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The Greenland ice sheet through the last glacial–interglacial cycle

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ABSTRACT

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The evolution of the Greenland ice sheet during the last 150,000 years, in response to a climate history derived from a Greenland ice-margin oxygen-18 record, is simulated by means of a three-dimensional, time-dependent ice-sheet model. The calculations indicate that the ice sheet displayed considerable thinning and ice-margin retreat during the last interglacial (the Eemian) and during a warm interstadial c. 100,000 yr B.P., resulting in a splitting up of the ice sheet into a central-northern and a southern part. However, the ice sheet in Central Greenland survived the warm stages with almost unchanged surface elevations as compared with the present.

Introduction

The Greenland ice sheet is the second largest ice mass in the world, with a volume of 2.82×10^6 km^3. The water stored as ice in Greenland is equivalent to a 6 m layer of water over the world oceans. During past warmer climates, part or all of the Greenland ice sheet may have disappeared, and contributed to a general sea-level change.

The possibility of a total deglaciation of Greenland in previous interglacials was discussed as early as 40 years ago (Flint, 1947; Charlesworth, 1957). However, the fact that at lower levels in the deep ice cores from Camp Century (Northern Greenland) and Dye3 (Southern Greenland) ice is found of higher δ^{18}O values than that of the present surface layers (indicating a warmer climate than the present one), has been taken as evidence that the ice sheet existed during the last interglacial, the Eemian (Dansgaard et al., 1971; Dansgaard et al., 1982).

However, recently this interpretation has been questioned due to difficulties in interpreting the same lower part of the Camp Century and Dye3 ice-core records (Reeh, 1990; Koerner,1989). The δ^{18}O values, the dust content of the bottom ice, and the snow-accumulation rates derived by ice-dynamic modelling have proved difficult to understand unless it is assumed that the ice sheet in the Camp Century and Dye3 regions had thinned considerably during the last interglacial. It has been further suggested (Reeh, 1990; Koerner, 1989) that a significant part (if not all) of the reported 6 m higher sea-level stand in the Eemian interglacial (Chappel and Shackleton, 1986) can be explained by melting of the Greenland ice sheet, either in part or in total.

Here we approach the question of the possible disappearance of the Greenland ice sheet in previous warm interglacial climates by means of a detailed ice-dynamic model study. The present mass balance distribution over Greenland, perturbed by climate variations as derived from a Central Greenland ice-margin oxygen-18 record (Reeh et al., in prep) is used as climate forcing for an ice-dynamic model during the last 150,000 years.

The ice sheet model

The model used is a three-dimensional, thermo-mechanical ice-sheet model originally developed for Antarctica (Huybrechts 1986, 1988, 1990, 1991), coupled with a mass-balance model for Greenland. The details of the ice-sheet model are described by Huybrechts. Here, we will only present the basic equations governing the model. For a thin ice sheet (horizontal dimensions are one order of magnitude more than the vertical dimension) with small bedrock and surface slopes, and a cold base, the ice deformation can be assumed to result from shear strain. The stress distribution is given by:

$$\tau_{xz}(z) = -\rho g(H+h-z)\frac{\partial(H+h)}{\partial x},$$
$$\tau_{yz}(z) = -\rho g(H+h-z)\frac{\partial(H+h)}{\partial y} \qquad (1)$$

where H is the ice thickness, h the bed elevation, ρ the ice density, g the gravity acceleration, x and y the horizontal coordinates, z the vertical coordinate. The rate of deformation is then related to the stress by Glen's flow law:

$$\dot{\epsilon}_{xz} = A(T)\tau^{n-1}\tau_{xz},\ \dot{\epsilon}_{yz} = A(T)\tau^{n-1}\tau_{yz} \qquad (2)$$

where $\tau = (\tau_{xz}^2 + \tau_{yz}^2)^{1/2}$, $n = 3$ (Paterson, 1981), and $A(T)$ is a temperature dependant parameter:

$$A(T) = m \cdot A_0 \cdot \exp\left(\frac{-Q}{RT^*}\right) \qquad (3)$$

where T* is the absolute temperature corrected for the pressure induced increase in the melting point, $A_0 = 1.14 \times 10^{-5}$ Pa^{-3} yr^{-1}, $Q = 60$ $kJmol^{-1}$ for $T < -10\,°C$, and $A_0 = 5.47 \times 10^{10}$ Pa^{-3} yr^{-1}, $Q = 139$ $kJmol^{-1}$ for $T > -10\,°C$ (Paterson, 1981, chapter 3), R is the gas constant ($R = 8.314$ J mol^{-1} K^{-1}), m is a tuning parameter. In order to simulate the ice thickness correctly, a value of $m = 3$ is used. This value is in accordance with the observed enhanced flow of the near-bottom ice layers (Dahl-Jensen and Gundestrup, 1987). The evolution of the ice thickness is given by the continuity equation for the mass flux:

$$\frac{\partial H}{\partial t} = -\nabla \cdot (\bar{v}H) + M \qquad (4)$$

where H is the ice thickness, $\nabla = (\partial/\partial x, \partial/\partial y)$, $\bar{v}$ the horizontal velocity field, and M the mass balance. The temperature distribution within the ice sheet comes from the thermodynamic equation, with the assumption that horizontal conduction is negligible as compared to the other terms:

$$\frac{\partial T}{\partial t} = -\left(\frac{k}{\rho C_p}\right)\frac{\partial^2 T}{\partial z^2} - \bar{V} \cdot \nabla T + \left(\frac{\phi}{\rho C_p}\right) \qquad (5)$$

where T is the temperature, k the thermal conductivity, C_p is the specific heat capacity, $\bar{V}$ the 3-dimensional velocity vector, $\nabla = (\partial/\partial x, \partial/\partial y, \partial/\partial z)$, and ϕ the heating provided by the deformation:

$$\phi = 2\dot{\epsilon}_{xz}\tau_{xz} + 2\dot{\epsilon}_{yz}\tau_{yz} \qquad (6)$$

The bedrock adjustment is described by a diffusion equation (Lliboutry, 1965; Oerlemans and Van der Veen, 1984):

$$\frac{\partial h}{\partial t} = D_a\nabla^2(h - h_0 + w) \qquad (7)$$

where h is the bedrock elevation at time t, h_0 the unloaded bedrock, and $w = H\rho/\rho_m$ the deflection of the lithosphere, with $\rho = 910$ kg/m^2 and $\rho_m = 3300$ kg/m^2 the ice- and the mantle density, respectively. $D_a = 5.10^7$ m^2 a^{-1} is a diffusion coefficient corresponding to an adjustment time L^2/D_a of 20,000 years for an ice sheet of a width $L = 1000$ km.

The main differences between the previous Antarctic model study and the present Greenland model study are concerned with ice shelves and mass balance. In Greenland, there are presently no ice shelves large enough to play a significant role for the mass balance of the ice sheet. Although ice shelves probably formed at some locations around the Greenland coast line in cold glacial climates, recent glacial-geological studies from North Greenland (the most likely regions for ice-shelf formation) indicate that the ice sheet, in spite of a likely lower sea-level stand during the last glacial period, did not expand far beyond the present coastline (England, 1985; Houmark-Nielsen et al., 1990). In the model experiment, the extent of the ice sheet is, accordingly, at all times limited to the present coastline. If the ice sheet reaches this boundary, then the model simulates calving by setting the ice thickness to 0. This means that the model does not account for details of the calving physics, but treats calving as a "passive" ablation process. That this assumption is reasonable is supported by the fact that a simulation of the ice sheet with the present mass-balance distribution and with the concept of calving as a "passive" ablation process results in an ice-sheet that looks surprisingly similar to the actual one (see Fig. 1), although calving accounts for about half the mass wastage from the present Greenland ice sheet. The ice sheet model does not simulate melting at the bottom. Experimental runs showed that that most of the ice sheet base was below the melting point.

The mass balance

The mass balance of the ice sheet depends on the climate. Here, we use a temperature record as climate forcing during 150,000 years. Since the two components of the mass balance, accumulation and ablation, response differently to changes in temperature, they are parameterized separately.

Accumulation

The distribution of the accumulation rate on the Greenland ice sheet is rather complex: it depends to a large extent on cyclonic activity and, therefore, cannot be modelled by using the saturation vapor pressure at the temperature prevailing

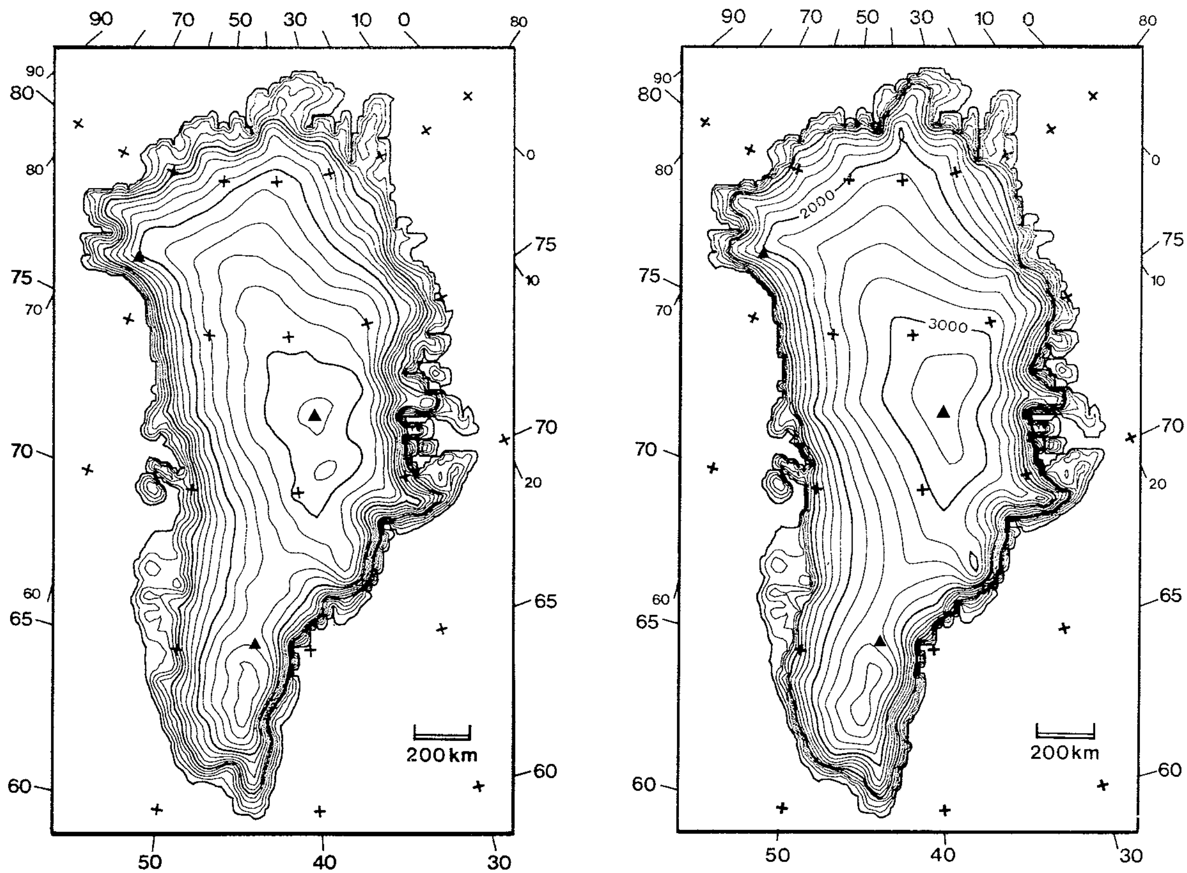

Fig. 1. Surface topography of Greenland. a. Measured. b. Simulated by the model, in steady state. The ice-margin contour is not explicitly shown, but can nonetheless be easily recognized, where the altitude contour lines (every 200 m) become very dense: this corresponds to the steep front of the ice margin. Triangles indicate deep drilling sites in Greenland: Camp Century in the northwest, Summit in the center, Dye 3 in the south.

in the atmosphere above the surface inversion layer, a method which works well for interior Antarctica (Lorius et al., 1985). One might think of using a general atmospheric circulation model to generate the accumulation distribution over Greenland. However, because atmospheric circulation models are not yet very successful in predicting precipitation, and also in order to keep computation times within acceptable limits, we simply used the present accumulation distribution (Ohmura and Reeh, 1991), corrected for climatic temperature variations according to the equations:

$$acc(t) = acc(0) * 1.0533^{DT(t)} \text{ for } DT(t) < 0, acc(t) = acc(0) \text{ for } DT(t) > 0 \quad (8)$$

where $acc(t)$ and $acc(0)$ are accumulation rates at time t and the present time, respectively. $DT(t)$ is the deviation of the temperature at time t from the present. The 5.33% change in accumulation rate per 1°C change in temperature is suggested by correlating annual accumulation rates derived from shallow ice cores in Central Greenland with the corresponding $\delta^{18}O$ values (Clausen et al., 1988), and converting $\delta^{18}O$ to temperature by means of the factor 0.62‰ $\delta^{18}O$ per °C valid for present-day Central Greenland conditions (Dansgaard, 1961). For a climatic temperature decrease of 10–12°C (representing full glacial climate conditions), accumulation rates calculated from the above equation are about half of the present value. This change agrees well with other estimates of accumulation-rate changes in Greenland during the last glacial/interglacial cycle (Reeh, 1990).

Ablation

Ablation is described by means of a positive degree-day model (Reeh, 1991). Although melting of ice and snow in the marginal areas of the ice sheet depends on the details of the surface energy balance (Ambach, 1963), we assume that air temperature and snow accumulation determine the melt processes. Studies on West Greenland ice-margin locations indicate a high correlation between positive degree-days (PDD) and melt rates (Braithwaite and Thomsen, 1984). The melt-rate model also includes a stochastic term accounting for short term temperature deviations from the average annual temperature cycle, which in the model is represented by a cosine function. This allows for the probability of having positive temperatures even when the average temperature cycle indicates temperatures below the freezing point (Braithwaite and Olesen, 1989):

$$PDD = \frac{1}{\sigma\sqrt{2\pi}} \int_0^A \int_0^{+\infty} \exp\left(- \frac{(T - TD)^2}{2\sigma^2} \right) \mathrm{d}T \, \mathrm{d}t \quad (9)$$

where PDD the number of positive degree-days, TD the daily temperature and σ the standard deviation of the daily temperature. The yearly temperature cycle is described with a sine:

$$TD = TMA + (TJ - TMA)\cos\left(\frac{2\pi t}{A}\right) \quad (10)$$

where t the time, A one year. TJ and TMA are the July and mean annual temperature, the present distributions of which are parameterized in terms of surface altitude and geographical latitude by means of the temperature distribution maps of Ohmura (1987).

By means of the positive degree days calculated as described above, ablation is computed in two steps: first the snow is melted, at a rate of 0.003 m/PDD. The first 60% of the melted snow refreezes to form superimposed ice, the rest runs off. Then superimposed ice and glacier ice is melted, using a higher melt rate of 0.008 m/PDD. The different melt rates used for snow and ice account for the albedo difference (Braithwaite and Olesen 1989). The production of superimposed ice creates a warming of the snow or ice surface that is taken into account in the modelling of the thermodynamics of the ice sheet.

Bedrock topography

The topography of the bed below the ice sheet as well as that of the surrounding ice-free land and sea bottom is needed as input to the model. The topography of the ice-sheet bed was constructed from ice-sheet surface and thickness data obtained by radio-echo sounding measurements (Electromagnetic Institute, Technical University of Denmark) and that of the surrounding ice-free land from the ETOP05 world data base. The data was interpolated and smoothed on a 20 × 20 km grid containing 141 × 83 = 11,703 points. In order to properly deal with large vertical variations of ice velocity and temperature, 14 vertical layers were used in the model, gradually decreasing in thickness towards the bottom. The thermodynamic equation was solved down to 2 km rock depth, where a geothermal heat flux of 42 mW/m^2 (1 heat flux unit, mean value for Precambrian rock) was used as boundary condition.

With a time step of 2 years (such a small time step is necessary to ensure computational stability on the fine spatial grid applied), a model run takes around 11 hours of CPU time for a 150,000-year integration on the Cray-2 computer of the University of Stuttgart, FRG.

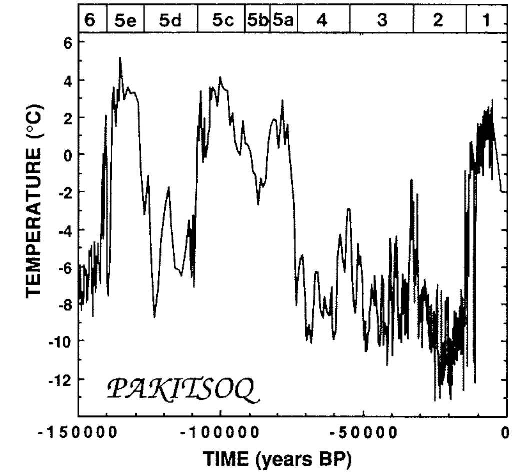

Fig. 2. Record of temperature deviations with respect to the present used to drive the Greenland ice sheet model, obtained by translating the $\delta^{18}O$ record obtained from the surface ice sampling in central West Greenland into temperature (Reeh et al., 1991). The Emiliani isotopic stages are also shown.

Steady-state reference

For the two-fold purpose of testing the model and producing a reference with which to compare the ice-sheet evolution during the past 150,000 years, the present ice sheet was simulated in steady-state (Fig. 1b). The similarity with the present ice sheet (Fig. 1a) is quite good. The simulated ice sheet is slightly larger and thicker, occupying 1.78×10^6 km^2 and containing 3.21×10^6 km^3 of ice as compared to $1.67 \times$ km^2 and $2.83 \times$ km^3 for the present ice sheet.

The shape of the ice sheet in a glacial climate was also simulated, applying a temperature 10° colder than at present and letting the program run until a steady-state was reached. This simulation was used as initial ice-sheet configuration for the calculation of the ice-sheet evolution during the past 150,000 years.

The climate history

The ice-dynamic model is driven by temperature perturbations derived from an oxygen-18 record measured on surface-ice samples collected from the ice margin at Pakitsoq, Central West Greenland (69° 26′N, 50° 16′W) (Reeh et al., 1991). This record was choosen in preference to one of the deep ice cores because the Pakitsoq record is the only one that seems to cover at least one full glacial/interglacial cycle of the Greenland history. The pre-Holocene part of this record was originally deposited in the central region of the ice sheet and, therefore, is only to a minor extent affected by the past advances and retreats of the ice margin. The record of past temperature deviations with respect to the present is shown in Fig. 2. It was derived by translating the oxygen-18 record (after correction for the $\delta^{18}O$ values presently found at the sites where the snow was originally deposited) into temperature deviations with respect to the present by means of the conversion factor 0.62‰ $\delta^{18}O$ per °C (Dansgaard et al., 1961). Thus, the curve in Fig. 2 represents the temperature history at the ice-sheet surface in Central Greenland.

The temperature record covers the period between 150,000 and 5000 yr B.P., on the time scale derived for the Vostok deep ice core (Lorius et al., 1985), i.e. from the end of Emiliani Isotopic Stage (EIS) 6 to the middle of EIS 1. As discussed by Lorius, there is good correspondence between the Vostok time scale and the time scale established for the deep sea isotopic records (Martinson et al., 1987) back to 110,000 yr B.P. Prior to this time, the two time scales may differ by as much as 10,000 years. Since the chronology of the ice

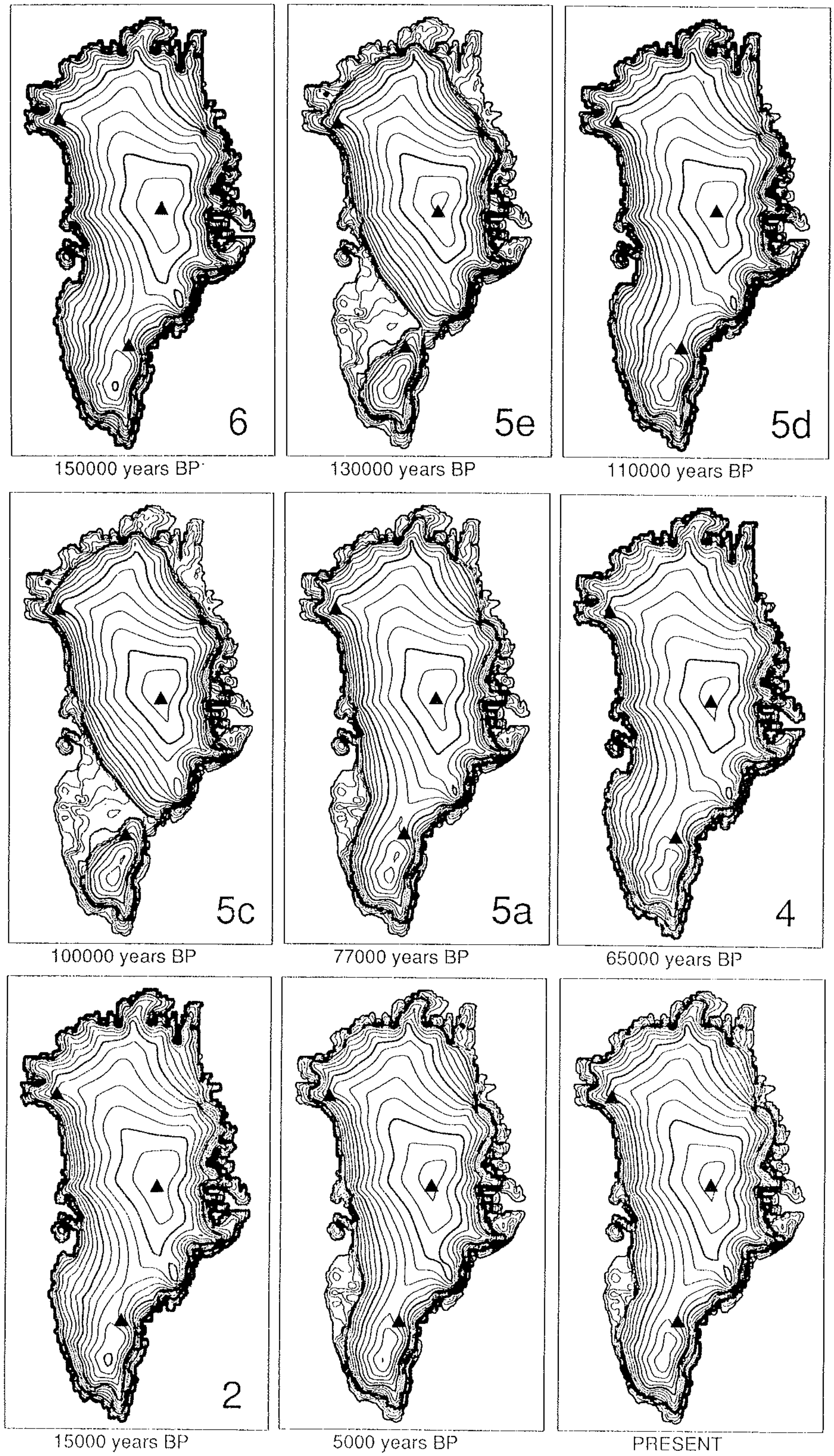

Fig. 3. Extent and surface topography of the Greenland ice sheet at selected times during the 150,000 years reconstructed evolution, with the corresponding Emiliani stages in the bottom right corners. Triangles show the location of the deep drilling sites (past and coming) on the ice sheet: Camp Century, Dye 3, and Summit.

margin isotopic-temperature record is established by correlation of isotopic stages, it is encumbered with a similar uncertainty.

From 5000 yr B.P. to the present, the temperature record is extended by means of palynological evidence from Greenland, summarized by Kelly (1980): from 5000 to 3000 yr B.P. the temperature deviation with respect to the present is changed linearly from circa 2°C to 0°C, and then kept constant until now.

Results

The evolution of the Greenland ice sheet during the past 150,000 years is illustrated in Figs. 3–5. Figure 3 shows the extent and surface topography of the ice sheet at selected times. The corresponding Emiliani isotopic stages are also indicated. Figure 4 shows the variation in surface altitude at the previous (Camp Century and Dye3) and coming (Summit) deep drilling locations in Greenland. Figure 5 shows the volume changes of the ice sheet. A scale showing the equivalent contribution to global sea-level changes is added to the right.

The simulated present surface elevations at the ice-sheet locations Camp Century, Dye3, and Summit are 1977, 2491, and 3431 m, as compared

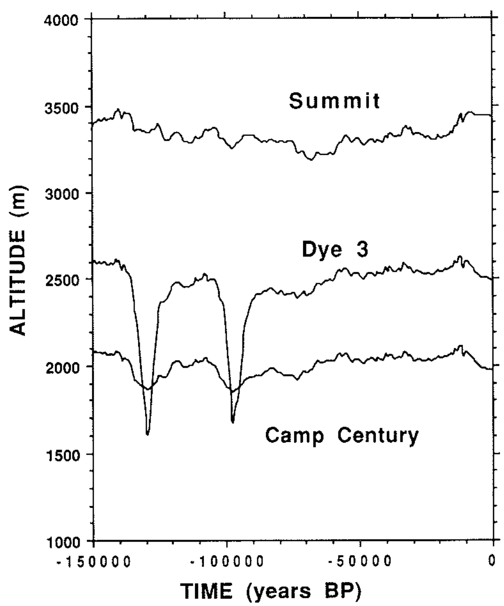

Fig. 4. Variation in surface altitude at the 3 deep drilling sites.

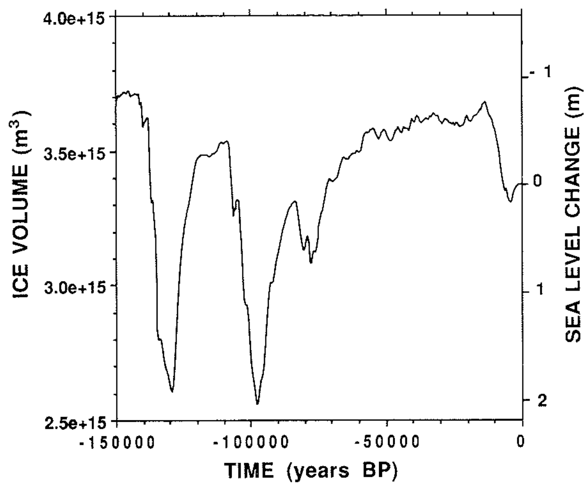

Fig. 5. Volume changes of the Greenland ice sheet during the 150,000 years of reconstructed evolution. The scale on the right indicates the corresponding sea level changes.

with the actual elevations of 1880 (Gundestrup et al, 1987), 2480 (Overgaard and Gundestrup, 1985), and 3233 m (Hodge et al, 1990), respectively. These results are encouraging, suggesting that the dynamic model is able to reproduce the essential features of the real ice-sheet behavior.

During the Eemian (the previous interglacial, 130,000 yr B.P., EIS 5e) and the interstadial 100,000 yr B.P. (EIS 5c) the warm climate, with temperatures up to 5°C warmer than at present, caused the ice margins to retreat and the ice sheet to split up into two parts, one main ice sheet covering the central and northern parts of Greenland, and a much smaller ice cap over the southern highlands. The reason for this ice configuration results from some characteristics of the bedrock altitude and the accumulation distribution: The ice-free area has a relatively low elevation as well as low accumulation. The position of the margin is then mostly controlled by the ablation, and will response quickly to a warming, even if the warming is moderate. However, on the southern dome, high altitudes as well as a very high precipitation rate create an ice cap that is not so sensitive to warming. The remaining central/northern ice sheet, especially on its eastern and western sides, is characterized by higher altitude and latitude (leading to lower ablation) and higher precipitation rate than the ice-free area.

The central/northern ice sheet is then also somewhat less sensitive to climate warming than the ice-free area.

It is interesting to note that, at that time, the deep-drilling location at Dye3 is in the ablation zone, very close to the margin (see Fig. 4). In terms of sea-level change, the reduced ice-sheet volume is equivalent to a sea-level rise of about 2 m above the present level.

The isotopic stages 4, 3 and 2 were a 60,000 year period with large climatic fluctuations, but significantly colder than at present. At the beginning of that period, the ice-sheet margins reacted quickly to the decrease in temperature, and within a few thousand years they had advanced to the present coastline. However, the basal ice has retained temperatures characteristic of the previous warm interstadial, creating a thinner ice sheet than in the steady-state initial configuration. A slower evolution then took place during the long cold period: the cold conditions imposed at the surface of the ice sheet created a cold wave that slowly propagated to the bottom, causing less deformable ice there, so that the ice thickness had to increase. At the end of the ice age, the ice sheet was very close to the glacial steady-state configuration used as initial condition for the simulation.

An important result of the model simulation is that even in periods when the ice-margin had advanced to the present coastline, the surface altitudes along the central ice divide were not much different from the present altitudes, see Fig. 3. This confirms the results of a previous model study with a much simpler perfectly-plastic ice-sheet model (Reeh, 1984). An equally important result is that, as far as the Central Greenland summit area is concerned, surface altitudes do not change very much, either, in periods when the ice margins had retreated considerably with respect to their present positions, see Figs. 3 and 4. This should make easier the interpretation of the ice-core records from the planned deep-drilling programs in this region of the ice sheet.

Discussion and conclusion

The uncertainty concerning the results of the model-simulation arises primarily from the climate forcing and to a much lesser extent from the imperfections of the ice-dynamic model. Our reasons for believing the results are that we have applied an ice-dynamic model which accounts for the most essential features of ice-sheet dynamics and thermodynamics, i.e. the non-linear rheology of glacier ice, the varying temperatures within the ice sheet caused by climatically induced changes of the temperature at the ice-sheet surface, the temperature dependence of the ice-flow properties, and the thermal inertia of the ice-sheet substratum. Moreover, glacial isostasy is considered. Melting is controlled by climatically varying air temperatures by means of a degree-day model that distinguishes between snow and ice melt. Also snow accumulation depends on the temperature in the way that cooler climates cause decreasing accumulation rates.

Not accounted for in the model is the change in ice-flow properties due to the fact that, during a glacial/interglacial cycle, the ratio between relatively soft glacial ice and relatively hard interglacial ice in the ice sheet varies because of flow-induced downward migration of the boundaries between the two kinds of ice (Reeh, 1985). A more serious problem is that probable changes in the accumulation distribution over the ice sheet resulting from likely changes of the atmospheric circulation pattern during the glacial/interglacial cycle are not accounted for in the model. However, sensitivity experiments show that the accumulation distribution has a moderate influence on the simulated evolution of the ice sheet (Letréguilly et al., 1991).

The result of the simulation depends strongly on the climate temperature history used to drive the model. We have used a temperature history derived from an ice-margin $\delta^{18}O$ record from central West Greenland (Reeh et al., 1991), which indicates that the climate in Greenland was warmer than at present, not only in EIS 5e (the Eemian interglacial), but also in EIS 5c and 5a (interstadials according to the deep sea isotopic records). Isotopic stages 5e and 5c were sufficiently warm and long to cause a substantial reduction in the extent and volume of the ice sheet, equivalent to a sea-level rise of about 1.5 m. The coral-reef sea-level curves respectively indicate a 6 m higher and

12 m lower sea level than at present in EIS 5e and EIS 5c (Chappel and Shackleton, 1986). We conclude, that in EIS 5e, glaciers, small ice caps, and the Antarctic ice sheet made a significant contribution to sea-level rise. However, to explain a 12 m sea level decrease in EIS 5c, it would have been necessary to store an ice volume corresponding to about twice the volume of the Greenland ice sheet on some other continent.

There is no indication of EIS 6 ice in the deep ice records from Dye 3 in South Greenland and Camp Century in Northwest Greenland. As to South Greenland, the model results show that during the Eemian interglacial, as well as during the stadial 5c, Dye 3 lies very close to the ice margin, and is in the ablation zone. Either a rise of 1° C during that period, or a lengthening of it (both of those possibilities lie within the uncertainty range of the temperature history) would be enough to make the ice disappear at Dye 3. Also the ice margin in the Camp Century region retreated substantially during EIS 5e and EIS 5c, and the ablation area was not very far from Camp Century. Increasing the amplitude of the temperature variations for North Greenland, as actually indicated by the much larger amplitudes of the Camp Century δ^{18}O record in respect to the Central and South Greenland records (Reeh et al., 1991; Dansgaard, 1984) would result in a larger ice-volume loss in North Greenland in the warm periods than shown in Fig. 3. This would make the margin of the ice sheet retreat to a position inland of Camp Century, which could then explain the missing EIS-6 ice in the Camp Century ice-core record.

The complete disappearance of the ice sheet never took place, the Central Greenland part remains in all the simulations tried. The ice sheet seems to have been more stable than had been suggested (Koerner, 1989). Part of this is due to isostatic uplift when the ice disappears, which amounts to up to 1000 m in Central Greenland. When the margin retreats, this is enough to bring it back to areas of higher elevations with reduced ablation, which slows down the retreat considerably.

In conclusion, we can say that, by combining the climatic record derived from a Greenland ice-margin δ^{18}O study with an advanced ice-sheet dynamic model, we have succeeded in simulating a history of evolution of the Greenland ice sheet which is consistent with some basic features of the ice-margin and ice-core records presently available. It shows that the ice sheet could not have disappeared during the Eemian interglacial, although the ice margin may have changed. This suggests that ice from EIS 6 (the previous glacial) still exists in Central Greenland, which supports the climate history derived from the ice-margin isotopic record (Reeh et al., 1991).

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