ\text{C}, 10^\circ\text{C}$; runs ggh16, ggh17, ggh18 with scenario `exp_S` and $\Delta T_{\text{ma}}^0 = 3^\circ\text{C}, 6^\circ\text{C}, 10^\circ\text{C}$; runs ggh19, ggh20, ggh21 with scenario

decr_S and $\Delta T_{\text{ma}}^0 = 3^\circ\text{C}, 6^\circ\text{C}, 10^\circ\text{C}$. Hereby, the setting **control_M** for the surface melting is kept.

In Fig. 4 the simulated total ice volume, V_{tot} , and sea-level equivalent, H_{sl} , at $t = 1000$ yr are shown for these runs together with the control runs. The results for scenarios **const_S**, **lin_S** and **decr_S** are not substantially different from one another. However, in scenario **exp_S** the retreat of the ice sheet is distinctly reduced. The difference between **exp_S** and **const_S** with respect to the present ice sheet is 4.3% ($+3^\circ\text{C}$), 12.7% ($+6^\circ\text{C}$) and 17.5% ($+10^\circ\text{C}$). The latter corresponds to an uncertainty of the sea-level equivalent of more than 1 m. The reason for this is that the exponential coupling between the snowfall S and the surface-temperature deviation ΔT_{ma} [Eq. (6)] stabilizes the ice sheet subjected to greenhouse-warming conditions via strongly increased snowfall.

It is difficult to judge which of these scenarios is the most appropriate. Scenario **const_S** (unchanged snowfall) is an ad-hoc assumption. As already mentioned above, scenarios **lin_S** and **exp_S**, which imply intensified snowfall under warmer climate conditions, were derived by extrapolating palaeoclimatic ice-core records into the future, and the assumed snowfall decrease in **decr_S** stems from the result of a $2 \times \text{CO}_2$ GCM simulation. The latter approach seems to be preferable, provided the GCM results are sufficiently accurate. As the results for **decr_S** and **const_S** are very close to each other, scenario **const_S** represents also a meaningful parameterization. In any case, the findings of this section demonstrate that a proper snowfall input is crucial for reliable predictions of the future evolution of ice sheets in warmer climates.

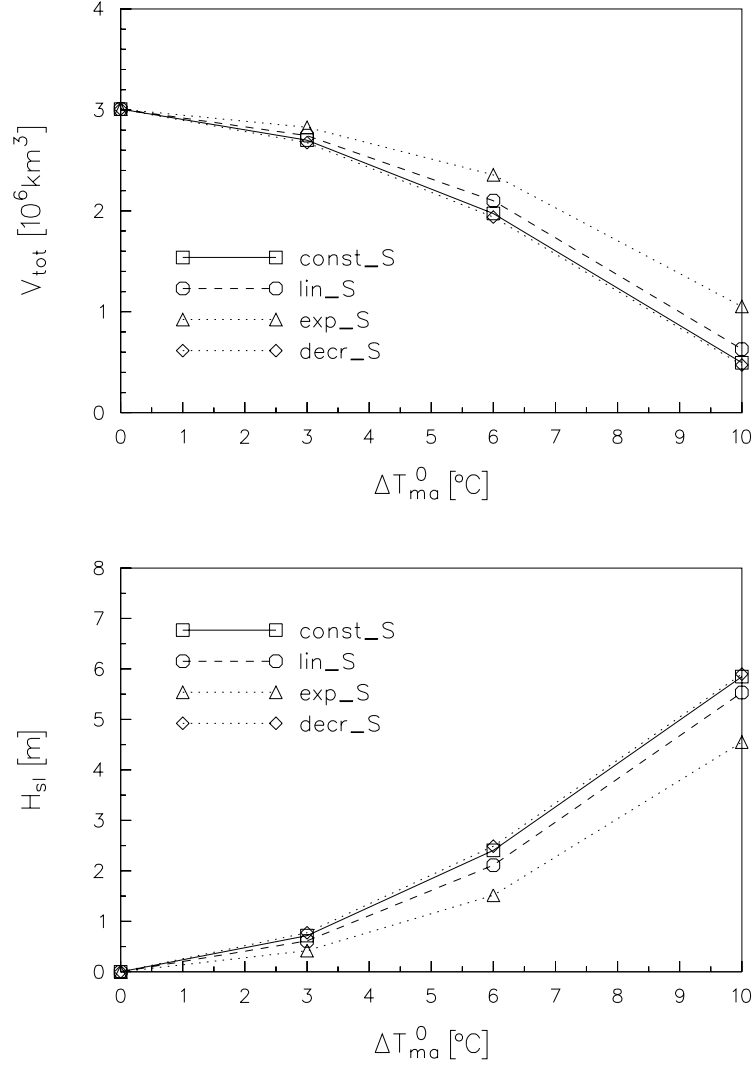


Figure 4: Simulated total ice volume, V_{tot} , and sea-level equivalent, H_{sl} , at $t = 1000$ yr, as functions of the temperature rise, ΔT_{ma}^0 , for different snowfall scenarios.

3.3 Variation of the melting rate

The mass balance at the surface of ice sheets is the difference between snowfall and surface melting. Since in the previous section a distinct influence of the prescribed snowfall on the ice-sheet dynamics was found, a similar influence of surface melting is expected. This is now investigated by modifying the degree-day factors of Eq. (4) within their range of uncertainty:

- Scenario **small_M**:

$$\beta_{\text{snow}} = 3 \frac{\text{mm WE}}{\text{d}\cdot\text{K}}, \quad \beta_{\text{ice}} = 7 \frac{\text{mm WE}}{\text{d}\cdot\text{K}}. \quad (8)$$

- Scenario **large_M**:

$$\beta_{\text{snow}} = 3 \frac{\text{mm WE}}{\text{d}\cdot\text{K}}, \quad \beta_{\text{ice}} = 10 \frac{\text{mm WE}}{\text{d}\cdot\text{K}}. \quad (9)$$

In these scenarios, the parameter β_{ice} which is most crucial for the parameterized melting rate is kept constant at the lower and upper value, respectively, of the latitude-dependent distribution given in Eq. (4).

Simulations are carried out with these settings as follows: Runs ggh22, ggh23, ggh24 with scenario **small_M** and $\Delta T_{\text{ma}}^0 = 3^\circ\text{C}, 6^\circ\text{C}, 10^\circ\text{C}$; runs ggh25, ggh26, ggh27 with scenario **large_M** and $\Delta T_{\text{ma}}^0 = 3^\circ\text{C}, 6^\circ\text{C}, 10^\circ\text{C}$. For the snowfall, the setting **const_S** is applied.

Fig. 5 displays the corresponding total ice volume, V_{tot} , and sea-level equivalent, H_{sl} , at $t = 1000$ yr. As it was expected, the above changes influence the ice-sheet disintegration to a similar extent as the changes of the snowfall rate do. With the present ice volume used as reference volume, the difference between the extreme cases **small_M** and **large_M** is 5.4% ($+3^\circ\text{C}$), 14.2% ($+6^\circ\text{C}$) and 18.6% ($+10^\circ\text{C}$). Consequently, accurate modelling of the processes that determine surface melting is very important for realistic simulations of ice-sheet dynamics under greenhouse-warming conditions. This claim contrasts with the simplicity of the degree-day model, which relates the melting rate empirically to the air-temperature excess above the melting point and does not account for the influence of wind speed, albedo and cloud cover. Hence, it is worth thinking of replacing the degree-day model by a more sophisticated energy-balance modelling as indicated by Braithwaite (1995). This could be achieved by embedding the ice-sheet model in a GCM.

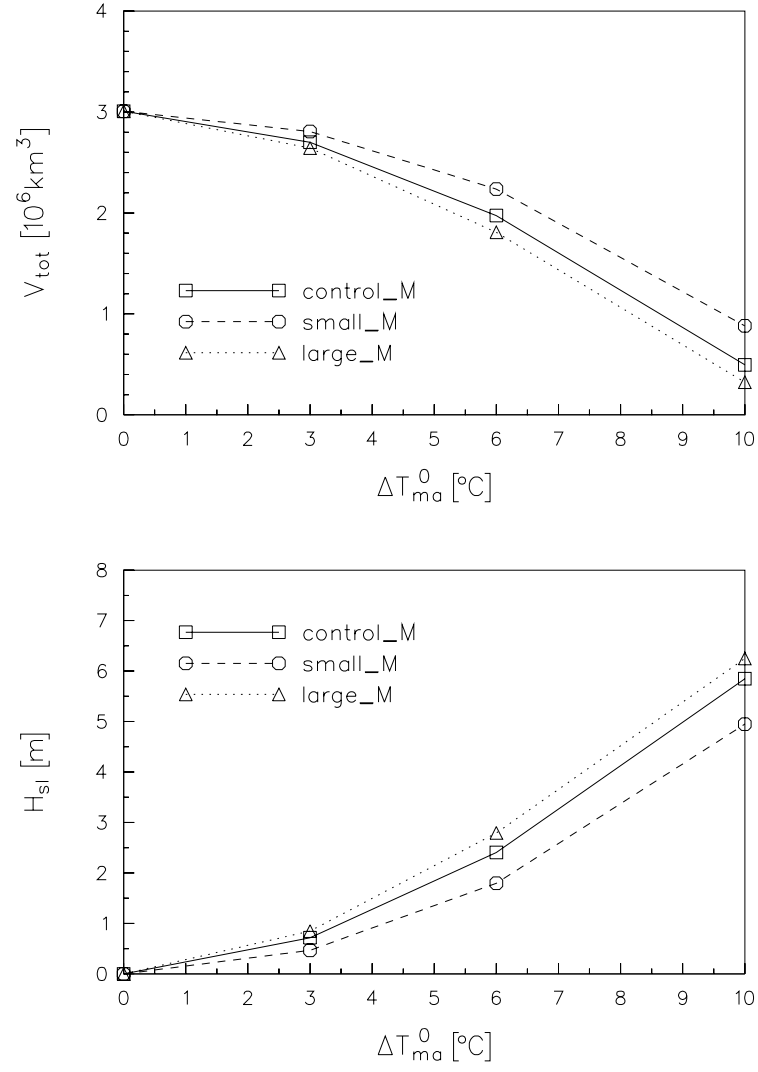


Figure 5: Same as Fig. 4, but for different melting scenarios.

4 Discussion

In the above section, a series of greenhouse-warming simulations for the Greenland ice sheet was presented, and the impact of the surface temperature and the surface mass balance (snowfall, melting) on the expected ice-sheet disintegration was elucidated. It was demonstrated that semi-quantitative predictions of the evolution of the ice sheet are possible for given surface-temperature scenarios. However, uncertainties associated with the snowfall and surface-melting input forbid really accurate results. The possibility of a critical threshold of the climatic input associated with a creep instability which was raised in the introduction and would manifest itself in a sudden volume drop of the ice sheet could not be confirmed. For all conducted simulations the ice-sheet response is a smooth function of the prescribed greenhouse warming.

In contrast to the Greenland ice sheet, the West Antarctic ice sheet may be more susceptible to discontinuous behaviour. The effect of a lubricating layer of rapidly deformable subglacial sediment (see introduction) may even lead to quasi-periodic disintegrations *in the absence of greenhouse warming* (MacAyeal, 1992), it is furthermore made responsible for several collapses of the glacial Laurentide (north American) ice sheet, which can explain the existence of distinct layers of ice-rafted debris in north Atlantic sediments (“Heinrich events”; Heinrich, 1988; Greve and MacAyeal, 1996). Since the onset and temporal evolution of a disintegration event triggered by a lubricating layer of basal sediment is strongly dependent on the (virtually unknown) initial sediment distribution below the ice, events of that type are very difficult to predict. This is a severe limitation of the reliability of simulated global ice-sheet responses to greenhouse-warming scenarios. Surprises are possible!

Further, feedbacks of the ice sheets on the atmosphere and the ocean can naturally not be described within the framework of isolated ice-sheet modelling. Such feedbacks comprise the effect of the changing ice topography and albedo on the air temperature and the predominant winds as well as the influence of the meltwater discharge on the pattern of ocean currents. The main consequence of this problem is that future work must aim at replacing the isolated ice-sheet-modelling approach by coupled atmosphere-ocean-ice-sheet modelling, because this provides in principle the means to account more adequately for the complexity of the system. First steps in this direction have been performed by e. g. Marsiat (1994), de Wolde et al. (1997) and Calov and Marsiat (1998) with coupled 2-d energy balance climate models and 3-d dynamic/thermodynamic ice sheet models, fully coupled simulations with 3-d atmosphere-ocean GCMs and ice sheet models have not yet been realized. A major problem with existing atmosphere models including GCMs is that they are relatively inconsistent in the simulated precipitation patterns; further,

computation of the ablation at the ice/atmosphere interface is a large source of error also in coupled simulations. This shows that great efforts are still required until precise predictions of the reaction of ice sheets within the climate system to future climate change can be made.

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