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The present evolution of the Greenland ice sheet: an assessment by modelling

Philippe Huybrechts

Alfred-Wegener-Institut für Polar- und Meeresforschung, Columbusstrasse, Postfach 120161, D-27515 Bremerhaven, Germany

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Abstract

The present state of balance of the Greenland ice sheet is an important initial condition to assess the ice sheet's contribution to future sea levels. In this paper this problem is investigated by first simulating the evolution of the ice sheet during the last glacial–interglacial cycle and subsequently analyzing the local thickness change which results for the present time. The 3-D, time-dependent ice sheet model includes thermomechanical coupling and the effect of the stiffness contrast between Holocene and Wisconsin ice. The model forcing consists of a temperature record for central Greenland which resolves features down to a period of 100 years.

The calculations indicate that the ice sheet as a whole is at present thickening slightly at a mean rate of almost 1 cm/yr, corresponding to a world-wide sea-level lowering of only a few mm during the last hundred years. However, large spatial differences occur. Central and northern parts high up in the accumulation area show a slight thinning which is most likely due to a slow but consistent basal warming in response to the last glacial–interglacial transition. Marginal thinning, most notably in the northeast and along major parts of the west coast, on the other hand, is probably due to increased melting caused by higher temperatures subsequent to the Little Ice Age. The most striking feature is a rather important thickening in the southwestern part of the ice sheet, which appeared to be robust against all changes in environmental and ice-dynamic factors tried. It is suggested that this represents a long-term trend caused by a purely dynamic reaction to the geometry which came out of the last glacial–interglacial transition.

A comparison of the calculated response pattern with all of the independent observations which exist for the southern part of the ice sheet shows a remarkably good agreement, in particular with respect to the sign of the evolution.

1. Introduction

During the last few years several estimates have been made of the possible reaction of the Greenland ice sheet to future greenhouse warming (Bindschadler, 1985; Ambach and Kuhn, 1989; Braithwaite and Olesen, 1990; Huybrechts et al., 1991; Oerlemans et al., 1991/1992). This ice sheet contains around 10% of the earth's ice volume and it is generally believed that a climatic warming will lead to increased melting and a smaller ice volume. According to these studies, a 1°C warming would result in a mass loss corresponding to a world-wide sea-level rise of between 0.25 and 0.5 mm/yr. If the current predictions of the magnitude of the CO_2-induced climatic warming

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hold true, this could lead to a Greenlandic contribution to sea-level rise of up to several tens of centimeters by the end of the next century.

A major unknown, however, is the present state of the ice sheet. At present, it is not really known whether the total mass of the ice sheet is increasing, decreasing or unchanging. From a physical point of view, regarding the very long time scales introduced by isostasy, thermo-mechanical coupling and the slow advection of harder Holocene ice in the basal shear layers, it seems unlikely that the Greenland ice sheet has adjusted completely to its past history. Even in the absence of any present-day climatic perturbations, the ice sheet is expected to respond to past changes of its surface boundary conditions for a long time to come, in particular to those changes associated with the last glacial–interglacial transition. Also mass balance variations caused by shorter-term climatic fluctuations will still influence the ice-dynamic response by their direct impact on the ice thickness distribution.

To date, several methods have been used to assess the imbalance of the Greenland ice sheet, yielding estimates on both global and regional scales. Based on only a limited amount of measurements, however, they give a rather confusing and very incomplete picture. For instance, in a survey of available mass budget estimates, where total accumulation estimates are compared with the mass loss terms of runoff and calving, Warrick and Oerlemans (1990) conclude that at present an imbalance of up to 30% of the annual mass turnover cannot be detected in a definite way. Other investigations have concentrated on a more limited area and have provided estimates based on local mass-balance studies. Reeh and Gundestrup (1985) conclude that the ice sheet at Dye3 is thickening by 3 ± 6 cm/yr, which is of a similar magnitude than the thickening rate of 6 ± 8 cm/yr derived by Kostecka and Whillans (1988) for the OSU-transect to the west of Dye3. Geodetic methods have been used along the EGIG-line at about 70°N in central Greenland (Mälzer and Seckel, 1975; Seckel, 1977). Repeat long-line leveling along this traverse showed that between 1959 and 1968 the ice thickened by around 10 cm/yr in the accumulation zone, while surface elevations in the ablation area below 1200 m of altitude lowered by 0.24 m/yr during the same period. On the basis of satellite altimetry, Zwally et al. (1989) found that the mean mass balance of the Inland Ice south of 72°N was positive in the period 1978–1986. They report that thickening occurred both in the accumulation and the ablation zones with a magnitude of 0.2–0.3 m/yr. There are however doubts regarding the accuracy of these results (Douglas et al., 1990). Direct observations on outlet glacier positions are few and mainly exist for the central and southern part of the west coast of Greenland. They show that the Neoglacial maximum usually occurred in the middle or last half of the 19th century and was followed by a general recession since then (Weidick, 1985). However, a recent update of the conditions to 1985 revealed a "turn of the tide", where the general recession since the last century is now substituted by a major tendency for advance (Weidick, 1991). Most of these observations, but not all, seem to be consistent with a recent warming trend, which would lead to higher accumulation rates and a thickening above the equilibrium line together with higher melting rates and a thinning in the ablation zone, but the picture remains far from unequivocal.

In this paper, the question of the present evolution of the Greenland ice sheet is approached in a different way. Here, the imbalance of the ice sheet is obtained by first modelling the ice sheet during a full glacial cycle and then investigating the evolution pattern which results for the present day. Theoretically, this alternative method to assess the imbalance would be exact provided both the past boundary conditions (mass balance, surface temperature, ...) over a time scale at least as long as the longest response time scale of the system and the ice dynamics were perfectly described. Although one cannot yet expect to fully meet these conditions with our present knowledge, a lot can be learned from such a study. In particular a sensitivity study in terms of environmental and ice-dynamic factors should help to better understand and explain the resulting patterns. Besides giving a global number for the volume change, this kind of experiments also yields information on the regional distribution of the ice thickness evolution.

2. Model description

An overview of the fundamental mathematical equations governing the model has been published in a previous paper (Huybrechts et al., 1991), and will not be presented again. Here, only a brief description will be given of the main components, with special attention to those new features that have been added since. The structure of the model is displayed in Fig. 1.

A basic assumption is that the ice flows in direct response to pressure gradients set up by gravity. Two processes contribute to the mass flow: internal deformation resulting from shear strain and sliding over the bed, which is assumed to be of Weertman-type (Weertman, 1964; Budd et al., 1979) and restricted to areas where the basal temperature is within 1°C of the pressure melting point. The softness parameter of ice, which determines the rate of deformation, depends on both the ice temperature and the age of the ice. For this reason, the three-dimensional temperature and velocity fields are calculated in the coupled mode. The age distribution is required because of the different mechanical characteristics of ice-age ice and Holocene ice found in Greenland ice cores, which is apparently related to a marked contrast in crystal size associated with varying concentrations of chloride and sulphate ions (Paterson, 1991). The ice age at any moment and location is therefore obtained by the inclusion of an age variable which is advected along particle paths as the ice sheet evolves with time. This represents a new feature of the model and is briefly commented upon in the appendix. The Wisconsin/Holocene transition is placed at 10,800 yr B.P. (Dansgaard et al., 1989), with the Wisconsin ice deforming three times more readily than the Holocene ice for the same stress and temperature conditions (Dahl-Jensen and Gundestrup, 1987). The possibility that harder interglacial ice from previous warmer periods is still residing in the ice column was not taken into account, mainly because of uncertain dating in the lowermost basal layers and the fact that these hard layers, if still there, would have become very thin anyway.

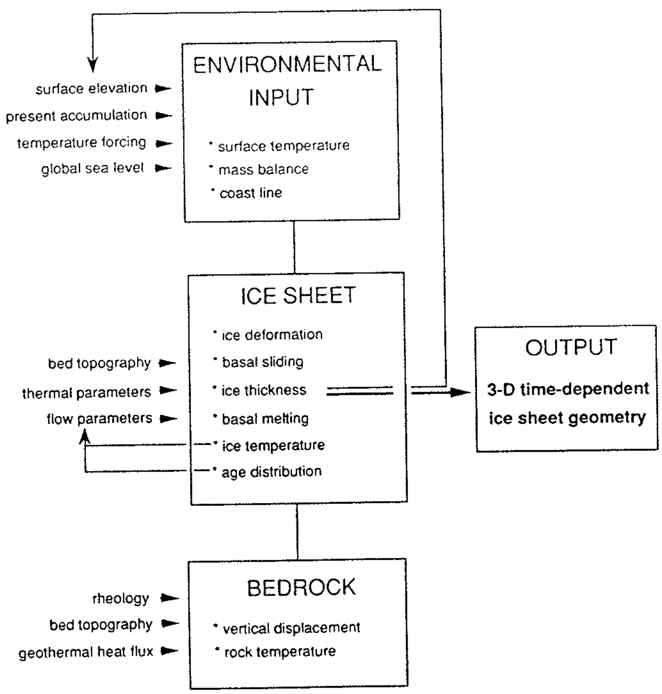

Fig. 1. Structure of the model used in this investigation. Inputs are given at the left-hand side. Prescribed environmental variables drive the model, which has grounded ice and bedrock adjustment as major components. Temperature and age of the ice determine the ice flow properties. Ice thickness feeds back on surface elevation, an important parameter for the calculation of the mass balance. The main output is the 3-D time-dependent ice sheet geometry, which is freely generated by the model.

Regarding the bedrock component, isostatic compensation in response to the changing ice loading is modelled in a simple way by assuming local depression in a damped fashion with a characteristic e-folding time scale of 3000 years. As temperature perturbations are also conducted further into the bedrock, implying that the heat flux at the ice–rock interface is not constant, the model also includes a temperature calculation down to a rock depth of 2 km.

All calculations are performed on the same grid. It has a horizontal spacing of 20 km. With an improved vertical resolution of 26 layers in the ice, with the closer spacing near to the base, and another 5 in the bedrock for the heat conduction calculation, this adds up to a total of over 300,000 gridpoints.

Model input consists of bed topography, surface temperature, mass balance, coast line, thermal and rheological parameters and an initial state. The environmental forcing is made up by both the eustatic sea level stand, which determines the coast line, so that the ice sheet can expand onto the shallow parts of the continental shelf during glacial periods, and a uniformly distributed background temperature change, from which the mass balance components (snow accumulation and runoff) are calculated. The presently observed precipitation rate is taken as a base and prescribed to vary by 5.3% for every degree of change in mean annual air temperature. This value is suggested by correlating annual layer thicknesses of shallow ice cores in Central Greenland with the corresponding $\delta^{18}O$ values (Clausen et al., 1988). The melt-and-runoff 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 in a way described earlier by Huybrechts et al. (1991) and Reeh (1989).

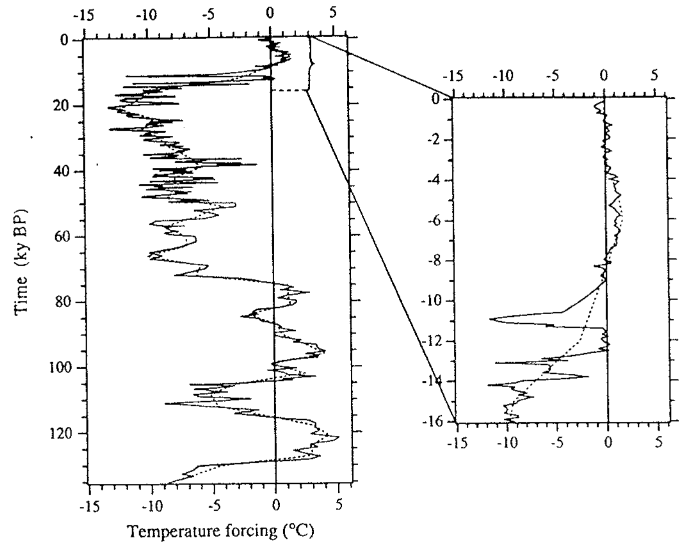

Fig. 2. Temperature record used as forcing during the last glacial–interglacial cycle. The part older than 10,800 yr B.P. originates from the Pakitsoq $\delta^{18}O$ record obtained from surface ice sampling in central West Greenland (Reeh et al., 1991). The Holocene part was collated from the Dye3 and Camp Century records (Dansgaard et al., 1984, 1989; Hammer et al., 1986; Reeh, pers. comm.). The dashed line shows a smoothed version of the temperature record. The enlargement at the right covers the period between the Last Glacial Maximum and the present-day.

3. The standard experiment

In order to minimize the effect of the startup conditions on the results, the ice sheet was first simulated through a complete glacial–interglacial cycle. This time span is considered to be of sufficient duration for the model to basically forget its initial conditions. Calculations started at the penultimate glacial maximum at 135,000 yr B.P. A simulation of the ice sheet in a glacial climate was used as initial configuration, by applying a temperature 10°C colder than at present and a sea-level lowering of 130 m, and letting the model run until a steady state was achieved.

3.1. Model forcing

Fig. 2 shows the temperature forcing used to determine the surface mass balance and drive the ice-dynamic model. This record, representative of central Greenland surface conditions, was assembled from several sources. The pre-Holocene part older than 10,800 yr B.P. was derived from an oxygen-18 record measured on surface-ice samples collected from the ice margin at Pakitsoq, central West Greenland (Reeh et al., 1991). This record was chosen in preference to one of the deep ice cores because the Pakitsoq record is presently the only one that seems to cover at least one full glacial–interglacial cycle of the Greenland history. The time scale prior to 110,000 yr B.P. was slightly modified from the original temperature series published in Reeh et al. (1991) to be in accordance with the marine isotopic records rather than the chronology established for the Vostok deep ice core. The temperature deviations with respect to the present were derived by translating the ^{18}O record (after correction for the $\delta^{18}O$ values presently found at the sites where the snow was originally deposited) by means of the conversion factor 0.62‰ $\delta^{18}O$ per °C (Dansgaard, 1961).

Since the Pakitsoq record covers only the early part of the present interglacial, the temperature history since the Younger Dryas was collated from the Camp Century and Dye3 oxygen isotope records, which strongly resemble each other during the Holocene (Dansgaard et al., 1984; Hammer et al., 1986). Except for the interval between c. 2500 and 6000 yr B.P., where data for Dye3 are lacking, the mean was taken of the respective temperature perturbations averaged over 100-yr intervals. Prior to this, the temperature variations in the Camp Century record were first reduced by a factor 1.5 to be compatible with the smaller temperature perturbations recorded elsewhere in Greenland. Conversion factors between $\delta^{18}O$ values and temperature deviations were set at 0.62‰ $\delta^{18}O$ per °C for the Camp Century data and 0.71‰ $\delta^{18}O$ per °C for the Dye3 oxygen-isotope data (Dansgaard et al., 1989). Additionally, a correction was made for the fact that the older ice was deposited further inland, and therefore, at a higher altitude with lower $\delta^{18}O$ values (corrections communicated personally by N. Reeh). The temperature forcing used here resolves features down to a period of 100 yr. The last value during the interval between 100 yr BP and today was kept constant at 0°C. This means that only the somewhat longer-term trend will show up in the results.

The sea level forcing, which is not so crucial for the evolution of the Greenland ice sheet, as the continental slope is rather steep, was taken from a sequence of raised marine terraces on the Huon Peninsula, New Guinea (Chapell and Shackleton, 1986). It reaches a minimum of −130 m at 17,000 yr B.P. and is constant at zero after 6000 yr B.P. These forcing functions together with the full model as described above define the standard experiment, which is also denoted as "R" further below.

3.2. Results

How the Greenland ice sheet looks like during the last glacial cycle in this experiment has been presented before (Letreguilly et al., 1991) and does not differ much from the present result. Despite some modifications in the model setup, it appears that the different treatments of accumulation rate and bedrock adjustment adopted for this paper have had largely counteracting effects. Of interest here is the ice thickness evolution during the final stages of the simulation. According to the numbers given in Table 1, which are averages over the last 200 yr consistent with the forcing resolution, the Greenland ice sheet would be still growing at a rate of about 14 km^3 of ice per year. With an ice sheet surface of 1.81×10^6 km^2, this is equivalent to a mean imbalance spread out over the entire ice sheet of the order of +8 mm/yr. However, compared to the mean annual mass turnover of around 35 cm/yr, this is only a small number, so that the Greenland ice sheet as a whole does not appear to be very far out of

Table 1
Overview of the ice thickness evolution rates for the various experiments performed. Numbers represent average values over the last 200 years. The standard model run refers to the one described in the text, the sensitivity experiments differ from the standard run by the items specified

Run	Description	Ice volume evolution (10^9 m^3/yr ice eq.)	Mean thickness change of ice sheet (mm/yr ice)	Contribution to global sea level (mm/century)
R	standard	+14.03	+7.76	−3.53
a	with smooth temperature forcing as shown in Fig. 2	+6.55	+3.66	−1.65
b	idem as a, but without the Climatic Optimum warming around 6000 yr B.P.	+6.52	+3.61	−1.64
c	without thermomechanical coupling and isothermal ice at −8°C	+28.79	+15.80	−7.23
d	without heat conduction in the bedrock below	+12.99	+7.18	−3.26
e	without the stiffness contrast between Wisconsin and Holocene ice	+13.01	+7.20	−3.27

balance due to long-term effects. Expressed in global sea level changes, the corresponding value is −3.5 mm per century. Putting it in another perspective, the calculated imbalance also corresponds to about 10–15% of the sensitivity of the Greenland mass budget to a 1°C warming (cf. values quoted in Warrick and Oerlemans, 1990), though is of opposite sign.

It is interesting to note that the modelled ice sheet is still growing in spite of the fact that the temperature forcing shows a rather sharp rise of 0.88°C during the last 200 years, implying increased melting rates subsequent to the Little Ice Age. The direct mass balance effect of this recent warming can be roughly estimated to be of the order of +0.10 mm/yr sea level equivalent. This number is obtained by summing the product between the mass balance sensitivity value of +0.22 mm $yr^{-1}°C^{-1}$ (Huybrechts et al., 1991) and the temperature departure for each year (totalling 0.44×200°C yr) and dividing the result by the number of years (200). The implication is that this recent temperature rise may well mask an underlying growing trend due to longer-term (thermo-) dynamic effects of perhaps 3 or 4 times as large. As a matter of a fact, the ice volume change averaged over the last 4000 years shows a mean growth rate close to 20 km^3/year, with a peak of over 30 km^3/yr between 700 and 200 years ago (see also Fig. 4).

The geographical distribution of the local surface elevation imbalance in this standard experiment, that is the present-day rate of change of the ice sheet's surface, is displayed in Fig. 3. Although the ice sheet as a whole is thickening slightly, a number of interesting spatial patterns occur. In central and northern areas high up in the accumulation area, it appears that the ice sheet is actually slowly thinning at a rate of at most a few cm/year (e.g. Summit area: −8 mm/yr). This region of lowering surface elevations is almost entirely enclosed by a belt of moderately rising elevations that extends further over most of the southern part of the Inland Ice (e.g. the Camp Century and Dye3 areas are both found to be rising at rates of between +1 and +2 cm/yr).

However, the most complex response pattern shows up in the marginal areas, where zones of

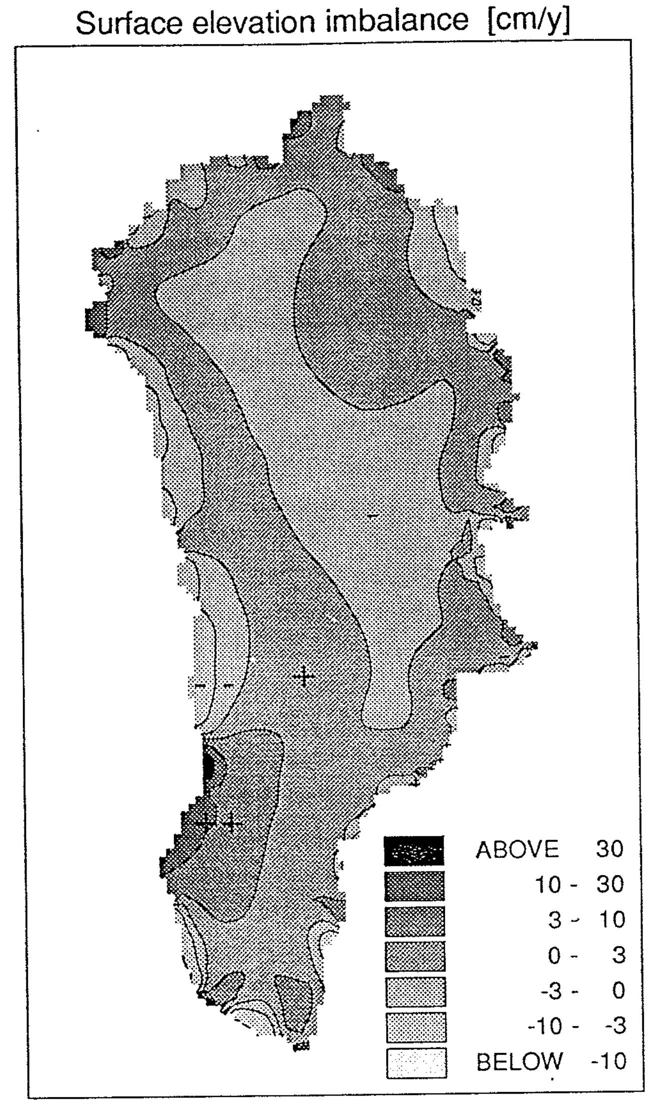

Fig. 3. Rate of surface elevation change averaged over the last 200 yr for the standard experiment. Positive values indicate a thickening, negative values are for a thinning, as marked by the respective symbols. The pattern for the ice thickness evolution resembles this one to a large extent, but values tend to differ slightly by a few mm/yr due to isostatic adjustments.

relatively large thinning of up to 10 cm/yr (e.g. in northeast Greenland between Lambert Land and the Storstrømmen area in Germania Land and along the west coast from Cape York to south of the Jakobshavn Isbræ catchment zone) are interspersed by zones where the ice is advancing. The most striking feature is the relatively large thickening in the southwestern corner of the Greenland ice sheet between about 68°N and 64°N, where the ice sheet margin is entirely land-based and glacier fronts are accordingly progressing. The largest ice thickening of up to 50 cm/yr actually occurs upstream the glacier tongues immediately north of the Søndre Strømfjord airbase. Closer inspection of the simulated ice sheet geometries revealed that this margin advance in the southwest represents a long-term trend, which commenced about 4000 years ago.

Also here, the result is likely to be most influenced by the most recent mass balance change associated with the recovery from the Little Ice Age cooling. Following a similar reasoning as above and assuming a steady state 200 years ago while neglecting dynamic effects, this should have led to a thickening of several cm/year in the accumulation area and a uniform thinning below the equilibrium line, at odds with the result shown in Fig. 3. Apparently, the underlying lower-frequency base trend is of sufficient strength to override these shorter-term mass-balance perturbations in many areas. This certainly demonstrates the impact of the ice sheet's (thermal) memory and the associated long response time scales.

It is of considerable interest to compare the pattern shown in Fig. 3 with the sparse measurements mentioned in the introduction. Although the model results are not necessarily comparable with these observations because of the time scales under consideration (a snapshot over a decade or so for the measurements versus 200-year mean values for the model, implying much care), there is actually a surprising agreement with all of the available measurements, in particular with regard to the sign of the evolution. The model results support the thickening derived for Dye3 and the OSU-line to the west as well as the differential behaviour of the EGIG-line further to the north. Moreover, the simulated imbalance is compatible with an average thickening south of 72°N, also in the ablation zone. This is in good agreement with the results obtained by Zwally et al. (1989), though their magnitudes are somewhat larger, and throws a new light on the apparent contradiction between the Zwally-findings and the observed thinning below 1200 m along the EGIG-line. In fact, the result presented in Fig. 3 supports both inferences. Also the observed recession of most outlet glaciers in the southwest up to 1950 may only have been a short interruption of a longer-term trend of advance (Weidick, 1991).

4. Sensitivity of the results to changes in the model setup

In an attempt to investigate how these response patterns depend on both internal (ice dynamics) and external (temperature forcing) factors, a number of sensitivity experiments were set up in which one item was changed at a time, while keeping all other things equal. Comparing pairs of experiments should then help to explain the observed features. Table 1 gives a short description of these sensitivity tests, together with a few large-scale variables summarizing their effects. The corresponding ice volume curves for the last 7500 years of the integration are displayed in Fig. 4.

Fig. 5 shows the resulting present-day response patterns for ice thickness. A remarkably robust feature found in all simulations appears to be the thickening of the southwestern part of the Greenland ice sheet, suggesting an explanation in terms of factors common to all runs. Also the major role played by both the forcing (and thus mass-balance changes) and thermo-mechanical effects shows up distinctly. For instance, the differences

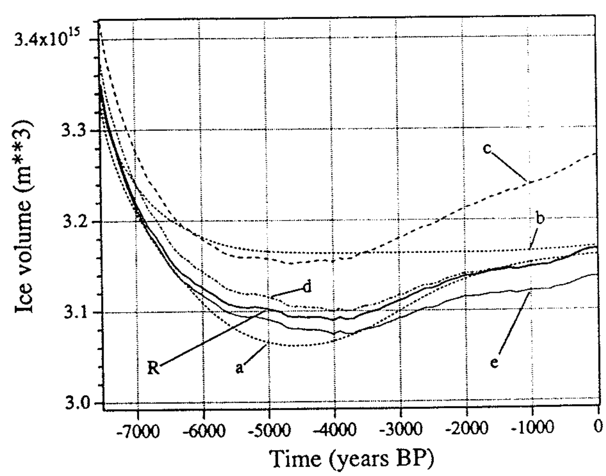

Fig. 4. Ice volume evolution since the middle Holocene. The solid curve denoted R refers to the standard experiment, the lower case characters correspond to the sensitivity runs specified in Table 1.

between the standard run (R) and the runs a (constant forcing at 0°C over the last 3000 yr) and b (idem, but over the last 8400 yr) must be due to the fact that the latter runs do not take into account the more recent temperature fluctuations such as those associated with the Little Ice Age.

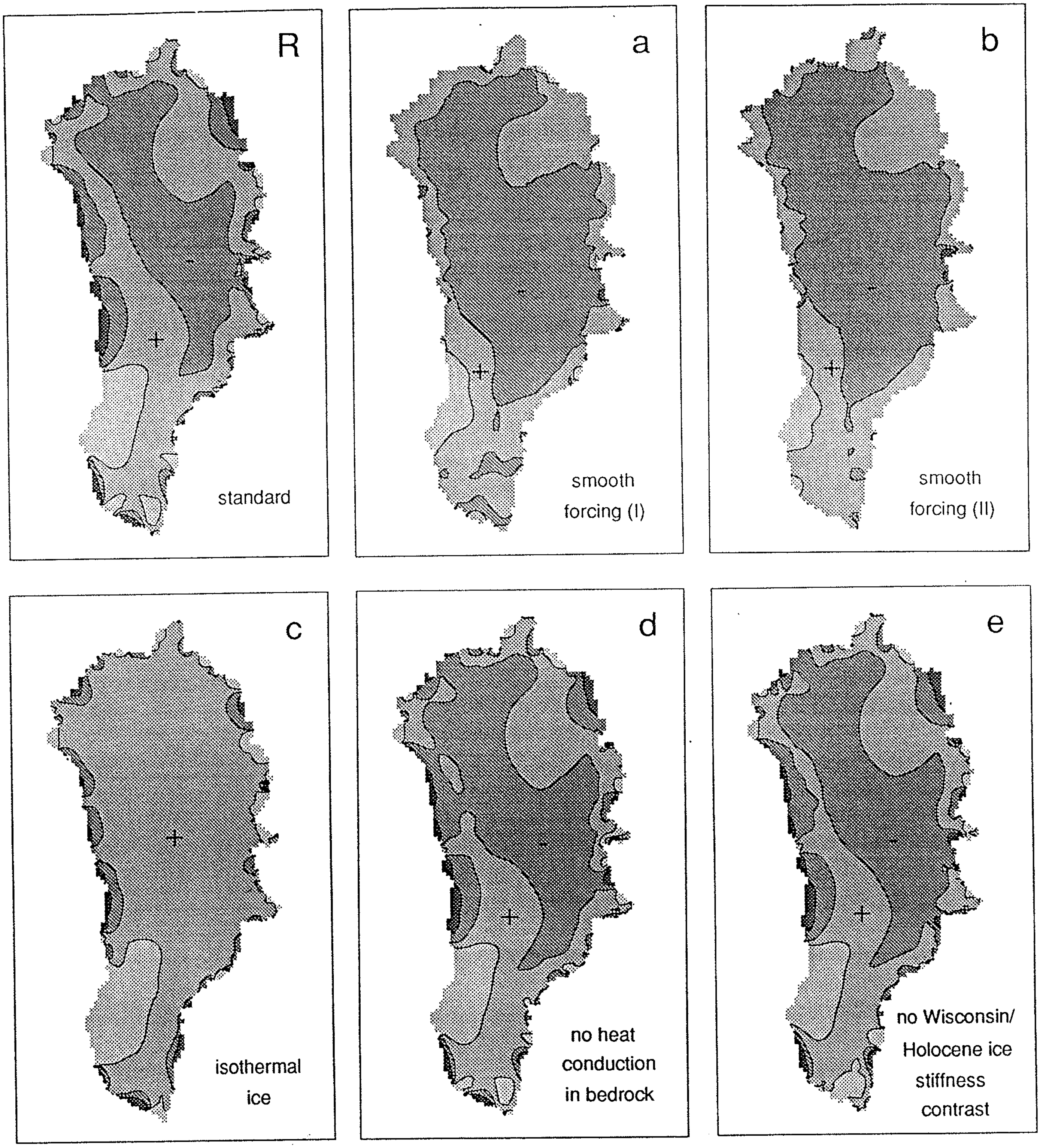

Fig. 5. Patterns of ice thickness evolution for the standard experiment (R) and the various sensitivity runs. Contour lines are for +3 cm/yr, 0 cm/yr and −3 cm/yr, respectively, yielding four classes. Plus and minus signs are indicated at both sides of the zero-evolution line. Lighter shading indicates a more positive value.

The effect is particularly evident at the margin, where the runs *a* and *b* do not exhibit any of the thinning linked to the recent warming in the standard run. The similarity between the runs *a* and *b* also indicates that the ice sheet has by now almost completely adjusted to the temperature rise which accompanied the Climatic Optimum. Lowering elevations in central areas in all three runs, on the other hand, points to a long-term response which can only be due to the last glacial–interglacial transition.

The impact of thermo-mechanical coupling becomes clear by comparing the standard run with run *c*. The latter experiment excludes the temperature dependence of the flow properties of ice, so that the ice sheet only reacts to mass-balance changes and driving stress (ice thickness and surface slope). Common features in the runs *R* and *c*, most notably the widespread thinning in the ablation area, must then be due to factors other than those related to ice temperature. Actually, the pattern for run *c* (isothermal ice) is much like what one would expect as a direct mass balance effect following the warming subsequent to the Little Ice Age, i.e. rising elevations above the equilibrium line and a thinning in the ablation area. The main effect of excluding thermo-mechanical coupling is then a doubling of the present growth rate (cf. Table 1 and Fig. 4) and the absence of the large thinning zone in the accumulation area common to all other experiments. A closer inspection of the temperature field in the standard experiment reveals that its smaller growth is caused by a warming of the basal shear layers. Warmer ice deforms more readily, leading to larger strain rates, so that smaller ice thicknesses result. Furthermore, it appears that this basal warming is at present occurring all over the ice sheet, though long thermodynamic time scales are involved and different areas generally exhibit different behaviour at different times (e.g. Huybrechts and Oerlemans, 1988). This is demonstrated in Fig. 6, showing to what extent the basal temperature field has adjusted to the last glacial–interglacial transition. The temperature response has been nearly completed at the margin, but interior regions are still far out of balance with the present interglacial climate. Actually, negative numbers down to −30% indicate that the net temperature effect has so far been a decrease, whereas the final response is positive. This feature is related to increased advection of cold ice towards the base, which in first instance dominates over the warming transferred by heat conduction. A more comprehensive analysis of the thermodynamics in these experiments, however, would be far beyond the scope of the present paper, and is to be presented elsewhere.

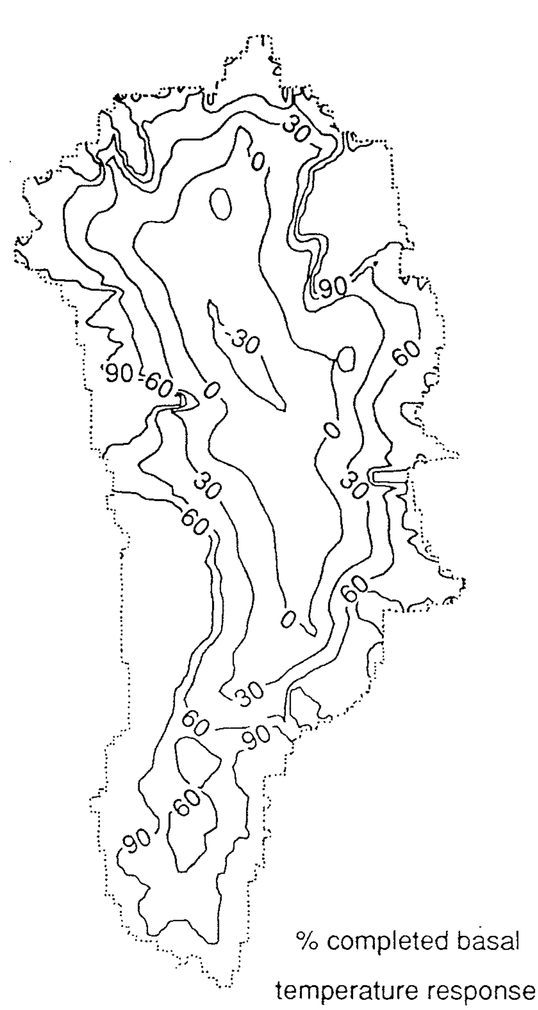

Fig. 6. Basal temperature change between the Last Glacial Maximum at 18,000 yr B.P. and the present time expressed as a fraction of the total temperature change between 18,000 yr B.P. and the one expected for steady state under present climatic conditions. The latter total temperature change ranges from about +5°C in interior areas to between 10 and 12°C in marginal zones where the ice is not at the pressure melting point. The percentages demonstrate the long thermodynamic response time scales involved.

As shown in run *d*, the inclusion of heat conduction in the bedrock, which introduces a thermal buffer and makes time scales longer, has only a small effect. A similar remark applies to the exclusion of the hardness difference between Holocene and Wisconsin ice (run *e*). Although this has led to progressively smaller ice volumes during the second half of the Holocene and a slower growth altogether as expected from qualitative reasoning (cf. curves *e* and *R* in fig. 4), its influence has not yet been very pronounced. This is because most of the velocity shear (over 80%) is concentrated in the lowermost 10% of the ice column, which is still largely made up by the softer Wisconsin ice. Apart from the southern margin, where Holocene ice accounts for more than 90% of the ice thickness, the Holocene–Wisconsin transition has at present only penetrated downwards to between 40 and 70% of the total depth elsewhere (Fig. 7). At Dye3, for example, the stiffness contrast between glacial and interglacial ice alone has so far resulted in a present-day rate of ice thickening of about 0.4 cm/yr, compared to less than 0.1 cm/yr in the Summit area. These values are a factor 3–5 less than the theoretical predictions given by Reeh (1985). The main reason for this discrepancy is the effect of thermo-mechanical coupling on the horizontal velocity distribution with depth, which in the present study results in a higher concentration of shearing in the lowermost layers. This is because near to the base temperatures are highest, implying a higher rate factor and hence, larger strain rates for the same stress conditions. The net result is that in the present study a proportionally lower fraction of deformation takes place in the hard-ice part of the ice column, resulting in a smaller thickening effect than the one reported in Reeh (1985).

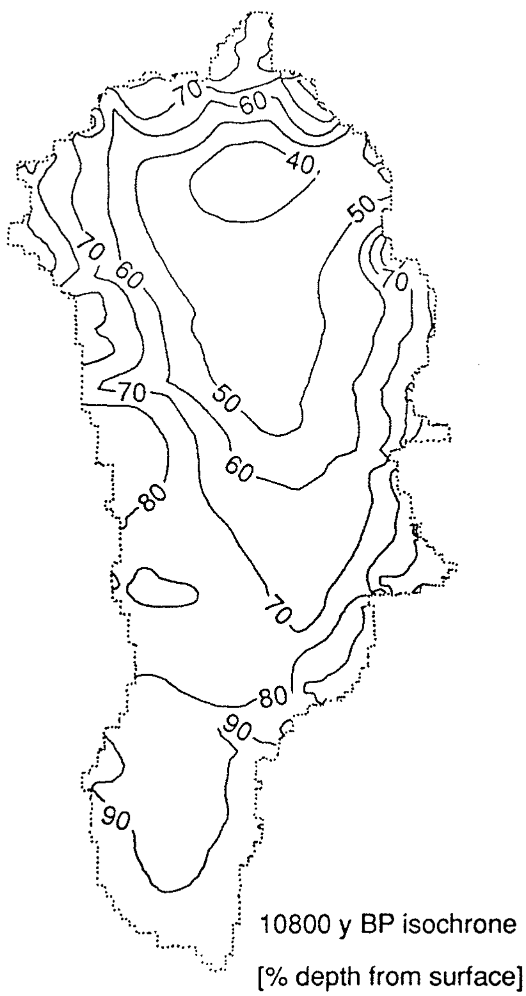

Fig. 7. Relative depth of the Wisconsin/Holocene transition in the ice sheet indicating how deep the stiffer Holocene ice has at present penetrated downwards.

5. Summary and conclusions

This paper addressed the question of the present evolution of the Greenland ice sheet by reconstructing the ice sheet's history through the last glacial cycle and analyzing the resulting imbalance field for the present-day. The numerical experiments clearly brought to light the role played by mass-balance and ice-dynamic effects as well as the time scales involved. As a general result, it was found that on the longer-term (> 100 yr) the ice sheet as a whole is almost stationary or thickening slightly, contributing to a global sea-level lowering of the order of a few mm per century only.

However, the response pattern revealed large geographical variations. The results of the sensitivity tests indicate that the thinning found in the central and northern parts of the accumulation area is a delayed response to the last glacial–interglacial transition, which is best explained in terms of a slow but consistent basal warming. This long-term response to past changes of the surface climate (surface temperature and mass balance) is furthermore dominant with respect to shorter-term fluctuations of the accumulation rate. The downward transport of the harder Holocene ice does not yet play a role of much significance. The thinning in marginal areas below the equilibrium line, on the other hand, is an immediate effect due to a recent increase in ablation rates subsequent to the Little Ice Age.

Most of the observed mean growth common to all simulations can be attributed to a long-term thickening trend of a relatively large magnitude in the southwestern part of the ice sheet. The reasons for this expansion are not entirely clear, but the sensitivity study strongly suggests that it is a purely dynamic reaction to the geometry which came out of the last glacial–interglacial transition. Most likely, the ice sheet west of the saddle area is still adjusting to the steeper slopes initially generated by a strongly retreated margin far into inland areas, supplemented with a thickening above the equilibrium line as accumulation rates increased into the Holocene. Such a mechanism would be much less effective in other parts of the ice sheet because the margin already coincides with the coast line, leaving little space for important changes of the ice sheet extent.

In this study only the longer-term base-trend was investigated. This does not mean, however, that recent climatic fluctuations on time scales shorter than the forcing resolution (i.e. on a decadal rather than on a century time scale) could not have pushed the ice sheet further out of balance. For instance, Warrick and Oerlemans (1990) estimate the direct mass balance effect of summer temperature deviations over the last hundred years to be of the order of $+0.23 \pm 0.16$ mm/yr of global sea level rise, which is a value largely in excess of the longer-term trend of -0.035 mm/yr found in this study. Although hard to quantify, uncertainties are also introduced by simplifications of the model itself. In real world, the response of the ice sheet will also depend on such factors as the anisotropy of the ice, variability in the calving speed of outlet glaciers, a non-uniform temperature forcing, a changing precipitation pattern with climate, etc... and these effects have not been properly dealt with mainly because a good understanding is still lacking. In all, it seems fair to state that at present the past surface mass-balance history cannot yet be reconstructed in a fully confident way. In spite of these uncertainties, however, the simulated pattern of ice-elevation change displayed a remarkable agreement with almost all of the existing measurements in the southern half of the ice sheet, in particular with respect to the sign of the evolution.

Finally, it should be noted that the magnitude of the simulated imbalance is likely to be very small when compared to the potential effects of a greenhouse warming of possibly as much as 10°C during the next century. Such large perturbations may easily lead to volume changes of up to two orders of magnitude larger than the trend found in this study. In view of the possibility of such large variations, it seems that the Greenland ice sheet may well be considered to be in equilibrium with the present Holocene climate.

6. Appendix: calculation of the age distribution

A new feature introduced in the model concerns the way the age distribution is obtained throughout the ice sheet. From a mathematical point of view, age can be considered as a scalar quantity which is conserved by the ice particle during its motion. By definition, in a Lagrangian description the following equation holds:

$$\frac{\mathrm{d}A}{\mathrm{d}t} \equiv 0 \tag{1}$$

where t is time (yr) and the ice age A (yr BP) refers to the time when the ice particle was deposited at the ice sheet's surface. In a Eulerian description this becomes:

$$\frac{\partial A}{\partial t} = -v \cdot \nabla A \tag{2}$$

where the velocity vector v and the ∇-operator are understood to be three-dimensional. This advective equation is solved subject to the boundary conditions:

$$A(t) = t \quad \text{at the surface} \tag{3}$$

$$\frac{\partial A}{\partial t} = S\frac{\partial A}{\partial z} - \left(u_b\frac{\partial A}{\partial x} + v_b\frac{\partial A}{\partial y}\right) \text{ at the bottom} \tag{4}$$

where S is the basal melting rate [m/yr], positive in the case of melting and u_b and v_b are the velocity components at the base. A flux boundary condition is not specified at the surface, so that Eq. 2 is strictly speaking only valid in the accumulation zone. An initial age distribution for the glacial ice sheet at 135,000 yr B.P. was obtained after integrating Eq. 2 for 265,000 yr using the velocity field corresponding to the glacial state. The maximum ice age is set at 400,000 yr B.P., which is also the final age at the bottom in those areas where cold ice has persisted throughout the evolution.

Numerically, explicit finite difference schemes of the type forward in time/central in space are known to be unconditionally unstable for pure advection equations. In order to remedy this behaviour, the following measures were taken. First of all, a small artificial diffusivity of 2×10^{-2} m^2/yr was added along the vertical to prevent transport errors from swamping the entire solution. This option is preferred to the usual practice of straightforward upwind differencing, which is only accurate to first order and can be shown to introduce an artificial diffusivity of up to three orders of magnitude higher. Secondly, the vertical age derivative, weighted for the variable grid size, was evaluated implicitly to relax the limitations on the size of the time step (Courant condition). This seems reasonable since the smallest grid spacings are along the z-axis. Finally, the horizontal advection terms were satisfactorily replaced by a dissymmetric upwind differencing scheme involving 3 gridpoints, which is accurate to second order and produces less numerical dissipation than a two-point upwind differencing scheme.

An advantage of the present ice age calculation is that transient effects, such as changes in ice thickness, accumulation rate or flow pattern are automatically taken care of. Though the dating becomes increasingly uncertain in the basal layers, a property in common with other ice-flow-modeling dating techniques (Reeh, 1989), the result shown in Fig. 7 seems to demonstrate that the present method works well for the purpose of determining the depth of the Holocene–Wisconsin transition. In the Greenland boreholes of Camp Century, Dye3 and Summit, this transition has been confidently established at a relative depth of 83.3%, 87.6% and 53.6% (Hammer et al., 1986; Dansgaard et al., 1989; Johnsen et al., 1992). The present method yields values of 84.4%, 86.2% and 54.4%. For comparison, the same numbers for a simple analytical model based on a constant vertical strain rate with constant ice thickness and accumulation rate (Nye-Haefeli time scale; Haefeli, 1961) are 94.8%, 94.6% and 55.4%, respectively. In these calculations, the Holocene–Wisconsin transition was set at 10,800 yr B.P. Parameters pertinent to the Nye-Haefeli time scale were: Camp Century ($H = 1387$ m, $M = 0.38$ m/yr); Dye3 ($H = 2037$ m, $M = 0.55$ m/yr) and Summit ($H = 3030$ m, $M = 0.227$ m/yr).

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