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Holocene paleoenvironmental reconstruction from deep ground temperatures: a comparison with paleoclimate derived from the δ^{18}O record in an ice core from the Agassiz Ice Cap, Canadian Arctic Archipelago

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ABSTRACT. Changes in ground-surface temperature for the past few hundred years have been derived from deep temperature profiles at three wells in the north-eastern Canadian Arctic Archipelago, and compared with the climatic history derived from the oxygen-isotope ratio $^{18}O/^{16}O$ measured in an ice core from the Agassiz Ice Cap, about 180–260 km to the east. Analysis of the ground-temperature profiles suggests that surface temperatures in the area decreased after the Little Climatic Optimum about 1000 years ago until the Little Ice Age (LIA). About 100 years ago, ground-surface temperatures appear to have increased by 2–5 K to reach today's values, while air temperatures increased by 2–3 K, according to the isotope record. Part of the larger ground-surface temperature change may be due to other paleoenvironmental effects, such as an increase in snow cover coincident with the end of the LIA.

The δ^{18}O climatic record was successful in predicting the general features of the ground-temperature profiles observed at two of the sites, but not the third. There is contemporary evidence that surface temperatures at the latter site may be substantially modified by other environmental factors such as snow cover.

INTRODUCTION

The δ^{18}O record from cores taken in present-day ice caps serves as a proxy paleoclimatic temperature record, through the association of the ratio of oxygen isotopes ^{18}O and ^{16}O to air temperatures at the time of condensation. Climatic change may also be preserved as a signal in temperature–depth profiles within the ground, not as a proxy indicator of past climate but as a direct consequence of the impact of past air-temperature variations and associated effects upon the ground surface (Lachenbruch and Marshall, 1986).

In the Queen Elizabeth Islands of the Canadian Arctic Archipelago (Fig. 1), there is an opportunity to compare climatic changes inferred from oxygen-isotope ratios in an ice cap to paleoenvironmental conditions derived from nearby ground-temperature profiles. δ^{18}O records for the past 10 000 years are available from the Agassiz Ice Cap in the northwest of the region (Fisher and others, 1983). Precision ground temperatures to depths of up to 1000 m are available from 40 petroleum exploration wells throughout the Archipelago (Fig. 1; Taylor and others, 1982) and three of these lie within 260 km of the Agassiz Ice Cap site. In this paper, we extract possible paleoclimatic or paleoenvironmental signals from ground temperatures measured at these three wells, and compare with the climatic signal inferred from the oxygen-isotope ratios measured in a core on the Agassiz Ice Cap. We restrict our study to the past few centuries.

The geothermal analysis is predicated on the fundamental hypothesis that the terrestrial heat flow, which arises largely from the decay of radioactive elements within the crust, is essentially constant in the upper few kilometres. But at many wells, the heat flow is observed to vary systematically with depth. While more random variations may be attributed to measurement errors in temperature or thermal conductivity, and corrections may be made for local effects like topography (e.g. Jaeger, 1965), the residual coherent "long-wavelength" variation may be attributed to transient effects arising from climate change.

Simple techniques are used to extract this signal, and we compare it with the climatic signal derived from the oxygen-isotope ratios. The analysis involves the familiar heat-conduction equations used in glaciological work for considering climate change (e.g. Robin, 1955; Paterson,

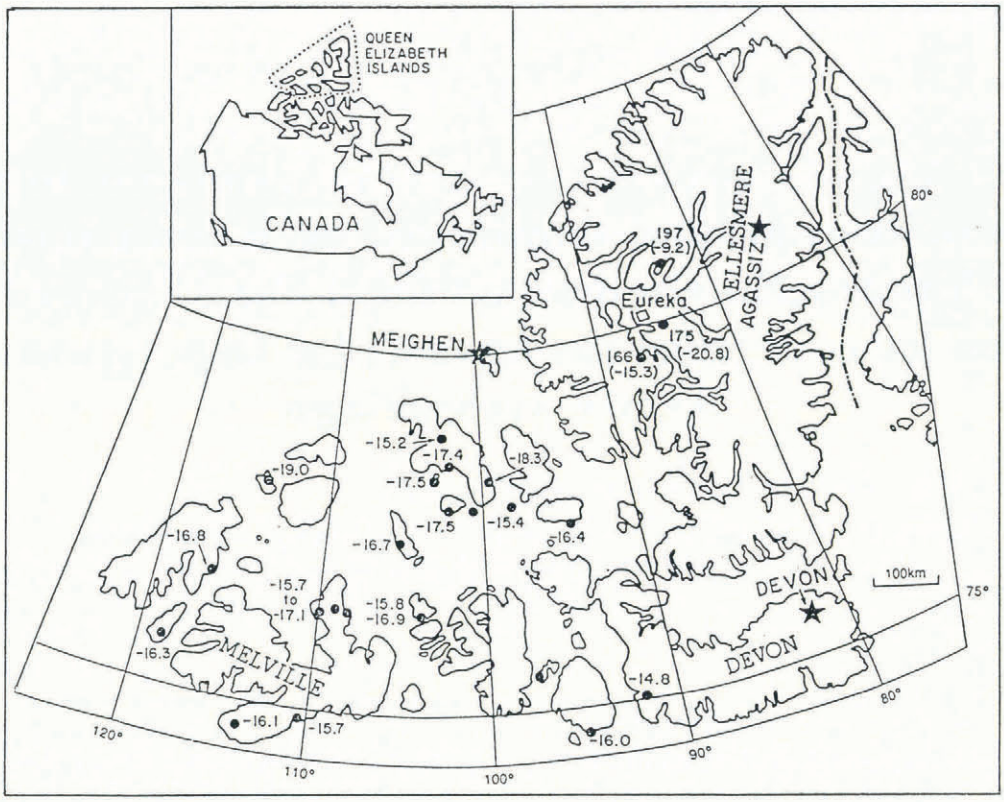

Fig. 1. Wells in the Queen Elizabeth Islands of the Canadian Arctic Archipelago where precise ground temperatures have been measured (solid dots). The mean ground-surface temperature obtained from the extrapolation of the upper 100 m of the temperature–depth profile is given at each site in °C. The location of drillholes in the Agassiz, Devon and Meighen Ice Caps are shown by asterisks. Wells used in this study are located in the northeastern region around the Eureka weather station and near the Agassiz Ice Cap; these wells are identified by a three-digit number (197, 175, 166), keyed to well numbers in Taylor and others (1982).

1968; Weertman, 1968; Paterson and Clarke, 1978; Budd and Young, 1983; Jenssen and Campbell, 1983), but the problem is much simplified since there are generally no time-dependent physical properties. Permafrost is 500 m or more deep at these wells, and we assume heat transport is by conduction (Lachenbruch and Marshall, 1986).

THE DATA

The oxygen-isotope data

We use the $\delta^{18}O$ record for the past 1000 years from the Agassiz 1979 drillhole (Fig. 2; Fisher and others, 1983). The suitability of oxygen-isotope data as indicators of regional paleoclimate is discussed in the literature (e.g. Paterson and Clarke, 1978; Fisher and others, 1983; Robin, 1983) and we assume a $\delta^{18}O$–T relationship of 0.6‰ per 1 K (e.g. Koerner and Russell, 1979).

Ground temperatures

Precision ground temperatures have been measured at 40 holes drilled in the course of petroleum exploration in the Canadian Arctic Archipelago, as part of a permafrost research program of the Canadian Department of Energy, Mines and Resources (Fig. 1). The preservation of these wells for temperature logging following drilling and the measuring technique have been reported by Judge (1974). Generally, following completion of drilling and testing operations on a well, permission to omit the regulatory surface plug was obtained and the drilling mud was displaced with diesel fuel. This permitted subsequent temperature logging of the well annually for 10 years or more, using a light-weight probe system; alternatively, a multithermistor cable was left in some wells by the operator (e.g. well #166). The temperature data have been published at regular intervals (e.g. Taylor and others, 1982). While a measurement accuracy of 0.01 K is attained through use of precision calibrated thermistors, several of the wells exhibit instabilities of up to 0.1 K due to convection in sections of high geothermal gradient; these are averaged in the measurement procedure and do not appreciably alter the general temperature variation that is analyzed here. Also, logs taken shortly after well completion reflect a considerable degree of thermal disturbance due to the drilling operations. This disturbance may be assessed and corrections made based on the temperatures taken over several years, if warranted (Lachenbruch and others, 1982). We used the last temperature log measured at each well, and assured ourselves that there was no residual curvature in the temperature–depth profiles attributable to the drilling about 10–13 years earlier.

In this study, we consider three wells, #197, #175 and #166, lying about 180–260 km from the Agassiz Ice Cap (Fig. 1). The temperature–depth profile and a simplified lithology for each of these wells are shown in Figure

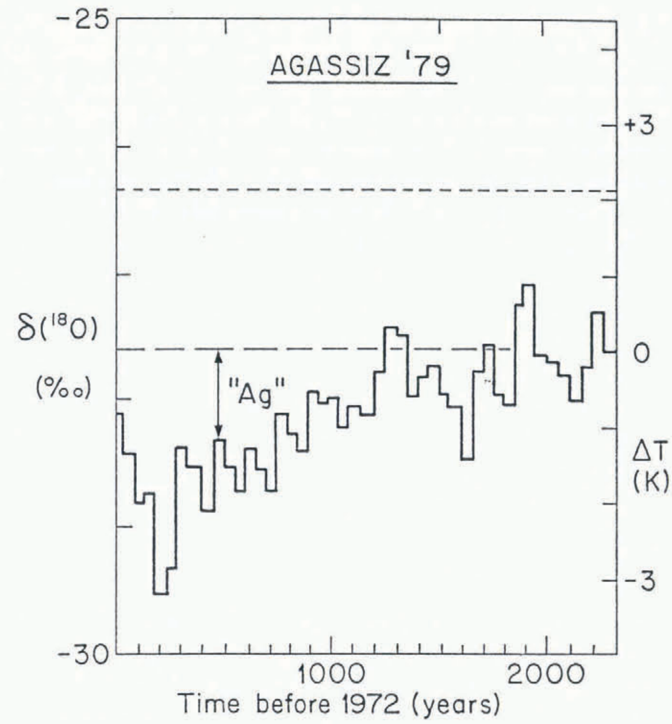

Fig. 2. The paleoclimate record for the past 2000 years derived from the oxygen-isotope record at the 1979 drillhole on the Agassiz Ice Cap (Fisher and others, 1983). The temperature conversion (righthand scale) assumes $0.6‰K^{-1}$. The difference between the $\delta^{18}O$ curve and the reference level, the long-dashed line, represents the past temperature variations relative to "today's" value. The short-dashed line represents a reference level 2 K higher (see text).

3. Well #197, Gulf Neil O-15, lies near the centre of a peninsula north of Greely Fiord, about 180 km west of the Agassiz 1979 borehole. Over the temperature–depth section considered, sandstones and siltstones predominate (70%) but are interbedded with shale (27%) and other lithologies. Well #175, Panarctic Gemini E-10, lies about 37 km east of the Eureka weather station. The lithology consists largely of shale (70%) and sandstone (24%). Well #166, Imperial Mokka A-02, lies about 4 km from the western coast of the Mokka Peninsula, Axel Heiberg Island. The lithology is shale.

Thermal conductivity of the formations

For the analysis, representative thermal conductivity measurements at approximately 30 m intervals in depth were made in our laboratory by the divided bar technique using chip samples recovered during the drilling procedure (Sass and others, 1971a; Jessop, 1990, chapter 2.9). While thermal conductivities of most lithologies can be measured by this technique within 10%, shales create special problems and conductivities measured in the laboratory are biased on the high side (see discussion and references in Taylor and others (1989)). Because shales dominate in wells #166 and #175, a synthetic conductivity profile was constructed for them using the lithologic description provided by the well operator (see below). Measured thermal conductivities for well #197,

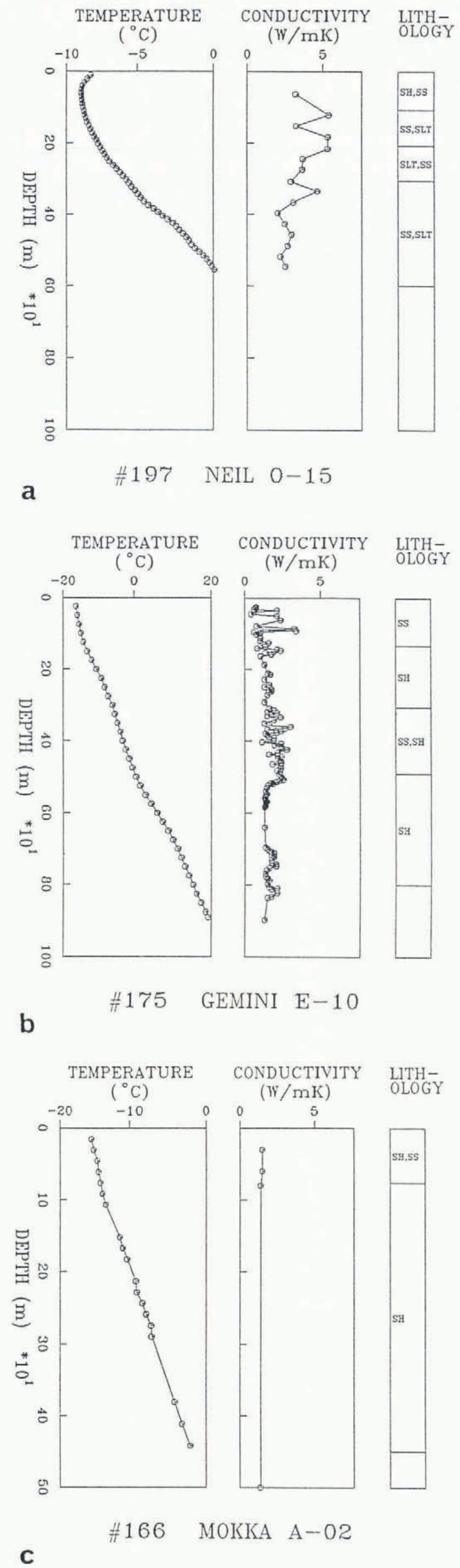

Fig. 3. Temperature, thermal conductivity and simplified lithology profiles at the three wells analyzed here. For locations, see Figure 1. For the lithology, SS, sandstone; SH, shale; SLT, siltstone.

and synthetic profiles for #166 and #175, are shown in Figure 3.

METHOD

The thermal regime of the upper few kilometres of the Earth is constrained by the deep equilibrium thermal regime and temperatures at the ground surface. The steady flow of heat Q from the interior constitutes the deep boundary condition:

$$Q = kG \tag{1}$$

where G is the equilibrium temperature–depth gradient and k is the thermal conductivity. Q is constant at any particular site. For a surface temperature T_0, the equilibrium temperature–depth profile $T(z)$ is given by

$$T(z) = T_0 + Gz \tag{2}$$

and is linear with depth z below the ground surface if k is a constant.

The equilibrium regime rarely is attained in practice. Changes in the ground-surface temperature occur due to climatic variations, environmental changes and other causes; these transient effects propagate slowly into the ground and persist for a period of time, due to the low thermal diffusivity of the ground and its high heat capacity. We assume that for time intervals $(0, t_1), (t_1, t_2), \ldots, (t_{n-1}, t_n)$ measured backwards from the present, surface temperatures differed from the present value by fixed amounts $0, T_1, \ldots, T_{n-1}$, respectively. Deep ground temperatures measured today are a superposition of the equilibrium state (Equation (2)) and a weighted summation of the effect of these changes in ground-surface temperatures (Jaeger, 1965, equation 28):

$$T(z) = T_0 + Gz + \sum_n t_n[\mathrm{erf}\{z/2(\alpha t_n)^{\frac{1}{2}}\} - \mathrm{erf}\{z/2(\alpha t_{n+1})^{\frac{1}{2}}\}] \tag{3}$$

where α is the thermal diffusivity of the ground and erf is the error function.

The extraction of past surface-temperature variations from ground temperatures measured today depends critically on the separation of equilibrium conditions (the first two terms in Equation (3)) from transient effects (the summation terms). There are two ways of specifying equilibrium conditions: (1) With independent knowledge of the deeper geothermal regime gained, for instance, from terrestrial heat-flow studies on deeper parts of the profiles considered here or on much deeper bottom-hole temperatures obtained during drilling, an equilibrium temperature profile may be hypothesized, based on Equations (1) and (2). This temperature profile may be constructed to pass through the surface-temperature intercept of the measured data or some deeper part of the measured profile. (2) A long, linear section of the measured temperature–depth profile may be assumed to represent a long-term trend and extrapolated to intersect the surface. In the present analysis, any departure of the measured temperatures from the constructed "equilibrium" profile is attributed to paleoenvironmental variation. The second approach has been used by Lachenbruch and Marshall (1986) to reconstruct surface-temperature variations during the past century, while the first has been used to demonstrate the sub-surface signature of recent changes in sea level (Taylor, in press).

In either approach, there is a possibility of falsely attributing variations in the temperature profile to past changes in surface temperature, when such variations may arise simply from changes in thermal conductivity (see discussion in Lachenbruch and Marshall (1986)). Because of the known lithologic variations in these wells, thermal conductivity is integrated in the analysis by replacing measured depth (units m) with a thermal depth (units $\mathrm{m^2\,K\,W^{-1}}$; Jaeger, 1965, equation 5):

$$z_{\mathrm{th}} = \int_0^z [\mathrm{d}z/k(z)] \tag{4}$$

where $k(z)$ is the thermal conductivity variation with depth. With this change of variable, a plot of equilibrium temperature versus thermal depth, the Bullard plot, is linear and the slope is the terrestrial heat flow, Q. Outside of errors in temperature and thermal conductivity measurement, any non-linearities may be attributed to local effects such as topography and to transient effects of past changes in surface temperature due to climatic variation and paleoenvironmental effects.

Occasionally, non-linearities are observed in the Bullard plot that bear a strong correlation with the lithology. This suggests error in the measurement of thermal conductivity. This is particularly noticeable in sections where shale formations are juxtaposed with formations predominantly of sandstone and other lithologies (Taylor and others, 1989).

In such cases, a synthetic conductivity profile was constructed for the well, based on the lithological description made by the well operator from samples of the drilling chips. This description provided the percentage of each rock type within the sample, and for each type we adopted a thermal conductivity, chosen from typical values in the literature (Table 1). The weighted mean conductivity for the interval was calculated as follows.

Each rock type i identified within the nth sample was assumed to occupy a discrete layer of thickness l_i based on its percentage occurrence such that

$$l_n = \Sigma l_i$$

where the summation is over all net rock types present and l_n is the thickness of the nth interval. A conductivity k_i was assigned to the rock type i and the effective conductivity for the nth interval is calculated (Jaeger, 1965, equation 7):

$$K_{\mathrm{eff}(n)} = \frac{\Sigma l_i}{\Sigma(l_i/k_i)}\,. \tag{5}$$

Considering successive sample intervals, a synthetic thermal conductivity profile is obtained for the well, and this is used for the analysis where shales dominate the lithology, here in particular for wells #166 and #175 (Fig. 3b and c; the operator's lithologic description was not detailed at #166).

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Table 1. Thermal conductivity values for individual rock types used in constructing synthetic conductivity profiles (Equation (5))

Rock type	*Thermal conductivity, k_i* $W m^{-1} K^{-1}$
Sandstone	3.72[1]
Shale	1.40[2]
Siltstone	2.68[1]
Limestone	2.80[1]
Dolomite	4.68[1]
Coal	0.30[3]

Notes: [1] Table 2 in Reiter and Jessop (1985). [2] Best estimate for shale, Canadian Arctic Archipelago (Taylor and others, 1989). [3] Majorowicz and others, 1988.

We identify the anomalous departure of measured ground temperatures from the hypothesized equilibrium or quasi-equilibrium ground temperatures, as described above, and derive a past ground-surface temperature model in the following manner. The past few centuries are divided into several time intervals and a physically reasonable range of ground-surface temperatures is identified with each interval. An iterative routine systematically takes all combinations of surface-temperature models within these ranges, calculates the sub-surface effect for each using Equation (3), subtracts it from the measured temperatures and calculates an estimate of the heat flow, thus "corrected" for the hypothesized transient effect. Of the many trials, the "best" model is that which reduces the variation with depth in the heat-flow value, as evidenced by a minimum in its standard deviation. Similar approaches have been used by Cermak (1971) and Vasseur and others (1983). The result may be depicted either as a linearization of the Bullard plot or, starting with the hypothesized quasi-equilibrium profile, as a model fit to the measured temperatures. The latter presentation is used here (Figs 4–7).

The surface-temperature model derived in this way is then compared with the climatic model derived from the oxygen-isotope data from the borehole in the Agassiz Ice Cap. We note that the geothermal model is not unique (see discussion in Vasseur and others (1983)). Indeed, a small suite of such models may be obtained by choosing a somewhat different division of past time, and by searching for a surface-temperature model that minimizes the variation in heat flow with depth for this particular division of time. Such alternative models tend to be similar, and illustrate the general trends of past surface-temperature variations that are consistent with the ground-temperature data.

As an alternative approach, we examine the degree to which the paleoclimate derived from the $\delta^{18}O$ data can predict the variations observed in deep ground temperatures. Using the Agassiz 1979 borehole data taken on a 25 year mean basis (Fig. 2), we take the difference between an average $\delta^{18}O$ value for the past 25 years (as reference) and the measured $\delta^{18}O$ values for the past 1000 years. The result is a time series of past departures in air temperature from "today's" value. These are used as the T_n in Equation (3) to predict the subsurface temperatures that would be expected if the air-temperature variations derived from the oxygen-isotope record were the same as the variation in ground-surface temperature at each well site.

THE AGASSIZ ICE CAP $\delta^{18}O$ RECORD AND PALEOENVIRONMENTAL RECONSTRUCTION AT THREE GROUND-TEMPERATURE WELLS

Three ground-temperature holes lie less than 260 km to the southwest of the Agassiz Ice Cap borehole and within 100 km of the Eureka weather station (Fig. 1). Wells #166 and #175 have similar temperatures and temperature–depth gradients, and the profiles intersect the ground surface between −15° and −17°C (Fig. 3). The latter range is generally typical of other sites in the Arctic islands (Fig. 1) and somewhat higher than the 30 year climatic mean temperature at Eureka (−19.4°C). In contrast, profile #197 departs markedly from the other two in gradient and in surface intercept (−8°C).

Well #197

Since sandstone and siltstone predominate in the upper section of this well, measured thermal conductivities were used in drawing the Bullard plot. To extract a surface-temperature history from this profile, we choose a quasi-equilibrium profile and identify the anomalous temperatures; this may be done in at least two ways: (1) A part of the profile immediately beneath the temperature–depth inversion is linear on the thermal depth plot, and may be extrapolated to the surface to represent a long-term trend (Fig. 4a). (2) A deeper section of the measured temperature profile might also be selected as a pseudo-equilibrium profile. If the linear section between 500 and 550 m is extrapolated to the surface, an intercept of −8.4°C, nearly identical to that observed today, is obtained (Fig. 4b). This choice is also supported by terrestrial heat-flow arguments. The heat flow calculated over this deep interval is $51 \pm 2\, mW\, m^{-2}$, comparable to a value of $48 \pm 12\, mW\, m^{-2}$ obtained through an independent analysis of industry temperatures up to 2400 m depth (Jones and others, 1989).

We attribute the departure from the hypothesized long-term profiles to past changes in surface temperature. For the first choice of quasi-equilibrium profile, the analysis (Equation (3)) shows that the higher temperatures in the upper 150 m are consistent with a cumulative increase in ground-surface temperatures of 3.3 K between 70 and 20 years ago (Fig. 4a). For the second choice, a distinct downward concavity to the temperature profile between 150 and 550 m (Fig. 4b) suggests that a period of decreasing temperatures preceded the more recent increase in surface temperature. It may be shown that changes earlier than 1000 years ago create negligible curvature below 500 m so the departure from the second choice of long-term profile will be attributed

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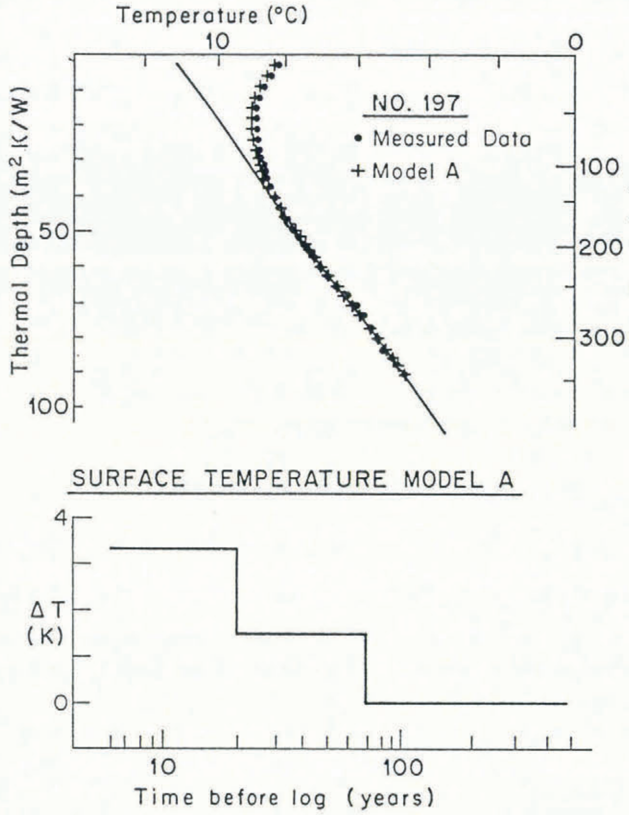

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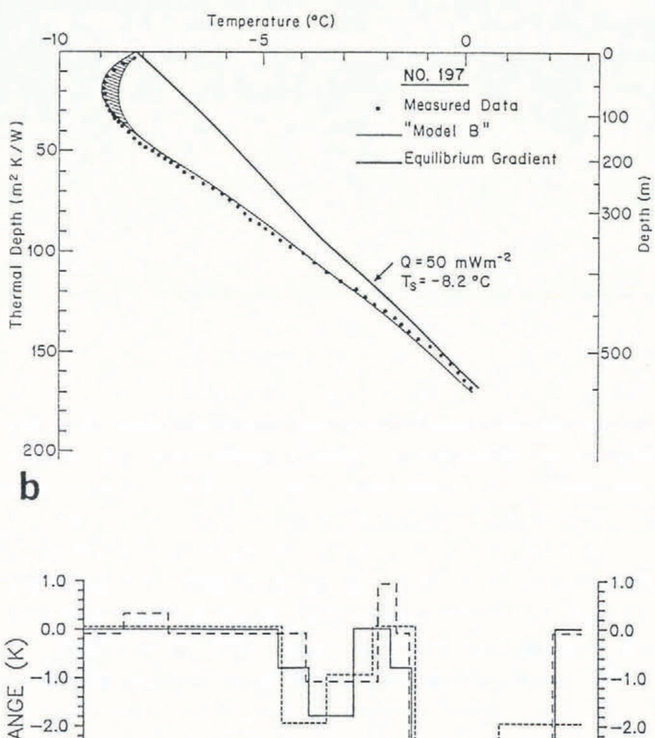

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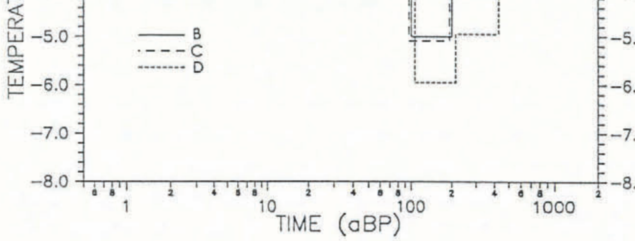

c

◀ *Fig. 4. Analysis of deep ground temperatures measured at well #197, about 180 km west of the Agassiz Ice Cap drillhole (Fig. 1). The temperatures are plotted versus thermal depth, obtained by normalizing depths for measured thermal conductivity; true depth is plotted along the righthand side. The equilibrium line is hypothesized to represent a long-term trend in temperatures. (a) Identification of the anomalous curvature in the upper 300 m and fit of the modelled temperatures to the measured data, for the surface-temperature history shown below. (b) Identification of anomalous curvature in the deeper data and fit of the modelled temperatures to the measured data, using model B (Fig. 4c). (c) Several surface-temperature models for the last 1000 years, all of which are consistent within the resolution of the technique with the curvature apparent in deep ground temperatures at well #197.*

to changes in the past millennium. Using Equation (3), several surface-temperature models have been extracted from the measured temperature profile by varying the time intervals (Fig. 4c). The fit of model B to the measured temperatures is shown in Figure 4b; the contribution of the past century is shown cross-hatched.

The models illustrate various possible past surface-temperature histories that are consistent with the measured ground temperatures. Their trends are similar and suggest that temperatures during the Little Ice Age were up to 5 K lower than today, with most of the recent increase occurring about 70–100 years ago.

We now assess the degree to which the climate inferred from the oxygen-isotope record at the Agassiz Ice Cap predicts the variations in ground temperatures measured at well #197. If the latest 25 year mean $\delta^{18}O$ value ($-27.64‰$, for 1971–46) is taken as a reference temperature for the oxygen-isotope record (Fig. 2), the predicted ground temperatures fall short of matching the measured temperatures (Ag; Fig. 5). Consider-

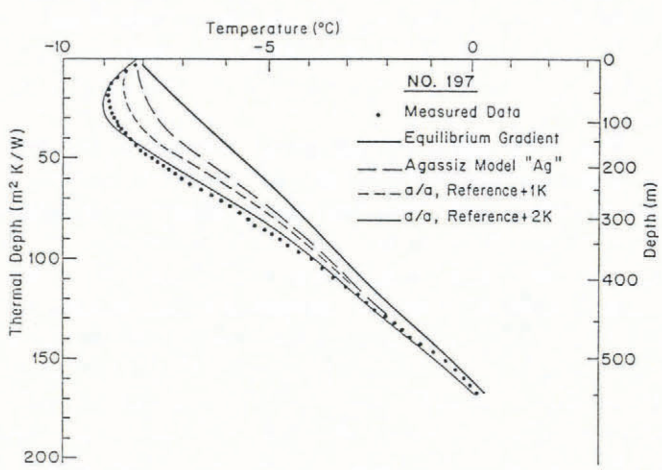

Fig. 5. Comparison of ground temperatures predicted using the Agassiz Ice Cap paleoclimate models (0, 1 and 2 K offsets in reference level; see Figure 2 and text), with measured temperatures at well #197.

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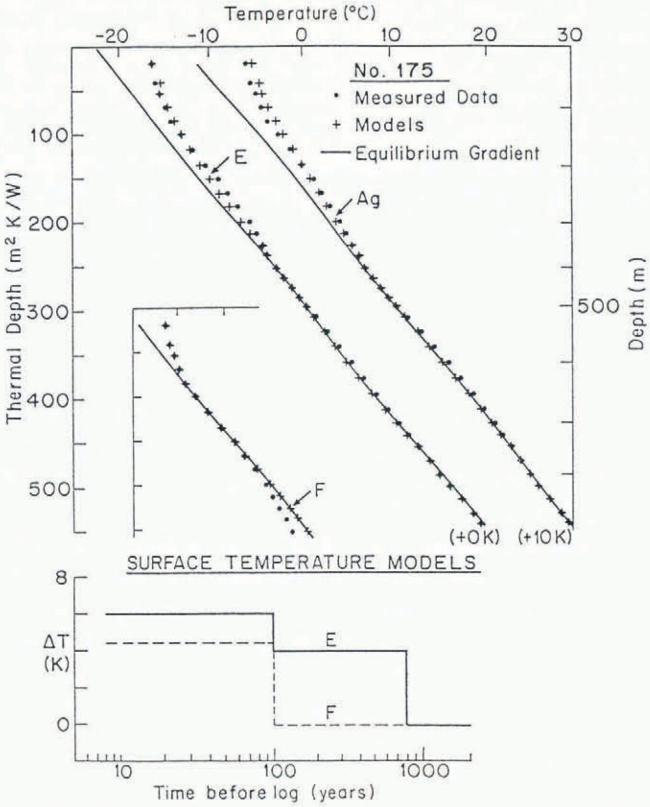

Fig. 6. Lefthand curve and inset: identification of anomalous curvature in ground temperatures at site #175 and fit of the modelled temperatures to the measured data, for surface-temperature histories E and F shown below. Righthand curve (offset 10 K): anomalous ground temperatures predicted from the Agassiz Ice Cap paleoclimate model, Ag, compared with the measured data.

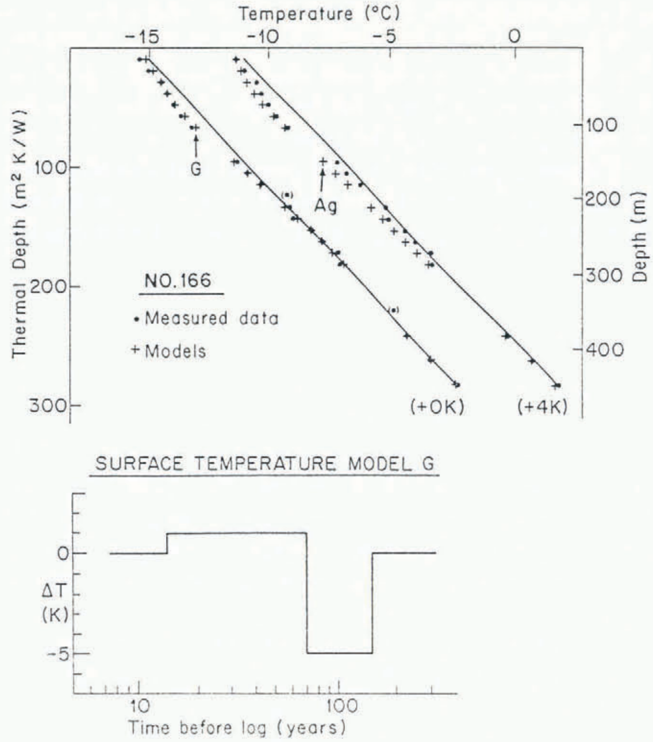

Fig. 7. Lefthand curve: identification of anomalous curvature in ground temperatures at site #166 and fit of the modelled temperatures to the measured data, for surface-temperature history G shown below. Righthand curve (offset 4 K): anomalous ground temperatures predicted from the Agassiz Ice Cap paleoclimate model, Ag, compared with the measured data.

ably less negative $\delta^{18}O$ reference values (e.g. $-26.4‰$ or 2 K higher; Fig. 2) are required to predict the ground temperatures measured at #197 (solid line fit; Fig. 5).

Well #175

The alternating sections of higher and lower gradients apparent in the temperature–depth profile correlate with formations alternating between shale and sandstone lithologies, respectively (Fig. 3b). In a Bullard plot, the correlation persists if measured conductivity values are used but largely disappears if the synthetic conductivity profile is taken (Fig. 6).

A long-term linear trend (400–900 m) may be extrapolated to intersect the ground surface at -23°C; above 400 m, temperatures are higher than this trend line and are consistent (Equation (3)) with an increase in surface temperature of 4 K about 800 years ago and a further 2 K about 100 years ago (model E; Fig. 6). Alternatively, the upper curvature may be modelled relative to a shorter linear part of the temperature profile between 150 and 300 m, which has a surface-temperature intercept of -21°C. Departures from this trend suggest an increase of 4.2 K in surface temperature 100 years ago (model F; Fig. 6 (inset)). The two surface-temperature models are compared in the lower part of Figure 6.

Finally, the Agassiz Ice Cap climatic record has been used with Equation (3) to predict the expected climatic perturbation to the ground temperatures (Ag; Fig. 6); note that the profile is offset 10 K.

Well #166

Some scatter in temperature appears on this log (Fig. 3c) that does not correlate with the coarse lithologic description and is not reduced in the Bullard plot using either measured or synthetic conductivities (Fig. 7). Unlike the other two wells, this profile was obtained from a multithermistor cable left in the well mud following drilling, and some deterioration of the sensors may have occurred because of freeze-back pressures. The synthetic profile has been used in the analysis, as it reflects a more reasonable value of shale conductivity.

A long-term profile may be constructed from a linear section between 200 and 400 m (Fig. 7); the slope gives a terrestrial heat flux of $47 \pm 2\,\mathrm{mW\,m^{-2}}$. From deeper industry temperatures, Jones and others (1989) estimated a heat flux of $48 \pm 12\,\mathrm{mW\,m^{-2}}$ for this well, supporting the choice of the linear trend as a long-term pseudo-equilibrium profile. Measured temperatures above 200 m depart from this trend and are consistent with ground temperatures about 5 K lower than today late in the Little Ice Age, but about 1 K higher for part of the present century. The change in curvature between the upper two

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data points and the points below is modelled in terms of a drop in surface temperature by 1 K about 14 years before the temperature log was taken; this lowering of the surface temperature may be associated with construction of the drilling pad at that time.

Ground temperatures predicted at this site from the Agassiz Ice Cap climate are a few tenths of a degree lower than the measured temperatures between 100 and 200 m (Ag; Fig. 7; offset 4 K).

DISCUSSION

The impact on ground temperatures of the Holocene climate and the Pleistocene glaciations has long been recognized (e.g. Lane, 1923; Hotchkiss and Ingersoll, 1934; Birch, 1948). However, rarely are the ground temperature and thermal property data good enough to extract paleotemperature fluctuations prior to the past few centuries or millennium (e.g. Vasseur and others, 1983); in many cases, there may be little obvious curvature in the temperature–depth profile attributable to a particular event (Sass and others, 1971b), or the background thermal regime may not be known.

The reconstruction of paleoclimate directly from the ground-temperature data lacks the temporal resolution of the $\delta^{18}O$ record, largely due to the property of the Earth to filter higher-frequency thermal signals and to integrate the effects of multiple smaller events. The search is limited to one or several climatic events that can be described in much more detail in the oxygen-isotope record. The earliest event is limited by the identification of some anomalous ground temperatures at greater depths. Also, as is clear from Figure 4c, a number of surface-temperature models that differ in detail may give acceptable fits to the measured data, and any criteria to identify the best model are at best subjective if the inherent resolution due to data spacing and inaccuracy is considered (e.g. Vasseur and others, 1983). Hence, only the general trends of the $\delta^{18}O$ climate model can be compared to a surface-temperature history derived from the ground temperatures.

The $\delta^{18}O$ record implies a 2–3 K increase in air temperatures between the end of the Little Ice Age (LIA) and 1971 (Fig. 2). In comparison, the ground temperatures at the three nearby wells suggest increases of 2–5 K at the end of the LIA (Figs 4a–c, 6 and 7). The decrease of temperatures that initiated the LIA appear to be reflected in ground temperatures only at wells #197 and #166.

An increase of 3.3 K in the last 70 years was inferred from the temperature inversion in the upper 150 m at #197 (Fig. 4a), while the analysis of the broad concave curvature observed to 500 m depth is consistent with a preceding period of seven centuries of decreasing temperatures (Fig. 4b and c).

Similarly, two possible long-term trends were hypothesized at #175. Curvature above 125 m is modelled in terms of an increase in ground-surface temperature of 4.2 K about 100 years ago, while departures of temperatures above 400 m from a deeper trend suggest a rise in ground-surface temperatures of 4 K around 800 years ago with a further increase of 2 K about 100 years ago (Fig. 6). In the latter case, the predicted temperatures fall short of matching the measured data between 200 and 350 m. We note, however, what appears to be a distinct break in slope of the Bullard plot between the trend above a depth of 300 m, as used for model F, and that below 350 m, as used for model E. This curvature is rather sharp to be climatically related and has the effect of forcing the onset of higher temperatures back to 800 years ago (model E), inconsistent with reconstructions from the other wells. The break is coincident with a major formation boundary (Fig. 3b), and it appears that the estimated thermal conductivity values do not reflect the change. We thus have considerably less confidence in model E.

The reconstruction at #166 suggests a short, intense LIA occurring from 150 to 70 years ago, when ground-surface temperatures appear to have been about 5 K lower than today (Fig. 7).

Using the mean 1971–46 $\delta^{18}O$ as a reference level, the paleoclimate record at Agassiz Ice Cap for the past 1000 years predicts the climatic related ground-temperature anomalies at sites #175 and #166 within ±0.5 K of the measured values (Figs 6 and 7), but falls short by more than 1 K at #197 (Fig. 5) and does not predict the large curvature and the temperature–depth inversion observed there. While Sass and others (1971b) noted that generalized climatic models from other disciplines may be inadequate to describe measured ground temperatures, it is instructive for the purposes of this paper to assemble evidence for these differences.

The lack of agreement at #197 might arise from several sources. The temperature logs used in the analyses were taken in 1986 (#197 and #166) and 1982 (#175), while the reference level used to determine a relative climatic history from the oxygen-isotope data was a mean value representative of a period about two decades earlier.

The two data sets are, then, non-contemporaneous, and possibly the reference level for the oxygen-isotope data needs adjusting to the later dates. However, it appears unlikely that the lack of fit at #197 is due to a subsequent increase in air temperatures by 2 K (Fig. 5), in view of the satisfactory agreement at #166 and #175. Such warming is also in conflict with a small decrease in air temperatures at Eureka weather station for the period 1972–86 relative to 1971–48 (Canada. Atmospheric Environment Service, 1973; Climatological summary sheets). Hence, we look to some other environmental factor unique to site #197 that can explain the temperature inversion.

As Lachenbruch and Marshall (1986) have pointed out, the surface-temperature history derived from deep ground temperatures refers to temperature changes at the base of the active layer in permafrost regions. The heat exchange between the top of the permafrost and the air is a complex process (e.g. Smith and Riseborough, 1983) and it is unlikely that an air-temperature variation is propagated to the base of the active layer without modification by seasonal freeze, thaw and snow cover.

We hypothesize that site #197 may have experienced an increase in mean snow cover coincident with the ending of the LIA such that the top of the permafrost would have experienced an increase in temperature due not only to an increase in mean air temperatures but also to increased thermal insulation against seasonal winter temperatures. Smith and Riseborough (1983) showed that variations in snow cover may result in differences in ground-surface temperature greater than that induced simply by air-temperature changes alone and that a simple correlation between air temperatures and ground temperatures is unlikely to occur in Nature. However, there appears to be no independent evidence that the snow cover at this site was less during the LIA.

There is, however, contemporary evidence for unusually deep snow at this well. In six visits to the well site during the Arctic spring in the decade preceding the 1986 log, 1 m of coarse, unpacked snow was encountered, in very marked contrast to the thin, hard-packed snow observed at other Arctic well sites. An early, deep snow cover might be expected to contribute several degrees (Lachenbruch, 1959; p. 28) to the significantly higher ground-surface temperatures measured at #197 compared both to other Arctic well sites (Fig. 1) and to Eureka mean air temperatures.

Alternatively, strong climatic inversions averaging 12 K in magnitude exist 80% of the time to heights of 120 m in the Eureka area over the winter months (Maxwell, 1982, fig. 6.2–6.20). Neil Peninsula is approximately 9 km in width and rises rapidly from the sea to an elevation of 497 m at well site #197 near the centre. If the topography penetrates the inversion, higher air temperatures may exist at the well site compared to the surrounding sea ice. This would complement the insulating effect of the present snow cover in raising the mean ground-surface temperature.

Hence, at the three wells, the greater rise in ground-surface temperature, 2–5 K, compared to the Agassiz Ice Cap climatic record, 2–3 K, could be attributed to a rise in air temperatures at the end of the LIA by that amount, or to a lesser rise in mean air temperature with other additive effects associated with the surface-energy balance. An increase in mean annual snow cover at these sites coincident with the end of the LIA could be one such effect. Sass and others (1971b) suggested that particular climatic events in air temperatures might be masked by coincident disturbances of comparable magnitude and opposite sign, when viewed from the perspective of ground temperatures. At #197, it appears that climatic effects are augmented by disturbances having the same sign that leave an enhanced signal in deep ground temperatures.

The rise in surface temperature at #197 10–20 years preceding the measured log (Fig. 4a and c), and the decrease about 14 years before the log at #166 (Fig. 7), may reflect changes induced by pad construction and site restoration, possible anthropogenic effects observed at some other wells in the region and in Alaska by Lachenbruch and Marshall (1986).

The variations in surface temperature over the past 1000 years have little effect on the base of the permafrost, because permafrost thicknesses are great: 500 m at wells #166 and #175, and 549 m at #197 (Taylor and others, 1982). A simple 5 K decrease in surface temperature 1000 years ago would result in an increase in permafrost thickness of only 10 m (Lunardini, 1980, equation 4.62). We conclude that the energy balance at the base of the permafrost, and particularly latent heat, may be neglected for the short climatic events considered here.

CONCLUSIONS

The paleoclimate record attributed to oxygen-isotope data in ice cores taken in the Agassiz Ice Cap, northeastern Canadian Arctic Archipelago, has been compared to changes in ground-surface temperatures derived from temperature profiles measured in three nearby wells (Fig. 1).

1. The $\delta^{18}O$ record from the Agassiz 1979 ice hole (Fig. 2) implies that air temperatures were 2–3 K lower than today during the Little Ice Age (LIA). Analysis of non-linearities in the upper several hundred metres of three ground-temperature profiles (Fig. 3) within 260 km of the Agassiz Ice Cap hole suggests that ground-surface temperatures were about 2–5 K lower than today during the LIA (Figs 4a–c, 6 and 7). This apparent larger increase in ground-surface temperatures at these wells following the LIA may be attributed to an air temperature increase of that magnitude, or more likely, to a smaller increase in mean air temperatures with a coincident increase in snow cover.

2. At site #197, a concave curvature between 150 and 450 m and a pronounced inversion in the temperature–depth profile above 150 m depth have been modelled as a gradual decrease in surface temperatures following the Little Climatic Optimum about 1000 years ago and a subsequent more rapid increase in temperature by about 5 K at the end of the LIA, about 100 years ago (Fig. 4b and c). In contrast, well #175 does not appear to have any signature of conditions prior to the LIA (Fig. 6).

3. The paleoclimate derived from the oxygen-isotope data predicts the sub-surface temperatures within ±0.5 K of the measured temperatures at sites #166 and #175 (Figs 6 and 7), but predicts neither the magnitude nor the nature of the observed temperature anomalies at site #197 (Fig. 5). A residual 2 K of surface-temperature increase since the LIA at #197 is not predicted by the oxygen-isotope record, and may arise from some paleoenvironmental event coincident with the end of the LIA, such as an increase in snow cover.

4. Mean ground-surface temperatures are about 6–10 K higher at well #197 compared to other wells across the Arctic Archipelago (Fig. 1). Simple modelling shows that part of this offset may arise from (1) the insulating effect of the 1 m of snow observed each spring over the past decade at this site, compared to the 0.1–0.2 m snow observed at the other wells, and (2) penetration of the topographic high at this well through the strong atmospheric inversion prevalent in the Arctic over the winter months.

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