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MODELLING OF PLEISTOCENE EUROPEAN ICE SHEETS: THE EFFECT OF UPSLOPE PRECIPITATION[1]

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ABSTRACT

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Results are presented from a numerical model of the Scandinavian Ice Sheet, in which the effect of upslope precipitation is included explicitly. The model is forced by changing the environmental conditions, formulated in terms of the annual mean temperature and the annual temperature range. These factors determine snowfall and melting rates, in dependence of the local conditions.

It appears that orographically induced precipitation, which causes zones of high precipitation to shift with the ice-sheet edge, and ocean temperature are very important with regard to the growth rate of the Scandinavian Ice Sheet. In particular, upslope snowfall causes the ice sheet to advance westwards into the North Sea region much more easily.

Stable equilibrium states of the ice sheet were calculated for various climatic conditions. A large ice sheet is only possible in cold conditions, whereas under very warm conditions no ice cover can be maintained. However, in between is a range of temperatures (2.5 to 6 K lower than present temperature) for which three stable equilibrium states exist: (i) no ice sheet, (ii) small ice sheets in the Scandinavian mountains, and (iii) a large ice sheet. This indicates that the response of the Scandinavian Ice Sheet to time-dependent forcing will be very complex.

INTRODUCTION

When inspecting precipitation maps of the midlatitudes, it is immediately obvious that the presence of mountain ranges has a pronounced effect on precipitation patterns. In Europe, for example, the mountainous areas of Scotland, Norway and central Europe have up to three times more precipitation than the surrounding flat regions. This abundance of precipitation is caused by the forced ascent of air up the mountain slopes. When the wind has a direction of preference, a mountain range creates a distinct asymmetry: high precipitation rates to windward, and low precipitation rates in the lee.

Of course, such processes also operate at the edge of a large continental ice sheet. The Antarctic Ice Sheet illustrates this very well (although there are other effects, too): some regions near the coast have an annual precipitation of up to 0.6 m, while flatter regions receive less than 0.05 m. In studying the growth of the Pleistocene European ice sheets, it is therefore of some importance to include the effect of upslope precipitation. A very interesting point against this background is that ice sheets form moving mountain ranges: an ice-sheet edge is capable of taking the high upslope precipitation with it. This may substantially affect the way in which an ice sheet grows.

In this study the flow model of the European Ice Sheet described in OERLEMANS (1981a) is linked to a precipitation model that includes the effect of surface slopes. The precipitation model consists of an advection equation for precipitable water, including a source (evaporation) and a sink (precipitation). The wind field advecting the moisture is prescribed. The advection equation is integrated until a steady state is reached, which is then supposed to reflect annual mean conditions. The model parameters are chosen in such a way that a reasonable simulation of the present-day precipitation pattern is obtained.

Once the precipitation model is tuned and linked to the computation of the annual ice accumulation rate (this also involves computation of the ice/snow melt), it is possible to study how the interaction of upslope snow fall and ice sheet evolution affects the creation of a large European ice sheet. Model experiments are discussed that shed some light on this. The model also allows sensitivity tests to changes in surface temperature of the adjacent ocean (which is the main source of moisture). Results of such tests will be presented.

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DESCRIPTION OF THE MODEL

The ice-sheet model

The description of the flow of ice is done by means of a nonlinear diffusion equation for ice thickness H (see OERLEMANS, 1981a for details):

$$\frac{\delta H}{\delta t} = \nabla D \nabla (H+h) + G \qquad (1)$$

$$D = AH^{m+1} \left[\left\{ \frac{(H+h)}{x} \right\}^2 + \left\{ \frac{(H+h)}{y} \right\}^2 \right]^{\frac{m-1}{2}} \qquad (2)$$

In these equations G is the yearly ice accumulation rate, H is ice thickness, h is bedrock elevation with respect to some equipotential level, A and m are flow constants, t is time and x and y are the space variables. In this study m=2.5 and $A=3\ m^{-3/2}\ yr^{-1}$.

To include the effect of ice load on the earth's crust, an equation is added that describes a local, damped return to isostatic conditions:

$$\frac{\delta \zeta}{\delta t} = -\alpha\,(qH+\zeta) \qquad (3)$$

ζ is the deflection of the bedrock with respect to the equilibrium value (no ice), q is the ratio of ice density to rock density and α^{-1} is the time scale on which the return to isostatic equilibrium takes place. Throughout this study values of 0.3 and 8000 yr are used for q and α^{-1}, respectively.

The precipitation model

Calculation of the precipitation rate is based on the conservation of moisture:

$$\frac{\delta W}{\delta t} = -\vec{v}.\nabla W - (f_0 + f_1 S)\,W + (W_m - W)/T_* + D_w \nabla^2 W \qquad (4)$$

W is the amount of water vapour in an atmospheric column extending from the surface to the 'top' of the atmosphere, W_m is its maximum value (it depends on temperature and surface elevation). W is advected by the (prescribed) mean wind field $\vec{v}$.

The second term on the right-hand side of eq. (4) models the precipitation. The precipitation rate is proportional to the moisture content W and consists of two parts: background precipitation determined by f_o, and upslope precipitation determined by f_1.

The third term is the moisture source. T_* is a characteristic time for evaporation from the surface to an atmospheric column and depends on the nature of the surface. It can be estimated from standard bulk formulas (e.g. KRAUS 1972).

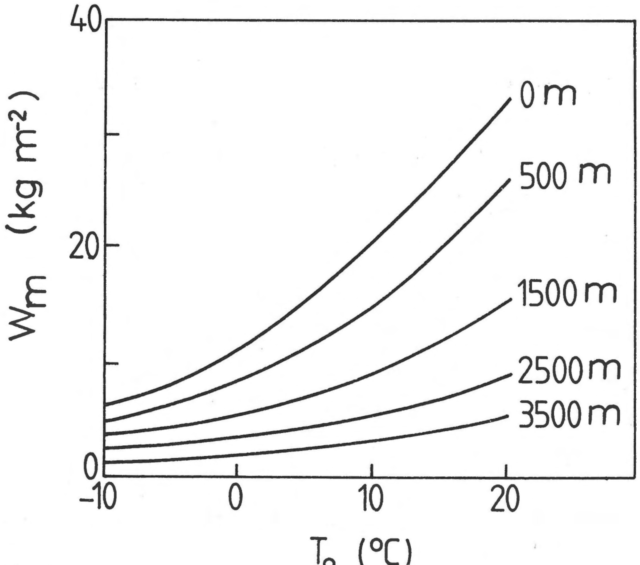

Fig. 1
Maximum water vapour content of an atmospheric column in dependence of sea level temperature T_o. Different curves are for different surface elevations.

Values used in this study are 3 days over water, 6 days over water, and 30 days over ice sheets.

The last term in eq. (4) accounts for horizontal mixing of moisture. It is very difficult to estimate the moisture diffusivity D_w for an annual mean precipitation model. Here a value of $5.2 \times 10^5\ m^2/s$ is used, which is an ambiguous choice. Fortunately, the results are not very sensitive to the particular value of D_w.

In order to solve eq. (4), it is necessary to relate W_m to surface temperature and elevation. This can be done by employing standard atmospheric thermodynamics (e.g. IRIBARNE & GODSON, 1973), using a fixed atmospheric lapse rate (6 K/km). Fig. 1 shows the result. In particular when sea level temperatures are high, the 'surface elevation effect' is substantial.

To tune the precipitation model, a number of experiments were carried out in which eq. (8) was integrated to a steady state, given the present orography and distribution of annual temperature (see p. 269 *Numerical method and bonndary conditions* for grid and numerical aspects). It appeared that the model performs best with a 10 m/s westerly wind, $f_o=0.188\ s^{-1}$ and $f_1=0.353\ s^{-1}$. In the tuning experiments, only the area north of 50 °N was taken into account.

A comparison between observed and calculated annual precipitation patterns is made in Fig. 2. The observed distribution of annual precipitation is derived from mean values over 100 by 100 km squares, and therefore local maxima of 2 m/yr, which appear in Norway and the Alps, are missing. For the same reason, the plotting routine produced a coastline that deviates from the actual one in regions with very small slope.

Comparing the maps it appears that the precipitation model produces a reasonable distribution of annual precipitation and simulates the upslope effect well. The pattern correlation

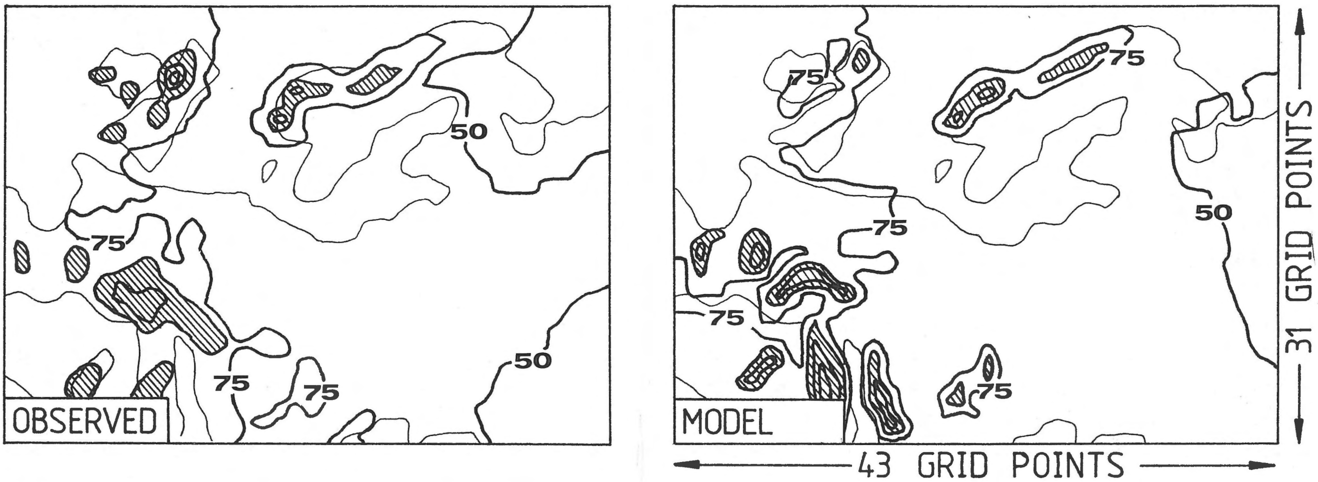

Fig. 2
A comparison between observed and computed amounts of annual precipitation (in cm) in Europe.

coefficient of the maps is 0.66. The model overestimates precipitation in southern Europe, for obvious reasons: the model makes no distinction between regions in the westerlies and regions at the fringes of the subtropical anticylone. In the absence of large scale circulation effects, the high water temperatures in southern Europe lead to high precipitation rates. Since these regions are not important in the present context, no attempt was made to improve the model for southern Europe.

To calculate the rate of ice accumulation, the fraction of precipitation that reaches the surface as snow has to be specified. From snow fall data from Northern America (*World Survey of Climatology,* 1974) the following empirical relation was calculated

$$SN = P(F_T+0.11). \qquad (5)$$

SN is the amount of snow, P the annual precipitation, and F_T the fraction of the year during which the surface temperature is below 1°C. This relation holds only for cold climates, of course. Seasonal differences in the rate of precipitation are not explicitly taken into account.

Computation of ice/snow melt

The ice/snow melt ME is computed from the monthly mean surface temperature T_m (in °C). So

$$ME = \max\,[o;\ aT_m+b], \qquad (6)$$

where a and b are empirical constants. According to eq. (6), ME equals zero if $T_m \leq -b/a$. The constant b is positive, accounting for the fact that even when the monthly mean temperature equals zero some melting nevertheless takes place. Fitting eq. (6) to mass-balance studies in Norway (PYTTE & ØSTREM, 1965) values of 0.1 m month^{-1} K^{-1} and 0.05 m month^{-1} were found for a and b.

The amount of melt can now be calculated if the seasonal cycle in T_m is known. However, one refinement was made to account for the effect of extensive melting on the environmental conditions. This was done by using

$$ME_y = [NME_o+(4-N)ME]/4. \qquad (7)$$

In eq. (7), ME_y is the ice/snow melt at the grid point y, ME is the ice/snow melt computed from a sinusoidal seasonal cycle in temperature at that point, and ME_o is the ice/snow melt computed from the same seasonal cycle except that it is cut off at T=0°C (so that the temperature is never above the freezing point). N is the number of grid points surrounding y which have ice ($N \leq 4$). This procedure account for the fact that over vast ice-covered regions surface temperatures are below or close to (in summer) the freezing point. So substantial melting can only occur near the edges of an ice sheet.

Numerical method and boundary conditions

The equations making up the model are solved on a grid of 43 to 31 = 1333 grid points, spaced at 100 km. This yields the domain shown in Fig. 2 (on a stereographic projection). Central differences are used to approximate derivatives in space. An explicit forward scheme performs the integration in time.

The ice-sheet model has a characteristic time scale which is much longer than that of the precipitation model. For the ice-sheet model a time step of 10 yr is used. The precipitation model (time step 2500 sec) is integrated to a steady state each 500 yr of simulated time. So the calculated distribution of precipitation is kept fixed during 500 yr.

Boundary conditions are as follows. For the ice sheet model:
if topography <-200 m (with respect to present sea level)

$$H(t)=0; \qquad (8)$$

on the boundary of the model domain

$$\delta H(t)/\delta n=0. \quad (9)$$

Eq. (8) states that the ice sheet cannot extend beyond the continental shelf. In eq. (9), n is the direction perpendicular to the boundary of the model. For the precipitation model:

$$\delta W/\delta n=0. \quad (10)$$

In addition to these conditions, present-day climatology is used to specify the annual sea level temperature field. Drops in temperature needed to initiate ice-sheet growth will always be expressed with respect to this field.

The annual temperature range (ATR) is parameterized as

$$\mathrm{ATR} = 15+0.4\lambda-2\times10^{-3}h\mathrm{K}, \quad (11)$$

were λ is longitude in °E and h is the surface elevation in m. Eq. (11) was derived from data on climatic conditions in northern Europe.

Forcing

The model described above can be forced by changes in surface temperature. Such temperature changes can for example be due to variations in the amount of solar radiation received at the top of the atmosphere (on which the Milankovitch theory of the Pleistocene glacial cycles is based).

Within the present model formulation, a temperature change may consist of a change in annual mean temperature and/or a change in the amplitude of the annual temperature range. Also, the temperature change can depend on the geographical location. In particular when solar radiation variations are considered, the magnitude of the temperature variations should increase with latitude. Here we use

$$\triangle T_{\mathrm{land}}=\triangle T_{70^\circ\mathrm{N}}\,(\varphi/50^\circ-0.4)\ \mathrm{K}, \quad (12)$$

where φ is latitude. Eq. (12) implies that at 70°N the temperature perturbation is twice that at 45°N.

Since the temperature of the ocean surface plays an important role in the moisture budget over Europe, care has to be taken in changing this model parameter. According to CLIMAP (1976), the drop in surface temperature over the North Atlantic Ocean during a full glaciation is about 6 K. Connecting this to a Scandinavian ice volume of $10^{16}\,\mathrm{m}^3$ gives a 'better than nothing' relation to calculate the drop in ocean surface temperature:

$$\triangle T_{\mathrm{sea}}=-6\times10^{-16}V\ \mathrm{K}. \quad (13)$$

Here V is the volume of the Scandinavian Ice Sheet. Of course this approach requires that a balance between ocean temperature and Scandinavian ice volume exists, which is somewhat doubtful. However, it seems that a sound alternative does not exist.

RESULTS

Growth rates

In a first experiment a number of runs were carried out to see how the model ice sheet approaches equilibrium for various values of $\triangle T_{70}$ (temperature change at φ=70°N) and the ATR (annual temperature range).

First all it appeared that a substantial temperature drop is needed to initiate ice sheet growth (in all experiments, the temperature drop was imposed instantaneously at t=0). The following criterion was found for ice-sheet inception:

$$\triangle T_{70}+7.5F\leqslant1.9. \quad (14)$$

F is the ATR divided by its present value, and $\triangle T_{70}$ is in K. So without any change in the ATR, a temperature drop of more than 5.6 K is needed to create ice sheets. However, these ice sheets remain very small ($V\simeq10^{14}\,\mathrm{m}^3$) and are restricted to the Norwegian mountains. A temperature drop of at least 7 K is required to obtain further growth of the ice volume. Then a huge ice sheet is formed with a typical volume of $10^{16}\,\mathrm{m}^3$. This type of behaviour will be discussed in more detail in the section on steady states.

Fig. 3 shows growth to equilibrium for four combinations of $\triangle T_{70}$ and F. The results clearly show how sensitive the growth

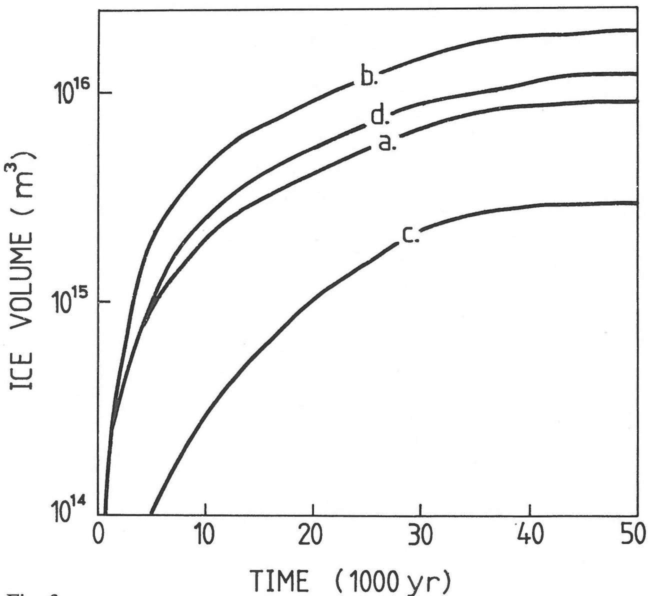

Fig. 3
Ice volume (on a logarithmic scale) as a function of integration time. Imposed conditions are as follows: Run a: $\triangle T_{70}=-10$ K and F=1; Run b: $\triangle T_{70}=-10$ K and F=0.75; Run c: $\triangle T_{70}=-7.5$ K and F=1; Run d: $\triangle T_{70}=-7.5$ K and F=0.75. F is the annual temperature range normalized to its present value.

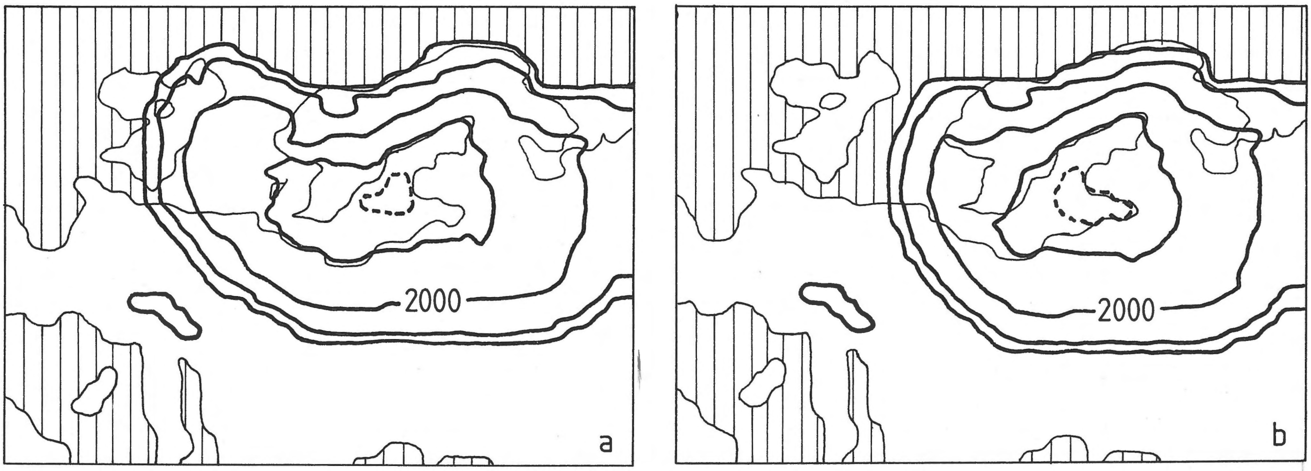

Fig. 4
Distribution of ice thickness after 45 000 years for the standard model (a) and a model in which interaction between shape of the ice sheet and orographically forced snowfall has been omitted (b). The contour interval is 1000 m. The imposed environmental conditions correspond to run d.

rate is to the annual temperature range. The reason for this is obvious: melting rates strongly increase when the annual temperature range becomes larger. In particular in case of a temperature drop close to the critical value (runs c and d) the dependence on the ATR is striking.

In Fig. 4a the distribution of ice thickness is shown for run d, at t=45 000 yr. The maximum ice thickness is about 3500 m and occurs over the Gulf of Bothnia. The steep edges of the ice sheet are a consequence of the nonlinear character of the flow law, and the dependence of the mass balance on elevation. The extent of the ice sheet resembles that during the Weichselian (e.g. LAMB, 1977), although at some places the southward penetration of the ice masses is somewhat more pronounced in the model.

To see the effect of upslope precipitation, run d was repeated with fixed orography in the calculation of snowfall. In that case the shape of the ice sheet does not affect the distribution of precipitation. The result is shown in Fig. 4b, also at t=45 000 yr. Comparing this to Fig. 4a shows that now the westward extention of the ice sheet is reduced: it has not crossed the North Sea. From this result it can be concluded that upslope precipitation favours the ice-sheet growth in the upstream direction (with regard to the dominating wind direction in the atmosphere).

Another sensitivity experiment involves the surface temperature of the North Atlantic Ocean. Two experiments were carried out to investigate the effect of ocean temperature on the growth rate of the ice sheet. In the first run temperature over land was lowered according to run a in Fig. 3, but the sea-surface temperature field was kept fixed to the present one. The growth rate curve for this run is shown in Fig. 5 and has the label e. Ice volumes are given relative to those of run a. Apparently, high ocean temperatures lead to a more rapid built-up of the ice sheet, and in fact the ice sheet soon covers an area which is unrealistically large (it extends to the Mediterranean Sea).

In the second run the temperature drop over sea was prescribed as

$$\Delta T_{sea} = -3(1+V/10^{16})K, \qquad (15)$$

where V is the ice volume in m^3. So in this case the ocean temperature is already lowered by 3 K when the ice sheet starts to grow. When the ice volume increases the environmental conditions become similar to those of run a. The ice volume curve of this run has label f in Fig. 5. As has to be expected, the growth rate is smaller now: the amount of precipitable water in the atmosphere is smaller and consequently snowfall is reduced.

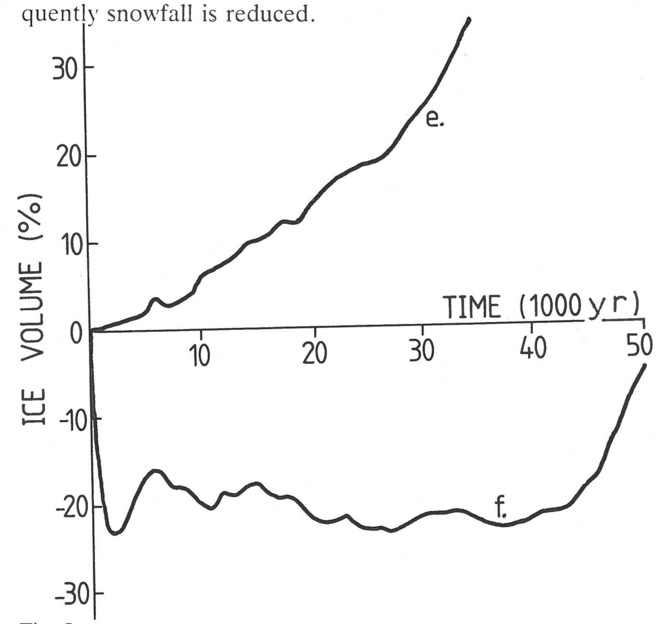

Fig. 5
Effect of ocean temperature on ice sheet growth. In run e the ocean temperatures are kept fixed to the present values. In run f the drop in ocean temperature lags behind the temperature drop over land (see text). Ice volume is given relative to that of run a in Fig. 3.

These experiments show that the evolution of the Scandinavian ice sheet is sensitive to how rapid the climate becomes colder. The best condition apparently is rapid cooling, so that the ice sheet can benefit from the thermal inertia of the oceans.

Steady states

It is of interest to know how the equilibrium state of the Eurasian ice sheet(s) depends on the climatic conditions. Because of the large amounts of computer time needed to run the present model, it is simply impossible to vary all model parameters and integrate until steady states are reached. Therefore, only the dependence of ice volume on an instantaneous temperature drop ΔT_{70} was considered as most interesting. The associated decrease in sea-surface temperature is as described in section 2e.

When searching equilibrium states one has to keep in mind that for some values of ΔT_{70} more than one stable equilibrium may exist. Due to the coupling of surface elevation and mass balance, a large ice sheet can survive moderately warm conditions while, if in these conditions no ice would be present at all, an ice sheet would not form. So large continental ice sheets show hysteresis with regard to the large scale climatic conditions (WEERTMAN, 1961; OERLEMANS, 1981b). This implies that, in order to find all stable equilibrium states, various initial conditions have to be used.

Runs were carried out for a wide range of values of ΔT_{70}, and starting from different initial conditions. The results are summarized in Fig. 6. It shows steady-state ice volume as a function of ΔT_{70}. The solid line refers to experiments in which V=0 has been used as initial condition. As $\Delta T_{70} \approx -6$ K, small ice sheets are formed in the mountains (arrow 1). At $\Delta T_{70} \approx -7$ K another critical point appears: the only possible equilibrium state now becomes a very large ice sheet (arrow 2). A further drop in temperature leads to a gradually increasing ice volume.

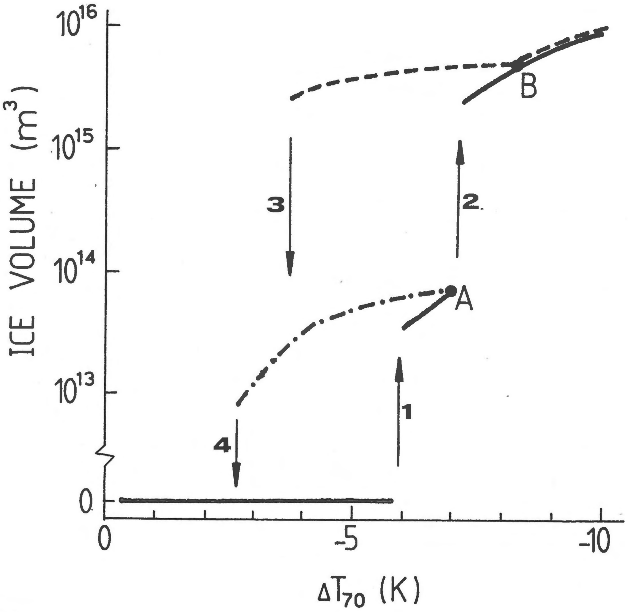

Fig. 6
Steady-state ice volume as a function of ΔT_{70}. Only stable equilibria are shown. See text text for further explanation.

The way back is different, as a consequence of the fact that the mass balance field not only depends on ΔT_{70} but also on the shape of the ice sheet itself. The temperature has to be raised until $\Delta T_{70} = -3.5$ K before the large ice sheet starts to decay. The decay is not complete: the model ice sheet jumps to a state (arrow 3) in which some ice cover in the mountains still exists. Only if $\Delta T_{70} > -2.5$ K the ice volume returns to zero (arrow 4). In this figure, B indicates the initial state for the integrations leading to the dashed line. The dot-dashed line was derived by using the ice sheet at A as initial state.

Fig. 6 clearly shows that the equilibrium ice volume depends on ΔT_{70} in a complex way. If ΔT_{70} is in the -2.5 to -6 K, even three stable equilibrium states exist! The consequence of this is that, given the environmental conditions, not much can be said about the volume of the Scandinavian ice sheet unless its history is known.

CONCLUSIONS AND DISCUSSION

The major conclusions of this study are:
- orographically-induced precipitation is important with regard to ice-sheet growth
- changes in the annual temperature range are just as important as changes in the annual mean temperature
- even when during a cooling event the ocean temperature lags behind the land temperature, at least 30 000 years are needed to create a full-grown Scandinavian ice sheet
- equilibrium states of the ice sheet depend in a complex way on environmental conditions.

It should be realized that the ice sheet and precipitation model used here is rather schematic. The restricted resolution of 100 km does not allow proper incorporation of high mountains with a horizontal scale less than 100 km. This probably is the reason why no ice sheet is initiated in Scotland.

No attempt was made to produce a simulation of the history of the Scandinavian ice sheet as can be inferred from proxy data. The reason for this is that the climatic boundary conditions are not very well known. The surface temperature over land can probably be linked to insolation, but the sea-surface temperature and the mean wind direction are unknown (they also depend on conditions in North America, for example). So one should accept the fact that at this stage only 'academic' experiments can be carried out.

The results presented here are concerned mainly with ice-sheet growth and steady states, not with the details of ice-sheet decay. Isochrones of retreat are not given because the effects of basal melting and associated sliding of the ice mass over the bed, which is important in ice-sheet decay, are not included in the ice sheet model. It would require a calculation of the heat budget of the ice sheet, which is still a problem for three-dimensional models. Experiments with a two-dimensional model including thermodynamics have shown that the process of extensive basal sliding is decisive in initiating ice-sheet decay (OERLEMANS, 1982).

Most important for future studies of this type is to know more about the relative phase of the evolution of the North American and European ice sheets. It seems to be unlikely that their growth and decay is completely isochronic, even if they are initiated at the the same time. An interesting experiment would be to couple models of the American and European ice sheets and to study their interaction through the climatic environment. However, before this can be done models should be constructed that are (computationally) more efficient than the present one.

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