Palaeogeography, Palaeoclimatology, Palaeoecology (Global and Planetary Change Section), 89 (1991): 399–412
Elsevier Science Publishers B.V., Amsterdam

The Greenland ice sheet and greenhouse warming

Philippe Huybrechts [a], Anne Letreguilly [b] and Niels Reeh [b]

[a] *Geografisch Instituut, Vrije Universiteit Brussel, Pleinlaan 2, B-1050 Brussels, Belgium*
[b] *Alfred Wegener Institut für Polar- und Meeresforschung, Columbusstrasse, Postfach 120161, D-2850 Bremerhaven, FRG*

(Received April 23, 1990; revised and accepted October 2, 1990)

ABSTRACT

Huybrechts, P., Letreguilly, A. and Reeh, N., 1990. The Greenland ice sheet and greenhouse warming. Palaeogeogr., Palaeoclimatol, Palaeoecol. (Global Planet. Change Sect.), 89: 399–412.

Increased melting on glaciers and ice sheets and rising sea level are often mentioned as important aspects of the anticipated greenhouse warming of the earth's atmosphere. This paper deals with the sensitivity of Greenland's ice mass budget and presents a tentative projection of the Greenland component of future sea level rise for the next few hundred years. To do this, the 'Villach II temperature scenario' is prescribed, and output from a comprehensive mass balance model is used to drive a high-resolution 3-D thermomechanic model of the ice sheet.

The mass balance model consists of two parts: the accumulation part is based on presently observed values and is forced by changes in mean annual air temperature. The ablation model is based on the degree–day method and accounts for the daily and annual temperature cycle, a different degree–day factor for ice and snow melting and superimposed ice formation. Under present-day climatic conditions, the following total mass balance results (in ice equivalent per years): 599.3×10^9 m^3 of accumulation, 281.7×10^9 m^3 of runoff and assuming a balanced budget, 317.6×10^9 m^3 of iceberg calving. A 1K uniform warming is then calculated to increase the runoff by 119.5×10^9 m^3. Since accumulation also increases by 32×10^9 m^3, this leads to reduction of the total mass balance by 87.5×10^9 m^3 of ice, corresponding to a sea level rise of 0.22 mm/yr. For a temperature increase larger than 2.7 K, runoff exceeds accumulation, and if ice sheet dynamics were to remain unchanged, this would add an extra amount of 0.8 mm/yr to the worlds' oceans.

Imposing the Villach II scenario (warming up to 4.2 K) and accumulating mass balance changes forward in time (static response) would then result in a global sea level rise of 7.1 cm by 2100 AD, but this figure may go up to as much as 40 cm per century in case the warming is doubled. In a subsequent dynamic model run involving the ice flow, the ice sheet is found to produce a counteracting effect by dynamically producing steeper slopes at the margin, thereby reducing the area over which runoff can take place. This effect is particularly apparent in the northeastern part of the ice sheet, and is also more pronounced for the smaller temperature perturbations. Nevertheless, all these experiments certainly highlight the vulnerability of the Greenland ice sheet with respect to a climatic warming.

Introduction

It is generally believed that the increasing atmospheric concentrations of radiative active trace gases (carbon dioxide, methane, etc..) will lead to a warmer climate in the future. Model predictions on the magnitude of this warming, in space as well in time, still differ to a wide extent, but according to the 'Villach II scenario', there is a 65% chance that in the year 2100 the global mean temperature will be between 2.7 and 5.7 °C higher than the pre-industrial level, with a higher value more likely in polar latitudes. At present, the anthropogenic signal might be of the order of 0.5 °C. Although the 'greenhouse warming' has not yet been 'detected' by observations in a statistically significant way, global temperature nevertheless shows a marked warming since the mid-1970's, and the

0921-8181/91/$03.50

3 warmest years between 1864 and 1984 have all occurred in the 1980's (Jones et al., 1986).

A frequently mentioned aspect of the greenhouse problem are changes in land ice volume and the associated consequences for global sea level. World-wide mean sea level has risen by 10–25 cm over the last hundred years (Barnett, 1983; Gornitz and Lebedeff, 1987; Peltier and Tushingham, 1989), and it is conceivable that larger changes will occur when global temperature rises by a few °C! At present, the largest contribution is thought to originate from thermal expansion of ocean water (Wigley and Raper, 1987), and in spite of the small ice volumes involved, the retreat of mountain glaciers and small ice caps could have contributed another 2–4 cm (Meier, 1984). This is because they have much shorter response times, and on a time scale of decades to centuries, they may even play a dominant role.

The largest amount of ice is stored in the Antarctic and this is also the ice sheet that has received most attention. Mercer (1978) has pointed out the fact that the West Antarctic ice sheet may be inherently unstable in such a way that a moderate warming may lead to a runaway situation in which the major part of the ice sheet disintegrates, involving changes in global sea level of up to 5 m in a few centuries. Later modelling, however, suggest that this view is probably too dramatic and that a sudden collapse of the West Antarctic ice sheet on these time scales is unlikely to happen (Van der Veen, 1986). A similar conclusion was also reached in Huybrechts and Oerlemans (1991), involving detailed calculations with a mass balance model coupled to a 3-D ice sheet model. These model experiments also indicated that as long as the warming does not exceed 8°C, the total surface mass balance of the Antarctic ice sheet would actually be *larger* than today. Imposing the Villach II scenario in a similar fashion as described in this paper would then lead to a reduction in global sea-level of 6 cm by 2100 AD. This response is mainly due to the very low air temperatures prevailing over the Antarctic and the associated low moisture content of the atmosphere. In case of a temperature rise, the increase in accumulation rates then far outweighs the increase in runoff.

The second largest ice mass in the world, and the largest ice body on the northern hemisphere, is the Greenland ice sheet. It has a volume of around 2.8 million km^3 of ice and contains about 9.5% of the earth's fresh water resources. The area of the ice sheet is 1.7 million km^2 and the average thickness about 1650 m. The largest thickness observed (by radio echo sounding) is 3420 m. The climate of Greenland, however, is very different from the Antarctic, with a temperature difference of 10–15°C in the annual mean. Whereas there is at present hardly any ablation on the Antarctic ice sheet due to the very cold conditions, the Greenland ice sheet is located in a region where temperatures are high enough to initiate widespread summer melting. As a consequence, at elevations below 1000 m in the north and 1600–1800 m in the southwest, the annual ablation exceeds the accumulation and a negative surface budget results. Mean ablation rates throughout the ablation zone are typically of the order of 0.5–1 m/yr, but may go up to 5 m/yr along some of the protruding glaciers at the western margin. Precipitation rates show a general decreasing trend from south to north, from about 2.5 m/yr in southeastern coastal areas to less than 0.15 m/yr in interior northeast Greenland. Estimates suggest a 60/40% ratio between ice mass lost by runoff and iceberg calving, although that only about 40% of the calving flux is actually based on concrete data from outlet glaciers, such as ice velocities and thicknesses (Weidick, 1985; Reeh, 1989).

The present state of balance is not really well known, although there seems to be no evidence that the ice sheet is too far out of equilibrium. This is confirmed by direct observations of the change in surface elevation along the EGIG line in central Greenland (Mälzer and Seckel, 1975). These investigations show a small lowering of the ice sheet surface in the ablation zone, but also a small increase farther up in the accumulation zone. A similar thickening above the equilibrium line, though of somewhat larger magnitude and possibly in the ablation zone as well, was also reported recently by Zwally et al. (1989) south of 72° N, but inferences (based on satellite altimetry) appear to remain speculative. In spite of the uncertainties involved, however, it is very likely that a climatic warming will lead to increased melting and several estimates on the possible contribution to world-wide sea level have been made. Bindschadler (1985), for example, estimated that if the predictions of the magnitude of the CO_2-induced climatic warming hold true, the mass loss from the Greenland ice sheet is likely to increase to between 2 to 4 times its present rate. This would make the ice sheet disappear in 2–5 ka. In this estimate, however, a likely effect of a climatic warming, that is, increased precipitation over the ice sheet and a likely increase of the meltwater that refreezes on the ice sheet, was not considered. Both effects will decrease the rate of disintegration of the ice sheet.

Other studies have concentrated more specifically on the sensitivity of the total mass budget. In an analysis of measurements obtained during the EGIG expeditions, Ambach and Kuhn (1989) came to the conclusion that a 1° C warming would lead to a sea level rise of 0.35 mm/yr. This value is very close to the 0.37 mm/yr obtained by Oerlemans (1989b). The latter value was based on calculations with an energy and mass balance model of the snow/ice surface. Due to the amount of calculations involved, however, Oerlemans only investigated the sensitivity of four mass balance profiles, and subsequently extrapolated results to obtain a value for the entire ice sheet. He also did not consider changes in accumulation.

In this paper a different method is followed. Explicit calculations with detailed models of the mass balance components (accumulation and runoff) are performed on a high-resolution grid (with 20 km spacing) covering the entire Greenland ice sheet. This should lead to a much better representation of the total mass budget and also allows use to be made of the best and most detailed data currently available. We also included ice dynamics. Although the response of the ice sheet on a short time scale (say, less than 100 years) can be considered in first approximation to be static, i.e., the flow does not react, this is probably not justified for a somewhat longer time integration. In that case, changing ice flow may start to affect the mass balance in a significant way and ice mechanics have to be considered. That is why in a subsequent stage a comprehensive 3-D thermomechanic ice sheet model has been employed, that was originally developed for the Antarctic ice sheet (Huybrechts, 1990, 1991), but has been adapted to simulate conditions on Greenland.

The mass balance model

In the present approach, the components of the mass balance and its perturbations are parameterized in terms of temperature, that is the principal forcing parameter. Although runoff and accumulation result from quite complex processes, involving the general circulation pattern in the atmosphere and the energy balance at the ice sheet surface, such an approach is necessary to keep calculation times on the fine grid in use (20 km spacing) within acceptable bounds. As will be demonstrated, however, this approach yields very reasonable results.

Accumulation

Precipitation is a difficult process to model. It not only depends on cyclonic activity, depression paths and moisture content, but accumulation rates over the Greenland ice sheet are also determined by such factors as temperature (since colder air can carry less precipitable moisture), continentality and orientation of the ice sheet surface with respect to prevailing winds (orographic effect). These factors are much to complicated to be parameterized. A better approach here is to use presently observed data and to perturb the resulting field by a prescribed change in a meteorological variable such as temperature. This is justified, because the greenhouse warming calculations do not involve really drastic changes of the ice sheet geometry. In the model, the annual precipitation has been digitized on the 20 km grid from a map published in Ohmura and Reeh (1990). This map includes all measurements up to date and is the best currently available. Given the present surface elevation distribution, these reference accumulation values are then first identified with a mean annual air temperature value TMA (0). The following relation has been derived from a recent compilation of temperature data, also by Ohmura (1987):

$$TMA = 49.13 - 0.007992H_{\mathrm{sur}} - 0.7576Lat + T_{\mathrm{for}}$$

$$\text{if } H_{\mathrm{sur}} \leq 300 \times \frac{Lat - 65}{15} = H_{\mathrm{inv}} \text{ then}$$

$$TMA = 49.13 - 0.007992H_{\mathrm{inv}} - 0.7576Lat + T_{\mathrm{for}} \tag{1}$$

where H_{sur} is surface elevation, Lat is the latitude in °N and T_{for} is the temperature forcing [zero for $TMA(0)$]. H_{inv} is the altitude, below which a temperature inversion is usually observed and it depends on latitude. It is accounted for in the parameterisation by keeping the gradient below H_{inv} at zero. TMA is expressed in °C. The accumulation rate is then prescribed to vary by 5.3% for every degree change in mean annual air temperature according to:

$$Acc(t) = Acc(0) \times 1.0533^{[TMA(t) - TMA(0)]} \tag{2}$$

This 5.3% figure is suggested by comparing layer thicknesses with the corresponding $\delta^{18}O$ value in Greenland ice cores and converting the result in a temperature change. For a 5 °C temperature shift, that may result in the model from a change in surface elevation and/or background temperature, the corresponding accumulation rate changes by around 30%. Although this relation has been derived from past colder climates, it is expected to be valid for the largest part of the Greenland ice sheet, where year-round temperatures are generally well below freezing, and will remain so even in case of a moderate warming. Such a dependence is also observed over the Antarctic, where there appears to be a particularly strong correlation between the precipitation rate and the temperature of snow formation (Fortuin and Oerlemans, 1990). Note, however, that the present approach does not account for changes in the circulation pattern and also neglects the amount of precipitation falling as rain. The latter is not considered a serious constraint, since only a small fraction of the yearly precipitation is involved anyway (during the summer months only) and a large part of the rainfall may be expected to refreeze into superimposed ice, before it can run off to the coast. Also, most of the rain probably falls on the ice-free areas in the lower reaches of the continent.

Runoff

Although melting actually depends on the energy balance at the surface, it is assumed that air temperatures and snow accumulation may be used to determine the annual melt rate, that will be calculated by means of a degree-day model. As shown by by Braithwaite and Olesen (1989), there is a high correlation between positive degree days and melt rate at west Greenland ice margin locations. The amount of positive degree days, that represents a melt potential, is then calculated in several steps. The average annual temperature cycle is assumed to be a cosine function, with amplitude $TMJ - TMA$ and a period A of 1 year, where TMJ is the mean July temperature. TD is the daily temperature:

$$TMJ = 30.78 - 0.006277H_{\mathrm{sur}} - 0.3262Lat + T_{\mathrm{for}} \tag{3}$$

$$TD = TMA + (TMJ - TMA)\cos\left[\frac{2\pi t}{A}\right] \tag{4}$$

Equation 3 has been obtained from temperature data on the ice sheet as listed in Ohmura (1987). Ohmura also included data from ice-free stations on the land, and made a difference between the east and west slopes of the ice sheet, but the data from the ice sheet alone are too scanty to justify such a differentiation.

A model is then developed to estimate positive degree-days in terms of TMA, TMJ, and a stochastic term, that accounts for temperature variations from the regular, long-term annual cycle. This is necessary, since even in case the temperature of the warmest summer month is below the freezing point, there are likely to be days when the temperature exceeds the zero-degree mark. Also random temperature fluctuations are likely to cause positive temperatures in the spring or in the fall, even though the average temperatures at these times may be way below freezing. Although the daily cycle contains a large deterministic component, it can also be accounted for by this stochastic term. The combined effects of all these terms are then approximated by a statistic, that is normally distributed and centered on the curve given by Eq. 4, and having a standard deviation σ. This leads to:

$$PDD = \frac{1}{\sigma\sqrt{2\pi}} \times \int_0^A \left\{ \int_0^{TD+2.5\sigma} T \times \exp\left[\frac{-(T-TD)^2}{2\sigma^2}\right] \mathrm{d}T \right\} \mathrm{d}t \qquad (5)$$

Values for the amount of positive degree-days are then obtained on the grid by numerical integration. We used a time step of 1 month and temperature steps $\mathrm{d}T$ of 2° C. σ is 5° C. More details on this melt rate model are presented in Reeh (1990b). Figure 1 shows resulting *PDD*'s for various locations on the ice sheet, for present conditions and in case of a climatic warming of up to 8° C.

The snow-and-ice-melt model is essentially similar to the model described by Braithwaite and Thomsen (1984), except that rainfall is neglected, i.e. precipitation is assumed to occur as snowfall, only. The available positive degree days

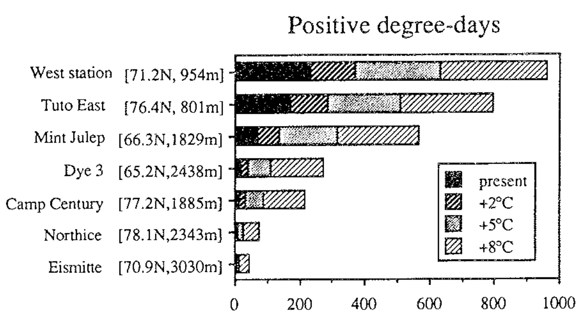

Fig. 1. Calculated amount of positive degree-days for various stations on the Greenland ice sheet.

(*PDD*) as calculated by means of Eq. 5 are used to melt snow and ice in the following order:

(1) snow is melted with a degree-day factor of 0.003 m ice melt/*PDD*, and the meltwater percolates into the snowcover and refreezes to form superimposed ice. Runoff occurs when the superimposed ice layer exceeds a given fraction (*pmax*) of the yearly snowfall. *pmax* has been set to 0.6

(2) superimposed ice is melted with a higher degree–day factor of 0.008 m ice melt/*PDD*. (This higher value is essentially related to a lower albedo for bare ice as compared to snow).

(3) glacier ice is melted.

This process may stop at any of the stages 1–3, depending on the magnitude of *PDD*. In case stage 3 is not reached, the amount of superimposed ice remaining at the end of the melting season (*SIR*) is retained to include the warming effect due to latent heat release for the upper boundary condition in the temperature calculations.

The ice sheet model

Basic description

The ice sheet model treats ice flow and its thermodynamics by solving the full set of coupled thermomechanical equations on a fine mesh in three dimensions. It is time-dependent and also includes the response of the underlying bedrock to a changing ice load. There is free interaction between climatic input and ice thickness, so that the entire geometry is internally generated. A full account of all the mathematical equations and numerical techniques governing the model will not be presented here, but can be found e.g. in Huybrechts and Oerlemans (1988), where a flowline version was employed to investigate the response of the East Antarctic ice sheet to a glacial–interglacial transition.

In short, the basic equations that are solved in the model are conservation equations for mass and heat:

$$\frac{\delta H}{\partial t} = -\nabla \cdot (\vec{v}H) + M \qquad (6)$$

$$\frac{\partial T}{\partial t} = \frac{k}{\rho c_p} \nabla^2 T - \vec{V} \cdot \nabla T + \frac{\Phi}{\rho c_p} \quad (7)$$

where H is ice thickness, $\vec{v}$ the depth-averaged horizontal velocity field, M the mass balance and t time (in years). The thermodynamic equation accounts for vertical heat conduction, three-dimensional advection and heat generation by internal deformation. Here, T is temperature, ρ ice density (910 kgm^{-3}), V three-dimensional ice velocity, Φ layer dissipation, and k and c_p are temperature dependent thermal conductivity and specific heat capacity, respectively.

A basic assumption is that the ice flows in direct response to pressure gradients set up by gravity. Ice deformation is assumed to result from shear strain and longitudinal deviatoric stresses are disregarded. These are the usual assumptions, and expressions for the horizontal velocity components can be derived by substituting equations for the shear stress distribution $\tau(z)$ in the flow law, that is of 'Glen-type' with exponent $n = 3$, and integrating the result with respect to the vertical. This yields:

$$\vec{\tau}(z) = -\rho g(H + h - z) \nabla(H + h) \quad (8)$$

$$\begin{aligned} \vec{v}(z) - \vec{v}(h) &= -2(\rho g)^n [\nabla(H + h) \cdot \nabla(H + h)]^{(n-1)/2} \\ &\quad \times \nabla(H + h) \int_h^z A(H + h - z)^n \, \mathrm{d}z \end{aligned} \quad (9)$$

with basal boundary condition $\vec{v}(h) = 0$. h is bed elevation. Note that since A depends on ice temperature (and therefore on position), Eq. 9 has to be evaluated numerically, given the temperature distribution.

The temperature dependence of the flow law coefficient $A(T)$, where T is corrected for pressure melting, is given by an Arrhenius relationship, and accounts for higher activation energies for ice temperature above $-10\,^\circ\text{C}$ (Paterson, 1981):

$$A(T^*) = m \cdot a \exp\left\{\frac{-Q}{RT^*}\right\} \quad (10)$$

$T^* < 263.15$ K $\quad a = 1.14\ 10^{-5}$ $\text{Pa}^{-3}\ \text{yr}^{-1}$

$Q = 60$ $kJmol^{-1}$

$T^* \geq 263.15$ K $\quad a = 5.47\ 10^{10}$ $\text{Pa}^{-3}\ \text{yr}^{-1}$

$Q = 139$ $kJmol^{-1}$

The enhancement factor m is a tuning parameter to obtain a correct height-to-width ratio and has to be evaluated experimentally. Its value is generally comprised between 1 and 10 and may be thought of as an implicit way to include the softening effect due to such factors as impurity content, meltwater and crystal fabric.

A second integration with respect to the vertical yields the mean horizontal ice mass flux:

$$\begin{aligned} \vec{v}H &= -2(\rho g)^n [\nabla(H + h) \cdot \nabla(H + h)]^{(n-1)/2} \\ &\quad \times \nabla(H + h) \\ &\quad \times \int_h^{H+h} \int_h^z A(H + h - z)^n \, \mathrm{d}z \, \mathrm{d}z \end{aligned} \quad (11)$$

where vertical motion, as a result of accumulation and vertical strain, is calculated from the incompressibility condition:

$$w(z) - w(h) = -\int_h^z \nabla \cdot \vec{v}(z) \, \mathrm{d}z \quad (12)$$

The model also takes into account bedrock adjustments in response to changing ice loading. The steady state deflection of the lithosphere, w is given by considering local isostatic equilibrium:

$$w = \frac{\rho}{\rho_m} H \quad (13)$$

where ρ_m is mantle density (3300 kgm^{-3}). In order to calculate the time evolution, a viscous asthenosphere model is employed (Oerlemans and Van der Veen, 1984):

$$\frac{\partial h}{\partial t} = D_a \nabla^2 (h - h_0 + w) \quad (14)$$

where h_0 is the undisturbed bed elevation in case the ice sheet is removed and D_a a diffusion coefficient (5×10^7 m^2 year^{-1}).

All calculations are performed on the same grid. It has a horizontal spacing of 20 km. With 14 layers in the vertical for the ice flow and thermodynamic calculations, this adds up to a total of somewhat less than 200,000 gridpoints. We used a scaled vertical coordinate in order to avoid boundary problems in the temperature calculations. The upper layer has a thickness of

10% of the ice thickness, the lowermost layer is 2%. This allows a much more refined description of deformation in the basal layers, where the shear concentrates. The finite difference equations are then solved numerically by the Alternating Direction Implicit method. This approach has the advantage that time steps can be taken up to 10 times larger than the more conventional explicit integration schemes. In the model, time steps are usually 4 yr for the ice thickness continuity calculations and 40 yr for the more slowly varying components, such as the bed adjustment and thermodynamic calculations. The program code has been extensively vectorized and is running on a CRAY-2 computer. On this machine, the model takes around 20 min CPU time for a 10,000-yr integration. In the greenhouse warming experiments discussed in this paper, that involve rapid changes in climatic input, the time step is 1 yr for all components, however.

Boundary conditions

Model inputs are bed topography, surface temperature, mass balance, thermal parameters and an initial state (that may be zero ice thickness but is taken here as the observed state). The model then essentially outputs the time-dependent three-dimensional ice sheet geometry, and the fully coupled velocity and temperature fields. Datasets for bedrock, surface topography and ice thickness have been digitized on the 20 km grid from a number of sources. Surface elevation and ice thickness data originate from a large number of flight tracks of an airborne radio-echo-sounding project carried out in the seventies by the Electromagnetic Institute, Technical University of Denmark. Whereas these data are the only source for the ice thicknesses, data for the surface elevation have been supplemented by satellite measurements from ETO (US military) and were further checked against the map published by Ohmura (1987). Resulting maps for ice thickness and bedrock topography are however published elsewhere (Letreguilly et al., 1990).

In the model, the present coastline serves as a model constraint and details of the calving physics are not taken into account. This means that the ocean is considered an infinite sink for ice, and in case the ice sheet reaches open water, ice thickness is automatically set to zero. This approach to modelling the boundary ice is admittedly the easiest one. Together with the fact that some of the outlet glaciers have lateral scales even smaller than the gridpoint distance, it may lead to a local underrepresentation of the calving flux, but an alternative seems difficult to construct.

In the central zone of the ice sheet (the dry-snow zone) the surface temperature is to a good approximation equal to the mean annual air temperature. However, in case melting occurs, refreezing of percolating meltwater will cause the near-surface and firn layers to warm up. This effect is taken into account in the upper boundary condition ('10 m- temperature') for the thermodynamic calculations. Data from all available Greenland ice sheet stations have been used, and the following relation has been obtained by plotting the amount of refrozen meltwater against the difference between the snow temperature at 10–15 m depth and the corresponding mean annual air temperature (Reeh, 1990b):

$$T_{\text{surface}} = TMA + 26.6\, SIR \quad r^2 = 0.75 \tag{15}$$

SIR is the amount of superimposed ice (if any) remaining at the end of the melt season. The basal boundary condition for the temperature calculations assumes a geothermal heat flux of $1 \text{HFU} = 42\ \text{mW/m}^2$, which is a value common for Precambrian rock.

The reference run

Studying changes in ice sheet geometry requires that a reference state is defined. Starting from the observed ice sheet and running the model forward in time with a prescribed temperature forcing is a much less meaningful way to proceed, because these input data are most likely not in full internal equilibrium with the model physics. This may be due to shortcomings in the description of ice mechanics, but may also indicate that the present ice sheet is just not in steady state, it is probably a combination of both. As a consequence, it would be impossible afterwards to distinguish between the 'natural model evolution' and the 'real' ice sheet response. One way to circumvent this ambiguity is to put forward a steady state and relax the model to equilibrium. The resulting steady state interglacial reference run, although differing from the presently observed data in some detail, should then be used as input in a climatic change experiment.

To do this, the coupled velocity and temperature fields have first been run forward for 100,000 years in a diagnostic way (thus keeping ice thickness fixed). This long time is needed to approach a more or less stationary temperature field. The tuning parameter m in Eq. 10 has been set to 3. This is surprisingly close to the value to be expected, as ice core studies seem to indicate flow enhancement by this factor for Wisconsin ice, that is probably making up for the bulk of the ice in the lower shear layers (Reeh, 1985). The resulting ice sheet after another 100,000 years, in which all degrees of freedom have come into play and the model has relaxed to a stationary state, is shown in Fig. 2, where also a comparison is made with the observations. Although the ice sheet has grown somewhat, the agreement between the two ice sheets is actually quite good, in particular if it is realized that the modelled ice sheet is in steady state and it is not precisely known how the present ice sheet is still responding to increased accumulation rates since the beginning of the Holocene. In the model, its volume is larger by 13% (from 2.825 million km^3 to 3.209 million km^3 of ice) and also the area of the ice sheet has expanded somewhat, from 1.67 million km^2 to

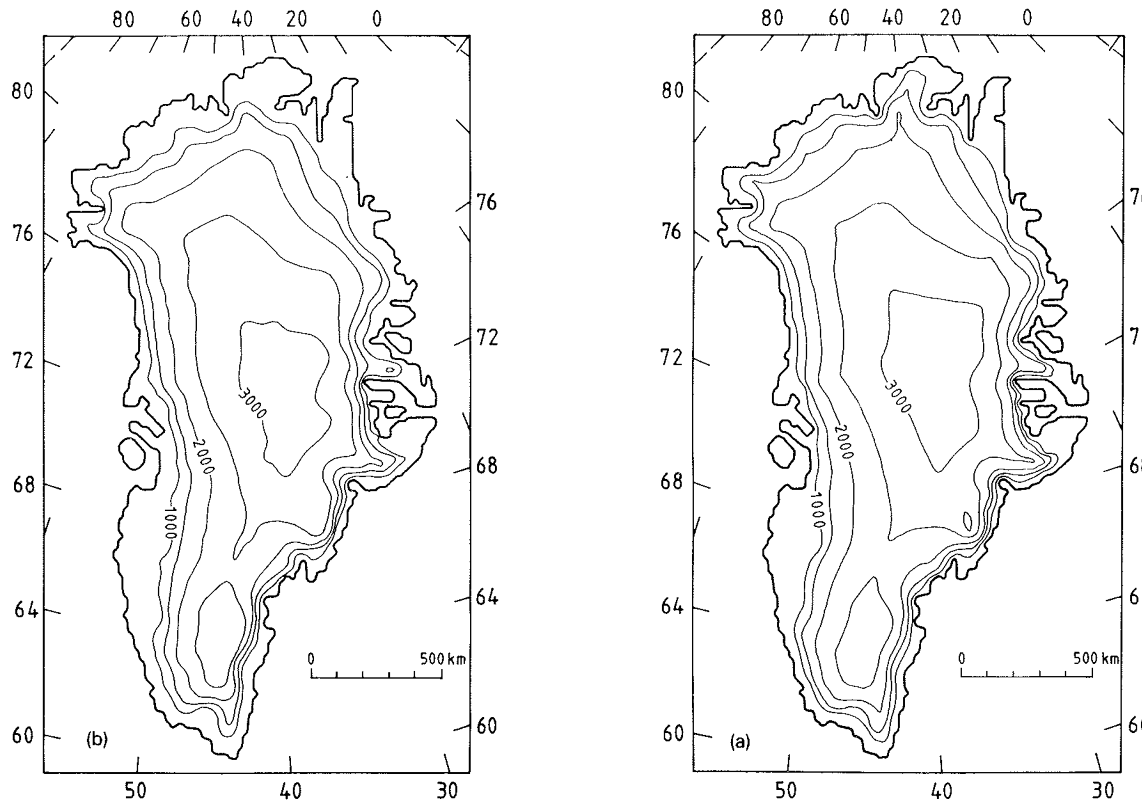

Fig. 2. The interglacial reference run as produced by the model (left panel) compared to the observations (right). Contour lines are for surface elevation (interval: 500 m).

1.78 million km^2. Closer inspection, however, revealed that this is to a large extent due to the situation in the north and northeast of Greenland, where the ice sheet now has a lobe expanding into Peary Land. Apart from a possibly poor representation of calving there, this may very well be due to an underestimation of the runoff in these areas. Nevertheless, the model does a really good job in catching the main features of the ice sheet, including more gentle surface slopes in the northern part due to smaller mass balance gradients, and a good representation of the ice-free areas, in particular in the southwest. Also the separation of the ice sheet in a northern and southern dome, that owes its existence to easy ice drainage to both east and west through the outlet glaciers Sermilik and Jakobshavn Isbrae, indicates that the model is able to characterize the main flow pattern well.

Sensitivity of the mass balance to climatic change

Using the presently observed surface elevation distribution as a boundary condition, we obtained the following values for the present total mass budget. A volume of 599×10^9 m^3 of solid precipitation (ice equivalent) is deposited on the ice sheet's surface every year. This amount of ice would lower global sea level by 1.5 mm/yr, were it not that each year a similar amount in the form of meltwater and wastage from outlet glaciers is transported back to the ocean. The runoff during the summer season is then calculated to represent an ice volume of 281.7×10^9 m^3. In case of equilibrium, accumulation should equal runoff plus iceberg calving. This requires a total calving rate of 317.3×10^9 m^3, implying that the quantity of produced calf-ice and the amount of mass lost by ablation are roughly of the same magnitude. These values appear to be in good agreement with previous estimates cited in the literature (Weidick, 1985). Assuming an area of 1.67 million km^2, the mean accumulation rate over the whole ice sheet expressed in water equivalent then equals 32.6 cm/yr, with a mean runoff of the order of 15 cm/yr.

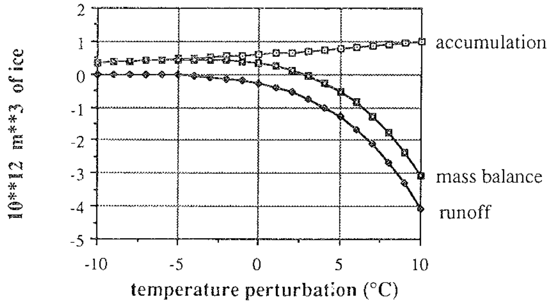

Fig. 3. Dependence of Greenland's mass budget on temperature perturbations relative to present. For a warming of more than 2.7 °C, the surface mass balance becomes negative.

The dependence of the Greenland mass budget on climatic conditions can now be investigated by forcing the mass balance model by changes in air temperature. The resulting sensitivity of the respective components is displayed in Fig. 3. Runoff appears to be negligible for temperature perturbations below −5 °C and is shown to increase exponentially with temperature. This higher order of dependence results because a warming climate will not only increase melting, but also the length of the ablation season. For instance, a 1 °C temperature rise with respect to present conditions is found to increase runoff by 40%. This corresponds to an additional mean melting rate spread out over the entire ice sheet of 65 mm/yr. Since the area of the world ocean is about 215 times larger than the area of the ice sheet, this would imply a rate of sea-level rise of 0.30 mm/yr. In view of the different methods followed, this is surprisingly close to the values found by Ambach and Kuhn (1989) and Oerlemans (1989b) cited in the introduction. Since in the model accumulation rates also rise, our value is partly set off by 0.08 mm/yr, so that 0.22 mm/yr corresponds to the sensitivity found in this study.

It is interesting to see that the total surface balance becomes negative for a (rather small) temperature increase of 2.7 °C only, i.e. at that stage the accumulation rate equals the ablation rate, and since calving still goes on, the ice sheet is likely to shrink considerably. In this case, an extra amount of 0.80 mm/yr would be added to the worlds' oceans, if ice dynamics were to re-

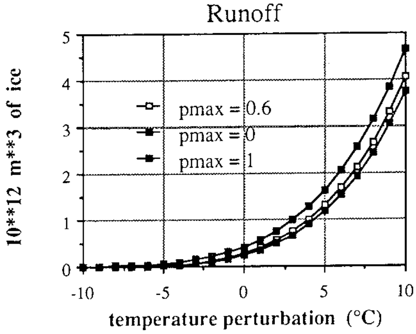

Fig. 4. The effect of superimposed ice formation on the predicted runoff. *pmax* is the fraction of the percolating snowmelt that refreezes.

main unchanged. If such a situation would persist long enough (say, a few thousand years), the ice sheet might even melt completely down, unless a new equilibrium can be established.

As displayed in Fig. 4, the process of superimposed ice formation, that is taken into account in the model by refreezing a specified part of the percolating snowmelt (*pmax*), is also of some importance. For present conditions, the amount of ablation, expressed as a fraction of total accumulation, may vary due to this process from 40% (the total snowcover first transforms into superimposed ice, $pmax = 1$) to 69% (no superimposed ice formation, $pmax = 0$). However, the general shape of the curve is well preserved and, fortunately, the sensitivity of the runoff (that is of real importance here) does not depend too heavily on the value of *pmax*, that has been set to 0.6 on an intuitive basis (Reeh, 1990b).

Figure 5 demonstrates how the surface mass balance changes for various values of the imposed climatic warming. In the interior regions and for the moderate warming case of 4° C, the mass balance increases, but runoff also becomes important here if this warming is doubled. In the ablation area, the exponential decrease of mass balance for warmer conditions is striking, and the mass budget may easily reach values of between −5 and −10 m/y. As a consequence, the balance gradient steepens in a warmer climate, which has a significant influence on the ice flow, a point discussed further below.

Response of the ice sheet to a climatic warming

The temperature scenario employed to force the mass balance model is taken from the Villach II meeting in the fall of 1987 (Oerlemans, 1989a). It is shown in Fig. 6 and predicts the temperature to rise by 4.2° C by the year 2100, starting from 1850, which is considered to be an undisturbed reference state. The probability that this figure is correct within 1.5° C is estimated to be 67%. This figure applies to the global mean surface temperature, however. As sensitivity studies with GCM's seem to point to a larger response in polar latitudes (e.g. Manabe and Stouffer, 1980), an alternative experiment was also carried out with a temperature rise twice as much (the 'high scenario'). Furthermore, temperature perturbations are assumed to be uniform through the year and independent of lati-

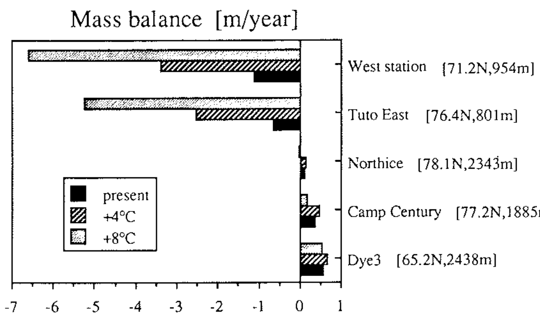

Fig. 5. Surface mass balance (in m/yr ice equivalent) for present conditions and in a warmer climate for some selected Greenland stations.

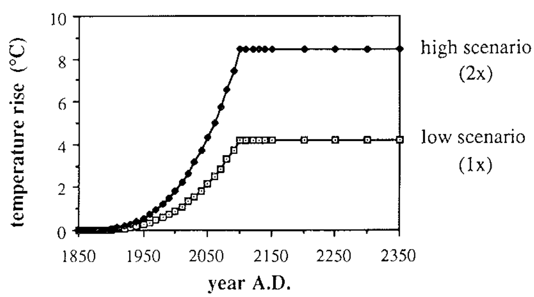

Fig. 6. The temperature scenario's ('*high*' and '*low*') used to force the mass balance and ice dynamics models. The lower curve (*1* ×) is the 'Villach II scenario', the upper curve (*2* ×) corresponds to a temperature rise twice as much.

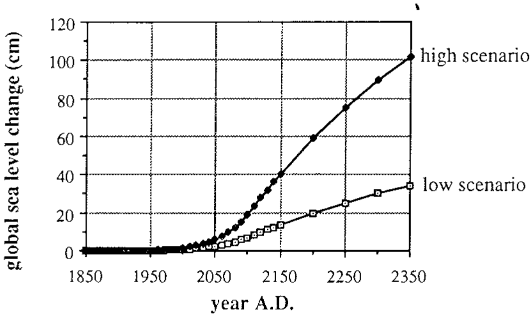

Fig. 7. Response of the ice sheet expressed in global sea level changes. In this experiment, ice dynamics are not considered (‘*static effect*’).

tude. In the interpretation of the results, we will distinguish between the ‘static effect’, that can be studied with the mass balance model solely, and the ‘dynamic effect’, in which also ice dynamics come into play.

Static response

In view of the long response time scales of large ice sheets, we first consider a ‘static’ experiment, i.e. the present ice sheet is assumed to be in equilibrium, and changes with respect to the present mass balance are accumulated forward into time. In these calculations, changing surface elevations have been taken into account as a contribution to local temperature changes (although this contribution is very small) and the maximum amount of ice that can melt away at any location is set equal to the present ice thickness. The results in terms of sea-level equivalent are shown in Fig. 7. For the low warming case, in which temperatures go up to 4.2 °C in 2100 AD, the Greenland ice sheet may contribute 0.7 cm by the year 2000, 7.1 cm by 2100 and even up to 34 cm in the year 2350, i.e. after 500 years since the onset of the warming. The latter number is very speculative, of course, since we don’t know what will happen after 2100, given the fact that such a warming occurs in the first place, but these figures certainly give an indication of the order of magnitude to be expected. In the high scenario, the corresponding figure for 2350 would even be 102 cm, i.e. the Greenland ice sheet has lost as much as 15% of its volume by then!

Dynamic response

Considering these important changes in ice thickness, the ice sheet is expected to show a dynamical response too. Increased balance gradients, in particular near and below the equilibrium line, can only be matched by larger ice fluxes and this should result in steeper surface gradients. This is also what happens and it is a counteracting effect. More ice is transported from above the equilibrium line to account for the increased loss of ice below this line, resulting in a thickening in the ablation area. As a result, the total area over which runoff can take place is reduced dynamically. This feature appears to be particularly effective in the north and northeast of Greenland, nowadays characterized by relatively gentle surface slopes. Calculated curves for sea-level rise are displayed in Fig. 8, where they are compared with a static control run, in which the *modelled* ice sheet topography of the reference run is used as a boundary condition (this makes the static response larger by some 20%). It appears that the dynamic effect is certainly not negligible, and introducing ice dynamics is found to reduce the response by nearly 50% for the low warming scenario, and to between 12% and 40% for the high scenario. Corresponding numbers for sea-level rise are now 0.37 cm by the year 2000, 4.8 cm in 21(0 and up to 26 cm at 2350 AD (low scenario), but the melting

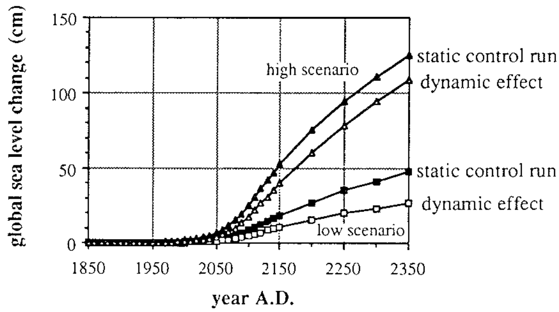

Fig. 8. Response of the ice sheet in case ice dynamics come into play. Also shown are results from a static control run.

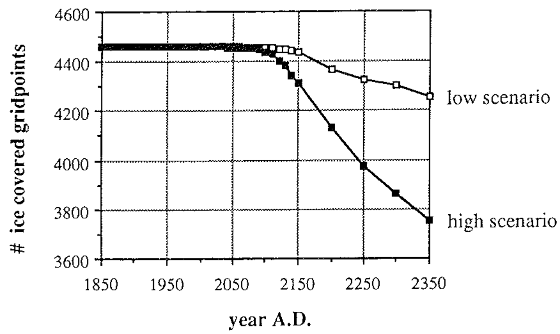

Fig. 9. Evolution of the ice sheet area expressed in number of (#) ice covered gridpoints. 1 gridpoint represents an area of 400 km^2.

rate may go up to 40 cm/century after 2100 if this warming is doubled!

Figure 9 shows the time evolution of the ice sheet area in these runs. Resulting ice sheet topographies by the year 2500 are presented in Fig. 10. As can be seen on the latter figure, interior regions appear to be relatively unaffected by the warming and, not surprisingly, margin retreat is most pronounced in those areas that are already now fully land-based and the 'calf-ice buffer' is absent. This implies that the height-to-width ratio increases in a warmer climate. One of the weakest part of the Greenland ice sheet then appears to be the saddle connecting the southern dome with the main ice sheet and this is also the place where the ice sheet is about to be separated in two smaller ice masses.

At 2500 AD, the ice sheet has by far not reached equilibrium, and an intriguing question at this stage was to find out whether the ice sheet could actually survive these higher temperatures, once increased melting and surface lowering have set in, especially in the light of the positive feedback between mass balance and elevation. In order to investigate this point in somewhat more detail, we finally conducted two additional runs in which the integration was carried further into

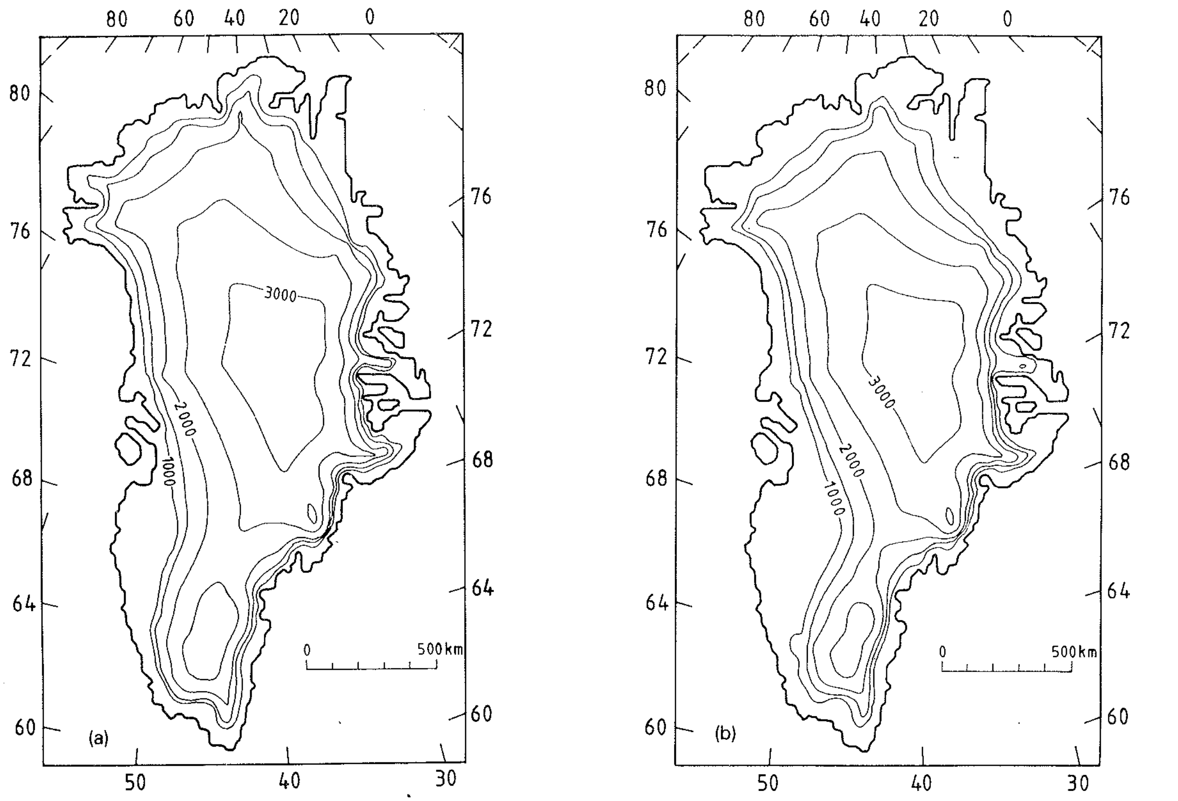

Fig. 10. Resulting ice sheet geometries after 650 model years have elapsed (i.e. at the year 2500) for the low (left) and high greenhouse warming scenario (right).

time. Surprisingly, this brought to light that the ice sheet may still be able to settle down to a new stationary state (though with about half of the present volume) for the low temperature rise (+4.2 °C), and this in spite of the fact that such a warming results in a largely negative mass balance for *present* surface elevations. This once more demonstrates the power of the dynamic effect as described above. For the high warming case, however, the only stationary state left is the one with no ice and a complete meltdown resulted in less than 5000 years. This type of behavior may have had far-reaching consequences for the glacial history of the ice sheet, and is particularly interesting in the light of ongoing speculations about the size of the Greenland ice sheet during the Last Interglacial (Koerner, 1989; Reeh, 1990a; Letreguilly et al., 1991). During that period, temperatures were several degrees higher than today, and partial or even complete melting of the ice sheet could have accounted for a significant part, if not all, of the 6 m higher-than-present sea level in the Eemian interglacial.

Conclusion

In this paper, we discussed the sensitivity of Greenland's mass budget and the response of the ice sheet in terms of future greenhouse warming. To do this, detailed models for the mass balance components were developed, that were in a subsequent stage coupled with a 3-D thermomechanic model for ice dynamics, allowing us to make use of the best data currently available. The results of the numerical simulations can be summarized as follows:

(1) For present conditions, an amount of 545.3 km^3 water equivalent is deposited on the ice sheet's surface very year. Of this amount, 47% is found to run off during the summer season.

(2) A 1 °C temperature increase would increase runoff by 40%, and since accumulation rates also go up, this would lead to a rise in the world-wide sea-level stand of 0.22 mm/yr.

(3) For a temperature rise of 2.7 °C only, runoff equals accumulation and a balanced surface budget results.

(4) Imposing the Villach II scenario, the model predicts a 5 cm sea level rise in 2100 AD due to changes on the Greenland ice sheet and up to 30 cm after 500 years since the onset of the warming (2350 AD).

It is interesting to compare the latter numbers with results obtained in a similar study conducted on the Antarctic ice sheet (Huybrechts and Oerlemans, 1991). A climatic warming on the Antarctic ice sheet would probably *lower* global sea level by a comparable amount, so that the combined effect of both ice sheets could be negligible for at least one century to come.

A number of uncertainties remain, however. In real world, the response of the ice sheet will also depend on such factors as snow surface albedo, meltwater transport in the firn layer, variability in the calving speed of outlet glaciers etc..., and these effects have only been dealt with in a schematic and implicit way. Also, model predictions on the magnitude of the warming still differ widely. Although really to be considered as an upper bound, experiments in which the temperature rise is taken twice as much, for instance, indicate that the sea-level rise may go up to as much as 40 cm/ century. This would result in a complete meltdown of the Greenland ice sheet in less than 5000 years. As a final conclusion we can say that our experiments have clearly demonstrated the sensitivity of the Greenland ice sheet with respect to a warming climate, and consequently, in case temperatures really go up, a significant wastage of the ice sheet has to be expected.

Acknowledgements

This study benefitted greatly from the excellent computing facilities (CRAY 2) offered by the Alfred Wegener Institut für Polar- und Meeresforschung, Bremerhaven (F.R.G.). We thank A. Ohmura (ETH Zürich) for communicating data that were unpublished at the time of the study. Useful discussions with Hans Oerlemans were also much appreciated. During this research, Philippe Huybrechts was supported by the Belgian National fund for Scientific Research (NFWO).

References

Ambach, W. and Kuhn, M., 1989. Altitudinal shift of the equilibrium line in Greenland calculated from heat balance characteristics. In: J. Oerlemans (Editor), Glacier Fluctuations and Climatic Change, Kluwer, Dordrecht, pp. 281–289.

Barnett, T.P., 1983. Recent changes in sea level and their possible causes. Clim. Change, 5: 15–38.

Bindschadler, R.A., 1985. Contribution of the Greenland ice cap to changing sea level: present and future. In: Glaciers, Ice Sheets and Sea Level. Effects of a CO_2-induced Climatic Change. Nat. Acad. Press, Washington, DC, pp. 258–266.

Braithwaite, R. and Olesen, O.B., 1989. Calculation of glacier ablation from air temperature, west Greenland. In: J. Oerlemans (Editor), Glacier Fluctuations and Climatic Change. Kluwer, pp. 219–233.

Braithwaite, R. and Thomsen, H.H., 1984. Runoff conditions at Paakitsup Akuliarusersua, Jakobshavn, estimated by modelling. Grønl. Geol. Unders. Gletscher-Hydrol., Medd. 84 (3), 22 pp.

Fortuin, J.P.F. and Oerlemans, J., 1990. Parameterization of the annual surface temperature and mass balance of Antarctica. Ann. Glaciol., 14: 78–84.

Gornitz, V. and Lebedeff, S., 1987. Global sea-level changes during the past century. In: D. Nummedal, O.H. Pilkey and J.D. Howard (Editors), Sea-level Change and Coastal Evolution. SEPM Publ., 41: 3–16.

Huybrechts, P., 1990. The Antarctic ice sheet during the last glacial-interglacial cycle: a 3-D model experiment. Ann. Glaciol., 14: 115–119.

Huybrechts, P., 1991. A 3-D model for the Antarctic ice sheet: a sensitivity study on the glacial-interglacial contrast. Clim. Dyn., in press.

Huybrechts, P. and Oerlemans, J., 1988. Evolution of the East Antarctic ice sheet: a numerical study of thermomechanical response patterns with changing climate. Ann. Glaciol., 11: 52–57.

Huybrechts, P. and Oerlemans, J., 1991. Response of the Antarctic ice sheet to future greenhouse warming. Clim. Dyn., in press.

Jones, P.D., Wigley, T.M.L. and Wright, P.B., 1986. Global temperature variations between 1861 and 1984. Nature, 322: 430–434.

Koerner, R.M., 1989. Ice core evidence for extensive melting of the Greenland ice sheet in the Last Interglacial. Science 244: 964–968.

Letreguilly, A., Huybrechts, P. and Reeh, N., 1990. Steady state characteristics of the Greenland ice sheet under different climates. J. Glaciol., in press.

Letreguilly, Reeh, N. and Huybrechts, P., 1991. The Greenland ice sheet through the last glacial–interglacial cycle. Palaeogeogr., Palaeoclimatol., Palaeoecol. in press.

Manabe, S. and Stouffer, R.J., 1980. Sensitivity of a global model to an increase of CO_2 concentration in the atmosphere. J. Geophys. Res., C90: 11,689–11,707.

Mälzer, H. and Seckel, H., 1975. Das geometrische Nivellement über das Inlandeis—Höhenänderungen zwischen 1959 und 1968 im Ost-West-Profil der EGIG. Z. Gletscherkd. Glazialgeol., 11: 245–252.

Meier, M.F., 1984. Contribution of small glaciers to global sea level. Science, 226: 1418–1421.

Mercer, J.H., 1978. West Antarctic ice sheet and CO_2 greenhouse effect: a threat of disaster. Nature, 271: 321–325.

Oerlemans, J., 1989a. A projection of future sea level. Clim. Change, 15: 151–174.

Oerlemans, J., 1989b. Greenland ice sheet and greenhouse warming. Inst. Meteorol. Oceanog., Univ. Utrecht (unpublished).

Oerlemans, J. and Van der Veen, C.J., 1984. Ice Sheets and Climate. Reidel, Dordrecht, 217 pp.

Ohmura, A., 1987. New temperature distribution maps for Greenland. Z. Gletscherkd. Glazialgeol., 23 (1): 1–45.

Ohmura, A. and Reeh, N., 1990. New precipitation and accumulation maps for Greenland. J. Glaciol., in press.

Paterson, W.S.B., 1981. The Physics of Glaciers. Pergamon, Oxford, 2nd ed., 380 pp.

Peltier, W.R. and Tushingham, W.R., 1989. Global sea level rise and the greenhouse effect: might there be a connection? Science, 244: 806–810.

Reeh, N., 1985. Was the Greenland ice sheet thinner in the late Wisconsinan than now? Nature 317: 797–799.

Reeh, N., 1989. Dynamic and climatic history of the Greenland ice sheet. In: R.J. Fulton (Editor), Quaternary Geology of Canada and Greenland, The (Geol. North America, K-1) Geol. Soc. Am. pp. 795–822.

Reeh, N., 1990a. Past changes in precipitation rate and ice thickness as derived from age–depth profiles from ice sheets; application to Greenland and Canadian Arctic ice-core records. In: U. Bleil and J. Thiete (Editors), Geological History of the Polar Oceans: Arctic versus Antarctic. Kluwer, Dordrecht, pp. 255–271.

Reeh, N., 1990b. Parameterization of melt rate and surface temperature on the Greenland ice sheet. Polarforschung, in press.

Van der Veen, C.J., 1986. Ice sheets, atmosphere CO_2 and sea level. Thesis. Univ. Utrecht, 185 pp.

Weidick, A., 1985. Review of glacier changes in west Greenland. Z. Gletscherkd. Glazialgeol., 21: 301–309.

Wigley, T.M.L. and Raper, S.C.B., 1987. Thermal expansion of sea water associated with global warming. Nature 330, 127–131.

Zwally, H.J., Brenner, A.C., Major, J.A., Bindschadler, R.A. and Marsh, J.G., 1989. Growth of Greenland ice sheet: measurement. Science, 246: 1587–1589.