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Notes



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ABSTRACT

The most prominent Holocene climatic event in Greenland ice-core proxies, with approximately half the amplitude of the Younger Dryas, occurred ~8000 to 8400 yr ago. This Holocene event affected regions well beyond the North Atlantic basin, as shown by synchronous increases in windblown chemical indicators together with a significant decrease in methane. Widespread proxy records from the tropics to the north polar regions show a short-lived cool, dry, or windy event of similar age. The spatial pattern of terrestrial and marine changes is similar to that of the Younger Dryas event, suggesting a role for North Atlantic thermohaline circulation. Possible forcings identified thus far for this Holocene event are small, consistent with recent model results indicating high sensitivity and strong linkages in the climatic system.

INTRODUCTION

Proxy records show that climatic variability was especially large during the most recent ice age (e.g., Johnsen et al., 1992), causing the warm Holocene in which we live to appear stable by

comparison. However, Holocene sediments often record more variability than typically observed instrumentally during this century (e.g., Little Ice Age records; Grove, 1988).

Holocene climatic shifts larger than those of the

Little Ice Age are recorded in many proxies, but typical dating uncertainties of a few centuries have limited our ability to correlate events that often lasted only a few centuries. Approximately correlative changes in proxy records from disparate regions may represent (1) a single, widespread, synchronous event; (2) a time-transgressive event; or (3) a number of local and uncorrelated events; with significantly different implications for our understanding of climatic change.

Ice cores help solve this difficulty because they record climatic conditions locally (e.g., temperature, snow accumulation), regionally (e.g., wind-blown sea salt and continental dust), and more broadly (via trapped-bubble records of concentrations of trace gases involved in global biogeochemical cycles). Local and regional indicators are delivered to the surface of the ice sheet synchronously by precipitation events. Gases are trapped in ice below the surface, but uncertainty in the age of trapped gas relative to the surrounding ice is less than 75 yr throughout the Holocene for central Greenland ice cores (Sowers et al., 1992). Because the troposphere is well mixed on decadal time scales, trapped-gas records provide a means of comparing local and regional climatic changes to various global biogeochemical cycles with minimal dating uncertainties.

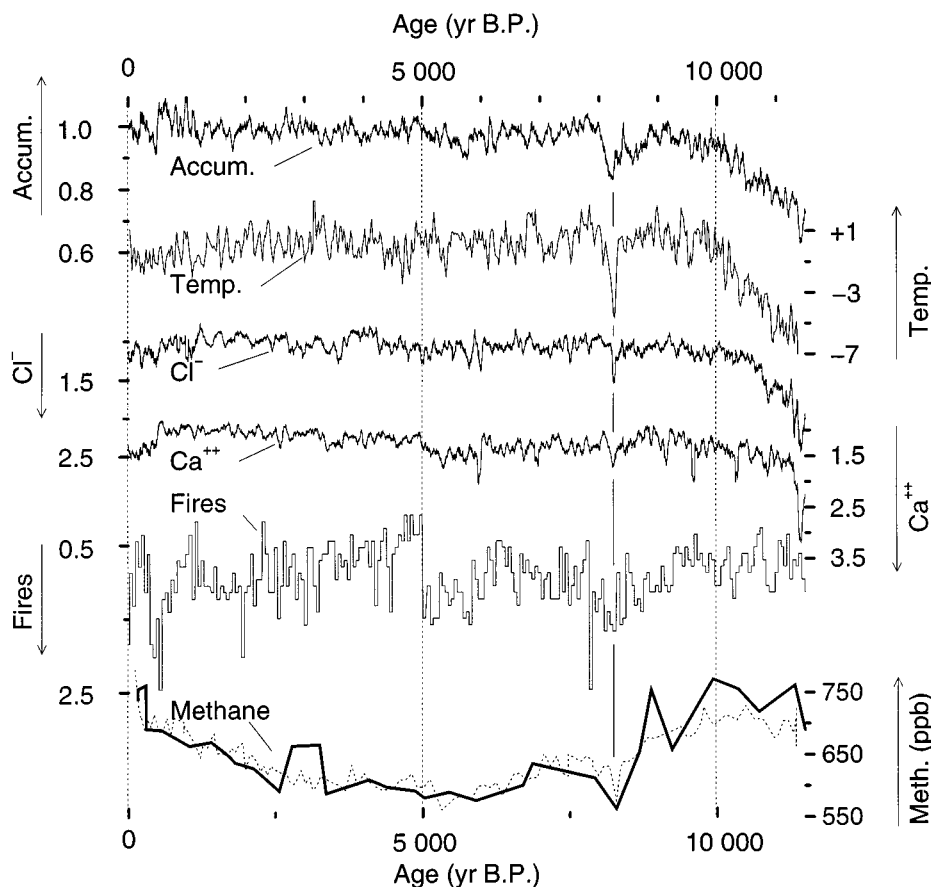


Figure 1. Climatic data from the GISP2 core, central Greenland. Approximately 50 yr running means of accumulation (Alley et al., 1993; Spinelli, 1996), chloride and calcium (O'Brien et al., 1995), and a 50 yr histogram of frequency of fallout from fires (Taylor et al., 1996), expressed as ratios to their average values during the approximately 2000 yr just prior to the Little Ice Age. Temperature ($^{\circ}\text{C}$) is calculated as deviation from average over same 2000 yr, from $\delta^{18}\text{O}_{\text{ice}}$ (Stuiver et al., 1995), assuming calibration of $0.33\text{‰}/^{\circ}\text{C}$ (Cuffey et al., 1995). Methane concentrations are shown in parts per billion by volume (ppb) for the GRIP core (thin, dashed line; Chapellaz et al., 1993; Blunier et al., 1995) and the GISP2 core (thick, solid line; Brook et al., 1996), with the GRIP data adjusted slightly for interlaboratory differences and time-scale differences to match GISP2 values following Sowers et al. (1997). Note that some scales increase upward and others downward, as indicated, so that all curves vary together at major events.

ICE-CORE DATA

Holocene climatic records from the Greenland Ice Sheet Project II (GISP2) ice core (Fig. 1) were dated by counting visible layers, with typically 1% errors (Alley et al., 1993; Spinelli, 1996). Snow accumulation was calculated by correcting annual layer thicknesses for ice flow following Alley et al. (1993); accumulation is controlled primarily by local temperature and local-to-regional atmospheric circulation patterns. Oxygen isotopic ratios of ice, $\delta^{18}\text{O}_{\text{ice}}$ (Grootes et al., 1993; Stuiver et al., 1995) primarily reflect the site temperature (Cuffey et al., 1995), although atmospheric circulation and other factors play a secondary role (e.g., Steig et al., 1994; Fawcett et al., 1997); our calibration to temperature (0.33‰/°C) is from Cuffey et al. (1995).

Chemical concentrations in the ice (O'Brien et al., 1995) reflect atmospheric loading, but fluctuations of concentrations in the ice may be magnified because colder, dustier times typically have reduced snow accumulation and thus reduced dilution of the chemicals. We have corrected for this effect, reducing the magnitudes of fluctuations in the chemical time series, following Alley et al. (1995). Because snow-accumulation variations are typically smaller than chemical-concentration variations, this correction is quite small. Chloride, primarily from sea salt, represents vigor of atmospheric circulation or distance from oceanic source areas; calcium, primarily from continental dust, is a signal of dust availability, dryness, and transport vigor from continental regions beyond Greenland and probably beyond the North Atlantic basin (e.g., Mayewski et al., 1994; Biscaye et al., 1996).

The record of forest fires is derived from the frequency of occurrence along the core of anomalously low electrical-conductivity values caused by fire-produced ammonium; this record integrates changes within source regions as well as changes in the geographic location of sources (Taylor et al., 1996). Methane, shown for the GRIP (Chappellaz et al., 1993; Blunier et al., 1995) and GISP2 (Brook et al., 1996) ice cores, was produced within the terrestrial biosphere under anaerobic conditions generally associated with wetlands. Holocene methane concentrations prior to the large human effects of recent centuries reflect moisture availability in broad land areas extending beyond the North Atlantic basin, probably including both high-latitude and low-latitude (monsoonal) regions (e.g., Chappellaz et al., 1993).

RESULTS AND DISCUSSION

The records in Figure 1 reveal large and abrupt climatic changes, and possible cyclicities (e.g., O'Brien et al., 1995; Stuiver et al., 1995). After the end of the deglaciation (the Preboreal warming) between ca. 11.5 and ca. 9.5 ka (calendar years before 1950), the largest deviations for almost all indicators occur between ca. 8.4 and ca.

8.0 ka, typically peaking at ca. 8.25 ka or ca. 7500 ^{14}C yr B.P. The pattern of this 8.2 ka event (cold, dry, dusty, and low methane occurring together) is the same as observed during the Younger Dryas and in many older stadial events (e.g., Johnsen et al., 1992; Alley et al., 1993; Chappellaz et al., 1993; Mayewski et al., 1994).

The assigned age for the 8.2 ka event is indistinguishable from that of the correlative $\delta^{18}\text{O}_{\text{ice}}$ minimum in other Greenland ice cores (8.21 \pm 0.03 ka for GRIP, Johnsen et al., 1992; and 8.23 \pm 0.05 ka for Dye 3 and Camp Century, Dansgaard, 1987). Some of the curves in Figure 1 could be interpreted to indicate that this event is part of a longer-lived anomaly from ca. 9.0 to 7.8 ka. Compared to later Holocene events, the 8.2 ka deviations are more anomalous in records of local conditions (temperature, accumulation) than for wind-blown indicators (dust, sea salt), suggesting that the Greenland climate stabilized relative to other regions in the latter Holocene.

To assess the magnitude of the 8.2 ka event, we calculate the deviation of each 8.2 ka "spike" from a baseline defined by values near 8.0 and 8.4 ka in 50-yr-average data (Fig. 1), and we similarly calculate Younger Dryas deviations from a baseline in the early part of the warm Preboreal that followed (Table 1). Forest-fire frequency increased 90% during the 8.2 ka event with a brief fivefold increase about a century later; however, insensitivity of the forest-fire proxy in the alkaline ice of the Younger Dryas (Taylor et al., 1996) precludes comparison between the events.

The $\delta^{18}\text{O}_{\text{ice}}$ change indicates a peak 8.2 ka cooling of 6 ± 2 °C when the Cuffey et al. (1994; 1995) deglacial, Holocene, and Little Ice Age calibrations are used, and is slightly more than half of the $\delta^{18}\text{O}_{\text{ice}}$ change of the Younger Dryas. Climatic deviations in the 8.2 ka event for several other indicators also are about half as large as those of the Younger Dryas, although the Younger Dryas conditions were sustained for more than a millennium whereas the maximum 8.2 ka deviations lasted for less than a century. The relatively larger Younger Dryas perturbations in wind-blown sea-salt and dust (roughly four times larger and ten times larger than at 8.2 ka, respectively) are at least in part explainable because entrainment of wind-blown materials is observed to increase much more rapidly than wind speed (reviewed by Petit et al., 1981).

CORRELATION WITH REGIONAL CHANGES

The 8.2 ka climatic perturbation has long been recognized as a significant event in Greenland (e.g., Dansgaard, 1987). Many climatic records from other regions show potentially correlative events (e.g., O'Brien et al., 1995), although problems with sample resolution and dating accuracy complicate deduction of the exact temporal relationships. Estimated calendar dates at these different sites for the midpoint of the "cor-

TABLE 1. CHANGES IN SEVERAL PALEOCLIMATIC PROXIES FOR THE 8.2 ka AND YOUNGER DRYAS EVENTS

Indicator	Event deviation from baseline	
	8.2 ka	Younger Dryas
accumulation	-20%	-50%
methane	-10 to -15%	-30%
Na ⁺	+60%	+260%
Cl ⁻	+60%	+200%
Ca ⁺⁺	+60%	+600%
Mg ⁺⁺	+40%	+370%
K ⁺	+110%	+230%
NO ₃ ⁻	+20%	+10%
$\delta^{18}\text{O}_{\text{ice}}$	-2‰	-3.5 to -4‰

Note: For the 8.2 ka event, the difference between the extreme value near 8.25 ka and a baseline defined by the values near 8.0 and 8.4 ka is expressed as a percentage of that baseline. The Younger Dryas is similarly compared to a baseline defined by the early Preboreal. (Percentage differences in $\delta^{18}\text{O}_{\text{ice}}$ are not especially meaningful, so they are left as per mil.) Most data are ~50 yr averages. Methane is sampled at discrete points. Younger Dryas chemical changes are from Alley et al. (1995) and average over several centuries, which improves statistical confidence without greatly changing the results.

relative" event range over a millennium, with estimated durations ranging from 100 to 1000 yr or more.

The synchronous occurrence of this event in local, regional, and global proxy climate records in central Greenland requires that the event also occurred in other places, thereby strengthening correlations suggested by other workers (e.g., Chappellaz et al., 1993; Gasse and van Campo, 1994). We consider it especially likely that this event correlates to (Fig. 2) cold conditions in northern Sweden (Karlén, 1976) and on the Agassiz Ice Cap, Ellesmere Island, Canada (Fisher et al., 1995); fresh, cool surface conditions in the North Atlantic (Duplessy et al., 1992; Keigwin and Jones, 1995; Bond, 1995), dry, windy conditions in the Laurentian Great Lakes and surrounding regions (Dean, 1993; Rea et al., 1994); dry conditions in broad monsoonal regions of Africa, the Arabian Peninsula, Tibet, and north-west India (Sirocko et al., 1993; Gasse and van Campo, 1994; Lamb et al., 1995; and references therein); and windy conditions in the Cariaco Basin, offshore Venezuela (Hughen et al., 1996). In general, the better-dated and more densely sampled records show greater similarity to the ice-core record in timing and duration.

The cause of this event is unknown. It may be related to periodicity in the climatic system (Bond, 1995; O'Brien et al., 1995) or other causes. The pattern in Greenland (cold, dry, windy, disturbed continental source regions, low methane) and elsewhere (cold North Atlantic basin, strong North Atlantic trade winds, dry monsoonal regions) closely

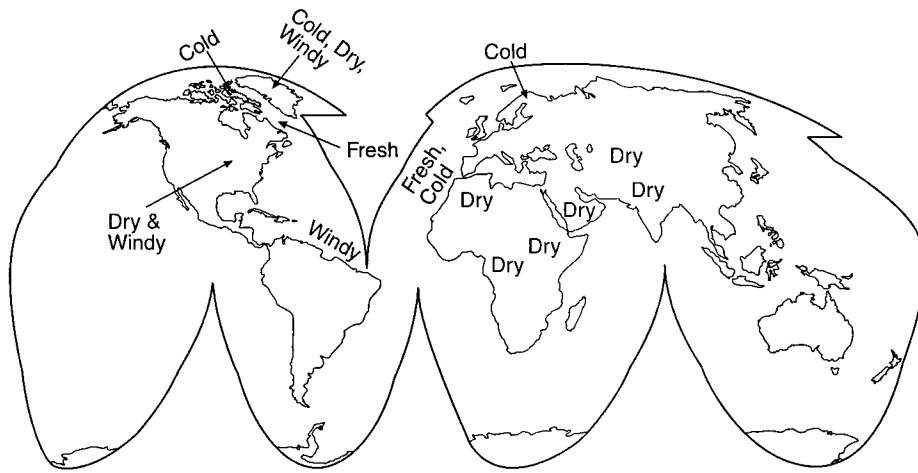


Figure 2. Regional changes probably correlative with prominent 8.4–8.0 ka event recorded in GISP2 ice core. References are given in text.

matches that of the Younger Dryas and stadial cold times of the ice age (e.g., Street-Perrott and Perrott, 1990; Peteet, 1995). Such a pattern is consistent with model simulations for a reduction of North Atlantic oceanic heat transport (e.g., Manabe and Stouffer, 1988; Mikolajewicz, 1996; Fawcett et al., 1997) and with climate-anomaly patterns observed this century in association with cold conditions in the North Atlantic (Parker et al., 1988).

Increased fresh-water supply to the North Atlantic triggering a slowdown of its thermohaline circulation is often suggested to explain this pattern (e.g., Broecker et al., 1990). However, no extraordinarily large increase in fresh-water flux to the North Atlantic has been identified at 8.2 ka. This may indicate for the 8.2 ka event that (1) increased fresh-water supply was not involved; (2) some large fresh-water source has been overlooked; or (3) the system was quite sensitive to small changes in fresh-water supply.

Based on the data and interpretations of Fairbanks (1990; also see Bard et al., 1996), 8.2 ka was a time of typically slow and slowing sea-level rise compared to the several preceding millennia (fresh-water delivery to the oceans of $\sim 0.07 \text{ Sv} = 0.07 \times 10^6 \text{ m}^3 \cdot \text{s}^{-1}$ if this was the only cause of sea-level rise, with most of the supply probably from the Antarctic, Peltier, 1994). The abrupt sea-level rise proposed by Blanchon and Shaw (1995) (but see Bard et al., 1996) is dated by them as having started at $7.6 \pm 0.1 \text{ ka}$, significantly more recently than the termination of the climatic event at $8.0 \pm 0.1 \text{ ka}$ in our records. The data of Larcombe et al. (1995) allow, but do not require, accelerated sea-level rise just before the event and a reduced rate of rise during the event. Deglaciation of Hudson Bay and retreat of the mid-latitude ice sheets from many marine margins significantly preceded the event, although calving margins in Foxe Basin and along Ungava Bay, Canada, probably persisted through the event (Andrews, 1989). The remnant of the Lau-

rentide ice sheet during the event is reconstructed as having been smaller than the modern Greenland ice sheet in volume and area (Dyke and Prest, 1987; Peltier, 1994).

Ice-marginal lakes that formed and drained around the Laurentide ice sheet in Labrador at about the time of the 8.2 ka event (Clark and Fitzhugh, 1990) might have been involved (Keigwin and Jones, 1995). However, those described by Clark and Fitzhugh (1990) could have produced a fresh-water flux only on the order of 0.05 Sv if they drained in a single year, possibly associated with additional fluxes from ice-flow changes linked to lake drainage (the actual drainage time and rate are not constrained precisely). Some models project that enhanced fresh-water fluxes of this magnitude to the North Atlantic will slow or stop deep-water formation if maintained long enough (e.g., Manabe and Stouffer, 1994); 0.015 Sv delivered to the Labrador Sea is sufficient to stop convection there in one model (Rahmstorf, 1995). If these lakes were significant in causing the event, then our data support those models showing large North Atlantic climatic sensitivity to small forcings.

It is important to verify this result, because an increase in fresh-water flux of similar magnitude may occur in the future. For example, a global warming of $3 \text{ }^\circ\text{C}$ with a twofold polar amplification in response to the coming doubling of atmospheric CO_2 is projected to increase the total fresh-water flux from the Greenland ice sheet by about 0.02 Sv and to maintain this level over centuries (Warrick et al., 1995). In addition, enhanced high-latitude precipitation and sea-ice melting in response to projected warming might cause an increase of similar magnitude in fresh-water flux to the North Atlantic (Manabe and Stouffer, 1994; Rahmstorf, 1995). Such an occurrence, although perhaps providing a negative feedback on greenhouse warming in the North Atlantic region, could cause major climatic per-

turbations on the societally important time scales of decades to centuries.

Available data thus show that a cold, dry, windy event in and around the North Atlantic basin occurred at ca. 8.2 ka, well within the Holocene during climatic conditions similar to those of today. The cause of this event is not known. We cannot exclude the hypothesis that it records a decrease in North Atlantic thermohaline circulation caused by an increase in fresh-water flux to the North Atlantic of the same magnitude as is expected from anthropogenic forcing.

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