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SHORT CONTRIBUTION

Greenland palaeotemperatures derived from GRIP bore hole temperature and ice core isotope profiles

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ABSTRACT

Modeling the temperature profile along the GRIP deep bore at the summit of the Greenland ice sheet leads to conversion factors that allow interpretation of the dated stable isotope profile as a climatic temperature record spanning the last 113,000 years. When corrected for surface elevation changes, the late glacial to Boreal temperature shift appears to have been 22°C in central Greenland. The warming at the end of the last glaciation probably began earlier in Greenland, than in Antarctica.

1. Introduction

It has been known for several decades that the isotopic composition, δ, of precipitation[†] changes with the temperature of formation of the precipitation (Dansgaard, 1953). A linear relationship between present mean annual values of temperature and δ ($^0/_{00}$) of precipitation was set up for temperate to polar climates (Dansgaard, 1964), later slightly modified for present Greenland ice sheet surface temperature T_S (°C) and precipitation (Johnsen et al., 1989) into

$$T_S = 1.50\delta + 20.3 \qquad (1)$$

by Johnsen et al. (1989). Based on deuterium excess considerations, they identified subtropical North Atlantic waters as the dominating moisture source for Greenland ice sheet precipitation under interglacial as well as glacial conditions, which contradicts the suggestion of Charles et al. (1994) that shifting moisture sources could be an important part of the explanation of the strongly varying δ values of ice formed during the last glaciation.

Due to the lack of a *temporal* T_S to δ relationship, δ records along Greenland ice cores have not been converted into palaeotemperature records in general. Nevertheless, if applied to δ records along deep Greenland ice cores (Dansgaard et al., 1971; 1982; Johnsen et al., 1992), eq. (1) suggests Late Glacial Temperature Minimum (LGTM) to Holocene shifts in T_S of 16°C at Camp Century, Northwest Greenland, 11°C at Dye 3 in South Greenland, and 9°C at the European Greenland Ice-core Project (GRIP) field station Summit in Central Greenland (72.57°N, 37.62°W, 3246 m a.s.l.). None of these temperature shifts were corrected for changing surface elevation, however, and they should therefore not be considered representative of the LGTM to Holocene climatic shift in Greenland.

However, studies on the temporal relationship have been made in recent years. Comparison of seasonal δ variations in snow on the ice sheet with observed air temperatures on the Greenland east coast (Steffensen, 1985), or with ice sheet surface temperatures (T_S) observed from satellites (Shuman et al., 1995), suggested the use of eq. (1) also for temporal changes, perhaps with a sensi-

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[†] Here denoted by $\delta^{18}O$, or just δ, which is the relative deviation of the $^{18}O/^{16}O$ ratio from the $^{18}O/^{16}O$ value of SMOW (Standard Mean Ocean Water).

tivity, $dT_S/d\delta$, slightly higher than 1.50°C per ‰. Modeling the measured temperature profile along the 400 m long Crête ice core also suggested a higher sensitivity (Johnsen, 1977), and a similar method applied to the upper part of the American GISP2 ice core indicated a sensitivity in the range 1.52 to 2.22°C per ‰ (Cuffey et al., 1994). Outside Greenland, a comprehensive study of the material collected under the IAEA/WMO precipitation network shows a sensitivity as high as 3°C per ‰ for seasonal variations at mid and high latitudes, but only 1.7°C per ‰ for long-term trends through the last three decades (Rozanski et al., 1993).

As to temperature profiles along deep ice cores, they depend on the geothermal heat flux, the ice flow pattern, and the surface temperature and accumulation rate histories. The latter parameters imply the possibility to deconvolute measured temperature profiles into climatic records. Dahl-Jensen and Johnsen (1986) reproduced the observed temperature profile along the deep Dye 3 bore hole within a measuring accuracy 0.03°C (Gundestrup et al., 1994) by using a combined ice flow/heat transport model, and assuming a mean ice age temperature in the 110 to 12 kyrs BP time interval 12°C colder than now, a mean accumulation rate 50% lower than now, and a geothermal heat flux of 42 mW/m^2. It should be noted, however, that the model did not account for surface elevation changes, and that the derived shift from mean glacial to mean Holocene temperature is not identical with the LGTM to Holocene shift dealt with here. The LGTM probably occurred a few thousand years prior to the late glacial extension maximum of the ice sheet (LGEM).

This work deals with the interpretation of the δ GRIP record (Dansgaard et al., 1993). The aim is to establish the central Greenland temperature history back to the marine isotope stage (MIS) 5d, 113 kyr BP, i.e., the time span through which the δ record is representative of climate changes in Greenland and the North Atlantic Ocean (Dansgaard et al., 1971; Grootes et al., 1993; Bond et. al., 1993; McManus et al., 1994). The preceeding warm MIS 5e (the Eem interglaciation) is disregarded, because the probable extinction of the southern dome of the Greenland ice sheet may have radically changed the climatic conditions on the northern dome, which complicates the modeling.

2. Results

The first step is to find the temporal relationship between δ and T_S by modeling the temperature profile along the 3029 m deep GRIP bore hole at Summit (Fig. 1), and the next step is to convert the δ profile into a palaeotemperature record, corrected for changes in surface elevation and isotopic composition of sea water.

A combined ice flow/heat transport model is developed on the basis of the following.

(a) An ice deformation model (Johnsen and Dansgaard, 1992) modified to account for the relationship between surface accumulation rate (λ) and δ, which amounts to 8% change in per ‰ change in δ during the Holocene (Clausen et al., 1988) and at least double as much under glacial conditions (Dahl-Jensen et al., 1993; Kaspner et al., 1995):

$$\lambda(\delta) = 0.23 \exp(-10.09 - 0.653\delta - 0.01042\delta^2)$$

m of ice per year. (2)

This relationship was used to slightly adjust the chronology applied by Dansgaard et al. (1993) into that used in this paper.

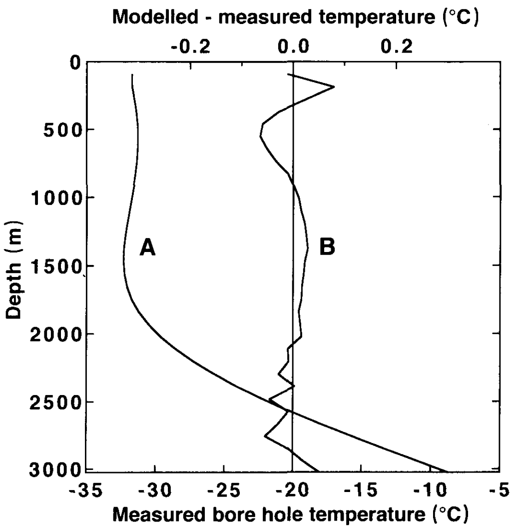

Fig. 1. Curve A: Measured temperature profile along the GRIP Summit bore hole (scale at bottom). The calculated profile is undistinguishable from A. The deviations are shown in curve B (extended scale on top).

(b) The non-steady state equation of heat transfer (Johnsen, 1977)

$$\frac{\partial T}{\partial t}=\kappa\frac{\partial^2 T}{\partial z^2}-v_z\frac{\partial T}{\partial z}+\frac{1}{c}\frac{dk}{dT}\left(\frac{\partial T}{\partial z}\right)^2, \qquad (3)$$

T being the in situ ice temperature, t the time, κ the thermal diffusivity of ice, z the depth, v_z the vertical ice velocity, c the specific heat capacity of ice, and k the coefficient of heat conduction in ice. The horizontal velocity and the internal heat production have both been put to zero, which should apply to the Summit location.

The surface temperature history at Summit is assumed to be related to the dated δ profile along the GRIP deep ice core as

$$T_S=\alpha+\beta\delta+\gamma\delta^2, \qquad (4)$$

δ being the $\delta^{18}O$ value of the ice, corrected for possible deviation of the isotopic composition of sea water from that of SMOW at the time of deposition (Sowers et al., 1993). α, β and γ are unknown constants to be determined along with the geothermal heat flux q, by seeking the best possible fit to the measured temperature profile along the GRIP deep hole (Fig. 1), which holds at least 8 degrees of freedom, according to a Backus-Gilbert type analysis.

In the first experiment, integration of eq. (3) is done with the assumption of constant ice sheet thickness h. The integration is extended through several past glacial cycles (repetitions of the last one), and through a 3000-m thick layer of bedrock, q being applied at the bottom of this layer. The result is a set of constants (of which $q=51.7\ \mathrm{mW/m^2}$) for best fit to the measured profile (standard error ± 0.092°C), corresponding to a $\mathrm{d}T_S/d\delta$ sensitivity of 2.7°C/‰ under present conditions ($\delta\approx -35.2$ ‰), and 3.5°C/‰ during the LGTM ($\delta\approx -42$ ‰).

In the second experiment, a simple procedure is introduced in order to account for the effect of changing ice surface elevation at Summit. We consider an elongated, idealized steady state ice sheet of half-width w, maximum thickness h at the ridge, and surface accumulation rate λ m ice per year. These parameters are constrained by the equation

$$\lambda\cdot w^4=b\cdot h^8 \qquad (5)$$

(Vialov, 1958), b being a constant to be determined from present day values.

The steady state half-width w is influenced, not only by λ, but also by the bottom topography and the melting in the coastal areas, i.e., by the coastal temperature, which, in turn, influences λ (cf. eq. 2), colder climate being associated with lower λ. These other factors are accounted for to some degree by defining the present and the most advanced position of the ice margin, particularly in the third experiment described below, where these positions are based on field observations (Weidick, 1993). For simplicity, w is made linearly dependent on the coastal temperature $T_C(t)$ at time t and low elevation. $T_C(t)$ is derived from the calculated Summit temperature at time t using a lapse rate of -1.0°C/100 m and the calculated elevation at Summit. The actual half-width w is subsequently adjusted with a time constant of 3000 years. We take $w=500$ km at present for $T_C=-5$°C, and a maximum value $w=550$ km, when eq. (4) gives $T_C\leqslant -15$°C, which essentially excludes melting in the summer time. Furthermore, λ is supposed to vary according to eq. (2). Eq. (5) is used within each time step $\mathrm{d}t$ with $w=w(t)$ to calculate the steady-state accumulation rate; the ice thickness change $\mathrm{d}h$ is subsequently found as $(\lambda(\delta)-\lambda_s)\,\mathrm{d}t$. The bedrock elevation in the Summit area is finally adjusted for isostatic movements with a time constant of 3000 years, and corrected for changing sea level (Sowers et al., 1993) to give the surface elevation. The narrow restraint on w makes h mainly dependent on λ, and therefore leads to a thinning of the ice sheet under glacial conditions (Cutler et al., 1995) relative to present, in our experiment 2880 m during LGTM. This possibility is not excluded, but the temperature profile modeled with these assumptions does have a poorer fit to the measured one (standard error ± 0.12°C). The geothermal heat flux comes out as $53.3\ \mathrm{mW/m^2}$, and the resulting $\mathrm{d}T_S/\mathrm{d}\delta$ sensitivities as 2.9 and 3.6°C/‰, for present and LGTM conditions, respectively.

In the third experiment, w is allowed to increase linearly with $T_C(t)$ to a maximum value of 700 km for $T_C(t)\leqslant -15$°C, corresponding to a position at the continental margin. The intervening calculated ice margin positions from the LGEM until today are in accordance with ice margin data of Weidick (1993). The calculated ice thicknesses/surface elevations vary from 3100/3300 m during LGTM

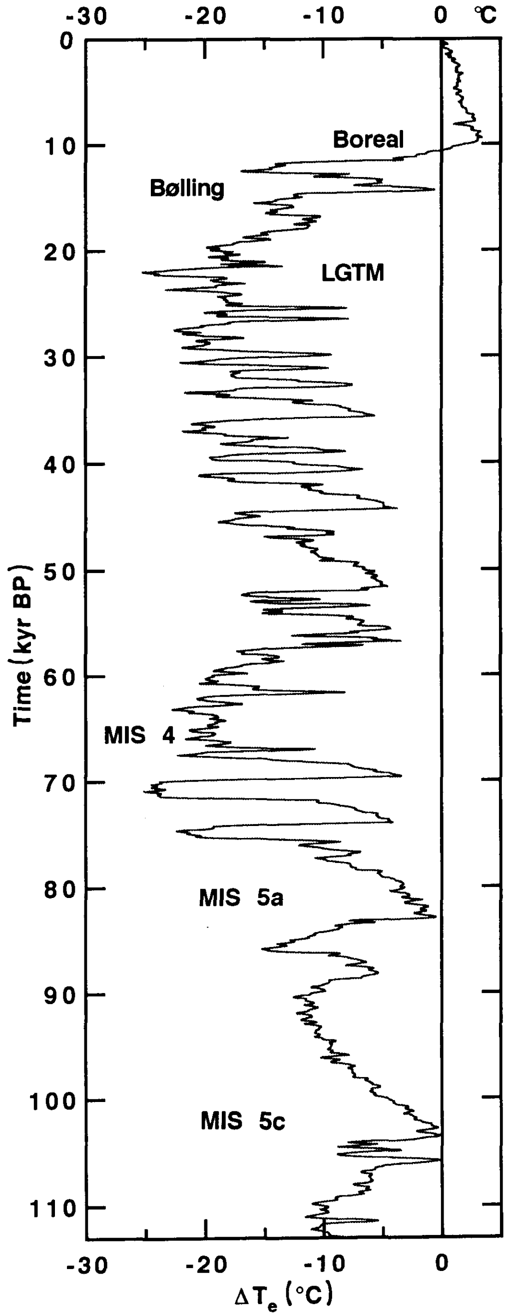

Fig. 2. Calculated central Greenland temperature deviations from present, adjusted to the present elevation (3240 m) of the summit of the Greenland ice sheet. The record spans the last 113 kyr (113,000 years). MIS stands for Marine Isotope Stage; LGTM for Late Glacial Temperature Minimum.

to 3240/3420 m in Boreal. It was only after Boreal that the ice sheet thinned considerably to the present thickness of 3029 m.

A highly improved best fit (standard error ± 0.036°C; Fig. 1) is obtained with the constants $\alpha = -211.4$°C, $\beta = -11.88$°C/‰, $\gamma = -0.1925$°C/(‰)^2, and $q = 49.2$ mW/m^2. The $dT_S/d\delta$ sensitivities become 1.67 ± 0.05 and 4.33 ± 0.04°C/‰, for Holocene and LGTM conditions, respectively.

The conversion of the GRIP δ record into a T_S record is now completed by eq. (4), and the T_S record is finally corrected for surface elevation changes to establish the palaeoclimate record for central Greenland shown in Fig. 2.

3. Discussion

A common result of the three experiments is the high sensitivity $dT_S/d\delta^{18}O$ (>3°C per ‰) for glacial conditions, and the high glacial to post-glacial temperature increase of more than 20°C (cf. Fig. 2), in agreement with an independent estimate by Cuffey et al. (1995), based on GISP2 data.

Several features strengthen the validity of the results of experiment 3.

(a) The extreme cold in the late stage of the glaciation makes the then Geenland look similar to present day Antarctica, i.e., little or no melting at the ice margin. This explains why the ice margin advanced to cover the continental shelf (Weidick, 1993), as assumed only in experiment 3.

(b) The fit between the measured and calculated temperature profiles is by far the best in experiment 3.

(c) When the changing isotopic composition of sea water is taken into account, the fit improves significantly (35%) in experiment no. 3, but it becomes worse in the two other experiments.

(d) Experiment no. 3 gives a δ to temperature calibration factor for the Holocene ($dT_S/d\delta^{18}O = 1.67$°C per ‰), which is remarkably close to other estimates based on recent mid- and high-latitude precipitation, as mentioned in the introduction.

(e) The calculated temperature deviations during the last glacial climatic oscillation, Oldest Dryas/Bølling/Younger Dryas/Boreal, are approximately $-15/-1/-15/+3$°C, which vary in accordance with the corresponding mean annual temperatures in Britain ($-9/+8/-7/+9$) based on studies of beetle remains (Atkinson et al., 1987). This supports the validity of eq. (4) also for short-term glacial changes. Furthermore, it shows that the dramatic temperature shifts during the glaciation (Fig. 2) are not restricted to Greenland. Finally, the agreement between calculated temperatures in Greenland and western Europe further weakens the argument (Charles et al., 1994) that the Greenland δ records could be influenced considerably by changing sources of Greenland precipitation.

It appears from Fig. 2 that:

(a) the post-glacial climatic optimum (Boreal) occurred 8000 to 10,000 years ago with Greenland temperatures up to 3 to 4°C higher than now;

(b) the temperature reached present day values in short periods during the Bølling, MIS 5a and MIS 5c;

(c) temperatures more than 20°C lower than now occurred several times during the last glaciation with absolute minima 25°C colder than today 21,500 and 71,000 years BP;

(d) the latter minimum ended with a 20°C warming within a century;

(e) the glacial to Holocene transition in Greenland already began 19,000 years B.P., i.e. slightly earlier than in Antarctica, according to the Sowers and Bender (1995) comparison of the GISP II and Byrd δ records, which are presented in a somewhat different chronology.

4. Conclusion

The technique outlined in this paper allows a calibration of central Greenland ice core isotope records in terms of past surface temperatures. When corrected for surface elevation changes, they provide the central Greenland, perhaps even the North Atlantic, climate history, which appears to have been more dramatic during the glaciation than hitherto assumed. Further modelling and total gas content data (Raynaud and Lebel, 1979) are needed, however, to improve the knowledge of surface elevation changes, and to make it possible to extend the convertion through the Eem interglaciation (MIS 5e).

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