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Stable isotopes and sediments from Pickerel Lake, South Dakota, USA: a 12ky record of environmental changes

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Abstract

Sedimentological parameters and stable O- and C-isotopic composition of marl and ostracode calcite selected from a 17.7-m-long core from the 8-m-deep center of Pickerel Lake, northeastern South Dakota, provide one of the longest (ca. 12ky) paleoenvironmental records from the northern Great Plains. The late Glacial to early Holocene climate in the northern Great Plains was characterized by changes from cold and wet to cold and dry, and back to cold and wet conditions. These climatic changes were controlled by fluctuations in the positions of the Laurentide ice sheet and the extent of glacial Lake Agassiz. We speculate that the cold and dry phase may correspond to the Younger Dryas event. A salinity maximum was reached between 10.3 and 9.5 ka, after which Pickerel Lake shifted from a system controlled by atmospheric changes to a system controlled by groundwater seepage that might have been initiated by the final withdrawal of Glacial Lake Agassiz. A prairie lake was established at approximately 8.7 ka, and lasted until about 2.2 ka. During this mid-Holocene prairie period, drier conditions than today prevailed, interrupted by periods of increased moisture at about 8, 4, and 2.2 ka. Prairie conditions were more likely dry and cool rather than dry and warm. The last 2.2 ka are characterized by higher climatic variability with 400-yr aridity cycles including the Medieval Warm Period and the Little Ice Age.

Although the signal of changing atmospheric circulation is overprinted by fluctuations in the positions of the ice sheet and glacial Lake Agassiz during the late Glacial-Holocene transition, a combination of strong zonal circulation and strong monsoons induced by the presence of the ice sheet and high insolation may have provided mechanisms for increased precipitation. Zonal flow introducing dry Pacific air became more important during the prairie period but seems to have been interrupted by short periods of stronger meridional circulation with intrusions of moist air from the Gulf of Mexico. More frequent switching between periods of zonal and meridional circulation seem to be responsible for increased climatic variability during the last 2.2 ka.

Introduction

Pickerel Lake (562 m above sea level., 45.51°N, 97.27°W) in Day County, northeastern South Dakota, is located on a steep climatic gradient that reflects the interaction of three dominant air masses in the north-central United States. Dry polar air and dry North Pacific air usually compete during the winter months, and moist, warm, tropical air from the Gulf of Mexico/Atlantic realm dominates during the summer months (Figure 1A; Bryson & Hare, 1974). The climate is characterized by hot summers, long and cold winters and evaporation exceeding precipitation.

The steep east-west climatic gradient across the north-central states is manifested in the chemistry of lake waters. Lakes throughout most of Minnesota are dominated by calcium and bicarbonate (Gorham et al., 1983), but the prairie lakes in western Minnesota and the Dakotas (Figure 1B) are more saline with higher concentrations of sodium, magnesium, and sulfate than Minnesota lakes. The change in water chemistry is the result of a combination of high net evaporation

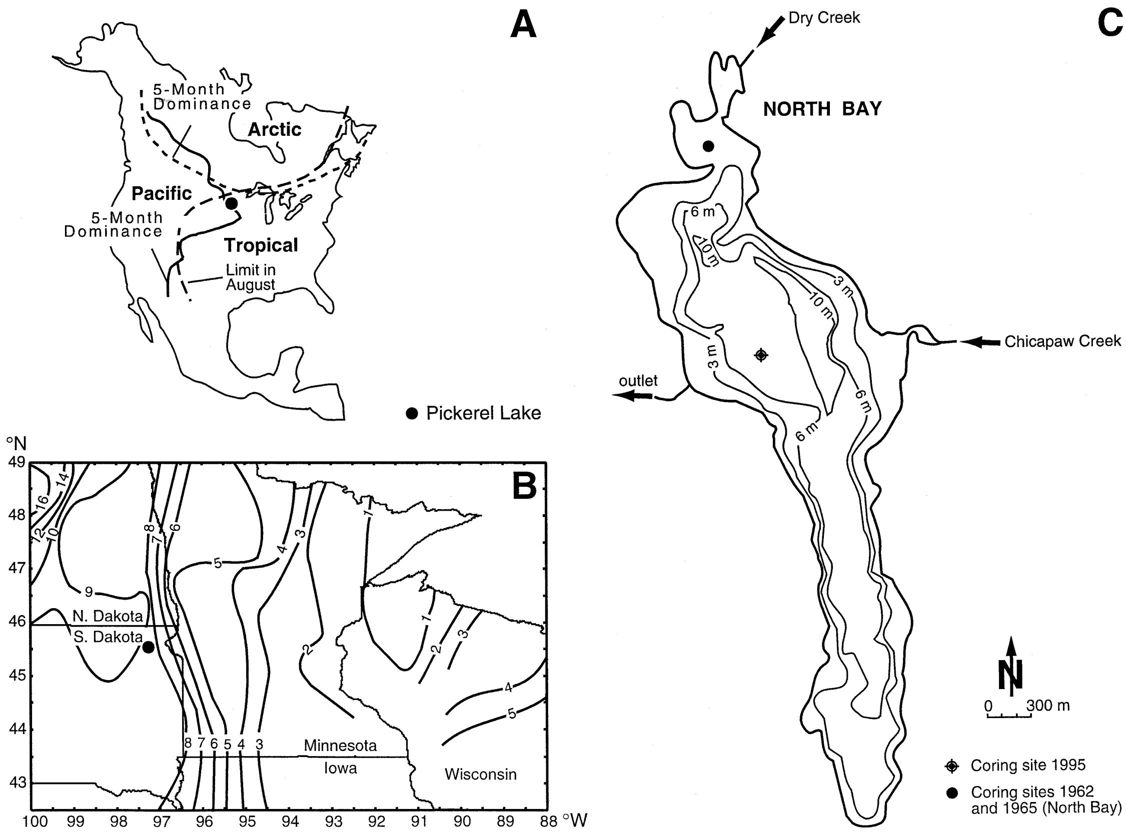

Figure 1. Maps showing (A) major airstream regions affecting the climate in North America (modified from Bryson & Hare, 1974); (B) contours of total cations in milliequivalents per liter in lakes in the north-central United States (data of Gorham et al., 1983); and (C) Pickerel Lake bathymetry and core sites discussed in this paper.

from the prairie lakes, loss of calcium and bicarbonate by precipitation of $CaCO_3$ in the lakes, and gain of sodium, magnesium, and sulfate from Cretaceous shales that underlie the prairie regions. All of the prairie lakes precipitate $CaCO_3$ during the summer, and many precipitate dolomite ($CaMg(CO_3)_2$; Dean & Gorham, 1976; Dean et al., 1993).

Pickerel Lake is one of many lakes that occupy depressions on calcareous glacial drift deposited by the Des Moines ice lobe that was channeled southward from the main Laurentide ice sheet in the lowland now occupied by the Red River. The high carbonate content of the drift is due to rock flour from Paleozoic carbonates and Cretaceous shale. Pickerel Lake is a Mg-Ca-HCO_3-SO_4-lake with a specific conductance of 450 μS. The low total ionic concentration and major ion chemistry indicate that the lake is probably a groundwater recharge site in the local flow system (Smith, 1991). Pickerel Lake is about 5 km long, up to 1500 m wide, and has a maximum water depth of 13 m (Figure 1C). The morphology is characterized by steep slopes and a flat bottom with an average depth of 6.5 m. Today, the lake has two intermittent inlets and only one outlet on the western shore (Figure 1C). Fluctuations in groundwater levels cause the lake level to vary by as much as a meter seasonally. During the drought of the 1930's, for example, the lake level was 2 m below the outlet, and North Bay was almost dry (Watts & Bright, 1968). Because the lake is well mixed by wind, the maximum difference between temperatures of surface water and subsurface water is 5 °C, and changes in surface-water temperature closely follow those in air temperature.

We selected Pickerel Lake for paleoclimatic studies because it is located on a steep climatic gradient, and the sediments should have recorded rapid and extreme climatic transitions during the Postglacial. The lake is just south of the plain of glacial Lake Agassiz, and the climate of the region was probably influenced by the presence of Lake Agassiz during deglaciation and early Holocene. In the 1960's, cores were taken from sites in about 1.6 m water depth in the sheltered habitat of North Bay of Pickerel Lake (Figure 1C) and were analyzed for pollen, seeds, and mollusks (Watts & Bright, 1968), stable isotopes of carbon and oxygen in mollusks (Stuiver, 1970), diatoms (Haworth, 1972), and ostracodes (Smith, 1987). Although the sediment sequences collected in these cores were only about 8 m long and incomplete, they clearly show a maximum of prairie development between approximately 9 ka and 4 ka. Values of δ^{18}O in mollusk shells were 2–3‰ higher for this period and were interpreted as increased mean annual temperatures (Stuiver, 1970). We conducted a study of carbon and oxygen isotopic compositions of marl and ostracodes from a 12 000-year sediment record from the profundal zone of Pickerel Lake to compare with Stuiver's (1970) mollusk isotopic study from North Bay. Our cores from the profundal zone should reflect more stable water temperature conditions in the deeper lake basin.

Methods

A 17.7-m-long sediment core was taken in 1995 from the ice surface in a water depth of 8.4 m using a modified Livingstone piston corer 5 cm in diameter (Wright, 1967). Whole-core magnetic susceptibility measurements were made with a Bartington susceptibility bridge prior to subsampling at 2 cm intervals. The core was then cut in half, and one half was logged for sedimentological features; colors were defined by using the Geological Society of America Rock Color Chart. Percentages of organic matter and total carbonate, calculated as $CaCO_3$, were determined by loss-on-ignition (Dean, 1974) in 1 cm^3 samples collected every 10 cm. The non-carbonate, inorganic fraction was calculated as the difference between 100% and the sum of % organic matter and % $CaCO_3$. This component is assumed to represent the detrital clastic fraction.

Percentages of total carbon (TC) and inorganic (carbonate) carbon (IC) were determined using a coulometer (Engleman et al., 1985) on dried and powdered samples of bulk sediment collected every 20 cm. Carbonate in the untreated sample was reacted with perchloric acid to liberate CO_2, which was then titrated in a coulometer cell to measure carbonate carbon. Total carbon was measured by liberating CO_2 by combustion of an untreated sample at 960 °C and titrating the CO_2 in a coulometer cell. Values of organic carbon (C_{org}) were determined by difference between TC and IC. The amount of $CaCO_3$ was calculated by dividing percent carbonate carbon by 0.12, the fraction of carbon in $CaCO_3$. This conversion probably overestimates the total carbonate content slightly because most samples contain minor amounts of dolomite, which contains 13 weight percent in organic carbon. For a sample that contains 8% IC, calculating total carbonate using the 0.12 conversion factor gives a total carbonate content of 66.7% (8% IC/0.12). However, partitioning the IC between 10% dolomite and 90% calcite gives a total carbonate content of 66.15% (dolomite = 0.8% IC/0.13 = 6.15%; calcite = 7.2% IC/0.12 = 60%) that is only 0.55% lower. Therefore, the low contents of dolomite do not make a significant difference for total carbonate content calculated as $CaCO_3$.

Semi-quantitative estimates of mineral contents of powdered bulk samples were determined by standard X-ray diffraction techniques (e.g. Moore & Reynolds, 1989). An aliquot of the powdered sample was packed into an aluminum holder and scanned from 15 ° to 50 ° 2Θ at 2 ° 2Θ min^{-1} using Ni-filtered, Cu-Kα radiation at 45 kv, 30 ma.

Ostracode valves were separated from 2-cm-thick slices, representing approximately 10 to 20 years of deposition, taken every 20 cm corresponding to LOI and carbon samples. Samples of 10 to 20 g wet sediment were placed in plastic wide mouth bottles and 250 ml of hot (ca 90 °C) deionized water were added to each sample followed by one teaspoon of baking soda. Once the sample had cooled to room temperature, half a teaspoon of commercial calgon was added and the contents were gently stirred. To allow full dispersal of the sediment, the sample remained several hours at room temperature. The sample was frozen solid and was then allowed to thaw at room temperature where it remained again for several hours. The disaggregated sediment was slowly sieved through a 250 μm sieve. Ostracode valves retained on the sieve were rinsed with deionized water and best preserved valves were selected for analyses of stable oxygen and carbon isotopes. Not all samples yielded ostracode valves, and additional samples were subsequently collected, processed, and analyzed to better define zones of rapid changes in isotopic composition. In this paper, we present results of isotopic analyses of valves of *Candona ohioensis*, the dominant species in Pickerel Lake sediments that lives at the sediment-water interface. Valves from both adult and juvenile ostracodes were analyzed because in some levels there were no adult ostracodes present. We found no significant difference in the ranges of isotopic values for both adult and juvenile specimens from the same sample. The ostracode valves and splits of bulk sediment from the carbon analyses were roasted in vacuum for 1 h at 380 ± 10 °C to remove volatile organic carbon prior to isotopic analyses. Isotopic analyses were obtained on 251 *C. ohioensis* samples from 94 stratigraphic levels. Each sample consisted of a minimum of one valve (adult) to as many as five valves (juveniles), with up to 8 samples per stratigraphic level. CO_2 was liberated from each sample using a Finnigan Kiel Automated Carbonate Extraction Device and analyzed in line with a Finnigan MAT 251 triple collector isotope-ratio mass spectrometer. In the Kiel device, four drops of 100% phosphoric acid were dripped on each sample in individual reaction vessels and allowed to react at 75 ± 1 °C to completion in 10 min. Evolved gases were cryogenically purified to remove water and noncondensible gases in the Kiel device. Purified CO_2 from the samples was introduced into the mass spectrometer through a capillary and measured against a reference standard of known isotopic composition. Samples are reported in the usual per mil (‰) δ-notation relative to the University of Chicago standard PDB (Pee Dee Belemnite) for carbon and oxygen:

$$\delta‰ = [(R_{sample}/R_{PDB})\text{-}1] \times 10^3,$$

where R is the ratio (^{13}C:^{12}C) or (^{18}O:^{16}O).

Results

Chronology

Pickerel Lake sediments in our core contained little material for radiocarbon dating, and pollen analyses are in preparation. Therefore, the chronology used in this study comes from two accelerator-mass spectrometer (AMS) ^{14}C ages on wood fragments (Figure 2). We used the uncalibrated ages that do not account for variations in ^{14}C production in the atmosphere to facilitate comparisons with uncalibrated radiocarbon ages published for other sites. Assuming a uniform sedimentation rate between ages, the average sedimentation rate between the top of the core (assumed to be 0 ^{14}C yr B.P.) and 859 cm (3290 ± 60 ^{14}C yr B.P.) is 0.26 cm yr^{-1}, and that between 859 cm and 1636 cm ($11\,880 \pm 60$ ^{14}C yr B.P.) is 0.09 cm yr^{-1}. Using these sedimentation rates, we estimated ages of sediment in the core at 1000-yr intervals (Figure 2).

Sediment stratigraphy, magnetic susceptibility and composition of sediments

The sediments recovered from Pickerel Lake are divided into six lithologic units (Figure 2) that are numbered from top to bottom. Basal Unit 6, below 17 m in the core, was deposited before approximately 12.2 ka. It consists of gray gravel and sand with large amounts of quartz, feldspars and dolomite, and is characterized by magnetic susceptibility values up to 400 SI, up to 85% detrital clastic material, less than 1% C_{org}, and up to 20% $CaCO_3$.

Unit 5, between 17 and 16.5 m in the core, was deposited between approximately 12.2 ka and 12 ka. The base of this unit is a 7-cm-thick dark gray layer that is a blend of terrestrial organic detritus, clay, silt, and sand. The basal layer is overlain by an olive-gray to medium-dark-gray, homogenous, clayey silt to silty clay with a few-centimeter-thick laminated sections. Values for magnetic susceptibility are about 25 SI, concentrations of C_{org} and $CaCO_3$ are less than 2% and 20%, respectively, and that of detrital clastic material averages about 75%.

Unit 4 extends from 16.5 m to 14.9 m and was deposited between approximately 12 ka and 10.3 ka. The base of this unit consists of a clayey silt characterized by alternations of 2- to 3-cm-thick olive-gray to light-olive-gray silty layers with 2- to 3-cm-thick bioturbated, yellowish-gray, carbonate-rich, clayey layers. The sediment contains very minor amounts of quartz and dolomite and is marked at the base by a sharp $CaCO_3$ increase from 20% to 80%. Concomitantly, the detrital clastic fraction decreases from 75% to 20%. The unit has relatively low magnetic susceptibility values that decrease throughout the unit from 25 to a minimum of 5 SI. The content in C_{org} is variable and remains mostly below 3%.

Unit 3, extending from 14.9 m to 13.4 m in the core, is a sandy, clayey, calcareous, olive-gray silt deposited between 10.3 ka and 8.7 ka that is characterized by increasing magnetic susceptibility values from 10 SI to about 20 SI, and increasing values of the detrital-clastic fraction from 20% to 60%. The $CaCO_3$ content decreases from 60% to about 40%, and that of C_{org} remains below 3%.

Unit 2, extending between 13.4 m and 5.8 m in the core, corresponding to an age of deposition of about 8.7 ka to 2.2 ka, is a massive, olive-gray to olive-gray-brown, calcareous, clayey silt with varying amounts of fine sand and visible ostracode valves. The main minerals identified by X-ray diffraction are calcite and

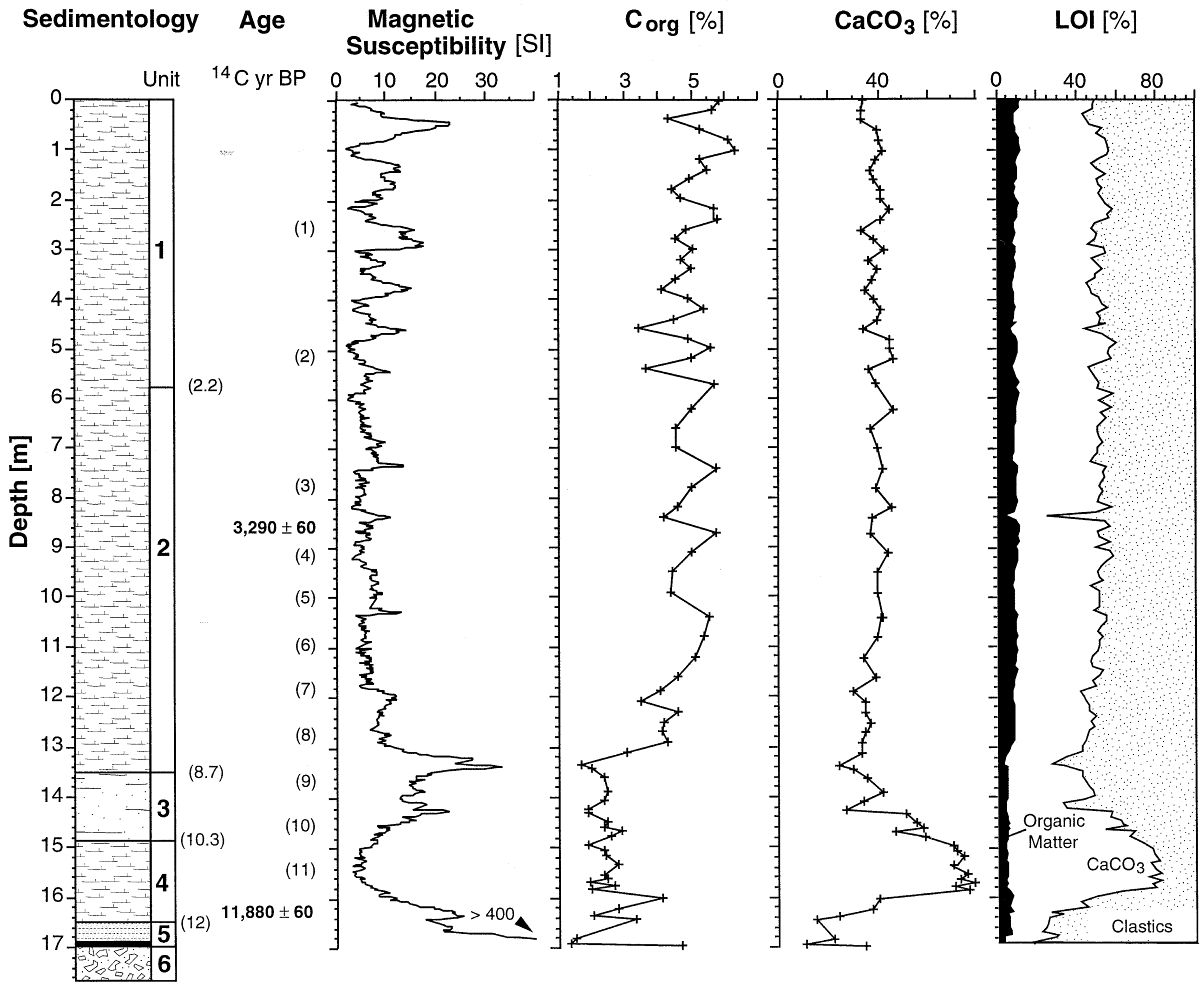

Figure 2. Sedimentology, ^{14}C chronology (numbers in parentheses are estimated interpolated ages in 1000 yr), magnetic susceptibility, percentages of organic carbon (C_{org}), calcium carbonate ($CaCO_3$), and organic matter, $CaCO_3$, and clastic material by loss-on-ignition (LOI). Units 1, 2 and 4: calcareous clayey silt; Unit 3: calcareous sandy silt; Unit 5: clayey silt with laminated sections; Unit 6: gravel and sand.

quartz with lesser amounts of dolomite and feldspar. The sediments in the lower 3 m of this unit are very compact. The bottom of the unit is marked by magnetic susceptibility values of about 35 SI, and a detrital clastic component of up to 70%. Above 13 m, values of magnetic susceptibility and detrital clastic material oscillate around 10 SI and 50%, respectively. The C_{org} content varies between 4% and 6%, and the content of $CaCO_3$ is about 40%.

Unit 1, representing the uppermost 5.8 m of sediment deposited during the past 2.2 ka, resembles Unit 2, except that this calcareous, clayey silt has distinctive 1-m-thick cycles in magnetic susceptibility. Peak values of magnetic susceptibility in each cycle generally increase upwards, and reach up to 25 SI. Values of C_{org} also vary cyclically, and generally increase from approximately 4% to 6%. Values of $CaCO_3$ and the detrital clastic fraction fluctuate between 35% and 45%, and 45% and 55% respectively.

Oxygen and carbon isotopes

Sedimentology and ostracode species assemblages as well as isotopic compositions of ostracodes and marl (see Figures 3 and 4) are used to define isotope zones, which correspond to the lithologic units. There were no ostracodes present in basal lithologic Unit 6, therefore there are only 5 isotope zones as presented in Figure 4. The values for $\delta^{18}O_{ostracode}$ are higher and fluctuate more than those for $\delta^{18}O_{marl}$, whereas the

$\delta^{13}C_{ostracode}$ values are lower than those for $\delta^{13}C_{marl}$. All four records, however, display similar major trends.

Unit 5 (17–16.5 m; approximately 12.2–12 ka) is characterized by $\delta^{18}O$ values between -7 and -5‰ for both ostracodes and marl. Values of $\delta^{13}C_{ostracode}$ range between -5‰ and -2‰ and those of $\delta^{13}C_{marl}$ range between -2‰ and 0‰. The most prominent ostracodes are *Candona ohioensis*, an indicator of freshwater lakes with low ionic concentrations (Smith, 1991), and *Cytherissa lacustris*, which today is associated with dilute, cold water in deep lakes ($>$3 m) in the boreal forests of Canada (Delorme, 1970).

In Unit 4 (16.5–14.9 m; approximately 12–10.3 ka), values of $\delta^{13}C$ and $\delta^{18}O$ in both marl and ostracodes show a covarying trend from higher to lower values in the middle of the unit, and then back to higher isotopic values at the top of the unit (see Figures 3 and 4). The difference between values in ostracodes and those in marl is about 4‰ for $\delta^{18}O$ and 5‰ for $\delta^{13}C$, the maximum differences in the entire sequence, due mainly to the low values of $\delta^{18}O$ and $\delta^{13}C$ in ostracodes. The dominant ostracodes in this unit are *C. ohioensis*, *C. lacustris* and *Candona candida*; the latter is an indicator of low water temperatures and lakes with cold groundwater input, typical for the late Glacial (Absolon, 1973), and cold, dilute lakes today of the Canadian prairie-forest transition (Forester et al., 1987).

At the base of Unit 3 (14.9–13.4 m; approximately 10.3–8.7 ka), values of $\delta^{13}C_{ostracode}$ increase almost 3‰ from about -6‰ to -3‰. The values for $\delta^{18}O_{ostracode}$ in Unit 3 first increase from -3‰ to -2‰, and then decrease abruptly at 14.2 m (approximately 9.5 ka) from -2‰ to -4.5‰. The marl data show a distinct covariant trend to higher values for both $\delta^{13}C_{marl}$ and $\delta^{18}O_{marl}$ (see Figures 3 and 4). Values of $\delta^{18}O_{marl}$ and $\delta^{13}C_{marl}$ increase from -7‰ to -6‰ and from -1.5‰ to -0.5‰, respectively. The difference between $\delta^{18}O_{ostracode}$ and $\delta^{18}O_{marl}$ is 3.5‰, and that between $\delta^{13}C_{ostracode}$ and $\delta^{13}C_{marl}$ is about 2‰. Ostracodes are mainly represented by *C. ohioensis* and also *Candona rawsoni*, which commonly occurs in permanent and ephemeral lakes and ponds with variable salinity in the prairie and in the prairie-forest transition in the United States and Canada (Forester et al., 1987). *C. rawsoni* indicates that moisture loss from the lake by evaporation is significant (Forester et al., 1994). Another characteristic ostracode species for Unit 3 is *Limnocythere herricki* that lives today in lakes located in the prairie and in the prairie-forest transition of Canada, where long and cold winters, warm to cool summers, and high drought frequency are common (Forester et al., 1987). Coincident with the decrease in $\delta^{18}O_{ostracode}$ at 14.2 m, *L. herricki* is replaced by *Limnocythere itasca*, a low salinity species (Smith, 1993).

The base of Unit 2 (13.4–5.8 m; 8.7–2.2 ka) is characterized by a 2‰ increase in $\delta^{13}C_{marl}$, and a peak in $\delta^{18}O_{marl}$ at about 13.4 m from approximately -6‰ to -2‰. Values of $\delta^{18}O_{ostracode}$ show a large range between -2‰ and -7‰. Values of $\delta^{13}C_{ostracode}$ increase from about -2‰ at the base of the unit to an average of 0‰ above 13 m. Both $\delta^{18}O_{ostracode}$ and $\delta^{13}C_{ostracode}$ reach minima around 13 m, 9 m, and 6 m. Values of $\delta^{18}O_{marl}$ vary between about -5.5‰ and -7.5‰, and values of $\delta^{13}C_{marl}$ vary between 2‰ and 0.5‰. An overall decrease of about 1‰ in values of $\delta^{13}C$ and $\delta^{18}O$ in both marl and ostracodes is visible from base to top of this unit. The difference between $\delta^{18}O_{ostracode}$ and $\delta^{18}O_{marl}$ in Unit 2 is about 4‰, and that between $\delta^{13}C_{ostracode}$ and $\delta^{13}C_{marl}$ is about 2‰. The ostracode species assemblage is dominated by *C. ohioensis* and *L. itasca*. *Darwinula stevensoni*, a species that has been linked to ground-water input (Schwalb et al., 1995), occurs sporadically, and is most abundant coincident with the low ostracode isotope values for *C. ohioensis* at 13 m, 9 m, and 6 m.

Unit 1 ($<$5.8 m; $<$2.2 ka) displays the highest variability in the core for values of $\delta^{18}O_{ostracode}$ and $\delta^{13}C_{ostracode}$ (see Figures 3 and 4). Values range from -2‰ to -6‰ for $\delta^{18}O_{ostracode}$ and -4‰ to 1‰ for $\delta^{13}C_{ostracode}$. The marl data are more restricted (see Figures 3 and 4) and range between -6‰ and -7.5‰ for $\delta^{18}O_{marl}$ and between -0.5‰ and 1‰ for $\delta^{13}C_{marl}$. Values of $\delta^{18}O_{ostracode}$ are 0.5–1‰ higher, and those of $\delta^{18}O_{marl}$ are about 0.5–1‰ lower, than those in Unit 2. Values of $\delta^{13}C$ are about 1‰ lower than those in Unit 2 for both ostracodes and marl. Values of $\delta^{18}O_{ostracode}$ are about 2.5‰ higher than those of $\delta^{18}O_{marl}$, and values of $\delta^{13}C_{ostracode}$ are 2‰ higher than those of $\delta^{13}C_{marl}$. Ostracode assemblages in Unit 1 are characterized by *C. ohioensis*, *D. stevensoni*, *Candona caudata*, the latter is a species that may be indicative for deep rapid circulation (R.M. Forester, pers. comm., 1997), and high abundances of nektic species.

Discussion

Advancement of the Laurentide ice sheet into South Dakota and Iowa persisted until 12.3 ka (Teller et al., 1980). Consequently, the sedimentary sequence from

Pickerel Lake with an estimated basal age of 12.2 ka would be one of the oldest paleoclimate and paleoenvironmental records recovered from lakes in the northern Great Plains. The results from sedimentological and isotope analyses, combined with the main ostracode species abundances, provide the proxies needed to understand the environmental changes recorded in the sediments. The comparison between marl and ostracode isotope data allows a comparison of changes in lake water temperature and primary productivity in the epilimnion (marl), versus changes at the sediment-water interface (ostracodes), averaged over 10 to 20 years. Values of $\delta^{18}O_{ostracode}$ are higher than those of $\delta^{18}O_{marl}$ (Figures 3 and 4) because higher surface-water temperatures result in greater oxygen-isotope fractionation during marl precipitation (by about 0.24‰ $°C^{-1}$; Stuiver, 1970). Assuming that ostracode valves calcify in approximate equilibrium with the isotopic composition of the lake water (Lister, 1988a), ostracode valves should reflect the isotopic signature of the lake water that may be controlled by changes in catchment hydrology (Lister, 1988b; Schwalb et al., 1994, 1995), air-mass history (Schwalb et al., 1995) and mean annual temperature of precipitation (Von Grafenstein et al., 1992, 1996). However, laboratory cultures of *C. rawsoni* by Xia et al. (1997a) show that values of $\delta^{18}O_{ostracode}$ can be 0.8–1.0‰ higher than those of inorganic carbonate. Although oxygen isotopic composition of precipitation is often more related to air-mass characteristics than temperature (e.g. Fritz et al., 1987; Lawrence & White, 1991), temperature today apparently is the controlling factor in the Southern High Plains (Nativ & Riggio, 1990) and Iowa (Simpkins, 1995), where precipitation depleted in ^{18}O dominates during the winter months and precipitation enriched in ^{18}O dominates during spring and summer months.

The trends in $\delta^{13}C_{ostracode}$ and $\delta^{13}C_{marl}$ are remarkably similar (Figure 3), but values of $\delta^{13}C_{marl}$ are consistently higher. Fractionation of carbon isotopes is not affected by changes in lake water temperature and, therefore, the higher values of $\delta^{13}C_{marl}$ must be caused by another factor. Plankton preferentially take up the lighter ^{12}C isotope leading to an enrichment of ^{13}C in the dissolved inorganic carbon (DIC) pool of the epilimnion and, consequently, in the marl (Stuiver, 1970; McKenzie, 1985). Organic matter depleted in ^{13}C is deposited and releases ^{13}C-depleted CO_2 to the hypolimnion and porewaters during decay (McKenzie, 1985) where it is taken up by benthic ostracodes. The carbon isotopic composition of carbonate seems mainly to reflect changes in vegetation in the catchment, namely the change between forests dominated by C3 plants using the Calvin-Benson photosynthetic pathway and prairie characterized by a mixture of C4 grasses, using the Hatch-Slack photosynthetic pathway, and C3 grasses. The C3 and C4 plants have very different $\delta^{13}C$ values averaging about -27 and -12‰, respectively (Cerling & Quade, 1993). Thus, C4 plants can provide abundant ^{13}C to soils, which leads to higher $\delta^{13}C$-values in lake sediments (Kelts & Schwalb, 1994). The history of Pickerel Lake can be divided into three sections; the late Glacial to early Holocene transition, the mid-Holocene, and the late Holocene. The overall parallel trend for both ostracode and marl $\delta^{13}C$ data may indicate that $\delta^{13}C$ is mainly controlled by changes in vegetation rather than productivity, and that Pickerel Lake seems to have always remained a well mixed oligotrophic lake.

Late Glacial to early Holocene transition (≥12.2 ka to 8.7 ka)

This section includes lithologic Units 6 through 3. The sand and gravels at the base of Unit 6 represent the glacial drift deposited prior to approximately 12.2 ka. The succeeding layer of terrestrial organic detritus mixed with sand, silt, and clay at the base of Unit 5 probably presents a forest floor formed on a buried ice block, either in place or slightly transported into a shallow depression filled with relatively warm water that melted the ice on the banks and on the bottom (Wright, 1993).

Late Glacial $\delta^{18}O$ values are not significantly lower than those for the Holocene. The similar $\delta^{18}O$ values for both ostracodes and marl in Unit 5, as well as the dominance of *C. ohioensis* and *C. lacustris*, indicate a boreal style climate with resulting cool, dilute, and seasonally stable water column. Relatively high values of $\delta^{18}O$ in late Glacial and early Holocene pedogenic carbonates and lake sediments also have been observed by Amundson et al. (1996) and Edwards et al. (1996), respectively. During the late Glacial, the summer insolation gradient from north to south was greater than today, resulting in a steep pressure gradient that focused the jet and associated cold fronts, and could have provided the mechanisms for increased precipitation in the central US (Wright, 1992). A similar pattern can be found today, when circulation is more zonal during winter. The main storm track is oriented from southwest to northeast, and cyclones pull warm, moist air from the Gulf of Mexico that causes heavy snowfalls

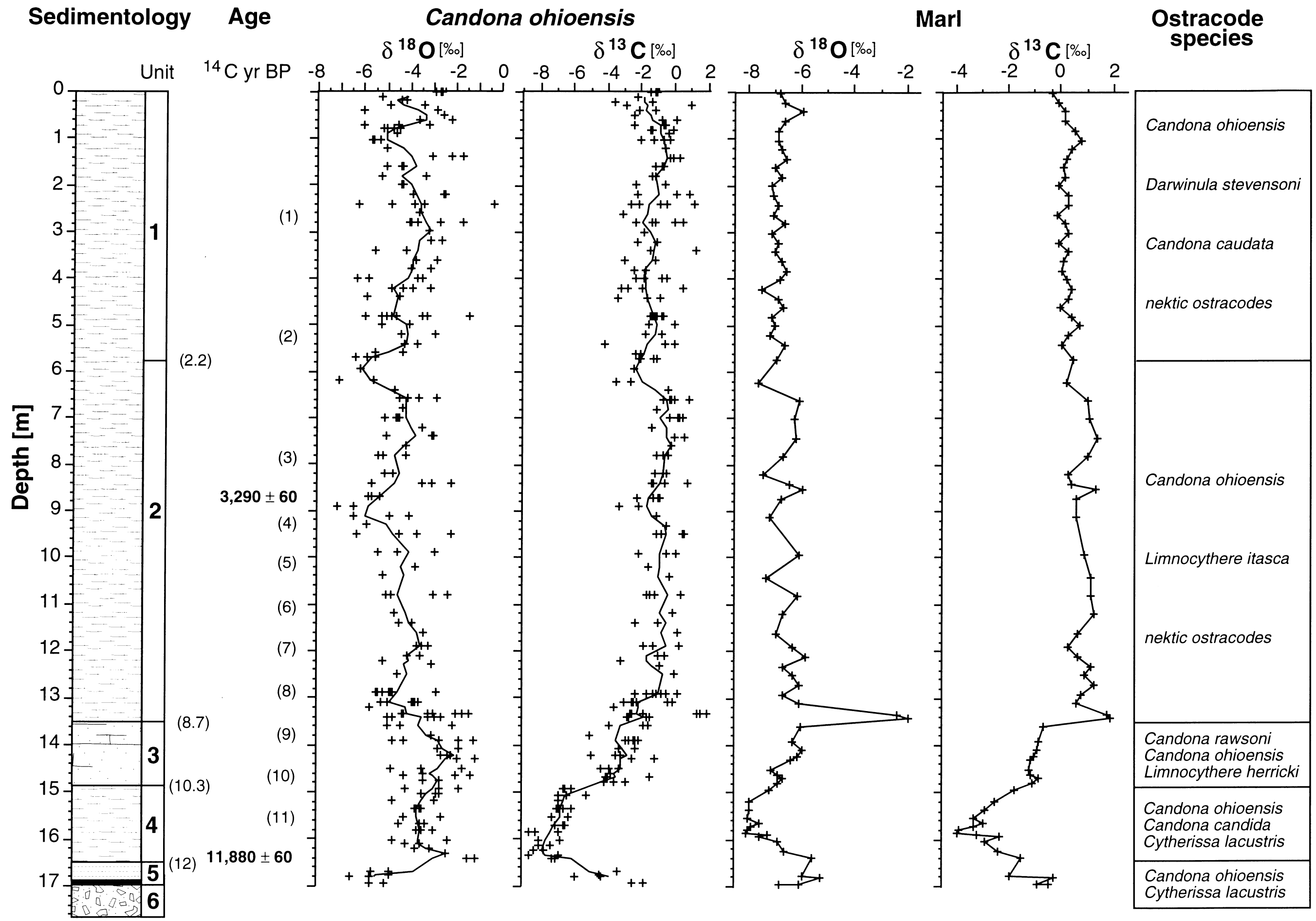

Figure 3. Sedimentology, ^{14}C chronology (numbers in parentheses are estimated interpolated ages in 1000 yr), and stable oxygen and carbon isotopic composition in valves of the ostracode *Candona ohioensis* and in marl; lines running through the ostracode data (plus symbols) represent 3-point running averages. Major ostracode species characteristic for each sedimentary unit are listed on the right. Units 1, 2 and 4: calcareous clayey silt; Unit 3: calcareous sandy silt; Unit 5: clayey silt with laminated sections; Unit 6: gravel and sand.

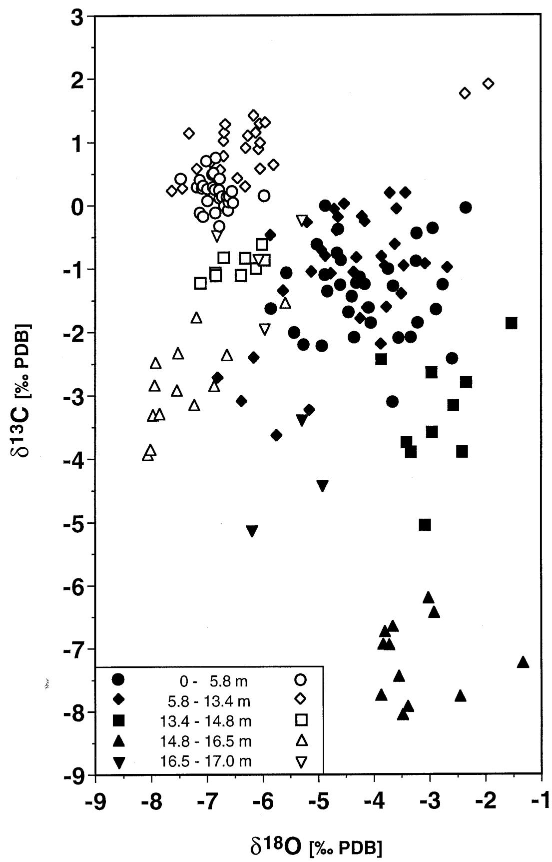

Figure 4. Cross plot of δ^{18}O versus δ^{13}C for ostracode calcite (solid black symbols) and marl (open symbols) divided into five groups that correspond to sedimentological Units 1 through 5. Data points represent the average values for each level.

in the Great Lakes area (Rodinov, 1994). In addition, the maximum summer insolation between 12 ka and 9 ka (Kutzbach, 1987) led to stronger summer monsoons that provided higher amounts of ^{18}O-enriched moisture from the Gulf of Mexico than today.

Additionally, the first highstand of nearby glacial Lake Agassiz (Lockhart phase, 11.6 ka to 10.9 ka, Teller, 1985 & 1987), whose meltwater entered the Gulf of Mexico through the Mississippi drainage until about 11.3 ka (Leventer et al., 1982; Fairbanks, 1990; Marchitto & Wei, 1995), could have influenced the regional climate, leading to cooler summers, warmer winters, and higher annual precipitation. As a result, low δ^{18}O–values in precipitation should be expected as seen on the leeside of the Great Lakes area today (Gat et al., 1994). Late Glacial $\delta^{18}O_{ostracode}$ values from Pickerel Lake are in general relatively high instead, suggesting that the presence of Lake Agassiz might have affected the local temperature but did not provide significant amounts of moisture.

Decreasing values of magnetic susceptibility and amounts of detrital clastic material as well as the trend to lower oxygen and carbon isotope values for the marl after 12 ka indicate that the detrital input decreased probably due to a denser or more vegetated catchment. The relatively low δ^{13}C values for both ostracodes and marl in Unit 4 reflect the change in vegetation from tundra grasses, with high percentages of C4 plants providing abundant ^{13}C to the soil and consequently to the lake, to spruce forest with predominantly C3 plants. Pollen from sediments collected in a core from shallow North Bay of Pickerel Lake (Figure 1) indicate that a boreal forest dominated by spruce surrounded the lake prior to 10 670 ^{14}C yr B.P. (Watts & Bright, 1968). The presence of a spruce forest also is indicated by diatom taxa that generally prefer acid environments (Haworth, 1972).

In Unit 4, between about 12 ka and 10.3 ka, the average difference in δ^{13}C between ostracodes and marl is about 3‰ (Figure 4). This suggests that photosynthetic activity in the surface water had begun to 'pump' ^{13}C-depleted organic matter from the epilimnion to the hypolimnion and sediments. The rapid increase in the carbonate content, mostly calcite, suggests either that the influx of detrital clastic material was suddenly reduced and/or that carbonate precipitation was so much greater that it considerably diluted both the detrital clastic as well as the organic matter fraction. A pulse of carbonate in late Glacial to early Holocene lake sediments is common in lakes in calcareous drift (e.g. Dean & Megard, 1993) due to rapid initial leaching of carbonate in fresh glacial drift followed by a gradual decline in supply of carbonate to the lake. The asymmetrical shape of the% $CaCO_3$curve for Unit 4 (Figure 2) suggests that such a pulse occurred in Pickerel Lake. This also agrees with diatom taxa from a core from Pickerel Lake North Bay indicating alkaline waters (Haworth, 1972). The fact that C_{org} remains relatively low suggests that photosynthetic activity was not the solely trigger for carbonate precipitation although the difference between $\delta^{13}C_{marl}$ and $\delta^{13}C_{ostracode}$ reaches a maximum in this unit (ca 3‰). The difference between $\delta^{18}O_{ostracode}$ and $\delta^{18}O_{marl}$ also is at a maximum (ca 5‰) in this unit suggesting that there was a large seasonal temperature difference. Assuming a decrease of the $^{18}O/^{16}O$ ratio of 0.24‰ for an increase in water temperature of 1 °C (Stuiver, 1970), bottom water temperatures of 4 °C and excluding vital effects, the surface water may have reached temperatures of up to 24 °C in summer, compared to modern maximum July surface-water temperatures of 26 °C to 27 °C (Weber, 1960). Warmer surface-water temperatures in summer also would have reduced the solubility of CO_2, perhaps enhanced by prolongation of the ice-free period, and subsequently increased carbonate precipitation. At the same time, the large bicarbonate reservoir could have supplied CO_2 at a rapid rate thus leading to a maximum of ^{13}C fractionation (Stuiver, 1975). The presence of *C. candida* and *C. rawsoni* attest to cold and dry conditions and suggest that the low $\delta^{18}O$ values might reflect a decrease in mean annual air temperature and a decrease in moisture from the Gulf of Mexico. A temperature decrease is likely because the low $\delta^{18}O$ values at 11–10 ka fall within the time period of the Younger Dryas interval. At that time, glacial Lake Agassiz had retreated into Canada (Moorhead phase) and drained east into Lake Superior (Lowell & Teller, 1994; Colman et al., 1994; Lewis et al., 1994). As a consequence of the withdrawal of Lake Agassiz, climate in the northern Great Plains might have been characterized by long, cold winters and short, relatively warm summers during the Younger Dryas as indicated by the presence of *C. lacustris*.

Between 10.3 ka and 8.7 ka, the trend to higher $\delta^{13}C$ values for both ostracodes and marl suggests a change in the DIC pool possibly related to the next vegetation change from mainly C3-dominated boreal spruce forests to a mix of C4 and C3 plants of the mixed deciduous forests with oak, elm and pine, and of the open savanna in the early Holocene (Watts & Bright, 1968). Increases in values of both $\delta^{18}O$ and $\delta^{13}C$ in marl and ostracodes between 10.3 and 9.5 ka suggest that evaporation and/or residence time increased as shown by Talbot (1990) for a series of lakes. In addition, longer summers could have led to an enrichment of $\delta^{18}O$ in the eplimnetic waters through evaporation, and an enrichment of epilimnetic waters in ^{13}C through the removal of ^{12}C-enriched organic matter by photosynthetic activity as described by Drummond et al. (1995) for a lake in Michigan. A higher evaporation rate might also have lowered the lake level leading to erosion of detrital clastic material from the littoral zone as suggested by the marked increase in magnetic susceptibility. A 50-cm thick sand layer was described just above a horizon dated at 9400 ^{14}C yr B.P. in a core from North Bay of Pickerel Lake (Figure 1C) used for diatom analyses (Haworth, 1972). High abundances for *C. rawsoni* suggest increased salinity indicating that evaporation rates were high, probably the highest in the entire sequence. The return to lower $\delta^{18}O_{ostracode}$ values at approximately 9.5 ka is accompanied by the replacement of *L. herricki* by *L. itasca*, suggesting a decrease of drought frequency and salinity, which is contrary to what is expected for the beginning of the dry mid-Holocene. Colder conditions are confirmed by the decreased difference in $\delta^{18}O$ between marl and ostracodes suggesting that summer temperatures were lower. This may have been the result of the proximity of Lake Agassiz that had returned to northern Minnesota and North Dakota to reach another high stand, the Emerson phase. Within a few hundred years between 10 ka and 9.5 ka, a readvance of the Laurentide ice sheet closed the eastern outlets of Lake Agassiz, the lake doubled in size (Teller, 1985, 1987), and meltwater was rediverted down the Mississippi River to the Gulf of Mexico (meltwater pulse 1b of Fairbanks, 1990).

High salinity phases are inferred from a series of lakes in the Dakotas. The record from Medicine Lake, about 60 km southeast of Pickerel Lake, shows a change from a dilute to a highly saline lake at approximately 9.5 ka (Kennedy, 1994). An increase in the Sr/Ca ratio in bulk carbonate in Coldwater Lake, North Dakota, between 10.8 and 8.9 ka suggests that the salinity of the lake increased (Xia et al., 1997b). The maximum salinity of Devils Lake, North Dakota, also occurred about 8 ka (Haskell et al., 1996). The salinity of Moon Lake, North Dakota, inferred from diatom assemblages increased from <2 g l^{-1} to >20 g l^{-1} between 10 ka and 7.3 ka as the hydrology of the lake went from an open lake to a closed lake (Laird et al., 1996a). This suggests that the lakes switched from systems controlled by atmospheric changes to systems controlled by local groundwater inputs during the early Holocene. This is contemporary with the change of drainage of Lake Agassiz through the McKenzie River System to the Arctic Ocean rather than to the Gulf of Mexico (Smith & Fisher, 1993) that began at about 9.9 ka and finally ended by 9.1 ka (Lowell & Teller, 1994). The termination of the drainage to the Gulf of Mexico is also indicated by the final episode of deep erosion by glacial River Warren, the outlet for Glacial Lake Agassiz, at about 9.5 ka (Eyster-Smith et al., 1991). The switch of the drainage from the Gulf of Mexico to the Arctic Ocean certainly would have changed the groundwater flow patterns of the lakes in the Great Plains from regional drainage to local drainage. Some lakes might have been cut off from the groundwater flow lines, whereas groundwater recharge areas became established close to other lakes as apparently happened in Pickerel Lake. The disappearance of Lake Agassiz and the retreat of the Laurentide ice sheet can also be linked to climate warming as indicated by deciduous forest (Watts & Bright, 1968).

Mid-Holocene (8.7 to 2.2 ka)

The peaks in values of both δ^{18}O and δ^{13}C in marl at approximately 8.7 ka (Figure 3) are probably due to increased influx of detrital carbonate from calcareous glacial drift, resulting from a thinned vegetation cover and/or from a drop in lake level as a consequence of high evaporation rates. Increased erosion is further supported by high values for magnetic susceptibility and detrital clastic material (Figure 2). The sharp increase in C_{org} concentration from 2% to 4% between 13.2 m and 12.8 m (Figure 2) marks the transition to a more productive mid-Holocene prairie lake. This change is recorded in the sediments in North Bay by marked increases at 5.8 m (ca 9 ka) in abundances of planktonic and limnophilous diatoms, and diatoms indicative of more eutrophic conditions (Haworth, 1972).

Pollen from North Bay indicate that from 8 ka to 4 ka the lake was surrounded by bluestem prairie dominated by grasses and a few openings with oak and elm, and possibly some pine (Watts & Bright, 1968). The trend in $\delta^{13}C_{ostracode}$ (Figure 3) resembles that of Stuiver's (1970) profile of δ^{13}C in mollusks from sediments in North Bay in which values of δ^{13}C increase from −7‰ at 10 ka to −1‰ at 7 ka, and then remain constant (Stuiver, 1970). Some of this increase may have been produced by the change from C3-dominated forest vegetation to mixed C3 and C4 prairie vegetation that would have provided more ^{13}C to soils and hence to the dissolved inorganic carbon pool in the lake. Values of $\delta^{13}C_{marl}$ also increase at the transition to prairie, and there is about a 2‰ difference between $\delta^{13}C_{ostracode}$ and $\delta^{13}C_{marl}$ that might be linked to increased burial of ^{13}C-depleted organic carbon causing the entire carbon reservoir of the epilimnion to become more enriched in ^{13}C (e.g. McKenzie, 1985).

The change to a dominance of freshwater ostracodes *C. ohioensis* and *L. itasca* in the mid-Holocene suggests that the lake became less saline between 8.7 and 2.2 ka. Lower values of $\delta^{18}O_{ostracode}$ and $\delta^{13}C_{ostracode}$ around 8 ka, 4 ka, and 2.2 ka point to periods of even shorter residence time possibly caused by enhanced groundwater input as indicated by the groundwater seepage-related ostracode *D. stevensoni*. We suggest that increased groundwater input resulted from increased precipitation when the predominant dry zonal flow was interrupted by short periods of stronger meridional circulation leading to the penetration of moist air masses from the Gulf of Mexico.

Fluctuations in abundances of seeds of aquatic vegetation in the sediments from North Bay led Watts & Bright (1968) to conclude that lake level there varied considerably in response to recurring periods of drought. High variability in diatom-inferred salinity in mid-Holocene sediments from Devils Lake, North Dakota (Haskell et al., 1996) can be explained by varying groundwater input that is characterized by large seasonal and annual variations of seepage directions (LaBaugh et al., 1987). Smith (1991) interpreted a peak in abundance of *Limnocythere ceriotuberosa*, an indicator for elevated total ionic concentrations, at about 5 ka in a core from North Bay as the result of desiccation leading to a hiatus in North Bay. *L. ceriotuberosa* has not been found in the core from the deep basin, attesting to increased evaporation in the North Bay that has not been observed with a comparable amplitude for the deep basin. There is a slight increase in the amount of dolomite in mid-Holocene sediments from the deep basin, perhaps suggesting higher evaporation rates. The maximum increase of only about 0.5‰ for $\delta^{18}O_{marl}$ in the deep-basin core compared with an increase of 2-3‰ in δ^{18}O in mollusks in a North Bay core (Stuiver, 1970), also attests to less variation in salinity in the North Bay than in the deep basin. Stuiver's δ^{18}O values from modern mollusks are about 2‰ lower than the ostracode values, even after correction for ^{18}O fractionation between aragonite (mollusks) and calcite (ostracodes). If this difference was due entirely to temperature, it would imply that waters in North Bay were about 8 °C warmer than those of the open lake. Various authors suggest decreased precipitation and increased temperatures in the mid-continent between 9 ka and 6 ka (e.g. Kutzbach, 1987; Bartlein et al., 1984; Dorale et al., 1992; Bartlein & Whitlock, 1993). Today, summer drought over the Great Plains is usually associated with higher temperatures linked to deep warm anticyclones periodically fed by dry subsiding tongues of air emanating from the westerlies (Namias, 1983). These periods of drought in the Great Plains occur when the dry Pacific airstream dominates over the Arctic and Gulf airstreams as a result of stronger westerly zonal winds (Bryson, 1966). However, ostracode and diatom assemblages from Elk Lake cores, northwestern Minnesota, show that Elk Lake was colder and more saline than at present until at least 6700 varve yr with conditions similar to those that exist today in cold prairie lakes of Canada (Forester et al., 1987; Dean & Stuiver, 1993). Bradbury et al. (1993) argue that winters in northwestern Minnesota may have been cold during the mid-Holocene because disintegrating northern ice sheets ceased to block winter outbreaks of Arctic air. We suggest that the increase in δ^{18}O in mollusks from mid-Holocene sediments in Pickerel Lake North Bay that Stuiver (1970) related to increased mean annual air temperature during the mid Holocene, is rather the effect of increased evaporation, because an increase in temperature should have led to higher water temperatures that should have been translated into lower values for δ^{18}O in the sediments. The increase in δ^{18}O in North Bay as well as the presence of *L. ceriotuberosa* also suggests that North Bay was more affected by increased evaporation than the open lake, indicating that North Bay may have been separated from the open lake as a result of a drop in lake level.

Late Holocene (<2.2 ka)

The uppermost 5.8 m of sediment, representing approximately the last 2.2 ka of deposition, show persisting prairie environment, although the presence of *D. stevensoni* indicates higher groundwater input that might have resulted from slightly moister conditions. The upland and lake vegetation was the same as now, with prairie dominating the upland, and with abundant oak and ash deciduous forests common around lakes and streams (Watts & Bright, 1968). The recurrence of deciduous forest has probably led to a slight decrease in $\delta^{13}C_{marl}$. The relatively small difference between ostracode and marl isotopes might be the result of a decrease in productivity and lower temperature and/or shallowing of the lake as indicated by diatoms characteristic of surf zones in lakes (Haworth, 1972) and increased numbers of littoral ostracodes.

The upper Holocene section is characterized by distinct 1-m cycles in magnetic susceptibility, % C_{org}, and δ^{13}C (Figure 5), suggesting that there were cyclic variations in the influx of detrital clastic material and in organic productivity. There are five cycles in the top five meters of section, estimated to have been deposited over the last 2000 years, yielding an average cycle period of 400 years. A late Holocene period of cyclic variations in salinity in Moon Lake in southeastern North Dakota (Laird et al., 1996a, 1996b), approximately 190 km north-northwest of Pickerel Lake, is inferred from diatom assemblages in the sediments. The Moon Lake record shows periods of extreme drought that were more frequent prior to AD 1200.

The 1-m cycles in the top four meters of the Pickerel Lake magnetic susceptibility record bears a remarkable resemblance to the record of aluminum (Al) concentration in varved sediments deposited over the last 1550 years in Elk Lake, northwestern Minnesota (Figure 5). The main allochthonous component in Elk Lake is detrital clastic material, as measured by bulk-sediment concentrations of aluminum, sodium, potassium, titanium, and quartz, that enters the lake mostly as eolian dust (Dean et al., 1996). Geochemical records of eolian activity from Elk Lake exhibit distinct cyclicities with dominant periodicities of 400 and 84 years (Dean, 1997). The most obvious Elk Lake cycles shown in Figure 5 are three groups of peaks about 400 years apart. The youngest and largest group of peaks (400 to 200 years ago) corresponds in time to the main phase of the Little Ice Age (LIA; AD 1550–1700; Lamb, 1977), and the highest concentrations correspond in time to the Maunder sunspot minimum (AD 1640–1710). If the proxy of Al for eolian dust is correct, then the Elk Lake record suggests that northwestern Minnesota experienced windier and dustier conditions during the last 1550 years in cycles with periods of about 400 years. We suggest that the prominent susceptibility peak centered at about 50 cm in the Pickerel Lake core corresponds to the LIA peak in Elk Lake, and the group of peaks between 240 and 300 cm in the Pickerel Lake core corresponds to the Medieval Warm Period (MWP; Figure 5). If the correlations between magnetic susceptibility in Pickerel Lake and % Al in Elk Lake are correct, then the varve-calibrated time scale for Elk Lake can provide a more precise time scale for the top 4m of the Pickerel Lake record than the crude extrapolation from one AMS ^{14}C date.

Although the ultimate mechanisms that control the 400-year cycles in Pickerel Lake sediments have not been resolved, the fact that lower values of both C_{org} and δ^{13}C (Figure 5) correspond to peaks in magnetic susceptibility suggest that productivity in the lake was lower during periods of greater influx of eolian clastic material. Perhaps influx of detrital clastic material produced turbidity that limited algal productivity. Relatively high values for δ^{18}O during periods of increased detrital input, as, for example, shown for the MWP and the LIA (Figure 5), indicate greater evaporation and/or cooler water temperatures associated with dry, windy conditions. We suggest, therefore, that peaks in magnetic susceptibility reflect increased aridity that occurred every 400 years, including the MWP and the

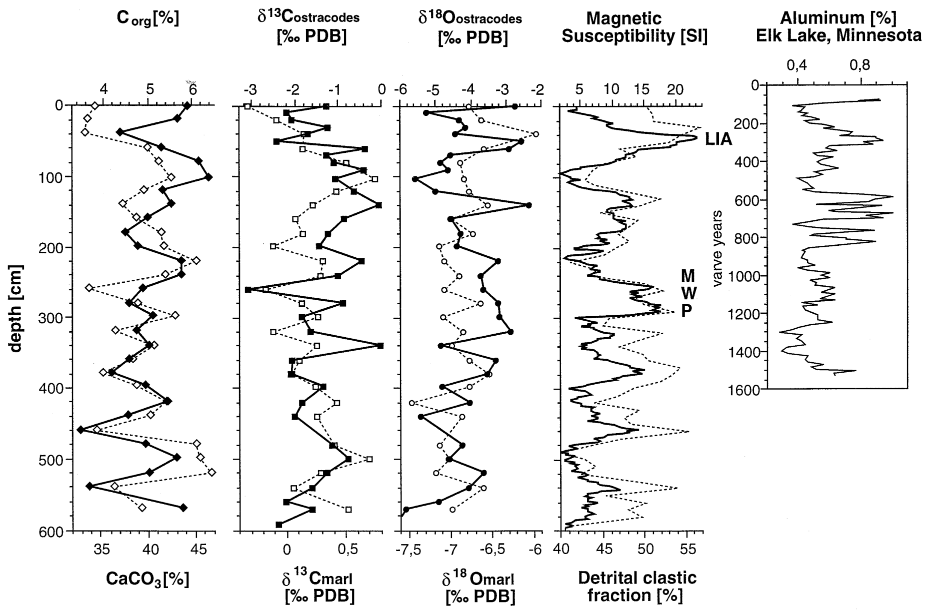

Figure 5. Cycles of organic carbon (C_{org}; solid symbols), carbonate ($CaCO_3$; open symbols), δ^{18}O and δ^{13}C for both ostracodes (solid symbols) and marl (open symbols), magnetic susceptibility (solid line), and detrital clastic fraction (dashed line) in upper 6 m of core. MWP and LIA locate the intervals of the Medieval Warm Period and the Little Ice Age. Peaks in magnetic susceptibility in the Pickerel Lake core correspond to peaks in aluminum from a varve-dated core from Elk Lake, Clearwater County, Minnesota.

LIA. Laird et al. (1996b) inferred dry conditions during the MWP and a complex LIA interval from the Moon Lake, North Dakota, record. Fritz et al. (1994) inferred an arid climate and steep climatic gradients between the Great Plains and the regions to both east and west for the LIA. Arid conditions for the LIA are also consistent with results from the Greenland summit ice core that has highest concentrations of sea salt and terrestrial dust since the Younger Dryas in the LIA interval of the core (O'Brien et al., 1995), thus indicating windier and a perhaps drier global climate.

Conclusions

The late Glacial climate in the northern Great Plains was controlled by the dynamics of a fluctuating ice sheet and glacial Lake Agassiz. The sediments from Pickerel Lake suggest that climate before 12 ka was cool, moist, and windy. A combination of strong zonal circulation and strong monsoons may have provided mechanisms for increased precipitation.

Between approximately 12 and 10.3 ka cold and dry conditions with relatively warm summers prevailed in response to the retreat of the ice sheet and subsequent drainage of glacial Lake Agassiz into Lake Superior during the Younger Dryas interval. Reduced detrital influx was the result of a change in vegetation from tundra grasses to a dense spruce forest.

Increases in both δ^{13}C and δ^{18}O in marl between 10.3 and 9.5 ka, as well as the presence of *L. herricki* and *C. rawsoni* in the ostracode assemblages, indicate that evaporation increased and that Pickerel Lake became more saline, probably reaching the highest salinity in the history of the lake. Increased evaporation may have caused a drop in water level resulting in erosion of the littoral zone, as suggested by increases in magnetic susceptibility and the detrital clastic fraction.

After 9.5 ka, the drainage of Lake Agassiz north into the McKenzie River system may have initiated the reorientation of the groundwater flow pattern in the northern Great Plains, Pickerel Lake seems to have shifted from a system controlled by atmospheric changes to a system controlled by local groundwater seepage. A sharp increase in C_{org} concentration between 8.5 ka and 8 ka marks the transition to a more productive mid-Holocene lake, and the vegetation in the drainage changed from spruce forest to mixed C3 and C4 plants in an open oak savannah. Zonal flow introducing dry Pacific air probably became more important during the prairie period.Lower values of $\delta^{18}O_{ostracode}$ and $\delta^{13}C_{ostracode}$ at about 8 ka, 4 ka, and 2.2 ka as well as the presence of *D. stevensoni* suggest periods of enhanced groundwater input, perhaps related to short periods of stronger meridional circulation introducing moist air from the Gulf of Mexico.

The increase of 2–3‰ in $\delta^{18}O$ of mollusk shells deposited during the mid-Holocene in shallow North Bay reported by Stuiver (1970) is not seen in values of $\delta^{18}O$ in either marl or ostracodes from our open lake core. This suggests that sheltered conditions in North Bay, presumably caused by a separation of North Bay from the open lake after a drop in lake level, resulted in greater evaporation at that time in response to drier climate.

Sediments deposited in Pickerel Lake over the past 2.2 ka are characterized by distinct 400-yr cycles in magnetic susceptibility, detrital clastic material, $\%C_{org}$, and $\delta^{13}C$. These cycles are interpreted as aridity cycles in which increased eolian activity caused an increase in the influx of detrital clastic material and reduced organic productivity. Relatively high values of $\delta^{18}O_{ostracode}$ during periods of increased detrital input (increased magnetic susceptibility) suggest that evaporation was greater in association with dry, windy conditions during periods that include both the MWP and the LIA. The cycles point to increased climatic variability provoked by frequent switching between periods of zonal and meridional circulation during the late Holocene.

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