The role of ice sheets in the Pleistocene climate

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Oerlemans, J.: The role of ice sheets in the Pleistocene climate. *Norsk Geologisk Tidsskrift*, Vol. 71, pp. 155–161. Oslo 1991. ISSN 0029-196X.

Northern hemisphere ice sheets have played an important role in the climatic evolution of the Pleistocene. The characteristic time-scale of ice-sheet growth has the same order-of-magnitude as that for the orbital insolation variations. The interaction with the solid earth, the importance of the thermal conditions at the base of ice sheets and feedback on the climate system (albedo feedback, precipitation regime) make the cryospheric response to climatic forcing complicated. Feedback of surface elevation on the surface mass balance allows northern hemisphere ice sheets to grow southward when cooler summer conditions prevail. Rapid ice-sheet decay, as observed in the paleo-record, must have involved one or more powerful destabilizing mechanisms like low accumulation rates at subpolar latitudes, high ice velocities due to water-saturated sediment beds, and high calving rates in proglacial lakes and seas. In terms of radiactive forcing of the *global* climatic fluctuations in the Pleistocene, the effects of northern hemisphere ice sheets (albedo), varying concentration of greenhouse gases (carbon dioxide, methane) and direct effect of orbital changes (insolation) are of similar magnitude.

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In the last few decades significant progress has been made in the modelling of land ice masses and in understanding the role they play in the climate system. Although many questions remain unanswered, a number of processes have been identified that establish a clear and important link between ice-sheet evolution and changes in the climate system. Several of these will be discussed in this contribution: direct implications of ice-sheet growth/decay for the annual energy budget, the albedo feedback, the elevation–mass balance feedback, and the effect on ocean circulation and precipitation rates.

Whether a glacier or ice sheet influences the climate depends very much on the scale: the effect of a mountain glacier will be restricted to the valley in which it flows. Many glaciers and small ice caps in a single mountain area may modify climate on a regional scale, but only for large ice sheets does one expect world-wide effects. The interesting aspect is that an effect on the local climate can still make an ice mass grow larger and larger, thereby gradually increasing its radius of influence.

At present, land ice covers an area of 15,861,766 km^2 (Haeberli et al. 1988), which is slightly more than 3% of the total area of the earth's surface and about 11% of the continents. At the peak of the last glaciation, the latter figure was in the order of 25%. By far the largest amount of present ice is stored on the Antarctic continent, about 90% of the total. Of the remainder, most is found on Greenland. Mountain glaciers and small ice caps contribute very little to the total ice volume, in fact. This total volume is estimated to be in the order of 32 million km^3, equivalent to approximately 70 m of sea-level rise when melted and spread uniformly over the world ocean. For an up-to-date summary of the distribution of glaciers and ice sheets over the globe, the reader may consult the reference quoted above.

The purpose of this paper is to discuss how ice sheets interact with the climate system. No attempt is made to review the paleoclimatic evidence of the Pleistocene glacial cycles, nor to discuss the wide variety of ice-age theories that have been proposed in the last few decades. A rather qualitative description of physical mechanisms will be given, showing that ice sheets form an *active* component of the climate system.

Ice and the energy budget of the climate system

Irrespective of the details of ice-sheet growth or decay, it is interesting to compare the amounts of energy involved with the basic energy flows in the climate system. The global balance between incoming solar radiation and outgoing terrestrial (infrared) radiation can be written as

$$(1 - A)S = \sigma T^4 = \tau\sigma T_s^4, \qquad (1)$$

where S (341 W/m^2) is the solar constant divided by 4 (because the area of the surface of the earth is four times the cross-section intercepting the sun's rays), A the mean planetary albedo (currently 0.3), σ the Stefan-Boltzmann constant (5.67×10^{-8} J/(m^2 K^4 s)), and T the effective radiative temperature of the planet. The outgoing terrestrial radiation is frequently expressed in the *surface* temperature T_s, which makes it necessary to introduce an effective transmissivity of the atmosphere for radiation emitted at the surface (denoted by τ). By inserting the values for A, S and σ it follows that $T \sim 255$ K, i.e. the radiative temperature is much lower than the mean surface temperature and the radiating level must be at several kilometres' height. This is typical for a greenhouse planet, where the atmosphere is not transparent for infrared radiation. To arrive at the observed mean surface temperature (about 288 K), it appears that $\tau = 0.61$.

The annual amounts of solar and terrestrial radiation passing the 'top of the atmosphere' are about

Table 1.

Absorbed solar radiation in climate system:	3.8×10^{24} J/yr
Absorbed at the earth's surface:	2.8×10^{24} J/yr
Heating the oceans by 1 K requires:	5.8×10^{24} J
Heating the atmosphere by 1 K requires:	5.1×10^{21}
Present mass of land ice is 2.7×10^{19} kg to melt this requires:	9.3×10^{24} J
Mass of ice in full glacial is 7.3×10^{19} kg return to interglacial requires:	1.6×10^{25} J

3.8×10^{24} J/yr. Table 1 shows how this compares to typical amounts of energy needed to heat the ocean or melt an ice sheet. Accumulating the solar radiation absorbed at the surface during 3.5 years would be sufficient to melt all present land ice. In fact, this would also be required to heat the ocean by about 1.7 K. So the amounts of energy involved in ice-sheet growth and decay are relatively small (see also Oerlemans & Van der Veen 1984). This is further illustrated by considering a rapid deglaciation: melting of all ice-age ice within 5000 years requires 3.8×10^{24} J per year, which is only 0.084% of the absorbed solar radiation in the present climate. So it is hard to imagine that the *direct* effect of ice-sheet growth or decay disturbs the global energy balance. The feedbacks must be much more important, in particular the albedo feedback.

From this simple argument it can also be concluded that the onset of deglaciation is not related to the global energy budget of the climate system, but only to the *local* climatological conditions at the surface of an ice sheet. When these become favourable for melting (e.g. larger summer insolation, lower surface elevation), the mass balance will become negative and the atmosphere will easily provide the energy for the melting process. For growing ice sheets the perturbations in the global energy budget are even smaller, as the process is slower.

Climate sensitivity and albedo feedback

To obtain a basic understanding of the albedo feedback, it is instructive to modify equation (1) and see what the temperature response is to a small perturbation of the radiation balance. In the atmosphere there is a tendency for the relative humidity to remain constant, implying a changing absolute humidity for changing temperature. As water vapour is one of the strong absorbers of infrared radiation in the atmosphere, the transmissivity τ in equation (1) will not be constant. A higher temperature will lead to more water vapour, and consequently to an increased greenhouse effect and a further temperature increase: the water vapour feedback. A simple way to deal (schematically) with this complication is to employ satellite data to relate surface temperature and outgoing terrestrial radiation R_{out}. In Fig. 1 values of T_s' and R_{out} for 10° wide latitude belts are plotted, and it appears that the relation can be fitted well by a straight line. As the amount of moisture in the atmosphere in this 18 point sample varies with temperature, the water vapour feedback

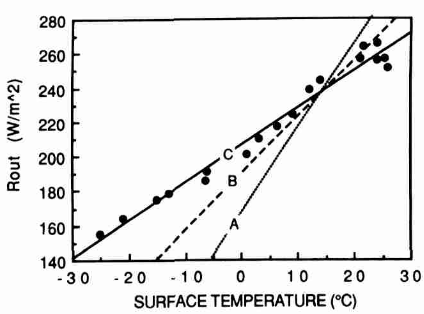

Fig. 1. A plot of outgoing infrared radiation (as measured by satellite) versus surface temperature. Each dot represents a 10°-wide latitude belt. Curve A shows the relation between surface temperature and outgoing radiation for an atmosphere without greenhouse effect; curve B the relation for an atmosphere without water-vapour feedback (both linearized around the present-day state). Curve C is a linear fit to the data. Based on Oerlemans & Van den Dool (1978).

is implicitly taken into account. So, instead of equation (1), we now use

$$(1 - A)S = a + bT_s', \qquad (2)$$

The coefficients derived from Fig. 1 are: $a = 206$ W/m^2 and $b = 2.2$ W/(m^2 K).

The sensitivity of surface temperature to a change in the absorbed amount of solar radiation is determined by the coefficient b. If this amount were to change by 10 W/m^2, a temperature change of 4.5 K would be needed to compensate for this in the outgoing infrared flux. So the smaller the b, the larger the 'climate sensitivity'. It is interesting to make a comparison with a straightforward linearization of the right-hand side of equation (1). Without a greenhouse effect ($\tau = 1$), it is found that $b = 5.42$ W/(m^2 K), and the temperature change would only be about 1.8 K. With a constant greenhouse effect, $b = 3.36$ W/(m^2 K), and the temperature change would be 3 K. This is for a linearization around the present climatic conditions, with a mean surface temperature of 288 K. Although this analysis is extremely simple, it clearly demonstrates the importance of the water vapour feedback.

We now turn to the albedo feedback. The change of mean planetary albedo for a small change in surface temperature is denoted by $\beta = \delta A/\delta T$. The linearized form of the radiation balance is then easily derived and the solution for the temperature perturbation δT reads:

$$(1 - A_o)\delta S - \beta S \delta T = b\delta T$$
$$\rightarrow \delta T = (1 - A_o)\delta S/(b + \beta S) \qquad (3)$$

In this expression A_o is the present-day mean planetary albedo of 0.3, and δS represents a change in the amount of solar radiation received by the earth. Equation (3) shows that climate sensitivity is enhanced by the albedo feedback ($\beta < 0$!). Theoretically, when the albedo feedback is so strong that $b + \beta S < 0$, a runaway situation occurs in which the entire earth would soon be covered by ice.

There are various ways to estimate the value of β. One is to compare the climate of an ice age as simulated by General Circulation Models (GCMs) of the atmosphere to the present climate, and compare the radiation budgets. Several studies of the ice-age climate have been undertaken, with different GCMs, and all based on the surface conditions as provided by CLIMAP (1976). A particular comprehensive analysis has been performed by Manabe & Broccoli (1985), who have compared the total radiation budget of the climate system for present and ice-age surface conditions with the same orbital forcing. During full glacial conditions the amount of solar radiation absorbed in the climate system is decreased by 3 W/m^2, which corresponds to a change in mean planetary albedo of slightly less than 0.009. With a decrease in mean surface temperature of about 3 K this implies that $\beta = 0.003\ K^{-1}$. A similar value was derived in Oerlemans & Van den Dool (1978) from experiments with an energy-balance climate model with a fairly detailed parameterization of the albedo. The value of βS is about $-1 W/(m^2 K)$, so the denominator in equation (3) becomes significantly smaller and climate sensitivity is enhanced.

In Table 2 some results are summarized in terms of the change in global mean surface temperature for a 1% increase in isolation. The significance of the albedo feedback is obvious, but it should be added that this is the combined effect of snow cover, sea ice and land ice.

Table 2.

No greenhouse effect	0.44 K
With greenhouse effect	0.71 K
With water vapour feedback	1.09 K
With water vapour and albedo feedback	2.03 K

Mass balance of the present ice sheets

As noted in the Introduction, the surface conditions at the ice/snow surface ultimately determine what will happen to an ice sheet. The surface mass balance, frequently expressed in m water equivalent per year, forms the link between ice sheet and meteorological conditions and thus deserves careful study. A natural start is to look at the two large ice sheets we can observe today. In Fig. 2, two mass balance 'profiles' for the ice sheets of Greenland and Antarctica are shown.

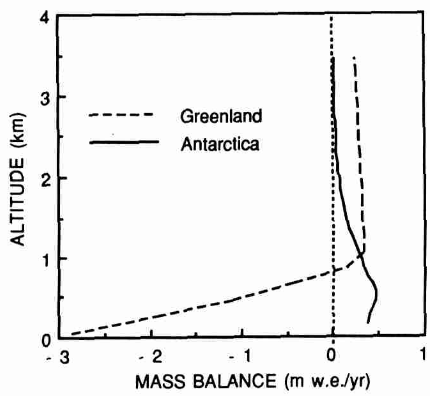

Fig. 2. Mass balance as a function of elevation for the Antarctic and Greenland ice sheets. The curves represent mean values for the entire ice sheets, local conditions may be quite different. Based on Drewry (pers. comm. 1988) and Oerlemans et al. (1990).

The mass balance of the Antarctic ice sheet is positive everywhere, at least when average values over elevation intervals are considered (there are places where the mass balance is locally negative because of very dry conditions and/or snowdrift). The largest values are found in the coastal regions, at elevations between 0 and 1000 m. The accumulation in the interior is very low: the polar desert. This is due to very low absolute humidity associated with the extremely low air temperature. The total annual accumulation on the ice sheet is estimated to be 1800 km^3. Virtually all loss of ice is by iceberg calving, but it is not known whether the ice sheet is close to equilibrium.

A totally different picture applies to the Greenland ice sheet. There is a large accumulation zone with a positive balance and a smaller but important ablation zone. The ablation rates on the lowest parts are very high, and in fact comparable to what is observed on the tongues of mountain glaciers in, for instance, southern Norway and the Alps. The low values of surface albedo due to accumulated atmospheric dust are the common factor here. A recent estimate of the total accumulation on the ice sheet is 535 km^3 (Ohmura & Reeh 1990). It is generally assumed that half of the loss is by iceberg calving, the other half by melting and runoff. As for the Antarctic ice sheet, it is not known to which degree gain and loss match.

The most important quantity governing the mass balance at a particular location is the annual mean surface air temperature T_{as}. The dependence of accumulation and ablation on T_{as} is shown in Fig. 3. The graph is schematic and it is supposed that cloudiness, seasonal temperature cycle, etc., are constant under a change of T_{as}.

Depending on local conditions, snowfall reaches a maximum for annual temperatures somewhere between -5 and $+5$°C. Ablation becomes significant when summer temperature exceeds the melting point, so the curve shown here applies to a location with an annual

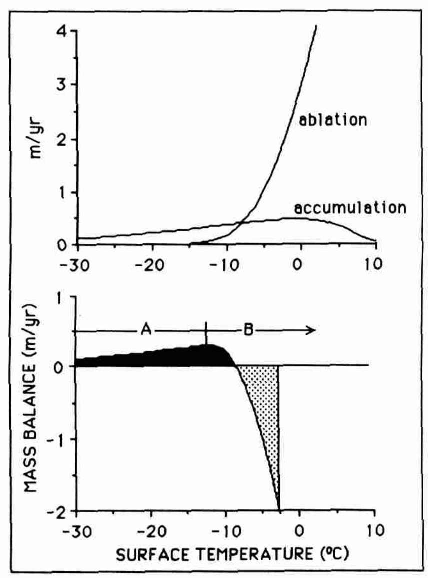

Fig. 3. Components of glacier mass balance at a fixed location in dependence of the annual mean surface temperature, with a given constant annual temperature range. The net balance (lower panel) shows a region where it *increases* with temperature (A) as well as a region where it *decreases* with temperature (B). The Antarctic ice sheet is in region A, the Greenland ice sheet mainly, but not entirely, in region B.

temperature range of 30 K. This is a large range, but not unusual in high latitudes. The resulting mass balance curve shows a peculiar behaviour: for $T_{as} < -10°C$ the mass balance increases with temperature (range A in the figure), for $T_{as} < -12°C$ it decreases with temperature (range B). The value of $-12°C$ should not be taken as a universal number, of course. Depending on local conditions it may vary in a range running from -20 to $-5°C$. *The important point to make here is that no monotonic, let alone linear, relation between glacier mass balance and annual air temperature exists.*

It is interesting to consider the Antarctic and Greenland ice sheets in the light of Fig. 3, as this may give a first indication about changes in ice-mass accumulation in a warmer world. Annual surface temperature on the Antarctic continent ranges from roughly $-10°C$ in the coastal areas to $-55°C$ on the interior plateau. So virtually all of the Antarctic ice sheet is in range A, and the mean surface balance would certainly increase when climate would get warmer. On the other hand, a significant part of the Greenland ice sheet is in range B. In fact, in several studies it has been shown that the mean surface balance of this ice sheet would become significantly smaller in a warmer climate (Ambach & Kuhn 1989; Oerlemans et al. 1990). Here the increase in snow accumulation is not large enough to compensate for the stronger melting.

Mass-balance conditions on the ice-age ice sheets of the northern hemisphere must have been similar to those of the Greenland ice sheet today, with significant mass turnover and large melting rates in the margin, and lower accumulation rates towards the north. In most ice-age climate simulations with GCMs the mass balance of the ice sheets has not been studied explicitly. However, Manabe & Broccoli (1985) give the geographical distribution of the 'net ice accretion rate' and this confirms the basic similarity with Greenland conditions: accumulation of the order of 0.5 m w.e./yr, ablation up to several m w.e./yr in the lowest parts of the southern margin.

A further remark concerns the fact that the accumulation rate is bounded while ablation can become very large for sufficiently high temperatures. This implies that ice sheets can decay much faster than they grow, which is in accordance with the palaeoclimatic evidence. As will be discussed later, other mechanisms exist that make this asymmetry even more pronounced.

Feedback between elevation and mass balance

As surface temperature is an important factor with regard to the mass balance, the feedback between increasing surface elevation and ice accumulation can be powerful. The full implications of this for the Pleistocene glacial cycles were first realized by Weertman (1961), who gave an elegant mathematical analysis of the problem.

The feedback is illustrated in Fig. 4 for typical northern hemisphere conditions. The continent is bounded in the north by the Arctic ocean, and climatic conditions are represented by an equilibrium line that slopes upward to the south. The intersection of the equilibrium line with sea level is termed the climate point P. Climatic change can now be considered as shifting the equilibrium line up and down, or moving the point P back and forth. In the event of climatic cooling, P shifts southward, and as soon as P is on the continent an ice sheet will form. Whether it will grow or not depends on the location of the equilibrium point E, the intersection of equilibrium line and ice-sheet surface. The surface elevation of a growing ice sheet increases, which automatically shifts E southwards, even without further cooling. In fact, we can be sure that the growth of the large ice-age ice sheets was only possible because of this powerful positive feedback mechanism.

In a (very) first approximation, the profile of a continental ice sheet only depends on its size L, and its mean elevation is given by $0.67 \times \sqrt{(\Lambda L)}$, see e.g. Oerlemans & Van der Veen (1984). Here Λ is a parameter determined by the stiffness (yield stress, actually) of the ice. The equilibrium size of an ice sheet can then be obtained from the condition that the total net balance is zero. For a sloping equilibrium line as shown in Fig. 4, this can be worked out to find L for any prescribed position of the equilibrium line, i.e. for any value of P. Fig. 5 shows a typical solution diagram. The heavy line represents stable steady states, the dashed line unstable steady states. Two critical

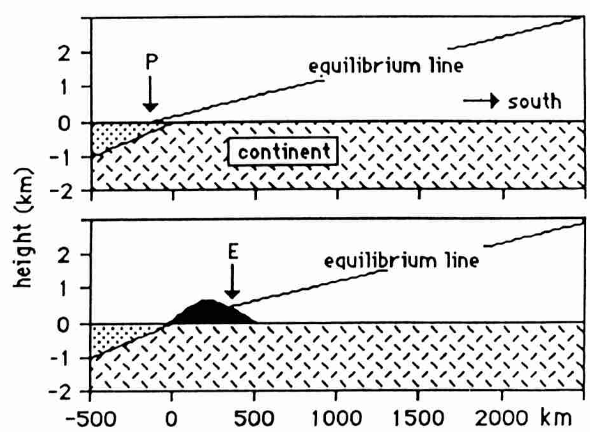

Fig. 4. An illustration of the feedback of surface elevation on mass balance. P is the 'climate point' (intersection of equilibrium line with sea level) and E the equilibrium point (intersection of equilibrium line with the ice-sheet surface). E may shift southward due to ice-sheet growth, even if the climate does not change (P stays in position).

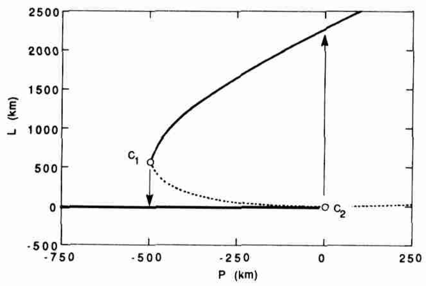

Fig. 5. Equilibrium size of a northern hemisphere ice sheet, in dependence of climatic conditions (position of P; $P > 0$ implies 'climate point' on the continent). Note that hysteresis occurs. The open circles indicate critical points and the dashed part of the curve represents unstable equilibria. From Oerlemans & Van der Veen (1984), original analysis by Weertman (1961).

points appear, reflecting the occurrence of hysteresis. There is a range of values of the climate point P for which two stable states are possible: no ice sheet or a large one, which is maintained by the elevation–mass balance feedback. When P is located south of the critical point C_2, there must be an ice sheet, when P is located north of the critical point C_1, a stable ice sheet cannot exist.

A few important conclusions can be drawn from Fig. 5. For convenience, we denote the characteristic response time of the northern hemisphere ice sheets by T_{ICE}, the characteristic time-scale of a forcing mechanism of the climate system by T_{CLIM}. If $T_{ICE} \gg T_{CLIM}$, then the ice sheet will hardly react to the forcing. Its equilibrium size will correspond to the mean forcing. In the case of $T_{ICE} \ll T_{CLIM}$ the ice sheet will always be in equilibrium, i.e. on the steady state curve of Fig. 5, thereby following the arrows. The most complicated case occurs when $T_{ICE} \approx T_{CLIM}$, because of transient effects. For the ice sheet to jump from one stable branch of the solution diagram to another it is necessary that P changes, *and* that this change lasts long enough.

When considering climatic forcing by changes in the earth's orbit (the Milankovitch theory, which is currently the most popular theory of the Pleistocene glacial cycles – for a review, see Berger 1988), we are in the situation that $T_{ICE} \approx T_{CLIM}$. The precession of the equinoxes has an apparent period of 22 ka (1 ka = 1000 years). A corresponding time-scale for the forcing then is 5 to 10 ka, which is somewhat less than the time ice sheets need in order to grow to a size where they start to affect the global climate. For insolation variations associated with changes in the obliquity (major period 44 ka), T_{ICE} and T_{CLIM} are quite comparable. The effect of changes in the eccentricity of the earth's orbit are on a somewhat larger time-scale (periods of ≈100 ka and ≈400 ka).

In view of these facts, *the response of the climate system to Milankovitch insolation variations cannot be linear, at least not when the build-up and decay of large ice sheets are involved.*

Weertman (1976) was the first to take the consequence of this observation. He forced a simple time-dependent model for a single northern hemisphere ice sheet with the Milankovitch insolation variations, by assuming that the equilibrium line moves up and down in proportion to changes in summer insolation at high latitudes. His attempt was partly successful. The model was able to reproduce realistic variations in ice-sheet size, but did not generate as much power at the longer time-scales (100 ka, in particular) as is normally seen in the deep-sea oxygen-isotope records. The slope of the equilibrium line and the accumulation rate are the most important parameters concerning the behaviour of a model ice sheet. The smaller the slope, the larger the ice sheet, whereas the time-scale for growth is mainly determined by the accumulation rate.

In several studies with ice-sheet models (Oerlemans 1980; Pollard 1982) it has been shown that a full-grown northern hemisphere ice sheet, generated with realistic model parameters, will not disappear because of high summer insolation alone. Rapid ice-sheet decay must have involved one or more destabilizing mechanisms.

Mechanisms that can destabilize ice sheets

The discussion in this section will be qualitative. Although several mechanisms described here have been studied by modelling, a discussion of this would be too technical in the present context. Also, many of those modelling studies are really sensitivity studies, of which the output still depends in a critical way on the boundary conditions and input parameters. They have shown, however, that the mechanisms depicted in Fig. 6 are potentially important. This figure summarizes, in a sketchy way, the possible history of a glacial cycle, without referring to a particular period or time-scale. The various factors are discussed below.

It is easy to imagine that isostatic sinking of the bed on which an ice sheet grows must affect the mass balance. The associated lowering of the surface implies higher air temperature and, when averaged over the ice sheet, smaller mass balance (most of a growing northern hemisphere ice sheet will be in region B of Fig. 3). The outflow of material in the asthenosphere takes time, so the isostatic adjustment lags the increasing load. Without detailed calculations we may conclude that bedrock sinking will slow down the growth of an ice sheet, or even stop it when the climatic cooling that initiated the ice sheet is not large enough or is of short duration only.

The same lagged response of the bedrock accelerates ice-sheet decay once it has been started. First, during the process of retreat of the ice edge, the bed will be much lower than the corresponding isostatic equilibrium state. The ice retreats in a region where ice thickness was large 'shortly' before, and the bed may still be below sea level. As the bed will slope downwards towards the ice sheet, large proglacial lakes and seas can form. Calving ice fronts, in combination with high sliding velocities of the ice lobes, can then lead to very fast retreat in certain places.

Time-dependent models of the

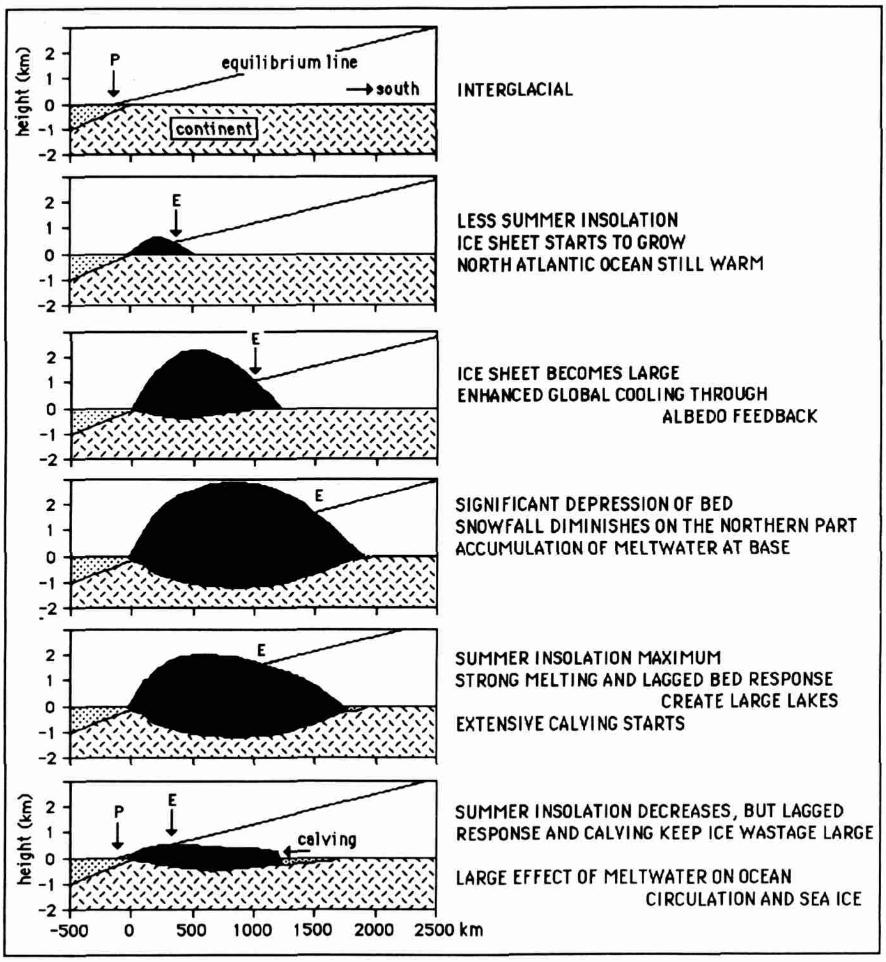

Fig. 6. An overview of processes affecting the evolution of ice sheets on the northern hemisphere continents, depicted in a north–south cross-section.

response of the solid earth to ice load range from extremely simple local isostatic adjustment (Oerlemans 1980), thin-channel models (Pollard 1982; Oerlemans 1982) to spherical visco-elastic representation of the mantle (Birchfield & Grumbine 1985; Hyde & Peltier 1985). The response of combined ice-sheet/geodynamic models to Milankovitch forcing is discussed in these papers, and significant differences have been found. Still, the results demonstrate that the interaction of geodynamics and ice-sheet growth/decay is capable of generating glacial cycles with a time-scale significantly larger than that of the forcing.

The thermal regime of an ice sheet can also contribute to destabilization. The temperature distribution is determined by advection (vertical and horizontal), internal heat generation by shearing, and geothermal heat input through the base of the ice sheet. The higher ice temperatures are always found at the base. During the process of growth, the basal ice layers get warmer and at some stage reach the (pressure) melting point. Subglacial meltwater is then produced. When the bed consists of rock, the meltwater may be discharged in sheet or channel flow (Weertman 1972), unless the entire warm zone is surrounded by ice, frozen to the bed. In that case subglacial lakes may occur (they have been found in Antarctica). The subglacial water may also saturate a till layer to generate a deformable bed, which may favour rapid ice discharge and lead to thinner ice sheets (e.g. Boulton & Jones 1979; Fisher et al. 1985). Recent investigations of the ice streams of West Antarctica have made clear that such till layers are present there, and seem to be the reason for the high ice velocities (e.g. Alley et al. 1987). To what extent similar situations occurred on the Laurentide and Fennoscandian ice sheets is not clear. 'Downdraw' of ice from the major ice domes through deep outlet glaciers has also been suggested as a significant destabilizing factor (Denton & Hughes 1981). In any case, there is theoretical and observational evidence for a number of processes that may accelerate the decay of an ice sheet when the base is sufficiently warm, having in common that they all effectively reduce the friction of ice flow at the bed.

When a glacial period approaches its maximum, sea ice cover in the Atlantic ocean extends much further southward, and the moisture supply to the more northerly parts of the Fennoscandian and Laurentide ice sheets will be restricted. Oerlemans & Vernekar (1981) conducted many experiments with a zonal climate model, including atmospheric dynamics and a full hydrological cycle, to see how the precipitation regime would change due to extensive glaciation. It was found that precipitation decreases by up to 50% north of about 60°N, and increases by up to 30% in the 40 to 60°N latitude belt, particularly in winter. The reason for this is the enhanced storm activity along the southern ice and sea-ice margin. A similar conclusion was reached by Manabe & Broccoli (1985), in an experiment with a general circulation model of the atmosphere, run for ice-age boundary conditions. The net effect appears to be a gradual drying out of northern hemisphere ice sheets when conditions get colder, and in particular when sea-ice cover becomes extensive.

The interaction between ice-sheet evolution and ocean circulation is complicated, and may also involve components that destabilize ice sheets. For example, when deglaciation has been initiated, a low-salinity meltwater layer may cover part of the North Atlantic Ocean and lead to more extensive sea–ice cover (for a more extensive discussion, see for example Ruddiman & McIntyre 1981). A likely consequence is the reduction in snowfall on the ice sheets, thereby contributing to the deglaciation process. On the other hand, the more stable stratification associated with input of meltwater affects the deep ocean circulation. There is evidence from proxy palaeoclimatic data and from modelling studies that the production of North Atlantic Deep Water (NDAW) is extremely sensitive to small changes in fresh-water input (Broecker et al. 1985; Manabe & Stouffer 1988; Maier-Reimer & Mikolajewicz 1989). In fact, it has been suggested that the Younger Dryas climatic event was forced by a substantial reduction in NADW production and the consequent decrease in poleward heat transport by the meridional circulation in the Atlantic Ocean.

Apart from processes directly linked to the shape and state of the northern hemisphere ice sheets, other feedback loops in the climate system had a major influence on the evolution of the Pleistocene climate. The atmospheric concentration of a number of greenhouse gases, in particular, appears to vary significantly between glacial and interglacial conditions (e.g. Barnola et al. 1987). Attempts to delineate the effects of albedo feedback, atmospheric CO_2 concentration and orbital insolation variations suggest that, in terms of radiative forcing of the climate system, these are all of the same order of magnitude (Broccoli & Manabe, 1987; Lorius, pers. comm. 1989).

Concluding remarks

There is general consensus that the processes outlined above are of direct importance to the growth and decay of northern hemisphere ice sheets, but the details remain unclear. More work with coupled models, in which atmospherics, ocean, land ice and solid earth are treated in a time-dependent (but not necessarily synchronous) way, is needed for a further diagnosis of 'the glacial cycle'. Such models should also comprise a component describing the geochemical cycles that have a direct effect on the energy balance of the climate system. It will be crucial to test models against field data. This must involve: (i) The use of large modern climatological data sets to test in detail the mathematical representations of physical processes included in the models. (ii) Further construction of palaeoclimatic data sets, mainly in the form of snapshots of the Pleistocene climate, for verification of climatic states generated by coupled models. This will imply large interdisciplinary efforts along the lines set by the COHMAP Project (1988).

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