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Abrupt changes in North American climate during early Holocene times

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Recent studies of the Greenland ice cores have offered many insights into Holocene climatic dynamics at decadal to century timescales[1–3]. Despite the abundance of continental records of Holocene climate, few have sufficient chronological control and sampling resolution to compare with the Greenland findings[4]. But annually laminated sediments (varves) from lakes can provide high-resolution continental palaeoclimate data with secure chronologies. Here we present analyses of varved sediments from Deep Lake in Minnesota, USA. Trends in the stable oxygen-isotope composition of the sedimentary carbonate indicate a pronounced climate cooling from 8.9 to 8.3 kyr before present, probably characterized by increased outbreaks of polar air, decreased precipitation temperatures, and a higher fraction of the annual precipitation falling as snow. The abrupt onset of this climate reversal, over several decades, was probably caused by a reorganization of atmospheric circulation and cooling of the Arctic airmass in summer that resulted from the final collapse of the Laurentide ice near Hudson Bay and the discharge of icebergs from the Quebec and Keewatin centres into the Tyrell Sea. The timing and duration of this climate reversal suggest that it is distinct from the prominent widespread cold snap that occurred 8,200 years ago in Greenland and other regions[1,5,6]. No shifts in the oxygen-isotope composition of sediment carbonate occurred at 8.2 kyr before present at Deep Lake, but varve thickness increased dramatically, probably as a result of increased deposition of aeolian dust. Taken together, our data suggest that two separate regional-scale climate reversals occurred between 9,000 and 8,000 years ago, and that they were driven by different mechanisms.

We retrieved three temporally overlapping sediment cores from Deep Lake (47° 41′ N, 95° 23′ W; Fig. 1), a topographically closed basin located ~45 km east of the prairie–forest border in northwestern Minnesota. The early Holocene sediments of these cores are varve couplets consisting of calcite precipitated through photosynthesis in summer paired with darker, clastic and organic debris deposited in other seasons. Annual laminae are easily identifiable under a low-power microscope, providing a high-quality chronometer. To minimize the accumulative error of varve counting from the surface, we anchored our varve counts on a reliable AMS (accelerator mass spectrometry) ^{14}C date of 8,090 ± 85 years (calibrated to 8.986 kyr calendar age[7]) on a large wood sample[8]. A second AMS ^{14}C date of 8, 740 ± 60 years, also on a piece of wood, is supportive of the annually laminated nature of early Holocene sediments from Deep Lake, but this date is less useful because it is on a ^{14}C plateau and yields ambiguous calibrated ages[7]. Varve thickness was measured, and subsamples from the cores were analysed for stable isotopes at multi-decadal resolution to reconstruct early Holocene climate. Our results provide an opportunity to test whether decadal to century scale climate changes observed in Greenland ice also occurred near the centre of the North American continent.

Bulk-carbonate $\delta^{18}O$ of the Deep Lake core shows distinct stratigraphic changes between 10.0 and 8.0 calendar kyr before present (cal. kyr BP; Fig. 2a). $\delta^{18}O$ increases by ~2‰ about 9.5 cal. kyr BP, reflecting the effects of regional climate warming related to increased summer insolation and the retreat of the Laurentide ice sheet[4], as well as the associated increase in evaporation from Deep Lake itself and the reduced influence on climate of glacial Lake Agassiz as it retreated into Canada[8,9]. Around 8.9 cal. kyr BP, $\delta^{18}O$

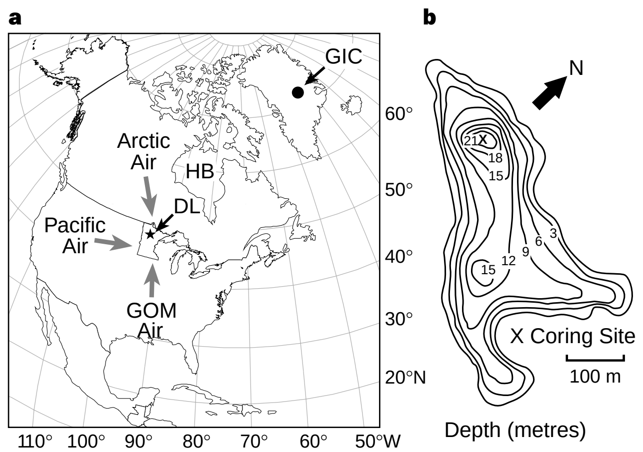

Figure 1 Air masses, site locations and lake bathymetry. **a**, General directions of the three main airmasses controlling the climate of Minnesota (GOM, Gulf of Mexico); also shown are the locations of Deep Lake (DL), Greenland ice cores (GIC), and Hudson Bay (HB). **b**, Bathymetry of Deep Lake.

declines abruptly by ~2‰ within <50 yr. It remains low at about −8‰ until 8.4 cal. kyr BP, when it gradually increases to ~−6.5‰ by 8.2 cal. kyr BP and stays constant at least until 7.8 cal. kyr BP. After 9.0 cal. kyr BP, $\delta^{13}C$ and $\delta^{18}O$ significantly co-vary ($R^2 = 0.820$, $p < 0.001$, $n = 42$) with similar stratigraphic patterns.

The abrupt decrease in $\delta^{18}O$ at 8.9 cal. kyr BP post-dates the forest–prairie transition at Deep Lake at ~9.5 cal. kyr BP[8], but it coincides with an overall increase in aeolian deposition inferred from increased varve thickness (Fig. 3), as discussed below. This coincidence suggests the possibility that input of detrital carbonate caused the observed isotopic changes. However, no consistent differences in either $\delta^{18}O$ or $\delta^{13}C$ exist between samples of thick-varve and thin-varve bundles within the period of interest (Fig. 2b). In addition, X-ray diffraction analysis of selected samples shows no corresponding changes in carbonate composition, and microscopic examination of sediments indicates the overall dominance of low-Mg authigenic calcite throughout the profile. Thus the observed isotopic changes are not caused by changes in carbonate composition due to detrital input.

Because lake-sediment $\delta^{18}O$ records from continental regions reflect a number of climatic and hydrologic factors[10,11], it is difficult to pinpoint the climatic causes of the $\delta^{18}O$ fluctuations at Deep Lake. However, this lake is today located in a climatically sensitive area near the junction of the Gulf-of-Mexico, Pacific and Arctic airmasses[12]. The $\delta^{18}O$ reversal at 8.9 cal. kyr BP probably resulted from the interplay of these airmasses. A change of moisture source is unlikely to have been an important factor in determining $\delta^{18}O$ values, because both the Pacific and Arctic airmasses carry little moisture and thus the dominant moisture source was probably the Gulf of Mexico throughout the Holocene. Nevertheless, as at present in the midwestern United States, the diminished influence of the warm Gulf-of-Mexico airmass in favour of the Arctic airmass would probably result in more frequent polar outbreaks, lower precipitation temperatures, and a greater proportion of precipitation as snow[12,13]. These climate changes should enhance ^{18}O depletion of precipitation[10,11,13], leading to the 2‰ decrease in $\delta^{18}O$ at Deep Lake.

The abrupt decline in $\delta^{18}O$ at ~8.9 cal. kyr BP coincides with the drainage at glacial lakes Agissiz and Ojibway to the Labrador Sea during the sudden collapse of the Hudson Bay ice dome[14–17], providing that the ^{14}C dates reported are calibrated[7]. The Hudson Bay ice dome surged repeatedly into glacial Lake Ojibway—the so-called Cochrane surges[15]—shortly before its collapse. As a result, the Hudson Bay ice dome disintegrated rapidly, and at 8.0 ^{14}C kyr BP (8.8 cal. kyr BP), lakes Agassiz and Ojibway drained northward catastrophically through the Hudson Strait[14,16]. The drainage resulted in a flux of $\sim 1.2 \times 10^5$ km^3 (~3.75 Sv if the drainage occurred within one year) of fresh water into the northwestern Atlantic Ocean and left distinct traceable sediment units in the southeastern Hudson Bay area as well as throughout the Hudson Strait. This freshwater influx would be sufficient to slow deep-water formation[18], thereby causing a significant reorganization in atmospheric circulation patterns[19]. One conceivable result of this reorganization would be that the polar front migrated southwards, and that the zonal flow was strengthened in the midwestern United States. These changes would have weakened the influence of the Gulf-of-Mexico airmass in the Midwest. In addition, after the collapse of the Hudson Bay ice dome, the Arctic airmass would have been cooled in summer as it passed over the Tyrell Sea, which was continually supplied by icebergs from the still active Quebec and Keewatin ice centres. The temperature gradient would have been steep from here to the mid-continent south and west of Minnesota, where temperatures were high because of greater summer insolation at that time. The jet stream and associated storm tracks must have been focused on this steep temperature/pressure gradient[20], and the cooled and strengthened Arctic air would prolong winter conditions. As a result, moist Gulf air entrained by storms passing over Deep Lake would have been lifted to higher altitudes and cooled to a grater extent, releasing precipitation with lower $\delta^{18}O$ values and contributing to the 2‰ decline in the Deep Lake profile at 8.9 cal. kyr BP. The icebergs that calved into the Tyrell Sea from the residual Keewatin and Quebec ice domes[16] probably sustained climate cooling until 8.3 cal. kyr BP.

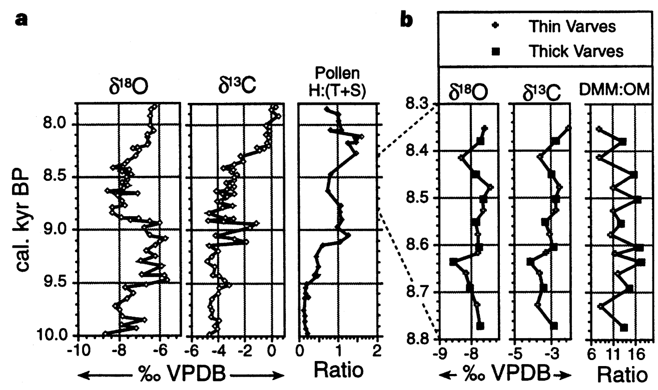

Figure 2 Proxy climate data from Deep Lake. **a**, Oxygen and carbon isotopes of bulk carbonates, and pollen ratios of total herbs to trees plus shrubs, H:(T+S), Deep Lake. **b**, Oxygen and carbon isotopes, and ratios of detrital mineral matter content (DMM) to organic matter content (OM) for thick-varve versus thin-varve bundles. For the analyses of oxygen and carbon isotopes, bulk-carbon samples were reacted at 25 °C for 1 h with 104% H_3PO_4 made from ultra-pure P_2O_5 and triply distilled H_2O (ref. 28); this reaction should not result in the substantial dissolution of detrital dolomite, which could complicate climatic interpretation of the data. The isotopic composition of the released CO_2 gas was analysed with a Finnigan MAT delta E triple-collector mass spectrometer. (The VPDB standard was used for the $\delta^{18}O$ and $\delta^{13}C$ values.) Pollen analysis followed standard procedures, and at least 300 pollen grains were counted at each level by K. Lease. Organic and carbonate contents were determined by loss on ignition at 500 and 1,000 °C; DDM = (100 − OM − carbonate content).

If the Hudson Bay deglaciation caused large changes in atmospheric circulation that resulted in the $\delta^{18}O$ reversal at Deep Lake, similar climate changes were probably widespread. A recent synthesis of stratigraphic data[19] indicates altered atmospheric circulation patterns at this time, as suggested by a significant decrease in marine productivity in the northwestern North Atlantic and by declined values of pollen influx and cellulose $\delta^{18}O$ in lake sediments from Quebec and adjacent New England. In the Greenland ice records, a number of proxies—including snow accumulation, $\delta^{18}O$-derived temperature reconstruction, and Ca and CH_4 contents—also show[1] an anomaly lasting from 9.0 to 8.0 cal. kyr BP, which may also be related to these inferred changes in atmospheric circulation resulting from the draining of lakes Agassiz and Ojibway. However, no evidence has been found that salinity and sea surface temperatures changed at this time in the northwestern North Atlantic; this is either because the freshwater flux was insufficient to cause such changes or because existing data lack adequate chronological control and sampling detail.

Abundant Holocene records of continental climate changes are available from the midwestern United States[21] and eastern North America[22], some of which show compelling evidence for marked climate reversals during the early Holocene[1]. Unfortunately, these previous records lack the dating accuracy and sampling detail necessary to test the geographical extent of the possible climate effects induced by the drainage of these large proglacial lakes. Stable-isotope analysis has been conducted on the varved sediments of Elk Lake (~60 km southeast of Deep Lake), Minnesota[23]. However, the coarse resolution of the early Holocene $\delta^{18}O$ data from this site is unsuitable for testing the $\delta^{18}O$ reversal at ~8.9–8.3 cal. kyr BP at Deep Lake. Nevertheless, a recent study of diatoms in a sediment

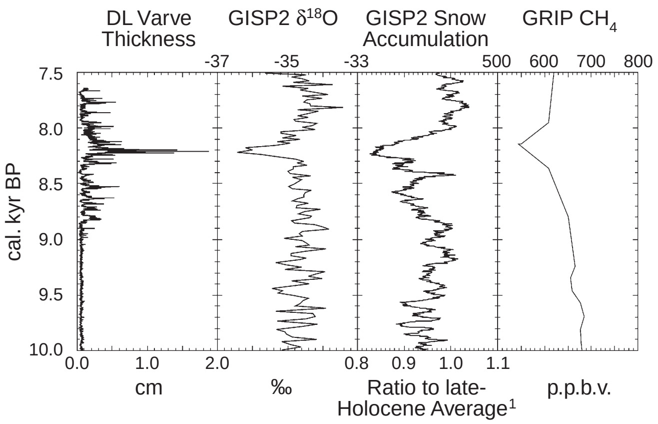

Figure 3 Palaeoclimate records from Deep Lake and Greenland. Varve-thickness data from Deep Lake are compared with records of $\delta^{18}O$ (refs 29, 30), snow accumulation ratio[1], and CH_4 concentration[2] from the Greenland ice. DL, Deep Lake; GISP2, US Greenland Ice-Sheet Project 2; GRIP, European Greenland Ice-Core Project. Note that the ^{14}C date on which the Deep Lake varve chronology is anchored has a standard error of ±85 years and that the climate reversal that peaked at 8.2 cal. kyr BP lasted from 8.0 to 8.4 cal. kyr BP in the Greenland ice-core record[1]. Varve thickness was measured with an Apple Macintosh PowerPC computer equipped with an Audio/Visual package and a Panasonic digital video camera to capture varve images. These images were then transformed with Photoshop version 2.5 (Adobe, San Jose, California, 1993) and digitized with Image (US Natl Institutes of Health, Bethesda, Maryland; available at http://rsb.info.nih.gov/nih-image).

core from Moon Lake (~100 km west of Deep Lake) in eastern North Dakota provides an early Holocene salinity record at multi-decadal resolution[24], the chronology of which is well constrained with multiple AMS ^{14}C dates of terrestrial plant materials. This record also shows a pronounced reversal at ~8.9–8.3 cal. kyr BP from the increasing salinity trend of the early Holocene. This salinity reversal was probably induced by the same climatic mechanism (particularly through decreased evaporation as a result of decreased summer temperature) responsible for the concurrent ^{18}O depletion at Deep Lake. In addition, our new initial results of high-temporal-resolution sediment analyses from Twin Lakes in northwestern Minnesota show an abrupt, pronounced decrease in $\delta^{18}O$ starting at ~9.0 cal. kyr BP, thus duplicating the Deep Lake record.

Coincident with the abrupt $\delta^{18}O$ decrease at 8.9 cal. kyr BP is an overall increase in varve thickness (Figs 2 and 3). Varve thickness in Minnesota lakes near the forest–prairie border is thought to be a sensitive recorder of aeolian activity[25]. Increased varve thickness at ~9.0 cal. kyr BP probably reflects increased aeolian dust input to Deep Lake from the northern Great Plains to the west as a result of strengthened zonal flow. This interpretation is supported by the fact that thicker varves generally have higher ratios of detrital mineral to organic matter (Fig. 2) and is consistent with the climatic interpretation of the $\delta^{18}O$ record discussed above. This climatic effect was probably strongest in the mid-continent, which was closer to the Tyrell Sea than the North Atlantic region.

The onset of the $\delta^{18}O$ reversal 8.9 cal. kyr BP at Deep Lake significantly pre-dates the prominent climate event that occurred between 8.4 and 8.0 cal. kyr BP and peaked at 8.2 cal. kyr BP. Evidence for this event has been found in various proxy records from the Greenland ice and other parts of the globe[1,5,6]. Nevertheless, a prominent peak in varve thickness occurs at 8.2 cal. kyr BP at Deep Lake (Fig. 3). Within the possible chronological errors of the Deep Lake and Greenland records, this varve-thickness peak correlates with changes in various climate proxies (for example, $\delta^{18}O$, snow accumulation, and CH_4 concentration) in the Greenland ice, providing strong evidence that the 8.2-cal. kyr event affected regions far from Greenland. If this event resulted from a temporary perturbation in thermohaline circulation in the North Atlantic circulation[1], it may have had a widespread effect on climate, causing enhanced drought intensity and aeolian activity in the North American mid-continent, as suggested by increased varve thickness and high pollen ratios of herbs to trees plus shrubs at Deep Lake (Fig. 2). Such an interpretation would seem to be supported by increased varve thickness at this time in the sediments of Elk Lake, Minnesota[25], although the early Holocene chronology of this record may be compromised by a substantial accumulative error[26] related to varve counting from the sediment–water interface. Values of $\delta^{18}O$ do not show oscillations corresponding to the varve-thickness peak at 8.2 cal. kyr BP at Deep Lake, implying that the controls are different from those for the reversal 8.9–8.3 cal. kyr BP. One possible explanation for the lack of an isotopic response at 8.2 cal. kyr BP is that $\delta^{18}O$ depletion due to climate cooling and associated reduced evaporation[10,11,13] was compensated by evaporative ^{18}O enrichment resulting from dry-climate conditions. As inferred from the much thicker varves and higher pollen ratios of herbs to trees plus shrubs at 8.2 cal. kyr BP than at 8.9–8.3 cal kyr BP, drought intensity was probably much greater at 8.2 cal. kyr after the cooling effect of icebergs in the Tyrell Sea terminated, which may have contributed to the difference of $\delta^{18}O$ signals between the two climate events.

The mechanism of the 8.2-cal. kyr BP event is a subject of growing controversy[1,5,6,19,27]. It has been a focus of recent debates whether climate cooling at this time was caused by the final collapse of the Laurentide ice near Hudson Bay and the drainage of large glacial lakes to the North Atlantic Ocean that altered the North Atlantic thermohaline circulation. Our high-temporal-resolution data of $\delta^{18}O$ and varve thickness from Deep Lake suggest that there existed two separate, abrupt climate events between 10.0 and 8.0 cal. kyr BP. The first event, characterized by decreased $\delta^{18}O$ with an overall increase in varve thickness 9.0–8.3 cal. kyr BP, was probably related to the deglaciation in the Hudson Bay region. The second event, characterized by peak varve-thickness values at 8.2 cal. kyr BP, was probably a manifestation of an abrupt climate event of unknown cause at a global scale. These data together provide evidence for the climatic complexity of the early Holocene and for the climatic linkages between the North Atlantic region and the mid-continent of North America. □

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Large-scale freshening of intermediate waters in the Pacific and Indian oceans

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Despite the central role of the oceans in the global hydrological cycle, direct observations of precipitation over the oceans are too sparse to infer global patterns of variability. For the regions of water-mass formation (the high latitudes), however, it is possible to obtain indirect information on changes in the surface salinity budget from salinity measurements elsewhere, as water masses in the ocean carry distinct signatures in temperature and salinity over long distances. Here we present a comparison of historical hydrographic data collected between 1930 and 1980[1,2] with six more-recent trans-oceanic hydrographic sections (1985–94) from the intermediate waters of the Pacific and Indian oceans[3,4]. North Pacific Intermediate Water and Antarctic Intermediate Water both show coherent basin-wide salinity decreases with time. The simplest explanation for these changes is a freshening of surface waters, over approximately 22 years, in the high-latitude North Pacific and Southern oceans, suggesting that precipitation (minus evaporation) has increased over the polar gyres. We estimate an increase by about 31 mm yr^{-1} for the Southern Ocean (between 55° S and 65° S), which is about three times larger than the values suggested by coupled atmosphere–ocean models with increasing atmospheric greenhouse-gas concentrations for the same period[5–8]. The patterns of change are, however, qualitatively consistent between models and observations, and our results provide evidence for an intensification of the global hydrological cycle over the past decades.

Observations of changes in the Atlantic Ocean's temperature and salinity structure[9–12] all suggest that the circulation and water-mass characteristics of the deep ocean have significant variability. Results from the South Pacific Ocean and Tasman Sea show coherent warming on basin-scales in Subantarctic Mode Water (300–700 m depth)[13,14]. Here we extend the geographical and depth coverage of these earlier results to North Pacific Intermediate Water (NPIW; 400–700 m) and Antarctic Intermediate Water (AAIW; 700–1,300 m), using six new zonal sections from the World Ocean Circulation Experiment (WOCE)[3,4] in the Pacific and Indian oceans (Fig. 1). These six trans-oceanic sections are located as follows: at 47° N (1985), 24° N (1985) and 10° N (1989) respectively in the subpolar gyre, subtropical gyre and equatorial region of the North Pacific; at 17° S (1994) in the subtropical gyre of the South Pacific; at 32° S (1987) in the Indian subtropical gyre; and at 43° S (1989) in the Tasman Sea[13].

Five of the sections show a well-defined salinity minimum of either NPIW (24° N) or AAIW (10° N, 17° S, 32° S, 43° S) (Fig. 2). The shape of the temperature–salinity curve is defined by the wintertime sea surface temperature and salinity where these waters sink into the ocean interior. Thus anomalies in wintertime sea surface temperature or salinity are reflected directly as anomalies in the temperature or salinity properties of the water after subduction. (We note that this is an over-simplification of the ventilation process because the mixed layer and its temperature–salinity characteristics are also sensitive to anomalies in the winds and mixing processes within the ocean interior[15–17]. However, this simplified model of ventilation does provide a guide to the interpretation of the results presented here.)

The differences between the historical and WOCE observations are clearly seen in the zonally averaged temperature–salinity curve for each section (Fig. 2) and are statistically significant at the 90% confidence level (Fig. 3). Section plots (not shown) of temperature and salinity differences on density surfaces all show coherent basin-wide changes as reflected in the averaged temperature–salinity curves. Along the five sections at 24° N, 10° N, 17° S, 32° S and 43° S, the salinity minimum has freshened (and cooled) on density surfaces, with the temperature–salinity curve now displaced to lower salinity values. Above the salinity minimum (that is, in the shallower layers), the displacement of the temperature–salinity curve to cooler/fresher values can be explained by a combination of warming and freshening of surface waters[18]. However, at the salinity minimum, changes in the wintertime sea surface temperature cannot displace the temperature–salinity curve to fresher values. Thus the observed freshening of the salinity minimum is explained most simply by a freshening of the surface waters in the source regions of NPIW and AAIW.

NPIW (indicated by the low-salinity tongue north of 15° N in Fig. 3) is confined to the North Pacific, and is observed only in the 24° N section. But along 47° N, significant freshening on density surfaces in the NPIW range (density surfaces 26.4–27.2 kg m^{-3}) is also present. The largest decreases are 0.1 practical salinity units (p.s.u.) along 47° N, and 0.017 p.s.u. along 24° N (Fig. 3a). For the 24° N section, the NPIW density surfaces have been displaced downwards on average by 22 m (Fig. 3b) and thickened. This thickening of the NPIW layer along 24° N has led to an increase in the vertical cross-sectional area of this water mass by 1.4%. The largest difference occurs in the 47° N section, with a weaker freshening signal in the 24° N section, consistent with the increased distance of 24° N from the source of NPIW in the Kuril Islands/Kamchatka regions and in the Sea of Okhotsk[19].

AAIW is found in all of the Southern Hemisphere oceans and parts of the North Atlantic and North Pacific oceans. AAIW (indicated by the low-salinity tongue south of the Equator in Fig. 3, and density surfaces 27.2–27.6 kg m^{-3}) is observed along the four sections at 10° N, 17° S, 32° S and 43° S, and shows significant freshening on density surfaces. The largest differences