Meddelelser om Grønland

A “deep” ice core from East Greenland

Sigfus J. Johnsen, Henrik B. Clausen, Willi Dansgaard, Niels S. Gundestrup, Margareta Hansson, Peter Jonsson, Jørgen P. Steffensen and Arny E. Sveinbjørnsdottir

Geoscience
29 · 1992

Meddelelser om Grønland

The series *Meddelelser om Grønland* was started in 1879 and has since then published results from all fields of research in Greenland. In 1979 it was split into three separate series:

Geoscience
Bioscience
Man & Society

The series should be registered as *Meddelelser om Grønland, Bioscience (Geoscience, Man & Society)* followed by the number of the paper. Example: *Meddr Grønland, Biosci.* 1, 1979.

The new series are issued by Kommissionen for Videnskabelige Undersøgelser i Grønland (The Commission for Scientific Research in Greenland).

Correspondence

All correspondence and manuscripts should be sent to:

The Secretary
Kommissionen for videnskabelige Undersøgelser i Grønland
Danish Polar Center
Hausergade 3
DK-1128 Copenhagen K.

Questions concerning subscription to all three series should be directed to the agent.

Agent

Geografforlaget, Fruerhøjvej 43, DK-5464 Brenderup. Tlf. +45 64 44 16 83.

ISSN 0106-1046
ISBN 87-601-3392-9

Meddelelser om Grønland, Geoscience

Meddelelser om Grønland, Geoscience invites papers that contribute significantly to studies in Greenland within any of the fields of geoscience (physical geography, oceanography, glaciology, general geology, sedimentology, mineralogy, petrology, palaeontology, stratigraphy, tectonics, geophysics, geochemistry). Papers primarily concerned with other areas in the Arctic or Atlantic region may be accepted, if the work actually covers Greenland or is of direct importance to continued research in Greenland. Papers dealing with environmental problems and other borderline studies may be referred to any of the series *Geoscience, Bioscience,* or *Man & Society* according to emphasis and editorial policy.

Scientific editor – Geology

S. Funder, Geological Museum, Øster Voldgade 5-7, DK-1350 Copenhagen K, Denmark. Telephone +45 33 11 42. Telefax +45 33 11 46 37.

Instructions to authors.
See page 3 of cover.

A "deep" ice core from East Greenland

Sigfus J. Johnsen, Henrik B. Clausen, Willi Dansgaard, Niels S. Gundestrup, Margareta Hansson, Peter Jonsson, Jørgen P. Steffensen and Arny E. Sveinbjørnsdottir

MEDDELELSER OM GRØNLAND, GEOSCIENCE 29 · 1992

Contents

Accepted July, 1992
ISSN 0106-1046
ISBN 87-601-3392-9
Printed in Denmark by AiO Print Ltd., Odense

A "deep" ice core from East Greenland

SIGFUS J. JOHNSEN, HENRIK B. CLAUSEN, WILLI DANSGAARD, NIELS S. GUNDESTRUP, MARGARETA HANSSON, PETER JONSSON, JØRGEN P. STEFFENSEN and ARNY E. SVEINBJØRNSDOTTIR

Johnsen, S. J., Clausen, H. B., Dansgaard, W., Gundestrup, N. S., Hansson, M., Jonsson, P., Steffensen, J. P. & Sveinbjørnsdottir, A. E. 1992. A "deep" ice core from East Greenland. Meddr Grønland, Geosci. 29, 22 pp. Copenhagen 1992-11-24.

Pilot studies on the Renland peninsula in Scoresbysund Fjord, East Greenland, indicated that the relatively small and separate Renland ice cap meets most of the criteria defining a favourable ice-core drill-site. In 1988, a Nordic expedition recovered a continuous surface-to-bedrock ice core from the summit. This relatively short core reaches deep into the past, probably throughout the last glaciation and through most of the preceding interglacial, Eem, 125,000 years B.P. The core contains detailed information on temporal changes of the coastal environment, and serves as a valuable complement to the new deep ice cores being drilled in Central Greenland. Core analyses suggest that (1) during Eemian time the East Greenland climate was at least 5°C warmer than now, and the precipitation 20% higher; (2) during the last glacial period, the precipitation decreased to a minimum, perhaps only 20% of the present value; (3) the post-glacial climatic optimum was 2.5°C warmer than now; (4) the long-term variability of the record is relatively low, due to isostatic movements in the area; and (5) from 70,000 years B.P., the Greenland glacial climate alternated between two quasi-stable stages. The latter point may reflect a chaotic feature of climate. If so, climate predictions will be difficult to access.

S. J. Johnsen and A. E. Sveinbjørnsdottir, Science Institute, University of Iceland, IS-107 Reykjavik, Iceland.
H. B. Clausen, W. Dansgaard, N. S. Gundestrup and J. P. Steffensen, Geophysical Institute, University of Copenhagen, Haraldsgade 6, Copenhagen 2200, Denmark.
M. Hansson, Department of Meteorology, Stockholm University, S-10691 Stockholm, Sweden, and Geophysical Institute, University of Copenhagen, Haraldsgade 6, Copenhagen 2200, Denmark.
P. Jonsson, Department of Technical Geology, Technical University of Lund, S-22100 Lund, Sweden.

Introduction

In many respects the polar ice cores so far recovered are complementary to deep sea sediment cores: Analyses of the ice cores give detailed information on temporal changes of the continental environment through a relatively short period of time (up to two glacial cycles), whereas analyses of deep sea sediment cores result in less detailed information on the general oceanic environment through much longer periods of time.

The first deep ice core was drilled at Camp Century, Northwest Greenland (Fig. 1) in 1966 (Hansen & Langway 1966). Analyses for stable isotopes (Dansgaard *et al.* 1969) and other parameters (Langway 1970) showed that the Greenland ice sheet contains a wealth of information on past environmental changes. The early science results from Greenland encouraged the drilling of other ice cores, both in the Arctic and in Antarctica. Until recently, the Vostok core drilled by Soviet in East Antarctica was the longest one obtained in terms of length (2546 m), and most likely in terms of time span (more than 185,000 years; Petit *et al.* 1990).

Fifteen years after Camp Century, the second deep Greenland ice core was drilled at Dye 3 in southeast Greenland under the American-Danish-Swiss Greenland Ice Sheet Program (GISP 1). Dye 3 was chosen as a drill site in preference to the more attractive summit of the ice sheet, because Dye 3 offered low-cost logistics, and because the Danish deep drill, ISTUK (Gundestrup *et al.* 1984), had not yet been tested. The ISTUK reached bedrock at a depth of 2037 m below surface in 1981, and new techniques were applied for processing and analyzing the core, which further confirmed the scientific importance and potential of ice core studies.

Dye 3 was not an ideal deep drill site, however, because (1) it is far from the ice divide and even farther from the summit of the south dome, and (2) the bedrock upstream is hilly. Since the area of deposition is distant, the ice in the deepest part of the Dye 3 core has had a long and complicated travel history over mountaneous areas (Overgaard & Gundestrup 1985). These are the

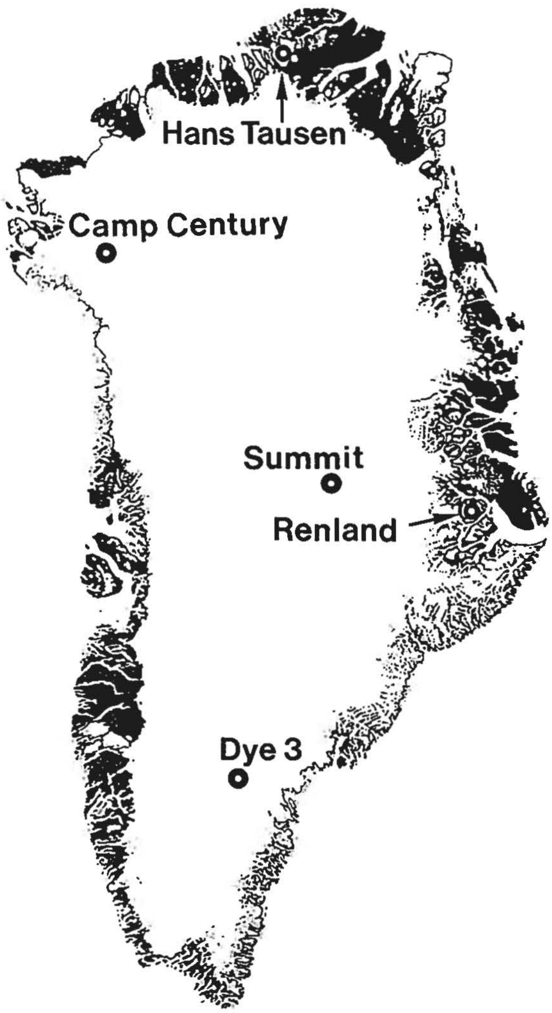


Fig. 1. Map of Greenland with the drill sites mentioned in the text.

reasons why the stratigraphy appears disturbed in the deep ice, and why the Dye 3 ice core represents less than 100,000 years of continuous deposition (Dansgaard *et al.* 1982).

In the summer of 1992, ISTUK reached bedrock 3029 m below surface at Summit, the highest point of the Greenland ice sheet. This was accomplished under the European Greenland Ice-core Project (GRIP). The core seems to span at least two full glacial cycles.

Drill site criteria

Deep ice-core drilling is technically complicated, and requires a comprehensive and expensive logistical support. Heavy ski-equipped airplanes are needed to transport the large amounts of cargo, fuel and all the other kinds of supplies necessary for running a remote, self-sustained field camp with a sufficiently high technical standard.

In 1985, a search was initiated for a shallow drill site, where long-term palaeo-climatic information could be gained by a modest technical and financial investment. A list of criteria was set-up as follows:

1. The drill site should be easily accessible by helicopters or small, ski-equipped airplanes. This excluded the central part of the main ice sheet.
2. In order to avoid disruptive effects in annual strata by seeping meltwater, which absorbs atmospheric soluble compounds, e.g. carbon dioxide or sulfates, there should be little or no summer melting on the surface. This excluded a broad marginal zone of the main ice sheet, as well as small isolated ice caps in south and west Greenland. Experience from other parts of Greenland shows that the mean annual surface temperature should be well below − 15°C.
3. The ice should be thin enough to be penetrated by the existing Danish lightweight shallow drill (Johnsen *et al.* 1980) originally designed for only 100 meter penetration depths. This essentially excluded the rest of the main ice sheet. If the criteria **2.** and **3.** were fulfilled, the bedrock temperature would be well below the pressure melting point. There would thus be no risk that the oldest strata were removed by bottom melting, and the ice close to bedrock might be very old.
4. The accumulation rate should be higher than 23 cm of ice equivalent per year to make it possible to identify annual layers and thereby date the core to considerable depths.
5. On the other hand, the accumulation rate should not be too high, because for a given total thickness of a cold ice cap, very high accumulation rate is associated with drastic thinning of the deep strata and, therefore, unfavourably low percentage of old ice.
6. The bedrock should preferably be flat in an area around the highest point of the ice cap surface. If so, this point should be chosen as a drill site, because the ice movement there is essentially vertical, which makes ice flow modelling relatively simple.

An ice core to bedrock on an ice cap meeting these requirements should render an environmental record through the last 10,000 years or more. This had previously proven possible in projects carried out on minor ice caps in the Canadian Arctic (Fisher 1982) and China (Thompson *et al.* 1989).

Pilot studies

The Renland peninsula is located in the Scoresbysund Fjord, only a few hundred km from airstrips in Mestersvig and Constable Pynt, and has an ice cap that is

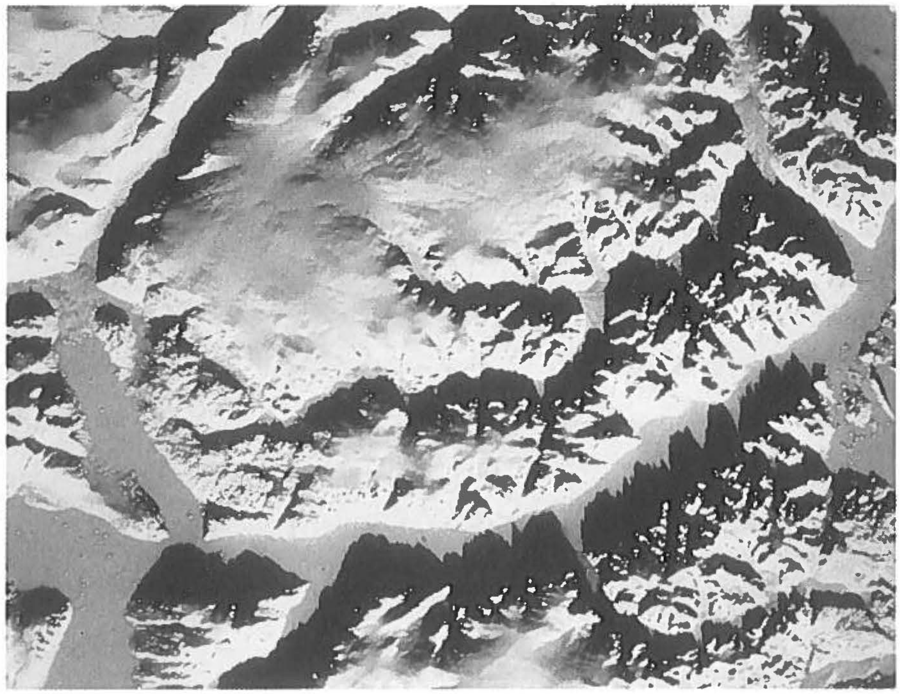
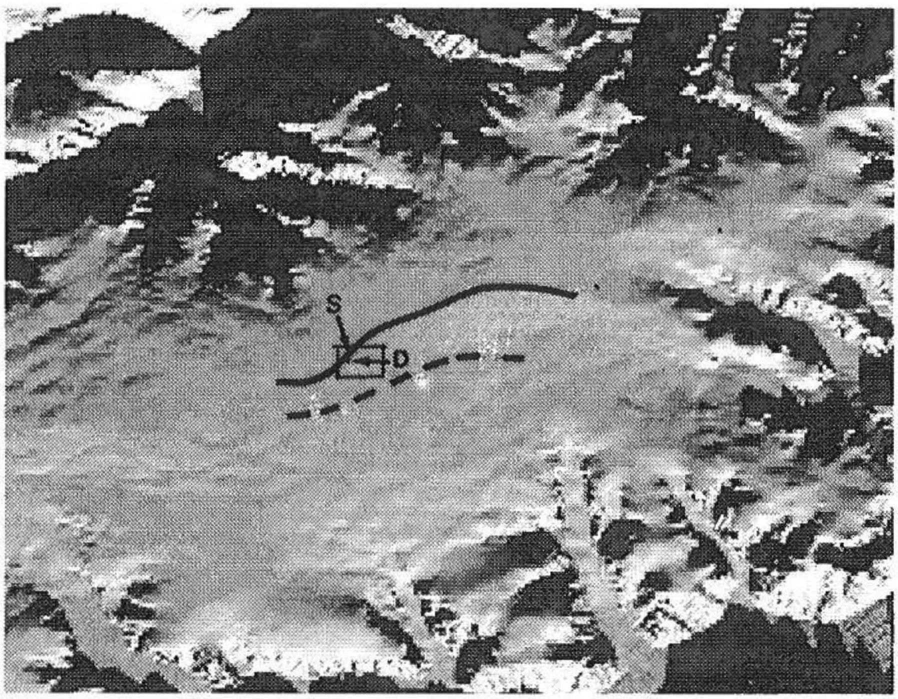


Fig. 2. A: Contrast enhanced Landsat photo showing most of the Renland peninsdula at low sun angle. Notice the mountain shadows on the 10 km wide fiord ice south of Renland.



B: Enlarged section of Fig. 2A showing the northeastern part of the Renland ice cap. The dashed line is a valley and the heavy line is the ice divide: the rectangle is the network area established for radar surveys in 1987 and 1988; and S and D are the highest point and the chosen drill site, respectively.

isolated from the main ice sheet (Fig. 1 and 2). Deep branches of the Scoresbysund Fjord nearly surround the peninsula. They drain the Inland Ice so efficiently that the ice cap may never have been overridden by the Inland Ice. The ice cap has an area of only 1200 km^2, and an ice thickness of only a few hundred meters. It covers the entire high elevation plateau (up to 2340 m a.s.l.) of the peninsula (Fig. 3 and 4), and it could therefore only have been considerably thicker in the past, if overridden by the Inland Ice.

The first pilot study was performed in 1985 with the objectives of mapping the bedrock topography of the

Fig. 3. The Renland ice cap covers the entire high elevation area of the peninsula.

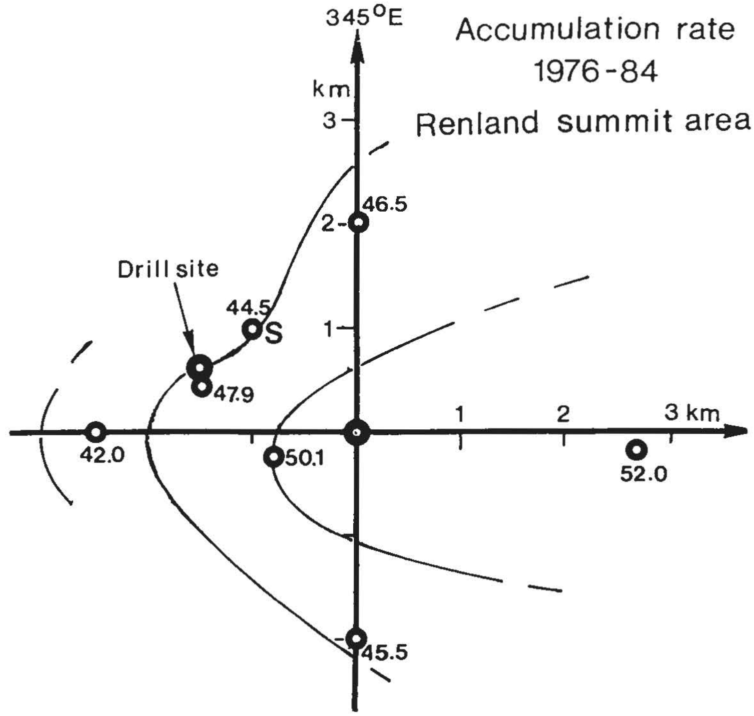
eastern part of the ice cap, where the elevation is highest, and measuring annual surface temperatures and accumulation rates.

For the airborne mapping, a 60 MHz pulse radar (Gudmandsen 1971) was installed in a Twin Otter aircraft from the Greenland Ice Reconnaissance Service. A quarter wave length antenna was mounted horizontally out from the aircraft body, which thus functioned as a vertical mirror. The antenna was tuned to a maximum output by adjusting its length. The nominal pulse width was 250 ns. Using visual navigation, seven North-South directed ice thickness profiles were obtained, having a total length of 250 km. The depth profiles revealed a stongly undulating bedrock topography. A deep East-West running valley (dashed line in Fig. 2B) covered by up to 700 m of ice drains the eastern part of the ice cap into the great Edward Bailey glacier on the South side of the peninsula (Fig. 4). North of the valley the ice thickness varies between 200 and 500 m.

The surface operation was initiated, when a Hughes 500 helicopter brought two men to the ice cap with an ice-core auger and satellite positioning equipment. An 11 m firn core was augered, and the temperature at the bottom of the hole indicated a mean annual surface

Fig. 4. The Edward Bailey glacier South of the ice cap drains the central and south-eastern part of the ice cap.

Fig. 5. Distribution of accumulation rate (in mm ice equivalent per year) in a high elevation area of the Renland ice cap, including the Summit S and the chosen drill site D.



temperature of (−18°C), and hence a bedrock temperature (ca. −13°C) far below the pressure melting point (> −0.5°C). The firn core was returned frozen to Copenhagen and a detailed $\delta^{18}O$ profile[1] was measured along its length. Based on previous experience concerning the signal to noise ratio, summer and winter layers were identified, and the core was found to represent 9½ years of accumulation, which provided a mean accumulation rate of 0.47 m of ice equivalent per year.

The conclusion of the 1985 pilot study was that the Renland ice cap fulfills the criteria mentioned above, except for No. 6. The summit of the ice cap is located in an area north of a valley, where the thickness varies between 300 and 500 m. The bedrock undulations of this magnitude required a more detailed radar survey prior to choosing the exact drill site.

A Landsat photograph (Fig. 2) taken at low sun angle (cf. the shadows of the mountains South of the fiord) confirmed the visual and radar observations of crevassed areas and the position of the ice divide (full line in Fig. 2B). This was the basis of the second pilot study carried out in 1987:

[1] The isotopic composition, δ, of water is defined as the relative deviation of the ^{18}O (or deuterium) concentration (c), from that (c_s) of Standard Mean Ocean Water, i.e.

$$\delta = (c/c_s - 1) \cdot 1000 \text{ per mille } (‰).$$

The δ value of a polar snow/ice sample is related to the temperature of formation of the snow/ice, δ being generally high in summer snow, and low in winter snow (Dansgaard et al. 1973).

(1) New shallow firn cores were augered in the summit area in order to estimate the spacial distribution of mean annual accumulation (the 1976 to 1984 mean values are plotted in Fig. 5).
(2) An automatic weather station was established. The meteorological data were used later to study the precipitation to air temperature relationship, and for interpretation of chemical ice core data.
(3) A strain network of 60 stations was established to cover a 2.5 × 4.5 km^2 area around the summit, cp. the rectangle in Fig. 2B.
(4) A Swedish broad band 120 MHz impulse radar, mounted on a Nansen sledge, was pulled a total distance of 70 km by a skidoo through the same network area. The annual stratification was mapped to 70 m depth throughout the network, but the range of the radar was too short to reach the bedrock.
(5) The Danish 60 MHz radar was installed in an Icelandic Flugfelag Nordurlands Twin Otter aircraft that flew at constant height over the snow surface in the network. The survey revealed ice thicknesses between 270 and 560 m and a generally hilly bedrock. A minor flat plateau appeared some 750 m southwest of the summit (S).

The main conclusion of the 1987 pilot study was that a suitable drill site could be found in the investigated area, at or close to the summit, but the 500 m mesh was too coarse to allow defining exactly the most favourable drill site.

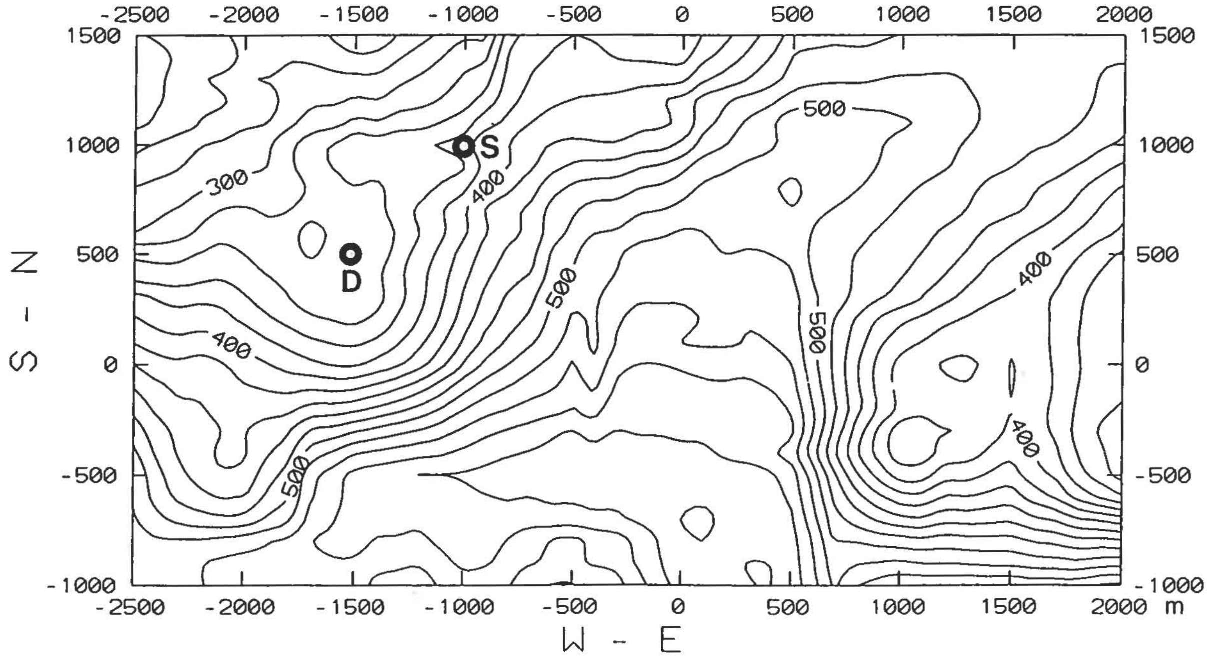


Fig. 6. Ice thicknesses (in meters) in a 2.5 × 4.5 km² network including the summit S and the drill site D. The map is based on impulse radar measurements, assuming an electromagnetic wave velocity of 1.71 · 10^8 m/s and an 8 m correction for firn densities lower than that of glacier ice. Equidistance 20 m.

RENLAND BEDROCK ELEVATION

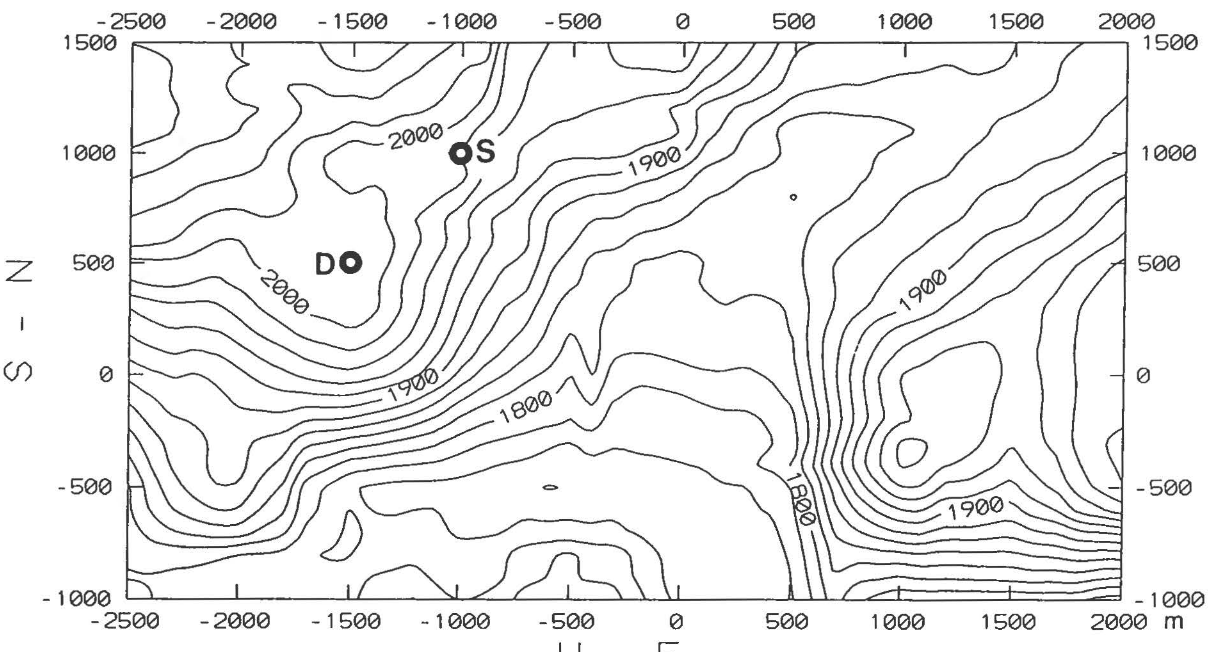


Fig. 7. Bedrock topography (in meters above sea level) in the same network as in Fig. 6, measured by the same technique and relative to the surface topopgrahy, which was determined by satellite position fixes combined with theodolite levelling. Equidistance 20 m.

The Renland expedition 1988

On July 5, 1988, a chartered Royal Swedish Air Force C-130 Hercules aircraft brought a five person field party, supplies and equipment, including the shallow drill, to the Constable Pynt air strip. A Greenlandair Twin Otter (and a helicopter, after the Twin Otter had "stranded" temporarily in the soft snow) took over and flew personnel and cargo to the Renland ice cap, not far from the summit.

The Swedish radar had now been modified to reach greater depths. It was used to survey a fine mesh subnet. Fig. 6 shows the ice thickness, Fig. 7 the bedrock topography, and Fig. 8 provides a 3-dimensional impression of the bedrock and surface topographies viewed from southwest. It was decided that under the very summit the ice is too thick (approximately 370 m) to ensure penetration by the shallow drill. At point D, however, 670 m Southwest of the summit S, the ice is 321 ± 5 m thick (Fig. 6) and, furthermore, the bedrock is relatively flat in that area (Fig. 7). Point D was therefore chosen as the drill site. Its geographic position was determined by a remote satellite fix at 71°18'17''N, 26°43'24''W, 2340 m a.s.l.

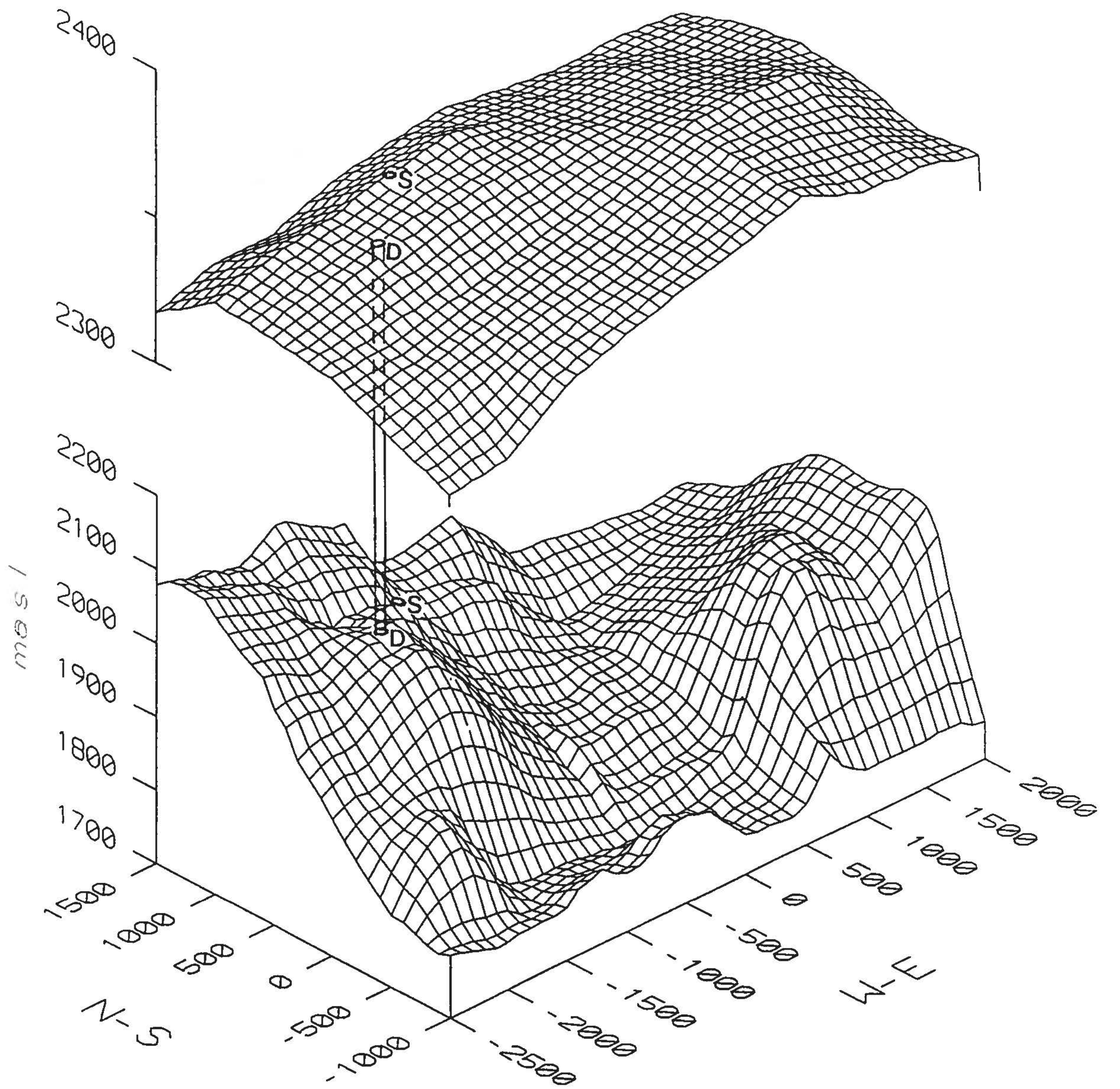


Fig. 8. 3-dimensional plot of surface and bedrock topographies in the same area as in Fig. 6, including the summit S and the chosen drill site D. Notice the different elevation scales.

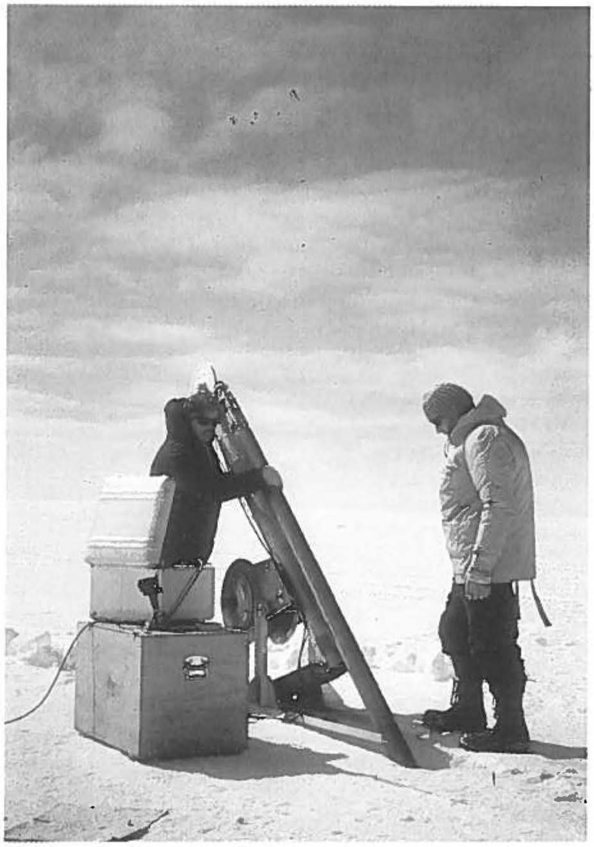


Fig. 9. The lightweight shallow drill used on Renland is operated by one or two persons. It is turned into horizontal position for maitenance and core retrieval. One meter of 7.5 cm diameter core is recovered per run. The drill, including generator, fuel and winch with 400 m of cable, can be mounted on a Nansen sledge and pulled by a skidoo.

The core drilling began on July 7 (Fig. 9) and terminated on July 14, when the drill got temporarily stuck at 324.4 m below the surface, very close to bedrock to judge from the radar soundings, and from the large particles in the deepest core increment recovered. The depth was more than twice the greatest depth previously reached by a shallow core drill. The core recovery was better than 99.9%, and the core quality was fair to good.

Thereafter, an additional 87 m core was drilled, and all personnel, equipment and core material was evacuated on July 28.

Accumulation rate

One of the most important parameters in ice flow modelling is the accumulation rate. The present day value measured during the preliminar operation in 1985 was essentially confirmed by a combined $\delta^{18}O$ and specific β activity study. Fig. 10 shows the two profiles spanning 18 annual cycles in the depth interval 18 to 30 m firn, or 9.97 to 18.20 m of ice equivalent, which corresponds to 0.46 m of ice equivalent per year. This interval is particularly interesting, because it contains at least two easily identifiable β reference horizons (Clausen & Hammer 1988): (1) the 1953 layer characterized by the first β level significantly higher than the background due to fall out of debris from the first hydrogen bomb test, and (2) the 1963 layer with much higher concentration of radioactive bomb debris than any other layer.

A better representation of long-term mean annual accumulation was obtained by considering the depths of well dated reference horizons. These are high acidity layers deposited shortly after great volcanic eruptions, which occurred somewhere in the northern hemisphere (Hammer *et al.* 1981). Some of the eruptions are dated by historic information, e.g. Tambora A. D. 1815 (fall out culminated in 1816), Laki 1783, or an unknown eruption 1601. In other cases (unknown eruptions 1259 and 1179, and Eldgjá 934) the signals were recognized from other ice cores dated by stratigraphic methods. The horizontal lines in Fig. 11 show the depths (in m of ice equivalent), where these reference horizons were identified in the Renland ice core. Each vertical line denotes the *in situ* mean annual layer thickness, λ, between two reference horizons. For example, the (Katmai) 1912 and (Tambora) 1816 layers were found at depths of 34.20 and 67.54 m ice equivalent, respectively. The 96 annual layers in between thus have a mean thickness of (67.54 – 34.20)/96 = 0.347 m. The heavy curve describes the plastic thinning of the layers, as they sink towards the bedrock 304 m of ice equivalent below the surface. The curve intersects the λ axis at λ_0 = 0.50 m, which is used as a representative of Holocene mean annual accumulation in the ice flow model.

The long-term δ record

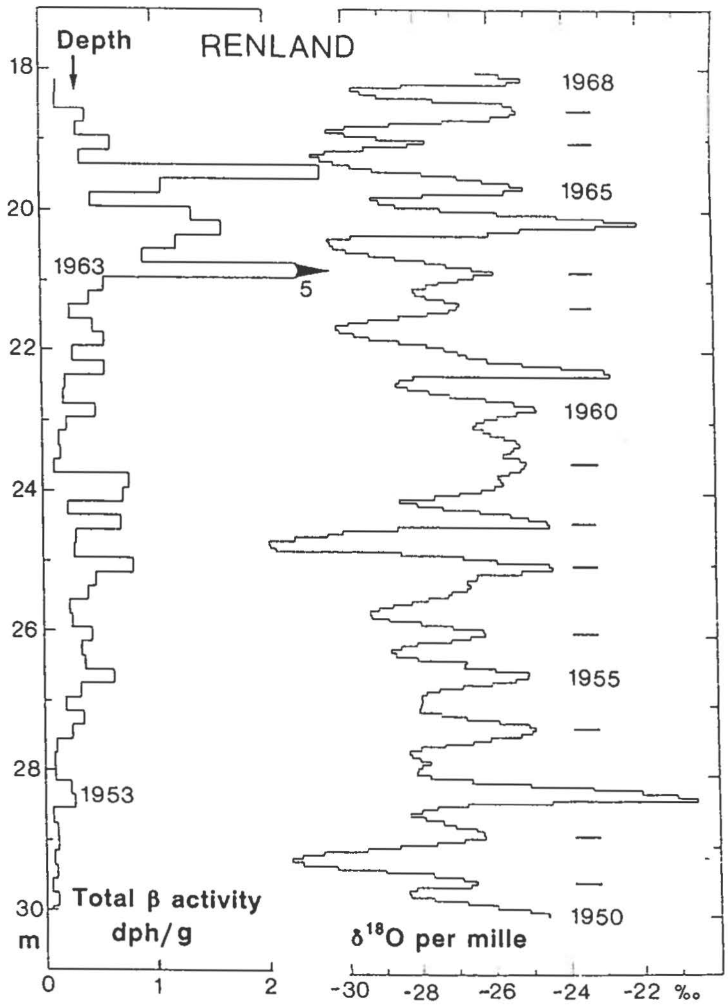
As mentioned in the footnote on p. 7 the δ value of high polar snow/ice depends mainly on the temperature of formation. The relationship between mean annual values of δ and surface temperature T_s °C is expressed as

$$\delta^{18}O = (0.67 \pm 0.02)\cdot T_s - (13.7 \pm 0.05)\ ‰ \qquad (1)$$

(Dansgaard *et al.* 1973; Johnsen *et al.* 1989). When averaged over many years, an 0.7‰ higher mean annual $\delta^{18}O$ value thus corresponds to about a 1°C increase in mean annual surface temperature. With adequate caution (Johnsen *et al.* 1972), smoothed δ profiles along ice cores may therefore be interpreted in terms of past climatic conditions.

Fig. 12 shows the complete $\delta^{18}O$ profile along the Renland ice core, measured as mean values of 55 cm

Fig. 10. Total specific β activity (in dph/g, i.e. disintegrations per hour per gram of ice) and $\delta^{18}O$ profiles spanning 18 years in the Renland ice core. The interpretation of the δ profile in terms of annual layers is supported by the 1953 and 1963 reference horizons identified in the total β activity record.



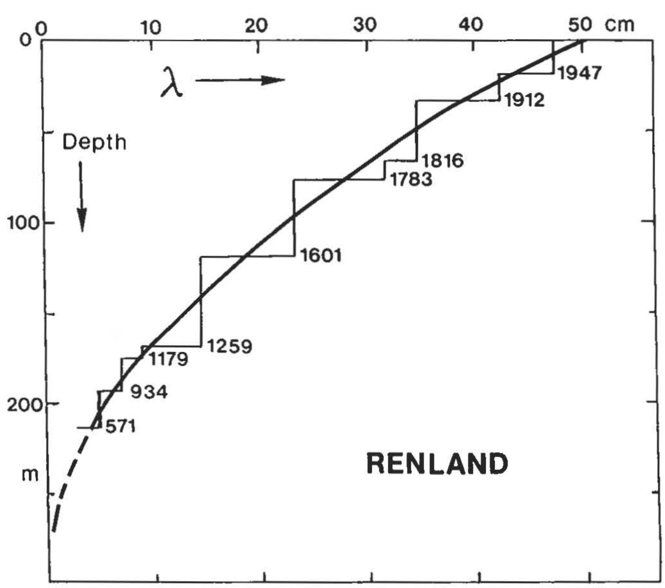


Fig. 11. Annual layer thickness in situ (λ cm ice equivalent) versus depth. The vertical lines indicate mean λ values in time intervals defined by dated reference horizons of acid volcanic fall out.

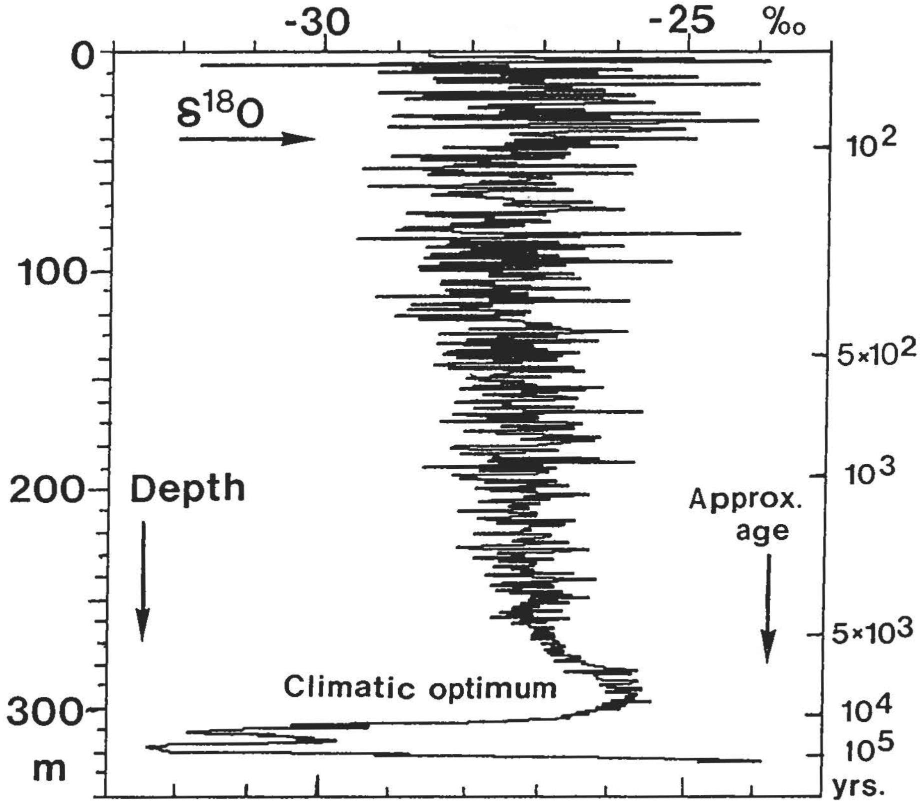


Fig. 12. The complete continuous $\delta^{18}O$ profile along the Renland ice core, measured on 55 cm increments, and plotted on the linear depth scale to the left. The deeper the increment, the wider its time span. This is the reason for the apparently decreasing scatter downward. The δ maximum between 280 and 300 m depth represents the postglacial climatic optimum. The deepest 18 meters represent the last glaciation and the preceding Eem interglacial, according to the calculated time scale to the right.

increments, and plotted on a linear depth scale. The apparently decreasing scatter downward is due to the progressing plastic stretching and thinning of the annual layers as they approach the bedrock. The deeper increments thus represent more annual layers. The deep δ minimum close to bedrock (depths between 306 and 320 m) indicates temperatures of original snow formation so low that this ice must have been formed under glacial conditions.

A detailed δ profile along the deepest 24 m of the Renland core is plotted in Fig. 13, and compared with that along the deepest 300 meters of the Camp Century ice core. The 4 m of extremely low δ ice between $z = 5$ and 9 m above bedrock in the Renland core will be treated below. Disregarding this feature, the correlation is generally good, even in detail, as indicated by the arrows, each of which points at strata for each core of presumably simultaneous deposition. The Renland core apparently reaches throughout the last glaciation and most of the preceding Eem interglacial.

The climatic warming from the coldest glacial to Preboreal conditions is reflected by increasing δ's. The total increase is as high as 9 ‰ at Camp Century in Northwest Greenland (where the shrinkage of the Laurentide ice sheet must have amplified the general climatic warming), and ca. 5 ‰ at Dye 3 in South Greenland (Dansgaard *et al.* 1982), at Summit (Johnsen *et al.* 1992), and in Renland (Fig. 13). According to equation (1), these figures correspond to 13 and 9 °C warming, respectively, and perhaps a degree more if the changing isotopic composition of sea water is accounted for. Using 1 ‰ as the total change of sea water composition up till the present (Shackleton *et al.* 1983), the δ curves further suggest that the present surface temperatures at Camp Century, Dye 3, Summit, and Renland are 21, 14, 11, and 9 °C higher, respectively, than during the coldest periods of the glaciation (cp. Table 2 in Johnsen *et al.* 1992). The relatively low figure for Renland may be due to isostatic uplift in East Greenland after the retreat of the main ice sheet (cp. Funder 1978), which would cool the surface by 1 degree per 100 m uplift.

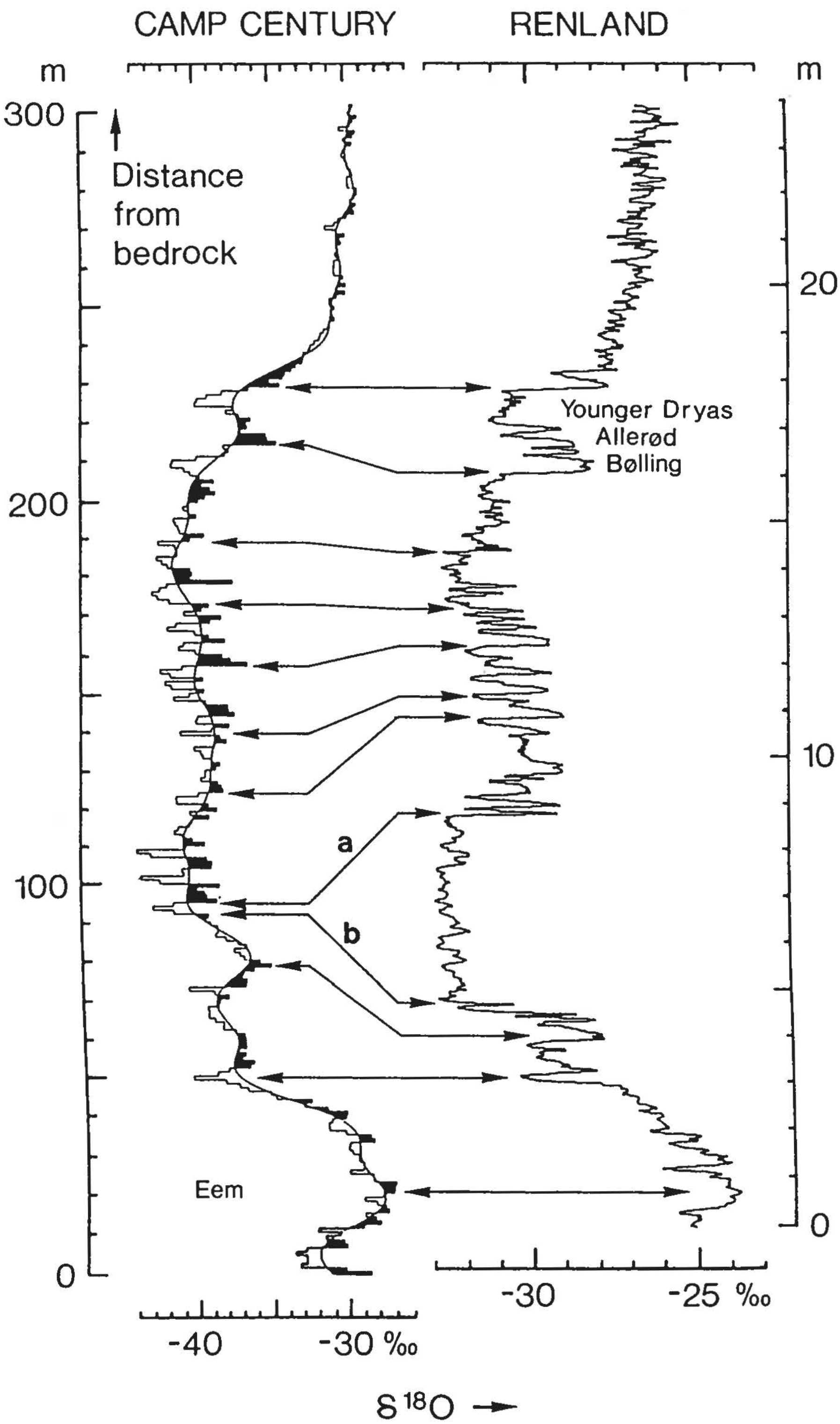


Fig. 13. The deepest 24 m of the Renland δ profile, to the right, compared with the deepest 300 m of the δ profile from Camp Century, North-west Greenland. Each arrow points at layers of presumably simultaneous deposition.

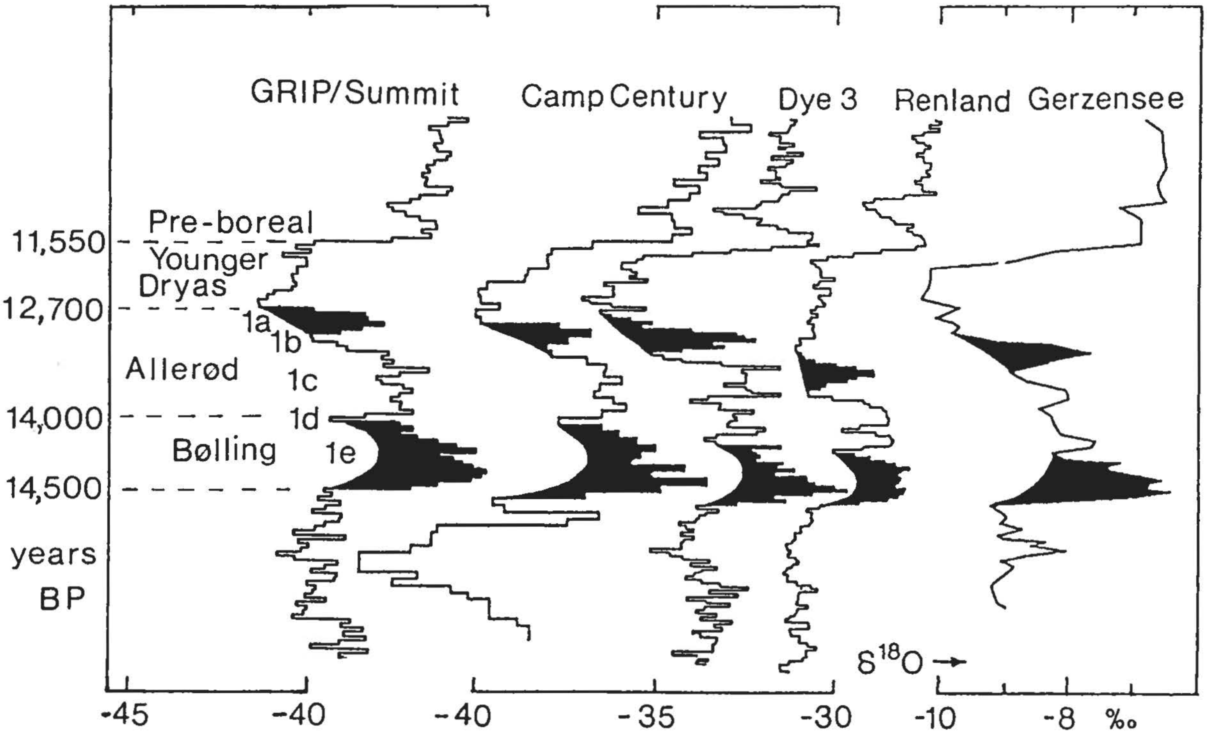


Fig. 14. The late-glacial climatic oscillation Bølling/Allerød-Younger Dryas recorded in four Greenland ice cores (Summit in central Greenland; Camp Century in Northwest; Dye 3 in South; and Renland on the East Greenland coast), and in a sediment core from Lake Gerzen, Switzerland. The first and the last spike in the Bølling/Allerød interstadial are set out in black. All of the records are plotted on linear depth scales (not shown). The non-linear time scale to the left is valid for Summit only. It has recently been established by Johnsen *et al.* (1992).

Bølling/Allerød – Younger Dryas

In Fig. 14 the last glacial climatic oscillation (Bølling/Allerød – Younger Dryas) is characterized by many common features in four Greenland records, and a record measured along a sediment core from the Swiss Lake Gerzen (Eicher & Siegenthaler 1976). Obviously, this oscillation went on the same way on both sides of the northern part of the North Atlantic Ocean, and since it has only been weakly recorded elsewhere, it was probably connected to changing flow direction and/or intensity of the North Atlantic Current (Dansgaard *et al.* 1989).

All of the five records are plotted on linear depth scales. The time scale to the outer left is valid for the Summit ice core, and has recently been derived by multiparameter stratigraphic methods (Johnsen *et al.* 1992). According to this time scale, the Bølling mild period began 14,500 ± 200 yr BP, and the Younger Dryas ended abruptly 11,550 ± 70 yr BP, *i.e.* earlier than previously estimated by simpler methods (Hammer *et al.* 1986), but in agreement with uranium-thorium dated sea level observations on Barbados corals (Bard *et al.* 1990).

The deuterium excess ($d = \delta D - 8 \cdot \delta^{18}O$) is an index of the evaporation conditions at the source region of the moisture that is deposited as snow at the high elevation regions in Greenland (Johnsen *et al.* 1989). The d value is as high as 8 to 10 ‰ in the Younger Dryas ice (like in the preceding low δ periods), which indicates high ocean surface temperature in the source region. This rules out the possibility that the low δ values in Younger Dryas were due to glacier melt water in the northernmost part of the North Atlantic being an important moisture source during Younger Dryas, which has been suggested (Fairbanks 1989). The Younger Dryas was a very cold period, in Greenland as well as in Europe.

Quasi-stationary glacial stages

The Bølling/Allerød-Younger Dryas climate oscillation was the last example of a long series of interstadials. Throughout the mid and late glacial, the Arctic climate shifted between two apparently quasi-stationary (*i.e.* temporarily stable) stages, cf. Fig. 13.

The shifts in δ are smaller in Renland (2.8 ‰) than at Camp Century (5.4 ‰), Dye 3 (4.8 ‰) and Summit (4.4 ‰, Johnsen et al. 1992), which may be ascribed to (1) the diffusion of the isotopic components, which is particularly efficient at smoothing temporal δ variations in the deep and extremely thinned Renland ice layers (this is also why the δ shifts in the interstadials appear less abrupt in Renland, than on the main ice sheet), and (2) the fact that orographic mechanisms are less important

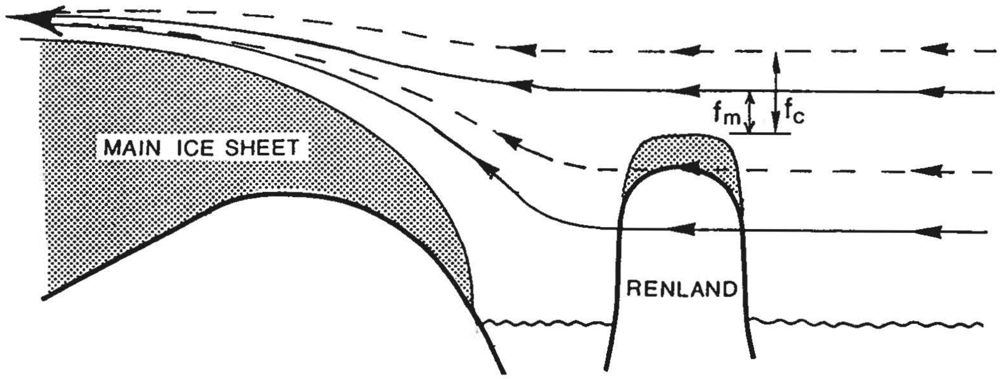


Fig. 15. Schematic drawing of the tracks of percipitating clouds feeding the Renland ice cap and the Greenland ice sheet under mild (thin, full curves) and cold (dashed curves) glacial conditions. During the cold periods, Renland received snow from a large fraction f_c of the cloud band with a mean temperature closer to that of the entire band, than was the case during interstadials, when Renland received precipitation from a minor top fraction, f_m, of the cloud band. Snow on the main ice sheet is created at approximately the same altitude under all circumstances. This may be the reason for the δ shifts being lower at Renland than on the ice sheet.

for the deposition of snow on the Renland ice cap than on the main ice sheet: The peninsula is nearly surrounded by deep branches of the Scoresby Sund Fjord, and the small ice cap is, and has probably been separated from the main ice sheet throughout the period characterized by abrupt δ shifts. During the cold stages, the polar front was displaced far south of its mild stage position. The cyclones had to travel further in order to reach Renland during the cold stages, with a consequently higher degree of occlusion and a higher elevation of the warm air upon arrival, than during mild stages (Fig. 15). The snow deposited on the high elevation Renland ice cap in the cold stages thus had relatively low δ's, but the snow represented a larger fraction f_c than in the mild stages (f_m), and its δ was therefore not as low as if the snow had originated in the same portion of the clouds as in the mild stages. This limiting factor is less important, and the δ shifts therefore higher, in the high elevation areas of the main ice sheet, were the precipitating clouds are thinner and always close to the snow surface.

The δ records indicate that the cold stages on the main ice sheet were some 7°C colder than the mild ones. Temperature fluctuations in Greenland of this order of magnitude must have been associated with significant environmental changes in the entire North Atlantic region, particularly in Europe, and may be compared with the botanical evidence of "swift climatic changes" during the last glaciation (Kolstrup 1990).

The duration of the late-glacial interstadials ranges between 500 and 2000 years, and they occur with irregular time intervals (Johnsen *et al.* 1992), associated with advances and abrupt retreats of the North Atlantic sea ice cover, which in turn are probably caused by shifts in the direction/intensity of the North Atlantic Current. The climatic shifts have not been satisfatorily explained, however. Even the mechanism behind the most thoroughly studied climatic oscillation, the Bølling/AllerødYounger Dryas, is still under discussion (Zahn 1992).

It cannot be excluded, therefore, that we are faced with a chaotic feature of climate, implying two or more possible atmospheric/oceanic circulation modes for a given set of primary climate parameters, *i.e.* Earth orbit parameters, chemical composition of the atmosphere, etc. The interstadials could then be the result of random, and more or less abrupt shifts between the two modes. If so, it would further complicate climate prediction. The interstadials were linked to changing circulation in a North Atlantic Ocean partly covered by sea ice, indeed, but it should be borne in mind that within the last millenium, the well documented medieval warmth in Europe was gradually replaced by "the little ice age", when Iceland was frequently surrounded by sea ice, and the succeeding warming that culminated in an abrupt temperature increase in the 1920'ies, too abrupt to be explained by the increasing greenhouse effect. This oscillation had smaller amplitude than its glacial predecessors, but similar sequence of events, *i.e.* gradual cooling followed by abrupt warming (Johnsen *et al.* 1992).

Comparison with dust and acidity profiles

In Fig. 16, the Renland δ profile from Fig. 13 is compared with dust concentration and acidity profiles measured by light scattering (Hammer 1977) and electrical conductivity (Hammer 1983) techniques, respectively.

The continental dust was brought to Greenland by violent storms mainly from loess covered regions in North America (Cragin *et al.* 1977). The dust concentration in the ice depends of course on the conditions in the source area, e.g. its magnitude, humidity and vegetation, but the storminess in the entire North Atlantic

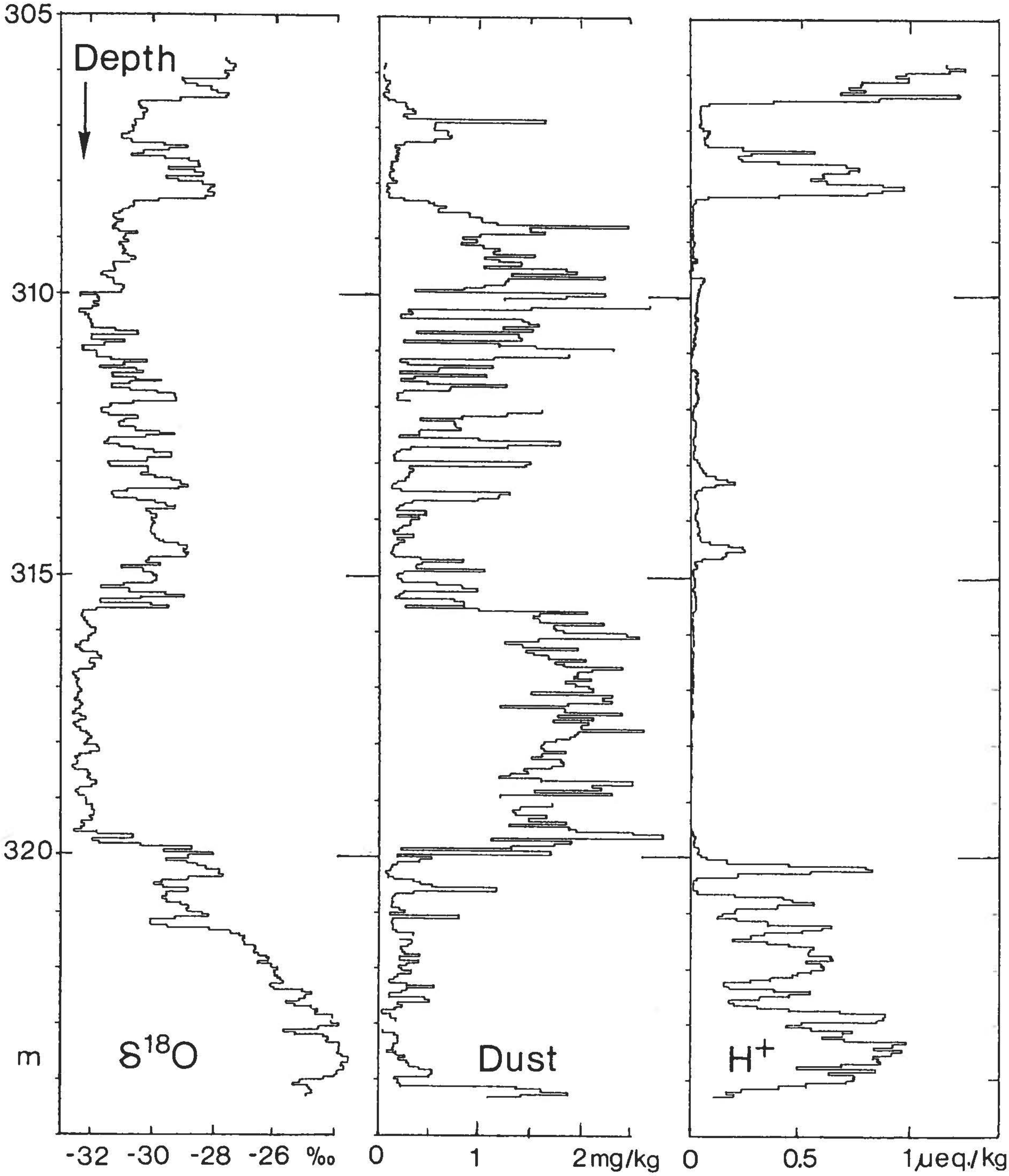


Fig. 16. $\delta^{18}O$, dust concentration and acidity (H^+ concentration in μequivalents per kg ice) profiles along the pre-Holocene part of the Renland core. The acidity is close to zero in most of the ice deposited during the glaciation.

region is undoubtly an important factor. The high dust concentrations associated with very low δ's (Fig. 16) suggest, therefore, that the coldest periods of the glaciation were dominated by frequent violent storms, which should also be expected in view of the high latitudinal temperature gradients that existed at mid latitudes (Dansgaard *et al.* 1989).

The electrical conductivity measured in the ice core is mainly due to fall out of strong acids. Particularly high conductivities, due to H_2SO_4 and/or HCl, are found in ice deposited shortly after great volcanic eruptions in the northern hemisphere (Hammer *et al.* 1981), but the deepest part of the Renland core is too compressed to allow tracing of individual eruptions. Throughout most of the glaciation the conductivity of the ice was close to zero (Fig. 16). This does not mean that the volcanic activity was very low, however, but rather that volcanic acids were neutralized in the atmosphere by an alkaline aerosol, mainly $CaCO_3$ blown up from the continental shelves that became increasingly exposed, when the sea level sank. We shall return to this point later.

The homogeneous layer

The isotopically homogeneous 4 m layer, at distance z = 5 m to 9 m above bedrock in the Renland core (Fig. 13), corresponding to 316 to 320 m below surface, has no analogue in any other climatic record. Its thickness must therefore be due to other processes or events than year-by-year deposition of snow followed by regular and undisturbed deformation of the annual layers. This has to be accounted for in time scale calculations.

At first glance, the δ profile seems to indicate a disturbed stratigraphy, in view of the small distance to the bedrock. However, the layers in the very deepest 5 meters are essentially undisturbed and perpendicular to the core axis. It looks, therefore, as if an originally very thick layer of homogeneous ice were deposited in a short and extremely cold period, after which the normal accumulation was resumed.

It can not be excluded that the high plateau of Renland was temporarily overidden by the main ice sheet, which left originally several hundreds of meters of ice from the inland, when it retreated in a succeeding milder period. This would be in accordance with observations indicating that the maximum extent of the ice sheet in North Greenland, including Scoresby Sund, occurred long before the global glacial maximum around 20,000 yrs BP (Funder *et al.* 1991). On the other hand, one might expect that (**1**) such ice from the inland would have even lower δ, and be isotopically inhomogeneous like other ice deposited during the glaciation, and that (**2**) overriding of huge amounts of Inland Ice would have removed most, if not all of the ice deposited previously. If nevertheless the 4 m of ice in question originates from the inland, its original thickness just after the deposition on Renland must have been of the order of 100 times greater than at present, *i.e.* of the order of 400 m. Hence, the surface elevation of the Inland Ice covering Renland at that time may well have been 2800 m above the then sea level.

Another possible explanation, which does not imply overriding by the main ice sheet, is that the isotopically homogeneous 4 m of ice be due to the Boudinage effect (Staffelbach *et al.* 1988) that causes strongly varying thickness of particularly soft layers exposed to long lasting local stresses around casual obstacles on the bedrock. If so, the thickness of the layer in question could be much smaller, perhaps close to zero, in an ice core drilled a few meters away. Its extremely low δ value is associated with extremely high impurity content (Fig. 16) and, therefore, extremely low "viscosity" (Shoji & Langway, 1984), which may be the explanation why no other low δ layer in the core has apparently undergone drastic deformation. On the other hand, it is not easy to visualize how the Boudinage effect could have multiplied the thickness of a low δ layer to make up 25 % of all Pleistocene ice in the core.

However, regardless of which one of the above interpretations that is realistic, if any, the nearly constant and extremely low δ value through the z = 5 to 9 m increment does not represent a very long period of stable climate, because lower temperature is generally associated with decreasing climatic stability and, furthermore, no other climatic record shows evidence of a very long lasting stable glacial period. Most of the δ profile along the increment has therefore been disregarded in the time scale calculation (Johnsen *et al.* 1992) outlined on p. 20.

The deepest 5 m of the core

Fig. 17 shows detailed $\delta^{18}O$, and Ca^{++} and H^+ concentration (acidity) profiles along the deepest 5.4 m of the core, derived by ion chromatography and a simple electrical conductivity method (Hammer 1983), respectively. We interprete the very deepest 2 m of ice as being deposited during the Eemian interglacial, because like post-glacial ice, it has (**1**) high δ values, in fact higher than recent snow, indicating generally warm climate; (**2**) numerous melt layers, indicating very warm summers; (**3**) low Ca^{++} and generally high H^+ values, indicating an acid aerosol without the significant $CaCO_3$ component that neutralizes the acids in the atmosphere under glacial conditions. The 322.4 to 321.3 m depth interval is interpreted as the gradual or stepwise transition of Eem (substage 5e) into the early part of the last glaciation, *i.e.* substages 5d to 5a in the marine oxygen isotope record (the 321.3 to 319.8 m depth interval). During the transition from Eem to full glacial severity, the Ca^{++} and H^+ concentrations vary in antiphase (the extremely low δ ice around 319 m depth has 0.8–1.0 mg Ca^{++} per kg). This may reflect the influence of a $CaCO_3$ aerosol windborne from the continental shelves and varying concurrently with the sinking and rising sea level.

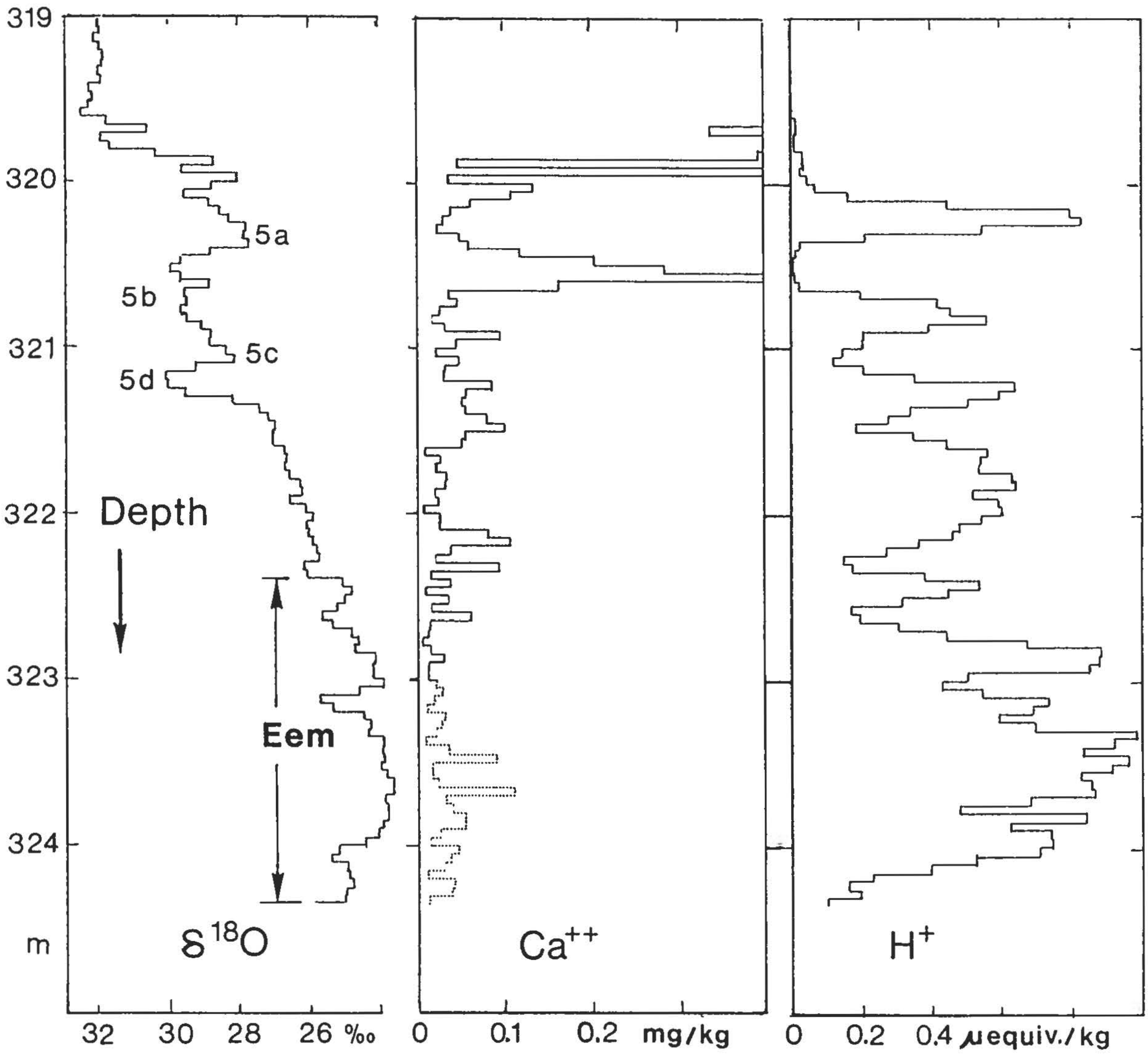


Fig. 17. $\delta^{18}O$, Ca^{++} and H^{+} concentration profiles spanning the deepest 5 m of the Renland ice core. The deepest 2 m is interpreted as being of Eemian origin. The decreasing δ's above it reflect the transition of the Eemian interglacial into the early stages of the glaciation (5d to 5a). During these stages, the acidity varies in antiphase with the Ca^{++} concentration, probably because the acid aerosol was neutralized by a $CaCO_3$ aerosol originating from the continental shelves that were exposed concurrently with the changing, but generally sinking sea level.

It should be noticed that, throughout most of the mid- and late-glacial time, which was dominated by frequent shifts between cold and mild stages, the acidity did not change very much, in contrast to the dust concentration. In fact the acidity was generally close to zero. This is consistent with the concept that the sea level, and therefore the global amount of continental ice, was not influenced considerably by the relatively short lasting interstadials, whereas the temperature and storminess varied drastically concurrently with retreats and advances of the sea ice cover in the North Atlantic Ocean.

Dating the core

Ice cores have previously been dated by **(1)** counting annual layers downward from the surface; this is a laborious, but accurate method, which has recently been applied throughout the entire Holocene and a considerable part of the last glaciation (Clausen *et al.* 1989; Johnsen *et al.* 1992); **(2)** by radioactive isotopes, in particular ^{14}C (Andrée *et al.* 1984), but even with the modern accelerator massspectrometer technique the ^{14}C range is limited to 30,000 years B.P.; **(3)** by correlation with other dated records, using reference horizons (Dansgaard *et al.* 1982); and (**4**) by ice flow calculations (Dansgaard & Johnsen, 1989; Reeh *et al.* 1985; Dahl-Jensen 1989).

As to the Renland core, a combination of the last two methods had to be applied. Volcanic reference horizons have been mentioned above. In addition, a few obvious pre-Holocene time markers were used as inputs to the ice flow model, which calculates the annual layer thickness versus depth relationship and, by numerical integration, a time scale along the core.

The flow model used (Johnsen & Dansgaard 1992) was modified from Dansgaard & Johnsen (1969) to account for (1) the "no shear stress situation" close to the ice divide, and (2) the "lubricating" effect of a thin and relatively soft bottom layer of silty ice, which corresponds to a small, but important sliding along the bedrock.

As mentioned above, most of the δ profile along the isotopically homogeneous 4 meter layer in the Renland core has to be disregarded, when calculating a time scale, and yet the ice flow model must account for its

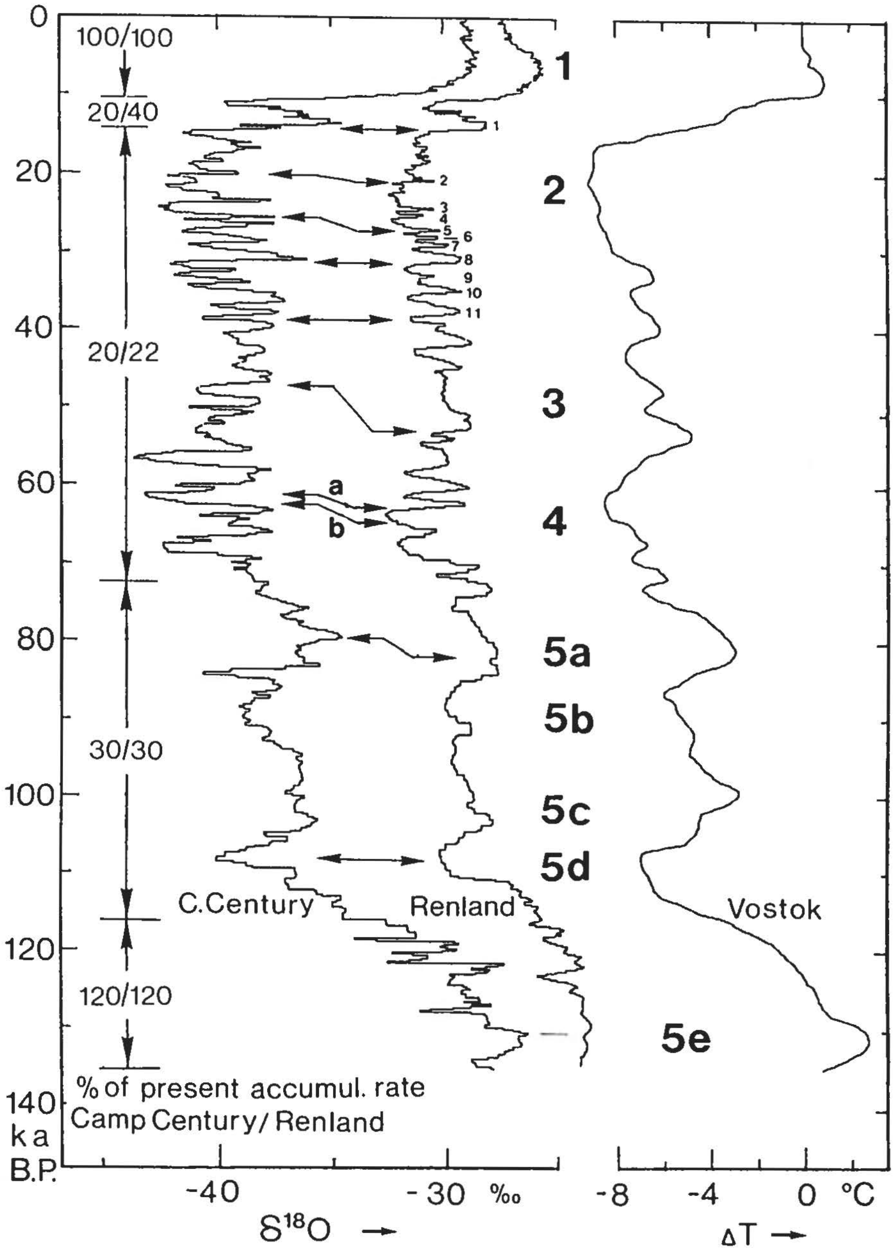


Fig. 18. *Outer left:* Accumulation rates in per cent of present values at Camp Century/Renland, used for time scale calculations. *Mid section:* Camp Century and Renland δ profiles plotted on the common time scale described in the text. The arrows between the two records are transferred from Fig. 13. A tentative reconciliation with the marine oxygen isotope stratigraphy (large figures) is added between the Renland record and, in the *right hand section,* Vostok temperature deviations (Petit *et al.* 1990) from the present temperature. The small figures indicate the middle and late glacial interstadials defined by Johnsen *et al.* (1992).

existence. Regardless of the origin of this layer, it has influenced the thinning of the rest of the strata in the core, perhaps since the formation of the adjoining layers, and in any case through many thousands of years. This was accounted for in the time scale calculation by, first, removing most of the $z = 5$ to 9 m interval from the δ record, then restablishing a continuous δ record over the entire core length by stretching the remaining parts of the glacial record to fill in the gap: 9.2 m of the record above the gap (up to the shift to the relatively rigid Holocene ice), and 1.8 m below the gap (down to the shift to the relatively rigid Eemian ice) were stretched uniformly and hooked together at $z = 6.3$ m. This procedure did not influence the age calculation outside the interval $z = 3.0$ m to 17.8 m.

For index reference, four time-scale fixed points were used: **(1)** zero year at the surface; **(2)** 10,720 yr BP for the end of the Younger Dryas event (Hammer *et al.* 1986) at $z = 17.8$ m (306.6 m depth); **(3)** 14,000 yr BP (Hammer *et al.* 1986) for the onset of the Bølling interstadial at $z = 15.9$ m; and **(4)** 125,000 yr BP for the "warmest" part of the Eemian ice at $z = 0.7$ m. The most recent estimates on **(2)** and **(3)** are 11,500 and 14,500 yr BP, respectively, but using these figures would not cause significant changes. This scheme put a tight constraint on the choice of model parameters. For example, fixed point **(4)** made it necessary to assume the past accumulation rates shown to the outer left in Fig. 18 as percentages of present values at Camp Century and Renland.

A similar procedure was applied to the Camp Century record. Nothing was removed from this δ profile, though, as it shows no evidence of disturbed stratigraphy due to the much higher distances to bedrock (Fig. 13). The primary reason for the low values of glacial accumulation at both Renland and Camp Century is that the mean position of the polar front in the North Atlantic Ocean was located some 1000 km farther South than now (Ruddiman & McIntyre 1981). In addition, precipitating clouds in the coastal regions had probably higher elevation than now, which reduced the orographic part of the precipitation at Renland (Johnsen *et al.* 1992), and at Camp Century, then positioned at a higher altitude and located on the lee side of the North American ice sheet, which must have reduced the precipitation rate considerably.

Both δ profiles are plotted on a common linear time scale in Fig. 18. The large figures refer to the conventional marine oxygen isotope stages. The small figures refer to the middle and late glacial interstadials IS1 to IS11, defined by Johnsen *et al.* (1992). The applied compensation for the homogeneous layer is supported by the common features between the fixed points, which are distinct in spite of diffusive smoothing of the strongly compressed Renland record, cp. the interstadials; the general trends; and the details during the stage 5e to 5d transition.

δ profile corrections

Interpreting a δ record in terms of surface temperature changes generally calls for several corrections, but some of them are negligible in the case of Renland: The correction for different deposition conditions upstream is zero, when a core is obtained from the summit of an ice cap. Furthermore, since the horizontal extention of the ice cap is limited, its thickness cannot have changed drastically, and surface elevation changes relative to the bedrock may be disregarded.

On the other hand, the Renland bedrock elevation relative to the sea surface did change significantly due to isostatic uplift and general sea level changes: ^{14}C dated raised beaches show a more than 110 m uplift of the Scoresbysund Fjord area, relative to the present sea level, from the end of the glaciation to 6000 yr BP (Funder 1978). The total uplift since the glacial maximum relative to the present sea level may very well have been at least twice as much, but relative to the changing sea level it has hardly been very much higher. It therefore corresponds to approximately 1.3°C colder surface temperature and, according to equation (1), to a δ correction of approximately 1 ‰, which has not been applied to the record in Fig. 7, though.

According to this figure, δ of snow deposited on the Renland ice cap 6000 yrs ago was 1.8 ‰ higher than it is now, and since the relative uplift (Funder 1978) as well as changes in δ of sea water have been insignificant in the intervening period, it may be concluded that the 1.8 ‰ corresponds to a 2.5 °C warmer climate in Eastern Greenland 6000 yrs ago, cp. formula (1). This is in agreement with a previous estimate based on pollen analyses (Funder 1978).

The δ records have neither been corrected for the higher δ of sea water existing during the glacial maximum. This correction is approximately −1 ‰ (Shackleton *et al.* 1983), which counteracts the approximately +1 ‰ correction for relative uplift. Hence, the glacial to post-glacial δ shift in Renland was approximately 7 ‰ (Fig. 7) as at Dye 3 in South Greenland, indicating a 10° to 11°C colder climate in Greenland 20,000 years ago. This agrees with measured and calculated Dye 3 temperature data (Dahl-Jensen & Johnsen 1986). Similarly, the Bølling/Allerød interstadial, and the isotope stages 5a and 5c were probably cooler in Renland than they appear in Fig. 18.

As mentioned above, the deepest ice in the Renland core is probably of Eemian origin. Since the main ice sheet at that time was smaller than today, and δ of sea water only slightly lower (Shackleton *et al.* 1983), a positive correction for isostatic uplift should probably be added. Therefore, it may be concluded that in Eemian time Renland δ's were at least 3.7‰ higher than today, and the East Greenland climate at least 5°C warmer, in essential agreement with previous estimates based on studies of plant and insect remains (Böcher & Bennike 1991).

The Camp Century core was not drilled on a summit, and the deeper layers were therefore deposited further upstream at higher elevations and, consequently, lower temperatures. A positive correction, increasing with the age of the ice, should therefore be put on the δ record. This would set off the postglacial maximum in the record, which is hardly noticeable in Fig. 13. So far, however, the data only justify the conclusion that there was also a pronounced post-glacial climatic optimum in Northwest Greenland.

The obvious similarity with the Vostok temperature record (Petit *et al.* 1990) in Fig. 18 suggests a roughly parallel climatic development in the Arctic and Antarctica, and it is quite possible that the "climatic generator" in the North Atlantic also influenced Antarctica in a modulated fashion. Demonstrating possible synchronism calls for better dating of the records, however, and for more details in the Vostok record.

Conclusion

The Renland study has shown that some of the small and isolated ice caps in the coastal areas of Greenland offer favourable conditions for ice coring and ice core studies: Their accessibility is tantamount to relatively low-cost logistics; the relatively thin ice caps can be penetrated by a light-weight core drill; and the time series obtained by analyzing the cores are valuable, not only as supplements to those measured along deep ice cores from the main ice sheet, but also as indicators of regional environmental changes. As an example of an obvious candidate for further studies, one may mention the Hans Tausen ice cap in Peary Land, located so close to the Arctic Ocean that its history is most likely settled in the ice cap.

Acknowledgements

The Renland field work and ice core studies were funded by the Nordic Council of Ministers, the Danish, Icelandic, and Swedish Natural Science Research Councils, and the Commission for Scientific Research in Greenland. Chartered aircrafts from Flugfelag Nordurlands, Greenlandair, Greenland Ice Reconnaissance Service, and the Royal Swedish Air Force were used for transport. We thank Dr. Eric Wolff, British Antarctic Survey, for measuring the β profile in Fig. 10, and Drs. Else Kolstrup and Svend Funder for helpful discussions.

References

Andrée, M., Moor, E., Beer J., Oeschger, H., Stauffer, B., Bonani, G., Hofman, H. J., Morenzoni, E., Nessi, M., Suter, M., & Wolfli, W. 1984. ^{14}C dating of polar ice. Nuclear Instruments and Methods in Physical Research 233: 380–384.

Bard, E., Hamelin, B., Fairbanks, R. G. & Zindler, A. 1990. U/Th calibration of the ^{14}C scale based on sea level changes observed on Barbados corals. – Nature 345: 405–410.

Böcher, J. & Bennike, O. 1991. Interglacial land biotas of Jameson Land, East Greenland. – Lundqua Report 33: 129–136.

Clausen, H. B. & Hammer, C. U. 1988. The Laki and Tambora eruptions as revealed in Greenland ice cores from 11 locations. – Annals of Glaciology 10: 16–22.

Clausen, H. B., Gundestrup, N., Hansen, S. B. & Johnsen, S. J. 1989. – In Ice Core Drilling: Proceedings of the 3'rd Internat. Workshop on Ice Drilling Technology, Grenoble, France, 1988: 14–20.

Cragin, J. H., Herron, M. M., Langway Jr., C. C. & Klouda, G. 1977. Inter hemispheric comparison of changes in the composition of atmospheric precipitation during the late Cenozoic era – In Dunbar, M. J. (ed). Polar Oceans. Proceedings of the Polar Oceans Conference, McGill Univ., Montreal, May 1974, Calgary, Alberta, Arctic Institute of North America, 617–631.

Dahl-Jensen, D. 1989. Two-dimensional thermo-mechanical modelling of flow and depth-age profiles near the ice divide in central Greenland. – Annals of Glaciology 12: 31–36.

Dahl-Jensen, D. & Johnsen, S. J. 1986. Palaeotemperatures still exist in the Greenland ice sheet. – Nature 320: 250–252.

Dansgaard, W. & Johnsen, S. J. 1969. A flow model and a time scale for the ice core from Camp Century, Greenland. – Journal of Glaciology 8: 215–223.

Dansgaard, W., Johnsen, S. J., Møller, J. & C. C. Langway, Jr. 1969. One thousand centuries of climatic record from Camp Century on the Greenland ice sheet. – Science 166: 377–381.

Dansgaard, W., Johnsen, S. J., Clausen, H. B. & Gundestrup, N. 1973. Stable isotope glaciology. – Meddelelser om Grønland 197: 1–53.

Dansgaard, W., Clausen, H. B., Gundestrup, N., Hammer, C. U., Johnsen, S. J., Kristinsdottir, P. M. & Reeh, N. 1982. A new Greenland deep ice core. – Science 218: 1273–1277.

Dansgaard, W., Johnsen, S. J., Clausen, H. B., Dahl-Jensen, D., Gundestrup, N., Hammer, C. U. & Oeschger, H. 1984. North Atlantic Climatic oscillations revealed by deep Greenland Ice Cores. – In Hansen, J. E. & Tagahashi, T. (eds.). Climatic Processes and Climate Sensitivity. Geophysical Monograph 29, Maurice Ewing Volume 5: 288–298.

Dansgaard, W., White, J. W. C. & Johnsen, S. J. 1989. The abrupt termination of the Younger Dryas climate event. – Nature 339: 532–534.

Eicher, U. & Siegenthaler, U. 1976. Palynological and oxygen isotope investigations on late-glacial sediment cores from Swiss lakes. – Boreas 5: 109–117.

Fairbanks, R. G. 1989. A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. – Nature 342: 637–642.

Fisher, D. A. 1982. Carbon-14 production compared to oxygen isotope records from Camp Century, Greenland and Devon Island, Canada. – Climatic Change 4: 419–426.

Funder, S. 1978. Holocene stratigraphy and vegetation history in the Scoresby Sund area. – Geological Survey of Greenland, Bulletin 129: 1–166.

Funder, S., Hjort, C. & Landvik, J. Y. 1991. Quaternary stratigraphy of Jameson Land – A first approximation. – Lundqua Report 33: 171–175.

Gudmandsen, P. E. 1971. Electromagnetic probing of ice – In: Wait, J. R. (ed.). Electromagnetic Probing in Geophysics. – Galem Press, Boulder Colorado: 321–348.

Gundestrup, N. S., Johnsen, S. J. & Reeh, N. 1984. ISTUK, a deep ice core drill system – Special Report No. 84–34 Ice drilling technology. United States Army Cold Regions Research and Engineering Laboratory, Hannover, New Hampshire, 7–19.

Hammer, C. U. 1977. Dust studies on Greenland ice cores. – Proceedings of Symposium on Isotopes and Impurities in Snow and Ice, International Association of Hydrological Sciences, Commission of Snow and Ice, International Union on Geophysics and Geodesy, XVI General Assembly, Grenoble Aug. Sept., 1975. IAHS-AISH publication 118: 365–370.

Hammer, C. U. 1983. Initial direct current in the build-up of space charges and the acidity of ice cores. – Journal of Physical Chemistry 87: 4099–4103.

Hammer, C. U., Clausen, H. B. & Dansgaard, W. 1981. Past volcanism and climate revealed by Greenland ice cores. – Journal of Volcanology and Geothermal Research 11: 3–10.

Hammer, C. U., Clausen, H. B. & Tauber, H. 1986. Ice-core dating of the Pleistocene/Holocene boundary applied to a calibration of the ^{14}C time scale. – Radiocarbon 28, No. 2A: 284–291.

Hansen, B. L. & Langway Jr., C. C. 1966. Deep core drilling in ice and core analyses at Camp Century, Greenland, 1961–66. – United States Antarctic Journal 1: 207–208.

Johnsen, S. J. & Dansgaard, W. 1992. On flow model dating of stable isotope records from Greenland ice cores. – In Bard, E. (ed). Proceedings of NATO workshop: The last deglaciation, absolute and radiocarbon chronologies. Erice, Italy, 1990. N.A.T.O. Advanced Studies Institute Series, Vol. I 2., 13–24.

Johnsen, S. J., Dansgaard, W., Clausen, H. B. & Langway, Jr., C. C. 1972. Oxygen isotope profiles through the Antarctic and Greenland ice sheets. – Nature 235: 429–434 and 236: 249.

Johnsen, S. J., Dansgaard, W., Gundestrup, N., Hansen, S. B., Nielsen, J. O. & Reeh, N. 1980. A fast lightweight core drill. – Journal Glaciology 25: 169–174.

Johnsen, S. J., Dansgaard W. & White, J. W. C. 1989. The origin of Arctic precipitation under present and glacial conditions. – Tellus 41B: 452–468.

Johnsen, S. J., Dansgaard, W., Clausen, H. B., Gundestrup, N., Jouzel, J., Stauffer, B. and Steffensen, J. P. 1992. Irregular interstadials recorded in a new Greenland ice core. – Nature 359: 311–313.

Kolstrup, E. 1990. The puzzle of Weichselian vegetation types poor in trees. – Geologie en Mijnbouw 69:253–262.

Langway Jr., C. C. 1970. Stratigraphic analysis of a deep core from Greenland – Geological Society of America, Special Paper 125: 1–186.

Overgaard, S. & Gundestrup, N. 1985. Bedrock topography of the Greenland ice sheet in the Dye 3 area. – In: Langway, C. C. Jr., Oeschger, H. & Dansgaard, W. (eds). Greenland Ice Cores: Geophysics, Geochemistry and Environment. American Geophysical Union Monograph 33: 49–56.

Petit, J. R., Mounier, L., Jouzel, J., Korotkevich, Y. S., Kotlyakov, V. I. & Lorius, C. 1990. Palaeoclimatological and chronological implications of the Vostok core dust record – Nature 343: 56–58.

Reeh, N., Johnsen S. J. & Dahl-Jensen D. 1985. Dating the Dye 3 deep ice core by flow model calculations. – In Langway Jr., C. C., Oeschger, H. & Dansgaard, W. (eds.). Greenland Ice Cores: Geophysics, Geochemistry and Environment: American Geophysical Union Monograph 33: 57–65.

Ruddiman, W. F. & McIntyre, A. 1981. The North Atlantic ocean during the last deglaciation. – Palaeogeography, Palaeoclimatology, Palaeoecology 35: 145–214.

Shackleton, N., Imbrie, J. & Hall, M. A. 1983. Oxygen and carbon isotope record of East Pacific core V19-30: implications for the formation of deep water in the late Pleistocene. – Earth and Planetary Science Letters 65: 233–244.

Shoji, H. & Langway Jr., C. C. 1984. Flow behavior of basal ice as related to modelling considerations. – Annals of Glaciology 30: 141–148.

Staffelbach, T., Stauffer, B. & Oeschger, H. 1988. A detailed analysis of the rapid changes in ice-core parameters during the last ice age. – Annals of Glaciology 10: 167–170.

Thompson, L. G., Mosley-Thompson, E., Davis, M. E., Bolzan, J. F., Dai, J., Yao, T., Gundestrup, N., Wu, X., Klein, L. & Xie, Z. 1989. Holocene-Late Pleistocene climatic ice core records from Qinghai-Tibetan plateau. – Science 246: 474–477.

Zahn, R. 1992. Deep ocean circulation puzzle. – Nature 356: 744–746.

Instructions to authors

Two copies of the manuscript, each complete with illustrations, tables, captions, etc. should be sent to the Secretary, Kommissionen for videnskabelige Undersøgelser i Grønland. Manuscripts will be forwarded to referees for evaluation. Authors will be notified as quickly as possible about acceptance, rejection, or desired alterations. The final decision on these matter rests with the editor.

Manuscripts corresponding to less than 16 printed pages (of 6100 type units) including illustrations are not accepted, unless they are part of a special theme issue. Manuscripts that are long in relation to their content will not be accepted without abridgement.

Manuscript

Language. – Manuscripts should be in English (preferred language), French, or German. Authors who are not writing in their native language must have the language of their manuscript corrected before submission.

Place names. – All Greenland place names used in the text and in illustrations must be names authorised by The Greenlandic Language Committee. Authors are advised to submit sketch-maps with all required names to the Secretary for checking before the manuscript is submitted. Names of Greenland localities outside the area with which the paper is concerned should be accompanied by coordinates (longitude and latitude).

Title. – Titles should be as short as possible, with emphasis on words useful for indexing and information retrieval.

Abstract. – An abstract in English must accompany all papers. It should be short (no longer than 250 words), factual, and stress new information and conclusions.

Typescript. – Typescripts must be clean and free of handwritten corrections. Use double spacing throughout, and leave a 4 cm wide margin on the left hand side. Avoid as far as possible dividing words at the right-hand end of a line. Consult a recent issue for general lay-out.

Page 1 should contain 1) title, 2) name(s) of author(s), 3) abstract, 4) key words (max. 10), 5) author's full postal address(es). Manuscripts should be accompanied by a table of contents, typed on separate sheet(s).

Underlining should only be used in generic and species names. The use of italics in other connections can be indicated by wavy line in pencil under the appropriate words.

Use at most three grades of headings, but do not underline. The grade of heading can be indicated in soft pencil in the left hand margin of one copy of the typescript. Avoid long headings.

Floppy disc. – It may be helpful in the printing procedure if, in addition to the hard copies, the manuscript is also submitted on a DOS-formatted floppy disc. However, editing will be made on the hard copy, and the text file on the disc must be identical to the final version of the manuscript.

References. – References to figures and tables in the text should have the form: Fig. 1, Figs 2–4, Table 3. Bibliographic references in the text are given thus: Shergold (1975: 16) ... (Jago & Daily 1974b).

In the list of references the following style is used:

Boucot, A. J. 1975. Evolution and extinction rate controls. – Elsevier, Amsterdam: 427 pp.

Sweet, W. C. & Bergström, S. M. 1976. Conodont biostratigraphy of the Middle and Upper Ordovician of the United States midcontinent. – In: Bassett, M. G. (ed). The Ordovician System: Proceedings of a Palaeontological Association symposium, Birmingham, September 1974: 121–151. University of Wales Press.

Tarling, D. H. 1967. The palaeomagnetic properties of some Tertiary lavas from East Greenland. – Earth and Planetary Science Letters 3: 81–88.

Meddelelser om Grønland, Geoscience (Bioscience, Man & Society) should be abbreviated thus: *Meddr Grønland, Geosci. (Biosci., Man & Soc.)*

Illustrations

General. – Submit two copies of all diagrams, map, photograph, etc., all marked with number and author's name. Normally all illustrations will be placed in the text.

All figures (including line drawings) must be submitted as glossy photographic prints suitable for direct reproduction, and preferably have the dimensions of the final figure. Do not submit original artwork. Where appropriate the scale should be indicated on the illustration or in the caption.

The size of the smallest letters in illustrations should not be less than 1.3 mm. Intricate tables are often more easily reproduced as text figures than by type-setting; when lettering such tables use "Letraset" or a typewriter with carbon ribbon.

Colour plates may be included at the author's expense, but the editor must be consulted before such illustrations are submitted.

Size. – The width of figures must be that of a column (76.5 mm), 1½ columns (117 mm), or a page (157 mm). The maximum height of a figure (including caption) is 217 mm. Horizontal figures are preferred. If at all possible, fold-out figures and tables should be avoided.

Caption. – Captions to figures must be typed on a separate sheet and submitted, like everything else, in duplicate.

Proofs

Authors receive two page proofs. Prompt return to the editor is requested. Only typographic errors should be corrected in proof; the cost of making alterations to the text and figures at this stage will be charged to the author(s).

Twenty-five copies of the publication are supplied free, fifty if there are two or more authors. Additional copies can be supplied at 55% of the retail price. Manuscripts (including illustrations) are not returned to the author after printing unless specifically requested.

Copyright

Copyright for all papers published by Kommissionen for videnskabelige Undersøgelser i Grønland is vested in the Commission. Those who ask for permission to reproduce material from the Commission's publications are, however, informed that the author's permission must also be obtained if the person is still alive.

Meddelelser om Grønland

Geoscience 29 · 1992

1992

Meddelelser om Grønland

Bioscience
Geoscience
Man & Society

Published by
The Commission
for Scientific
Research
in Greenland

AiO Print as, Odense, Denmark