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A MODEL FOR THE SURFACE BALANCE OF ICE MASSES:
part II.
APPLICATION TO THE GREENLAND ICE SHEET

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With 7 figures

ABSTRACT

The mass budget of the Greenland ice sheet is studied with an energy balance model of the ice/snow surface. The effect of refreezing is taken into account in a schematic way. Four regions are considered, with different climatic characteristics, and the model is applied at 15 points with 200 m spacing in elevation.

With air temperature, total precipitation and cloudiness as input, the following total mass budget is obtained (in water equivalent): 532 km^3/yr of accumulation, 398 km^3/yr of ablation, and 134 km^3/yr of iceberg calving. This result is extremely sensitive to the climatological input data: taking a 1°C lower value of the mean air temperature would lead to an ablation of 252 km^3/yr and a calving rate of 283 km^3/yr (in case of equilibrium). The mean equilibrium-line altitude is found to be 905 m.

The sensitivity to climatic change varies significantly from region to region — it is largest in the warmest parts of the ice sheet. The mean change in the specific surface balance of the entire ice sheet found for a 1°C warming is -0.076 m w. e./yr. This corresponds to a sea-level rise of 0.36 mm/yr. This value does not depend very much on how the present state is diagnosed. Assuming that in a warmer model climate precipitation will increase in proportion to the amount of precipitable water in the atmosphere, the model effect of a 1°C temperature rise is reduced by about 25 % due to increased snow accumulation and the consequent albedo feedback.

EIN MODELL FÜR DIE OBERFLÄCHENBILANZ VON EISMASSEN, TEIL II: ANWENDUNG AUF DAS GRÖNLÄNDISCHE INLANDEIS

ZUSAMMENFASSUNG

Die Massenbilanz des grönländischen Eisschilds wird mit einem Energiebilanzmodell der Schnee- und Eisoberflächen untersucht, wobei das Wiedergefrieren von Schmelzwasser schematisch mitbehandelt wird. Das Modell wird auf vier Gebiete mit verschiedenen Klimabedingungen für jeweils 15 Punkte in 200 m Höhenabstand angewendet.

Mit den Eingangsgrößen Lufttemperatur, Jahresniederschlag und Bewölkung werden folgende Massenumsätze für ganz Grönland berechnet: 616 km^3 Wasseräquivalent Akkumulation pro Jahr, 388 km^3 Ablation pro Jahr und 228 km^3 Kalbung von Eisbergen. Dieses Ergebnis reagiert sehr empfindlich auf die klimatologischen Eingangsgrößen: wenn man eine um 1 K niedrigere Jahresmitteltemperatur einsetzt, werden für Gleichgewichtsbedingungen nur noch 220 km^3 Ablation und 399 km^3 Kalbung pro Jahr berechnet.

Die Empfindlichkeit auf Klimaänderungen ist in den vier Gebieten deutlich verschieden und ist in den wärmsten Gebieten am größten. Die mittlere Änderung der spezifischen Massenbilanz des ganzen Eisschilds nach einer Erwärmung um 1 K beträgt −0,076 m Wasseräquivalent pro Jahr, was einem Anstieg des Meeresspiegels um 0,36 mm pro Jahr entspricht. Diese Empfindlichkeit hängt nicht sehr von der Diagnose des jetzigen Klimazustands ab. Die Annahme, daß in einem wärmeren Klima der Niederschlag proportional zur Menge des niederschlagbaren Wassers in der Atmosphäre steigt, führt im Modell zu einer Verringerung der Empfindlichkeit um 25 %, weil Schneeakkumulation und Albedo steigen.

1. INTRODUCTION

Investigation of the total mass budget of the Greenland ice sheet is seriously hampered by the limited amount of measurements. Accumulation has been measured on some traverses, see Radok et al. for an overview (1982), which unfortunately do not give good coverage of the entire ice sheet. More detailed work, involving both the ablation and accumulation area, has been done during the EGIG expeditions, see for instance Ambach (1963, 1979). This work has given a basic insigth into the nature of the surface processes in the ablation and refreezing zone, and has also stirred up many questions. A large number of glacio-hydrological studies was then carried out in the ablation zone of West and Southwest Greenland by the Geological Survey of Greenland (Copenhagen), e. g. Braithwaite and Olesen (1984, 1989). Running a mass balance programme is a costly matter, because of the logistic problems involved (Olesen and Braithwaite, 1989). It is thus not surprising that mass-balance conditions in the inaccesible regions of the Greenland ice sheet are only very crudely known, and that, consequently, the total balance as estimated by various authors shows large differences (e. g. Robin, 1986).

Application of an energy balance model is not going to solve the problem, of course. It forms an alternative approach, worthwhile to try because the calculation of the total mass budget is then based on an entirely different data set: climatological data from a number of meteorological stations instead of direct mass balance observations. Also, it is interesting to find out how the sensitivity of the mass budget to climatic change, as obtained from a process-oriented approach, compares to more empirically derived estimates (e. g. Ambach and Kuhn, 1989).

The energy balance model described in part I is applied here to four sets of climatological input data, thought to represent average conditions for four regions of the Greenland ice sheet, see fig. 1 and table 1. One may criticize this approach for yielding a too small resolution of the varying conditions on the ice sheet, but in our view it is a useful step. A distinction is made between the dry north and northeasterly part (region 1), the moist southwesterly part including the southern tip (region 2), the drier and relatively warm part of the west coast between approximately 65 and 69° N (region 3) and the moister and colder area bordering the Baffin Bay (region 4). The relative area covered by these regions is listed in table 2. To obtain budgets for the entire ice sheet, the area-elevation distribution is needed for each region. These have been calculated from a map published by the Geological Survey of Greenland (Weidick, 1971). The elevation intervals considered in this study are 0—200 m, 200—400 m, 400—600 m, etc. The energy-balance model is then applied at 100 m, 300 m, 500 m, etc.

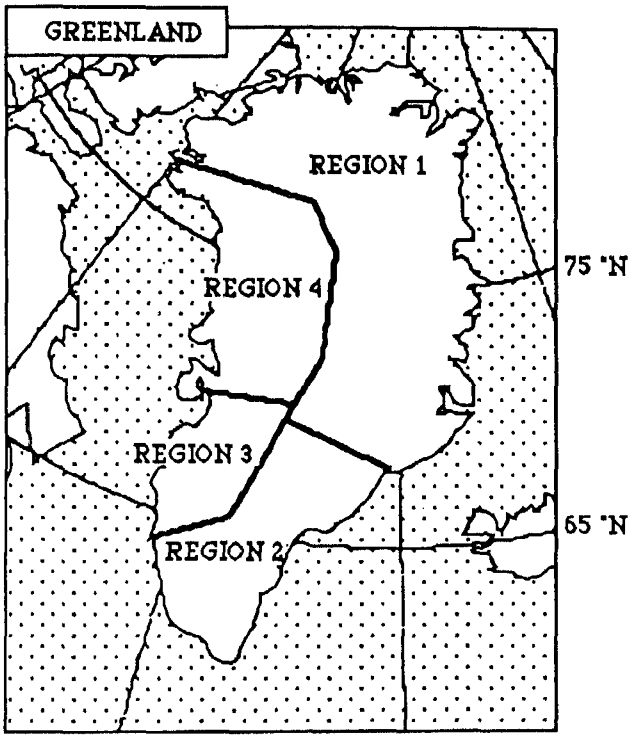

Fig. 1: Map of Greenland, showing the four regions considered in this study

Table 1: Climatological input data used for the energy-balance calculations

region	altitude (°N)	Θ_{ann} (°C)	Θ_{range}	Γ_{jul} (K/100 m)	Γ_{dec}	clouds	precipitation (m/yr; h in m)
1 (40 %)	76	−10	28	0.4	0.9	0.3	$0.25\exp(-h/8000)$
2 (25 %)	65	− 1	18	0.4	0.9	0.7	$0.86\exp(-h/3000)$
3 (15 %)	67	− 6	28	0.65	0.9	0.4	$0.16+0.00007\,h$
4 (20 %)	75	− 9	34	0.6	0.9	0.5	$0.21+0.00021\,h$ $-5.1\text{x}10^{-8}h^2$

Concerning surface elevation, there is a large uncertainty in many regions of the ice cap. The treatment of the outermost margin is particularly difficult, and much 'interpolation' was done to derive the distributions. For a region in central West Greenland, a comparison was made between an area-elevation distribution calculated from the 20 km gridding of Radok et al. (1982), and the planimetry applied here. There was a reasonable agreement, except in the lower few hundred meters. Here the gridded data cannot resolve the small outlet glaciers. This is serious, as the ablation rates on these glaciers pushing into the tundra are very high (up to 5 m w. e./yr) and significantly affect the total mass budget.

The area-elevation distributions for the four regions ultimately used in the present study are shown in fig. 2. The differences are not very large, except that in regions 1 and 4 significantly more ice is found with surface elevation below 700 m.

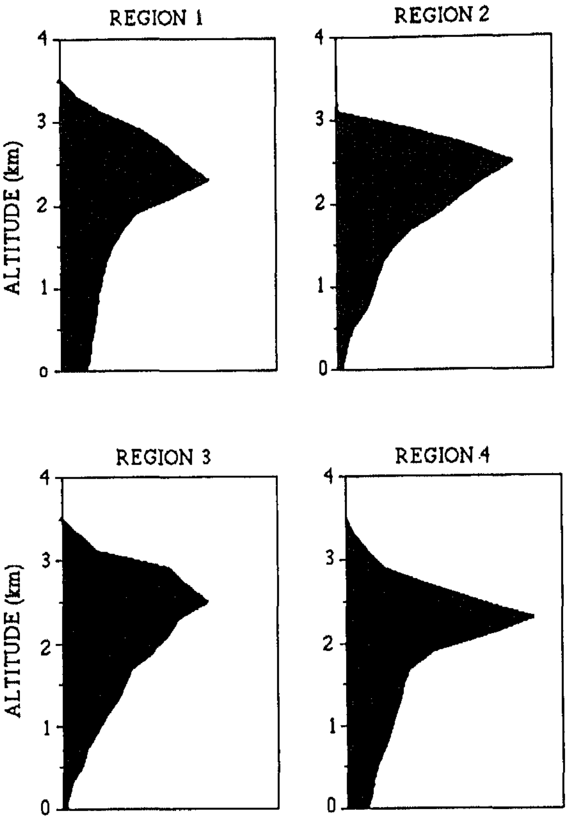

Fig. 2: The area-elevation distributions for the four regions

2. MODIFICATION OF THE ENERGY/MASS BALANCE MODEL

To become applicable to the Greenland ice sheet, the model described in Part I needs to be adjusted, because refreezing of melt water may be an important term in the energy balance. Ambach (1963) has shown that on large parts of the Greenland ice sheet heating of the upper ice/firn/snow layers by refreezing and conduction consumes a significant amount of energy in late spring/early summer. From his measurements and calculations, one can estimate that, in a summer season, a layer equivalent to 2 m of solid ice is heated up from the annual mean temperature to the melting point. This order-of-magnitude estimate is used here to model the implication for the energy balance of the refreezing process. No attempt is made to calculate a detailed temperature/ density profile.

Here the following formulation was chosen for the difficult treatment of runoff. Denoting the surface energy balance by B, the amount of energy available for melting and runoff (R) and the amount used for heating up the upper ice/snow layers (H_{ice}) are written (for $B > 0$):

$$\begin{aligned} R &= B \exp(\Theta_{ice}) \\ H_{ice} &= B - R = B\{1 - \exp(\Theta_{ice})\} \end{aligned} \tag{1}$$

Here Θ_{ice} is the mean temperature (in °C) of the equivalent 2 m ice layer mentioned above. In the model, it is set equal to the annual mean air temperature at the beginning of the ablation season. So, for instance, when the temperature of the upper ice/firn layer is −4 °C, only 2 % of the melt water runs off; for a temperature of −1°C, this is 37 %. Some tests with other formulations for the partitioning of the energy between R and H_{ice} were also carried out. With regard to the annual mass balance, it appeared that the total amount of energy needed to heat up the cold ice layer is the most important parameter. This can be easily understood: when a smaller runoff rate would be chosen for a specific ice/firn temperature, heating up of the ice/firn proceeds faster and the reduced runoff will be compensated later, as the upper layers reach temperate conditions earlier. An example of how the formulation of refreezing affects the runoff is discussed in a later section.

The parameterization of the albedo a also needs some adjustments. The dependence of the background albedo on elevation as used for the Alps (see eq. (11), Part I), is too strong for the major part of the Greenland ice sheet. The type of formulation was not changed, but the constants were modified to give elevation h and equilibrium-line altitude E in m, *snow depth* d *and accumulated melt* M_m *in m* w. e.:

$$\alpha = \max\,[0.12\ ;\ \alpha_{sn} - (\alpha_{sn} - \alpha_b)\, e^{-5d} - 0.015\ M_m] \tag{2}$$

$$\alpha_b = 0.089\ \mathrm{arctg}\{(h - E + 400)\ /\ 500)a\} + 0.48 \tag{3}$$

Compared to the values in Part I the mean albedo for snow (α_{sn}) was changed from 0.72 to 0.75.

Whether this type of albedo parameterization is really satisfactory is difficult to judge, as albedo measurements on the ice sheet have been carried out at a few places only. Hopefully, satellite images may help to test and improve this type of scheme in the future.

3. CLIMATOLOGICAL INPUT DATA

Most of the input data are derived from compilations of A. Ohmura (ETH Zürich). In Ohmura (1987) an extensive discussion is given on Greenland temperature. Annual mean temperature and annual temperature range used here are based on the maps presented in this work. Lapse rates vary enormously through the year, so a seasonal cycle is included in the present model. The values were *not* taken from Ohmura's table 2, as these are mean values for the 1—3 km altitude range. Vertical cross sections (Ohmura's fig. 2) show that, in summer, the lapse rate in the lower 1 or 2 km, is relatively small [this is typical for high latitude summer conditions, see also Ohmura and Müller, 1977)]. It is important to take this into account in the calculation of the enery balance.

Table 2: Model output of the mass budget obtained with the climatological data of table 1. The column 'total' gives values for the entire ice sheet, i. e. weighted according to area as listed in the first row

	region 1	region 2	region 3	region 4	total
part of total area (%)	40	25	15	20	100
accumulation (m/yr)	.194	.443	.304	.397	.313
specific balance (m/yr)	.036	.240	−.091	.096	.079
ablation (m/yr)	.158	.203	.395	.301	.234
E (m)	718	925	1349	920	905

The input values used for the various regions are listed in table 1. Interpreted as mean values over the regions, we estimate the error in the annual temperature to be of the order of 2° C, in the annual temperature range of the order of 3° C.

Annual precipitation is prescribed by simple formulas expressing dependence on altitude, to match the precipitation map compiled by Ohmura (Ohmura and Reeh, 1991). 'Sampling' of precipitation rates along isohypses on the map resulted in the expressions listed in table 1. In region 1 there is a slight decrease with altitude, in region 2 a more pronounced decrease with altitude, in region 3 an increase with altitude, and in region 4 precipitation reaches a maximum at an elevation of about 2000 m. The profiles are plotted in fig. 3.

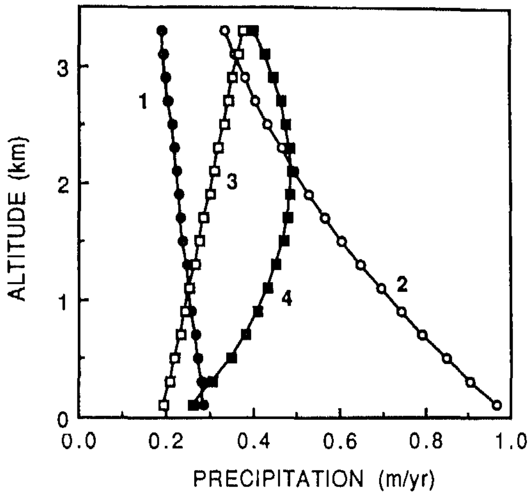

Fig. 3: Prescribed variation of annual pecipitation with elevation, as derived from a map provided by A. Ohmura. The labels refer to the four regions shown in fig. 1

4. THE PRESENT STATE

The calculation of the mass balance is performed in the same way as described for alpine glaciers in Part I. Integrations are extended over three years of simulated time, to let the background albedo profile (α_b) adjust to the equilibrium-line altitude. Convergence to an equilibrium solution is even faster than for alpine conditions, because initial ice temperature is set to the prescribed annual mean air temperature (there is much to say about such an approach, but this falls outside the scope of the present study).

The effect of refreezing is considerable, as is illustrated for region 1 in fig. 4. With refreezing the mean surface balance of region 1 is 0.036 m w. e./yr, without only −.035 ! This difference is equivalent to the effect of a 1.4°C increase in annual mean temperature, as will be discussed later. In fig. 4b the cumulative balance is shown for the point at 700 m a. s. l., that is, close to the equilibrium line. The daily energy balance becomes positive during the last days of May. However, it is not until the 10th of July that runoff starts to exceed the rate of accumulation. It is hard to say whether these effects are realistic. Current and future field work in the refreezing zone of the ice cap may soon give the answer, and it could turn out that the refreezing/runoff problem cannot be treated in the present schematic way.

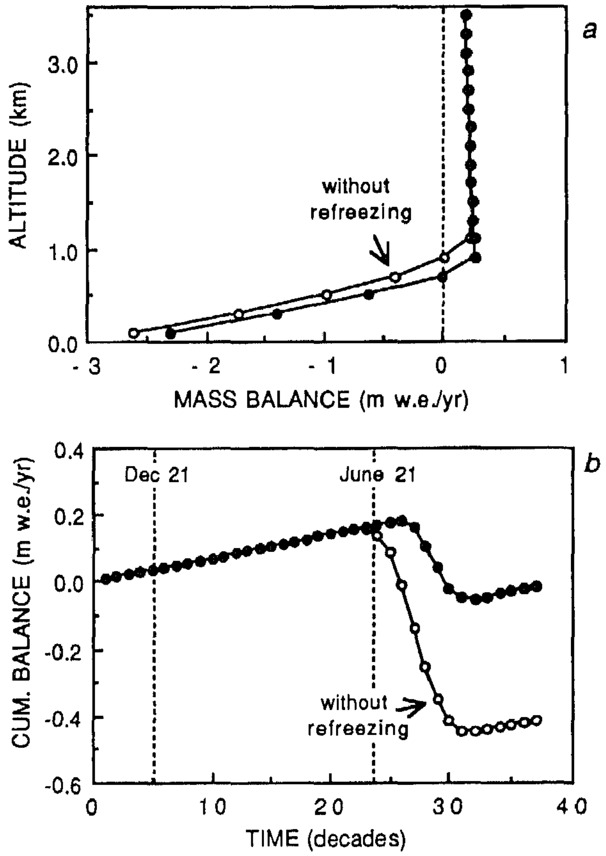

Fig. 4: The effect of refreezing melt water on the mass balance calculations. In (a) the profiles are shown, in (b) the cumulative balance (sampled at 10-day intervals) is plotted for the equilibrium line at 700 m above sea level in region 1

The calculated mass-balance profiles for the four regions are plotted together in fig. 5. The differences appear to be large. For region 3 the model predicts an equilibrium-line altitude which is much higher than the average. Interestingly, this is the region where glaciological studies have been concentrated during the past few decades! The mass balance profiles of regions 2 and 4 are surprisingly similar. Here, the higher temperature for region 2 is compensated by a higher precipitation rate. The balance gradients in the ablation zones are not too different, which is of course to a large extent due to the use of the same background albedo profile. The mean value of the gradient, about 0.4 m/yr per 100 m elevation, is in good agreement with the measurements as summarized by Reeh (1985).

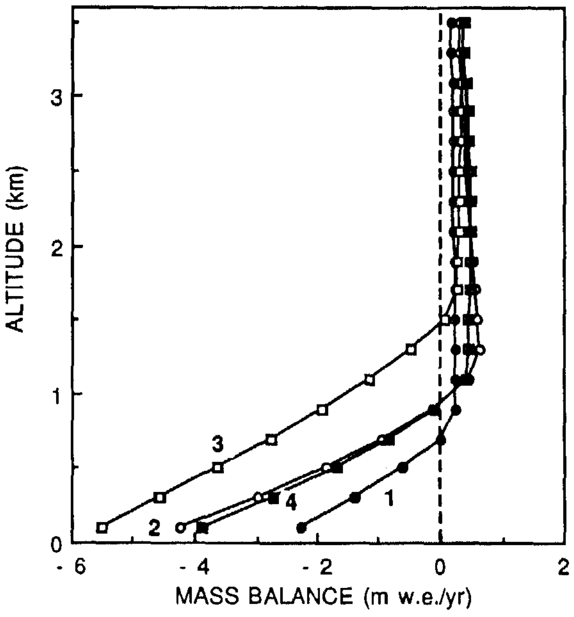

Fig. 5: Profiles of the specific balance as calculated for the four regions

The budgets are listed in table 2. The total accumulation on the ice sheet is found to be 0.313 m/yr, the net specific balance 0.079 m/yr. The mean ablation thus equals 0.234 m/yr. Multiplying the specific balance with the area of the ice sheet (1.73×10^6 km^2, Haeberli et al, 1989) yields about 137 km^3 water equivalent of ice to be removed by calving in case of equilibrium. The individual regions used in the present analysis are not drainage basins from the viewpoint of ice dynamics, and thus cannot be expected to be in balance. Region 3, in particular, has a negative specific balance and imports large amounts of ice from the other regions.

The model output for the equilibrium-line altitude E is also listed in table 2. The weighted mean value is 905 m. This is several hundreds of meters lower than generally assumed (e. g. Ohmura et al. 1986; Ohmura, 1987; Weidick, 1984), and such a difference requires an explanation.

First of all, mass balance measurements spanning the entire range from lower ablation zone to upper accumulation zone have hardly been made. In fact, only along the

EGIG profile in region 3 have such measurements been done, although still on a limited scale. It is also here, however, that the calculated equilibrium-line altitude is close to the observed one. The estimate of the equilibrium-line altitude of Ohmura et al. (1986) is based on a regression analysis where much of the input comes from different regions on the earth. Weidick's map is based mainly on observations on small isolated glaciers and ice caps in the tundra and foreland, which will not be representative for the main ice sheet as air temperature in summer is probably higher. In view of this, there are no sufficient arguments to reject the values found in the present model calculations. When pushing some model parameters to their extreme values within the range of uncertainty they have, it would be possible to obtain a much higher equilibrium line, but this would lead to a very large total amount of ablation.

Table 3 presents a comparison of the calculated budget for the entire ice sheet with earlier studies (as listed by Robin, 1986). According to the present model, total ablation is larger than found in all other studies [more than twice as large as in the study by Reeh (1985)], and, consequently, calving much smaller. Although one should probably consider the ablation of 404 km^3 as too large, it does not imply that the energy balance model is largely in error: sensitivity test showed that the result is very sensitive to the input data. A uniform decrease of the annual mean temperature by only 1°C makes the balance comparable to other studies (last line in table 3). So it is simple to match any 'observations' on the mass budget by adjusting the input parameters within their range of uncertainty. This implies that the present state of balance cannot be derived from existing climatological data. Fortunately, the sensitivity of the mass balance to climatic change, defined by, say, $\delta M/\delta \Theta_{ann}$, will turn out to be rather independent of how the present state is diagnosed.

Table 3: A comparison of the calculated mass budget of the Greenland ice sheet with earlier studies

source	accumulation	ablation	calving	balance
Bader (1961)	+630	−120 to −270	−240	+270 to +120
Benson (1962)	+500	−272	−215	+13
Bauer (1968)	+500	−330	−280	−110
Weidick (1984)	+500	−295	−205	0 (assumed)
Reeh (1985)	+487	−169	−318	0 (assumed)
Ohmura & Reeh (1990)	+535			
this study	+541	−404	−137	0 (assumed)
this study (−1K)	+548	−292	−256	0 (assumed)

5. CLIMATE SENSITIVITY

Many experiments were conducted to study the dependence of the mass budget to climatic parameters. Small temperature changes were considered first. In broad sense, the change of the specific balance in the ablation zone is proportional to the temperature change and to the ablation rate in the reference state. It is thus not surprising that the sensitivity of the regions exhibits large differences. For a change in Θ_{ann} of +1 K, the specific balances change by −.050, −0.090, −0.130 and −0.076 m/yr for the regions 1, 2, 3 and 4, respectively. For the entire ice sheet a value of −0.077 m/yr is then found. The associated total amount of water is 133 km^3, which is rather close to the value of about 120 km^3 found by Ambach and Kuhn (1989) in a new analysis of the

EGIG data. The dependence of the mean specific balance for larger perturbations of mean air temperature (at sea level) is shown in fig. 6. Apparently, the sensitivity of the balance to temperature does not critically depend on the temperature itself, at least not in the range considered here. A tentative conclusion could be that, although the absolute balance between accumulation, ablation and calving is not accurately known, the estimate of the sensitivity is not too bad. *This does not imply, however, that we know whether the ice sheet is in balance with the present climate.*

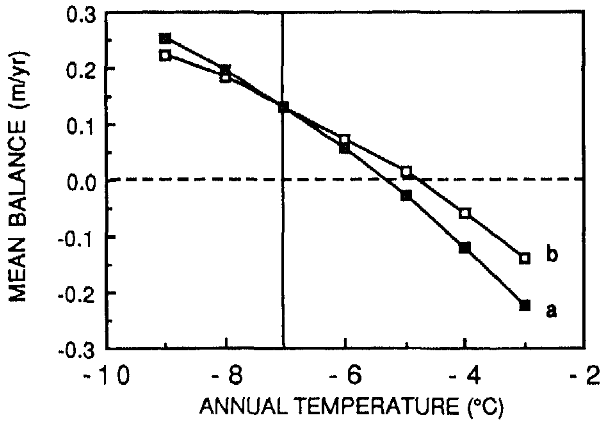

Fig. 6: The dependence of the mean specific balance on annual mean air temperature (curve a). Accepting $\Theta_{\mathrm{ann}} = -7^\circ C$ as the present state, the balance would become negative if temperature would rise by more than 2.2 K. Curve b is with 'precipitation feedback' as explained in the text

For a +10 % change in annual precipitation, the specific balances change by +0.030, +0.054, +0.036 and +0.050 m/yr for the regions 1, 2, 3 and 4. The mean change amounts to +0.041 m/yr. It has been argued that precipitation on the Greenland ice sheet would increase when air temperature goes up (e. g. Reeh et al., 1979). Here it is assumed that precipitation varies in proportion to the amount of maximum precipitable water in an atmospheric column. Following Oerlemans and Van der Veen (1984) [p. 140, where precipitable water is given as a function of sea-level temperature and lower bound of the atmospheric column] the increase in precipitation would then be of the order of 4 % per degree K temperature increase. Using such a coupling of temperature and precipitation in the input for the energy-balance model then yields the second curve in fig. 6. As expected, the change of mean specific balance with temperature is somewhat smaller now (−0.060 m/yr instead of −0.077 m/yr).

Fig. 7 summarizes the implications of the present results for sea level change. As the area of the Greenland ice sheet is 0.00471 of that of the world ocean, equivalent changes of sea level are easily calculated. A 1 K temperature increase would then increase sea level by 0.36 mm/yr. The influence of cloud cover is twofold: more clouds reduce the amount of solar radiation reaching the surface, but also increase the longwave radiation balance. It thus depends on the albedo whether the net effect is an increase or decrease of the energy balance at the surface, as shown already by Ambach (1974). In the present model, the break-even point is found for an albedo between 0.6 and 0.7. For the total mass budget of the ice sheet, decreasing solar radiation for larger cloudiness slightly dominates.

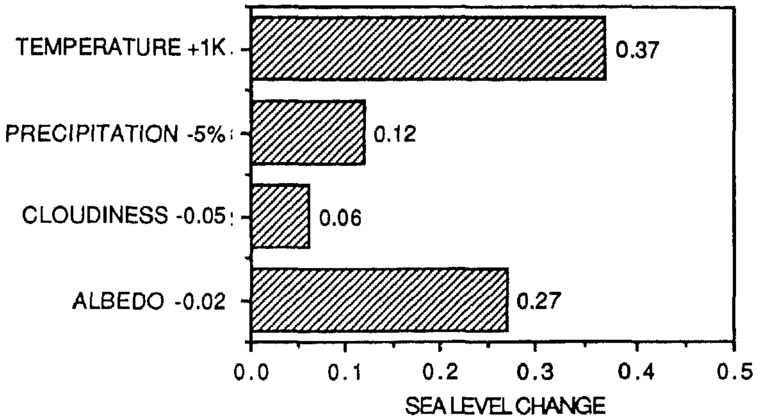

Fig. 7: Rate of sea-level change associated with climatic forcing of the Greenland ice sheet, as found from the energy balance model

As was found for alpine glaciers, the effect of a small change in the mean surface albedo is impressively large. Whether externally caused changes of α of this order of magnitude occur on a decadal or century time scale is unknown, but it seems very well possible.

Finally, some experiments were done in which the Greenland ice sheet was represented in a single mass balance profile, with weighted mean values for the input data from table 1, and applied to the mean area-elevation distribution. The calculated mass balance was very close to the result of table 2 (as listed in the column 'total'), and even the sensitivity to climatic change was within a few % of the earlier result. So the energy balances of the different regions appear to be additive to a large extent: it is really the altitude dependence that should be represented well. This result implies that running the energy-balance model in a version with high spatial resolution is not useful — the change in the mass budget would be very small compared to a change in the input data (within their range of uncertainty).

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