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Century-scale paleoclimatic reconstruction from Moon Lake, a closed-basin lake in the northern Great Plains

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Abstract

Estimates of past lake-water salinity from fossil diatom assemblages were used to infer past climatic conditions at Moon Lake, a climatically sensitive site in the northern Great Plains. A good correspondence between diatom-inferred salinity and historical records of mean annual precipitation minus evapotranspiration (P − ET) strongly suggests that the sedimentary record from Moon Lake can be used to reconstruct past climatic conditions. Century-scale analysis of the Holocene diatom record indicates four major hydrological periods: an early Holocene transition from an open freshwater system to a closed saline system by 7300 B.P., which corresponds with a transition from spruce forest to deciduous parkland to prairie and indicates a major shift from wet to dry climate; a mid-Holocene period of high salinity from 7300 to 4700 B.P., indicating low effective moisture (P − ET); a transitional period of high salinity from 4700 to 2200 B.P., characterized by poor diatom preservation; and a late Holocene period of variable lower salinity during the past 2,200 yr, indicating fluctuations in effective moisture.

Potential changes in water availability in important agricultural areas are a major concern, especially in drought-prone regions, such as the northern Great Plains, where even minor climatic change might increase the frequency and severity of drought. To measure or predict possible anthropogenic influences on the climate system, we must first understand natural climatic variability. Proxy records of past climatic conditions can provide the long time series needed to establish these natural patterns.

Lakes are intricately linked to the hydrological system and can be sensitive recorders of climatic change. Patterns of continental-scale lake-level fluctuations have been used to infer past changes in general circulation patterns in North America (Harrison and Metcalfe 1985; Harrison 1989), Europe (Guiot et al. 1993), and Africa (Street and Grove 1976). Within topographically closed lakes, fluctuations in the water budget affect both lake level and lake chemistry. Because diatoms are sensitive to lake chemistry, fossil diatom assemblages preserved in lake sediments record past hydrochemical conditions. In this study, we use quantitative salinity-inference techniques based on modern diatom distribution in relation to water-chemistry gradients in the northern Great Plains (Fritz 1990; Fritz et al. 1991, 1993) to reconstruct changes in effective moisture (precipitation minus evapotranspiration, P − ET) for a climatically sensitive site, Moon Lake, North Dakota. Century-scale climatic fluctuations at this site are then compared with other paleoclimatic records from the northern Great Plains to assess broad-scale regional trends vs. local-scale variability.

Study area

The northern Great Plains is a vast region of prairie and steppe in central North America characterized by level to rolling topography (Fig. 1). Prior to agricultural settlement, the natural vegetation was composed of three major plant communities: eastern tall-grass prairie, central mixed-grass prairie, and western short-grass plains (steppe) (Carpenter 1940). Climate is extreme and variable, often with winter temperatures as low as −29°C and summer temperatures as high as 38°C and varying pre-
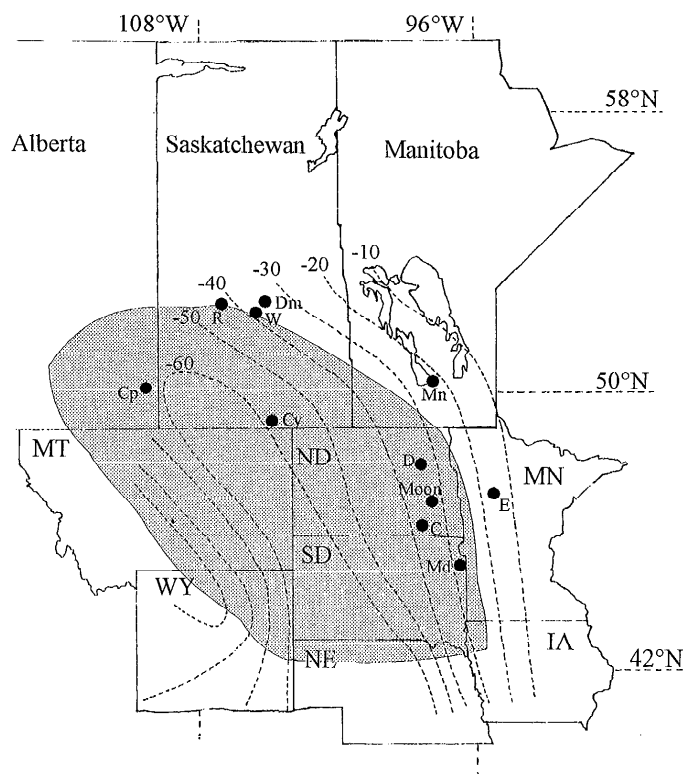
Fig. 1. Map of the northern Great Plains (shaded) showing the location of Moon Lake and other sites referred to in the text: Ceylon—Cy; Chappice—Cp; Coldwater—C; Deadmoose—Dm; Devils—D; Elk—E; Manitoba—Mn; Medicine—Md; Redberry—R; Waldsea—W. Isolines (cm yr^{-1}) are equal precipitation minus evapotranspiration.

Fig. 2. Morphometric map of Moon Lake showing core locations. Short core—X1; long core—X2. Depths in meters.

cipitation regimes as a result of the continental location and intrusions of air masses of differing characteristics: warm dry flow from the Pacific, cold dry Arctic air, and moist tropical air from the Gulf of Mexico (Bryson 1966). Much of the region was ice covered during the last glaciation, resulting in the formation of thousands of lakes. Many of these are closed-basin lakes, ranging in salinity from fresh to hypersaline, with a diversity of hydrochemical, sedimentological, and biological characteristics (Fritz et al. 1993; Last 1992).

Moon Lake (46°51′27″N, 98°09′30″W, elevation 444 m), Barnes County, North Dakota, has an area of 35 ha and a relatively small watershed (1,166 ha) (Fig. 1). At Valley City, ~8 km northeast of Moon Lake, the average annual temperature is 6°C (range, −29 to 38°C) and the average annual precipitation is 45.9 cm, most of which falls during the growing season (April–September). Surficial geology of Barnes County is dominated by ground moraine with gently undulating hills and by linear belts of end moraine with higher and more strongly rolling hills interspersed with numerous closed depressions. Moon Lake lies within the Kensal-Oakes drift, an end moraine associated with a readvance of the Des Moines ice lobe of the last glaciation (Kelly and Block 1967), and is in the Stoney Slough outwash channel that cuts across parts of the Kensal-Oakes end moraine. The outwash overlies

glacial till, which in turn overlies the Pierre Shale, the bedrock unit in much of Barnes County.

The Stoney Slough Aquifer, 3–30 m below land surface, is a surficial aquifer of glacial outwash that is recharged primarily by precipitation and runoff and secondarily by subsurface flow (Kelly 1966). The Pierre Shale and glacial drift are hydrologically connected and are similarly recharged by precipitation and runoff, normally immediately after spring thaw (Kelly 1966). Only during periods of above-normal rainfall are the aquifers recharged during the growing season. Although most of the annual precipitation falls during this period, much of it is lost through evaporation and evapotranspiration and never reaches the water table.

Methods

A core 1.2 m long, encompassing approximately the past 250 yr, was collected on 20 August 1991 at a water depth of 12 m (Fig. 2) with a modified piston corer operated with rigid drive rods equipped with a clear plastic tube to ensure an undisturbed mud–water interface (5-cm i.d.) (Wright 1991). The core was held upright and extruded in the field at 1-cm intervals from 0 to 30 cm,

Table 1. Radiocarbon AMS dates (Lawrence Livermore Laboratory) for Moon Lake. Core depth indicates depth from which dated material was collected.

Core depth (cm)	Material	CAMS (lab No.)	^{14}C date (B.P.)	Error (1 SD B.P.)
110–114	charcoal flakes	6821	460	60
188–192	charcoal flakes	14337	830	60
245–246	charcoal flakes	14338	1080	50
342–347	charcoal flakes	6822	2300	60
468–476	charcoal flakes	15294	4590	60
564–568	*Ambrosia* seed	6824	5410	70
564–568	charcoal flakes	6823	5420	60
660–668	charcoal flakes	15295	6310	70
830–838	charcoal flakes	6825	6610	90
900–908	charcoal flakes	9156	7300	80
978–981	charcoal flakes	6826	9530	130
1,034–1,039	*Picea* needle	6827	10,780	100
1,096–1,100	woody twig	6828	11,770	80
1,120–1,125	woody twig	6829	11,830	80

2 cm from 30 to 90 cm, and 4 cm from 90 to 120 cm. Core chronology is based on ^{210}Pb dating using the constant rate of supply model (Appleby and Oldfield 1978; Jacobson and Engstrom 1989). *Salsola,* an annual non-native weed, was used as an independent stratigraphic marker to verify the accuracy of the ^{210}Pb chronology (B. Hansen unpubl. data; e.g. Jacobson and Engstrom 1989). A paleosalinity reconstruction from this core (*see below*) was compared to the climate records from Valley City and Jamestown, ~100 km west of Moon Lake. Mean annual P − ET was calculated from monthly precipitation and temperature records. Valley City temperature and precipitation records are incomplete for the early years of weather recording, consequently records for 1893–1917 are based on weather data from Jamestown. There is a highly significant correlation ($r = 0.63$, $P < 0.01$) between the annual P − ET of Jamestown and Valley City from 1918 to 1988. The Turc equation was used to calculate potential evapotranspiration from recorded precipitation and temperature data (Wovcha 1989). A 5-yr weighted running average of P − ET was calculated for comparison to the diatom-inferred salinity record.

A 11.2-m sediment core, with a basal date of 11,800 B.P., was collected in 13.2 m of water in the deepest part of the basin (Fig. 2) on 31 January 1986 with a modified Livingston corer (Wright et al. 1984). Chronology is based on 14 accelerator mass spectrometry (AMS) radiocarbon dates of terrestrial charcoal and plant macrofossils (Table 1). Macrofossils are infrequent throughout most of the core, except for the late-glacial *Picea* zone; however, charcoal is abundant. Each dated charcoal sample consisted of 50–100 small flakes of charcoal picked from the screened sediment under a stereomicroscope. Care was taken to select charcoal and not lignite, which is frequent in the core. Charcoal is distinguished from lignite by its shiny black color with obvious cellular structure; lignite is a dull, very dark brown and lacks obvious cellular structure. Most of the charcoal was grass epidermis. The paired
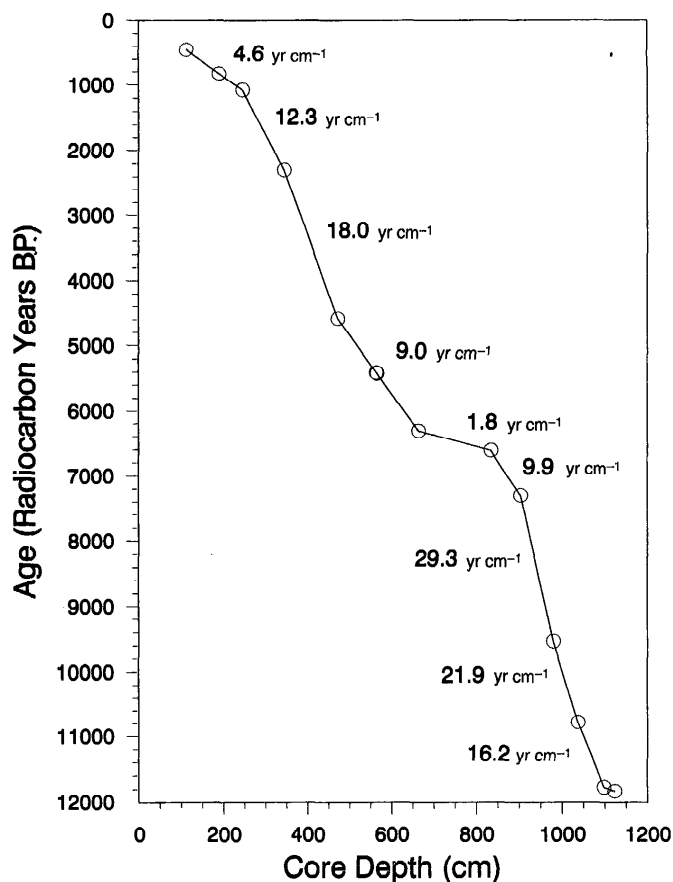

Fig. 3. Rate of sediment deposition in the long core, based on 14 radiocarbon AMS dates (Table 1), with degree of resolution (yr cm^{-1}).

dates of charcoal and an *Ambrosia* seed, which differ by only 10 yr, demonstrate the suitability of charcoal for dating (Table 1). For discussion of the long core, unless otherwise stated, all ages are given in radiocarbon years B.P. To compare the last 4,000 yr of the record with other calendar-year chronologies, the radiocarbon ages were calibrated to calendar years with the CALIB 3.0 program (Stuiver and Reimer 1993) with the bidecadal calibration curve (Stuiver and Pearson 1993).

Subsamples (1 cm) for diatom and pollen analysis were taken at 4–8-cm intervals with a temporal resolution of ~50–200 yr (Fig. 3). Sediment samples for pollen analysis were treated with sodium pyrophosphate, sieved through a 7-μm Nitex screen (Cwynar et al. 1979), and then treated with 10% KOH, hot 10% HCl, and acetolysis solution for 1 min (Fægri et al. 1989). A minimum of 400 pollen grains were counted on a Leitz Dialux microscope at 400×, with critical determinations made under oil at 1,000×. Botanical nomenclature follows that of the Great Plains Flora Association (1986).

Sediment samples for diatom analysis were processed in cold hydrogen peroxide (30%) with potassium dichromate to oxidize organic matter and rinsed several times with distilled water until free of peroxide. The resultant slurries were settled onto coverslips, and the dried cov-

erslips were mounted onto slides with Naphrax. Diatoms were counted in transects under oil immersion either on an Olympus BH2 microscope with a 100× plan apochromatic objective (numerical aperture, NA = 1.4) and interference contrast or on a Leitz DMRB microscope with a 100× fluotar objective (NA = 1.3) with brightfield. A minimum of 300 valves identifiable to species were counted unless specified otherwise. In a few samples, diatom concentrations were low or diatom preservation was poor, and fewer valves were enumerated. Recognizable valve fragments (¼, ½, ¾) of centric diatoms (primarily *Cyclotella quillensis*) and ½ fragments of araphid diatoms (e.g. *Synedra fasciculata*) were counted and included in the final counts. Identifiable central areas of biraphid diatoms were counted as a single valve (e.g. *Rhoicosphenia curvata, Mastogloia smithii,* and *Mastogloia elliptica*). Taxonomic names and authorities follow those of Fritz et al. (1993).

Salinity reconstructions were calculated by a salinity-weighted-averaging regression and calibration model with the program WACALIB v. 3.3 (Line et al. 1994) and based on a transfer function developed by Fritz et al. (1991, 1993). Mean bootstrap estimates of salinity calculated from 1,000 cycles with the above program are highly correlated with the weighted-average salinity estimates (r = 0.99). Nine salinity estimates for the period between 4700 and 2200 B.P. were based on counts of <50 valves and were consequently eliminated. Samples were not enumerated at the base of the core because of poor diatom preservation. The diatom-inferred salinities for the last 4,000 yr were smoothed with a fast Fourier transform (FFT) filter with a 5-point smoothing window (Press et al. 1988).

As a means of evaluating whether the transfer function provided suitable analogs for the salinity reconstruction of Moon Lake, a canonical correspondence analysis (CCA) ordination constrained to only salinity was undertaken with the computer program CANOCO v. 3.10 (terBraak 1988, 1990) to calculate the squared residual distance of each fossil diatom sample from the salinity axis (CCA axis 1). The degree of fit was determined by comparing the squared residual distance of the fossil samples with those of the modern samples used to derive the transfer function (Fritz 1990; Fritz et al. 1991, 1993). The distribution of squared residual distances of the modern samples was skewed, and consequently values from both the modern and fossil samples were log-transformed before determining 90 and 95% confidence intervals. Fossil samples with residual distances within the 90% confidence intervals of residual distances in the modern samples are considered to have a good fit to salinity, those in the extreme 10% are considered to have a poor fit, and those in the extreme 5% are considered to have a very poor fit (Birks et al. 1990).

Results and discussion

The historical relation between diatom-inferred salinity and climate—The relations between lake-water salinity and climate are complex and depend on the geologic,
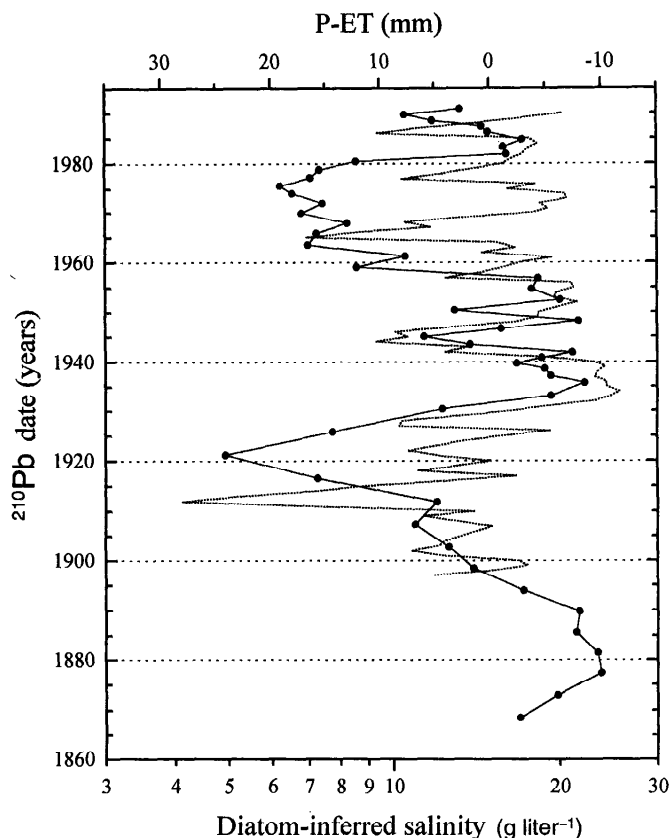

Fig. 4. Comparison of diatom-inferred salinity (●——●) and instrumental climate record (dotted line). The climate record is a weighted 5-yr running average of precipitation minus evapotranspiration. Chronology is based on ^{210}Pb dating.

hydrologic, and climatologic setting of the lake. Hence, to reconstruct past climate from the paleolimnological record the relation between salinity fluctuation and climate should be determined for each lake.

Moon Lake has several salient features that make it a suitable system for paleoclimatic reconstruction, including a small catchment, a topographically closed basin, a continuous Holocene sediment record, and a hydrological setting in which groundwater recharge is primarily driven by precipitation. Comparison of the diatom-inferred salinity of Moon Lake with historical records of mean annual P − ET, calculated from monthly temperature and precipitation records from the Valley City and Jamestown climate stations, suggests that Moon Lake responds hydrologically to fluctuations of effective moisture (Fig. 4). The correspondence between diatom-inferred salinity and climate after 1940 is extremely good, with the exception of the mid-1970s. In 1964 the North Dakota Department of Game and Fish began a stocking program, primarily rainbow trout, in Moon Lake. In 1975, Moon Lake underwent intense chemical treatment to eliminate a green sunfish population (VanEeckhout 1987). This disturbance may account for the failure of diatom-inferred salinity to record the 1970s drought. Before 1940, diatom-inferred salinity does not track the short-term climate
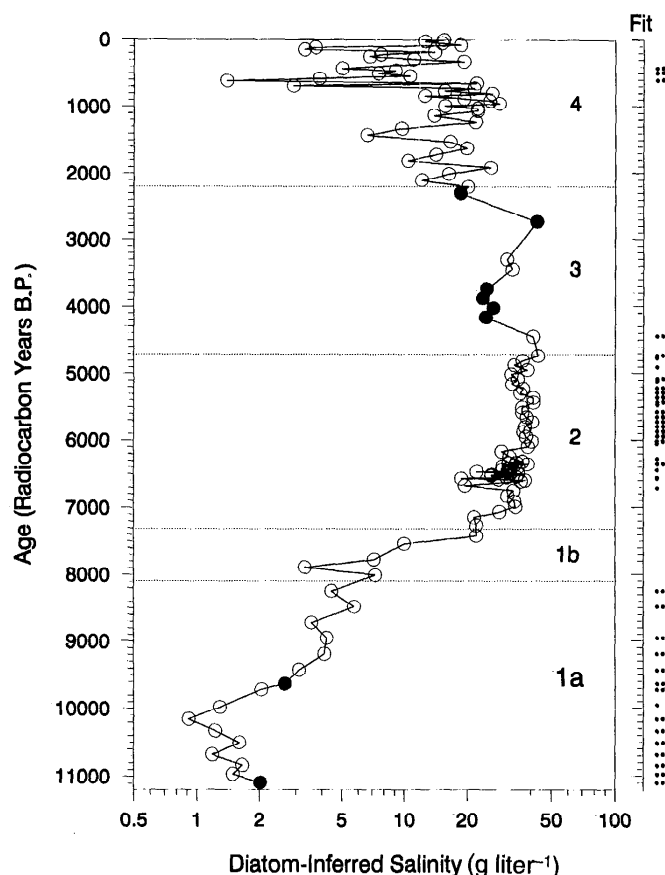
Fig. 5. Holocene diatom-inferred salinity reconstruction. Reconstructions based on counts of at least 100 but <150 valves—●; reconstructions based on counts of at least 300 valves—○. The four zones are described in the text. The "fit" diagram (on right) is the degree of fit or reliability of the salinity reconstruction based on the northern Great Plains training set: samples that fall in the extreme 10%—●; samples that fall in the extreme 5% distribution—●● (further details given in text). Chronology is based on linear interpolation between AMS dates (Table 1).

fluctuations as well, probably because of the decreased resolution of the sediment samples from 1- to 2-cm intervals below 30-cm depth (1940 ^{210}Pb yr); however, it does record the larger scale trends. Although the hydrological response of the lake may have been altered somewhat by the breakup of prairie sod and later agricultural practices and most recently by fisheries management, we believe that the close correspondence between diatom-inferred salinity and P − ET justifies the use of paleosalinity fluctuation as a proxy for moisture availability.

Century-scale analysis of the Moon Lake Holocene sediment record—Century-scale analysis of the Holocene diatom record indicates four major hydrological periods (Fig. 5): an early Holocene transition from an open lake of low salinity (<2 g liter^{-1}) to a closed saline system (salinity >20 g liter^{-1}) between 10,000 and 7300 B.P., which corresponds with the transition from spruce forest to deciduous parkland to prairie (Fig. 6) and suggests a

major shift from wet to dry climate; a mid-Holocene period (7300–4700 B.P.) of high salinity, indicating low effective moisture (P − ET); a transitional period from 4700 to 2200 B.P. of high salinity, characterized by poor diatom preservation; and a late Holocene period of variable lower salinity (past 2,200 yr), indicative of fluctuations in effective moisture.

The transition of Moon Lake from a freshwater system to a closed saline system is probably a consequence of decreasing lake depth and volume as a result of increasing aridity. This shift from fresh to saline conditions occurred in two stages: first, a gradual transition between ~10,000 and 8100 B.P. from ~1 to 4 g liter^{-1} (Fig. 5, period 1a), followed by a more rapid transition from 5 to >20 g liter^{-1}, culminating by ~7300 B.P. (Fig. 5, period 1b). Freshwater taxa (salinity optima <3 g liter^{-1}) dominate the first stage of the transition and are subsequently replaced by hyposaline taxa (salinity optima, 3–10 g liter^{-1}) (Fig. 7). Moon Lake was likely open (possessing an outlet) during the freshwater stage of the early Holocene with a much larger volume and surface area than today. With a lake-level rise of ~3 m above the present level, Moon Lake would coalesce with St. Marys Lake to the south and more than double its surface area. A rise of ~5 m from the present level would be required for the combined lake to have an outlet.

The early stages of lake development were characterized by a relatively cool and moist climate, as indicated by the presence of spruce forest until ~10,300 B.P. (Fig. 6). Spruce forest was replaced by a parkland of mixed deciduous forest dominated by elm and oak, with openings of grasses (Poaceae) and *Artemisia*. If Moon Lake were open, then the regional water table would have been considerably higher, and a much larger area would have been flooded with a greater percent coverage of poorly drained soils. This implication is consistent with the early Holocene abundance of *Ulmus* (elm), which thrives best on poorly drained soils. Elm gradually disappeared by 8000 B.P., when *Ambrosia* began to increase sharply. *Quercus* (oak) remained until 7000 B.P., suggesting a climate not sufficiently dry to eliminate scattered oak groves. The onset of very high salinity at ~7000 B.P., as inferred from the diatoms, correlates with the disappearance of oak and the appearance of *Ruppia*, a taxon that suggests shallow high-salinity waters (e.g. Vance et al. 1993). There is a similar correspondence between vegetation and salinity at Medicine Lake, South Dakota (Radle et al. 1989).

The preservation of aragonite and gypsum laminations (B. Valeros-Garcés unpubl. data) during the mid-Holocene (Fig. 5, period 2) suggests at least seasonally anoxic conditions at the sediment–water interface, indicating possibly a highly productive or a meromictic lake. The presence of aragonite in the sediments suggests a high Mg : Ca ratio in the lake water (Kelts and Talbot 1990). Evidence for a meromictic state is suggested by the dominance of *Cyclotella choctawhatcheeana* (Fig. 7A), which is most abundant in modern-day surface samples from meromictic lakes (Fritz et al. 1993). Similarly, Medicine Lake, a meromictic lake, became dominated by *C. choctawhatcheeana* with the expansion of prairie conditions (Radle et al. 1989). *C. choctawhatcheeana* (Prasad et al.
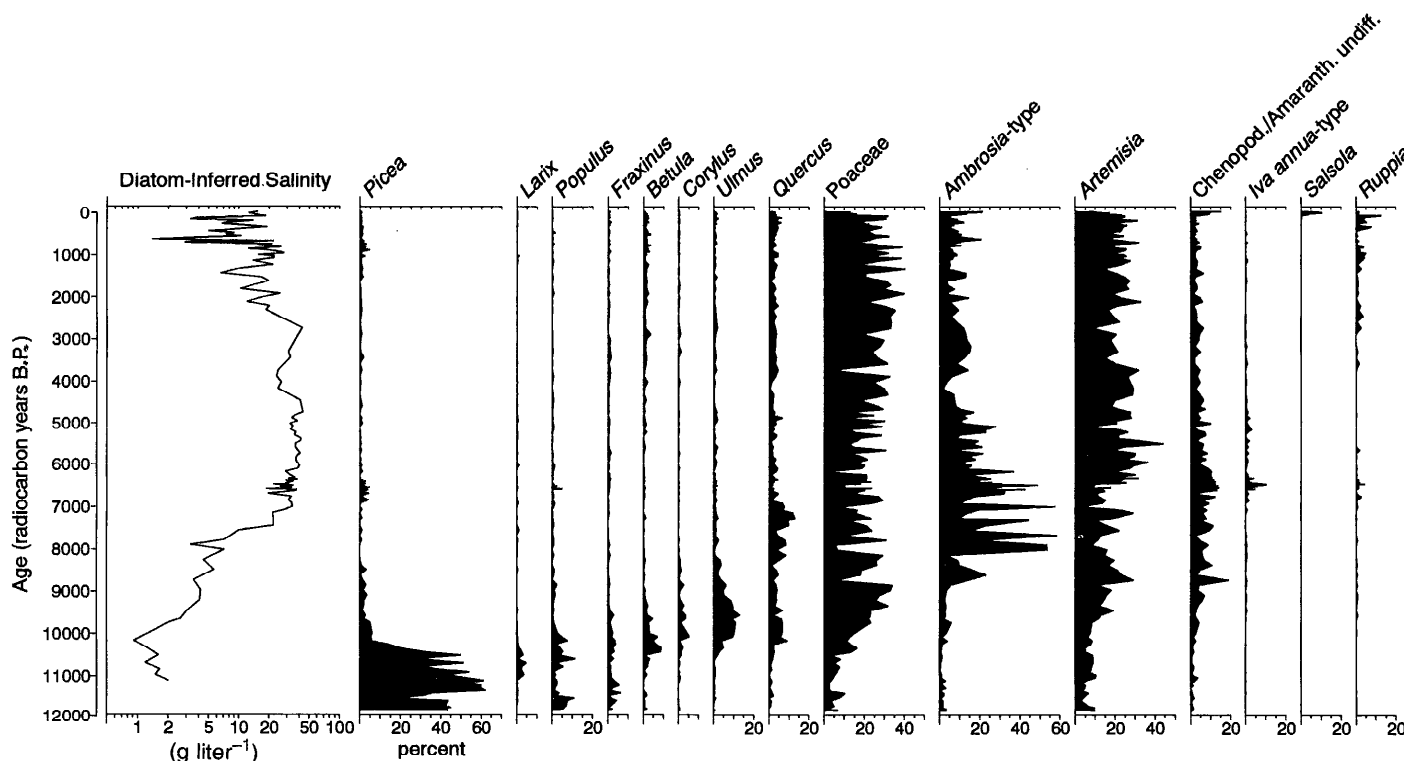
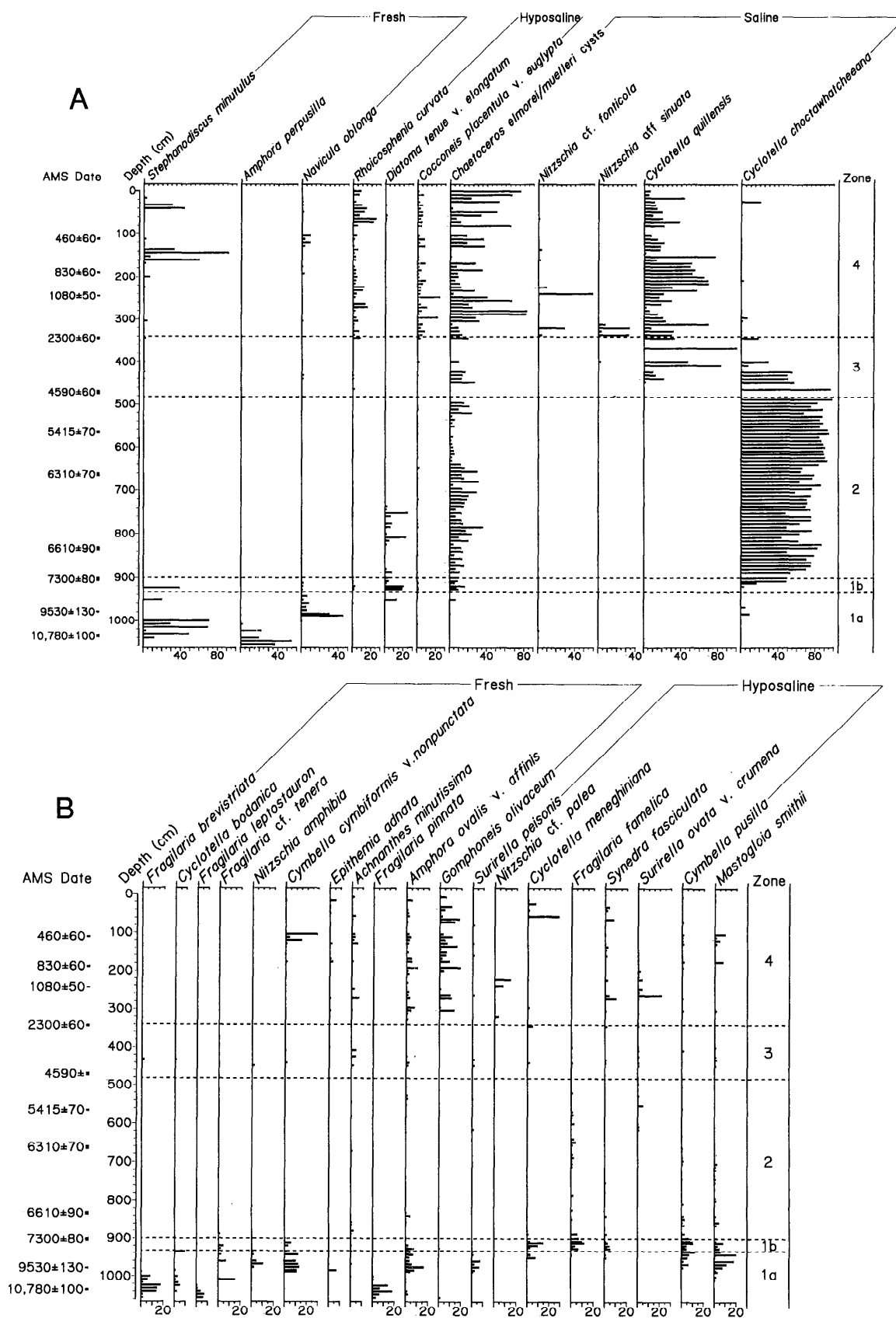Fig. 6. Summary pollen diagram (%) from Moon Lake (P. G. Mueller analyst).

1990) is currently recognized as the correct epithet for the taxon referred to as *Cyclotella caspia* by Radle et al. (1989), Fritz (1990), and Fritz et al. (1991, 1993) (Carvalho et al. 1995).

During the mid-Holocene, a short interval of rapid sediment deposition (~6600–6300 B.P.) coincides with peaks in *Iva annua*, *Ruppia*, and *Picea* (spruce) pollen. Except for a few outliers, the modern range of *I. annua* does not extend north of mid-Nebraska (Great Plains Flora Assoc. 1977), and its presence in the sediment record suggests significantly warmer temperatures during this period. *I. annua* typically grows on moist, disturbed sites, such as exposed lake muds. The peak in *Ruppia* suggests shallow water, and the peak in spruce suggests erosion and redeposition of exposed older nearshore sediments at times of lowered lake levels, with *I. annua* growing on the exposed muds. Whereas the pollen spectra and presence of gypsum muds (B. Valero-Garcés unpubl. data) suggest the lowest effective moisture of the Holocene, the diatoms do not particularly distinguish this interval. The lack of a distinct diatom response is a result of the dominance of *C. choctawhatcheeana* throughout the mid-Holocene (Fig. 7A) and the limited number of hypersaline taxa in the modern-day samples; as a consequence, the current inference model is insensitive to extreme salinity excursions beyond the tolerance range at which this taxon dominates. However, in this interval there is a higher abundance of *Chaetoceros elmorei/muelleri* cysts (Fig. 7A), a taxon that seems to reach its highest percent abundance in shallow saline lakes (K. Laird unpubl. data; J. Johansen pers. com.). Even so, there is not a significant relationship to depth and percent abundance of this taxon in surface

samples of the modern-day northern Great Plains lake data set (Fritz et al. 1993).

A change in lithology from thin aragonite laminations to a massive unit of gray lake sediment with several thick laminae of mixed aragonite-calcite (B. Valeros-Garcés unpubl. data) coincides with the onset of period 3 (Fig. 5). The hydrochemical explanation for this period is unclear: it could be a result of changes in lake-water chemistry, a lowering of lake level, a decrease in the extent of seasonal anoxia, increased water-column mixing, or some combination of these. Poor diatom preservation during this period may be caused by increased turbulence and sediment mixing as a result of lowered lake level or increased wind action. Nine samples in this interval had very low diatom concentrations (<50 valves enumerated) and were excluded from the salinity reconstruction. Species preserved during this period are the saline taxa *C. elmorei/muelleri*, *C. quillensis*, and *C. choctawhatcheeana* (Fig. 7A). The first two taxa are heavily silicified and tend to preserve well, whereas the third taxa has a sufficiently distinctive central area to allow identification of even poorly preserved specimens. Thus, these fossil assemblages may not accurately represent the living community of this entire period. During period 3 the assemblage switches from *C. choctawhatcheeana* dominance to either *C. quillensis* or *Chaetoceros* spp. dominance. Although the limnological mechanism is unclear, coincident changes in sedimentology, stratigraphy, and preservation suggest a major environmental change.

The past 2,200 yr of the record represent typically high-salinity conditions, with small fluctuations from 2200 to 750 B.P. (mean, 18 g liter^{-1}) and larger fluctuations be-

tween fresh and saline conditions from 750 to 100 B.P. (mean, 8.5 g liter^{-1}). The fresh periods are dominated by *Stephanodiscus minutulus,* a common planktonic taxon often found in eutrophic freshwater lakes. Either *C. quillensis* or *Chaetoceros* spp. dominates the saline periods (Fig. 7A). The variability in salinity during this interval suggests that the climate was also highly variable. The apparent higher variability of salinity during the past 2,200 yr in comparison to the mid-Holocene, however, is the result of the log-scale plot (Fig. 5), which is necessary because of the log-based salinity inference model and for determination of confidence intervals. As a consequence, fluctuations at higher salinities (>20–30 g liter^{-1}) are dampened, reflecting the loss of power of the inference method and the inability to evaluate the significance of the high-salinity variations.

Evaluation of the reliability of suitable analogs for the salinity reconstruction is represented by the degree of fit of the core samples to the northern Great Plains training set, in which the training set is the modern-day surface samples with corresponding chemical data (Fritz et al. 1993). Of fossil samples, 67% fall within the 90% C.I. of the modern-day samples and therefore are adequately represented in the training set, whereas 9% of the samples fall within the extreme 10%, and 24% of the samples fall within the extreme 5%. Most of these extreme samples (41 out of a total of 44) occur in periods 1a and 2 (Fig. 5). The lack of fit of the samples in the freshwater stage (period 1a) is a result of the current deficiency of freshwater sites in the northern Great Plains training set. The samples with poor fit in the mid-Holocene (period 2) are highly dominated by *C. choctawhatcheeana* (>75–80% relative abundance). Although the current training set is fairly representative of this taxon's distribution, its abundance never exceeds 38%. Although the absolute salinity values for samples with poor fit may be somewhat unreliable due to inadequate representation in the current training set, the general trends would remain the same and, in the case of periods 1a and 2, would probably be amplified (e.g. the fresh periods becoming "fresher").

*Regional patterns of climatic change: early to mid-Holocene—*The long transition of Moon Lake from a freshwater to a closed saline system in the early Holocene, characterized by a gradual increase from fresh to hyposaline conditions between ~10,000 and 8100 B.P., followed by a rapid and large increase in salinity by ~7300 B.P., is in contrast to an apparent abrupt early Holocene transition from freshwater to saline at ~9200 B.P. at Medicine Lake (Radle et al. 1989; Kennedy 1994), and ~8000 B.P. at Devils Lake (Fritz et al. 1991). A summary of the lakes discussed is given in Fig. 8. The differences among lakes in the timing and rate of the early Holocene salinity increase may reflect differential hydrological responses to a changing climate, regional variability in moisture availability or may be a result of differences in chronological control and sampling resolutions between sites. The Medicine Lake record in particular is poorly dated (Radle et al. 1989) and considerable uncertainty surrounds the age estimate of 9200 B.P. for the beginning of the high-salinity episode. Nonetheless, the transition seems to have occurred somewhat earlier than at sites to the north. The Devils Lake chronology is based on only four AMS dates. However, an AMS date of 6855 ± 110 B.P. at ~1.5 m above the base of the high-salinity interval suggests that an estimated age of 8000 B.P. is reasonable for the transition to high salinity. The initial transition from fresh to hyposaline conditions at Moon Lake culminates at ~8100 B.P. and correlates with the loss of elm and increase in *Ambrosia*.

At all three sites, periods of diatom-inferred salinity change correlate with pollen evidence of climatic warming and drying; thus, it is unlikely that the differences in the timing of increased salinity are a function solely of differential hydrological responses or changes in lake sensitivity (e.g. Campbell et al. 1994). However, the pattern and rate of salinity change is variable from site to site, and the differences undoubtedly do reflect hydrological variability. At Moon Lake, which has a small surface area and catchment, the transition from fresh to saline conditions is gradual; whereas at Devils Lake, with a very large catchment, large surface area, and complex hydrology, the transition is sharp. Differences in basin morphometry, particularly the lake surface area (e.g. Benson and Paillet 1989), and in initial hydrochemistry may have resulted in differential hydrochemical thresholds and variable patterns and rates of change. The Moon Lake diatom data show a continued and rapid increase in salinity to ~7300 B.P. and a correlative shift in the pollen record indicates increased aridity. Isotopic and trace-element analyses of ostracod shells at nearby Coldwater Lake (North Dakota) indicate a fresh to saline transition similar to the pattern at Moon Lake, with a gradual change beginning ~8800 B.P. and a maximum salinity by ~6500 B.P. (J. Xia unpubl. data).

Last (1990) concluded, based on geochemical and sedimentological evidence, that Ceylon Lake in southwestern Saskatchewan underwent a similar transition from a fresh, likely hydrologically open and relatively deep lake, to increasing salinity with an accompanying change from carbonate- to sulfate-dominated brine, likely associated with hydrological closure and lowering lake level ~7000–6000 B.P. Similarly, ~230 km to the east at Elk Lake pollen, diatom, and ostracode evidence suggests increasing aridity from ~7700 to 7000 B.P. (Bartlein and Whitlock 1993). Although the data suggest regional variability

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Fig. 7. Stratigraphic diagrams of diatom taxa (% in sample) driving the salinity reconstructions. A. Taxa > 15% relative abundance in at least two samples. B. Taxa between 5 and 15% relative abundance in at least two samples. Taxa are arranged by increasing salinity optima (*see* Fritz et al. 1993). Salinity categories based on Rawson and Moore (1944): fresh (<3 g liter^{-1}), hyposaline (3–10 g liter^{-1}), saline (10–100 g liter^{-1}). Zones are the same as in Fig. 5 and are described in the text.

Age (B.P.)	Moon	Medicine	Devils	Ceylon	Elk	Chappice	Waldsea	Manitoba	Redberry
0	gen. lower salinity (mean = 8.5 g liter⁻¹)	generally high salinity	high salinity; high salinity peaks	high lake stand	warm, moist	high lake stand; low water, high sal.	relatively deep stand; slightly lower stand	mild, moist	warm/dry; humid/cool; humid/cool
1000–2000	generally high salinity (mean = 18 g liter⁻¹)	high salinity	large fluctuations	high lake stand		large, relatively fresh lake	relatively deep stand	relatively high lake stand (present-day level)	relatively dry/warm
2000–3000	high salinity (mean = 29 g liter⁻¹)	low salinity	generally low salinity with episodes of higher salinity	increasing lake level	transition	high lake stand; gradually increasing lake level	shallow lake; high lake stand; increasing lake level	increasing lake level	
3000–5000	(poor diatom preservation)	gradually decreasing salinity		very low lake levels	generally warm, dry prairie period	highly saline shallow lake	hypersaline shallow lake	dry lake stand	
5000–6000	very high salinity (mean = 33 g liter⁻¹)	very high salinity	high salinity with fresh episodes			large fluctuations between relatively high stands and desiccation		warm with fluctuating moist and dry conditions	
6000–7000				high salinity with episodes of desiccation	forest to prairie transition				
7000–8000	low to high salinity transition		very high salinity		relatively cool, moist			relatively high lake stands with dry episodes	
8000–9000	fresh to low salinity transition		fresh to saline transition		spruce to pine transition				
9000–10,000		fresh to saline transition	fresh	fresh	cold, dry (spruce)			cool	
10,000–11,000	fresh	fresh							
Indicators	diatoms, pollen	diatoms, pollen, geochemical	diatoms	lithology, mineralogy	pollen	paleobotany, mineralogy	lithology, mineralogy	pollen, lithology, mineralogy	mineralogy, ¹⁸O, pigments
References	this paper	Radle et al. 1989; Juggins et al. 1994	Fritz et al. 1991; Fritz et al. 1994	Last 1990; Teller and Last 1990	Bartlein and Whitlock 1993	Vance et al. 1993	Last and Schweyen 1985; Teller and Last 1990	Teller and Last 1981; Last and Teller 1983	Stempvoort et al. 1993

Fig. 8. Summary of Holocene lake and climate conditions discussed in the text. Age is in radiocarbon years B.P. The lake names are indicated at the top of the respective columns. Summaries are based on the indicators below and follow the interpretation of the investigators cited.

in moisture availability, additional well-dated sites are needed to characterize the early Holocene climatic change within the region with confidence.

Several Dakota sites indicate a mid-Holocene arid period. However, the magnitude of fluctuations in effective moisture and the temporal extent of the period differs among sites. Both Moon Lake and Medicine Lake (Juggins et al. 1994) paleosalinity records inferred from fossil diatom assemblages suggest a relatively stable period of high salinity, indicating drier conditions. At Moon Lake this period spans from ~7300 to 4700 B.P. and at Medicine Lake from ~9200 to 4500 B.P. Devils Lake, on the other hand, shows a prolonged period of very high inferred salinity and thus an extreme deficit in effective moisture from ~8000 to 7000 B.P., followed by a period of high salinity from ~7000 to 5000 B.P., interrupted by episodes of freshwater indicating large fluctuations in effective moisture.

Several studies of Canadian prairie lakes also indicate a mid-Holocene period of high salinity and low effective moisture. Mineralogical and paleobotanical data from

Chappice Lake in the Cypress Hills of southeastern Alberta indicate a shallow lake from 7300 to 4400 B.P., with large fluctuations from relatively high stands to desiccation prior to 6000 B.P. (Vance et al. 1993). The Ceylon Lake record indicates high-salinity conditions with episodes of desiccation from ~8000 to 6000 B.P. and continued low lake levels until 4000 B.P. (Last 1990; Teller and Last 1990). Although the Waldsea Lake record extends only to 5000 B.P., lithological and mineralogical evidence indicates a shallow lake from ~5000 to 4000 B.P. (Last and Slezak 1986; Teller and Last 1990). Lithological, mineralogical, and chemical analyses in Lake Manitoba indicate a period of fluctuating wet and dry conditions from 9200 to 4500 B.P., however, climatic interpretation is complicated by the effects of isostatic rebound (Teller and Last 1981). In addition, in central Alberta an early Holocene arid period occurs from ~12,000 to 6000 B.P. (Hickman and Schweger 1993), with many paleolimnological indications of lower lake levels and increased salinity from ~8000 to 3000 B.P. (Schweger and Hickman 1989).

The sources of variability in pattern among these sites in the Great Plains are numerous. Much of the variability in timing is undoubtedly a function of poor chronological control attributable to a small number of radiocarbon dates or to problems associated with radiocarbon dates from bulk sediment, such as the hard-water effect. Differences in the hydrological sensitivity of lake basins to climatic change and in the responses of varied paleoecological proxies to both limnological and climatic change may also produce variability in timing, magnitude, and rate of response. Without good dating control, multiproxy analyses, and hydrological evaluation of each site, it is impossible to draw conclusions about century-scale moisture variability within the region from the available data. However, the data do confirm the pattern of a drier mid-Holocene in the northern Great Plains, and in most cases the period from ~8000 to 4000 B.P. is comparatively dry, suggesting regionally recurring drought.

Regional droughts in the Great Plains have occurred when dry Pacific air dominates over tropical and Arctic air flow as a result of stronger westerly zonal flow (Bryson 1966). Changes in the relative importance of these three air masses are believed to have caused shifts in the major ecotones bordering the Great Plains grassland (Bryson 1966; Bryson et al. 1970; Ritchie 1976; Bartlein et al. 1984), including the well-documented expansion and contraction of the eastern prairie-forest border (Winter 1962; Wright et al. 1963; McAndrews 1966; Webb et al. 1983) and the coincident shift from temperate grassland (tall grass) to more xerophytic grassland (short grass) prairie (Barnosky et al. 1987). General circulation patterns enhanced by the last Milankovitch insolation maximum, modulated by periglacial effects of the waning ice sheet, in large part can explain these patterns (Wright 1992).

Regional patterns of climatic change: Late Holocene—
Many sites in the northern Great Plains indicate a climatic change from a more arid mid-Holocene (low effective moisture) to one of increasing effective moisture ~4000 B.P., which continues until ~2000 B.P. Medicine Lake shows gradually decreasing salinity from 30 to 5 g liter^{-1} between 4500 and 3000 B.P. and remains low until ~2500 B.P. (Juggins et al. 1994). Devils Lake also indicates a generally low salinity (1–10 g liter^{-1}) with short episodes of slightly higher salinity between 5000 and 2500 B.P. (Fritz et al. 1991). Waldsea and Ceylon lake paleorecords show increasing lake levels ~4000 B.P., culminating in deep-lake stands by ~3000 and ~2000 B.P., respectively (Last and Schweyen 1985; Teller and Last 1990). Similarly, evidence exists for a gradual increase in effective moisture from 4400 to 2600 B.P. at Chappice Lake (Vance et al. 1993) and from ~4000 to 3200 B.P. farther east at Elk Lake (Bartlein and Whitlock 1993). Some sites (e.g. Medicine, Redberry, and Waldsea Lakes) show an interruption of the wet climatic regime, with a period of high salinity and decreased effective moisture from ~2500 to 1500 B.P. at Medicine (Juggins et al. 1994) and Redberry (Stempvoort et al. 1993) and from ~2800 to 2000 B.P. at Waldsea (Last and Schweyen 1985).

In contrast, the Moon Lake record indicates a relatively high-salinity period from 4700 to 2200 B.P. (Fig. 5, period 3), indicating a relatively dry climate. Interpretation, however, may be biased by the generally poor preservation of diatoms during this period and the preservation of heavily silicified saline taxa, which may not adequately represent the actual community composition of this entire period (as discussed earlier). However, *Ruppia* appears again between ~3000 and 1000 B.P., which corroborates the inference of an arid climate. Just south of Moon Lake at Coldwater Lake the stratigraphic record (J. Xia unpubl. data) also suggests continued drought through ~2000 B.P. As in the previous periods there are broad similarities in inferred climate from many of the sites; however, differences also exist among sites suggesting local-scale variability in effective moisture.

With the establishment of the climatic regime of the past 2,200 yr, the Moon Lake record shows generally high salinity with small fluctuations between ~2200 and 750 B.P. (mean, 18 g liter^{-1}) and generally lower salinity (mean, 8.5 g liter^{-1}) between ~750 and 100 B.P. Reconstructions from Medicine Lake show stable high salinity (15 g liter^{-1}) with small fluctuations for the past ~1,500 yr (Juggins et al. 1994), whereas Devils Lake indicates large fluctuations over the same period with high salinity peaks (>30 g liter^{-1}) at ~1000 and ~600 B.P. (Fritz et al. 1991). Low lake level at ~1000 B.P. is evident in several other sites, including Chappice Lake (~1000–600 B.P.) (Vance et al. 1993), Waldsea Lake (~1000–700 B.P.) (Last and Slezak 1988), Redberry Lake (~1000–900 B.P.) (Stempvoort et al. 1993), and Deadmoose Lake (Last and Slezak 1986, 1988). These low lake stands roughly correspond to the Medieval Warm Period (MWP), ~950–750 B.P. In contrast, Ceylon Lake indicates a high lake level at this time (Teller and Last 1990). Evidence at some sites indicates an arid climate in the northern Great Plains during the Little Ice Age (LIA), ~500–100 B.P., including Devils Lake (Fritz et al. 1994) and Redberry Lake, which indicates a decrease in effective moisture from ~500 to 200 B.P. (Stempvoort et al. 1993). In contrast, there is evidence of a high lake stand at Chappice Lake from ~600 to 100 B.P. (Vance et al. 1993), suggesting wetter conditions during the LIA interval.

In calendar years (Fig. 9), the salinity of Moon Lake decreased from ~2300 to 1300 B.P.; it then rose to a peak at ~850 B.P., which is coeval with the MWP (950–750 B.P.) (Lamb 1977). The salinity then decreased rapidly from ~700 to ~550 B.P. The onset of the LIA is clearly evident in the salinity reconstruction, with the lowest salinity since the late glacial-early Holocene. The details in the salinity record from the end of the MWP to the present correlate remarkably well with the record of fire frequency from northwestern Minnesota, ~230 km to the east. Clark (1988, 1990) reconstructed fire frequency from charcoal in annually laminated lake sediments for the past 750 varve years. Fire frequency was reduced during periods 710–510 B.P. and since 350 B.P., whereas it was higher during the intervening interval 510–350 B.P. The two periods of reduced fire frequency are contemporaneous with the two freshwater peaks during the LIA at Moon Lake, and the intervening period of greater fire
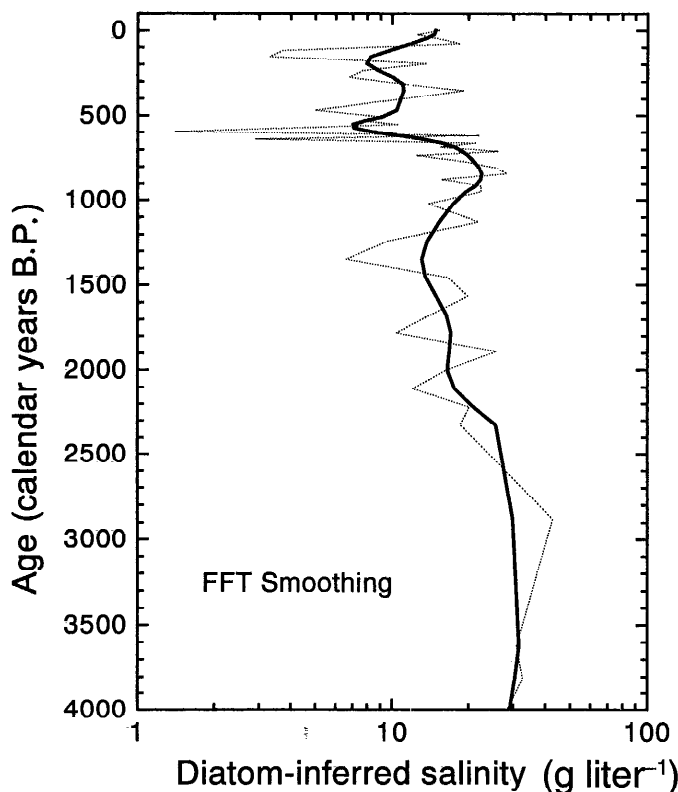
Laird et al.

Fig. 9. Diatom-inferred salinity for the past 4000 B.P. Dotted line shows the original salinity reconstruction; solid line shows the data smoothed with a fast Fourier transform (FFT) filter.

frequency matches the time of more saline conditions. Thus, these records from eastern North Dakota and adjacent Minnesota from the grassland and forest biomes indicate drought during the MWP and generally moist or cool intervals during the LIA. Climate, however, was variable during both of these periods and although the LIA contains intervals fresher than any in the previous ~8,000 yr, it also has intervals of drought equal to those of the 20th century Dust Bowl.

In the forested region east of Moon Lake, major vegetation changes occurred during the LIA. Forest invaded prairie on outwash plains in north-central Minnesota (Almendinger 1992), and the mixed deciduous "Big Woods" developed from oak woodland in south-central Minnesota (Grimm 1983). Farther east in Wisconsin, quantitative analysis of pollen data from annually laminated lakes indicates forest response to MWP and LIA climatic changes (Gajewski 1987, 1988). The pollen record from Moon Lake, however, shows little evidence of these changes. Either the prairie vegetation did not respond to these changes, or they are not visible palynologically. Unfortunately, the grasses are distinguishable only to family level, and important changes in the populations of prairie grasses might not be evident in the pollen record. Thus, more well-dated high-resolution analyses of sensitive climatic proxies are needed to confidently characterize the regional patterns of climate during the MWP and LIA intervals in the northern Great Plains.

The diatom and pollen proxies show varying sensitivities to climatic change during different time periods. During the middle Holocene when Moon Lake was hypersaline, a species-poor diatom assemblage did not clearly respond to varying salinities, whereas the pollen did record changing water levels. However, during the late Holocene the diatoms show more sensitivity to change than the pollen. Together, the diatom and pollen proxies provide a much more detailed picture of climatic change.

Future research

Although broad-scale patterns of climatic change during the Holocene in the northern Great Plains are evident, variability in inferred effective moisture exists amongst sites. This variability may be a function of poor chronological control, differences in hydrological sensitivity of individual lake basins, varied responses of the different paleolimnological proxies to climatic change, or local-scale climatic variability. Current research in the northern Great Plains includes paleolimnological analyses of several sites with good chronological control in which comparison of modern climatic records to the various proxy records will provide information on the hydrological sensitivity of each lake and insight into the rate and magnitude of response of the different proxies to climatic change. Multiple proxies (i.e. diatom, ostracod, isotopes, and pollen) are being analyzed at century scale for the entire Holocene and at subdecadal resolution for the last 2,000 yr to generate a more regional picture of climatic patterns, particularly drought frequency. Diatom analysis of additional surface samples from freshwater lakes in this region are being added to the modern-day training set to improve quantitative estimates of salinity optima. At some point, we hope to augment the surface-sample analyses with physiological and ecological studies of the timing and controls of diatom blooms, which may allow us to decipher changes in seasonal patterns of lake response to climatic change.

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