Mass balance parameterisation for Hans Tausen Iskappe, Peary Land, North Greenland

By Niels Reeh, Ole B. Olesen, Henrik Højmark Thomsen, Wolfgang Starzer, and Carl Egede Bøggild

Abstract

Reeh, N., O. B. Olesen, H. H. Thomsen, W. Starzer, and C. E. Bøggild 2001. Mass balance parameterisation for Hans Tausen Iskappe, Peary Land, North Greenland. Copenhagen, Danish Polar Center. Meddelelser om Grønland Geoscience 39, pp. 57-69.

A mass balance model based on a positive degree-day approach is used to model the mass balance-elevation relationship for Hans Tausen Iskappe (82.5°N, 37.5°W). Model parameters are estimated by means of field data from a glacier basin in the north-east corner of the ice cap. The mass balance model is run for the total Hans Tausen Iskappe by using additional information on the snow fall distribution based on accumulation data from three local ice cores. The model indicates that, in the period 1975-1995, the total balance of Hans Tausen Iskappe was negative (–0.14 m/y of ice equivalent averaged over the ice cap, corresponding to –104% of the annual average accumulation in the period). During the same period, the central part of the ice cap was very likely thickening. The sensitivity of the total mass balance to changing summer temperature is –0.17 m ice/y/K. With a 5% increase of snow fall per degree increase of summer temperature, the sensitivity is changed to –0.14 m ice/y/K. Model studies with larger deviations of the summer temperature from the present value indicate that a 5 K warmer temperature would result in ablation over the entire ice cap, which would then melt away completely in a few hundred years. Even a three-fold, simultaneous increase of the accumulation rate would not in general restore the mass balance, but might secure survival of small isolated ice caps in the northern mountainous landscape. The firn warming (the increase of the temperature at 10 m depth above the mean annual air temperature) of the central area of the ice cap is also studied. Extreme changes of climate conditions as those mentioned above, are needed in order to change the thermal regime of the central region of Hans Tausen Iskappe from cold to temperate. This indicates that Hans Tausen Iskappe was a cold glacier during most of its existence.

Keywords: Glacier mass balance; firn warming; North Greenland

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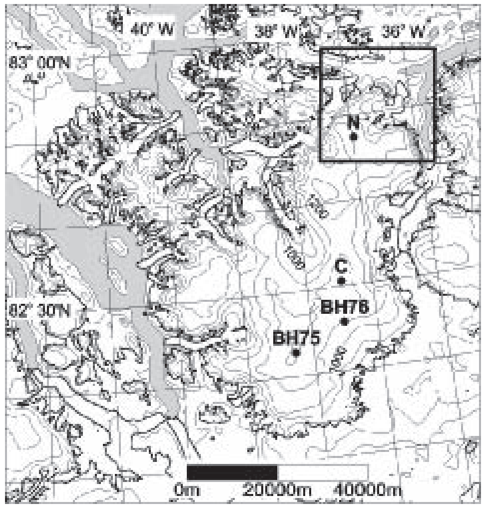

Fig. 1. Map of Hans Tausen Iskappe. North Dome (N), Central Dome (C), and drill locations BH75 and BH 76 are indicated. The square delimited by a thick line in the north-east corner of the map shows the area covered by the map in Fig. 2.

Introduction

The main purpose of this paper is to set up a model for estimating the present mass balance of Hans Tausen Iskappe and its sensitivity to climate change. Hans Tausen Iskappe is a local ice cap, about 75 km from north to south and 50 km from east to west, located in western Peary Land (Fig. 1). Covering an area of 3975 km^2 (Starzer and Reeh 2001), it rests on one of the widespread c. 1000 metre high plateau of Silurian rocks, dominating the landscape of Peary Land. The ice cap has several domes (outflow centres) which reach elevations of 1200 to 1300 m above sea level (a.s.l.). Several outflow glaciers drain the ice cap to the west, north and east, often terminating at elevations of a few hundred metres or reaching sea level with a calving front. The southern margin of the ice cap can be characterised as a "quiet" sector, covered by snow drifts that often survive the summer melt period. Here, the sub-ice landscape is relatively smooth. The maximum ice thickness is around 300 metre, except in a marked, deep valley trending north-south, where the thickness locally reaches a value of 600 metre. In contrast, the northern part of the ice cap rests on a mountainous terrain with ice-thickness variations typically between 100 and 450 metre (Starzer & Reeh 2001).

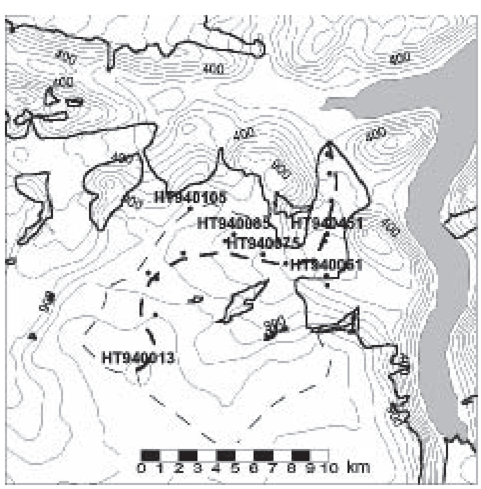

Fig. 2. Map of Hare glacier basin. Basin limit is shown by a thin dashed line. The central flow line is shown by a thick dashed line. Stake and pit locations for mass balance measurements are shown as dots.

Mass Balance Observations

In the summers of 1994 and 1995, field work was carried out on the tongue and in the drainage basin of an outlet glacier located in the north-east corner of the ice cap, in the following referred to as Hare glacier (Fig. 2). An important part of the field work was mass balance studies and related measurements of glacier climate and englacial temperatures. The mass balance measurements were performed as snow-pit/firn-core studies and stake readings along the centre line of the drainage basin. They show that considerable summer melting takes place even on the higher parts of the ice cap. In peak summer, several decimetres thick slush develops over extended areas of the glacier basin. A significant part of the melt water penetrates into the snow/firn pack where it re-freezes and creates high-density layers or ice layers in the snow pack. This process releases energy (latent heat) causing local warming of the surface-near layers of the glacier, as confirmed by repeated measurements of ice temperatures down to 10 m depth at several locations on the glacier. From the local dome at 1320 m a.s.l. to the glacier terminus at c. 100 m a.s.l, the following zones are represented (in descending order): A snow zone (percolation zone, wet and soaked snow), a superimposed ice zone (covered by slush in peak summer), a second snow zone, a second superimposed ice zone, and an ice ablation zone.

In Fig. 3, 'winter balance' (August 1994-June 1995), 'summer balance' (June-August, 1995), and annual balance (August 1994-August 1995) along the central flow line of Hare glacier are shown as functions of elevation. All mass balance-elevation curves show large fluctuations in respect to smooth trend lines. The fluctuations are mainly due to uneven distribution of the snow accumulation caused by slope dependent wind drifting. The 1994/95 annual equilibrium line altitude was located near an elevation of 700 m a.s.l.

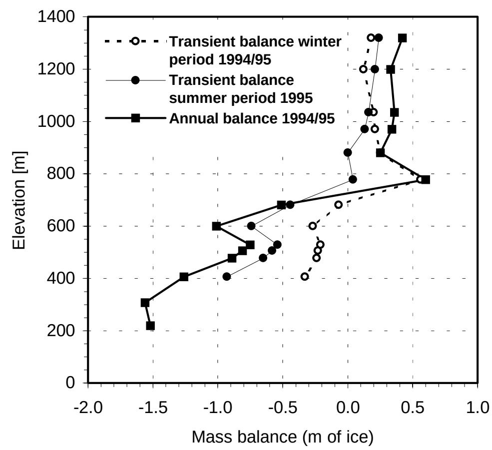

Fig. 3. Measured mass balance as function of elevation in the Hare glacier basin.

Mass Balance Model for Hare Glacier

The model, which is based on the approach described by Reeh (1989), has proven to be a useful driver for time dependent ice-dynamics models (e.g. Huybrechts *et al.*, 1991). The model calculates melt rate and snow/ice surface temperature, using parameterisations of annual snow fall, mean annual air temperature and mean July air temperature as input. Melt rates are calculated using positive degree-days, and firn warming (i.e. the positive deviation of the temperature at 10-15 m depth in the firn from the mean annual air temperature) is estimated by means of the model-calculated amount of refrozen melt water in the firn. The parameters of the melt model are determined by means of field data collected on the tongue and in the drainage basin of Hare glacier. The limited extent of this basin justifies to parameterise temperature and snow accumulation only in terms of elevation. For the modelling of the total mass balance of Hans Tausen Iskappe, the parameterisations must be generalised to account also for latitudinal gradients of temperature and snow accumulation.

Annual snow fall

The annual snow fall needed as input to the mass balance model is determined as follows: In the accumulation area, the annual snow fall is put equal to the annual balance. Since part of the annual balance is probably supplied by freezing of rain in the summer season, this results in a slight overestimation of the annual snow fall. In the ablation zone, the annual snow fall is estimated based on soundings of snow depths in June 1995.

Air-temperature

The 1995-summer air temperature records at three sites HT940061 (574 m a.s.l.) HT940075 (777 m a.s.l.), and HT940013 (1320 m a.s.l.) (for location, see map in Fig. 2) are displayed in Fig. 4. Unfortunately, successful recording of air temperature at HT940013 on the local dome lasted only from June 20 to June 25, whereas temperatures were successfully recorded at the two other locations during the entire measurement period from June 20 to August 14. The temperature record for the same period from Kap Molkte, located about 100 kilometres to the south-east of Hans

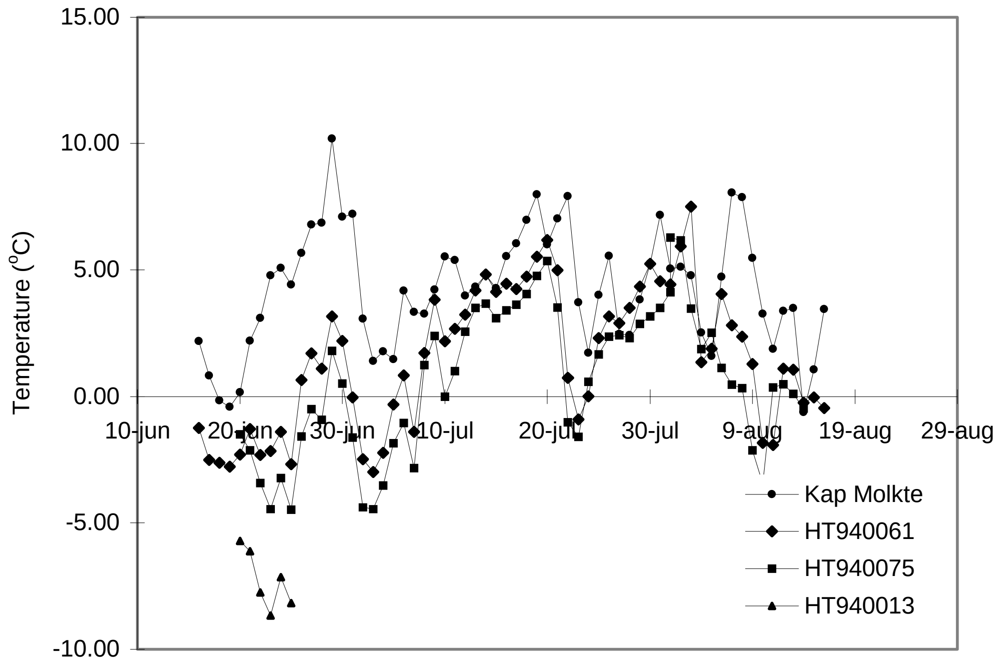

Fig. 4. Air-temperature records observed in the summer of 1995 at stations in the Hare glacier basin (for location, see Fig. 2.) and at Kap Molkte located about 100 kilometre south-east of Hans Tausen Iskappe at an elevation of 13 m.

Tausen Iskappe at an elevation of 13 m, is also shown in Fig. 4. In general, the records show synchronous variations, although discrepancies occur. In Fig. 5 average air temperatures for two periods (June 20-25 and July 1-31) are plotted versus elevation. The first period includes observations from all four sites. The second period, representing peak summer, includes only Kap Moltke, HT940075 and HT940061. It appears that the air-temperature/elevation gradient is more negative in the former period (–0.0081 K/m) than in the latter period (–0.0047 K/m), probably due to generally more humid weather in July than in June. If the distant Kap Molkte data are excluded, the corresponding gradients become –0.0071 K/m and –0.0056 K/m, respectively. We choose the latter value to represent the average gradient in the summer period. This gives the following parameterisation of July air-temperature in the Hare glacier basin:

$$TJM = 5.62 - 0.0056\ E, \qquad (1)$$

where *E* is elevation.

Unfortunately, mean annual air temperature has not been measured on Hans Tausen Iskappe. Reeh (1989) derived the following parameterisation of mean annual air temperature on the Greenland ice sheet by linear regression of observed ice-sheet mean annual air temperatures *TMA* on elevation *E* and latitude *L*:

$$TMA = 48.4 - 0.751\ L - 0.00792\ E. \qquad (2)$$

The temperature inversion occurring at low altitudes in North Greenland (Ohmura, 1987) was accounted for by putting the temperature gradient below 300 m equal to zero north of 75°N. The

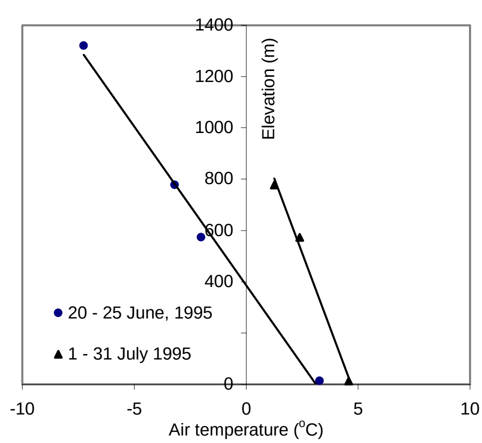

Fig. 5. Air-temperature/elevation relationship observed in the summer of 1995 in the Hare glacier basin. Summer temperatures at Kap Molkte, located about 100 kilometre south-east of Hans Tausen Iskappe at an elevation of 13 m, is also shown.

temperature-latitude gradient of 0.751 K/degree latitude appearing in Equation(2) is in good agreement with the corresponding gradient of 0.7 K/degree latitude derived from temperature records at climate stations in the ice free area in East Greenland (Funder et al, 1998). For Kap Moltke, Equation (2) yields a mean annual air temperature of –15.7°C which is 0.6°C cooler than the 1974-1995-average of the mean annual temperatures measured at Kap Molkte, and 0.8°C cooler than the average (-14.9°C) of the mean annual temperatures at Kap Molkte for the two years of mass balance observations 1994 and 1995. At the mean latitude of Hare Glacier about 0.6 degree of latitude north of Kap Molkte, the mean annual air temperature at sea level for 1994-1995 is accordingly estimated at –14.9 – 0.6*0.751= –15.4°C. Altogether, this leads to the following parameterisation of the mean annual air temperature in the Hare glacier basin for the balance year 1994-1995

$$TMA = \begin{cases} -15.4 \ \textit{for } E < 300\ m \\ -15.4 - 0.00792\ (\mathrm{E} - 300) \\ \qquad \textit{for } E > 300\ m \end{cases} \tag{3}$$

10-metre firn/ice temperature

The relationship expressed by Equation(3) is displayed in Fig. 6 together with observed 10-metre firn/ice temperatures. Three sites (HT940013, HT940105, and HT940075) (for location, see map in Fig. 2) show remarkably large positive deviations from the air-temperature curve. These sites are all located in snow-accumulation areas where snow-pit/firn core studies indicate substantial re-freezing of meltwater, causing warming of the firn. Site HT940085 is located in the upper superimposed ice zone, where most of the latent heat produced by the re-freezing process is released to the atmosphere. Consequently, the 10-metre temperature at this site is expected to deviate little

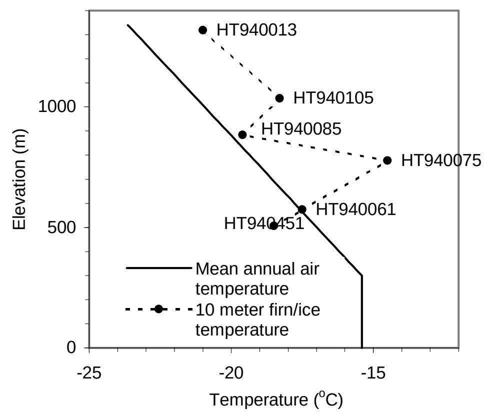

Fig. 6. Model estimated mean annual air temperature and measured 10-metre firn/ice temperature versus elevation in the Hare glacier basin.

from the mean annual air temperature as actually seen in Fig. 6. Sites HT940061 and HT940451 are both in the ablation area, where firn warming does not occur. The negative deviation of the 10-metre temperature from the mean annual air temperature at site HT940451, located several kilometres down-stream from the equilibrium line, is probably due to upward advection of cold ice originating from a site in the higher elevated and consequently colder accumulation region.

Degree-day factors

A degree-day factor for ice melt of 0.0065 m of ice/degree-day (0.0059 m of water/degree-day) is derived from a local detailed stake-farm study (CEC, 1995, Table 2, p7). For snow-melt, the degree-day factor is set to 0.0027 m of ice/degree-day in accordance with the degree-day factor of 0.0025 m of water/degree-day published by Braithwaite (1995).

In order to account for the daily temperature cycle and random temperature deviations from an assumed regular sine-variation of the average annual temperature cycle, a statistic with standard deviation *s* is introduced in the degree-day calculation (Reeh, 1989). In the Greenland ice sheet study of Reeh (1989), *s* was set equal to 4.5 K. For reasons to be explained in the following

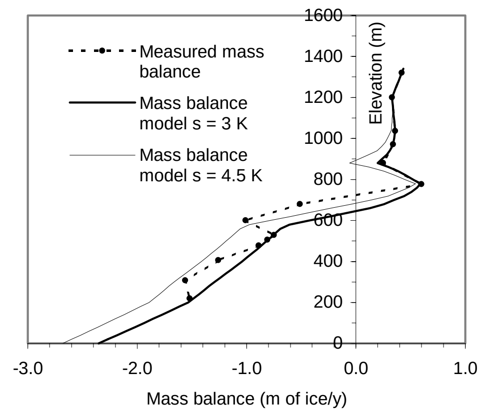

Fig. 7. Comparison of measured and model-estimated mass-balance/elevation relationships for Hare glacier basin.

section, we choose s = 3 K in the melt-rate model for the Hare glacier basin. Another parameter in the mass balance model is the amount of ice (expressed as a fraction P_{MAX} of the total snow fall) formed by re-freezing of melt water in the upper firn before run-off begins. We choose P_{MAX} = 0.6 in accordance with Reeh (1989).

Results

In Fig. 7, the model calculated annual mass-balance elevation relationship for the Hare glacier basin (heavy full curve) is compared to the observed mass balances. The close agreement between model and observations in the accumulation zone reflects the fact that the model predicts that there is no run-off (mass loss) above c. 800 m elevation. This is in accordance with our snow-pit/firn-core observations. In the ablation zone, the model generally underestimates the ablation. By increasing the value of the standard deviation of the statistic s from 3 K to 4.5 K (the value used by Reeh (1989) for the Greenland ice sheet) a better fit (although slightly too high) to the observations in the ablation zone is obtained (thin full curve in Fig. 7). However, with this s-value, the model predicts substantial run-off in the altitude interval 780-1100 m in disagreement with our field observations. Moreover, the model-predicted amount of refrozen melt water in the upper firn greatly exceeds the amount observed in our snow-pit/firn-core study, see Table 2.

In Tables 1 and 2, further comparisons between observations and model results are given. Table 1 shows positive degree-days in the period June 20 to August 14, 1995 calculated by means of the temperature records shown in Fig. 4. The positive degree-days estimated by the mass-balance model are also shown. The increasing excess with decreasing elevation of the model-estimated PDD over the observed PDD can be explained by the limited length of the observation period. The lover the elevation, the more likely is the occurrence of positive temperatures before June 20 and after August 14. The contribution from these temperatures is missing in the calculation of the "observed" PDD, but are included in the model estimate giving the PDD for a full year.

In Table 2, observed 10-metre firn/ice temperature, mean annual air temperature as calculated by Equation(3), and their difference (firn warming) are listed. The weight percentage of ice layers as observed in the snow-pit/firn-cores on June 23 and August 16 1995 and the change of this percentage are also shown in the table. At stake HT940085 located in the upper superimposed ice zone, the change is 100%. The observed amount of re-frozen melt-water is derived by multiplying the annual mass balance with the percentage change of ice layers. Model derived amounts of refrozen melt-water are shown for two different values of the standard deviation s. For s = 3K there is excellent agreement be-

Table 1. Observed (20 June – 14 August, 1995) and model-estimated positive degree-days (PDD).

Station	Elevation (m)	PDD (observation)	PDD (model)
Kap Moltke	13	253	350
HT940061	574	127	150
HT940075	777	91	95

Stake	Elevation (m)	10-metre temp. (°C)	Mean annual air temp. (°C)	Firn warming (K)	Percen-tage ice June 23	Percen-tage ice August 16	Change of ice-per-centage	Mass balance (m ice/y)	Refrozen melt-water (m ice/y) Observed	Modelled	
HT940451	506	–18.5	–17.0	–1.5						s=3 K	s= 4.5 K
HT940061	574	–17.5	–17.6	+0.1							
HT940075	776	–14.5	–19.2	+4.7	10	53	43	0.60	0.26	0.26	0.35
HT940085	883	–19.6	–20.0	+0.4	0	100	100	0.16	0.16	0.15	0.15
HT940105	1035	–18.3	–21.3	+3.0	47	72	25	0.36	0.09	0.13	0.22
HT940013	1318	–21.0	–23.5	+2.5	0	13	13	0.42	0.055	0.055	0.14

Table 2. Observed firn/ice 10-metre temperature, model-calculated mean annual air temperature, and observed and model-calculated amounts of refrozen melt water.

tween observed and modelled values, whereas the modelling with s = 4.5K overestimates the amount of refrozen melt water. In Fig. 8, the amount of refrozen melt water that remains after the summer melt season (expressed as a fraction of the annual snow fall) as estimated from the mass-balance model with s = 3K, is shown together with the observed amount.

Disregarding location HT940085 in the superimposed ice zone where most of the latent heat produced in the refreezing process is released directly to the atmosphere and therefore causes little – if any – warming of the near-surface layer, a regression of firn warming on amount of refrozen melt water (forced to pass through the origin (0,0)), yields a regression coefficient of b_{FW} = 20 K /m ice. This value is close to the corresponding value (27 K /m ice) found in the study by Reeh (1989) using all available 10-metre temperature observations on the Greenland ice sheet. Since the regression coefficient determined in the total Greenland study is based on c. 170 10-metre temperature observations as compared to 3 observations in the present study, we will use the formerly derived value b_{FW} = 27 K /m ice to estimate firn warming also in thus study.

Mass Balance Model for Hans Tausen Iskappe

The modelling of the total mass balance of the Hans Tausen Iskappe requires generalisations of the parameterisations derived for the Hare glacier drainage basin.

Accumulation

The spatial distribution of the annual snow fall (accumulation) on Hans Tausen Iskappe is derived from the snow-pit and stake observations in the Hare glacier basin discussed above. Information is also provided by three firn/ice cores drilled at the central dome in 1994 and 1995 and by two firn/ice cores from the southern part of the ice cap drilled in 1975 and 1976 (Clausen *et al.* 2001). The accumulation data are summarised in the Table 3.

Table 3 displays large accumulation-rate variations both spatially and tempo-

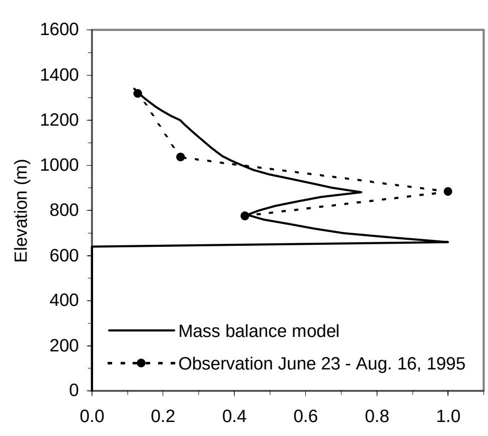

Fig. 8. Observed and model-estimated refrozen amount of melt-water remaining at the end of the 1995-summer season in the Hare glacier basin expressed as a fraction of the annual snow fall.

Table 3. Hans Tausen Iskappe: Mean annual accumulation rates (m of ice per year) in different periods.

Site	°N	°W	Altitude (m)	1994	1912-1975	1783-1975	1975-1995
North Dome	82.79	36.98	1318	0.425			*0.306*
Central Dome	82.51	37.47	1275	0.132	0.120[1)]	0.109[2)]	0.095
BH75	82.38	38.28	1150	*0.171*	0.155	0.147	*0.123*
BH76	82.43	37.50	1125	*0.158*	0.147[3)]	0.141[3)]	*0,116*

Comments: Numbers in regular style are measured values. Italicised numbers are inferred values. [1)] calculated by using mean values for the periods 1912-1995 and 1975-l995. [2)] calculated by using mean values for the periods 1783-1995 and 19751995. [3)] 1976-accumulation included.

rally. Because of the large temporal variations, accumulation-rate values at the different sites must be referred to the same period of time when used to derive the spatial accumulation rate distribution over the ice cap. Unfortunately, this requirement cannot be met by using directly measured accumulation-rate values on Hans Tausen Iskappe for any period of time. However, the accumulation rate record from the central dome has at least one overlapping period with the records from all other sites, and can therefore be used to scale accumulation rates from one period to another.

It is clear from Table 3, that the latitudinal accumulation rate gradient is much larger between the northern and central domes than south of the central dome. A likely explanation is that the storms delivering the major part of the precipitation on Hans Tausen Iskappe come from the north, thus releasing an increasing amount of precipitation as the air masses move up the north-facing mountain-fringed part of the ice cap. The southern slope is at the lee side of the ice cap, and thus receives less precipitation. Here catabatic winds probably play a large role by re-distributing the snow, leading to increasing snow accumulation in the direction of the southern margin, as actually indicated by the large snow drifts seen on aerial photographs.

The different accumulation regimes require different parameterisations of the annual snow fall north and south of the central dome located at 82.51°N. North of this latitude, a snow fall elevation gradient of 0.00036 m ice equiv./m is determined by linear smoothing of the accumulation and winter-snow data from the Hare glacier basin discussed in the previous section. As mentioned previously, surface-slope variations in this basin cause substantial local deviations of snow accumulation from the linear distribution. Although similar deviations are expected all over the ice cap, the limited number of accumulation observations can only justify simple parameterisations in terms of linear dependencies on elevation and latitude. Local deviations from the simple parameterisation will certainly occur, but the deviations will to some extent cancel out in estimates of the overall mass balance of the ice cap.

Using the 1994-values of accumulation rate at the northern and central domes, the latitudinal gradient north of 82.51°N is determined as 0.989 m ice equiv./degree latitude. South of 82.51°N, a linear regression of the accumulation values at the central dome, BH75 and BH76 on latitude and elevation shows that the elevation gradient is insignificant, supporting the idea that orography is of little importance for the snow-distribution on that part of the ice sheet. The snow fall distribution south of 82.51°N is therefore represented by a linear regression of accumulation rate on latitude.

Altogether, this results in the following parameterisation of snow fall on Hans Tausen Iskappe for 1994:

$$ACC=\begin{cases}0.00036\,E + 0.989\,L - 81.93 & \text{for } L \text{ (} 82.51°\text{N} \\ -0.307\,L + 25.46 & \text{for } L < 82.51°\text{N}\end{cases} \quad (5)$$

Table 3 shows that the snow fall on Hans Tausen Iskappe was excessive in 1994. At the central dome, the snow accumulation rate in 1994 (0.132 m ice/year) was 20% higher than the long term (934-1995) mean accumulation rate of 0.110 m ice/year (Clausen *et al*. 2001). 1994 was a year with high accumulation also at the northern dome. This follows from comparing the accumulation-rate value for 1994 (0.425 m ice/year, see Table 3) with the significantly smaller long-term balance values that can be derived from the measured submergence velocity (0.2-0.3 m ice/year) and the local strain net (0.24 m ice/year) at the dome.

The parameterisation of the snow fall on Hans Tausen Iskappe for other periods (e.g. 1975-1995) is determined by multiplying Equation(5) by the ratio between the accumulation rate at the central dome for the period in question and the 1994-value of the accumulation-rate at this location.

Parameterisation of air temperature

The parameterisation of the July air temperature over Hans Tausen Iskappe is obtained by combining the temperature elevation relationship derived for the Hare glacier basin around 82.8°N (Equation(1)) with the July air temperature gradient of --0.4K per degree latitude derived from East Greenland coastal climate stations (Funder *et al*. 1998). This results in the following generalised parameterisation of the 1994 July air temperature

$$TJM = 38.74 - 0.4\,L - 0.0056\,E \quad (6)$$

For the 1994-1995 mean annual air temperature, we use Equation(3) combined with the mean annual temperature gradient of –0.751K per degree latitude given in Equation(2) to obtain

$$TMA=\begin{cases}46.8 - 0.751\,L & \text{for } E \text{ (} 300\text{ m} \\ 49.2 - 0.751\,L - 0.00792\,E & \text{for } E > 300\text{ m}\end{cases} \quad (7)$$

For the period 1975-1995, air-temperature data are available from Kap Moltke. For this period, mean July and mean annual air temperatures over Hans Tausen Iskappe are derived by adjusting the temperatures given by Equations (6) and (7), assuming that the interannual temperature variations at Kap Moltke are representative also for Hans Tausen Iskappe.

Model Results for Hans Tausen Iskappe

All model runs for the Hans Tausen Iskappe are performed with the same degree-day factors, *s*- and P_{MAX} -values that were used for the Hare glacier basin. Calculations are performed for 1994 and for the period 1975-1995 (in the following used as a reference period) applying the parameterisations of snow fall and air temperature discussed in the previous section.

Mass Balance

The model calculation indicate that, in 1994 the total balance of Hans Tausen Iskappe was negative, viz. –0.08 m ice equivalent/y averaged over the ice cap, corresponding to –42% of the average accumulation rate (0.192 m of ice equivalent/y). For the reference period 1975-1995 with a mean accumulation rate of 72% of the 1994 value and a mean summer temperature 0.26 K lower than the 1994 summer temperature, the total balance was even more negative (–0.14 m of ice equivalent/y). This corresponds to –104% of the mean accumulation averaged over the ice cap. The total mass balances of the ice cap for the year with the warmest summer (1989), and the

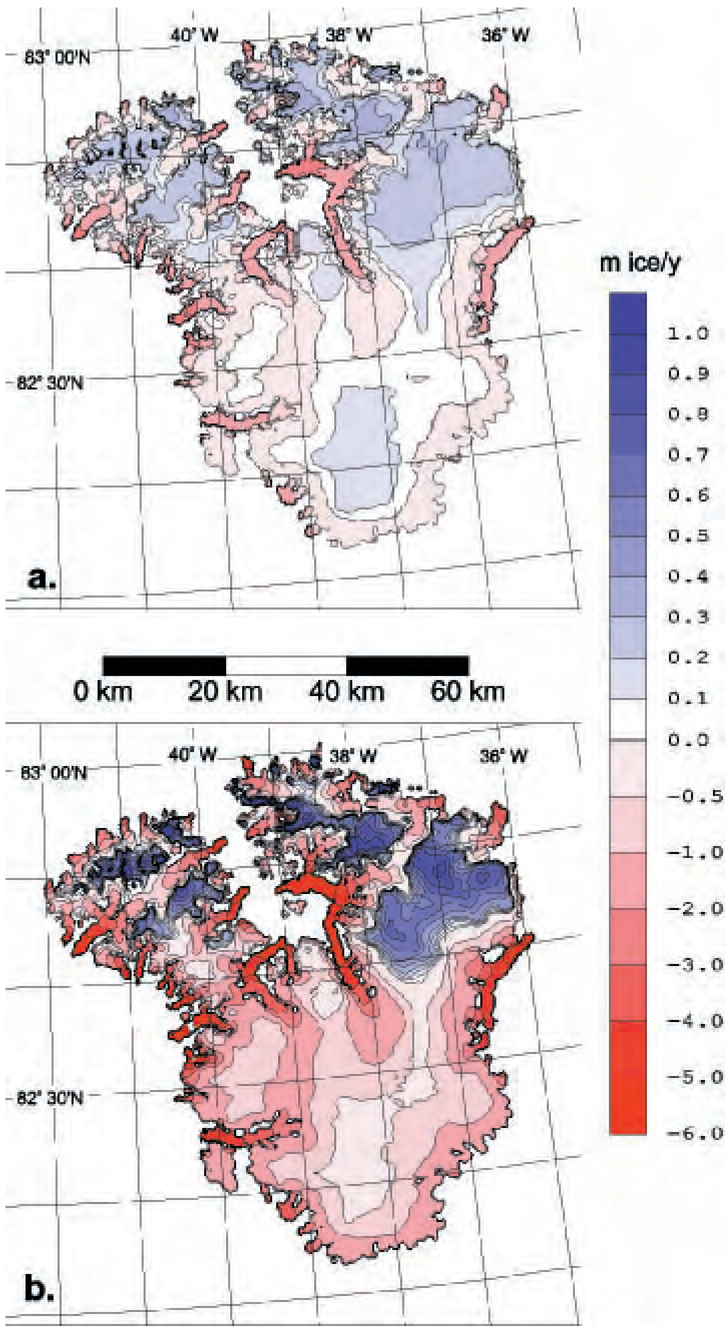

Fig. 9. Model-estimated distribution of mass balance on Hans Tausen Iskappe. a: 1975-1995 mean climate conditions. b: Holocene climatic optimum conditions (5K warmer temperatures and 3 times higher accumulation rate than at present).

year with the coldest summer (1982) are calculated at –0.32 m ice and –0.035 m ice, respectively.

At first sight, these numbers appear rather alarming for the present state of health of the ice cap. Furthermore, the results seem to contradict the conclusion of Keller *et al.* (2001) that the surface elevation of the central dome is presently increasing and probably has been increasing for a long period of time. However, inspection of Fig. 9a, showing the calculated mass balance distribution for Hans Tausen Iskappe for the period 1975-1995, reveals that the mass loss is concentrated at a number of extended outlet glaciers with elevations below 500 metre. These glaciers might be relicts formed in the cool climate of the Little Ice Age and may presently be subject to down-wasting, which seems to be a wide-spread mass-loss mechanism – rather than horizontal recession – for the cold based glaciers in North Greenland (Weidick 2001). Hence, the present mass balance condition for Hans Tausen Iskappe might be thickening of the central accumulation area due to increased snow accumulation associated with the temperature increase after the Little Ice Age, and thinning of the extended low elevated outlet glaciers due to the increased summer warmth, altogether leading to a steeper ice cap. Whether or not Hans Tausen Iskappe with time is able to reach a stable situation under present climate conditions cannot be settled by mass balance considerations alone, but will have to await the results of a planned glacier dynamics model study. Our preliminary conclusion is that, in spite of possible thickening of the central region, Hans Tausen Iskappe is most likely loosing mass under present climate conditions.

Mass Balance Sensitivity

The sensitivity of the mass balance of Hans Tausen Iskappe to climate change is illustrated in Table 4 showing the calculated total mass balance for different changes of the mean annual temperature (T) and different percentage changes of snow accumulation per degree change of temperature $(p/(p(T)$. All changes are in respect to the 1975-95 mean values. The table shows that a summer temperature 1.3 K cooler than present will bring the ice cap in balance with the mean

$\Delta p/(p\Delta T)$	ΔT=–2 K	ΔT=–1 K	ΔT=0 K	ΔT=+1 K	ΔT=+2 K	ΔT=+5 K
0 % K-1	0.046	–0.021	–0.141	–0.335	–0.606	–1.874
5 % K-1	0.031	–0.031	–0.141	–0.318	–0.568	–1.759
10 % K-1	0.016	–0.040	–0.141	–0.302	–0.529	–1.620
20 % K-1	–0.008	–0.057	–0.141	–0.270	–0.449	–1.266

Table 4. Total mass balance (m ice/y) of Hans Tausen Iskappe for different changes of the mean annual temperature ΔT and different percentage changes of snow accumulation per degree change of temperature $\Delta p/(p\Delta T)$ in respect to the 1975-1995 mean values.

snow accumulation in the period 1975-1995.

With a temperature increase of 1-2 K in respect to the present temperature, the balance becomes exceedingly negative even if accumulation increases by 10-20% per degree warming. For temperatures 5 K warmer than now and a 20% increase of accumulation per degree warming (a scenario for the Holocene climatic optimum), there is a significant mass loss even from the highest elevated areas of the ice cap. The total annual mass loss for this scenario amounts to 4.3 km^3/y. With a total volume of the present ice cap of 763.5 km^3 (Starzer and Reeh, 2001), the ice cap would under such climate conditions disappear in a few hundred years. Thus the results of our mass balance modelling adds to the evidence that Hans Tausen Iskappe melted away completely during the Holocene climatic optimum.

The mass-balance distribution over the ice cap for an extreme Holocene climate optimum scenario with air temperature increased by 5K and accumulation rate increased by a factor of 3 as compared to their 1975-1995 mean values is shown in Fig. 9b. The figure shows that, even with such a large increase of accumulation rate, most of the ice cap would experience massive melting. Whether or not small areas in the northern part of the ice cap with positive mass balance might survive as small local ice caps cannot be settled by mass balance modelling alone, but will have to await the results of the previously mentioned, planned glacier dynamics model study.

Firn warming

In Fig. 10, calculated firn-temperature distributions of Hans Tausen Iskappe are shown for the climate conditions of the 1975-1995 reference period and for the extreme Holocene climate optimum scenario with temperature 5K warmer than now and accumulation rate three times higher than now. The calculations are performed with a firn-warming coefficient b_{FW} equal to 27 K /m. Fig. 10b shows that even in this extreme climate optimum scenario, firn temperatures in the central region of the ice cap would still be below the freezing point. Lower surface elevations either during the period of deterioration of the ice-age ice cap or during the build-up of the existing ice cap may in periods have caused air/firn temperatures a few degrees higher than those shown in Fig. 10b. Moreover, the geothermal heat flux causes higher temperature at the base of the ice cap than at the surface. Therefore, we cannot exclude the possibility that parts of the central region of Hans Tausen Iskappe – particularly the high accumulation region to the north – were temperate or experienced basal melting during certain periods of the Holocene climatic optimum.

Conclusions

Our study has shown that it is possible to model the observed mass-balance and firn-warming distributions of an outlet glacier basin of Hans Tausen Iskappe by using reasonable values of the model parameters of the degree-day model of Reeh (1989). Application of the mass-

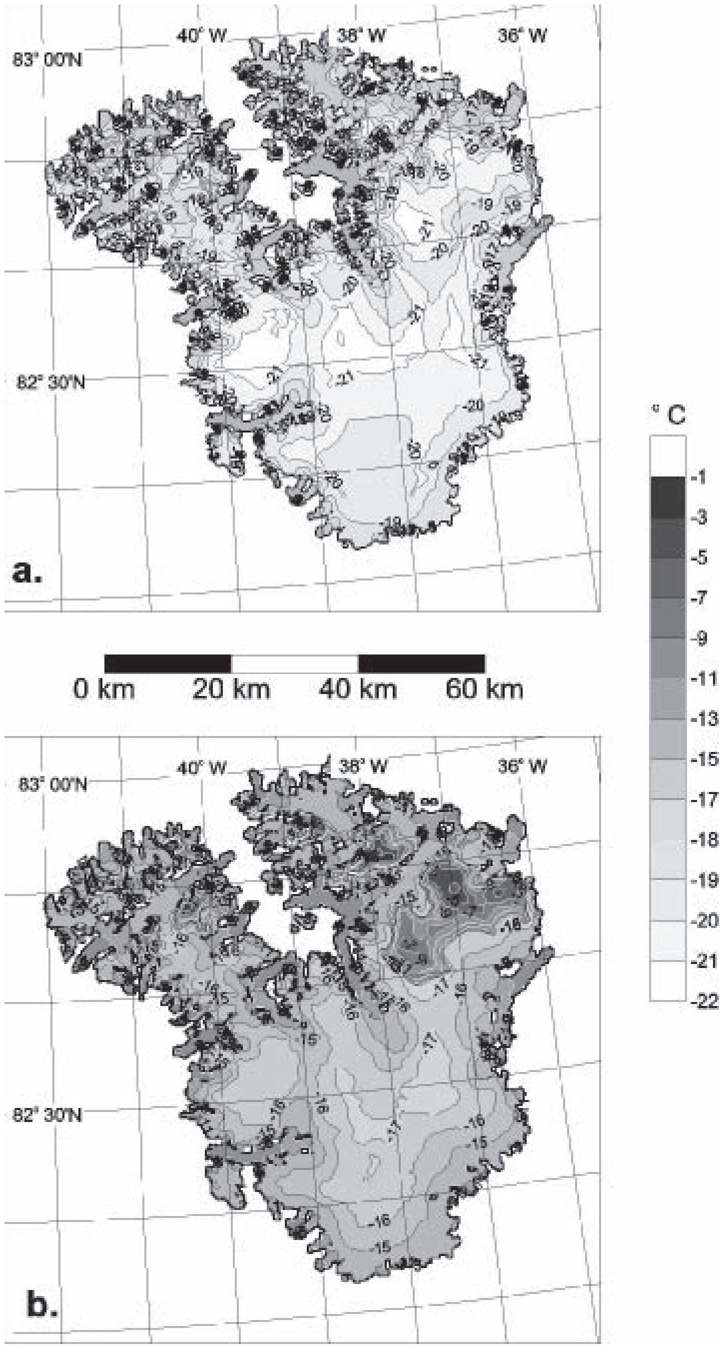

Fig. 10. Model-estimated distribution of firn/ice temperature on Hans Tausen Iskappe. a: 1975-1995 mean climate conditions. b: Holocene climatic optimum conditions (5K warmer temperatures and 3 times higher accumulation rate than at present).

balance model indicates that in 1994 the total balance of Hans Tausen Iskappe was negative, viz. –0.08 m ice equivalent/y averaged over the ice cap, corresponding to –42% of the average accumulation over the ice cap for that year (0.192 m of ice equivalent/y). For the period 1975-1995, the total balance was even more negative (–0.14 m of ice equivalent/y, corresponding to –104% of the mean accumulation for that period averaged over the ice cap).

Keller *et al.* (2001) conclude that the surface elevation of the central dome is presently increasing and probably has been increasing for a long period of time. Hence, at present, Hans Tausen Iskappe might be thickening in the central accumulation area probably due to increased snow accumulation associated with the temperature increase after the Little Ice Age, and at the same time experience thinning of extended low-elevated outlet glaciers due to the increased summer warmth. This will altogether lead to a steeper ice cap. Our preliminary conclusion is that, in spite of the thickening of the central region, Hans Tausen Iskappe is most likely loosing mass under present climate conditions. However, with time, the icecap will probably be able to reach a stable situation with reduced ablation area. A summer temperature only 1.3 K cooler than present will bring the ice cap in balance with the mean snow accumulation in the period 1975-1995, demonstrating the high sensitivity of the icecap's mass balance to summer temperature.

For a temperature 5 K warmer than now and a 20% increase of the accumulation per degree warming (a scenario for the Holocene climatic optimum) model calculations show a significant mass loss even from the highest elevated areas of the ice cap. For this scenario, the total annual mass loss amounts to 4.3 km^3/y. With a total volume of the present ice cap of 763.5 km^3 (Starzer and Reeh 2001) the ice cap would deteriorate in a few hundred years. This supports other evidence that Hans Tausen Iskappe melted away completely during the Holocene climatic optimum.

A calculation of the firn-temperature distributions on Hans Tausen Iskappe, using a climate scenario with a temperature 5 K warmer than now and an accumulation rate three times higher than now (an extreme scenario for the Holocene climatic optimum), gives firn temperatures in the central region of the ice cap a few degrees below the freezing point. This indicates that Hans Tausen Iskappe was a cold glacier during its lifetime. Lower surface elevations, either during the period of deterioration of the ice-age ice cap or during the build-up of the existing ice cap, may have further increased the firn temperature. Also, geothermal heat flow causes higher temperature at the base of the ice cap than at the surface. Therefore, we cannot exclude the possibility that some regions of Hans Tausen Iskappe – particularly the high accumulation region to the north – were temperate or experienced basal melting during certain periods of the Holocene climatic optimum.

Acknowledgement

This research was supported by *The Nordic Environmental Research Programme 1993-1997* of the *Nordic Council of Ministers*. Comments by N. Tvis Knudsen and an anonymous reviewer led to improvements of the paper. Results based on GEUS material is published with permission of the Geological Survey of Denmark and Greenland.

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