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Early Holocene change in atmospheric circulation in the Northern Great Plains: an upstream view of the 8.2 ka cold event

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Abstract

Elk Lake, in northwestern Minnesota, contains numerous proxy records of climatic and environmental change contained in varved sediments with annual resolution for the last 10,000 years. These proxies show that about 8200 calendar years ago (8.2 cal. ka; 7300 radiocarbon years) Elk Lake went from a well-stratified lake that was wind-protected in a boreal forest to a well-mixed lake in open prairie savanna receiving northwesterly wind-blown dust, probably from the dry floor of Lake Agassiz. This change in climate marks the initiation of the widely recognized mid-Holocene "altithermal" in central North America. The coincidence of this change with the so-called 8.2 cal. ka cold event, recognized in ice-core and other records from the circum-North Atlantic, and thought by some to be caused by catastrophic discharge of freshwater from proglacial lakes Agassiz and Ojibway, suggests that the two "events" might be related. Our interpretation of the Elk Lake proxy records, and of other records from less accurately dated sites, suggests that change in climate over North America was the result of a fundamental change in atmospheric circulation in response to marked changes in the relative proportions of land, water, and, especially, glacial ice in North America during the early Holocene. This change in circulation probably post-dates the final drainage of proglacial lakes along the southern margin of the Laurentide ice sheet, and may have produced a minor perturbation in climate over Greenland that resulted in a brief cold pulse detected in ice cores.

1. Introduction

An early Holocene cold pulse has been documented in the oxygen-isotope records from Greenland ice cores (Alley et al., 1997). This cold pulse, dated precisely by counting annual ice layers at 8300–8100 calendar years ago (about 7400–7200 radiocarbon years ago), has an amplitude of about one half that of the Younger Dryas cold event (13,000–12,000 calendar years ago; Fairbanks, 1990). The really anomalous part of the early Holocene event lasted less than a century (R.B. Alley personal communication). This event, referred to as the "8.2 cal. ka event", has been reported from other sites around the northeastern North Atlantic Ocean (e.g., Klitgaard-Kristensen et al., 1998; von Grafenstein et al., 1998; Tinner and Lotter, 2001), the Labrador Sea (see review by Barber et al., 1999), and elsewhere (Alley et al., 1997; de Vernal et al., 1997). Barber et al. (1999) suggested that this cooling in the circum-North Atlantic region was forced by the rapid discharge of freshwater from proglacial Lakes Agassiz and Ojibway through Hudson Bay and the Hudson Strait into the Labrador Sea that marked the final demise of the Laurentide ice sheet. As postulated, this discharge was similar to the earlier episode of melt-water flow down the St. Lawrence that triggered the Younger Dryas cold event (Kennett, 1990). Those 8.2 cal. ka "events" coincided with a profound change in the relative proportions of land, water, and ice in North America, as illustrated by Dyke and Prest (1987) and Dyke et al. (1996; Fig. 1). These dramatic changes in geography also would have produced major changes in atmospheric circulation and the climate of central North America. The 8.2 cal. ka "event" in the circum-North Atlantic region is but one manifestation of the post-glacial changes in geography of North America.

Understanding the potential causes of the event on the continent requires that it be examined with several proxies with an accurate calendar-year chronology. The most desirable record would be from a continuous sequence of lake sediments with annual laminations (varves) and numerous climatically and environmentally sensitive proxy components. One lake that meets these

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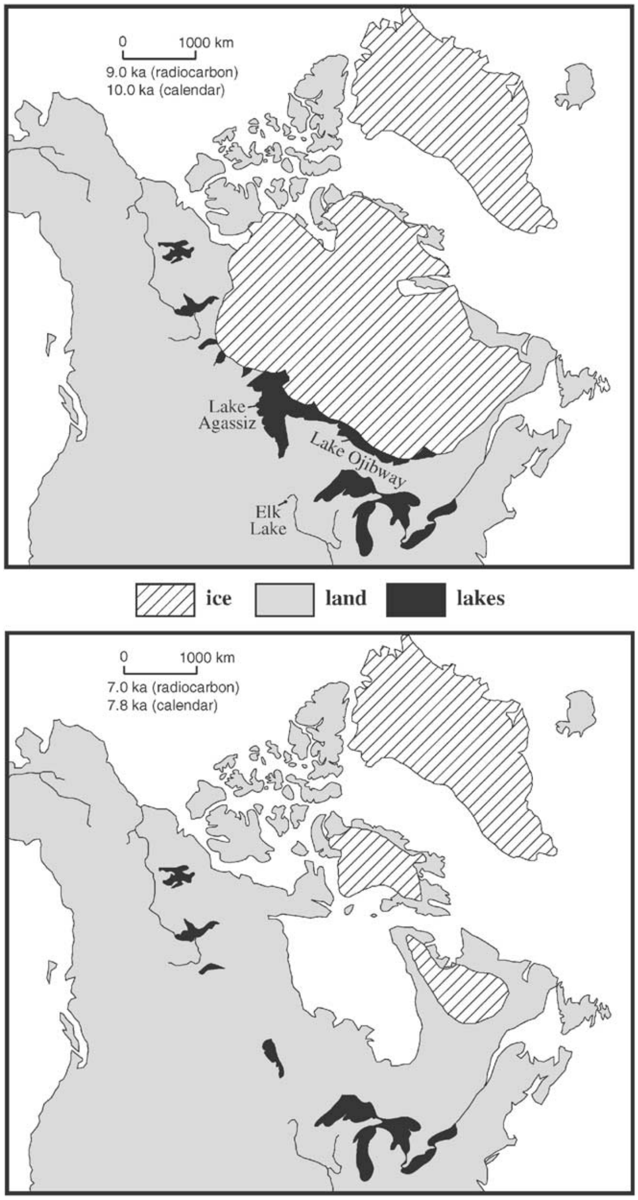

Fig. 1. Maps showing the relative distribution of land, water, and ice at 9 and 7 ka (corrected radiocarbon; 10 and 7.8 cal. ka). Modified from Dyke et al. (1996).

criteria is Elk Lake located in northwestern Minnesota in Itasca State Park at the headwaters of the Mississippi River. The lake is presently surrounded by pine-hardwood forest, but is only 80 km east of the present prairie-forest border that marks the eastern edge of the Northern Great Plains (Fig. 2A). This region is located at the intersection of the cold arctic airmass to the north, a warmer and wetter Gulf of Mexico-Atlantic airmass to the south, and dry Pacific air from the west (Fig. 2B), resulting in a steep climatic gradient across Minnesota. Variations in position, duration, and strength of these airmasses occur on all time scales from seasonal to millennial. This region is, therefore, a climatically sensitive area that is ideal for recording climate change.

The purpose of this paper is to describe climatic changes for the period centered on 8.2 cal. ka based on the varve-calibrated, multiproxy sediment record from Elk Lake. As we will show, the collective proxies imply that the climate in central North America was dominated by polar air prior to 8.2 cal. ka, and then by strong westerly atmospheric flow after 8.2 cal. ka. These major changes in airmass frequency, duration, and strength were triggered by the demise of the Laurentide ice sheet, and the reduction in the size of Hudson Bay.

2. Elk Lake, Minnesota

2.1. Setting

Elk Lake is a dimictic lake with a seasonally anoxic hypolimnion. Temperatures in the epilimnion and littoral-zone respond to seasonal air temperatures, commonly reaching $>20°C$ during the summer (Megard et al., 1993). The lake water is derived mainly from precipitation and ground-water flow, with little input from surface-water flow. Dissolved ions in the lake are dominated by calcium and bicarbonate, with dissolved solids (TDS) of about $260\,mg\,l^{-1}$. Because the surface waters of the lake are near calcite ($CaCO_3$) saturation most of the year, and above calcite saturation during the summer months (Dean and Megard, 1993), chemically precipitated (endogenic) carbonate minerals, dominated by low-magnesium calcite always were major sediment components.

2.2. Varve chronology

What makes Elk Lake unusual is that the entire Holocene sediment section in the deepest part of the lake (29.6 m) consists of millimeter-thick varves (Anderson, 1993; Anderson et al., 1993). Because the sediments are varved, the timing and rates of change of climatic and limnologic events have been determined accurately by counting annual laminae couplets, independent of the complications associated with radiocarbon age determinations. The accuracy of the Elk Lake varve chronology has been confirmed by radiocarbon dating (Anderson et al., 1993), pollen zonation (Whitlock et al., 1993), correlation between magnetic secular variation curves from Elk Lake with those from other Minnesota Lakes (Sprowl and Banerjee, 1993), and by independent varve counting in other cores (Sprowl and Banerjee, 1985; Sprowl, 1993). Significantly, the base of the varve chronology in the 1978 Elk Lake cores is 10,400 cal. years, whereas the base of the varve chronology in oriented cores taken in 1982/1983 for paleomagnetic studies is 10,120 cal. years, a difference of only 3% (Sprowl, 1993).

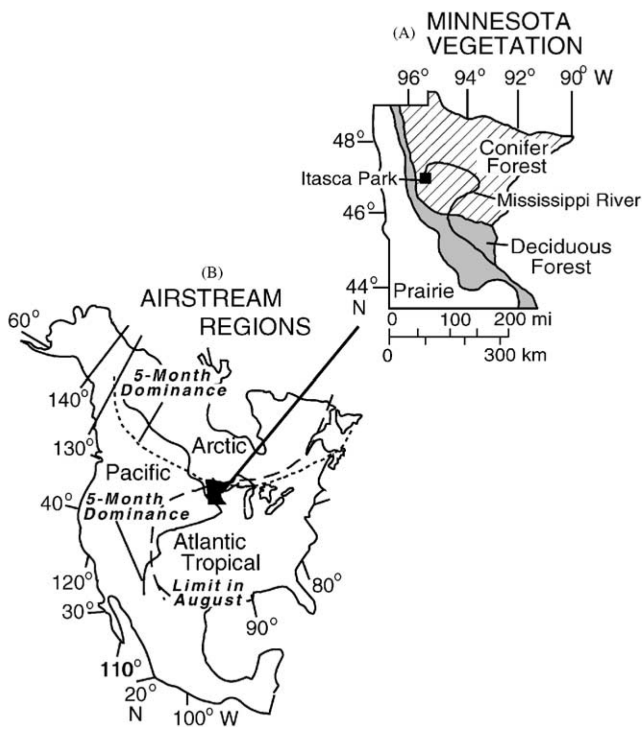

Fig. 2. Maps showing (A) vegetation zones of Minnesota and (B) air stream regions in North America (modified from Bryson and Hare, 1974).

2.3. Sediment components

The sediment components that make up the varve couplets are predominantly endogenic or biogenic ($CaCO_3$, biogenic SiO_2, organic matter, iron and manganese oxyhydroxides, and iron phosphate); allochthonous siliciclastic components are minor except in sediments deposited during the mid-Holocene (ca. 8 to 4 cal. ka; Dean, 1993). The varves form because these components are deposited as distinct seasonal pulses, with a distinct, white, summer $CaCO_3$-rich lamina. The seasonal succession of sediment components remains undisturbed because the bottom waters are seasonally anoxic, which prevents bioturbation by benthic organisms. Because no streams flow into Elk Lake, the siliciclastic material that does enter the lake is interpreted as being mostly wind-borne (eolian; Bradbury et al., 1993; Dean et al., 1996; Dean, 1997), thus providing a continuous record of wind conditions in the region.

Air currents also deliver pollen and spores that record regional vegetation changes. Pollen records throughout Minnesota show that during the mid-Holocene, the prairie-forest border migrated at least 100 km east of its present position, at which time Elk Lake was surrounded by sagebrush/grass prairie and oak savanna (Whitlock et al., 1993), and had an average annual precipitation about 100 mm lower than at present (Bartlein and Whitlock, 1993). Ostracode and diatom data from Elk Lake (Forester et al., 1987; Bradbury and Dieterich-Rurup, 1993) suggest that the early part of mid-Holocene in northwestern Minnesota was cooler throughout the year than at present, under conditions similar to those found today in lakes of the Canadian prairie, until at least 6.7 cal. ka. We use Wright's (1976) term "prairie period" for the mid-Holocene dry episode from about 8 to 4 cal. ka in northern Minnesota, marked simply by dry conditions with no necessary connotation of temperature.

2.4. External versus internal indicators of change

Several proxies in the sediments of Elk Lake record an eolian signal that is external to the lake (e.g., varve thickness, magnetic susceptibility, quartz, % Al) whereas other proxies record internal chemical and biological signals at about 8200 varve years ago (8.2 cal. ka; e.g., Mn, Fe, P, diatoms, plant pigments; Fig. 3). Some of these signals are manifested as short pulses, but most are manifested as a step-wise change in environmental conditions. Today the hypolimnion of Elk Lake is anoxic during most of the summer. However, during the early Holocene the lake may have been as much as 20 m deeper (the thickness of Holocene sediments), and may have been meromictic, as suggested by high concentrations of manganese, phosphorus (Figs. 3D and E), iron, and molybdenum in sediments deposited prior to 8.2 cal. ka.

2.4.1. Diatoms

Many planktic diatom species with heavily silicified frustules rely on turbulence in the epilimnion to keep them in the photic zone. Therefore, the kinds of diatom species present provide a direct indication of wind mixing and turbulence in the epilimnion. Early Holocene diatoms indicate low turbulence and strong stratification (Bradbury and Dieterich-Rurup, 1993), thus limiting nutrient recycling and keeping productivity low. After 8.2 cal. ka, less persistent stratification (possible break down of meromixis) and increased wind-driven circulation recycled phosphorus, which previously had been sequestered in the sediments (Fig. 3E), greatly increasing diatom productivity (Fig. 3G), dominated by the phosphate-dependent diatom *Stephanodiscus minutulus* (Bradbury and Dieterich-Rurup, 1993). The eutrophic diatom *Stephanodiscus niagarae* first appeared at about 8.2 cal. ka (Fig. 3H) indicating increased nutrient availability and lake circulation (Brugam, 1983; Bradbury and Dieterich-Rurup, 1993). Increased productivity at this time is further indicated by an abrupt increase in concentrations of chlorophyll and carotenoid pigments in the sediments (Fig. 3F; Sanger and Hay, 1993).

2.4.2. Detrital clastic material

The wind-mixed lake also received larger amounts of detrital clastic material as recorded by quartz and numerous lithophile elements such as Al, Si (Figs. 3B and C), Mg, Ti, K, and many trace elements interpreted as having been transported by eolian dust (Bradbury et al., 1993; Dean, 1993). Other manifestations of increased detrital influx are increased varve thickness (Fig. 3A) and magnetic susceptibility (Bradbury et al., 1993; Sprowl and Banerjee, 1993; Dean et al., 1996). Note that increases in Al, Si and varve thickness at 8.2 cal. ka appear as pulses but do not return to levels recorded prior to 8.2 cal. ka, and, in fact, remain high throughout the mid-Holocene section until about 5.3 cal. ka (Fig. 3). The high concentrations of Si (Fig. 3C) after 8.2 cal. ka are thus the result of a combination of higher biogenic silica production, as well as increased influx of siliciclastic material, mostly eolian quartz (Dean, 1993). The source of this siliciclastic material was probably the dried, exposed floor of Lake Agassiz, which was north and west of Elk Lake, in the path of the dominant northwesterly winds during the mid-Holocene (Dean et al., 1996).

2.4.3. Pollen

The vegetation response to the drier mid-Holocene prairie period was much more gradual, and preceded the change in state of Elk Lake at 8.2 cal. ka. Prairie vegetation that defines the prairie period, indicated by *Artemisia* (sagebrush; Fig. 3I) and Gramineae (grass; not shown), began to replace the early Holocene pine forest (Fig. 3J) at about 8.8 cal. ka, and was most

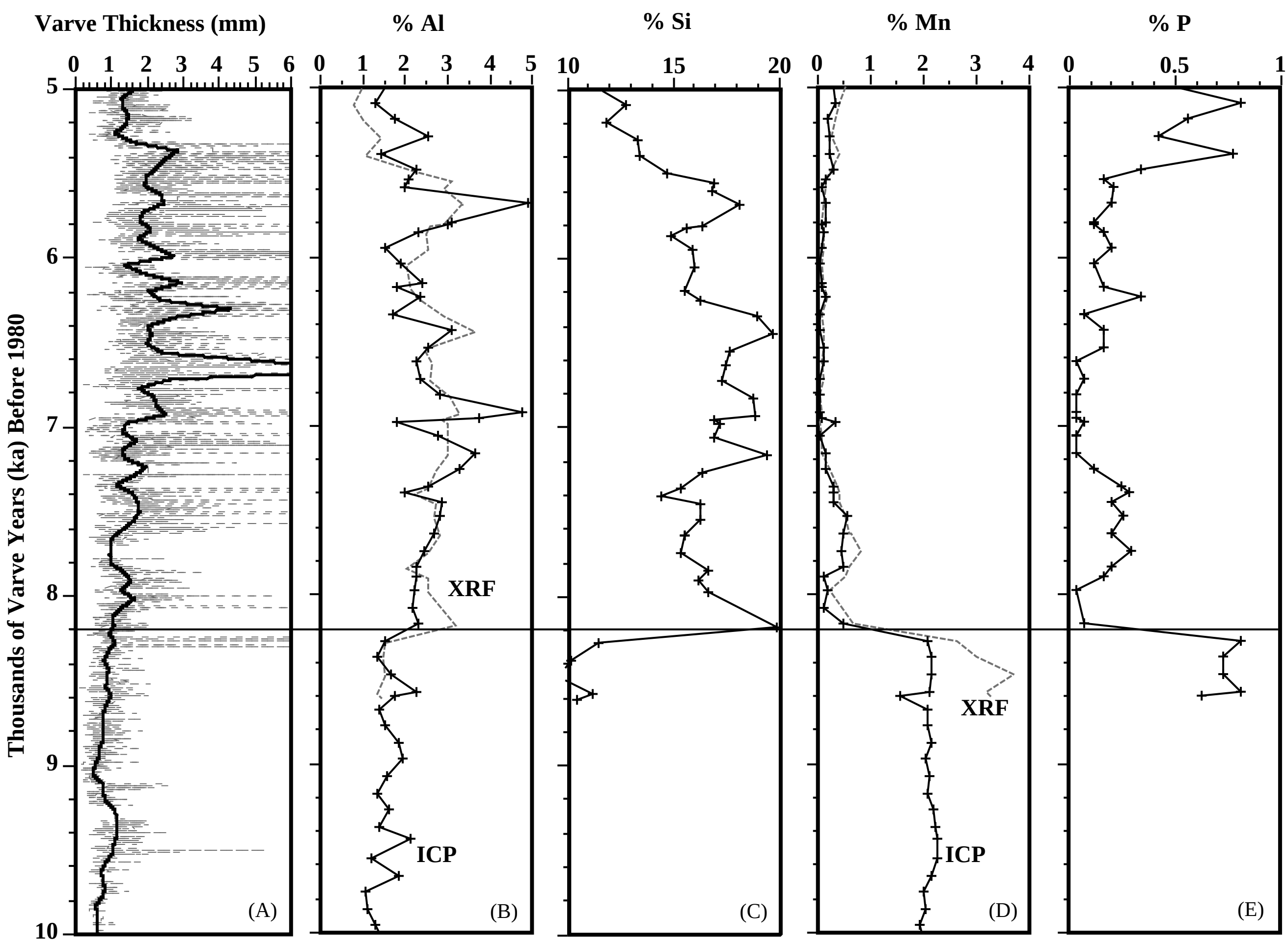

Fig. 3. Profiles of (A) varve thickness; (B) %Al; (C) %Si; (D) %Mn; (E) %P; (F) chlorophyll plus carotenoid pigments in units per gram organic matter; (G) total diatom flux in diatoms/cm^2/yr; (H) % *Stephanodiscus niagarae*; (I) % *Artemesia*; (J) % *Pinus bansiana* and % *Quercus*; (K) values of δ^{18}O; (L) values of δ^{13}C in sediments deposited in Elk Lake between 10 cal. ka and 5 cal. ka. XRF = X-ray fluorescence. ICP = inductively coupled argon plasma emission spectrometry. High concentrations of iron and manganese in sediments deposited before 8.6 ka caused fluorescence and interference with the XRF analyses, and hence there are no XRF results for those samples. Values of δ^{18}O and values of δ^{13}C are expressed in the usual per mil (‰) δ-notation relative to the Vienna carbonate standard VPDB (Vienna Pee Dee Belemnite) for carbon and oxygen, $\delta‰ = [(R_{\text{sample}}/R_{\text{PDB}}) - 1] \times 10^3$, where R is the ratio ^{18}O:^{16}O or ^{13}C:^{12}C. A–E are from Dean, 1993. F is from Sanger and Hay, 1993. G and H are from Bradbury and Dieterich-Rurup, 1993. I and J are from Whitlock et al., 1993. Values of δ^{18}O and δ^{13}C for the sublittoral core (K and L) are from Dean and Stuiver (1993). See those publications for methods of analyses.

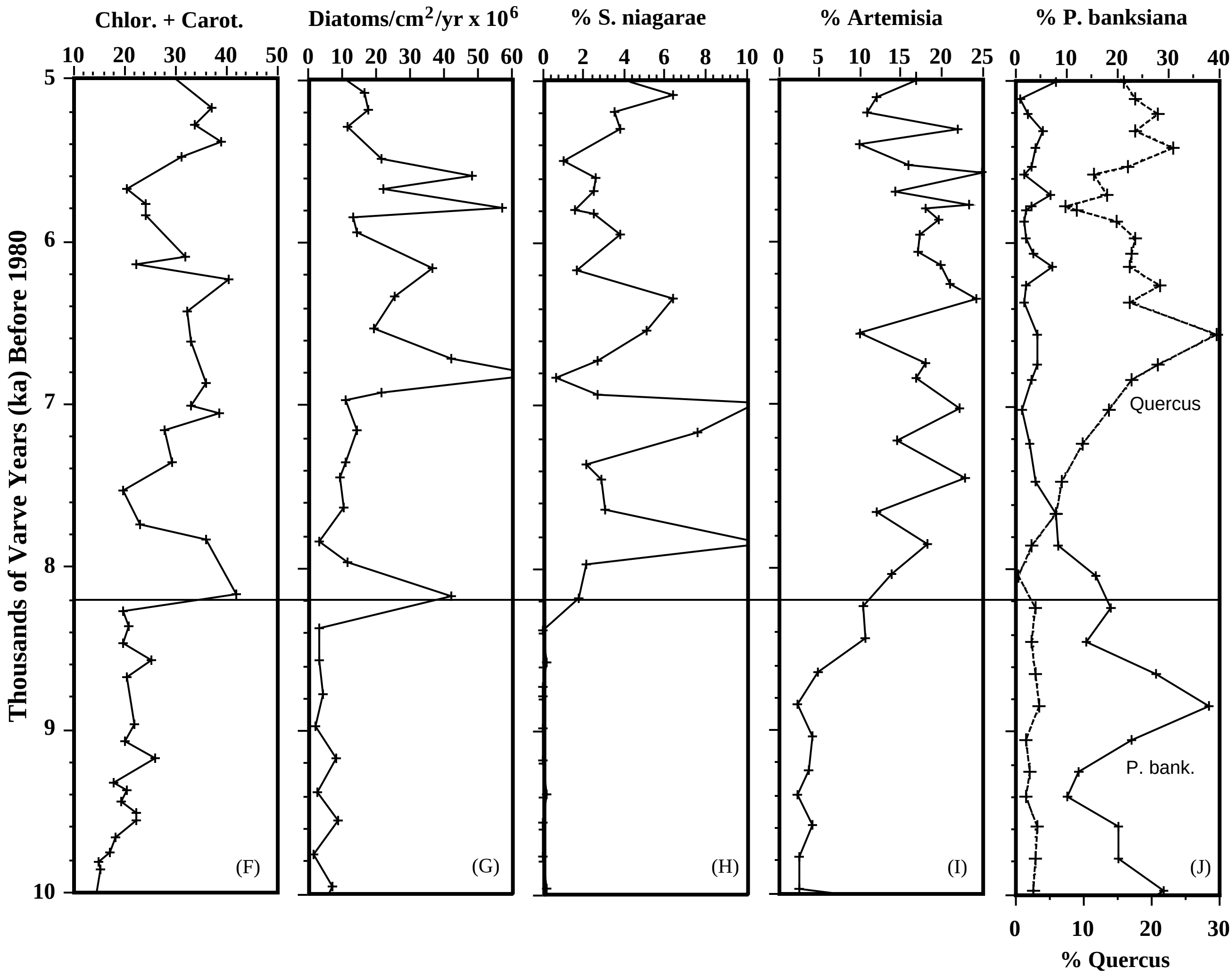

Fig. 3 (*continued*).

abundant after 8.0 cal. ka (Whitlock et al., 1993). An increase *Quercus* (oak; Fig. 3J) also began about 8.0 cal. ka, signaling the initial development of an oak-grassland with scattered sagebrush that characterized most of northwestern and north-central Minnesota (McAndrews, 1966; Wright, 1976, 1992). Reconstruction of annual precipitation by comparing the Elk Lake data with modern pollen data shows an abrupt decrease in precipitation at 8.0 cal. ka of about 200 mm/yr (Bartlein and Whitlock, 1993). From the above evidence, it appears that between about 8.6 and 8.0 cal. ka, Elk Lake went from a well-stratified lake in a pine forest that protected the lake from wind mixing, to a well-mixed lake in dry, open prairie and oak savanna receiving wind-blown dust.

2.4.4. Stable isotopes

The stratigraphic profile of $\delta^{18}O$ in calcite (Fig. 3K) shows stratigraphic changes that fall into ranges of about 1–2‰ or less for millennial to century scale episodes. However, there is no major change around 8.2 cal. ka that might reflect wind driven evaporation or other climate. A range of 2‰ in $\delta^{18}O$ throughout the Holocene is very small considering the location of the site, where climate change was great in terms of major climate states (glacial to interglacial), moisture sources (polar to tropical), temperature regimes (cold annual to large seasonal changes), and evaporation (boreal forest to prairie).

There are two probable factors that explain most of the variations in $\delta^{18}O$ in the Elk Lake data. The first, and most important factor, is that Elk Lake is a through flowing, hydrologically open system, and probably was always that way throughout the Holocene. The volume of Elk Lake is likely small relative to the ground-water flow supporting it, and consequently the residence time of the lake is short. As a through flowing system, the $\delta^{18}O$ value of the lake water will typically remain close to the value of local ground-water recharge under any climate. Although not measured, recharge in this location probably comes largely from winter precipitation, because dense vegetation cover should take up

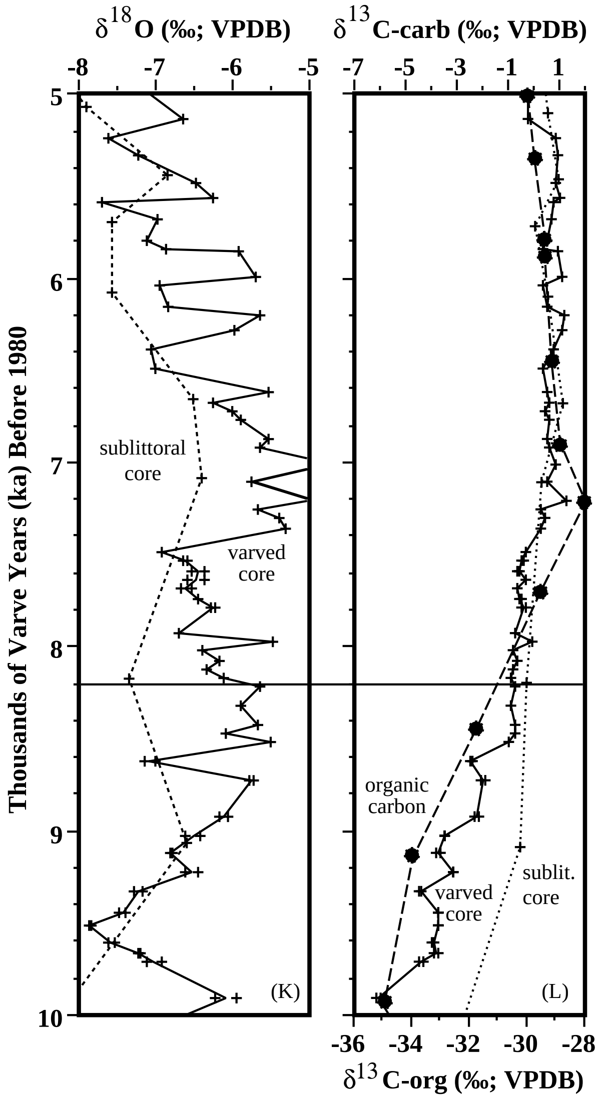

Fig. 3. Continued

much of the summer precipitation. The $\delta^{18}O$ value of winter precipitation will vary from year to year, but the values of the resulting recharge likely will not vary greatly, typically no more than a few per mil, so through flowing lake water also will have a narrow range.

The second factor is water temperature, where a 1‰ shift is equal to about a 4–5°C change in summer water temperature. Temperature changes are likely coupled with small changes in the $\delta^{18}O$ values of the water, where during cold periods lake water was close to the recharge $\delta^{18}O$ value, and during warm periods the $\delta^{18}O$ value of the water was somewhat elevated relative to the recharge value due to evaporation. For example, during the episode from about 8.2 to 8.6 cal. ka, *Cytherissa lacustris* dominates the ostracode assemblage, and the $\delta^{18}O$ values of marl in the varved core are high suggesting cold summer temperatures. The loss of *C. lacustris* after 8.2 cal. ka and the decrease in $\delta^{18}O$ values of marl in the varved core likely signifies a rise in summer water temperatures. The abrupt increase of about 1.5‰ beginning at 7.5 ka, however, may signify a decrease in recharge (increase in residence time) and increased evaporation. That time corresponds to a period in the lake's history when ostracodes indicating warmer conditions (*Physocypria* spp.) were common (Forester et al., 1987), and aragonite is recorded in the varved core (Dean, 1993). The shift from calcite to aragonite precipitation likely reflects increased evaporation and residence time that elevated the Mg/Ca ratio of the water. This change in carbonate mineralogy, together with greater warmth implied by the ostracodes, suggests that evaporation was more intense at that time. The mineralogy of carbonate shifts back from aragonite to calcite at about 6.3 ka, and this is the only interval in the core that contains aragonite (Dean, 1993). Values of $\delta^{18}O$ in marl in the sublittoral core generally parallel those in the deep varved core (Fig. 3K; Dean and Megard, 1993), but are up to 1‰ lower than in the varved core suggesting that sublittoral waters were generally warmer than those in the open epilimnion.

The $\delta^{13}C$ values of marl (Fig. 3L) increase from –6‰ at 10 cal. ka to about 0 ‰ at 8.5 cal. ka, and then remain between 0 and +1‰ until 5.2 cal. ka, when they gradually decrease to modern values of about −4‰ (Dean, in press). The $\delta^{13}C$ values of the organic matter follow a similar structureless trend (Fig. 3L). The change from low to higher $\delta^{13}C$ values could reflect a change in the catchment from predominately C_3 plants (e.g., pine forest) to predominantly C_4 plants (grasses) having values of $\delta^{13}C$ that average about −27 and −12‰, respectively (Cerling and Quade, 1993). Therefore, the dissolved inorganic carbon (DIC) derived from C_4 plants would have a much higher $\delta^{13}C$ value than C_3-derived DIC. This ^{13}C -enriched DIC discharging into Elk Lake, might produce the observed calcite $\delta^{13}C$ (−6 to +1‰) without other carbon sources. By contrast, earlier C_3-produced DIC must mix with a ^{13}C -enriched carbon source such as dissolved marine carbonate ($\delta^{13}C$ about +4‰) in order to produce the observed lower calcite $\delta^{13}C$ values. Because there are Cretaceous and Paleozoic marine carbonate rocks in the calcareous tills around Elk Lake (Wright, 1993), some portion of that higher $\delta^{13}C$ source always must have contributed to the DIC pool in Elk Lake. Therefore, if C_4 plants played a role in the Elk Lake carbon budget, they probably did so to a limited degree.

Alternatively, the gain in $\delta^{13}C$ values could simply reflect an increase in productivity within Elk Lake. The most commonly cited model for ^{13}C enrichment by increased productivity is the cultural eutrophication of Lake Greifen, Switzerland (McKenzie, 1985; Hollander and McKenzie, 1991). As a result of increased productivity of Lake Greifen during the 20th century and burial of ^{13}C-depleted organic carbon, the entire carbon reservoir of the lake became progressively enriched in

^{13}C, as did algal organic matter and precipitated $CaCO_3$ (McKenzie, 1985; Hollander and McKenzie, 1991). Applying this model to Elk Lake, the high $\delta^{13}C$ values from about 8.4 to 5.2 cal. ka reflect the high wind turbulence, higher nutrient flow, and other factors that led to a gain in lake productivity as discussed above, even though Elk Lake remained hydrologically open. Independent evidence for increasing productivity is the fact that the organic carbon accumulation rate increased from <2 to $5\,mg\,C\,cm^{-1}\,yr^{-1}$ between 10 and 7 cal. ka (Dean, 1993). It appears that an equilibrium, perhaps between productivity and through flow, was reached in Elk Lake about 8.5 cal. ka, although slight enrichment continued until about 7 cal. ka. (Fig. 3L). By 10 cal. ka, calcite in the sublittoral core was already enriched by about 3‰ relative to the varved core (Fig. 3L), perhaps reflecting early establishment of macrophyte productivity in the littoral zone. By 8.5 cal. ka, ^{13}C-enrichment of calcite in the varved core had reached that of calcite in the sublittoral core, and the two records are essentially identical thereafter. In either of the above scenarios the change in $\delta^{13}C$ from low to high values occurs before 8.2 cal. ka, and although increasing wind activity may contribute to the higher $\delta^{13}C$ values after 8.5 cal. ka, it does not result in a noticeable increase in $\delta^{13}C$ at 8.2 cal. ka.

2.4.5. Ostracodes

The ostracode record from Elk Lake comes from the sublittoral core (10 m water depth; Forester et al., 1987), in which the chronology was established by comparing selected bulk-metal analyses, but especially sodium, from this core to similar analyses from the long varved core taken in the deepest part of the lake (Fig. 4). The comparison between the two sodium time series was made using the AnalySeries software package (Paillard et al., 1996). This comparison provides a means of transferring the varve-year ages from the varved core to the unlaminated sublittoral core. In Fig. 3 we only show the records from 5 to 10 cal. ka, but in Fig. 4 we show the entire Holocene time series from both cores to show all points of correlation, particularly those between 4 and 5 ka. We use sodium as a proxy for low available moisture and, therefore, reduced decomposition of plagioclase feldspar, the dominant mineral residence of sodium (Dean, 1997). The interval between 8.2 and 5.8 cal. ka is the only interval in the varved core where plagioclase feldspar was detected by X-ray diffraction (Dean, 1993). Extrapolation of ages from the base of the varved section to the non-varved sediments deposited before 10.4 cal. ka places the base of the sediment section at 11.6 cal. ka (Whitlock et al., 1993), which is consistent with the age of 11.9 cal. ka of radiocarbon dated wood at the base of the sublittoral core (Fig. 4; Anderson et al., 1993).

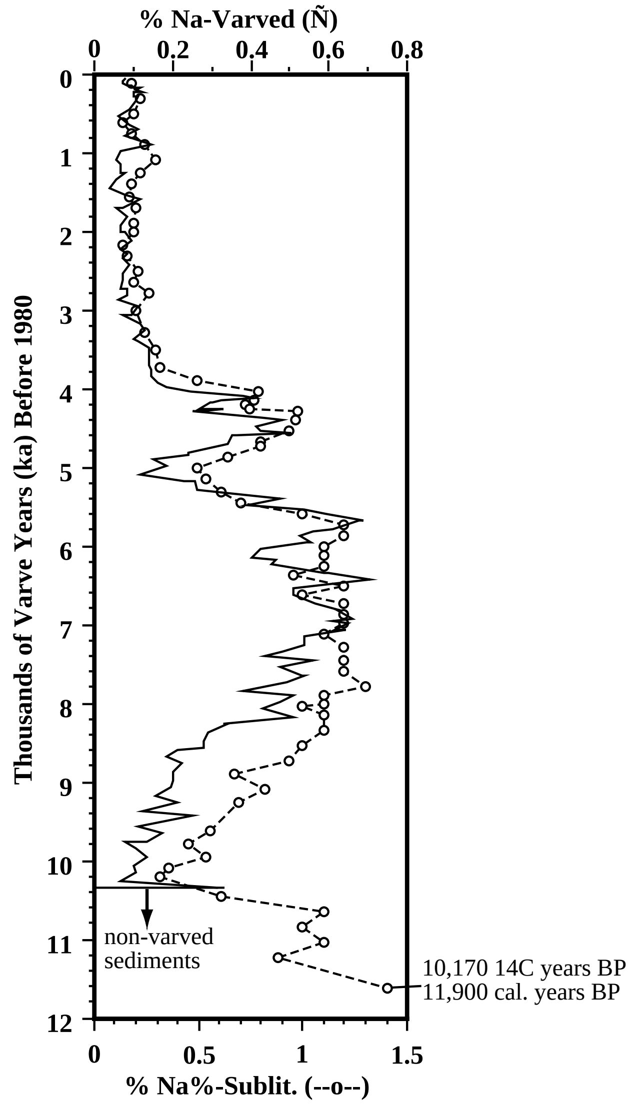

Fig. 4. Bulk-sediment values of percent sodium (Na) in the deep varved core and shallow sublittoral core from Elk Lake. Values in sediments from the sublittoral core were adjusted using AnalySeries software (Paillard et al., 1996 and available electronically at: www.agu.org/eod_elec/96097e.html) to fit with those from the varved core in order to transfer the measured varve-year time scale in the varved core to the sublittoral core.. A piece of wood at the bottom of the sublittoral core was radiocarbon dated at 10,170 years B.P. (11,900 cal. years B.P.).

The distributions of ostracode species living in lakes and wetlands are strongly related to latitude, indicating that air-temperature is a factor determining their biogeography (Delorme, 1969; Forester, 1987). Ostracode species occurrences are also related to water chemistry, in terms of both composition and concentration of the solutes (Forester, 1983, 1986, 1991; Smith, 1993). Thus, stratigraphic changes in species are linked to changes in both air temperature and effective moisture, at least as those factors are related to the limnology and hydrology of a lake or wetland (Smith and Forester, 1994).

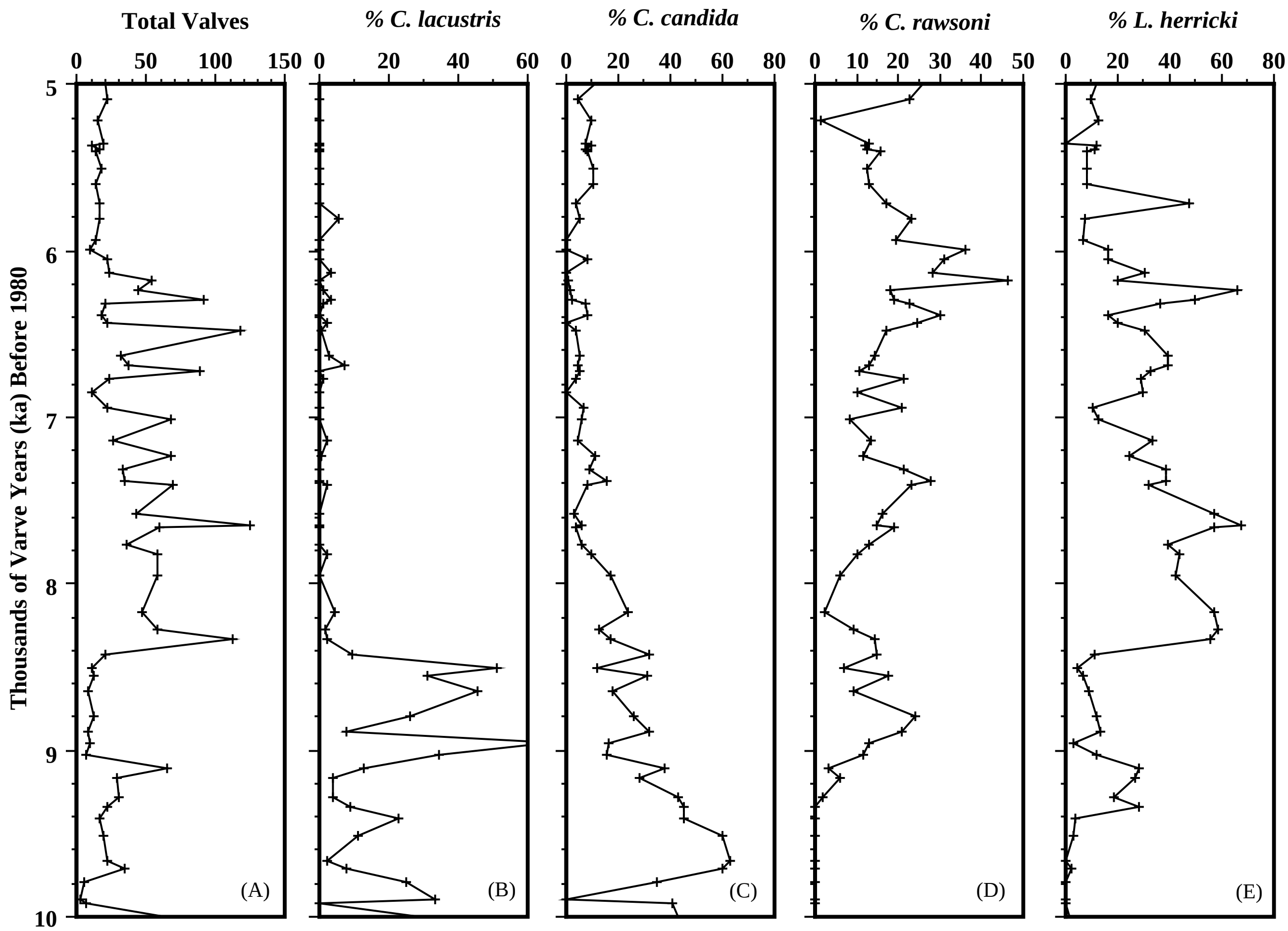

Fig. 5. Profiles of (A) % total ostracode valves counted per unit volume of sediment; (B) % *Cytherissa lacustris*; (C) % *Candona candida*; (D) *Candona rawsoni*; and (E) % *Limnocythere herricki* all relative to total valves counted.

Four species of ostracodes, *Candona rawsoni*, *Candona candida*, *Cytherissa lacustris*, and *Limnocythere herricki*, are especially important for reconstructing past climates of central North America. In general, *C. rawsoni* is a prairie-lake species that lives in both permanent and ephemeral, saline to fresh-water lakes in United States and Canada. It is well adapted to highly variable physical environments. *L. herricki* also is a prairie species that does well in highly variable, but relatively fresh-water, environments. It appears to be largely restricted to the Canadian prairies, although it is also known from high intermountain grasslands in Colorado. *C. lacustris* lives in cool to cold fresh-water lakes in the boreal forests of Canada and Alaska. As far as is known, it cannot tolerate physical and chemical environmental variability. Lastly, *C. candida* lives in lakes in Canada and Alaska, and rarely in forested areas of northern United States. It lives in lakes with higher salinities than those inhabited by *C. lacustris* (>300 mg/l; Delorme, 1989), and tolerates limited environmental variability compared to *C. lacustris*.

Ostracode valves are abundant in Elk Lake sediments deposited prior to about 9.1 cal. ka and after about 8.4 cal. ka (Fig. 5A). In the interval where ostracodes are uncommon (9.1–8.4 cal. ka), the dominant ostracode is *C. lacustris* (Fig. 5B), but the other three species also are present. Low total ostracode abundance (Fig. 5A), and the dominance of *C. lacustris* (Fig. 5B) and *C. candida* (Fig. 5C) in Elk Lake prior to 8.4 cal. ka suggest a wet, cold, polar-air dominated annual climate as occurs today in Canada and southern Alaska (Forester et al., 1989). Conversely, the dominance of *C. rawsoni* (Fig. 5D) and *L. herricki* (Fig. 5E) after 8.3 cal. ka implies a prairie-lake setting with polar-air domination in winter, and westerly flow in summer like the present-day prairies of Canada.

3. Records from other lakes in Midwestern USA

The oxygen-isotope record for the early Holocene in Deep Lake, another lake with varved sediments just north of Elk Lake, shows an increase in values of $\delta^{18}O$ in bulk carbonate from about −8‰ to −6‰, then a decrease of about 2‰ (Hu et al., 1999). This 2‰ maximum in occurs over a time interval from about 9.8

to 8.8 cal. ka. Thus, the Deep Lake oxygen-isotope "event" apparently predates the prominent 8.2 cal. ka cold event of the circum-North Atlantic. Rates and timing of events in the lower part of the Deep Lake core are calibrated by varves, but the absolute age scale is fixed by one radiocarbon date on wood (8090 ^{14}C ka), which could be off by thousands of years if the wood was reworked. The 2‰ maximum in the Deep Lake oxygen-isotope profile is similar in shape and magnitude (−8 to −6‰) as the maximum in the Elk Lake profile beginning about 9.8 cal. ka, but the Elk Lake maximum is more than 1000 years longer, ending at about 7.5 ka (Fig. 3L). A peak in varve thickness in the Deep Lake core at 8.2 cal. ka may correspond to the peaks in varve thickness in Elk Lake between 8.3 and 7.7 cal. ka (Fig. 3A), but uncertainties of a floating varve chronology and the short segment (ca. 2 kyr) of the Deep Lake time series presented by Hu et al. (1999) make comparison difficult.

A comparison of pollen profiles from about 26 lakes, ponds, and bogs in Minnesota, South Dakota, and Iowa by Whitlock et al. (1993) suggested that the increase in prairie forbs began about 1000 radiocarbon years earlier in South Dakota and western Minnesota (ca. 9 ^{14}C ka, 10 cal. ka) than in eastern Minnesota (8 ^{14}C ka, 9 cal. ka). The exception was Elk Lake, which, when the varve chronology is converted to radiocarbon chronology, shows that the beginning of the prairie period in Elk Lake was about the same (8.8 cal. ka; Fig. 3I) as in the better dated eastern Minnesota sites. Whitlock et al. (1993) concluded that the beginning of the prairie period was not time transgressive, but that the radiocarbon dates from the other western sites were too old because of a carbonate error.

Estimates of past lake-water salinity from diatoms in a core from Moon Lake in eastern North Dakota inferred that the lake went from an open lake of low salinity to a closed saline lake between 10.0 and 7.3 ^{14}C ka (11 and 8 cal. ka), corresponding to the transition of the region from spruce forest to prairie (Laird et al., 1996). This long increase in salinity in Moon Lake is in contrast to an apparent abrupt increase in salinity of Medicine Lake, South Dakota, at about 9.2 ^{14}C ka (10.2 cal. ka; Radle et al., 1989), and of Devils Lake, North Dakota about 8.0 ^{14}C ka (9.0 cal. ka; Fritz et al., 1991). Laird et al. (1996) suggested that the differences in rate and timing of early Holocene salinity among lakes in the Northern Great Plains might be due to differences in hydrology or regional variability of available moisture, but that differences in chronological control were more likely. Paleolimnological proxies (diatoms, ostracodes, isotopes, geochemistry, and mineralogy) for a suite of lakes in western Minnesota indicate a gradual increase in aridity beginning in the early Holocene with an accelerated rate of increase after 8 ^{14}C ka (9 cal. ka) (see Fritz et al., 2001 for which lakes and references). We agree with the conclusion of Whitlock et al. (1993) that the expansion of aridity that marks the beginning of the prairie period in the Northern Great Plains was synchronous at about 8.8 cal. ka (e.g., rise in sagebrush, decline in pine, Figs. 3I and J), and we further suggest that this expansion intensified at 8.2 cal. ka as defined by the varve-calibrated proxies in Elk Lake (e.g., increases in Si, Al, magnetic susceptibility, and varve thickness; decrease in Mn, Fe, and P, Fig. 3).

An early Holocene melt-water event associated with low levels of the Great Lakes was recognized from low values of δ^{18}O in ostracodes and mollusks collected from cores from Lake Huron and Georgan Bay (Rea et al., 1994a, b; Moore et al., 2000). This event was dated at 8.0–7.5 ^{14}C ka, but when corrected (Rea et al., 1994a) and calibrated to calendar years, the age of melt-water outflow becomes 8.9–8.5 cal. ka (Moore et al., 2000), and thus postdates by several thousand years the last melt-water pulse down the Mississippi River and into the Gulf of Mexico (melt-water pulse IB of Fairbanks, 1989, 1990; 11.5–10.5 cal. ka), and predates by several hundred years the final draining of the proglacial lakes as defined by the 8.2 ka event (Barber et al., 1999). Ostracode and oxygen-isotope evidence (Forester et al., 1994), and sedimentologic evidence (Colman et al., 1990, 1994a, b) from cores from southern Lake Michigan suggest that the level of the lake was lower and more saline at 8.2 cal. ka due to increased evaporation. The evidence for a melt-water flood event at 8.2 ka in the Great Lakes is, therefore, equivocal, and subject to the usual complications associated with radiocarbon chronologies.

4. Discussion

The 8.2 cal. ka cold event, originally identified in ice cores from Greenland, and apparently identified in other terrestrial and marine records from the circum-North Atlantic, is believed to have been caused by a catastrophic glacial-lake melt-water event, which resulted in a cold climate episode in the region (Barber et al., 1999). This melt-water event has been repeatedly dated at about 8 ^{14}C ka, mainly by radiocarbon dating of marine molluscs (e.g. Dyke et al., 1996; Andrews et al., 1999; Licciardi et al., 1999). The data presented by Barber et al. for the timing of this event, and especially of a "red bed" in the Hudson Strait, interpreted to have been deposited during a catastrophic outburst flood from Lakes Agassiz and Ojibway, average about 8200–8100 conventional (uncorrected) radiocarbon years ago. Using the reservoir corrections given by Barber et al., these radiocarbon ages yield calibrated ages of about 8500 calendar years ago (8.5 cal. ka). A date of 8.5 cal. ka for the discharge event is more in line with the 1 kyr time slice maps of Dyke et al. (1996) which show that the lakes are gone by 8000 corrected (but not calibrated) radiocarbon years ago (corrected by subtracting a 400 year marine reservoir effect from raw radiocarbon dates on marine molluscs). A date of 8000 corrected radiocarbon years calibrates to an age of about 8.5 cal. ka, which is also consistent with the melt-water event identified in Lakes Huron and Michigan at 8.9–8.5 cal. ka (Moore et al., 2000). Even though drainage of the large proglacial lakes must have had a significant influence on climate, the last record of any outflow apparently was several hundred years before 8.2 cal. ka. However, all of these calibrated radiocarbon ages are dependent upon what reservoir correction was used, and calibrated ages could easily be off by several hundred years in either direction.

The proxies from Elk Lake that allow reconstruction of the lake's physical and chemical paleolimnology (diatoms, concentrations of redox-sensitive trace metals, and $\delta^{18}O$ values) show that prior to about 8.2 cal. ka the lake was a stable, dimictic lake that was strongly stratified, at least seasonally. The same proxies show that after 8.2 cal. ka the lake was a turbulent, seasonally well-mixed and shallower lake than it was prior to 8.2 cal. ka.

The proxies that are related to climate factors external to the lake (dust as % Al and % Si, varve thickness, and pollen) show that prior to 8.2 cal. ka the lake was receiving relatively little dust, implying little wind activity. At 8.2 cal. ka, % Si rises dramatically (Fig. 3C) implying a large influx of dust and, by inference, a dramatic increase in wind activity. Another consequence of greater wind activity was greater wind mixing in the lake with weaker stratification allowing phosphorus to be recycled and diatoms to bloom, which also contributes to the high silica values. The varves are relatively thin prior to 8.2 cal. ka, but begin to increase just prior to 8.2 cal. ka (Fig. 3A), reflecting the higher dust flux into the lake, and from that a change from a protected lake to a less protected lake after 8.2 cal. ka. The pollen data indicate that between 8.6 and 8.0 cal. ka, the pine forest around Elk Lake was replaced by sagebrush/grass prairie and oak savanna. Lastly, the ostracode faunal assemblages, which provide information about the limnology and watershed characteristics, indicate that, prior to 8.2 cal. ka, the lake was a stable, dilute environment with characteristics typical of lakes in boreal forests. At 8.4 cal. ka, the ostracode assemblage abruptly shifts to an assemblage typical of relatively fresh Canadian prairie lakes with large seasonal variability in physical characteristics.

The climate proxies from Elk Lake all point to a substantial change in the annual/seasonal physical and chemical dynamics of the lake, as well as significant change in wind energy. For conceptual purposes, one might envision those changes in Elk Lake by imagining a lake in a present-day setting similar to northern Michigan or southern Ontario from about 9.1 cal. ka to about 8.4 cal. ka. Then, about 8.4 cal. ka, the setting changed in roughly one or two centuries to one similar to the present-day Manitoba prairie, where it stayed for about 3000 years, and then shifted to more or less its present setting in a few centuries. These environmental changes were not unique to Elk Lake because other lacustrine records in the Northern Great Plains also record environmental change around 8.2 cal. ka (with variations of about ± 1 ka due to uncertainties in dating; see review by Fritz et al., 2001), indicating that the driver of these changes must be regional or global in extent.

Examination of continental glacial ice, large lakes, and Hudson Bay from about 10 cal. ka (9 ^{14}C ka) through about 5 cal. ka (4.2 ^{14}C ka) (Dyke and Prest, 1987; Dyke et al., 1996), shows that much of the ice and water were gone before the 8.2 cal. ka, and that modern configurations of ice and water were reached shortly after 7.8 cal. ka (7.0 corrected ^{14}C ka). The loss of the last major quantity of continental glacial ice and the associated reduction in volume of Hudson Bay water was around 8.2 cal. ka (Dyke and Prest, 1987), although owing to the constraints on radiocarbon ages, the exact timing of those loses is unknown. However, the timing and rates of environmental changes in the Elk Lake sedimentary record indicate that a significant change in atmospheric circulation occurred rapidly about 8.2 cal. ka. Because, the last drainage of proglacial lakes occurred earlier (ca. 8.9–8.5 cal. ka), the environmental changes recorded in Elk Lake about 8.2 cal. ka must be due to factors other than the drainage of proglacial lakes. If the dust source attributed to thicker varves in the Elk Lake record did indeed come from the dry floor of Lake Agassiz, then that too establishes the sequence of events. The losses of continental ice and Hudson Bay water around this time suggest that this change in geography may have been responsible for the changes in atmospheric circulation interpreted from the Elk Lake proxy records.

The distribution of continental ice (Fig. 1) and large cold water bodies influence the temperature and pressure of the overlying atmosphere, and when the ice and cold water are widely distributed the entire polar airmass is affected. The ice and water distributions also provide general indications of the summer locations of the polar front relative to Elk Lake. For example, the continental ice sheet was just north of the Great Lakes at 10 cal. ka (9 ^{14}C ka), so the summer position of the polar front was south of the ice, associated with the steep atmospheric thermal gradient at the ice-land boundary. By 9 cal. ka, continental ice had retreated northward, but Hudson Bay, which was in ice contact, was much larger than today. Perhaps Hudson Bay at that time sustained cold, high-pressure air over it, and, if so, the summer position of the polar lows would have remained south of Hudson Bay. The dominance of *Cytherissa lacustris* in Elk Lake from about 9.1 to about 8.4 cal. ka suggests that polar air masses persisted in the region during summer, just as they do in the region where this ostracode lives today in Canada and Alaska.

By 8 cal. ka, Hudson Bay was smaller than at 9 cal. ka, but was larger than it is today and still had limited ice contact (Dyke et al., 1996). The smaller and perhaps warmer Hudson Bay likely did not support as cold an airmass over it as before. So during the summer, polar air was likely warmer, resulting in weaker polar high-pressure cells and a more northerly position of the polar lows than at 9 cal. ka. This northward movement of the polar front in summer would allow westerly flow to move northward into the Upper Midwest of the United States, as is indicated by the ostracodes and wind proxies in Elk Lake. Thus, the rapid (century scale) transition from low-wind, polar-low-dominated summers to summers dominated by strong westerly winds appears to be associated with the reduction in the extent of Hudson Bay that, in turn, is associated with the final stage of the loss of continental glacial ice.

This interpretation of the 8.2 cal. ka event is partly consistent with paleoclimate model simulations for the last 18 in 3 kyr steps. Those simulations indicate that during the 9–3 ka interval, westerlies were stronger (than present) in July, reaching a maximum about 6 ka (Kutzbach and Guetter, 1986, Kutzbach, 1987, COHMAP members, 1988). The Elk Lake wind proxies suggest the strongest winds occurred between 7.0 and 5.5 cal. ka (Bradbury et al., 1993; Dean et al., 1996), and then abruptly decreased within a decade at 5.3 cal. ka (Fig. 3A).

Some of the Elk Lake proxies show a gradual change around 8.2 cal. ka rather than an abrupt change. In particular the stable isotope and pollen records reflect gradual change. The $\delta^{18}O$ record likely shows a gradual change in this lake because $\delta^{18}O_{H_2O}$ was hydrologically buffered in this lake whose volume is likely small relative to its ground water sources, and a lake that would not store an evolving isotopic signature. The gradual nature of the pollen record is more problematic. It may be that the terrestrial plant community responded differently to climate forcing. Or it could be that the prairie pollen from the west arrived at Elk Lake before the actual prairie vegetation due to a strengthening wind field. The latter process would then blur or average the pollen record of plant activity around and near the lake.

The 8.2 cal. ka event recorded in the varved Elk Lake sediments was a sharp inflection point in a stepwise pattern of changing atmospheric circulation. Changes in circulation were predominately responding to a contraction of polar air resulting in a northward movement of the westerlies. In contrast, the Greenland Ice core record of the 8.2 cal. ka event indicates a shift to colder conditions for only a few centuries implying a brief expansion of polar air. Why should one record indicate an abrupt change in atmospheric circulation, and the other a climate pulse? The Elk Lake record provides an accurately dated initiation of a mid-Holocene dry, warm, windy period, documented in other records with less accurate dating, that was a manifestation of a change in circulation following the final collapse of the Laurentide ice sheet. By contrast, the Greenland ice-core has recorded a distinct cold pulse. We don't know why the two regions show opposite signs of climate change, but in the context of what we propose here, the answer has to be in the atmosphere, because it is not recorded in deep-sea sediments. One possibility is that due to changes in atmospheric circulation around 8.2 cal. ka, the average position of the Icelandic Low shifted from its typical location southeast of Greenland to some other place for several centuries. If, for example, it moved to the northeast toward northern Europe the counter clockwise circulation would deliver colder air to the ice sheet. The Icelandic Low might have moved due to colder water in the North Atlantic near Greenland or to other changes affecting the polar-air pressure field. The movement of the Icelandic low back to its preferred Holocene position due to, for example, the return of warmer North Atlantic waters would then complete the observed ice-core pulse. In this example, the 8.2 cal. ka event recorded in Greenland ice cores would have been a consequence of the change in atmospheric circulation over North America.

Note: Many of the data described in this paper and plotted in Fig. 3 are from papers in Bradbury, J.P. and Dean, W.E. (Eds.), Elk Lake, Minnesota: Evidence for Rapid Climate Change in the North-Central United States, Geological Society of America, Special Paper 276. Those data and more are available in digital form as the "Elk Lake Dataset" at World Data Center-A for Paleoclimatology, NOAA/NGDC 325 Broadway, Boulder, CO 80303 (phone: 303-497-6280; fax: 303-497-6513), or on the Internet at (http://www.ngdc.noaa.gov/paleo/paleodata.html).

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