Stable isotope studies on ice margins in the Thule area

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The main purpose of the glaciological programme carried out by the NORDQUA 86 expedition to the Thule area, Northwest Greenland was to investigate to what extent stable isotope studies in the marginal areas of the Inland ice and local ice caps could provide information about the dynamic and climatic history of the Northwest Greenland ice cover during the last glacial/interglacial cycle.

The expectations that stable isotope studies on the ice margins could actually contribute such information were based on results of previous isotope studies at other Greenland ice-margin locations. Such studies have documented that ice, deposited in the inland regions of the ice sheet during the Weichselian glaciation (here considered equivalent to the Wisconsinan of North America), is now exposed at the surface of the ice sheet in a several hundred metre wide band near the ice edge (Reeh *et al.* 1987a). Also, studies on Barnes ice cap, Arctic Canada (72°W, 69.5°N) have revealed that ice of Wisconsinan origin is still present at the margins of this relatively small ice cap (Hooke & Clausen 1982).

The transition from the Weichselian to the Holocene is characterised by a large step-like change in $\delta^{18}O$ from large negative values in the Weichselian to less negative values in the Holocene. This has been found in deep ice-core records from the inland regions of the Greenland ice sheet (e.g Dansgaard *et al.* 1984), as well as in ice-margin records (e.g. Hooke & Clausen 1982, Reeh *et al.* 1987a, Clausen & Stauffer 1988). Even though there are other sources for the variations found in $\delta^{18}O$ records (Fisher & Alt 1985), $\delta^{18}O$ variations on the Greenland ice sheet are believed to be mainly controled by the condensation temperature at the snow-deposition site (Dansgaard *et al.* 1973). The size of the $\delta^{18}O$ shift at the Weichselian/Holocene transition, therefore, not only depends on the magnitude of the climatic temperature change from glacial to interglacial conditions, but also on the change of surface elevation at the snow-deposition site. A comparison of the $\delta^{18}O$ values at the margin of the local ice cap at Tuto Ramp (see map in Fig. 27) with the $\delta^{18}O$ values at the margin of the Greenland ice sheet at Nuna Ramp, and with the $\delta^{18}O$ values of the deep core-ice from Camp Century located *c.* 150 km inland of the ice margins, could therefore reveal possible changes in the dynamics of the Northwest Greenland ice cover at the end of the Weichselian.

However, these expectations were not fulfilled, because only ice of Holocene origin was found at the ice margins in the Thule area. Therefore, the isotope studies can only provide information about the present geographical $\delta^{18}O$ distribution in the Thule area, and to some extent throw light on the inland deposition site for the ice at the margins, and on the thermal conditions at the base of the ice covers.

The results of the isotope analysis are also used to derive input data to ice dynamic models for the ice margins, and to check the results of model calculations. The calculations indicate that a few hundred metres from the ice margins, the surface ice is 2000–3000 years old, and, therefore, that ice of Weichselian origin, if present at all, can at most constitute only a narrow band near the margin.

Calculations also show that, very likely, a significant part of the Weichselian ice has been melted away at the bottom of the ice covers or has simply been removed by increased surface ablation during the Holocene climatic optimum. It is even possible that increased ablation during this period could have resulted in complete obliteration of local ice caps in the Thule area.

Glaciological setting

The ice margins in the Thule area have been intensively studied in the 1950's and 1960's by U.S. Army Snow Ice and Permafrost Research Establishment (SIPRE) and U.S. Army Cold Regions Research & Engineering Laboratory (CRREL). The glaciological investigations comprised observations of mass balance, ice temperature, surface-ice velocity and deformation, and investigations in two 400 m tunnels excavated into the ice margin at the Tuto Ramp. The results were published in SIPRE and CRREL reports, e.g. Schytt (1955) and Nobles (1960). A more recent study was carried out by Hooke (1970).

The Tuto Ramp is situated *c.* 20 km southeast of Thule Air Force Base (see map in Fig. 27). From the ice margin at an elevation of *c.* 500 m, the surface displays a gentle, upward slope towards a local dome located *c.* 10 km east of the ice margin at an elevation of 900–1000 m (Figs 29 and 30a). The dome is connected to the main Greenland ice sheet by a saddle and a more than 30 km long, undulating ridge with elevations between 700 and 850 m (Benson 1962). Therefore, at present, the local dome is dynamically independent of the main ice sheet, and ice from there cannot reach the Tuto Ramp area. This must have been the case as long as the "Tuto ice dome" has been separated from the main ice sheet by a saddle.

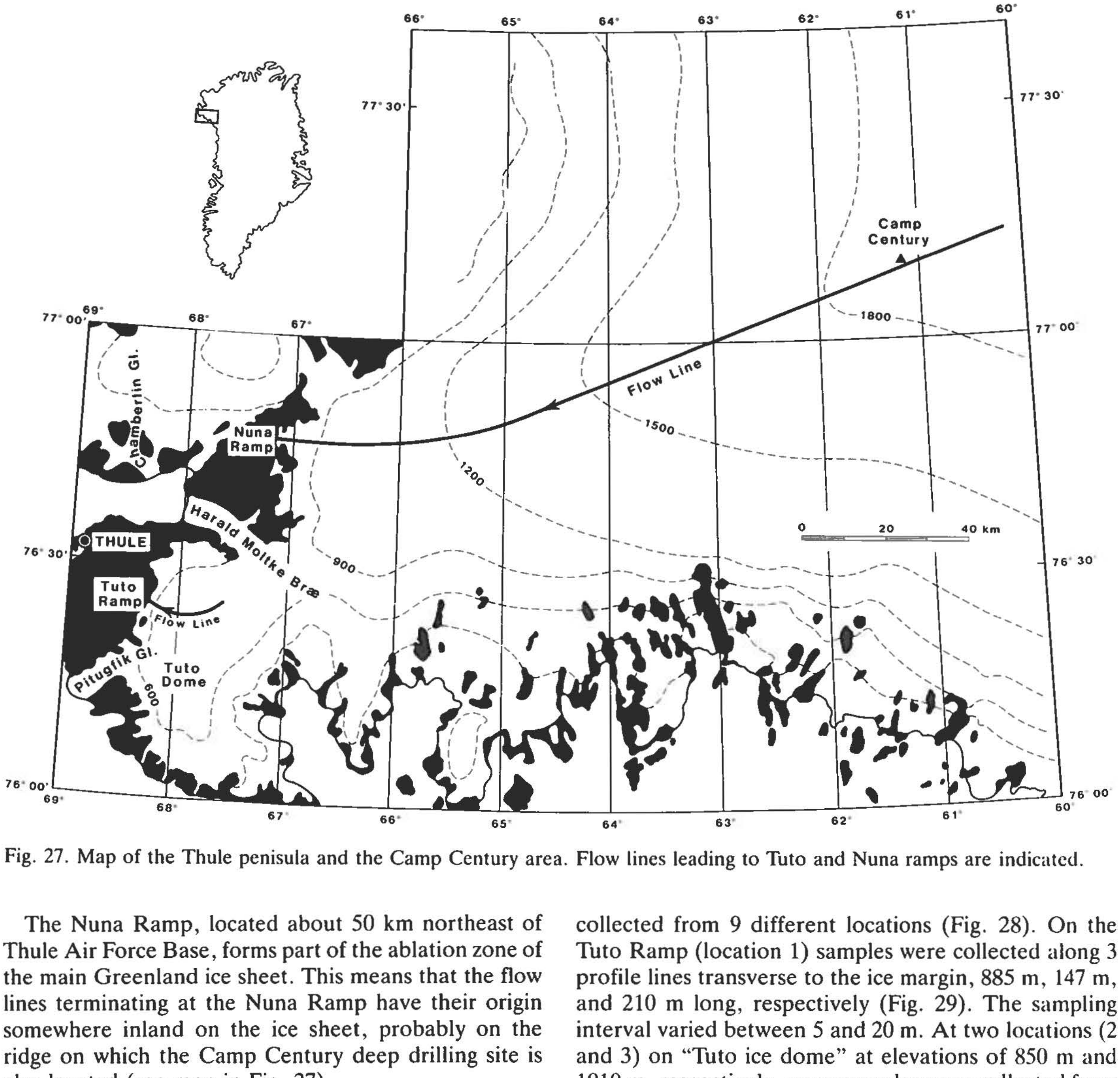


Fig. 27. Map of the Thule penisula and the Camp Century area. Flow lines leading to Tuto and Nuna ramps are indicated.

The Nuna Ramp, located about 50 km northeast of Thule Air Force Base, forms part of the ablation zone of the main Greenland ice sheet. This means that the flow lines terminating at the Nuna Ramp have their origin somewhere inland on the ice sheet, probably on the ridge on which the Camp Century deep drilling site is also located (see map in Fig. 27).

Therefore, the ice at the two ramps should originate from very different locations, and this must have been the case for as long as the present flow pattern has remained essentially unchanged. However, if the main ice sheet, for example during a stage of more extensive glaciation, had overridden the "Tuto ice dome" then the ice deposited as snow during that period would have originated further inland at a high elevation and consequently would carry a relatively low $\delta^{18}0$ value.

As already mentioned, only ice of Holocene origin was found at the surfaces of the ramps. This also applies to the other sampling locations in the area. Possible explanations are discussed in a later section.

Sampling programme

More than 450 samples of water, ice and snow were collected from 9 different locations (Fig. 28). On the Tuto Ramp (location 1) samples were collected along 3 profile lines transverse to the ice margin, 885 m, 147 m, and 210 m long, respectively (Fig. 29). The sampling interval varied between 5 and 20 m. At two locations (2 and 3) on "Tuto ice dome" at elevations of 850 m and 1010 m, respectively, snow samples were collected from 2 m deep snow pits, to provide a present-day reference $\delta^{18}0$ value from the accumulation area feeding the Tuto Ramp.

Some 60 samples were collected from the bottom to the top of an 11 m, near-vertical ice cliff at the southern part of Store Landgletscher (location 4). The ice cliff had a layered structure with numerous dirt bands and dirt inclusions, supporting the interpretation that the cliff consists of ice layers formed by refreesing of meltwater at the base of the ice cap at some distance behind the ice margin. At the same location, ice samples were collected along a 500 m profile at the ice surface leading to the cliff.

Further ice samples were collected from a local piedmont glacier at the northern margin of Store Landgletscher (location 5), from two profiles, respectively 110 m and 75 m long, transverse to the ice margin at the

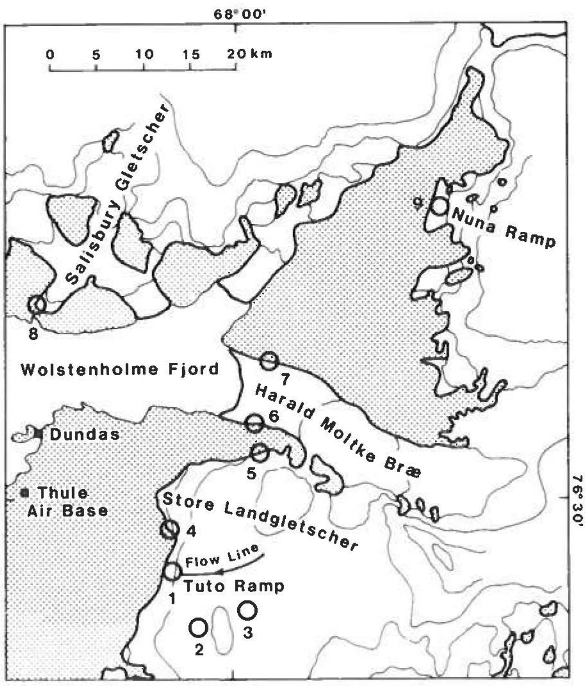


Fig. 28. Location map of the Thule area. Sample locations are marked with circles. Flow line leading to Tuto Ramp is indicated.

southern and northern part of the calf-ice producing outlet glacier, Harald Moltke Bræ (locations 6 and 7), and from the moraine covered ice in front of Salisbury Gletscher (location 8).

Finally, samples were collected along a profile, 1001 m long, transverse to the ice margin at Nuna Ramp at intervals between 4 and 20 m (location 9).

The summer of 1986 was unusually cold in the Thule area, and in the period of sampling around mid August, even the lowermost parts of the ice margins were still covered by up to 0.5 m of snow from the previous winter's accumulation. This impeded the sampling enormously, since the ice surface had to be cleared for snow before the samples could be taken. In fact, the deep snow cover prevented sampling of a 200–300 m section of the surface nearest to the margin. Also, due to the presence of the previous winter's snow, one cannot completely exclude the possibility that some samples were taken from a layer of superimposed ice below the snow cover, even though the upper ice layer was cut away down to the depth of the cryochonite horizon before the samples were taken. According to Schytt (1955), this should ensure that real glacier ice was reached.

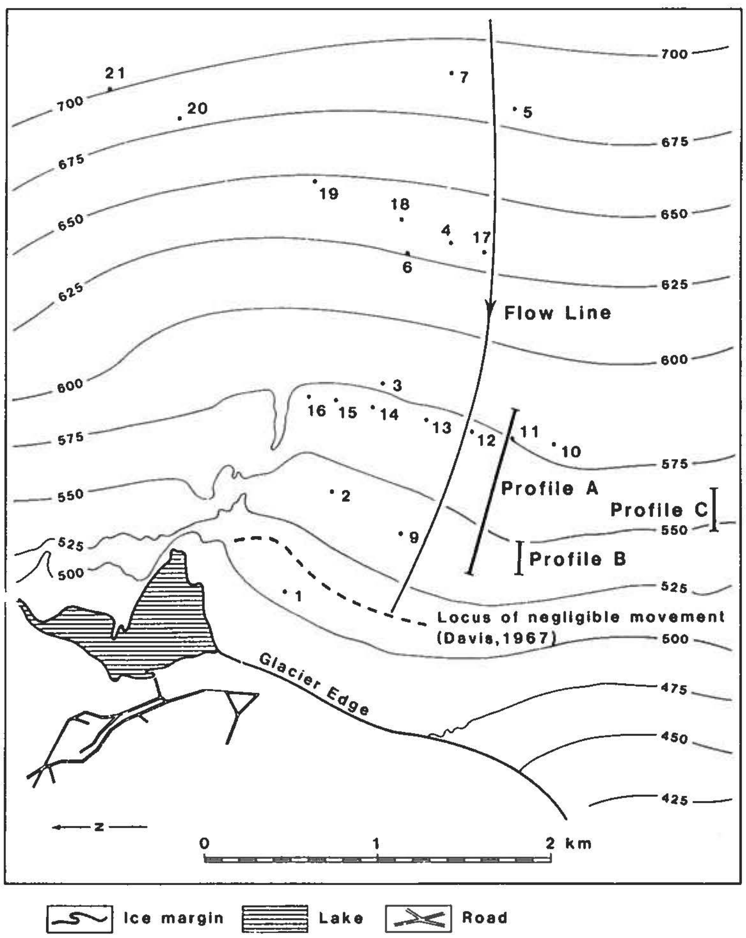


Fig. 29. Map of the Tuto Ramp. Surface elevations are in metres. Points marked by numbers are mass balance and velocity points. Locations of $\delta^{18}O$ sampling profiles are indicated. Modified from Schytt (1955).

Geographical $\delta^{18}O$-distribution

The results of the $\delta^{18}O$ analysis are summarised in Table 11. Disregarding the snow samples from profile A, which are not representative of the annual precipitation, it appears from the table that the range of mean $\delta^{18}O$ values is from −18.4 to −24.0 0/00. Moreover, if the $\delta^{18}O$ means from Nuna Ramp and the northern margin of Harald Moltke Bræ are left out, the range is even narrower, i.e. from −18.4 to −21.7 0/00. The ice with $\delta^{18}O$ means in this range can most reasonably be referred to an origin in the accumulation areas of the local ice caps around Wolstenholme Fjord. This interpretation is supported by the $\delta^{18}O$ mean values of -20.3 ± 0.8 0/00 and -20.5 ± 1.0 0/00 for the samples from the two snow pits on the "Tuto ice dome", which both fall within this range.

The mean $\delta^{18}O$ values from Nuna Ramp and the northern margin of Harald Moltke Bræ (−23.0 and −24.0 0/00) are lower than the rest of the mean $\delta^{18}O$ values by 2.5–3.5 0/00. However, these values are certainly not low enough to be interpreted as ice-age values (ice age $\delta^{18}O$ values from the deep Camp Century record are in the range from −38 to −42 0/00, i.e. 8.5–12.5 0/00 lower than the present-day $\delta^{18}O$ value at Camp Century). The ice, carrying $\delta^{18}O$ values around −23 to −24 0/00, is more likely to originate from the lower part of the accumulation area of the main Greenland ice sheet.

As regards Harald Moltke Bræ, the different origin of the ice at the southern and northern margins indicated by the different $\delta^{18}O$ values, is in accordance with the flow pattern inferred from the surface-elevation contours of the ice-sheet sector draining to Harald Moltke Bræ: The southern part is fed by ice originating at the local dome to the northeast of Store Landgletscher (see map in Fig. 2), whereas, the northern (and also the central) part is fed by ice from the main ice sheet.

Also in terms of variability (expressed as the root-mean-square (rms) value), there are characteristic differences between the various $\delta^{18}O$ records. The snow-pit

Table 11. Mean, standard deviation, and variability of $\delta^{18}O$ samples.

Location	Location No.	Mean $\delta^{18}O$ (0/00)	Standard deviation of mean (0/00)	Variability (r.m.s.) (0/00)	Number of samples
Tuto Ramp	1				
Surface ice (Prof. A)		-19.12	0.10	0.90	81
Snow (Prof. A)		-17.09			
Snow (Prof. A)		-34.70			
Snow (Prof. A)		-30.36			
Surface ice (Prof. B)		-18.41	0.15	0.90	36
Surface ice (Prof. C)		-19.42	0.21	1.07	26
Water from Lake Tuto		-20.34	0.13	0.18	2
Tuto ice dome	2				
Surface snow		-24.37	0.46	0.92	4
Snow pit		-20.52	1.01	3.19	10
Tuto ice dome	3				
Surface snow		-25.36	0.47	0.94	4
Snow pit		-20.34	0.81	2.56	10
Store Landgletscher	4				
Ice cliff		-18.77	0.05	0.37	55
Surface ice		-20.06	0.27	0.97	13
Piedmont glacier	5				
Surface ice		-20.85	0.18	0.31	3
Harald Moltke Bræ					
Southern margin	6				
Surface ice		-20.97	0.13	0.64	24
Northern margin	7				
Surface ice		-24.00	0.27	1.12	17
Salisbury Glacier	8				
Moraine covered ice		-21.65	0.11	0.27	6
Nuna Ramp	9				
Surface ice		-23.02	0.14	1.36	95

records display the highest variabilities of 3.2 and 2.6 0/00, respectively. Then follows in decreasing order the Nuna Ramp and the northern Harald Moltke Bræ records with rms values of 1.36 and 1.12, respectively. The rms values of the $\delta^{18}O$ records inferred to be of local origin are all around 0.9, except for the rms values from the ice-cliff record from Store Landgletscher, and the records from the piedmont glacier and Salisbury glacier which are around 0.3–0.4 0/00. Due to the few degrees of freedom the rms value from the piedmont glacier is encumbered with a large uncertainty, and therefore the low rms value of *c.* 0.3 0/00 is not significantly different (F-test) from the rms value of about 0.9 0/00 for the other local $\delta^{18}O$ records. The records from the ice cliff and Salisbury glacier, on the contrary, have rms values of 0.37 0/00 (54 degrees of freedom) and 0.27 0/00 (5 degrees of freedom) which are significantly lower than the rms value of the other local $\delta^{18}O$ records.

The differences of the rms values can be explained as follows: In the snow-pit records derived from the uppermost firn layers, a relatively large part of the annual $\delta^{18}O$ cycle is still preserved. As the layers sink into the ice cap under continuous thinning, the $\delta^{18}O$ variations are smoothed by diffusion (Johnsen 1977), and probably also due to melting-refreezing processes. When the ice resurfaces in the ablation zone at the ice-sheet margin, it therefore displays lower $\delta^{18}O$ variations than do the upper firn layers.

The tendency for the $\delta^{18}O$ records from Nuna Ramp and the northern margin of Harald Moltke Bræ to have slightly larger rms values than the records of local origin, could be due to the fact that the ice at these locations originates from higher elevated and consequently colder areas than does the local ice.

Consequently, the annual $\delta^{18}O$ cycles are better preserved in this ice, resulting in higher variability of the $\delta^{18}O$ records. This is in agreement with previous findings of ice-margin studies in central West Greenland (Reeh *et al.* 1987b).

The low variability of the $\delta^{18}O$ records from the ice cliff and Salisbury glacier support the hypothesis that the ice at these locations has undergone melting-refreezing, probably at the glacier bed, causing additional smoothing of the $\delta^{18}O$ variations.

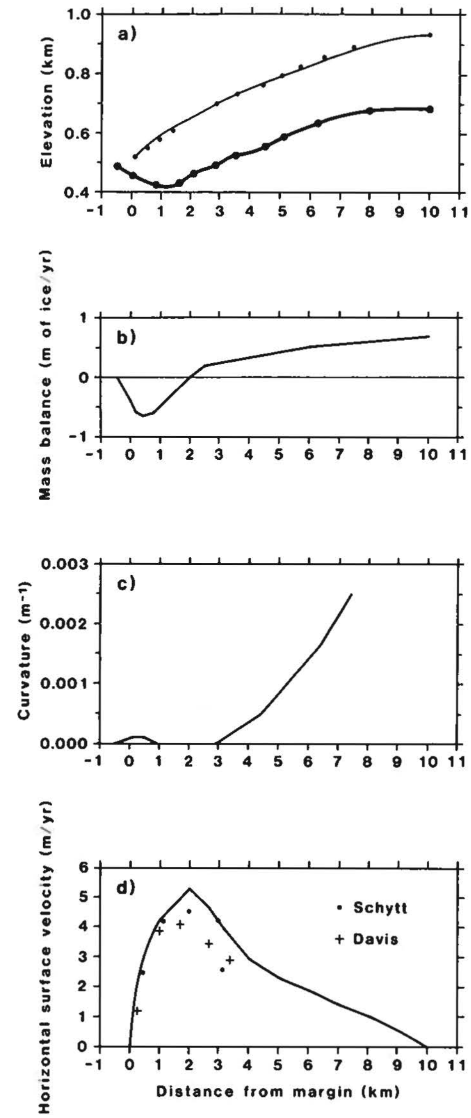


Fig. 30. Input data and results of ice-dynamic calculation for Tuto Ramp. a) shows observed base elevations (dots connected by thick curve), observed surface elevations (points marked by dots), and calculated surface elevation profile (thin curve). Observations based on data from Schytt (1955), Bishop (1957), Benson (1962), and Davis (1967). b) shows observed mass balance along flow line. Data from Schytt (1955) and Griffiths (1960). c) Curvature of elevation contours at their point of intersection with the flow line. Data from Schytt (1955) and Bishop (1957). d) Observed (points marked by dots and crosses) and calculated (curve) horizontal surface velocities along the flow line.

Ice-dynamic model for Tuto Ramp

Further interpretation of the $\delta^{18}O$ records depends on the possibility of dating them. This is possible for the long $\delta^{18}O$ record from Tuto Ramp, since information is available about ice-cap geometry, mass balance, and dynamics. This information can be used to set up an ice-dynamic model for theoretical dating of the ice exposed at the surface of the ramp. The model applied is the flow-line model described by Reeh (1988), which has previously been applied to date the surface ice at an ice-margin location in central West Greenland (Reeh *et al.* 1987b).

Ice cap geometry. – Surface-elevation maps of the Tuto Ramp area have been published by Schytt (1955) (see Fig. 29) and Davis (1967), and surface- and base-elevation maps of part of the "Tuto ice dome" are shown by Bishop (1957). Further surface-elevation data are given by Benson (1962) and in this work. Based on this information, the flow line shown in Figs 28 and 29, the curvature of the surface-elevation contours at their intersection with the flow line (Fig. 30c), and the observed surface- and base-elevation profiles indicated in Fig. 30a have been determined.

Mass balance. – The mass-balance distribution shown in Fig. 30b and the observed surface velocities shown in Fig. 30d are based on data from Schytt (1955), Griffiths (1960), and Davis (1967). Whereas the velocities measured in the different studies are very similar, the ablation rates measured by Schytt (1955) for the balance year 1953/54 are much higher than those observed by Griffiths (1960) for the balance year 1955/56. Griffiths' ablation data are nearly in balance with the velocity data, indicating that the Tuto Ramp was close to a balanced state in 1955/56. Schytt's ablation data, on the other hand, indicate that Tuto Ramp had a large negative balance in 1953/54. Since Griffiths' ablation-rate data are nearly in balance with the observed velocities, these data are more likely to represent the mass balance over a longer period, and, consequently, Griffiths' data are used in the modelling.

Ice temperatures. – Schytt (1955) also published observed temperatures at 10 m depth at one location in the ablation zone of the Tuto Ramp, and at two locations in the accumulation area further inland at elevations of 800 and 850 m respectively. Whereas the 10-metre temperature from the site in the ablation area is around $-13°$ C, i.e. close to the mean annual air temperature (Ohmura 1987), the temperature at the sites in the accumulation area are *c.* $-3°$ C and $-5°$ C, respectively, i.e. up to 10° C higher than the mean annual air temperature. The reason for these high firn temperatures is the release of latent heat by refreezing of percolating melt water formed at the surface during the summer period (Schytt 1955).

With such high firn temperatures combined with an ice thickness, of *c.* 200 m and an accumulation rate of *c.* 0.5 m/yr, a simple temperature-profile calculation indicates that the ice-sheet base is at the melting point in the accumulation area and probably also in the upper part of the ablation area, thus creating conditions for basal melting. In the lower part of the ablation area the cooling of the near-surface layers will eventually penetrate to the base, and thus create conditions for refreezing of the basal melt water formed further upstream. Temperatures measured in the wall of an ice tunnel excavated into the near-bottom part at the margin of the Tuto Ramp glacier were in the range from -5.5 to $-9.5°$ C (Rausch 1958), thus confirming that sub-freezing temperatures exist in the basal ice near the margin.

Results. – Measured and predicted surface profiles are shown in Fig. 30a. Discrepancies are small. The fit is obtained with a flow-law parameter of A ($-5°$ C)= 4.6 10^{-16} kPa^{-3} s^{-1}, which is about half the minimum value recommended by Paterson & Budd (1982). In view of

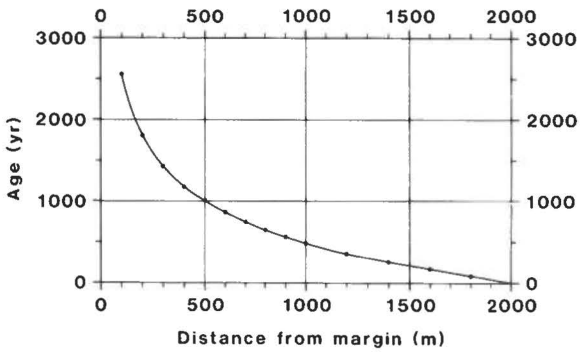


Fig. 31. Calculated age of surface ice along the near-margin part of the flow line on Tuto Ramp.

the uncertainty of the input data to the model, and the large variation in measured flow-law parameters, the disagreement between the predicted value and the recommended value is not discouraging.

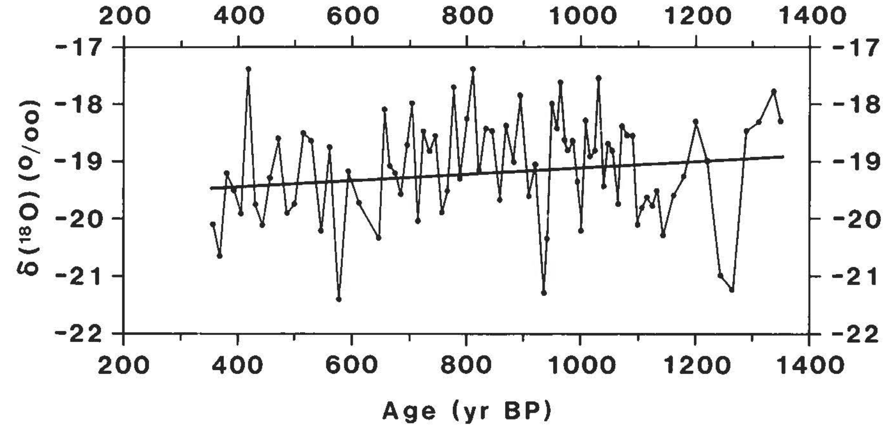
Fig. 30d shows measured and predicted horizontal surface velocities. The variations along the flow line of measured and predicted velocities are similar, but the model overestimates the velocities between 2 and 3.5 km from the margin by about 25%. Unfortunately, observations of velocities are not available further inland than 3.5 km, but it is unlikely that the model predictions deviate more than 25%, from the actual velocities anywhere along the flow line. This uncertainty includes possible velocity fluctuations due to temporal mass-balance variations which are not accounted for by the steady-state ice-dynamic model. Therefore, the calculated ages of the surface ice displayed in Fig. 31 should be accurate to within ±25%, with higher ages being more likely than younger ages.

It appears from Fig. 31 that the age of the surface ice is around 2500 years at a distance from the margin of 100 m. (The ice margin is defined as the locus of negligible (horizontal) movement as observed by Davis (1967)). This indicates that ice of Weichselian origin (older than 11 000 years), if present at all at the surface, extends for at most some tens of metres inwards from the margin. Unfortunately, this part of the surface was covered by deep snow in the summer of 1986 which prevented sampling of ice.

The calculated ages can be compared with radiocarbon datings performed on carbon dioxide extracted from air inclusions in ice from a tunnel excavated in the near-bottom layers of the ice cap close to Tuto Ramp. Radiocarbon ages were *c.* 3000 and 5000 yr, respectively, for ice taken 300 and 200 m from the tunnel entrance (Oeschger *et al.* 1966). Taking into account that the age of the ice increases with increasing distance below the surface, these ages are compatible with the calculated ages of the surface ice.

Fig. 32 shows the $\delta^{18}O$ record from Tuto Ramp plotted against age BP. The time scale is the one given by the distance-age diagram of Fig. 31. The interval covered by

Fig. 32. $\delta^{18}0$ profile from the surface of Tuto Ramp. Age scale is based on the age-distance relationship of Fig. 31. Thick line indicates the linear trend.



the surface $\delta^{18}0$ record is *c.* 350–1350 years BP. The ages can be considered accurate to within ±25%. It might be worth noting that the positive trend in $\delta^{18}0$ with increasing age (0.5 o/oo in 1000 years), though not statistically significant, is compatible with a trend of 0.8 o/oo during the same period indicated by the $\delta^{18}0$ record from Camp Century (Robin 1983).

Age estimate for the Nuna Ramp $\delta^{18}0$ record

The data coverage upstream of Nuna Ramp is not good enough to justify application of advanced ice-dynamic models for dating the ice. Some simpler considerations will be applied.

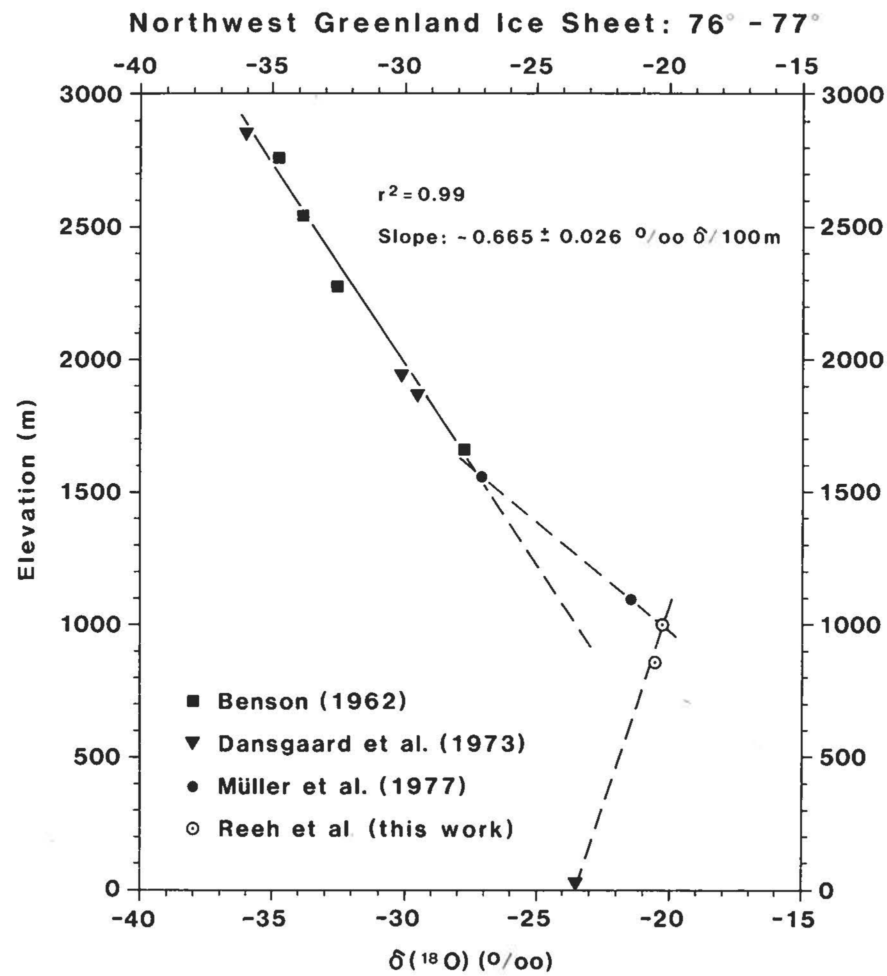


Fig. 33. δ^{18}O-elevation relationship for Northwest Greenland between 76 and 77°N.

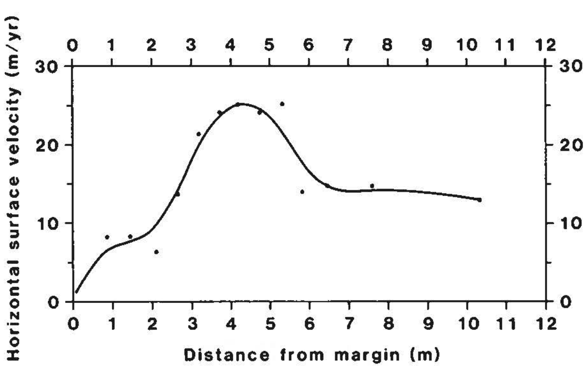


Fig. 34. Observed horizontal surface velocities along the Nuna Ramp flow line (points marked by dots). The curve represents the smoothed velocity distribution used in the estimate of the age of the surface ice.

By means of the $\delta^{18}O$ elevation diagram for the northwest Greenland sector between 76° and 77°N (Fig. 33), the elevation of the deposition site for the snow forming the ice of the sampling location on Nuna Ramp ($\delta^{18}O$ = -23.0 0/00) is estimated to be *c.* 1250 m. According to data published by Nobles (1960) this elevation is reached on the Greenland ice sheet at a distance of *c.* 25 km from the margin of Nuna Ramp. From the margin and 10 km inland, horizontal surface velocities have been measured by Nobles (1960) (Fig. 34), which can be used to estimate the time used by an ice particle to travel from 10 km upstream of the margin to the sampling profile located 300 to 1300 m from the margin. This time has been estimated as 800 ± 50 years. Between 10 and 25 km upstream from the margin, velocities are not known, but most likely the velocities decrease going inland. A reasonable estimate of the surface velocity 25 km inland is 5 m/year with estimated error limits of 2 and 13 m/year, the latter value being the surface velocity measured *c.* 10 km from the margin. Assuming a linear velocity variation along the 10 – 25 km section results in an estimated travel time along this section of 1800 yr with upper and lower limits of 2500 and 1150 yr respectively. Adding up, the age of the ice at the sampling location is estimated to be around 2500 yr with upper and lower limits of 3500 and 2000 yr, respectively. As shown in Fig. 35, the profile starts about 300 m from the ice margin, and therefore it can be concluded that ice of Weichselian origin if present at all at the surface of the Nuna Ramp, is limited to a narrow band near the margin, probably less than 100 m wide. Unfortunately, deep snow and slush prevented sampling of ice from this part of the surface in the summer of 1986.

The $\delta^{18}O$ record from Store Landgletscher

The $\delta^{18}O$ record from the ice cliff of Store Landgletscher is shown in Fig. 36. It is not possible to give even a rough estimate for the age of the ice in the cliff, if only for the reason that the cliff has probably been formed by freeze-on at the base of melt water produced by basal melting further inland. The layered structure of the ice cliff with numerous dirt layers and inclusions, and the low variability of the $\delta^{18}O$ record point to this conclusion. Furthermore, it was shown in a previous section that the thermal conditions required for melting/refreezing to take place are in fact present at the ice-cap base. Comparison of the $\delta^{18}O$ values from the ice-cliff samples with the mean $\delta^{18}O$ value of the samples collected on the near-horizontal glacier surface upstream of the cliff provides further support for the freeze-on hypothesis. The mean $\delta^{18}O$ value of the clean debris-free surface ice which exibits a sharp contact to the ice-cliff ice, is significantly lower than the ice-cliff $\delta^{18}O$ values. The explanation could be that the $\delta^{18}O$ values from the ice cliff are increased due to fractionation during the freeze-on process (Jouzel & Souchez 1982).

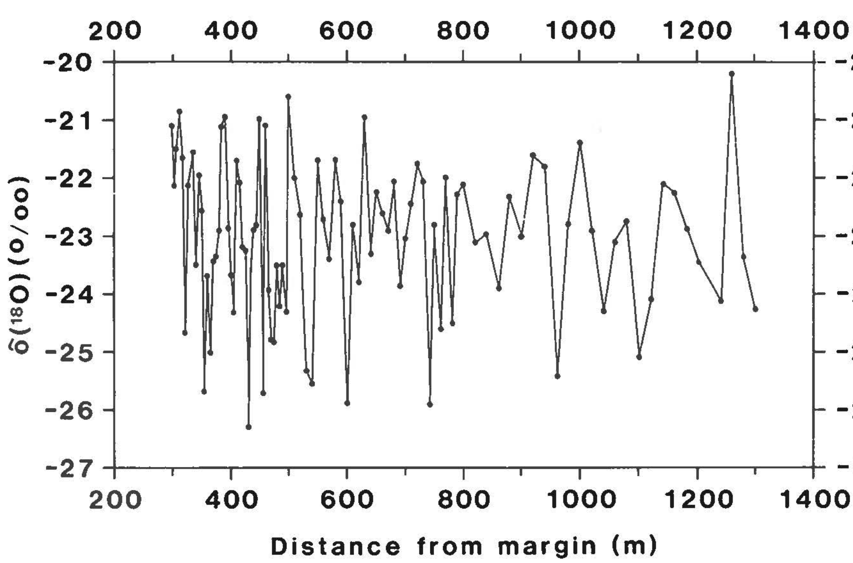


Fig. 35. $\delta^{18}O$ profile from the surface of Nuna Ramp.

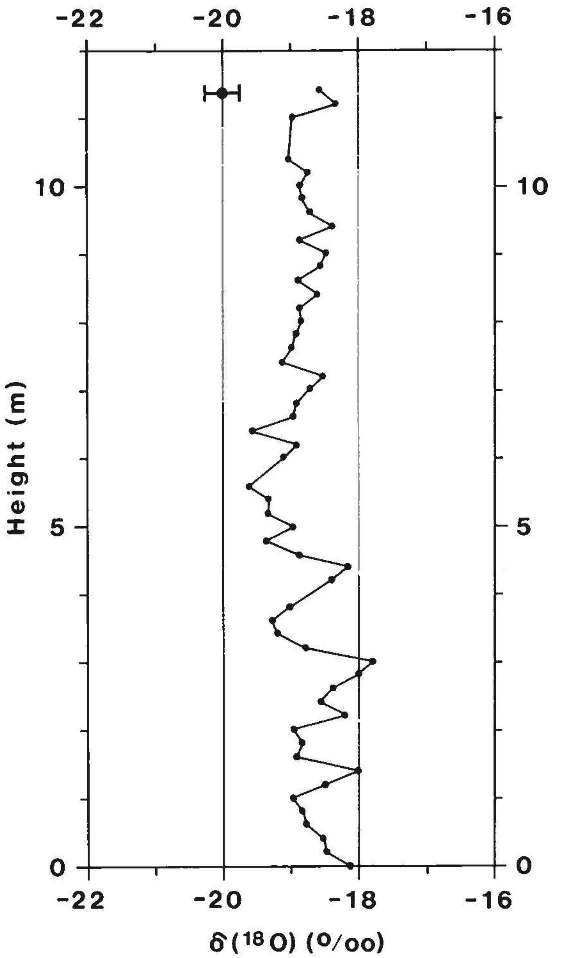


Fig. 36. $\delta^{18}O$ profile from the ice cliff of Store Landgletscher. The mean $\delta^{18}O$ value for the surface ice above the cliff with error limits (1 standard deviation) is also indicated.

In connection with the present discussion of the origin of the ice cliff, a comment is appropriate about the formation of the moraines on, and along, the ice-cap margin, even though these moraines were not studied during our expedition. Bishop (1957) and Swinzow (1962) argued that the moraines were formed by shear deformations in the ice-cap margins. This explanation, however, was opposed by Weertman (1961) and Hooke (1970), who considered a freeze-on process to be a more likely mechanism for incorporating debris in the basal ice layers. The present study can be taken as support for the Weertman-Hooke point of view, since it provides evidence that basal freeze-on of ice is likely to take place near the ice margins.

The $\delta^{18}O$-elevation relationship

A plot of mean $\delta^{18}O$ values from the upper firn layers on the ice sheet against elevation (E) (see Fig. 33) seems to distinguish between three different elevation ranges, each with its particular $\delta^{18}O$-elevation relationship. Above 1500 m the $\delta^{18}O$-elevation relationship is closely approximated by the regression equation (r = 0.99)

$$\delta^{18}O = -31.4 - 0.00665\,(E - 2180)$$

The $\delta^{18}O$-elevation gradient indicated by this equation (-0.665 ±0.026 o/oo per 100 m) is in good agreement with the value of -0.66 o/oo per 100 m found by Dansgaard *et al.* (1973) for locations above 1500 m on the ice sheet in central West Greenland.

Below an elevation of 1000 m the data indicate a small, but positive, $\delta^{18}O$-elevation gradient of *c.* 0.3 o/oo per 100 m. Similar small, positive gradients were found by Koerner (1979) for the same elevation range on high-arctic Canadian ice caps. Koerner (1979) explained the small gradient in terms of precipitation from a single condensation level. Reversed $\delta^{18}O$-elevation gradients were also found in Antarctica below elevations of *c.* 1000 m (Lorius & Merlivat 1977).

Between 1000 and 1500 m elevation, the data indicate a negative gradient about twice the value found above 1500 m. It is tempting to interpret this higher-than-normal gradient as being caused by fractionation in connection with melting/refreezing processes in the upper firn layers (Arnason 1969). The lower the location in the accumulation area, the larger is the fraction of melt water that will escape by runoff, and the more enriched in heavy isotopes will be the remaining refrozen melt.

Comparison of mean annual air temperatures and ten-metre firn temperatures from the Northwest Greenland ice sheet (to be presented elsewhere) supports this hypothesis: Above 1500 m there is good agreement between the two temperatures. Below 1500 m, however, the ten-metre firn temperature deviates more and more from the mean annual air temperature with decreasing elevation, indicating increasing warming due to refreezing of melt water formed at the surface in the summer period, in accordance with results previously presented by Hooke *et al.* (1983) and Ohmura (1987).

Discussion and conclusions

Even though the main purpose of the NORDQUA 86 glaciological programme was not achieved because ice of Weichselian origin was not found at the investigated ice margins in the Thule area, the study demonstrated the usefulness of applying stable isotope methods in glaciological studies of ice margins. The isotopic signature of the ice, in terms of mean value as well as variability, is characteristic for the site of formation of the ice and for its thermal and phase-change history since the time of formation.

The model calculations of the ice flow of the Tuto and Nuna ramps indicate that ice of Weichselian origin, if present at all, is limited to narrow bands near the margins, probably less than some tens of metres wide. However, basal melting, which is likely to occur upstream of Tuto Ramp and probably also upstream of Nuna Ramp, may, partly or fully, have removed the Weichselian ice. For Tuto Ramp the effect of basal melting on the age of the ice can be estimated by means of Fig. 32 in a paper by Reeh (1989) which shows calculated ages at the base of an ice sheet as a function of the ratio of basal melt rate to surface-accumulation rate. For "Tuto ice dome", the basal melt rate is probably only a fraction of the value (0.5 cm/yr) corresponding to the case of a temperate glacier subject to an average geothermal heat flux, since part of the heat flows to the upper surface, where the mean annual snow temperature is a few degrees below the freezing point. For a basal melt rate of 0.1 cm/yr, for example, the maximum age of the ice is calculated to be *c.* 14 000 yr, whereas the maximum possible basal melt rate of 0.5 cm/yr results in a maximum age as young as *c.* 5000 yr. These estimates indicate that the extent of Weichselian ice at the margin of the Tuto Ramp may be even less than indicated by the theoretical dating mentioned previously. Moreover, increased surface melting in the Holocene climatic optimum *c.* 8000–3000 yr B.P. could also have seriously affected the ice margins in the Thule area, and could in fact have threatened the very existence of the local ice caps in this period. Calculations by a climate/mass-balance model (Roger Braithwaite, personal communication) indicate that an estimated 2–3°C warmer summer climate could have resulted in a rise of the equilibrium line on the "Tuto ice dome" from the present elevation of 650 m to an elevation between 800 and 900 m. Moreover, the mass wastage by ablation is likely to have increased by about 1 m/yr in average over the ablation area, which, furthermore, would have been much larger than at present because of the rise in equilibrium-line elevation. The thin ice caps in the Thule area could hardly have survived such conditions for a period of several thousand years.

The effect of a likely increase in snow accumulation is not accounted for in the above estimates. However, a study by Paterson & Waddington (1984) indicates no more than a 10-20 % increase in accumulation rate in the Camp Century area in the period 5000-3000 yr B.P. A similar small increase in snow accumulation on the local ice caps in the Thule area would probably not be sufficient to ensure the survival of these ice caps in a climate with summer temperatures 2–3°C warmer than present.

If the local ice caps in the Thule area suffered such severe losses or maybe even became extinct in the Holocene climatic optimum, why did Barnes ice cap, located *c.* 750 km further to the south on Baffin Island, Canada (69.5°N, 72°W) not suffer the same fate? That Barnes ice cap survived the Holocene climatic optimum is proven beyond doubt by the detection of ice of Wisconsinan origin in deep bore holes as well as on the margins of the ice cap (Hooke & Clausen 1982). The present dimensions and surface elevations of Barnes ice cap are similar to those of "Tuto ice dome", and also the mass balance-elevation relationships are similar for the two ice caps, although with somewhat lower accumulation rates for Barnes ice cap. (The location of Barnes ice cap far to the south of "Tuto ice dome" is nearly compensated, as far as temperatures are concerned, by the much colder climate at the same latitude on the Canadian side than on the Greenland side of Baffin Bay (Wilson 1969)).

Therefore, there is nothing in the present conditions of the two ice caps that can explain why "Tuto ice dome" might have melted away in the Holocene climatic optimum, whereas Barnes ice cap survived. However, the explanation may be found in the very different conditions at the two ice-cap locations at the end of the Weichselian glacial. Whereas the late glacial ice cover in the Thule area was only slightly more extensive than the present (Fig. 10), Baffin Island was in the Early Holocene covered by the thick ice masses of the Laurentide ice sheet. According to Prest (1969), Barnes ice cap still had an extent of 4–5 times its present size even as late as 6000 B.P. This means that whereas "Tuto ice dome" was probably a small ice cap at the time of the onset of the Holocene climatic optimum (8000–9000 B.P.), and therefore from the very beginning was likely to suffer severely due to the warm climate, Barnes ice cap was still part of a large ice mass at that time. Therefore, even though the melting was intense also in the Baffin Island area, there was so much more ice to melt, and Barnes ice cap could survive as a small relic of the Laurentide ice sheet.

Also the margins of the Greenland ice sheet in the Thule area seem to have been affected by the warmer climate in the Holocene climatic optimum, as documented by findings of mollusc shells, radiocarbon dated to between 2650 and 7050 yr BP, in the marginal moraines along Harald Moltke Bræ, Chamberlain Gletscher, and Pitugfik Gletscher (see map in Fig. 28), thus indicating a significant retreat of the ice margins behind their present positions sometime during the Holocene (Kelly 1980b)