ON THE RESPONSE OF THE GREENLAND ICE SHEET TO GREENHOUSE CLIMATE CHANGE

RALF GREVE
Institut für Mechanik III, Technische Universität Darmstadt, Hochschulstraße 1, D-64289 Darmstadt, Germany

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 $\Delta T = 1^\circ\mathrm{C}$ to $12^\circ\mathrm{C}$, 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 $\Delta T \leq 3^\circ\mathrm{C}$, whereas the most extreme scenario, $\Delta T = 12^\circ\mathrm{C}$, 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 vapor, 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, $\Delta T_{2\times\mathrm{CO}_2}$. According to the Intergovernmental Panel on Climate Change (IPCC, 1996), an estimate for this value is $1.9^\circ\mathrm{C} \leq \Delta T_{2\times\mathrm{CO}_2} \leq 5.4^\circ\mathrm{C}$, 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, the 3-d ice-sheet model SICOPOLIS (see below)

Climatic Change **46:** 289–303, 2000.

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 sediments 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, at present, the prerequisites for large-scale surging and grounding-line instability are only fulfilled 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 Modeling 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 to 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, rate factor, basal sliding) 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 ($t = 1000\,\mathrm{yr}$) are discussed. The simulations are driven by surface-temperature increases from $+1^\circ\mathrm{C}$ to $+12^\circ\mathrm{C}$, with the sea level kept constant. A variety of snowfall and melting scenarios are applied, and the geothermal heat flux is kept constant at $Q^{\perp}_{\mathrm{geoth}} = 65\,\mathrm{mW/m^2}$. 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 palaeoclimatic simulations (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 of 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:

$$T_{\mathrm{ma}}(x,y,z,t) = T^{\mathrm{today}}_{\mathrm{ma}}(x,y,z) + \Delta T_{\mathrm{ma}}(t), \tag{1}$$

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, $T^{\mathrm{today}}_{\mathrm{ma}}$, 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^0_{ma} and the delay time τ_{gw}:

$$\Delta T_{\mathrm{ma}}(t) = \Delta T^0_{\mathrm{ma}}\left(1 - e^{-t/\tau_{\mathrm{gw}}}\right), \qquad \tau_{\mathrm{gw}} = 100\,\mathrm{yr}. \tag{2}$$

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), and Wilhelms (1996):

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

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 $72^\circ\,34'$ N, $37^\circ\,38'$ 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):

$$\beta_{\mathrm{snow}} = 3\,\frac{\mathrm{mm\,WE}}{\mathrm{d\cdot K}}, \qquad \beta_{\mathrm{ice}}(\phi = 60^\circ\mathrm{N}) = 7\,\frac{\mathrm{mm\,WE}}{\mathrm{d\cdot K}}, \tag{4}$$

$$\beta_{\mathrm{ice}}(\phi = 80^\circ\mathrm{N}) = 10\,\frac{\mathrm{mm\,WE}}{\mathrm{d\cdot K}},$$

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 modeling 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 $\tau_{\mathrm{V}} = 3000\,\mathrm{yr}$.

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

Figure 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 $+3^\circ\mathrm{C}$ 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 ($+6^\circ\mathrm{C}$ temperature rise) and ggh10 ($+10^\circ\mathrm{C}$ temperature rise): for these simulations, the molten ice volume is 34.4% ($+6^\circ\mathrm{C}$) and as large as 83.5% ($+10^\circ\mathrm{C}$), corresponding to sea-level equivalents of 2.4 m ($+6^\circ\mathrm{C}$) and 5.8 m ($+10^\circ\mathrm{C}$).

It is striking that these retreats are virtually not accompanied by changes in the maximum surface elevation. Even in run ggh10 ($+10^\circ\mathrm{C}$), h_{max} changes merely

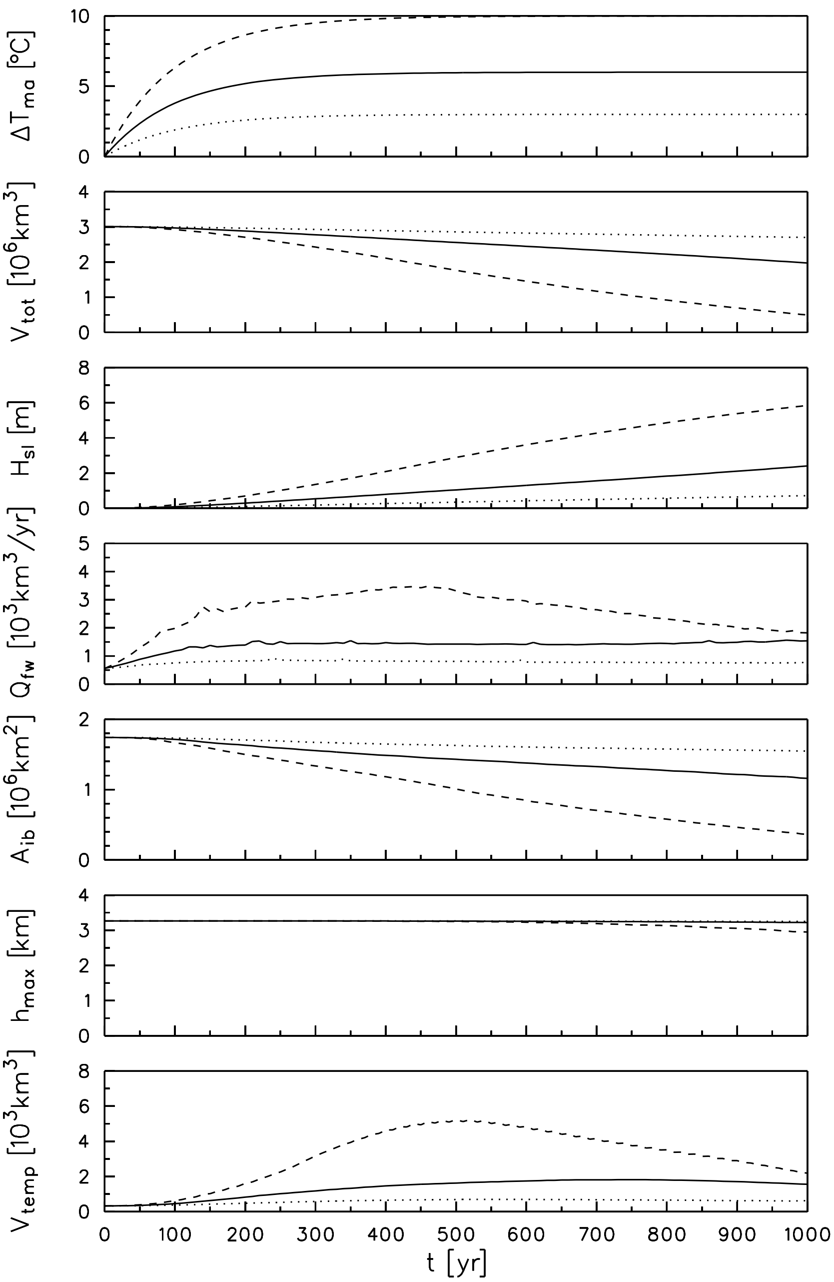

Figure 1. Temporal evolution of the mean annual air temperature (external forcing), ΔT_{ma}, the total ice volume, V_{tot}, the sea-level equivalent, H_{sl}, the freshwater discharge, Q_{fw}, the ice-covered area, A_{ib}, the maximum elevation, $h_{\max}$, and the volume of temperate ice, V_{temp}, for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

from 3267 m at $t = 0$ to 2950 m at $t = 1000\,\mathrm{yr}$. This is so because the eastern part of the ice sheet situated between approximately $70°\mathrm{N}$ and $76°\mathrm{N}$ is very stable as a consequence of the high bedrock elevation in this region [see Figure 2 which depicts the simulated present topography of Greenland (Greve, 1997c) used as the initial condition for the greenhouse simulations, and the simulated topographies at $t = 1000\,\mathrm{yr}$ for the three control runs ggh03, ggh06, ggh10].

The simulated present value of the freshwater discharge, $559\,\mathrm{km}^3/\mathrm{yr}$, increases to the maxima of $916\,\mathrm{km}^3/\mathrm{yr}$ (+3°C), $1559\,\mathrm{km}^3/\mathrm{yr}$ (+6°C) and even 3484 $\mathrm{km}^3/\mathrm{yr}$ ($+10°\mathrm{C}$) during the modeled 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-modeling approach as it is pursued in this study. This shortcoming indicates clearly that future work should aim at modeling the entire coupled system atmosphere/ocean/ice-sheets in order to deal more accurately with the dynamic effects that 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 that make 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.

Figure 3 gives for the twelve control runs ($\Delta T^0_{\mathrm{ma}} = +1°\mathrm{C} \ldots + 12°\mathrm{C}$) the total ice volume, V_{tot}, at $t = 1000\,\mathrm{yr}$, the sea-level equivalent, H_{sl}, at $t = 1000\,\mathrm{yr}$ and the time required to produce a sea-level equivalent of $40\,\mathrm{cm}$, t_{sl40}. $40\,\mathrm{cm}$ is a threshold value for the sea-level rise in the North Sea that, when exceeded, would entail a cost-intensive reconstruction of the North German harbors (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_{sl40} is not the time at which the global sea level increases by $40\,\mathrm{cm}$.

Evidently, the response functions for $V_{\mathrm{tot}}(t=1000\,\mathrm{yr})$ and $H_{\mathrm{sl}}(t=1000\,\mathrm{yr})$ are smooth s-shaped functions of the asymptotic temperature increase ΔT^0_{ma}, which vary from an unaffected ice sheet for $\Delta T^0_{\mathrm{ma}} = 0°\mathrm{C}$ to an almost disintegrated ice sheet for $\Delta T^0_{\mathrm{ma}} = 12°\mathrm{C}$. Catastrophic events that would manifest themselves in discontinuous responses of the ice sheet do not appear. $40\,\mathrm{cm}$ sea-level equivalent is reached within the modeled period (1000 yr) for $\Delta T^0_{\mathrm{ma}} \geq 3°\mathrm{C}$, and the required time, t_{sl40}, decreases down to 122 yr for the extreme scenario with $\Delta T^0_{\mathrm{ma}} = 12°\mathrm{C}$. 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 melt-

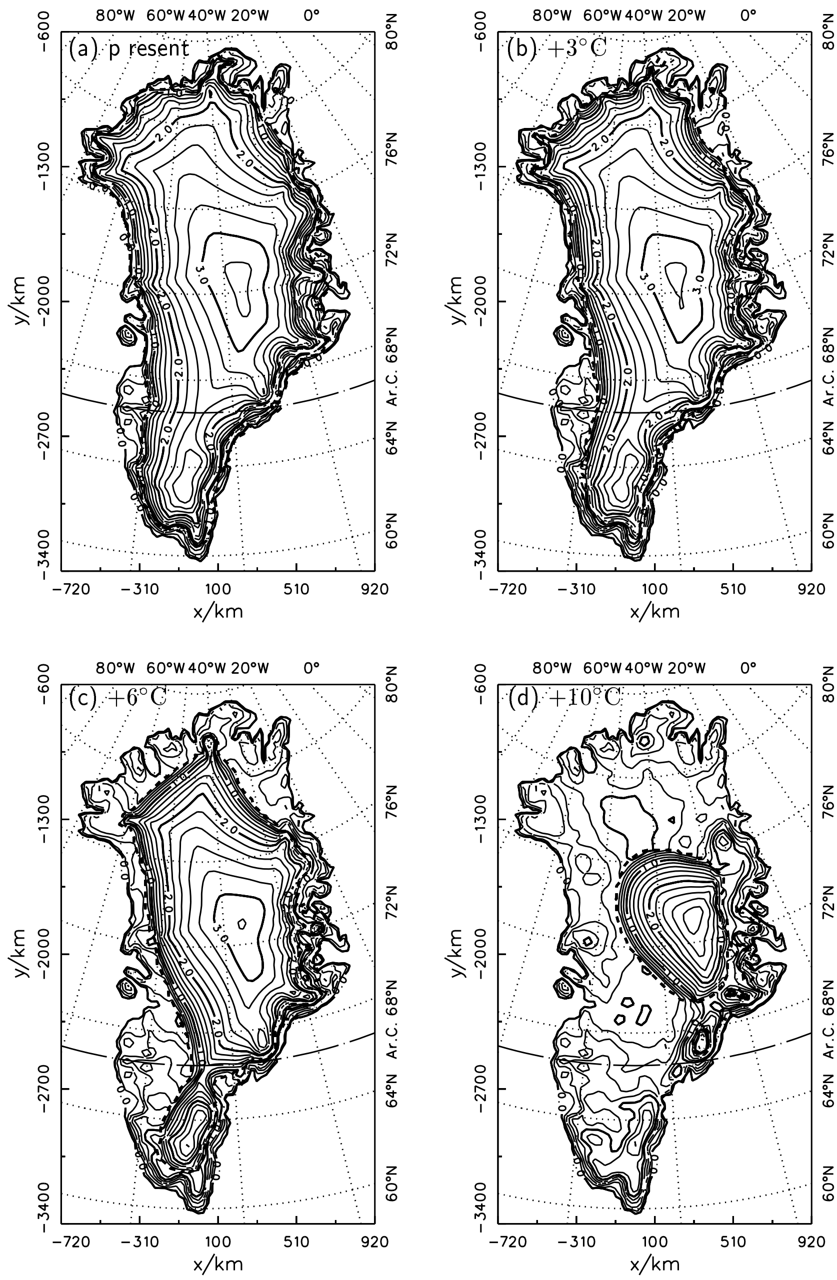

Figure 2. Simulated surface topography of Greenland, in km a.s.l. (spacing 200 m). (a) Simulated present state (Greve, 1997c), used as initial condition for the greenhouse simulations. (b) Control run ggh03 ($\Delta T^0_{\mathrm{ma}} = 3°\mathrm{C}$), at $t = 1000\,\mathrm{yr}$. (c) Control run ggh06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

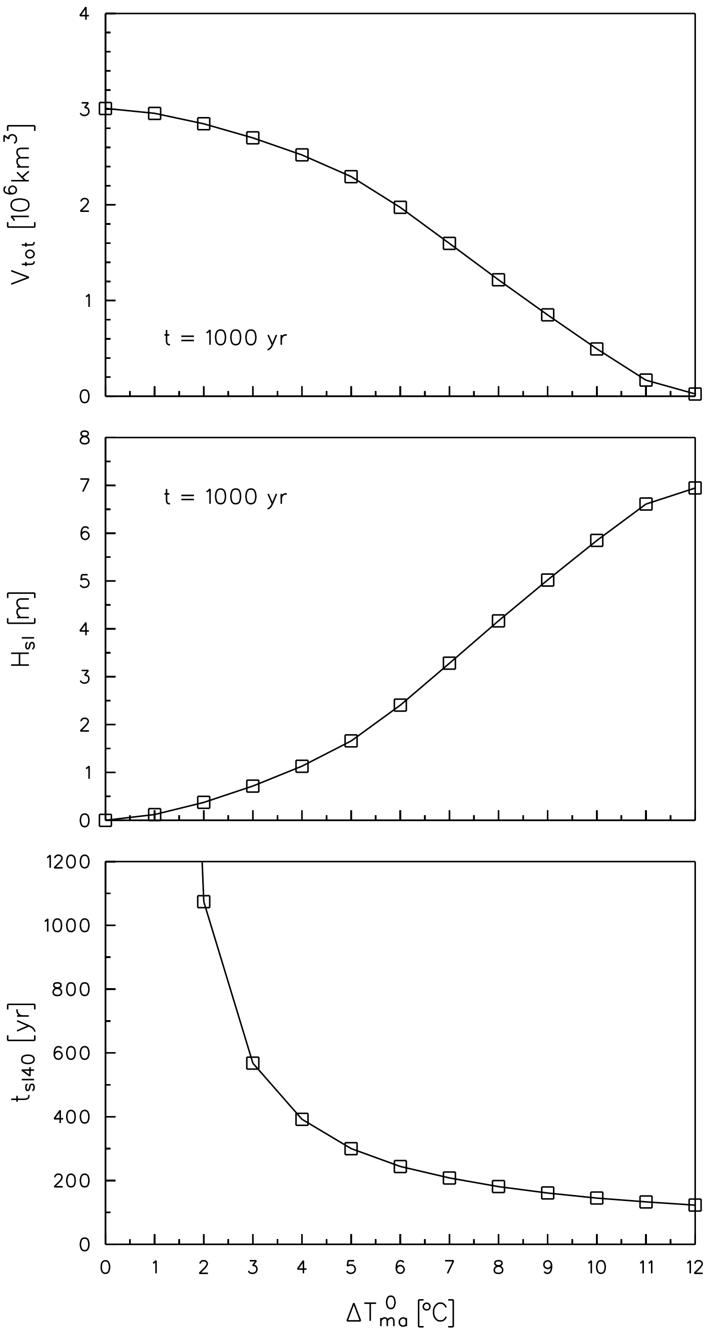

Figure 3. Simulated total ice volume, V_{tot}, sea-level equivalent, H_{sl} (both at $t = 1000\ \mathrm{yr}$), and time at which the sea-level equivalent reaches 40 cm, t_{sl40}, as functions of the temperature rise, ΔT^0_{ma}, for the control runs ggh01-ggh12.

ing of ice sheets as a consequence of the climate change can affect considerably the life on earth in the near future.

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 Equation (3) are defined:

– Scenario lin_S:

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

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 $\Delta T_{\text{ma}} = -25°\text{C}$) and the assumption of a linear relationship between S and ΔT_{ma} (see Greve, 1997c, and references therein). It is extrapolated here for the greenhouse scenarios.

– Scenario exp_S:

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

This is a similar reconstruction of the accumulation history derived from the GRIP core; however, 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 changes in accumulation rate are very likely of dynamic rather than 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:

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

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 $\Delta T^0_{\text{ma}} = 3°\text{C}, 6°\text{C}, 10°\text{C}$; runs ggh16, ggh17, ggh18 with scenario exp_S and $\Delta T^0_{\text{ma}} = 3°\text{C}, 6°\text{C}, 10°\text{C}$; runs ggh19, ggh20, ggh21 with scenario decr_S and $\Delta T^0_{\text{ma}} = 3°\text{C}, 6°\text{C}, 10°\text{C}$. Hereby, the setting control_M for the surface melting is kept.

In Figure 4 the simulated total ice volume, V_{tot}, and sea-level equivalent, H_{sl}, at $t = 1000\,\text{yr}$ 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

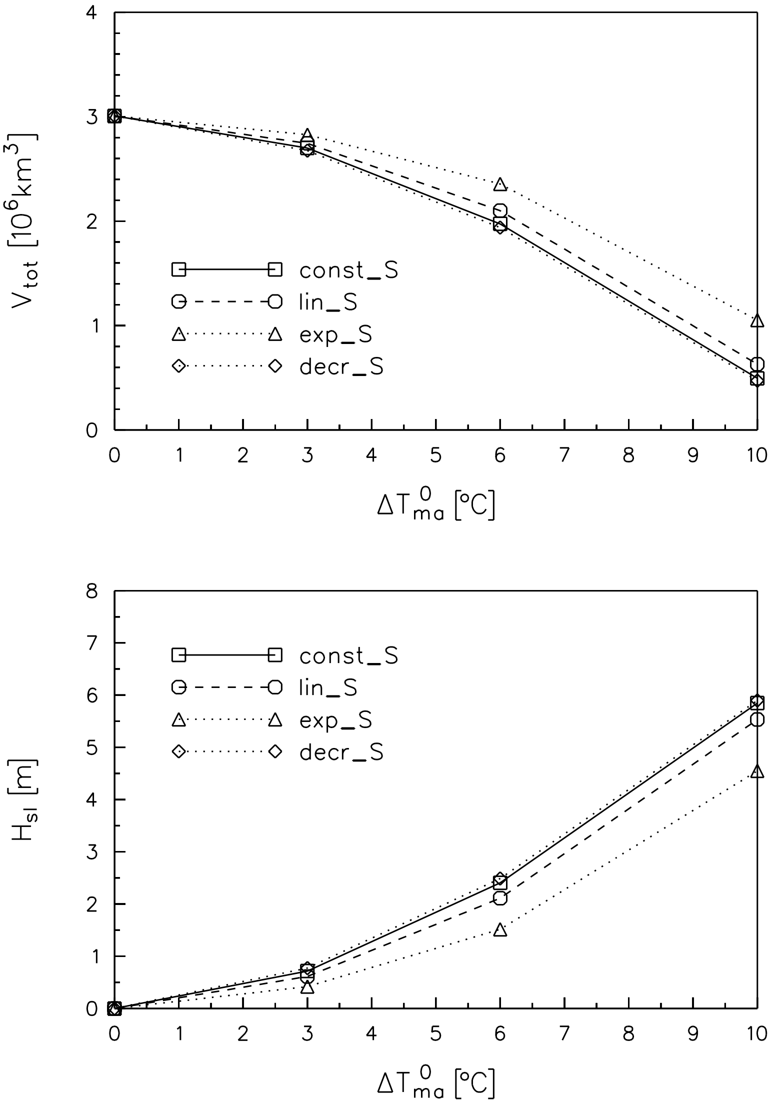

Figure 4. Simulated total ice volume, $V_{\rm tot}$, and sea-level equivalent, $H_{\rm sl}$, at $t = 1000$ yr, as functions of the temperature rise, $\Delta T^0_{\rm ma}$, for different snowfall scenarios.

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% ($+3°$C), 12.7% ($+6°$C) and 17.5% ($+10°$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 $\Delta T_{\rm ma}$ (Equation (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 that 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 $2\times CO_2$ 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 also represents 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.

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 Equation (4) within their range of uncertainty:

– Scenario small_M:

$$\beta_{\mathrm{snow}} = 3\,\frac{\mathrm{mm\,WE}}{\mathrm{d\cdot K}}, \qquad \beta_{\mathrm{ice}} = 7\,\frac{\mathrm{mm\,WE}}{\mathrm{d\cdot K}}. \tag{8}$$

– Scenario large_M:

$$\beta_{\mathrm{snow}} = 3\,\frac{\mathrm{mm\,WE}}{\mathrm{d\cdot K}}, \qquad \beta_{\mathrm{ice}} = 10\,\frac{\mathrm{mm\,WE}}{\mathrm{d\cdot K}}. \tag{9}$$

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 Equation (4).

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

Figure 5 displays the corresponding total ice volume, V_{tot}, and sea-level equivalent, H_{sl}, at $t = 1000\,\mathrm{yr}$. As it was expected, the above changes influence the ice-sheet disintegration to a similar extent as changes in 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% ($+3°\mathrm{C}$), 14.2% ($+6°\mathrm{C}$) and 18.6% ($+10°\mathrm{C}$). Consequently, accurate modeling 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 model as indicated by

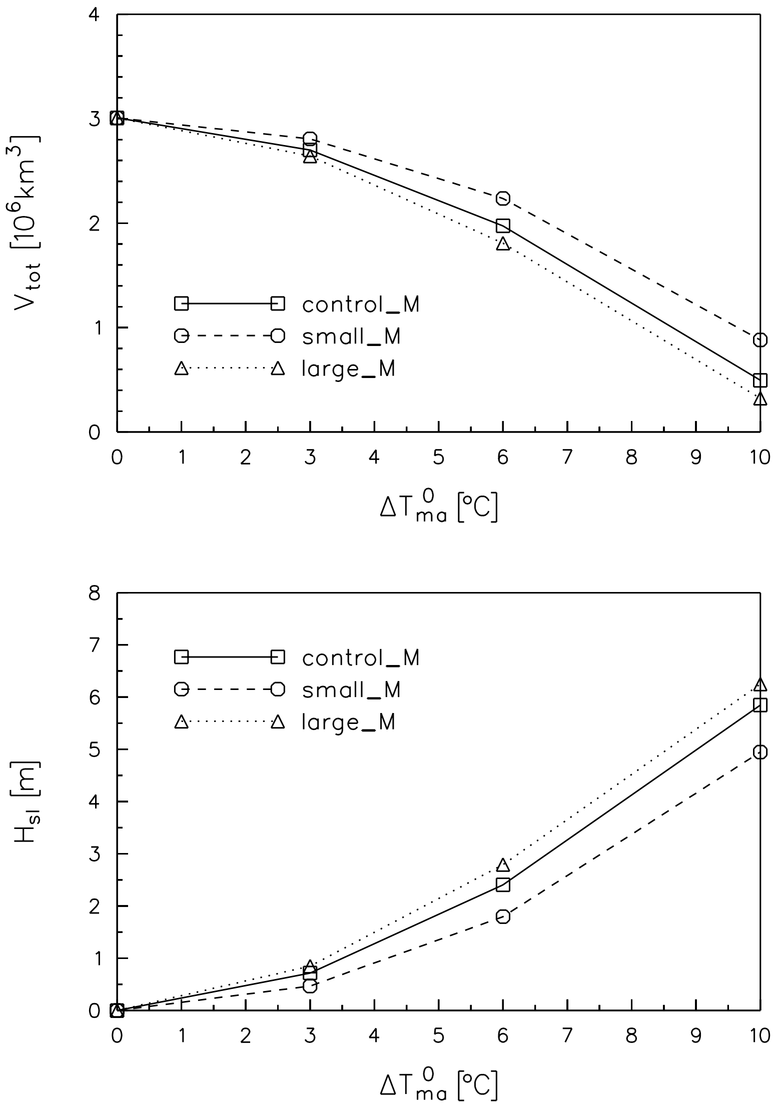

Figure 5. Same as Figure 4, but for different melting scenarios.

Braithwaite (1995). This could be achieved by embedding the ice-sheet model in a GCM.

4. Discussion

In the previous 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 that 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, that 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 this type are very difficult to predict. This is a severe limitation on 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 modeling. 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-modeling approach by coupled atmosphere-ocean-ice-sheet modeling, 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.

Acknowledgements

This paper is the outcome of a presentation held by the author on the “International Workshop on Climate Change: Thresholds and Response Functions” carried out by the Potsdam Institute for Climate Impact Research (January 23-24, 1997; Potsdam, Germany). The fruitful interdisciplinary discussions on this workshop led to a considerable improvement of the original version.

Thanks are due to Prof. Kolumban Hutter for carefully reading and correcting the manuscript, and to two anonymous reviewers whose comments improved considerably the clarity of this paper.

References

Bolzan, J.F. and Strobel, M.: 1994, ‘Accumulation-rate variations around Summit, Greenland’, *Journal of Glaciology* **40**(134), 56-66.

Braithwaite, R.J.: 1995, ‘Positive degree-day factors for ablation on the Greenland ice sheet studied by energy-balance modelling’, *Journal of Glaciology* **41**(137), 153-160.

Braithwaite, R.J. and Olesen, O.B.: 1989, ‘Calculation of glacier ablation from air temperature, West Greenland’, in J. Oerlemans (ed.), *Glacier fluctuations and climatic change*, Kluwer, Dordrecht, pp. 219-233.

Calov, R. and Hutter, K.: 1996, ‘The thermomechanical response of the Greenland ice sheet to various climate scenarios’, *Climate Dynamics* **12**, 243-260.

Calov, R. and Marsiat, I.: 1998, ‘Simulations of the northern hemisphere through the last glacial-interglacial cycle with a vertically integrated and a 3-d thermomechanical ice sheet model coupled to a climate model’, *Annals of Glaciology* **27**, 169-176.

Clarke, G.K.C., Nitsan, U., and Paterson, W.S.B.: 1977, ‘Strain heating and creep instability in glaciers and ice sheets’, *Reviews of Geophysics and Space Physics* **15**(2), 235-247.

De Wolde, J.R., Huybrechts, P., Oerlemans, J., and van de Wal, R.S.W.: 1997, ‘Projections of global mean sea-level rise calculated with a 2D energy-balance climate model and dynamic ice sheet models’, *Tellus* **49A**, 486-502.

Fabré, A., Letréguilly, A., Ritz, C., and Mangeney, A.: 1995, ‘Greenland under changing climates: sensitivity experiments with a new three-dimensional ice sheet model’, *Annals of Glaciology* **21**, 1-7.

Fowler, A.C. and Larson, D.A.: 1978, ‘On the flow of polythermal glaciers. I. Model and preliminary analysis’, *Proceedings of the Royal Society London* **A-363**, 217-242.

Greve, R.: 1997a: ‘A continuum-mechanical formulation for shallow polythermal ice sheets’, *Philosophical Transactions of the Royal Society London* **A-355**(1726), 921-974.

Greve, R.: 1997b, ‘Application of a polythermal three-dimensional ice sheet model to the Greenland ice sheet: Response to steady-state and transient climate scenarios’, *Journal of Climate* **10**(5), 901-918.

Greve, R.: 1997c, ‘Large-scale ice-sheet modelling as a means of dating deep ice cores in Greenland’, *Journal of Glaciology* **43**(144), 307-310; Erratum **43**(145), 597-600.

Greve, R. and MacAyeal, D.R.: 1996, ‘Dynamic/thermodynamic simulations of Laurentide ice sheet instability’, *Annals of Glaciology* **23**, 328-335.

Greve, R., Weis, M., and Hutter, K.: 1998, ‘Palaeoclimatic evolution and present conditions of the Greenland ice sheet in the vicinity of Summit: An approach by large-scale modelling’, *Palaeoclimates* **2**(2-3), 133-161.

Heinrich, H.: 1988, 'Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130 000 years', *Quaternary Research* **29**, 142-152.

Hutter, K.: 1982, 'A mathematical model of polythermal glaciers and ice sheets', *Journal of Geophysical and Astrophysical Fluid Dynamics* **21**, 201-224.

Hutter, K.: 1993, 'Thermo-mechanically coupled ice sheet response. Cold, polythermal, temperate', *Journal of Glaciology* **39**(131), 65-86.

Huybrechts, P.: 1994, 'The present evolution of the Greenland ice sheet: an assessment by modelling', *Global and Planetary Change* **9**, 39-51.

Huybrechts, P., Letréguilly, A., and Reeh, N.: 1991, 'The Greenland ice sheet and greenhouse warming', *Palaeogeography, Palaeoclimatology, Palaeoecology (Global and Planetary Change Section)* **89**, 399-412.

IPCC, 1996, *Climate Change 1995: The Science of Climate Change*, Cambridge University Press, Cambridge.

Johnsen, S.J., Dahl-Jensen, D., Dansgaard, W., and Gundestrup, N.: 1995, 'Greenland palaeotemperatures derived from GRIP borehole temperature and ice core isotope profiles', *Tellus* **47B**, 624-629.

Kapsner, W.R., Alley, R.B., Shuman, C.A., Anandakrishnan, S. and Grootes, P.M.: 1995, 'Dominant influence of atmospheric circulation on snow accumulation in greenland over the past 18 000 years', *Nature* **373**, 52-54.

Le Meur, E. and Huybrechts, P.: 1996, 'A comparison of different ways of dealing with isostasy: examples from modelling the Antarctic ice sheet during the last glacial cycle, *Annals of Glaciology* **23**, 309-317.

Letréguilly, A., Huybrechts, P., and Reeh, N.: 1991, 'Steady-state characteristics of the Greenland ice sheet under different climates', *Journal of Glaciology* **37**(125), 149-157.

MacAyeal, D.R.: 1992, 'Irregular oscillations of the West Antarctic ice sheet', *Nature* **359**, 29-32.

Marsiat, I.: 1994, 'Simulation of the northern hemisphere continental ice sheets over the last glacial-interglacial cycle: Experiments with a latitude-longitude vertically integrated ice sheet model coupled to a zonally averaged climate model', *Palaeoclimates* **1**, 59-98.

Ohmura, A.: 1987, 'New temperature distribution maps for Greenland', *Zeitschrift fr Gletscherkunde und Glazialgeologie* **23**, 1-45.

Ohmura, A. and Reeh, N.: 1991, 'New precipitation and accumulation maps for Greenland', *Journal of Glaciology* **37**, 140-148.

Ohmura, A., Wild, M., and Bengtsson, L.: 1996, 'Present and future mass balance of the ice sheets simulated with GCM', *Annals of Glaciology* **23**, 187-193.

Reeh, N.: 1991, 'Parameterization of melt rate and surface temperature on the Greenland ice sheet', *Polarforschung* **59**(3), 113-128.

Ritz, C.: 1997, 'EISMINT intercomparison experiment. Comparison of existing Greenland models. Unpublished manuscript, Laboratoire de Glaciologie et de Géophysique de l'Environnement, Saint Martin d'Hères, France. [Available online from ftp anonymous glaciog.ujf-grenoble.fr, directory /pub/EISMINT-INTERCOMP/GREENLAND.]

Schönwiese, C.-D.: 1992, *Klima im Wandel. Tatsachen, Irrtümer, Risiken*, Deutsche Verlags-Anstalt, Stuttgart, Germany.

Thomas, R.H.: 1979, 'The dynamics of marine ice sheets. *Journal of Glaciology* **24**(90), 167-177.

Van de Wal, R.S.W. and Oerlemans, J.: 1997, 'Modelling the short-term response of the Greenland ice sheet to global warming', *Climate Dynamics* **13**, 733-744.

Wilhelms, F.: 1996, 'Leitfähigkeits- und Dichtemessung an Eisbohrkernen', *Berichte zur Polarforschung* **191**, 224 pp.

(Received 30 December 1997; in revised form 18 April 2000)