

# An ice-marginal $\delta^{18}\text{O}$ record from Barnes Ice Cap, Baffin Island, Canada

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**ABSTRACT.** Barnes Ice Cap, Baffin Island, Canada, is a remnant of the Laurentide ice sheet that separated from it about 8500 years ago. Owing to recession of the ice cap during the Holocene, Pleistocene-age ice is now exposed along the margin in a distinctive bubble-rich white band.  $\delta^{18}\text{O}$  variations across the white ice resemble those in Canadian Arctic ice cores, suggesting that Barnes Ice Cap preserves a climatic record through the last glacial period, possibly reaching back into the previous (Sangamon) interglacial. The  $\delta^{18}\text{O}$  shift at the Wisconsin–Holocene transition (15‰) exceeds that in other Canadian and Greenland records and cannot be explained solely in climatic terms. A steady-state model reconstruction of the Laurentide ice sheet during the Last Glacial Maximum suggests that Late-glacial strata in Barnes Ice Cap originated high up (>2400 m a.s.l.) and far “inland” on the ice sheet, along a ridge that extended between the ancestral Foxe and Keewatin ice domes.

## INTRODUCTION

Most ice caps in the Norwegian, Russian and Canadian Arctic melted completely (or nearly so) during the early-Holocene thermal maximum (Koerner and Fisher, 2002). In the Canadian Arctic the larger ice caps retreated but survived, as indicated by the presence of pre-Holocene ice in cores. A cooler climate in the late Holocene led to regrowth

of ice caps. As a consequence, Pleistocene ice which would otherwise appear at the margins of the larger Canadian ice caps was buried when equilibrium lines dropped. One exception is Barnes Ice Cap, central Baffin Island (Fig. 1), which, like Penny Ice Cap, came into existence when it separated from the northeastern sector of the Laurentide ice sheet about 6000<sup>14</sup>C years ago. This is borne out by geological evidence (isochrones of deglaciation on Baffin Island; Dyke and Prest,

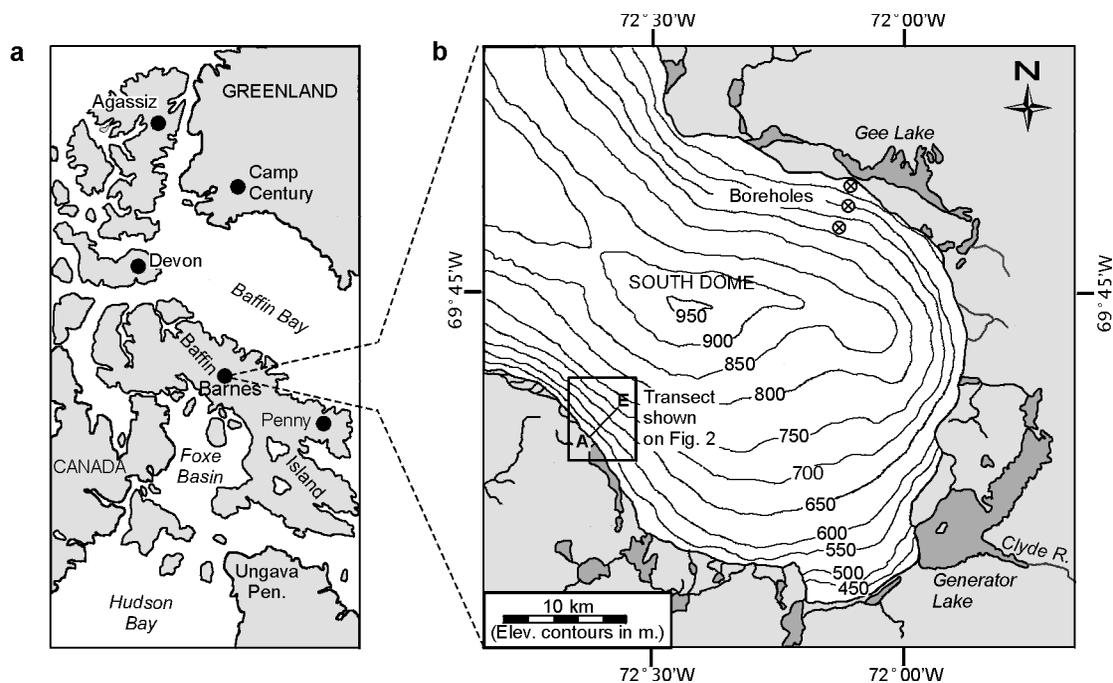


Fig. 1. (a) Location map of Barnes Ice Cap, Canadian Arctic. Locations of other ice-core sites mentioned in the text are also shown. (b) South dome of Barnes Ice Cap, showing location of sampling transect A–E (this work) and boreholes studied by Hooke and Hudleston (1980).

1987; Dyke and Hooper, 2001) and by Pleistocene-age ice outcropping along the margin of the ice cap (Hooke, 1976; Hooke and Clausen, 1982). This ice can be identified by its white, bubble-rich aspect, and on air photographs and Landsat imagery appears nearly continuous around the ice cap. In this way Barnes Ice Cap differs from the Queen Elizabeth Islands (QEI) ice caps which were part of a separate late-Pleistocene ice sheet (Dyke and others, 2001). While some QEI ice caps contain ice of Pleistocene age, this did not originate from the Laurentide ice sheet proper. Recession of Barnes Ice Cap during the late Holocene has left exposed the deep ice layers of Pleistocene age along its margin. At present, it is the only ice cap in the Canadian Arctic that allows sampling of relict Laurentide ice for paleoenvironmental research without deep-coring.

Studies on the Greenland ice sheet have shown how  $\delta^{18}\text{O}$  variations in marginal ice can be correlated with those in deep cores from the center of the ice sheet, thereby allowing marginal ice to be dated (Reeh and others, 1993, 2002). Barnes Ice Cap may also preserve an ice-marginal  $\delta^{18}\text{O}$  record that can be correlated with cores. In June 2000, we made a reconnaissance expedition to the ice cap to collect surface samples in order to establish a preliminary isotopic ( $\delta^{18}\text{O}$ ,  $\delta\text{D}$ ) chronology and evaluate the paleoenvironmental record.

## SITE DESCRIPTION

The study site (69°39' N, 72°39' W) is located along the southwestern margin of Barnes Ice Cap's South Dome (Fig. 1). It was selected for ease of access, lack of crevasses and its wide exposure of white ice, as determined from air photographs. The slope of the ice margin here averages 12°, and ice layers dip up-glacier at 21–29°. The lowermost ~50 m of the ice-margin zone is littered with rock debris that blends into broad Neoglacial moraines. The white ice is 165–180 m wide over ~1 km. As observed by Hooke (1976), the white ice owes its color to abundant air bubbles. The blue ice (upslope) is comparatively bubble-poor and is thought to have formed as superimposed ice in the early Holocene under conditions as warm as present or warmer (Hooke and Clausen, 1982). From the dip of foliation, the thickness of the white ice band is estimated to attain 100 m (locally), exceeding the 40 m measured by Hooke (1976) at another site.

## METHODS

Air temperature during fieldwork (13–24 July, 2000) varied from -2° to 8°C. Fog, rain or snow were frequent. A total of 555 ice samples were collected on flowline transects extending above and below the white ice. Samples were collected with an ice chisel at 1 m intervals from ~20 cm below the surface after scraping off any slush or snow. They were melted in sealed polyethylene bags and poured into high-density polyethylene vials with airtight screwcaps which were sent to Ottawa and kept frozen until analysis. Samples were measured for stable isotopes of O and H at the University of Ottawa using a Micromass SIRA 12 gas source mass spectrometer with an analytical precision of  $\pm 0.1\text{‰}$ . Only data for oxygen are reported here. Results are expressed as  $\delta^{18}\text{O}$  relative to Standard Mean Ocean Water.

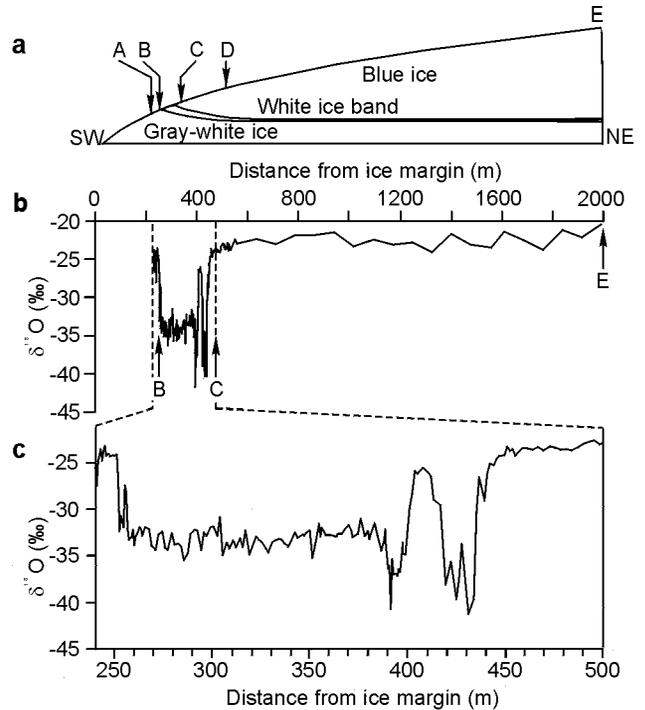


Fig. 2. (a) Schematic southwest–northeast cross-section through the margin of Barnes Ice Cap. A–E are control points on sampling transect. Sampling interval from A to D is 1 m, and from D to E is 25 m. (b)  $\delta^{18}\text{O}$  profile along transect A–E. Segment B–C is enlarged in (c).

## OXYGEN ISOTOPE STRATIGRAPHY

A preliminary  $\delta^{18}\text{O}$  profile across the southwestern margin of Barnes Ice Cap is presented in Figure 2. While some details are yet unresolved, results clearly show that the white ice is consistently isotopically “colder” (i.e. more negative  $\delta^{18}\text{O}$  values) than the blue ice (upslope) or grayish white ice (downslope). The median  $\delta^{18}\text{O}$  in the white ice is -33.1‰, compared with -22.7‰ in blue ice and -24.7‰ in grayish white ice. Our findings therefore support Hooke and Clausen’s (1982) contention that the Pleistocene-age ice is laterally continuous around the Barnes Ice Cap margin. Sampling of the grayish white ice was hindered by meltwater near the margin, so the transect only extended ~20 m downslope from the white ice. However, our data reveal that the contact between the grayish white ice (downslope) and the white ice (upslope) is marked by a transition, over a distance of ~5 m, from warmer (-24.2‰) to colder (-33.5‰)  $\delta^{18}\text{O}$  values (Fig. 2). Hooke and Clausen (1982) suggested that the relatively “warm”  $\delta^{18}\text{O}$  signature of the grayish white ice may be either primary, reflecting a warmer climatic interval that preceded the formation of the isotopically “cold” white ice, or secondary, reflecting an isotopic enrichment due to regelation at the ice–bedrock interface. They also pointed out that the bubble content of the grayish white ice appears too great for it to have formed as superimposed ice during the warmer, preglacial climate.

We have insufficient data to unequivocally resolve how the grayish white ice formed. However, we can gain insights into the early history of Barnes Ice Cap by comparing the preliminary  $\delta^{18}\text{O}$  profile with profiles from ice cores (Fig. 3). The Barnes Ice Cap  $\delta^{18}\text{O}$  profile shows a close resemblance to that from Penny Ice Cap, the only other ice cap to preserve relict Wisconsinan ice from the Laurentide ice sheet

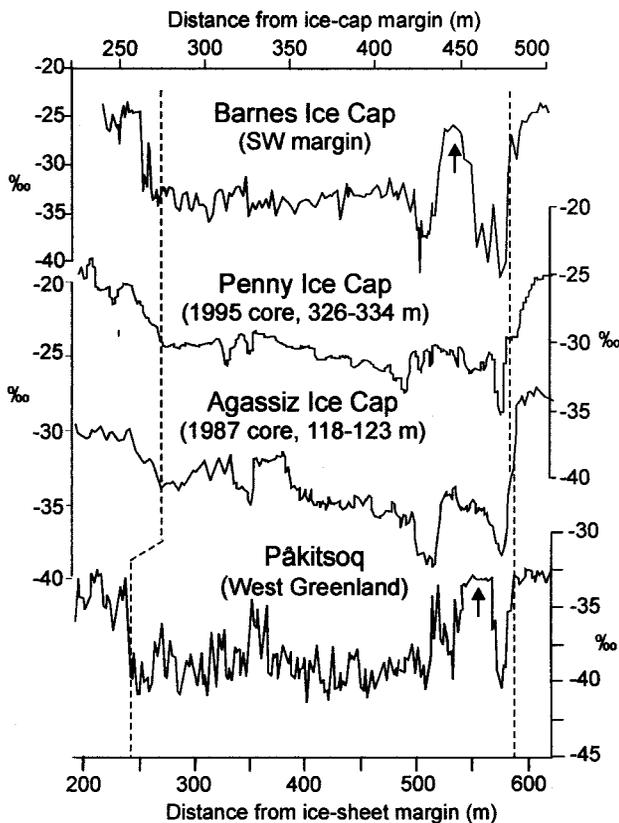


Fig. 3. Comparison of ice-core and ice-marginal  $\delta^{18}\text{O}$  profiles from the Canadian Arctic and Greenland. Tentative correlations between the profiles are shown as dashed lines. Arrows point to “anomalous” segments in the Barnes Ice Cap and Pákitsoq  $\delta^{18}\text{O}$  profiles (see text). Data for Penny Ice Cap: Fisher and others (1998); for Agassiz Ice Cap: Fisher (1988); for Pákitsoq: Reeh and others (1991).

(Fisher and others, 1998). There is also good agreement between the  $\delta^{18}\text{O}$  profiles from Barnes and Agassiz Ice Caps, although these ice caps were linked to separate ice sheets during the last glaciation (Dyke and others, 2001). At present, no reliable pre-Holocene chronology exists for Canadian Arctic cores owing to the considerable attenuation of Pleistocene ice in them (typically <10 m thick). However, comparisons between properties ( $\delta^{18}\text{O}$ , ice texture, impurity content) of deep ice layers in Canadian and Greenland cores suggest that the larger QEI ice caps (Devon, Agassiz) probably began their existence late in the Last Interglacial (Sangamon) or in the Wisconsin glaciation (Koerner and Fisher, 1979, 2002). Likewise, Barnes and Penny Ice Caps may have formed during the initial growth phase of the Laurentide ice sheet in the late Sangamonian. This hypothesis could account for the relatively high  $\delta^{18}\text{O}$  values (averaging  $-24\text{‰}$ ) found in the deepest layers of all these ice caps (Koerner and Fisher, 1979). If so, the relatively bubble-rich, yet isotopically “warm”, grayish white ice found at the margin of Barnes Ice Cap could have formed during early glacial time when regional temperatures had not yet dropped to full glacial levels, but the climate had cooled sufficiently to allow for ice accumulation through dry-firm densification.

The dispersed dirt in the grayish white ice, which may indicate regelation at the glacier bed (Hooke and Clausen, 1982), could also be explained, in part, by deposition of wind-blown dust during the early stages of the ice cap, when its elevation was low. A similar explanation was proposed to

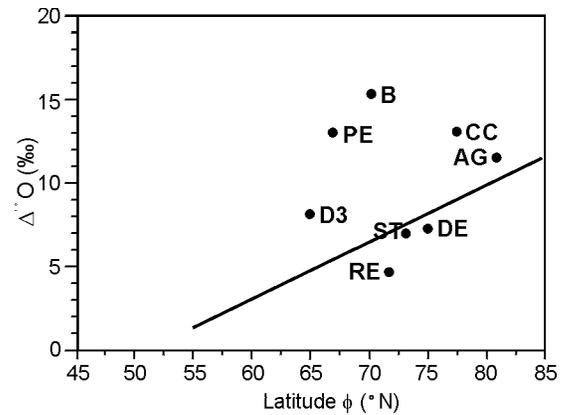


Fig. 4. Variations of  $\Delta^{18}\text{O}$  ( $=\delta^{18}\text{O}_{\text{LGM}} - \delta^{18}\text{O}_{\text{PRESENT}}$ ) in ice cores from the Canadian Arctic (AG, B, DE) and Greenland as a function of latitude  $\phi$ . AG, Agassiz Ice Cap; B, Barnes Ice Cap; CC, Camp Century; DE, Devon Ice Cap; D3, Dye 3; RE, Renland ice cap; ST, Summit (GRIP). The solid line is the theoretical  $\Delta^{18}\text{O}-\phi$  relationship for  $\phi > 55^\circ\text{N}$ . Modified from Fisher and Alt (1985).

account for dirt pockets in the basal layers of the cold-based Devon Ice Cap (Koerner and Fisher, 1979). Alternatively, the grayish white ice may also have formed through tectonic mixing of dirt-bearing basal ice with cleaner englacial ice, as described by Souchez and others (1995).

Between 428 and 430 m from the margin, the Barnes Ice Cap  $\delta^{18}\text{O}$  profile shows an abrupt shift from  $-41.7\text{‰}$  to  $-26.7\text{‰}$ , a difference of  $15\text{‰}$  over  $\sim 2\text{ m}$ . The  $\delta\text{D}$  profile (not shown here) mirrors the  $\delta^{18}\text{O}$  profile closely. The abrupt  $\delta^{18}\text{O}$  shift at the Wisconsin–Holocene transition is a characteristic feature in long ( $>10^4$  years) ice-core records from the Northern Hemisphere (e.g. Dansgaard and others, 1982; Johnsen and others, 1992). Curiously, the  $\delta^{18}\text{O}$  step in Barnes Ice Cap does not correspond to the estimated point of contact between the white (bubble-rich) and blue (bubble-poor) ice, but occurs  $\sim 40\text{ m}$  upslope from it. The most likely explanation for this mismatch is that we misidentified the exact location of the white-ice/blue-ice contact. This contact is clearly visible from a distance, but becomes indistinct up close due to surface irregularities and runoff.

A close-up of the  $\delta^{18}\text{O}$  profile across the white ice reveals a shift to near-Holocene  $\delta^{18}\text{O}$  values (Fig. 2c) amidst the isotopically “coldest” part of the sequence. It is tempting to identify this “warm” interval with the Bølling–Allerød event, and the subsequent peak of “cold”  $\delta^{18}\text{O}$  values with the Younger Dryas. However, in Canadian ice cores the Bølling–Allerød/Younger Dryas oscillation is typically faintly expressed (Fig. 3). The “warm”  $\delta^{18}\text{O}$  interval in the Barnes Ice Cap profile stands out as an anomaly with respect to these records. Similar anomalies occurring at or near the Wisconsin–Holocene transition are seen in ice-marginal  $\delta^{18}\text{O}$  profiles from Greenland (e.g. Pákitsoq (Fig. 3)). They have been attributed to folding of ice due to the large viscosity difference between layers of Holocene and late-Wisconsin age (Reeh and others, 1993). The “warm–cold” oscillation in the Barnes Ice Cap  $\delta^{18}\text{O}$  profile could also reflect the presence of recumbent folds in the ice, as hypothesized by Hudleston (1976). Clearly, caution must be observed in interpreting details of the Barnes Ice Cap marginal  $\delta^{18}\text{O}$  profile that may be of structural, rather than climatic, origin. Stratigraphic disturbances need not be

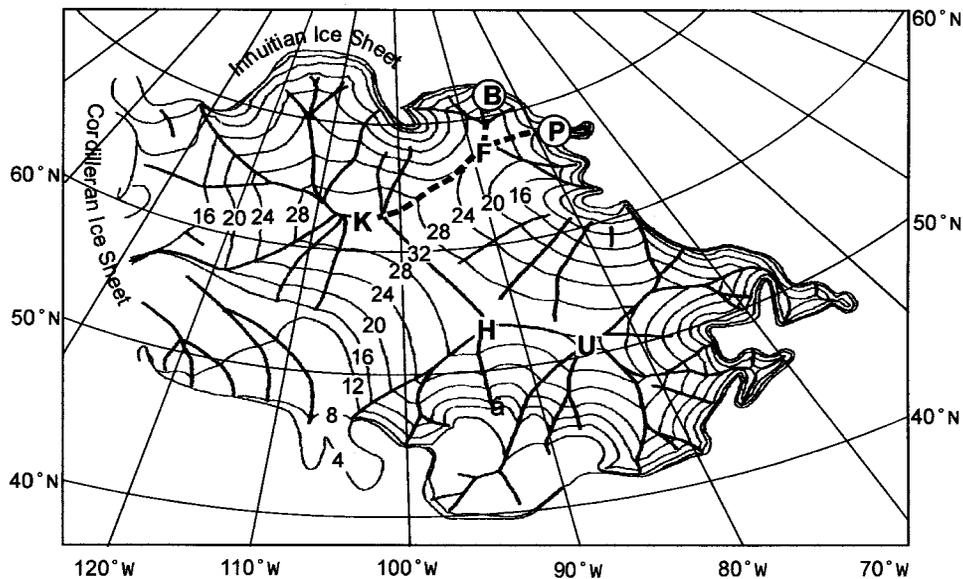


Fig. 5. Steady-state model reconstruction of the Laurentide ice sheet for the LGM, showing topography (contour lines in hundreds of metres) and major ice flowlines (bolder lines). Ice centres: F, Foxe; K, Keewatin; H, Hudson; U, Ungava. Modern location of Barnes (B) and Penny (P) Ice Caps is shown. After Fisher and others (1985, fig. 2).

continuous around the ice-cap margin, however, and a careful site selection may yield an undisturbed  $\delta^{18}\text{O}$  record.

The Barnes Ice Cap  $\delta^{18}\text{O}$  profile shows an unusually large isotopic shift at the late-Wisconsin–Holocene transition (15‰). This shift is greater than in other Canadian Arctic and Greenland cores (8–11‰), except for Penny Ice Cap (about 14‰; Fisher and others, 1998). Part of this shift could be accounted for by differences in  $\delta^{18}\text{O}$  of precipitation formed under late-Wisconsin and Holocene climatic conditions. Fisher and Alt (1985) used a multiple-source, global isotope model to investigate changes in the  $\delta^{18}\text{O}$  of precipitation as a function of latitude under Last Glacial Maximum (LGM) and present conditions ( $\Delta^{18}\text{O} = \delta^{18}\text{O}_{\text{LGM}} - \delta^{18}\text{O}_{\text{PRESENT}}$ ). If the model-predicted  $\Delta^{18}\text{O}$ –latitude relationship is compared with  $\Delta^{18}\text{O}$  values in Canadian Arctic and Greenland records (Fig. 4), it becomes apparent that the  $\Delta^{18}\text{O}$  in Penny and Barnes Ice Caps exceeds markedly the expected range of values.

Hooke and Clausen (1982) estimated that 5‰ of the 15‰ Wisconsin–Holocene  $\delta^{18}\text{O}$  shift in Barnes Ice Cap can be accounted for by the paleo-elevation isotope fractionation effect on precipitation that fell in Late-glacial time, when the ice cap was an extension of the Laurentide ice sheet and its surface was considerably higher than at present. We take this inference further by considering the hypothetical topography of the Laurentide ice sheet during the LGM. We use a steady-state reconstruction of the Laurentide ice sheet that assumes non-deformable subglacial sediments in the Hudson Bay region (Fisher and others, 1985). This reconstruction gives an LGM ice-sheet configuration that agrees well with geologically constrained reconstructions (Dyke and Prest, 1987). Assuming a standard altitude fractionation effect of  $-0.6\text{‰}$  per 100 m, the model suggests that late-Wisconsin ice in Penny and Barnes Ice Caps may have originated thousands of km from their modern location, up on a ridge that extended from the ancestral Foxe Dome ( $\geq 2400\text{ m}$  above present sea level) to the central Keewatin Dome (3200 m a.s.l.; Fig. 5). These paleo-elevation estimates are supported by a reappraisal of the Laurentide ice-sheet configuration during the LGM which gives greater and thicker ice cover than previously thought (Dyke and Hooper, 2001; Dyke and others, 2001).

Our extended-flowline hypothesis has the additional merit of providing a plausible interpretation for  $\delta^{18}\text{O}$  and major-ion chemistry records in Penny Ice Cap, which differ markedly from those of QEI ice caps (Fisher and others, 2002).

## CONCLUSIONS AND FUTURE WORK

Marginal strata on Barnes Ice Cap preserve a paleoenvironmental record that may extend into early-Wisconsin or Sangamonian time. The oldest ice in Barnes Ice Cap is speculated to have formed during initial build-up of the Laurentide ice sheet. The Barnes ice-marginal  $\delta^{18}\text{O}$  profile is strikingly similar to that from Penny Ice Cap, which supports a common origin in the northeastern sector of the Laurentide ice sheet. The unusually large Wisconsin–Holocene  $\delta^{18}\text{O}$  shift (15‰) suggests a high-elevation point source for Late-glacial ice in Barnes Ice Cap, possibly on the Foxe–Keewatin sector of the Laurentide ice sheet. This extended-flowline hypothesis also provides the basis for a plausible interpretation of  $\delta^{18}\text{O}$  and major-ion records from Penny Ice Cap.

Some features of the Barnes  $\delta^{18}\text{O}$  profile suggest stratigraphic disturbances in the record, probably due to folding of ice layers. More exhaustive sampling along the ice margin is needed to confirm our preliminary findings and establish a robust chronology for the  $\delta^{18}\text{O}$  record. Future work should include measurements of other parameters such as trace gases ( $\text{CO}_2$ ,  $\text{CH}_4$ ) or radionuclides (e.g.  $^{10}\text{Be}$ ,  $^{36}\text{Cl}$ ) that can be used to constrain the age of ice strata through correlation with Greenland ice-core records.

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REFERENCES

Dansgaard, W. and 6 others. 1982. A new Greenland deep ice core. *Science*, **218**(4579), 1273–1277.

Dyke, A. S. and J. Hooper. 2001. *Deglaciation of northwest Baffin Island, Nunavut*. Geological Survey of Canada. (Map 1999A, scale 1:500,000)

Dyke, A. S. and V. K. Prest. 1987c. *Paleogeography of northern North America, 18 000–5000 years ago*. Geological Survey of Canada, Ottawa, Ontario, map. (GSC Map 1703A, scale 1:12,500,000)

Dyke, A. S. and 6 others. In press. The Laurentide and Innuitian ice sheets during the Last Glacial Maximum. *Quat. Sci. Rev.*

Fisher, D. A. and B. T. Alt. 1985. A global oxygen isotope model — semi-empirical, zonally averaged. *Ann. Glaciol.*, **7**, 117–124.

Fisher, D. A., N. Reeh and K. Langley. 1985. Objective reconstructions of the Late Wisconsinan Laurentide ice sheet. *Géogr. Phys. Quat.*, **39**(3), 229–238.

Fisher, D. A. and 12 others. 1998. Penny Ice Cap cores, Baffin Island, Canada, and the Wisconsinan Foxe Dome connection: two states of Hudson Bay ice cover. *Science*, **279**(5351), 692–695.

Fisher, D. A. and 7 others. 2002. The effects of flowline length evolution on the chemistry- $\delta^{18}\text{O}$  profiles from Penny Ice Cap, Baffin Island, Canada. *Ann. Glaciol.*, **35** (see paper in this volume).

Hooke, R. LeB. 1976. Pleistocene ice at the base of the Barnes Ice Cap, Baffin Island, N.W.T., Canada. *J. Glaciol.*, **17**(75), 49–59.

Hooke, R. LeB. and H. B. Clausen. 1982. Wisconsin and Holocene  $\delta^{18}\text{O}$  variations, Barnes Ice Cap, Canada. *Geol. Soc. Am. Bull.*, **93**(8), 784–789.

Hooke, R. LeB. and P. J. Hudleston. 1980. Ice fabrics in a vertical flow plane, Barnes Ice Cap, Canada. *J. Glaciol.*, **25**(92), 195–214.

Hudleston, P. J. 1976. Recumbent folding in the base of the Barnes Ice Cap, Baffin Island, Northwest Territories, Canada. *Geol. Soc. Am. Bull.*, **87**(12), 1684–1692.

Johnsen, S. J. and 7 others. 1992. A “deep” ice core from East Greenland. *Medd. Grøn., Ser. Geoscience* 29.

Koerner, R. M. and D. A. Fisher. 1979. Discontinuous flow, ice texture, and dirt content in the basal layers of the Devon Island ice cap. *J. Glaciol.*, **23**(89), 209–222.

Koerner, R. M. and D. A. Fisher. 2002. Ice-core evidence for widespread Arctic glacier retreat in the Last Interglacial and the early Holocene. *Ann. Glaciol.*, **35** (see paper in this volume).

Reeh, N., H. Oerter, A. Letrégouilly, H. Miller and H.-W. Huberten. 1991. A new, detailed ice-age oxygen-18 record from the ice-sheet margin in central West Greenland. *Palaeogeogr., Palaeoclimatol., Palaeoecol., Global Planet. Change Sect.*, **90**(4), 373–383.

Reeh, N., H. Oerter and H. Miller. 1993. Correlation of Greenland ice-core and ice-margin  $\delta^{18}\text{O}$  records. In Peltier, W. R., ed. *Ice in the climate system*. Berlin, etc., Springer-Verlag, 481–497. (NATO ASI Series I: Global Environmental Change 12)

Reeh, N., H. Oerter and H. H. Thomsen. 2002. Comparison between Greenland ice-margin and ice-core oxygen-18 records. *Ann. Glaciol.*, **35** (see paper in this volume).

Souchez, R., M. Lemmens and J. Chappellaz. 1995. Flow-induced mixing in the GRIP basal ice deduced from the  $\text{CO}_2$  and  $\text{CH}_4$  records. *Geophys. Res. Lett.*, **22**(1), 41–44.