A multi-proxy limnologic record of rapid early-Holocene hydrologic change on the northern Great Plains, southwestern Saskatchewan, Canada

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Abstract: Clearwater Lake, Saskatchewan, is a relatively shallow, topographically closed, perennial lake centrally located in the Palliser Triangle region of western Canada. Despite its present-day small size, Clearwater Lake is one of only a few lacustrine basins in this vast 400 000 km^2 region of the northern Great Plains whose stratigraphic record extends to the early Holocene. Multiple proxy indicators are used to interpret the early-Holocene limnologic and hydrologic conditions in the basin. Changes in solute chemistry and concentration are reflected by siliceous microfossils and the endogenic and authigenic mineralogy of the sediment. The diatoms also provide evidence of fluctuations in water depth and water column stratification. Plant macrofossil assemblages, sediment texture and bedding features, and detrital mineralogy help to reconstruct drainage basin conditions and water level changes. Stable O- and C-isotopic composition of inorganically precipitated, endogenic carbonate minerals provide insight into the hydrologic budget of the basin and the residence time of the water. An early-Holocene shallow, clastic-dominated, freshwater lake, probably surrounded by boreal forest type vegetation, was followed by a mesosaline to hypersaline phase between 9800 and 8600 BP. A fresh to hyposaline lake with an open basin hydrology returned after 8600 BP and persisted until at least 6600 BP except for a short episode of higher salinity, closed-basin evaporitic conditions about 8200–7700 BP.

Key words: Palaeolimnology, lithostratigraphy, mineralogy, diatoms, palaeohydrology, plant macrofossils, stable isotopes, aragonite, evaporites, early Holocene, northern Great Plains, Canada.

Introduction

In order to gain a better appreciation of how future global change may affect the strategically important agricultural heartland of Canada, a considerable effort has been made in recent years to decipher past environmental changes that have occurred in the southern reaches of the Canadian prairie provinces (Lemmen, 1996; Lemmen *et al.,* 1993). Sediments preserved in the many and diverse lakes in this vast area of western Canada offer an excellent opportunity to unravel the history of local hydrologic systems. Ultimately, when regional hydrologic fluctuations are well defined, a history of climatic variability (the defining characteristic of regional climate), may be developed. Lakes that occupy topographically closed basins are particularly responsive to even minor changes in regional and local hydrology, short- and long-term climatic fluctuations, and watershed weathering characteristics. Such lakes have been targeted for detailed multidisciplinary stratigraphic investigations (Vance, 1997; Vance *et al.*, 1995; Vance and Last, 1994).

Notwithstanding this increased attention by palaeolimnologists, to date only three of the extant lakes in the northern Great Plains of Canada have provided complete, uninterrupted Holocene sequences. Indeed, as discussed elsewhere (e.g., Yansa, 1995;

Geological Survey of Canada Contribution No. 1997178; Palliser Triangle Global Change Contribution No. 36.

Last, 1995; Schweger and Hickman, 1989), very few lacustrine stratigraphic sections in this region extend to the early Holocene. The early Holocene was clearly a time of dramatic and rapid climatic and hydrologic changes in the northern Great Plains. Maximum postglacial warmth and aridity were established very early in the Holocene (10 000–9000 BP) at the northwestern margin of the region in central Alberta (Schweger and Hickman, 1989; Hickman *et al.*, 1984). By 8000 BP, the same conditions prevailed on the eastern margin of the plains (Curry, 1997; Last *et al.*, 1994; Teller and Last, 1981). In both these peripheral areas the early-Holocene warm and dry conditions gave way to a somewhat cooler and moister climate and more positive hydrologic budgets after about 8000 BP. However, relatively little information is available on the nature and timing of these early-Holocene climate changes in the central part of the northern Great Plains. The objective of this paper is to present the results of multidisciplinary palaeolimnological studies of a well-dated early-Holocene sediment core from Clearwater Lake, which is located in the midst of the most arid part of the northern Great Plains known as the Palliser Triangle.

The modern lake

Clearwater Lake, located about 75 km north of Swift Current in southern Saskatchewan (Figure 1), is one of many small, closed-basin lakes situated on the Missouri Coteau. The Missouri Coteau, a major geomorphic feature of the northern Great Plains, is a 50–100-km-wide, discontinuous band of knob and kettle topography (hummocky moraine) that extends for over 1200 km from South Dakota into west-central Saskatchewan.

The climate of the northern Great Plains is continental subhumid, and is characterized by very cold winters, short, but warm summers, low precipitation, and high evaporation (Environment Canada, 1993). Mean annual temperature in the vicinity of Clearwater Lake is about 1.5°C, with mean January temperature −15°C and mean July temperature +20°C. With an average annual precipitation of only 380 mm and an annual precipitation to potential evapotranspiration ratio of less than 0.6, Clearwater Lake is located in the most arid part of the region. Although most of the land around Clearwater Lake is cropped or used for grazing, the prevailing vegetation prior to agricultural disturbance was mixed grass prairie (Risser *et al.*, 1981). A small regional park and seasonal-use cottages were developed on the south and west side of the lake during the 1920s, and these remain a popular recreation site today, despite declining water quality and reduced fish populations in the last five years.

Clearwater Lake is small (0.5 km^2) and relatively shallow (mean depth: 2.6 m; maximum depth: 9.3 m). There are no permanent streams entering the lake, but several ephemeral creeks drain a small watershed of about 16 km^2. The discharge of these creeks has not been quantified but is assumed to be minor relative to the

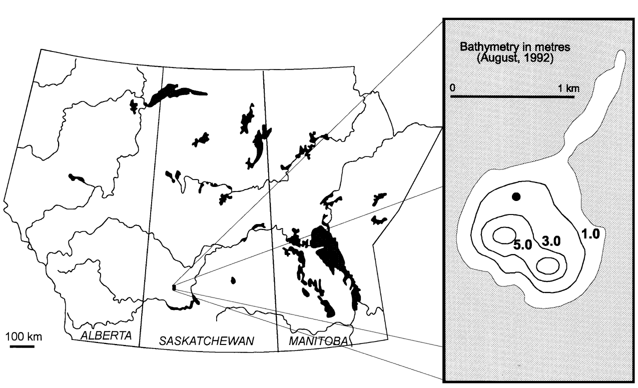

Figure 1 Map showing the location of Clearwater Lake in western Canada. Inset shows the bathymetry of the lake and the location of the core.

contribution of groundwater into the lake basin. Although details of groundwater dynamics at Clearwater Lake have not been investigated, the existence of this relatively low-salinity lake in a topographically closed basin in an area of high evaporation/precipitation implies that the lake today is a site of both groundwater recharge as well as discharge. This type of topographically closed but hydrologically open wetland setting is not uncommon in the hummocky terrain of the Missouri Coteau (Winter, 1992; 1995; Rozkowski, 1969).

Clearwater Lake is currently subsaline (TDS: 0.9–1.6 ppt), alkaline (pH: 8.4–9.25), and dominated by Mg^{2+}, Na^+, and HCO_3^- ions (Table 1 and Figure 2). The water is at all times of the year strongly supersaturated with respect to most carbonate minerals and, because of the very high Mg/Ca molal ratios (20–60), aragonite occurs in the modern sediment as a primary inorganic precipitate. Both magnesian calcite and protodolomite also occur in small amounts in the modern offshore sediments, probably as very early or penecontemporaneous diagenetic products generated at the sediment-water interface. In addition to these endogenic and authigenic carbonate minerals, which comprise about 25% of the modern offshore deposits, Clearwater Lake bottom sediment also consists of detrital carbonates (mainly dolomite), clay minerals, feldspars, and quartz. The organic matter content of the offshore sediment is about 15%.

Previous palaeoenvironmental research

Clearwater Lake has been the subject of previous palaeoenvironmental investigations. Mott (1973) conducted a palynological study of a 12.7-m-long core from the basin. Dating of this core proved inconclusive, however, and it was estimated that the entire recovered sequence extended to less than 6500 BP.

Prior to the detailed multidisciplinary analyses presented here, Cywinska and Delorme (1995) conducted a preliminary survey (each sample containing 5 cm of core was collected every 15 cm) of ostracode stratigraphy on the same Clearwater Lake core (CW2) that we have analysed. With the exception of samples below 720 cm, ostracodes were present in sufficient numbers to outline significant limnological changes, despite the large sample size and coarse sampling interval employed. The lower metre of sediment, although containing sparse numbers of ostracodes, indicates an early episode with low salinities and low water levels (<1.0 m), during which the lake experienced high rates of stream-derived detrital sedimentation. This was followed by a lengthy period of alternating saline (Na-SO_4 brine) and freshwater conditions recorded in the core between depths of 680 cm and 220 cm. The authors suggest that these repeated oscillations between saline and fresh water were in response to fluctuations between a relatively drier and a relatively wetter climate. Finally, ostracodes in the upper 2 m of this core indicate a sustained episode of low salinity and calcium bicarbonate enrichment, presumably due to cool and wet climatic conditions.

The most recent 400 years of sedimentation in Clearwater Lake was investigated by Leavitt *et al.* (1998). They conducted a multiproxy study (algal pigments, diatom valves, plant macrofossils, physical stratigraphy, and mineralogy) on a metre-long ^{210}Pb-dated core from the offshore area of the basin in an attempt to decipher recent climatic, hydrologic and limnological changes. Surprisingly, despite well-documented historical water-level fluctuations in the basin, and significant climatic changes in the region over the past millennium, there is little stratigraphic evidence of major changes in limnological or hydrologic conditions during the most recent 400 years.

Methods

This report is based on a 7.7-m-long sediment core (CW2) taken from Clearwater Lake in March, 1992, using a 3-inch diameter

Table 1 Hydrochemistry of Clearwater Lake (modified from Vance and Last, 1996a). All samples from surface water. Ion concentrations in mmol L^{-1} with mg L^{-1} in parentheses; na indicates not analysed; TDS is total dissolved solids (in mg L^{-1})

Property	July 1938	Nov 1966	Feb 1967	June 1967	Sept 1967	Aug 1994	Jan 1995
Ca^{2+}	0.22 (8.9)	0.99 (4)	0.18 (7)	0.27 (11)	0.19 (8)	0.14 (5.6)	20.3 (11.6)
Mg^{2+}	5.59 (136)	5.19 (126)	6.51 (158)	4.73 (115)	5.48 (133)	6.71 (163)	8.69 (211)
Na^{+}	3.65 (84)	2.96 (68)	3.31 (76)	2.35 (54)	2.74 (63)	3.82 (87.6)	5.32 (122)
K^{+}	na	0.63 (22)	0.64 (25)	0.48 (19)	0.54 (21)	0.71 (27.8)	0.07 (2.9)
$HCO_3^- + CO_3^{2-}$	8.87 (541)	9.79 (579)	12.03 (733)	9.28 (566)	8.81 (537)	9.64 (585)	13.94 (847)
SO_4^{2-}	1.39 (134)	1.65 (158)	1.94 (187)	1.45 (139)	1.72 (165)	2.59 (248)	3.25 (312)
Cl^{-}	0.58 (20.5)	0.68 (24)	0.82 (29)	0.54 (19)	0.65 (23)	0.92 (32.5)	1.16 (40.9)
TDS	934	999	1215	923	950	1159	1565
pH	na	8.75	8.55	8.4	9.0	9.25	8.98

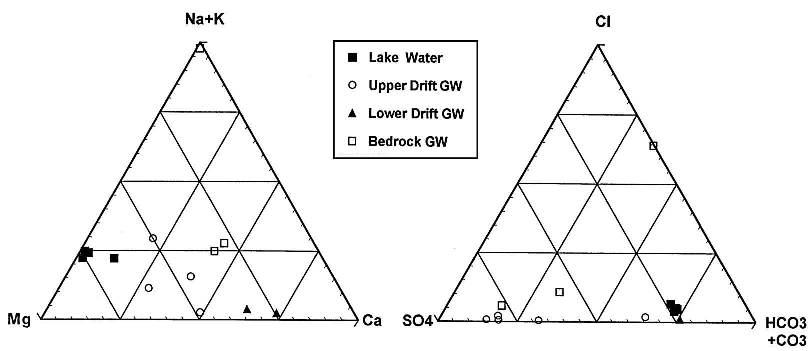

Figure 2 Triangular plots showing the composition of Clearwater Lake water and groundwater (in meq%; after Vance and Last, 1996a).

vibracorer (Vance and Last, 1994). Water depth at the core site was 1.5 m. After collection, the core was cut into sections approximately 2 m long, capped with plastic and fibre tape and stored in a cool room (4°C) until sampled. The core was split longitudinally, logged, photographed (visible and xradiography) and subsampled for various physical, geochemical and biological analyses.

Samples for moisture content, organic matter content, total carbonate content, particle size and mineralogy were taken at approximately 5-cm intervals. Where bedding or other sedimentary features dictated, sample spacing was reduced. Moisture, organic matter, and total carbonate mineral contents were evaluated by weight loss on heating to temperatures of 85°C, 500°C and 1000°C respectively (Dean, 1974). Particle-size spectra for each subsample were determined using the Galai CIS-1 automated laser-optical particle size analyser (Syvitski *et al.*, 1991; Aharonson *et al.*, 1986) after removal of organic matter by hydrogen peroxide treatment. Because of the high sulfate mineral content of many samples, a 1% solution of Triton X100 was used as the dispersion agent, rather than disodium hexametaphosphate. Particle-size statistics (mean, median, standard deviation) were calculated by the moment method (Allen, 1981). Duplicate samples were prepared and analysed every 25 samples and a standard run every 50 samples. These replicate analyses indicate precision of the particle size data is approximately $\pm 3\%$.

All samples analysed for mineralogy were air-dried at room temperature, disaggregated in a mortar and pestle, and passed through a 62.5-μm sieve. Bulk mineralogy and detailed carbonate and evaporite mineralogy were determined using standard x-ray diffraction techniques (Klug and Alexander, 1974; Müller, 1967). Mineral identification was aided by the use of an automated search-match computer program (Marquart, 1986). Percentages of the various minerals were estimated from the bulk mineral diffractograms using the intensity of the strongest peak for each mineral as outlined by Schultz (1964). Non-stoichiometry of the dolomite and calcite was determined by examining the displacement of the d_{104} peak on a detailed (slow) XRD scan (Goldsmith and Graf, 1958) and calculated according to Hardy and Tucker (1988). Duplicate samples were prepared and analysed for bulk mineralogy every 10 samples. These replicate analyses indicate precision of the mineralogical data is approximately $\pm 6\%$.

The $\delta^{13}C$ and $\delta^{18}O$ stable isotope composition of $CaCO_3$ (aragonite plus calcite) was measured for the <62.5 μm fraction of 25 samples (approximately 30-cm sample spacing) using a VG PRISM mass spectrometer and standard preparation methods (Drimmie *et al.*, 1992). Selective acid extraction provided data for $CaCO_3$ with minimal dolomite interference (Al-Aasm, 1990). Replicate analyses were within ± 0.4‰. Aragonite $\delta^{13}C$ and $\delta^{18}O$ values were calculated by correcting for detrital calcite as outlined in Van Stempvoort *et al.* (1993). The small isotopic fractionation differences between aragonite and magnesian calcite (Romanek *et al.*, 1992; Tarutani *et al.*, 1969) were ignored.

Sediment samples, generally spanning 5-cm core increments (amounting to between 75 and 100 ml of wet sediment per sample), were prepared for macrofossil analysis by first measuring sample volume and then gently washing the sediment with tap water through nested sieves (16, 60 and 80 mesh). Macrofossils picked from the organic residue on each screen were identified using a Leica MZ8 stereoscopic microscope and keys, drawings and photographs in Beijerinck (1976), Martin and Barkley (1961) and Montgomery (1977), and verified with the Geological Survey of Canada (GSC) macrofossil reference collection. Where preservation permitted, species were distinguished in the Chenopodiaceae and Betulaceae families, as well as within members of the genus *Potamogeton*. Once identifications were completed, the macrofossil content of each sample was converted to a standard 100 ml equivalent.

Sediment sample preparation for diatom analyses followed procedures outlined in Wilson *et al*. (1996), with the exception that samples were placed in a lower temperature water bath (65°C) for 8 h, and the dry coverslips were mounted on glass slides with Naphrax® mounting medium.

Diatoms were identified and enumerated using a Leitz DMRB microscope with differential interference contrast optics (1000 × oil immersion magnification, numerical aperture = 1.30). A minimum of 350 diatom valves was counted from individual samples spaced at 10-cm intervals, except where valves were virtually absent (from 350 to 310 cm, and 770 to 680 cm). Taxonomic identifications are based on (Cumming *et al*., 1995; Hakansson *et al*., 1993; Carvalho *et al*., 1995; Krammer and Lange-Bertalot, 1986; 1988; 1991).

Palaeosalinity reconstructions were generated from relative abundance data of the fossil diatom assemblages using a salinity transfer function developed from 208 lakes in southern British Columbia and 11 in the northern Great Plains (Wilson *et al*., 1996), and using the computer program WACALIB v. 3.3 (Line *et al*., 1994). The cyst:diatom ratio (Smol, 1985) was also considered as an indicator of palaeosalinity. To assess the reliability of individual reconstructed salinity values, a canonical correspondence analysis with axis 1 constrained to salinity was performed on the modern (those in transfer function) and fossil diatom samples, using the program CANOCO v. 3.1 (ter Braak, 1990). The squared residual distance of each fossil sample from the salinity axis was compared to the squared residual distances of the modern samples, such that fossil samples with residual distances equal to or larger than the extreme 5% of the training set were considered to have a very poor fit to the salinity axis (Birks *et al*., 1990). The estimated bootstrap standard error of prediction for each inferred salinity value was also plotted as a means of showing the reliability of the reconstruction.

Results

Chronostratigraphy

To minimize chances of dating redeposited organic matter, only well-preserved plant macrofossils from upland and shoreline species should be selected for accelerator mass spectrometry (AMS) radiocarbon dating (Table 2). Unfortunately, much of the macrofossil collection obtained from this near-shore coring site displays some evidence of reworking. Although the samples selected for dating represent the best-preserved macrofossils in the core CW2 macrofossil collection, the *Scirpus* cf. *americanus* seeds were not in pristine condition, as on one bristles had been removed from the style base (105–110 cm), and on others the apex tip was either broken or rounded (190–195, 345–350, 475–480, 585–590 cm). Although the *Picea* cf. *glauca* needles were complete, most were dry, friable and rust-coloured. Such characteristics raise the possibility that these samples were subaerially exposed and/or reworked.

It is evident from this sequence of ^{14}C dates that the stratigraphic record recovered from Clearwater Lake at this location in the basin is truncated and includes only early-Holocene sediments (Figure 3). Although there is no visually obvious unconformity within the top metre of core, the uppermost 20 cm of sediment is mineralogically distinctive and is interpreted to represent modern deposition. Using a second-order best fit line of depth versus age (Figure 3) the age of the sediment at 20-cm depth is 6600 ^{14}C yr BP and the base of the core (7.7 m depth) is 10 100 ^{14}C yr BP.

Core description

Core CW2 consists of massive to well-bedded, dark greenish-grey (Munsell: 5Y 2.5/1) to olive-grey (5Y 4/2), dark greyish-brown (2.5 YR4/2), and dark grey (2.5 Y 2.5/1) clayey silt. Lighter laminae (10 YR 4/1) and faint light-dark banding on a scale of several centimetres occur throughout the core, but the most well-defined laminae generally occur in the lower 3 m (Figure 4). One of the most distinctive structural features visible in the core is a pedogenic-like zone at 310–315 cm (Figure 5). This zone is characterized by an angular blocky to pelletal structure, a very sharp upper contact and gradational lower boundary, and a mottled and gleyed (5 GY4/1) colour.

Although overall there is little systematic change in particle size (mean size: 18 μm; average: 4% sand, 75% silt, 21% clay), the lower 150 cm and upper 260 cm of core CW2 are somewhat coarser grained (more silt-rich) than the rest of the section. Like the modern bottom sediments, organic matter content is relatively low, ranging from less than 1% to about 20% (average: 4.5%). Stratigraphically, the amount of organic matter is closely correlated with the silt content of the sediment and inversely related to clay content.

The core is composed of subequal proportions of detrital siliciclastics, clay minerals, carbonate minerals, and endogenic/authigenic carbonates and evaporites. As shown in Figure 6, the detrital fraction of the sediment dominates the sequence (average 68%), except for the interval between about 6.2 and 6.8 m which is almost entirely endogenic and/or authigenic in origin. Quartz (average: 24%) and feldspar minerals (19%) are the major detrital components, with smaller proportions of clays (15%), dolomite (8%), calcite (4%), and amphiboles ($<$1%). Generally, sodium-rich plagioclase is more abundant than potassic feldspars.

The fraction of the sediment that originated within the lake basin itself (i.e. endogenic and authigenic minerals) consists of a complex mixture of aragonite ($CaCO_3$), gypsum ($CaSO_4 \bullet 2H_20$), disordered, nonstoichiometric dolomite [$Ca_{0.x}Mg_{0.y}(CO_3)_2$], magnesian calcite, bloedite [$Na_2Mg(SO_4)_2 \bullet 4H_20$], thenardite ($Na_2SO_4$, or the hydrated species, mirabilite, $Na_2SO_4 \bullet 10H_20$), magnesite ($MgCO_3$), hydromagnesite [$Mg_5(CO_3)_4(OH)_2 \bullet 4H_20$], and pyrite ($FeS_2$). The magnesian calcite present in the core shows a narrow range of composition, from about 10 to 16 mol% $MgCO_3$. The disordered dolomite (protodolomite) has a similarly narrow composition, averaging about 17 mol% excess $CaCO_3$.

There are few clear stratigraphic trends in these mineralogical constituents over the entire length of the core. Variation in proportions of the various siliciclastic components (quartz, K-feldspar, plagioclase, amphiboles) are all significantly correlated with one another (linear correlation coefficients significant at 0.95 confidence level) and inversely correlated with the abundances of organic matter, aragonite and gypsum. Similarly, detrital calcite and dolomite covary and are negatively correlated with the endogenic carbonate minerals. The occurrence and abundance of the soluble evaporite minerals (gypsum, Na sulfates, Mg + Na sulfates) are, as expected, significantly correlated, but, except for gypsum, do not show either a positive or negative covariance with the detrital or other endogenic components.

Table 2 Clearwater Lake core CW2 AMS radiocarbon age determinations

Lab no.	Depth (cm)	Material[1]	^{14}C age (yrs BP)
CAMS[2]-19172	105–110	1 *Scirpus* cf. *americanus* seed	7320 ± 70
CAMS-17431	190–195	1 *Scirpus* cf. *americanus* seed	7310 ± 60
CAMS-17430	345–350	1 *Scirpus* cf. *americanus* seed	8840 ± 60
CAMS-17429	475–480	1 *Scirpus* cf. *americanus* seed	8930 ± 70
CAMS-8372	585–590	3 *Scirpus* cf. *americanus* seeds	9340 ± 70
CAMS-19171	730–735	2 *Picea* cf. *glauca* needles	9980 ± 70

[1]Assumed $\delta^{13}C$ of −25 for all samples.
[2]Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory.

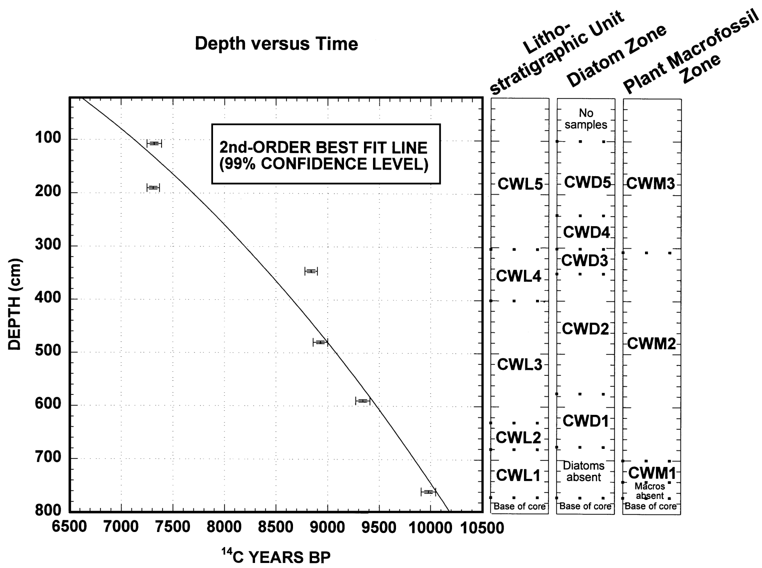

Figure 3 Best fit line of depth versus ^{14}C time in Clearwater Lake core CW2. Also shown are the positions of the various stratigraphic units identified in the core.

Lithostratigraphy

Based on the mineralogical composition, particle size characteristics, organic and moisture contents, and bedding features, this 7.7-m-long, early-Holocene lacustrine sequence recovered from Clearwater Lake can be subdivided into five lithostratigraphic units (Table 3; Figures 4, 6 and 7).

The lowermost unit penetrated (Lithostratigraphic unit CWL1), from the base of the core to 680 cm depth, is a firm, compact, structureless, siliciclastic-rich clayey silt with relatively high sand and low moisture and organic matter contents. In addition to the high proportions of detrital quartz and feldspars, this lower metre of section is also characterized by relatively high levels of protodolomite. The composition of this disordered carbonate is very consistent throughout the unit (66 mol% $CaCO_3$).

Sharply overlying this basal unit is a 50-cm-thick, finely laminated, organic-rich gypsite (CWL2). Aragonite and various soluble sodium and magnesium sulfate salts also occur in this thin evaporitic unit (Figure 7). These Na and Mg sulfates are most common at the base of the unit. In contrast to the lower unit, CWL2 contains very little disordered dolomite. Although minor, the siliciclastic and clay mineral content of CWL2 grades from relatively high at the base to very low at the top of the unit. Both the upper and lower contacts are sharp.

CWL3, extending from 630 cm to 400 cm depth in the core, is a finely laminated to indistinctly bedded, calcareous, clayey silt. The unit is distinguished from the underlying evaporitic CWL2 mainly by its finer grain size (higher clay content) and higher carbonate mineral content. The endogenic carbonates in CWL3 are dominated by aragonite and protodolomite with a much lower proportion of magnesian calcite relative to the underlying units (Figure 7). The aragonite content increases upward from about 10% at the base to over 60% at about 500 cm depth and then decreases to less than 10% near the top. Disordered dolomite similarly increases upward to 500 cm depth and decreases further upward in the unit. The total soluble salt content as well as organic matter both show a gradual but sporadic decrease upward. Although the content of gypsum and other soluble salts is lower in CWL3 relative to CWL2, nonetheless, there are numerous thin zones within the unit that are salt-rich (Figure 7).

The aragonite-rich CWL3 grades upward into an unusual siliciclastic unit, CWL4. The sediment from 400 cm to 310 cm depth is composed mainly of non-sandy, detrital quartz and feldspars with very low clay mineral content. Despite the paucity of clay minerals, CWL4 has the greatest proportion of clay-sized material. The unit contains no soluble salts and very low proportions of aragonite and organic matter. The lower contact with CWL3 is gradational but the upper contact of CWL4 is very sharp and marked by a distinctive dry, blocky structured, variegated pedogenic horizon at 310 cm (Figure 5).

Immediately overlying the pedogenic horizon developed into the top of CWL4 is the uppermost early-Holocene lithostratigraphic unit, CWL5. This unit is composed of relatively coarse-grained, structureless to faintly bedded and finely laminated, calcareous, clayey silt. Like the underlying CWL3, the endogenic carbonate content is dominated by aragonite, with very high levels (>80%) of this endogenic carbonate mineral occurring in a finely laminated zone at the base of the unit. Magnesian calcite is very low or absent except in several thin zones near the top of the section. Authigenic pyrite occurs sporadically in CWL5 and the organic matter content is relatively high. Gypsum content is also

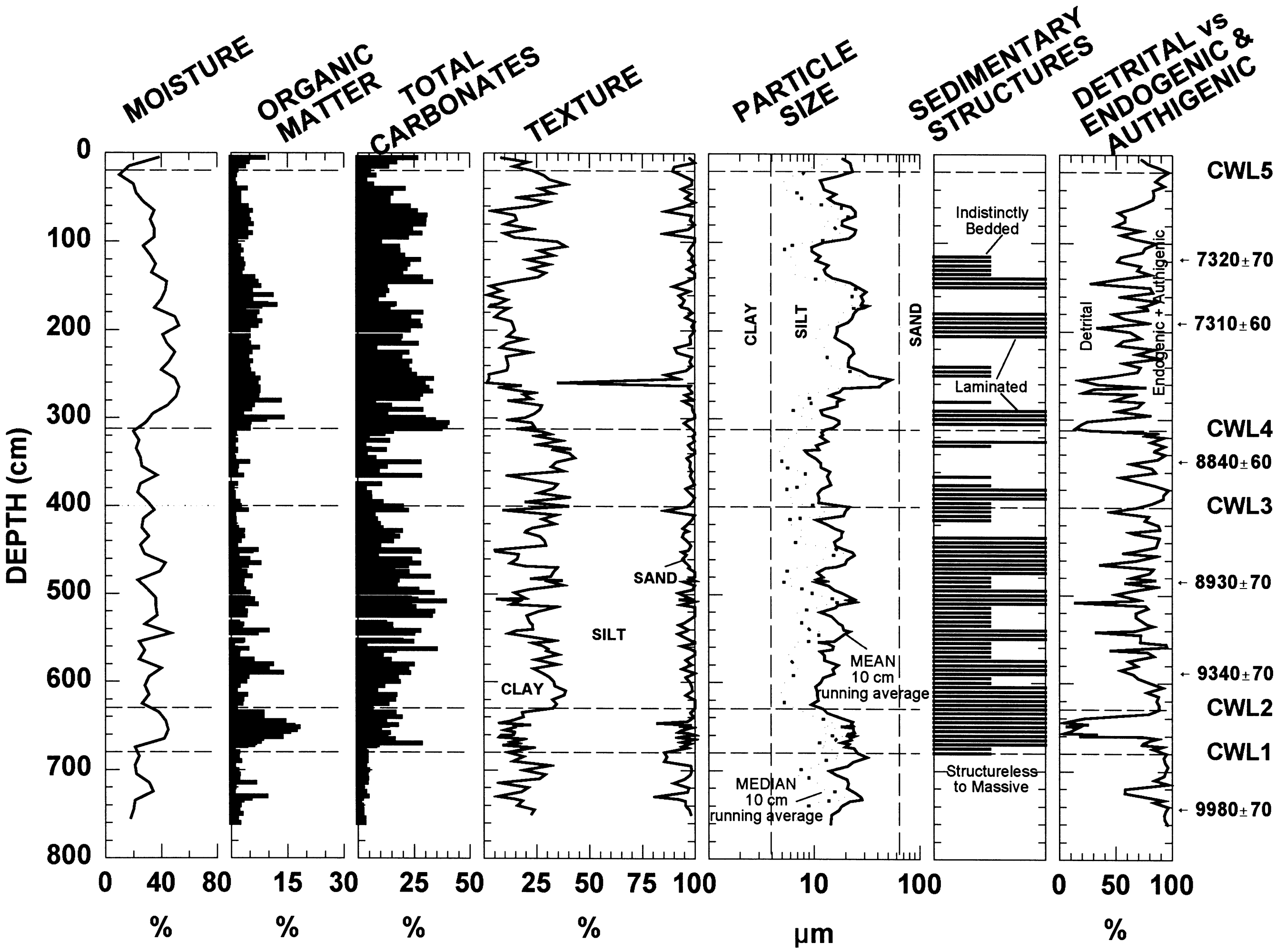

Figure 4 Stratigraphic variation in moisture, organic matter, and total carbonate mineral contents, texture (% sand, % silt, % clay), particle size (mean and median), sedimentary structures, and detrital versus endogenic + authigenic mineral components. Radiocarbon dates are shown on right with lithostratigraphic units (Table 3). Dashed lines indicate lithostratigraphic unit boundaries. The upper 20 cm of this core is modern sediment.

relatively high in the lower 50 cm of the unit but decreases upward in this unit to only sporadic occurrences of less than 10% above about 200 cm depth. In contrast to the generally decreasing upward trend of the various endogenic components within CWL5, quartz and feldspar minerals both increase upward in the unit. The feldspar weathering index, a measure of the intensity of chemical weathering of clastic sediments in the watershed, shows a gradual but irregular increasing upward trend in this unit (Figure 6). There are also significant fluctuations in the particle-size parameters within CWL5 (Figure 4). Although the unit is overall coarser than the underlying deposits, zones of finer-grained sediment occur at the base (310–260 cm) and in the upper metre of the unit (120–100 cm and 60–20 cm).

Stable carbon and oxygen isotopes of aragonite

The stratigraphic variation in stable isotope ratios of aragonite is shown in Figure 7. The $\delta^{18}O$ of this endogenic carbonate ranges from a low of −9.7‰ (PDB) to as high as −4.3‰ (mean = −7.8‰). The more negative $\delta^{18}O$ values are within the lowermost section of the core and in the upper two lithostratigraphic units (CWL4, CWL5), while the highest ratios (least negative) are in CWL2. The $\delta^{13}C$ data range from −1.4% to +3.4‰ (mean = +0.9‰), with the lowest values occurring in the basal unit (CWL1) and generally increasing values upward in the section.

Overall, the $\delta^{18}O$ and $\delta^{13}C$ values of aragonite exhibit a weak positive covariation. However, the two isotopes show a very strong positive correlation in lithostratigraphic units CWL2 and CWL3 and a strong inverse correlation in the basal unit. Above about 250 cm depth, there is relatively little variation in $\delta^{18}O$ whereas the $\delta^{13}C$ continues to increase upward in the core.

Macrofossils

The diverse macrofossil assemblage in core CW2 is subjectively divided into four zones, according to shifts in the concentration of the major constituents (Figure 8). Taxa are arranged on the diagram to approximate an ecological gradient from upland trees and shrubs (left) through shoreline taxa to aquatic taxa (right). Macrofossils were relatively abundant throughout the core, with the exception of the basal 32 cm (772–740 cm), where no macrofossils were recovered.

The lowermost macrofossil assemblage (CWM1; 740–700 cm) is distinguished by a paucity of aquatic taxa (only a single *Chara* oogonium was tallied) and a relative abundance of *Picea* cf. *glauca* (Moench) Voss needles, *Andromeda polifolia* L., *Rubus idaeus* L., *Arctostaphylos uva-ursi* L., and *Chenopodium rubrum* L. seeds. Occurrences of *Picea* and *Fragaria vesca* L. seeds are restricted to CWM1. Shoreline taxa representation is low throughout.

A relative abundance of *Chara* oogonia and *Scirpus* achenes distinguishes macrofossil zone CWM2 (700–310 cm). Cyperaceae and Chenopodiaceae seed concentrations are relatively high throughout CWM2. *Chenopodium salinum* Standl. and *Corispermum* seed concentrations peak in the middle portion of the zone (*c.* 600–400 cm). Betulaceae seed concentrations also attain peak representation in the middle portion of zone CWM2 (*c.* 575–400 cm). *Potamogeton pectinatus* L. representation is relatively high

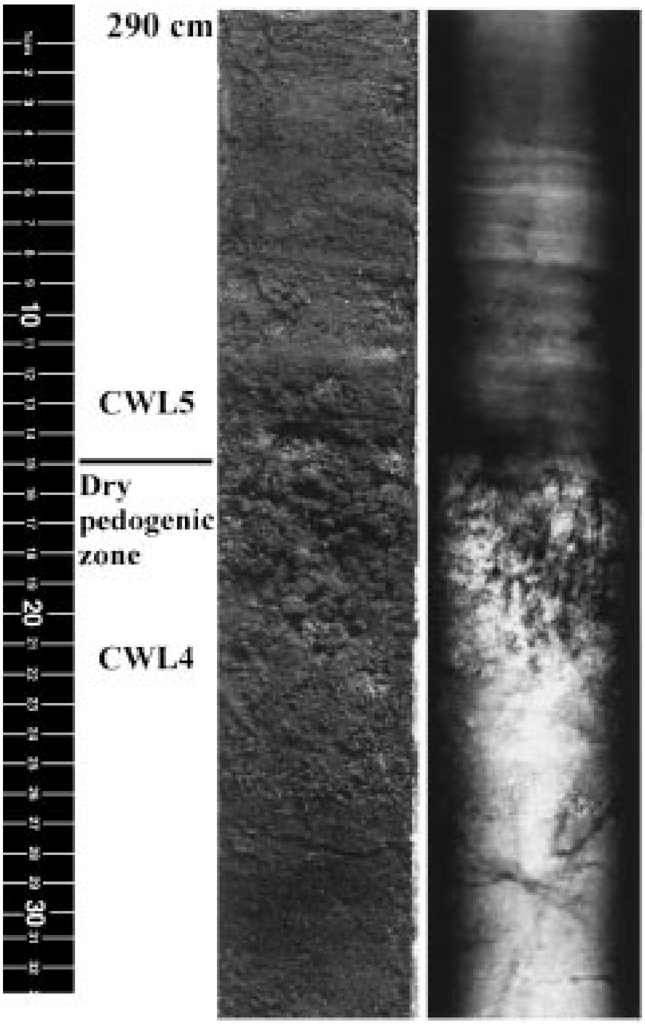

Figure 5 Light (left) and xradiograph (right) imagery of core CW2 showing the distinctive angular to blocky and pelletal structure of the pedogenic zone at the top of Lithostratigraphic Unit CWL4. Up is toward the top of the figure.

over roughly the same interval. *Ruppia* seeds appear only in CWM2, albeit sporadically in low concentrations in the upper half of the zone. Sporadic occurrences of *Juniperus*, *Symphoricarpus occidentalis* Hook. and *Rosa* are generally confined to CWM2. *Cornus stolonifera* Michx., *Viburnum*, *Euphorbia* and Gramineae seeds, as well as a single *Viburnum* fruit, were recovered only in CWM2.

Macrofossil deposition is generally low throughout CWM3 (310–30 cm). Low concentrations of Chenopodiaceae and Cyperaceae seeds occur on occasion. *Chara* oogonia are abundant throughout CWM3, and a consistently high percentage coated with carbonate. All other aquatic taxa are absent throughout the zone.

Relatively high concentrations of Chenopodiaceae, Cyperaceae, *Hippuris vulgaris* L., and *Potamogeton* spp. seeds and peak *Chara* oogonia concentrations (albeit with a reduced proportion, compared to CWM3, bearing a carbonate coating) are the defining characteristics of the uppermost macrofossil zone CWM4 (30–0 cm). Sporadic occurrences of *Juniperus*, *Rosa*, *Smilacina racemosa* (L.) Desf. and *Viola* further distinguish CWM4.

Diatom stratigraphy

The early-Holocene diatom record from Clearwater Lake has been subjectively divided into five zones (Figure 9). Although diatoms were absent or too sparse to count in the lower metre of the core, and in the interval 310–340 cm depth, diatom preservation throughout the remainder of the core was very good, with almost no evidence of diatom dissolution.

In diatom zone CWD1 (675–575 cm), taxa having a wide range of salinity optima occur together. The predominant freshwater/subsaline taxa include *Amphora libyca* Ehrenb., *Navicula oblonga* (Kütz.) Kütz., *Cocconeis placentula* var. *euglypta* (Ehrenb.) Grun., *Navicula* aff. *veneta*, *Navicula capitata* var. *hungarica* (Grun.) Ross and *Mastogloia smithii* Thwaites. The dominant taxa in the hyposaline-mesosaline range include *Cyclotella choctawhatcheeana* Prasad *et al*., *Chaetoceros muelleri* Lemm. (cysts) and *Navicula* aff. *rhyncocephala*. The saline taxa *Navicula bulnheimii* Grun. in V.H., *Brachysira aponina* Kütz and *Fragilaria famelica* (Kütz) Lange-Bertalot, present at extremely low abundance in this zone, show a marked increase to ~10% relative abundance at 600 cm depth.

CWD2 (575–345 cm) has a virtual absence of freshwater diatoms, and is dominated by five meso- to hyposaline taxa: *C. choctawhatcheeana*, *B. aponina*, *Amphora* aff. *coffeaeformis* (= *Amphora* sp. 6 PISCES; Wilson *et al*., 1996), *Achnanthes thermalis* (Rabh.) Schoenf., and *Navicula* aff. *salinicola*. Other less abundant saline and subsaline taxa that consistently occur throughout the zone include *Amphora coffeaeformis* (Agardh) Kütz; *Cymbella pusilla* Grun. ex A. Schmidt, *Denticula valida* (Ped.) Grun. in V.H., *M. smithii* and *Epithemia argus* (Ehrenb.) Kütz.

CWD3 (345–305 cm) is defined by a near absence of diatoms. Overlying this barren interval is zone CWD4 (305–235 cm), where *N.* aff. *salinicola* resumes abundance, along with *M. smithii*, *D. valida* and *E. argus*, the latter two appearing in high abundance for the first time. Saline and freshwater taxa that were present together in zone CWD1 (e.g., *C. choctawhatcheeana*, *Ch. muelleri*, *Navicula* aff. *rhyncocephala*, *A. libyca* and *N. oblonga*) also co-occur in the early stages of CWD4. Other taxa that appear in abundance for the first time include *Rhoicosphenia abbreviata* (Agardh) Lange-Bertalot, *Cocconeis placentula* var. *lineata* (Ehrenb.) V.H., *Gomphonema angustum* Agardh and *Achnanthes minutissima* Kütz., all of which have salinity optima in the freshwater to subsaline range.

The dominant taxa in CWD5 (235–100 cm) include mainly freshwater forms such as *Co. placentula* var. *lineata*, *G. angustum*, *A. minutissima*, *Fragilaria crotonensis* Kitt. and *Cymbella microcephala* Grun. in V.H. The subsaline taxa *M. smithii* and *Cymbella* sp. 1 PISCES (Cumming *et al*., 1995) persist at moderate abundances throughout the zone, along with some hyposaline taxa present at lower abundances.

Discussion

Clearwater Lake probably originated as an ice block meltout basin or kettle hole associated with deglaciation and the formation of the hummocky terrain of the Missouri Coteau about 11 000 years ago. The 7.7 m of sediment recovered at this relatively shallow-water, near-shore site records dramatic fluctuations in lake-water chemistry, drainage basin conditions and limnological setting of Clearwater Lake during the early Holocene (Table 4). Although a comparable stratigraphic sequence has not yet been retrieved from a more basinal offshore area of the lake, the tightly constrained chronology based on AMS ^{14}C dates (Table 2) clearly demonstrates that the basin contained water for much of the early Holocene.

The structureless, black muds of the basal lithostratigraphic unit (CWL1), deposited between about 10 000 BP and 9800 BP, represent sedimentation in a relatively shallow non-productive lake. High sedimentation rates (0.45 cm yr^{-1}), combined with the predominance of siliciclastic and detrital carbonate minerals and the absence of soluble salts, indicates that this earliest Holocene lake

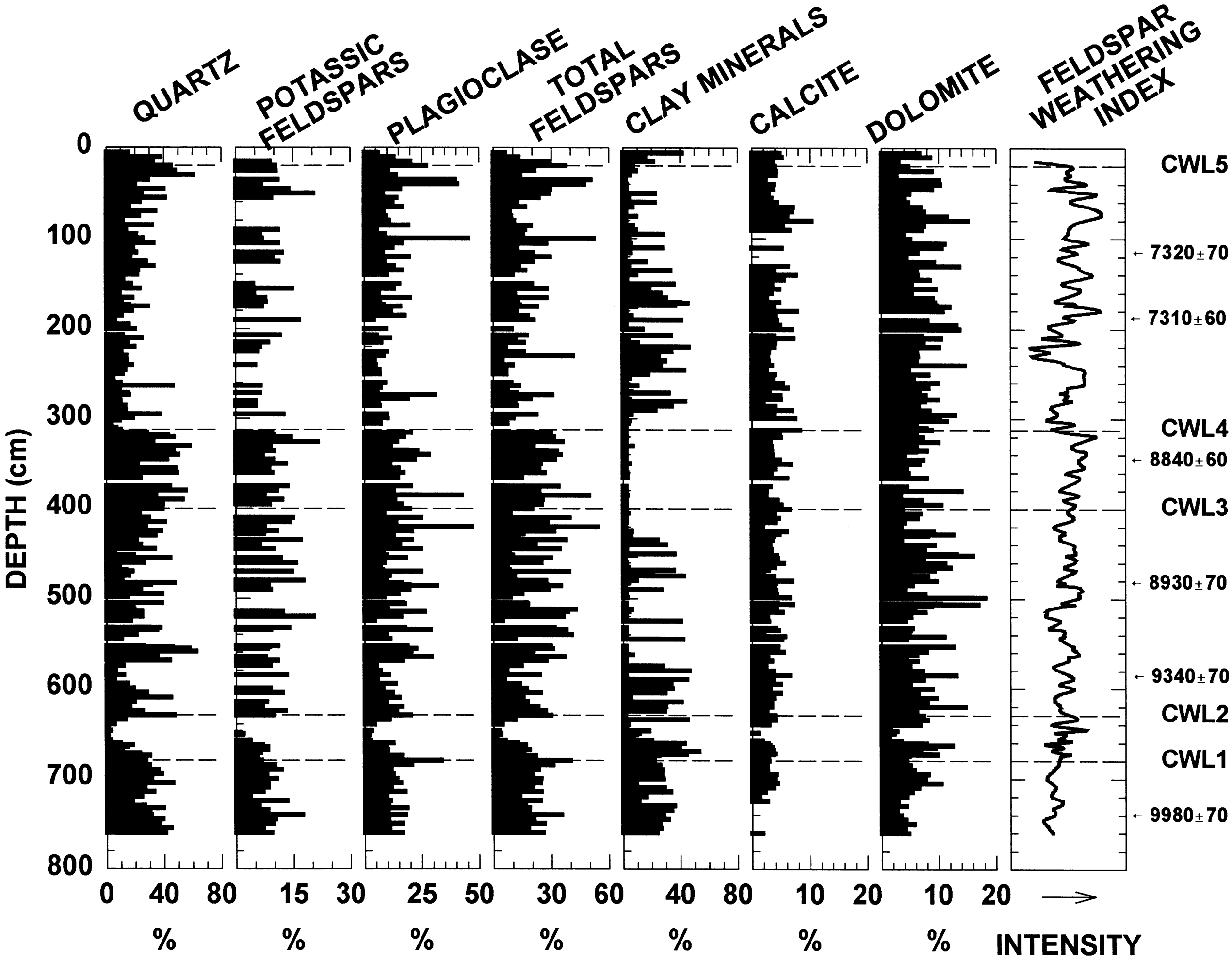

Figure 6 Stratigraphic variation in detrital quartz, feldspar minerals, total clay minerals, calcite, dolomite, and feldspar mineral weathering index. The feldspar weathering index is a ratio of the relatively chemically stable potassic feldspars to the less stable plagioclase feldspars. This ratio increases (to the right in the figure) under more intense chemical weathering conditions. Radiocarbon dates (Table 2) and lithostratigraphic units (Table 3) are shown on right. The upper 20 cm of this core is modern sediment.

was relatively fresh, turbid, and dominated by stream and sheet flood influx of clastic sediments derived from the newly deposited, unstabilized tills of the watershed. Although there is little in the sediments to indicate water depth, the relatively coarse grain size and the absence of lamination implies shallow water and/or disruption of bedding by waves or bioturbation. The inverse covariance of the stable oxygen and carbon ratios of the small amount of endogenic magnesian calcite present in these basal sediments and the relatively low $\delta^{18}O$ values suggest an open hydrologic system and non-evolved (non-evaporated) waters.

The occurrence and abundance of disordered dolomite in CWL1 is puzzling. Although the precise thermodynamic and geochemical conditions responsible for primary and/or penecontemporaneous dolomite formation are unknown (Warren, 1989; Morrow, 1982), lacustrine protodolomite is generally associated with elevated carbonate alkalinities ($> \sim 5000$ mg l^{-1}), high Mg^{2+}/Ca^{2+} and HCO_3^-/Ca^{2+} molal ratios ($> \sim 20$), and moderate to high ionic strengths (Last, 1990a; Folk and Land, 1975; Müller *et al.*, 1972). There is also some evidence that dolomite formation is enhanced in solutions having low sulfate ion activities (Baker and Kastner, 1981). This sulfate inhibition of dolomite precipitation is still controversial, however (Burton *et al.*, 1992; Morrow and Ricketts, 1986; 1988). It seems unlikely that limnological and geochemical conditions in this early lake were conducive to the formation of the relatively high amount of primary or penecontemporaneous dolomite found in CWL1. A more likely explanation is that this nonstoichiometric carbonate mineral is a diagenetic product, probably associated with a later, more saline, evaporitic episode in the basin. This postdepositional origin is supported by the occurrence of small nodules and rounded intrasedimentary clasts visible within the unit on xradiographs and by the remarkably constant composition of the nonstoichiometric material.

Unfortunately, the absence of diatoms in the sediment deposited before 9700 BP (Figure 9; Table 4) and the lack of macrofossils in the lower 30 cm of the section (Figure 8; Table 4) preclude a more detailed reconstruction of this early episode in the Clearwater basin. Macrofossils do appear just after 10 000 BP and provide important information about the watershed conditions at that time. The presence of *Picea* cf. *glauca* remains (the poor condition of the needles precludes positive species identification) in CWM1 indicates that this early unit was deposited at or near the time that *Picea* was migrating northward from glacial refugia to the south as climate ameliorated during deglaciation (Ritchie and MacDonald, 1986). *P. glauca* macrofossils, including logs, have been recovered from excavations at the Andrews dugout near Moose Jaw, Saskatchewan (Yansa, 1995) and *P. glauca* needles and cones have been identified at the Hafichuk site (Ritchie and De Vries, 1964), both situated on the eastern flank of the Missouri Coteau in south-central Saskatchewan, some 300 km east of Clearwater Lake. In both these cases the *P. glauca* remains have been dated at greater than 10 200 BP. Although local conditions may have prolonged *P. glauca* habitation in the Clearwater Lake basin, the generally poor condition of the macrofossils and the slightly later dates compared to the other sites indicates that they

Table 3 Summary of average mineralogical, geochemical and particle-size characteristics of lithostratigraphic units. nd = not determined; np = not present

Characteristic	Lithostratigraphic unit					
	CWL1	CWL2	CWL3	CWL4	CWL5	Core average
% Total clay minerals	25.9	23.3	12.4	1.5	15.8	14.9
% Quartz	33.1	9.1	25.8	36.7	19.8	23.7
% Total feldspar minerals	22.3	9.8	22.4	24.6	16.1	19.0
% Plagioclase	13.7	7.0	15.0	16.0	11.1	12.7
% K-feldspars	8.6	2.8	7.4	8.6	5.0	6.3
% Amphibole minerals	0.1	np	0.5	0.5	0.2	0.3
% Total calcite	3.0	2.9	4.8	4.7	5.2	4.6
% Low-Mg calcite	1.7	1.8	3.9	3.7	4.0	3.5
% Magnesian calcite	1.3	1.2	0.5	1.0	1.1	1.0
Mol% $MgCO_3$ in magnesian calcite	13.9	11.7	11.4	11.5	10.7	11.7
% Total dolomite	8.6	6.2	10.5	11.2	10.3	9.6
% Well-ordered dolomite	5.1	5.1	7.8	7.7	8.5	7.5
% Disordered dolomite	3.5	0.6	2.7	3.5	1.9	2.4
Mol% $CaCO_3$ in protodolomite	65.7	64.7	68.3	64.9	66.1	66.7
% Magnesite	np	np	0.2	0.2	0.1	0.1
% Aragonite	np	4.5	18.8	18.7	22.8	16.8
$\delta^{18}O$ of endogenic $CaCO_3$ (ppt PDB)	–8.4	–5.5	–7.1	–8.9	–9.0	–7.8
$\delta^{13}C$ of endogenic $CaCO_3$ (ppt PDB)	–1.1	1.6	0.6	0.4	1.7	0.9
% Gypsum	2.1	43.8	6.4	np	8.6	10.2
% Mg + Na sulfate salts	np	4.6	0.7	np	0.4	0.9
% Na sulfate salts	np	0.4	0.2	np	0.1	0.1
% Pyrite	np	np	np	np	0.6	0.2
% Detrital fraction	89.2	49.2	72.1	74.1	64.1	68.5
% Endogenic + authigenic fraction	10.8	50.8	27.9	25.9	35.9	31.5
% Moisture	23.9	35.5	31.3	26.7	38.7	32.7
% Organic matter	2.6	9.8	3.6	1.3	5.2	4.5
Mean particle size (μm)	20.8	21.7	15.7	11.9	20.6	18.1
Standard deviation of mean size	23.7	23.3	18.5	15.2	19.2	19.4
Median particle size (μm)	11.3	14.2	8.7	6.2	14.9	11.4
% Sand	6.1	5.0	3.0	1.6	4.5	3.9
% Silt	74.6	80.8	71.5	67.5	78.5	74.9
% Clay	19.3	14.2	25.5	30.8	17.0	21.3
Sedimentation rate[1] (cm 100 yr^{-1})	45	25	25	19	17	22

[1]Calculated on the basis of a best fit line of depth versus time (Figure 3) using the six ^{14}C dates and assuming no interruption in deposition.

may have been redeposited. However, the presence of pristine *Andromeda polifolia* seeds, as well as *Arctostaphylos uva-ursi*, *Fragaria vesca* and *Rubus idaeus* seeds, lends a distinctly boreal character to this macrofossil assemblage. *A. polifolia* occurs today in circumpolar peat bogs (Moss, 1983) and boreal swamps and muskegs (Budd, 1979), over 200 km north of Clearwater Lake. Although *A. uva-ursi*, *F. vesca* and *R. idaeus* are found on occasion in the Clearwater Lake area today, they are all more common in the aspen parkland and boreal forest regions to the north. In addition to the plant macrofossils, a right elytron of an *Elaphrus clairvillei* beetle was recovered. This beetle is commonly found today in wet organic mud in the shade of sedges, shrubs, or forest (Goulet, 1983). In sum, the CWM1 macrofossil assemblage indicates the presence of a boreal setting (possibly with *P. glauca*), including areas of peat accumulation, in the watershed at ~10 000–9800 BP.

Lithostratigraphic unit CWL2 records a significant change in limnological conditions at the core site between 9800 BP and 9600 BP (Table 4). The depositional environment shifted from an open, freshwater setting with high clastic sediment influx to an evaporitic, highly productive, saline-hypersaline system dominated by chemical sedimentation. The endogenic mineralogy of CWL2 suggests that water salinities were generally within the range of 15 to 70 g L^{-1} but may have occasionally reached higher levels. Initially the brine was dominated by magnesium and sodium cations and bicarbonate and sulfate anions. As discussed above, this saline to hypersaline solution, having relatively high carbonate alkalinities and elevated Mg/Ca and HCO_3/Ca ratios, was also probably responsible for diagenetic alteration of the underlying muds and generation of authigenic dolomite in CWL1. The upward gradation from Na+Mg salts at the base of CWL2 to Ca precipitates at the top of the unit indicates that the lake brine changed to a Mg-Ca-SO_4 system, with relatively low carbonate alkalinities and Mg/Ca ratios of about 2–20. Closed basin and saline, evaporitic conditions at the core site are also indicated by the strong positive covariance of the aragonite $\delta^{18}O$ and $\delta^{13}C$ (Figure 7) and the high $^{18}O/^{16}O$ values.

This gypsum-aragonite-organic matter assemblage of CWL2 is a common sedimentary facies type found in numerous modern saline lakes in the region (Last, 1989; 1993), as well as Holocene stratigraphic sequences of other basins in the Great Plains (Xia *et al.*, 1997; Valero-Garcés *et al.*, 1995; Sack and Last, 1994; Last and Schweyen, 1983). Despite the fine, undisturbed lamination, which is characteristic of this facies in CWL2 and elsewhere, deep-water conditions are not *necessarily* required. This aragonite and gypsum-rich organic mud was probably deposited in shallow water or perhaps on saline evaporitic mudflats surrounding a deeper water basin.

Generally, diatom assemblages contemporaneous with CWL2 (the basal 45 cm of CWD1) indicate fluctuating hydrologic conditions, with both relatively high and low salinity episodes evident (Figures 9 and 10). Lower salinity intervals inferred in diatom zone CWD1 are characterized by decreases in the relative abundance of the mesosaline taxon *Cyclotella choctawhatcheeana*, and

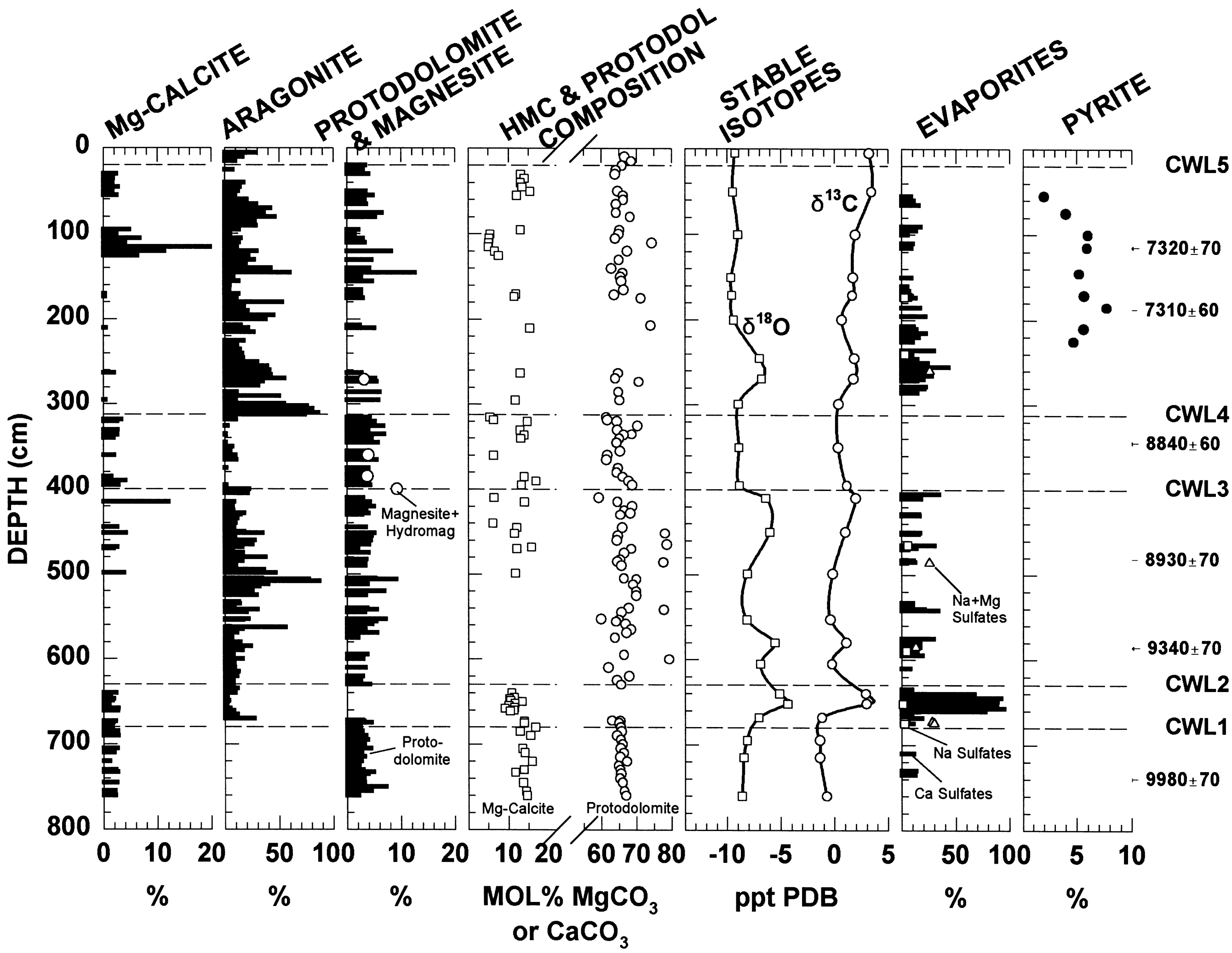

Figure 7 Stratigraphic variation in magnesian calcite, aragonite, protodolomite, magnesite (plus hydromagnesite), Ca sulfates, Na sulfates, Na + Mg sulfates, and pyrite. Also shown are the variations in composition of the magnesian calcite and protodolomite, and the stable oxygen and carbon isotopic composition of the aragonite. Radiocarbon dates (Table 2) and lithostratigraphic units (Table 3) are shown on right. The upper 20 cm of this core is modern sediment.

corresponding increases in *Chaetoceros muelleri*, *Navicula* aff. *rhyncocephala* (both hyposaline), *N.* aff. *veneta* (subsaline) and *Amphora libyca* (freshwater). During two of these low salinity intervals, the cyst:diatom ratio is also much higher (Figure 9), perhaps reflecting the lower salinity conditions (Cumming *et al.*, 1993). An abundance of *C. choctawhatcheeana* in this zone suggests that the lake was fairly deep at times and probably experienced episodes of meromixis, as this taxon is abundant in the modern sediments of deep (11–28 m), meromictic lakes of the northern Great Plains (Fritz *et al.,* 1993).

Macrofossil assemblages contemporaneous with lithostratigraphic zone CWL2 comprise the lower 70 cm of CWM2 (Figure 8). Here, the onset of relatively continuous *Chara* oogonia representation indicates the establishment of limnic conditions similar to present-day Clearwater Lake. Boreal elements, no longer part of the macrofossil assemblage, are replaced by a diverse assemblage that remain part of the local flora today. In particular, periodic inputs of *Juniperus, Symphoricarpus occidentalis, Rosa, Cornus stolonifera* and Betulaceae seeds, as well as *Viburnum* seeds and a single fruit (Figure 8), indicate the establishment of parkland-type shrubs (Budd, 1979; Moss, 1983) that now exist around the margin of Clearwater Lake. The near-continuous deposition of *Chara* oogonia, at times with a high percentage bearing a carbonate coating, indicates the presence of carbonate-rich water. The abundance of Cyperaceae seeds (mainly *Scirpus*), suggests shoreline proximity, as is the case today. These aspects of CLM2 collectively indicate the establishment of a groundwater-fed lake similar to present-day Clearwater Lake.

The major defining characteristics of CWL3 (9600–8700 BP) are the fine grain size, relatively high proportions of aragonite and protodolomite and complementary low soluble salt contents, and increased siliciclastic mineral abundances. These features are consistent with a decrease in salinity, probably associated with increased water depth at the core site, and a shift toward more alkaline conditions with higher Mg/Ca ratios. The sporadic occurrence of gypsum-rich zones throughout the unit, however, suggests that elevated salinities still periodically affected sedimentation at the core site and the significant positive covariance of the $\delta^{18}O$ and $\delta^{13}C$ indicates continued closed basin, evaporitic conditions. The upward decrease in aragonite and protodolomite and complementary increase in magnesian calcite above about 5 m depth indicate a gradual moderating of Mg/Ca ratios after about 8900 BP.

Although it is tempting to view each of the three lower lithostratigraphic units identified in this core as representing discrete phases or different episodes of lacustrine sedimentation in the Clearwater basin, a more sedimentologically reasonable interpretation is that the three units simply represent an onlap of facies in a basin that is transgressing or experiencing rising water levels. In this deepening-upward scenario, the core site was affected by progressively deeper water conditions, beginning at the base with the shoreline clastics of CWL1 (10 000–9800 BP), followed by

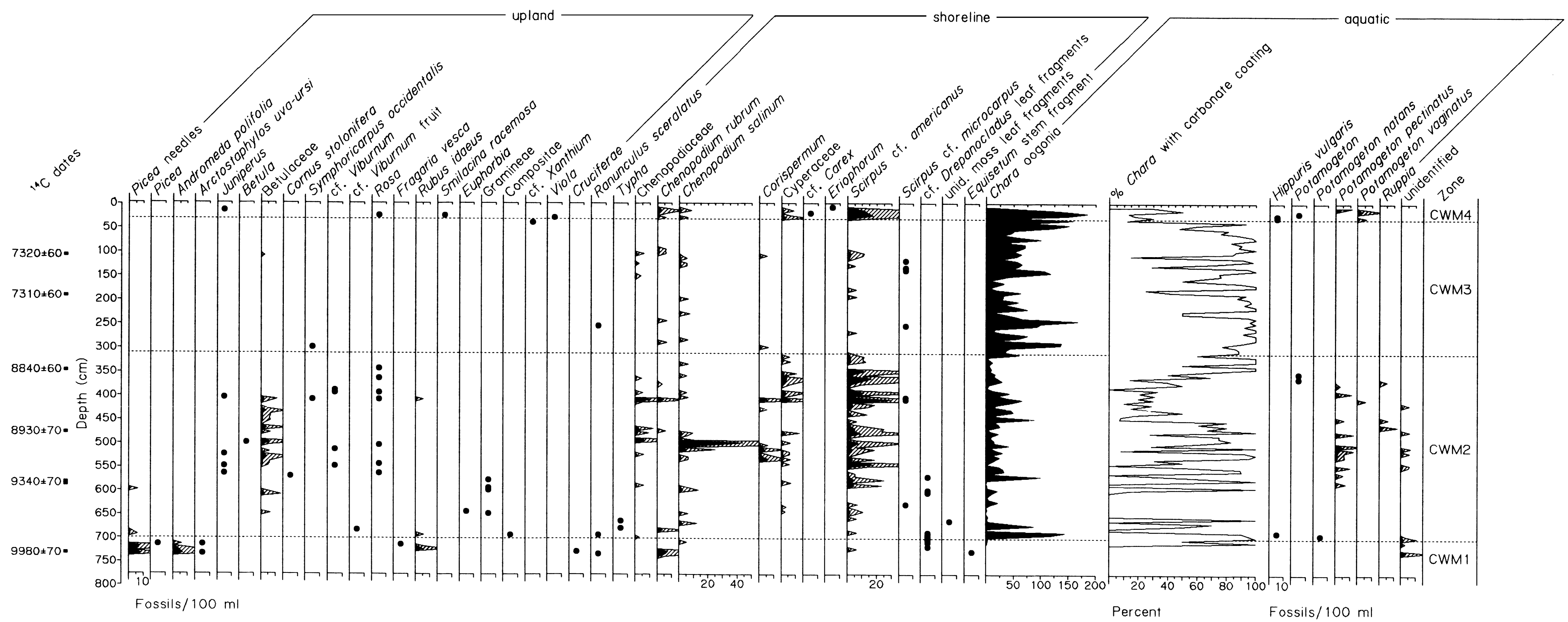

Figure 8 Stratigraphic variation in plant macrofossils, Clearwater Lake core CW2 (note scale changes). Unless otherwise indicated, all macrofossil remains are seeds (i.e. one reproductive unit). Radiocarbon dates (Table 2) are shown on right, macrofossil zones on left. Solid circles indicate presence (<3.5 fossils/100 ml), hatching indicates 5 × exaggeration.

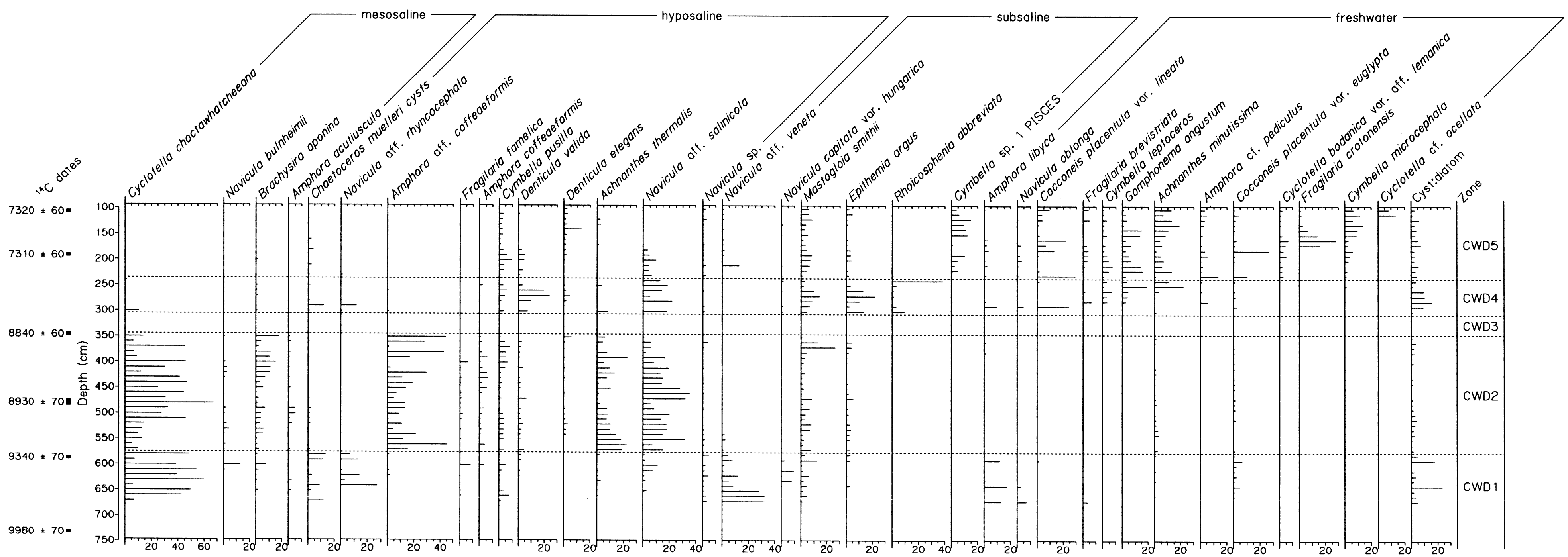

Figure 9 Stratigraphic variation in diatom taxa between 1 and 7 m depth in the Clearwater Lake core CW2. Taxa present at ≥5% relative abundance in at least two samples are shown. The freshwater to mesosaline designations are based on the known (Wilson *et al*., 1996) or suspected salinity optima of the taxa, and are based on Hammer's (1986) system of saline lake classification: mesosaline = 50–20 g L^{-1}; hyposaline = 20–3 g L^{-1}; subsaline = 3–0.5 g L^{-1}; freshwater = <0.5 g L^{-1}. Taxa with designated PISCES numbers were documented from lakes sampled in the Palaeolimnological Investigations of Salinity, Climatic and Environmental Shifts (PISCES) project, and are depicted in Cumming *et al*. (1995). Radiocarbon dates (Table 2) are shown on the left, biostratigraphic diatom zones on right.

Table 4 Summary of inferred limnological and hydrological changes during the early Holocene at Clearwater Lake

^{14}C yr BP	Lithostratigraphy	Isotopes	Plant macrofossils	Diatoms
6500	fresh to subsaline becoming more clastic dominated			not analysed
7000	watershed weathering intensity increasing possibly stratified water column with reducing conditions at sediment/water interface	open system fresh non-evaporitic	shallow low salinity carbonate-rich water	fresh
7500	continued high to moderate Mg/Ca			$<$0.7 g L^{-1}
	hyposaline to saline shallow high SO_4	saline evaporitic		shallow transitional from saline below to fresh above
8000	fresh, relatively deep, high Mg/Ca high alkalinity			$<$4 g L^{-1}
	subaerial exposure ~8250 BP	open system low salinity non-evaporitic		diatoms absent
8500	fresh to subsaline low SO_4 & alkalinity			
			saline to subsaline significant fluctuations in water level	hyposaline to mesosaline ~4–20 g L^{-1} fluctuating water level
9000	hyposaline to saline deep high alkalinity high to moderate Mg/Ca	closed system evaporitic	carbonate-rich groundwater-fed	probably some meromixis fluctuating salinity
9500				subsaline to saline (~2–20 g L^{-1}) fluctuating water level deep, episodic meromixis transitional from subsaline below to saline above
	saline to hypersaline 15–70 g L^{-1} Mg-Ca-SO_4	closed system, saline evaporitic		
	fresh, turbid, shallow, high clastic sedimentation	open system non-evaporitic	Boreal forest and peat accumulation	Diatoms absent
10 000			Macrofossils absent	

the mudflat and shallow water gypsite (CWL2; 9800–9600 BP), and finally the deeper-water, aragonite-laminated muds of CWL3 (9600–8700 BP). The endogenic and authigenic mineralogy of the three units, although diverse, is entirely in keeping with modern sedimentary facies found in numerous other perennial saline lakes in the Great Plains today. Indeed, remarkably similar modern facies spatial relationships have been delineated in other saline lakes of the Great Plains such as Waldsea Lake (Schweyen, 1984; Schweyen and Last, 1983), Deadmoose Lake (Last, 1991; Last and Slezak, 1986), and Freefight Lake (Last, 1993; Slezak, 1989). Unfortunately, with a single core from Clearwater Lake it is not possible to confirm these expected spatial relationships in the stratigraphic record.

However, there is no clear-cut support for the aforementioned deepening-upward scenario in either the diatom or plant macrofossil stratigraphies, or the previously published ostracode stratigraphy of core CW2 (Cywinska and Delorme, 1995). In contrast, a generally variable lacustrine setting, with possible low lake-level episodes, is indicated. As previously mentioned, meromictic conditions are inferred from periodic abundances of *Cyclotella choctawhatcheeana* at the base of diatom zone CWD1 (Figure 9), and through the upper half of CWD1 this interpretation is supported by the presence of *Navicula bulnheimii, Brachysira aponina* and *Fragilaria famelica* – taxa that are common in modern-day assemblages of saline meromictic lakes in British Columbia and the northern Great Plains (Wilson *et al.*, unpublished data; Fritz *et*

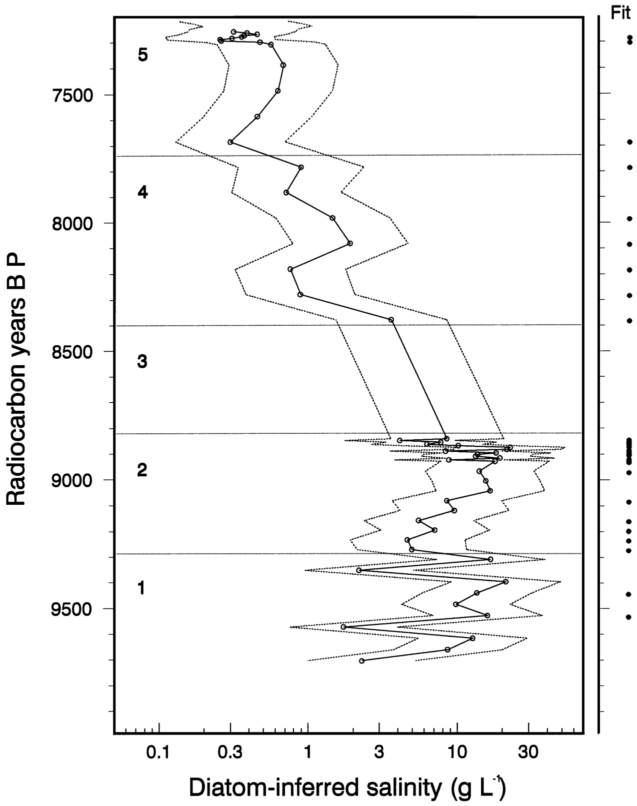

Figure 10 A plot of the diatom-inferred salinity for the early Holocene in Clearwater Lake. Solid circles on the Fit axis indicate fossil diatom assemblages that were not well represented in the modern diatom assemblages of lakes used to develop the transfer function (see text for details). The inferred salinity values are mean bootstrap estimates and the errors (dotted lines) are estimated standard errors of prediction provided in the WACALIB output.

al., 1993). In particular, *N. bulnheimii* is found only in lakes with a chemocline, but can be very abundant in those as shallow as 2 m. *Chaetoceros muelleri* cysts, also present in this zone, are abundant in many of the deep meromictic lakes from the northern Great Plains (Fritz *et al.,* 1993), but in British Columbia appear almost exclusively in shallow lakes (0.5–2 m), some of which have a chemocline (Wilson *et al.,* unpublished data). Thus, although meromictic conditions are implicated, the diatom stratigraphy does not necessarily demand interpretation of deep water conditions.

Aspects of the CWM2 macrofossil assemblage suggest a drier setting than exists today, with lake levels similar to present, or even slightly lower. Periodic inputs of Chenopodiaceae seeds, in particular *Chenopodium salinum*, indicate that at times salt-encrusted mudflats were present (Moss, 1983), possibly maintained by seasonal fluctuations in water levels. The only appearance of *Euphorbia*, combined with periodic inputs of *Corispermum*, indicate the presence of dry habitat in the watershed, as both taxa favour dry, sandy soil (Budd, 1979). Periods with higher than present rates of evaporation are implicated, since parkland shrubs were well established around the lake (indicative of relatively constant groundwater inputs), but saltflats and dry sandy habitat, not common features of the watershed today, were more widespread. Ostracode assemblages also indicate fluctuating salinities through this portion of the record, in this case interpreted as a result of variable precipitation (Cywinska and Delorme, 1995).

The relative abundance of *Potamogeton pectinatus* seeds, combined with the appearance of *Ruppia* seeds in the middle to upper portion of the CWM2 assemblage, indicates that lake-water salinity was higher than at present (*c.* 3–15 g L^{-1}). Both species are common in potholes and shallow lakes of the area today (Kantrud, 1990; 1991), but *P. pectinatus* is generally replaced by *Ruppia* when salinity exceeds *c.* 15 g L^{-1} and water level fluctuations are pronounced. *Ruppia* growth is enhanced by a clear water column, aerobic and low H_2S conditions (Kantrud, 1991), and it is known to reproduce vigorously when salinity increases (Husband and Hickman, 1985). In general, the CWM2 macrofossil assemblage indicates the presence of a carbonate-rich lake, similar in size to its current extent but at times more saline and subjected to greater fluctuations in water level than at present. The appearance of *Ruppia* seeds in the upper portion of CWM2 indicates that peak salinity (*c.* 10–15 g L^{-1}) occurred between *c.* 9000–8800 BP.

The CWD2 diatom record (Figure 10) supports the inference that Clearwater Lake was continuously saline during lithostratigraphic zone CWL3, although lower relative abundances of *Cyclotella choctawhatcheeana* and corresponding increases in the benthic saline taxa at the beginning of this zone suggest that water levels may have been lower at this point. Similarly, dramatic fluctuations in the abundances of *C. choctawhatceeana* and *Amphora* aff. *coffeaeformis* towards the end of this zone may also indicate substantial changes in water level, consistent with fluctuations inferred from the macrofossil record. However, high sedimentation rates in the latter half of this zone (each 1-cm interval represents seven years of sedimentation history) and subsequent higher stratigraphic resolution may be partially responsible for these apparently large fluctuations. The very low cyst:diatom ratio throughout this period (Figure 9) is probably indicative of higher salinities (Cumming *et al.,* 1993).

The degree of fit to the salinity axis for the assemblages in this section of the core is very poor (Figure 10), and two of the dominant taxa, *Achnanthes thermalis* and *Navicula* aff. *salinicola*, were too scarce in the modern calibration lakes to be included in the salinity transfer function (Wilson *et al.,* 1996). The salinity preferences assigned to these two taxa are estimates from what is known of their distribution in the calibration lakes and from the literature (e.g., Kaczmarska and Rushforth, 1983). *A. thermalis* is also known to occur in thermal springs (Kaczmarska and Rushforth, 1983; E. Stoermer, P. Bradbury, personal communication, 1995). The salinity optimum of *Amphora* aff. *coffeaeformis* should also be considered tentative, for although this taxon was included in the transfer function it had very few occurrences in the calibration lakes and was present at much lower abundances ($<3\%$) than observed in this core (Wilson *et al.,* 1996). Considering the analogue problems for this period of the lake's history, the salinity reconstructions should be regarded with some caution, and although it is quite clear that the lake was saline throughout this zone, salinities may have actually been higher than indicated in Figure 10.

The palaeoenvironmental significance of lithostratigraphic unit CWL4 and diatom zone CWD3 is enigmatic. It is clear from both the mineralogy and siliceous fossil remains that after about 8700 BP water salinity significantly decreased and deposition at the core site was once again dominated by fine-grained detrital sediments. The absence of gypsum and other soluble sulfate salts implies that salinities were generally less than about 10 g L^{-1} and/or that calcium, sodium, and sulfate ion activities were low. Similarly, $\delta^{18}O$ values decrease significantly in CWL4 and CWD3, probably in response to the lower salinity. In contrast, the occurrence of magnesite, the relatively high values of protodolomite, and low aragonite and magnesian calcite contents point toward high carbonate alkalinities and greatly elevated Mg/Ca ratios. The paucity of diatoms and of clay mineral components in this fine-grained unit is also problematic. It is possible that the immediate source of the detrital sediment deposited at the core site between 8700 and 8250 BP was not the glacial tills of the watershed but rather the influx of clastics from reworking of a previously deposited strandline from which the clay minerals had already been winnowed. However, this does not explain the absence or lack of preservation of siliceous microfossils through diatom zone CWD3. Low macrofossil concentrations also characterize this portion of the record (uppermost 40 cm of CWM2), and the few remains recovered do not confer a strong ecological signal. Only a few Cyperaceae seeds were encountered, and aquatic macrofossils are rare, including *Chara* oogonia, which attain some of the lowest concentrations recorded in the diagram.

The low moisture content and distinctive blocky and pelletal structure developed into the top of CWL4 and diatom zone CWD3 indicates that water levels in the basin dropped sufficiently to subaerially expose the sediment at the core site about 8250 BP (Table 4). The duration of this low-water episode is not known, but may have been very short. Teller and Last (1982) indicate that pedogenic horizons like this can form in a matter of a few years. Observations (WML) during the 1980s at Old Wives Lake (a large, shallow, saline lake near Moose Jaw, Saskatchewan) confirm that dry, blocky-structured zones can develop in fine-grained offshore bottom muds after just two summer seasons of exposure.

The finely laminated, calcareous, organic-rich sediments immediately overlying this blocky-structured pedogenic zone also indicate that a major change occurred in Clearwater Lake at this time. The rising lake that reflooded the core site was initially relatively fresh, with Mg/Ca ratios of about 10–20 and high carbonate alkalinities. This freshening, combined with high levels of organic productivity, created strongly supersaturated conditions in the lake resulting in several hundred years of abundant aragonite precipitation. However, between about 8000 and 7700 BP, shallow water, evaporitic and saline conditions again returned as evidenced by the reappearance of gypsum and other soluble salts, relatively coarse grain size, and an abrupt increase in $^{18}O/^{16}O$ ratios of aragonite.

After this short episode of evaporative concentration and high salinity, the weak inverse correlation between endogenic carbonate $\delta^{18}O$ and $\delta^{13}C$ and the much lower $^{18}O/^{16}O$ ratios suggest a change toward more open-basin, fresher water conditions. Greater runoff from the watershed is also indicated by gradually increasing levels of siliciclastics upward in CWL5. Similarly, pyrite, which occurs only in CWL5 above about 2 m depth, is consistent with deeper water conditions. This reduced iron sulfide is most likely an early diagenetic product formed in association with bacterial degradation of organic matter at or near the sediment-water interface under moderate to strongly reducing conditions. Although FeS_2 can form under shallow, nonaerated ‘swamp’ conditions, in the northern Great Plains today it is found exclusively in relatively deepwater perennial lakes.

The diatom stratigraphy supports this interpretation of generally freshwater conditions after 8250 BP, but does not contain evidence of high salinity and evaporative conditions between 8000 and 7700 BP (similar trends are apparent in the ostracode stratigraphy; Cywinska and Delorme, 1995). Diatom zone CWD4 (305–235 cm; 8200–7900 BP) records a gradual transition to a less saline water body where salinities fluctuate just above and below 1 g L^{-1} (Figure 10). Numerous subsaline and freshwater diatoms appear in abundance for the first time in this zone, including *Epithemia argus*, *Rhoicosphenia abbreviata*, *Cocconeis placentula* var. *lineata*, *Gomphonema angustum* and *Achnanthes minutissima* (Figure 9). Higher cyst:diatom ratios at this time corroborate the lower salinities inferred by the diatoms.

Modern analogues for this period of the lake’s history are once again scarce, largely due to the high abundances of *Navicula* aff. *salinicola*. The inferred salinity values (Figure 10) should therefore be regarded as tentative, and are probably underestimates of the actual salinity concentrations, since *N.* aff. *salinicola* may have a salinity optimum in the high hyposaline to mesosaline range (Wilson *et al.*, unpublished data).

In the absence of good salinity estimates, it is difficult to establish precisely when the transition from saline/subsaline to fresh water took place, although it is clear that the lake was fresh ($\leq$0.5 g L^{-1}) by 7500 BP (Figure 10). The common taxa in this freshwater phase (zone CWD5) include the benthic and epiphytic forms *Cymbella* sp. 1 PISCES, *Cocconeis placentula* var. *lineata*, *Gomphonema angustum*, *Achnanthes minutissima*, *Cymbella microcephala*, *Mastogloia smithii*, and the planktonic forms *Fragilaria crotonensis* and *Cyclotella* cf. *ocellata* Pant. (Figure 9). The dominance of benthic taxa in zones CWD4 and CWD5 suggests that water levels were probably lower than 9800–8400 BP, as indicated by the lithostratigraphic and isotopic records.

The macrofossil record supports inferences of increasing water levels after 8250 BP, but, like the diatom record, shows no indication of the brief period of heightened salinity between 8000 and 7700 BP inferred from the lithostratigraphic and isotopic records. The abundance of *Chara* oogonia in the sediments deposited after 8250 BP (macrofossil zone CWM3) indicates the continued existence of carbonate-rich waters, although the absence of *Potamogeton pectinatus* and *Ruppia* indicates that salinity had declined from peak levels attained in CWM2. Thick *Chara* beds are now established in shallow water (<2 m) in the Clearwater Lake littoral zone (Vance, unpublished data), including the area where core CW2 was collected. The presence of *Chara* throughout zone CWM3 suggests shallow water conditions were maintained, although the relative paucity of other macrofossils in such a setting is puzzling. Since so few macrofossils were recovered in this zone, interpretations are limited.

As indicated previously, the uppermost 20 cm of core CW2 are mineralogically distinct and interpreted as representing modern deposition. The onset of a diverse, high concentration macrofossil assemblage in zone CWM4 supports this interpretation. With the data at hand it is impossible to address either the cause of the underlying sedimentary hiatus or its duration. However, given the abundant evidence for widespread middle-Holocene drought conditions in western Canada (Vance *et al*., 1995), it is likely that the hiatus is the result of a lake-level decline *c*. 7000 BP sufficient to subaerially expose coring site CW2.

Employing a multiple-proxy approach in palaeolimnological studies offers the possibility of developing more detailed reconstructions than are typically achieved utilizing a single indicator. However, because each proxy indicator has inherent strengths and limitations related to varying thresholds and sensitivities, the use of multiple proxies can at times lead to what appear to be contradictory inferences. Although the proxies utilized in this study, and the previously published ostracode stratigraphy (Cywinska and Delorme, 1995), are in general agreement in terms of major early-Holocene hydrological developments, it is clear that discrepancies between indicators do occasionally arise. Such discrepancies underscore the necessity for cautious interpretation of records based on a single proxy and highlight the need for improved understanding of the environmental factors driving current responses of various proxy indicators. Nevertheless, areas of general agreement amongst various proxies used in the Clearwater Lake analyses significantly improve confidence in the reconstructions developed here, allowing a preliminary assessment of regional hydrologic and climatic conditons in the early Holocene.

Mineralogical and isotopic evidence from this early-Holocene record indicates that maximum water salinities (~20–70 g L^{-1}), probably associated with evaporitic, closed-basin conditions, were attained in Clearwater Lake about 9800–9600 BP. The diatom and macrofossil records from this core further suggest that, despite significant water level fluctuations, moderate salinities (~3–20 g L^{-1}) persisted in the lake under closed-basin hydrologic conditions until at least 8800 BP. This early-Holocene high-salinity phase corresponds well with several other late glacial/early-Holocene lacustrine records from the region. Valero-Garcés *et al.* (1995), Radle *et al.* (1989) and Kennedy (1994) indicate that Medicine Lake, South Dakota, changed from a dilute, freshwater lake at about 10 000 BP to a highly saline basin at 9500 years ago. Similarly, the geochemistry of ostracod shells and endogenic inorganic carbonates from Coldwater Lake, North Dakota, shows that an increase in salinity occurred about 9000 BP (Xia *et al.*, 1997). Oro Lake, a closed-basin, saline lake in south-central Saskatchewan, records an abrupt change from shallow but relatively freshwater conditions at 9400 BP to a deepwater, hypersaline, meromictic basin at about 9000 BP (Vance and Last, 1996b; Last, 1996). A 10 000-year-long sequence of salts from a hypersaline playa in the Great Sandhills area of southwestern Saskatchewan, approximately 100 km southwest of Clearwater, indicates much lower relative humidity about 9800 BP (Shang and Last, 1998). Diatom and pollen stratigraphies from Moore Lake, east-central Alberta, indicate a shift from fresh to saline conditions between 10 000 and 9000 BP (Hickman and Schweger, 1993).

In contrast, the stratigraphic records at Devils Lake, North Dakota (Fritz *et al.*, 1991; Haskell *et al.*, 1996), Moon Lake, North Dakota (Laird *et al.*, 1996), Ceylon Lake, Saskatchewan (Last, 1990b), and the Andrews dugout near Moose Jaw, Saskatchewan (Yansa, 1995), indicate that the transition to fully saline conditions occurred considerably later in the Holocene (~8000–6000 BP). Likewise, at many of these sites, this high-salinity phase persisted considerably longer into the Holocene than is evident at Clearwater Lake, although the truncated record at Clearwater does not permit a complete comparison of Holocene events.

At most of the above sites, high salinities during the early to middle Holocene have been interpreted as periods of lower water level resulting from evaporative concentration under a warmer and/or drier climate. This warm period has been explained by increased summer insolation at high latitudes between 12 000–9000 BP, induced by Milankovitch earth/sun orbital variations (Barnosky *et al.*, 1987; Ritchie and Harrison, 1993), and generally occurred somewhat later in the midwestern and eastern parts of the continent where the climate was controlled for a longer period by the retreating Laurentide ice sheet (Barnosky *et al.*, 1987). The fact that the highest salinities at Clearwater Lake occurred prior to about 8800 BP probably does, indeed, reflect this early-Holocene maximum summer insolation. However, the relatively abrupt termination of the high-salinity phase and the transition to a generally fresher-water environment in Clearwater Lake after 8700 BP may reflect the initiation and increased influence of a local groundwater flow system, although it may be related to regional climate change. Between 9400 and 8400 BP, both downslope expansion of *Pinus* and *Picea* forests in western Alberta (MacDonald, 1989) and interruption of high-salinity conditions with fresh water in Moore Lake, east-central Alberta (Hickman and Schweger, 1996), have been posited as possible responses to a period of more mesic regional climate, compared to conditions before and after this time period. Since the freshwater phase in Clearwater Lake beginning at 8700 BP falls within this interval, it may also have been climatically induced. However, until this possible climatic event is documented in other well-dated sites in the region, the climatic linkage will remain speculative. At this time, however, the dominance of the groundwater discharge-recharge system on the hydrologic budget of the Clearwater Lake basin, a system established early in the Holocene, created a situation similar to that of today in which the lake was able to maintain relatively low salinities and thereby overwhelm any regional climatic signal that may have been preserved by the proxy indicators (Leavitt *et al.*, 1998).

Conclusion

A need for more lacustrine sequences that record early-Holocene environmental conditions in the vast northern Great Plains region of western Canada is evident. Centrally located in the Palliser Triangle, the most arid part of this 400 000 km^2 region**,** Clearwater Lake is strategically situated to provide an important link between the well-studied marginal areas of the prairie region in Manitoba and west-central Alberta. The various proxy indicators investigated in this study (sediment texture, bedding, mineralogy, stable isotopes, plant macrofossils and diatoms) exhibit somewhat different responses, which probably reflect inherent differences in the way the components respond to lacustrine and hydrological conditions in the basin. Although the earliest sediments in the lake (deposited prior to ~9800 BP) are devoid of fossil remains, the physical, mineralogical and isotopic parameters indicate that a shallow, freshwater, clastic-dominated lake occupied the Clearwater basin. The earliest plant macrofossils preserved in the Clearwater record suggest that the landscape was dominated by boreal forest components. After 9800 BP, the basin entered a 1000-year-long episode of relatively high salinity, closed, evaporitic conditions and fluctuating water levels, as indicated by the diatoms, endogenic mineralogy, and the δ^{18}O values. The plant macrofossils provide evidence of the existence of a groundwater-fed lake that was probably subjected to periodic fluctuations in water level and of the establishment of salt-encrusted mudflats and sandflats in the marginal areas of the basin. Fluctuating salinities, from subsaline to saline conditions, probably with episodes of meromixis, are also evident from the diatom record. After about 8600 BP, the mineral, diatom, and isotope records show that the lake became mainly fresh but still maintained high alkalinities and elevated Mg/Ca ratios. Except for a short period of shallow and saline conditions at about 8000–7700 BP, a freshwater, open-basin environment persisted in the lake until at least 6600 BP. The macrofossils similarly point toward the establishment of a low-salinity, carbonate-rich, groundwater-fed lacustrine setting after about 8250 BP. At some point subsequent to 6600 BP, water levels in the basin dropped and the remaining Holocene record at the core site was lost through either nondeposition or erosion.

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References

Aharonson, E.F., Karasikov, N., Roitberg, M. and **Shamir, J.** 1986: GALAI-CIS-1 – A novel approach to aerosol particle size analysis. *Journal of Aerosol Science* 17, 530–36.

Al-Aasm, I.S. 1990: Stable isotope analysis of multiple carbonate samples using selective acid extraction. *Chemical Geology* 80, 119–25.

Allen, T. 1981: *Particle size measurement* (third Edition). Toronto: Chapman & Hall.

Baker, P.A. and **Kastner, M.** 1981: Constraints on the formation of sedimentary dolomite. *Science* 213, 214–16.

Barnosky, C.W., Grimm, E.C. and **Wright, H.E.** 1987: Towards a postglacial history of the northern Great Plains: a review of the paleoecologic problems. *Annals Carnegie Museum* 56, 259–73.

Beijerinck, W. 1976: *Zadenatlas der Nederlandsche Flora: Ten Behoeve van de Botanie, Palaeontologie, Bodemcultuur en Warenkennis*. Amsterdam: Backhuys & Meesters, 316 pp.

Birks, H.J.B., Line, J.M., Juggins, S., Stevenson, A.C. and **ter Braak,**

C.J.F. 1990: Diatoms and pH reconstruction. *Philosophical Transactions of the Royal Society London* B327, 263–78.

Budd, A.C. 1979. *Budd's flora of the Canadian Prairies*. Research Branch, Agriculture Canada, Publication 1662.

Burton, E.A., Haln, T. and **Machel, H.G.** 1992: Magnesium sulfate brine lakes that precipitate dolomite: evidence that Mg^{2+} dehydration is more important than removal of SO_4^{2-} inhibition in dolomite formation. In Kharaka, Y.K. and Maest, A.S. editors, *Water-Rock Interaction*, Rotterdam: Balkema, p. 617–20.

Carvalho, L.R., Cox, E.J., Fritz, S.C., Juggins, S., Sims, P.A., Gasse, F. and **Battarbee, R.** 1995: Standardizing the taxonomy of saline lake *Cyclotella* spp. *Diatom Research* 10, 229–40.

Cumming, B.F., Wilson, S.E., Hall, R.I. and **Smol, J.P.** 1995: *Diatoms from British Columbia (Canada) and their relationship to salinity, nutrients, and other limnological variables.* Bibliotheca Diatomologica Band 31. Berlin/Stuttgart: Gebrüder Borntraeger, 207 pp.

Cumming, B.F., Wilson, S.E. and **Smol, J.P.** 1993: Paleolimnological potential of chrysophyte cysts and scales and of sponge spicules as indicators of lake salinity. *International Journal of Salt Lake Research* 2, 87–92.

Curry, B.B. 1997: Paleochemistry of Lakes Agassiz and Manitoba based on ostracodes. *Canadian Journal of Earth Sciences*, 34, 699–708.

Cywinska, A. and **Delorme, L.D.** 1995: *Paleoenvironmental interpretations using fossil freshwater ostracodes from Clearwater Lake core site, Saskatchewan, and Max Lake, Manitoba*. National Water Research Institute, Burlington, Ontario, NWRI Contribution 95-1-7, 43 pp.

Dean, W.E. 1974: Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition: comparison with other methods. *Journal of Sedimentary Petrology* 44, 242–48.

Drimmie, R.J., Heemskerk, A.R., and **Patton, M.E.** 1992: *^{13}C and ^{18}O in carbonates*. Technical Procedure 11.0, Rev. 01. Environmental Isotope Laboratory, Department of Earth Sciences, University of Waterloo, 10 pp.

Environment Canada 1993: *Canadian climate normals 1961–1990 (Volume 2) – Prairie Provinces*. Ottawa: Canadian Climate Program.

Folk, R.L. and **Land, L.S.** 1975: Mg/Ca ratio and salinity: two controls over crystallization of dolomite. *American Association of Petroleum Geologist Bulletin* 59, 60–68.

Fritz, S.C., Juggins, S. and **Battarbee, R. W.** 1993: Diatom assemblages and ionic characterization of freshwater and saline lakes of the northern Great Plains, North America: a tool for reconstructing past salinity and climate fluctuations. *Canadian Journal of Fisheries and Aquatic Sciences* 50, 1844–56.

Fritz, S.C., Juggins, S., Battarbee, R.W., and **Engstrom, D.R.** 1991: Reconstruction of past changes in salinity and climate using a diatom-based transfer function. *Nature* 352, 706–708.

Goldsmith, J.R. and **Graf, D.L.** 1958: Relation between lattice constants and composition of the Ca-Mg carbonates. *American Mineralogist* 43, 84–101.

Goulet, H. 1983: The genera of holarctic Elaphrini and species of *Elaphrus* Fabricus (Coleoptera: Carabidae): classification, phylogeny and zoogeography. *Quaestiones Entomologicae* 19, 219–482.

Hakansson, H., Hajdu, S., Snoeijs and **Loginova, L.** 1993: *Cyclotella hakanssoniae* Wendker (Bacillariophyceae), and its relationship to *C. caspia* Grunow and other similar brackish water *Cyclotella* species. *Diatom Research* 8, 333–47.

Hammer, U.T. 1986: *Saline lake ecosystems of the world*. Dordrecht: Dr W. Junk Publishers, 616 pp.

Hardy, R. and **Tucker, M.** 1988: X-ray powder diffraction of sediments. In Tucker, M., editor, *Techniques in sedimentology,* Oxford: Blackwell Scientific Publications, 191–228.

Haskell, B.J., Engstrom, D.R. and **Fritz, S.C.** 1996: Late Quaternary paleohydrology in the North American Great Plains inferred from the geochemistry of endogenic carbonate and fossil ostracodes from Devils Lake, North Dakota, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 124, 179–93.

Hickman, M. and **Schweger, C.E.** 1993: Late glacial-early Holocene paleosalinity in Alberta, Canada – climate implications. *Journal of Paleolimnology* 8, 149–60.

—— 1996: The Late Quaternary paleoenvironmental history of a presently deep freshwater lake in east-central Alberta, Canada and paleoclimatic implications. *Palaeogeography, Palaeoclimatology, Palaeoecology* 123, 161–78.

Hickman, M., Schweger, C.E. and **Habgood, T.** 1984: Lake Wabamun, Alta.: a paleoenvironmental study. *Canadian Journal of Botany* 62, 1438–65.

Husband, B.C. and **Hickman, M.** 1985: Growth and biomass accumulation of *Ruppia occidentalis* in three lakes differing in salinity. *Canadian Journal of Botany* 63, 2004–14.

Kaczmarska, I. and **Rushforth, S.R.** 1983: *The diatom flora of Blue Lake warm spring, Utah, USA* Bibliotheca Diatomologica Band 2. Vaduz: A.R. Gantner Verlag Kommanditgesellschaft, 123 pp.

Kantrud, H. A. 1990: Sago pondweed (*Potamogeton pectinatus* L.): a literature review. Washington, DC: United States Department of the Interior, Fish and Wildlife Service, Resource Publication 10, 91 pp.

—— 1991: Wigeongrass (*Ruppia maritima* L.): a literature review. Washington, DC: United States Department of the Interior, Fish and Wildlife Service, Resource Publication 10, 58 pp.

Kennedy, K.A. 1994: Early-Holocene geochemical evolution of saline Medicine Lake, South Dakota. *Journal of Paleolimnology* 10, 69–84.

Klug, H.P. and **Alexander, L.E.** 1974*: X-ray diffraction procedures for polycrystalline and amorphous materials.* New York: John Wiley, 416 pp.

Krammer, K. and **Lange-Bertalot, H.** 1986, 1988, 1991: Bacillariophyceae 1–4. In Ettl, H., Gärtner, G., Gerloff, J., Heynig, H. and Mollenhauer, D., editors, *Susswasserflora von Mitteleuropa*, Band 2/1, 2/2, 2/3, 2/4. Stuttgart/New York: Gustav Fischer Verlag.

Laird, K.R., Fritz, S.C., Grimm, E.C. and **Mueller, P.G.** 1996: Century-scale paleoclimatic reconstruction from Moon Lake, a closed-basin lake in the northern Great Plains. *Limnology and Oceanography* 41, 890–902.

Last, W.M. 1989: Continental brines and evaporites of the northern Great Plains of Canada. *Sedimentary Geology* 64, 207–21.

—— 1990a: Lacustrine dolomite – an overview of modern, Holocene, and Pleistocene occurrences. *Earth Science Reviews* 27, 221–63.

—— 1990b: Paleohydrology and paleochemistry of Ceylon Lake, Canada. *Journal of Paleolimnology* 4, 219–28.

—— 1991: *Sedimentology, geochemistry and evolution of saline lakes in the Saskatoon area – excursion guidebook*. International Conference on Sedimentary and Paleolimnological Records of Saline Lakes, Saskatoon, Canada, 192 pp.

—— 1993: Geolimnology of Freefight Lake: an unusual hypersaline lake in the Northern Great Plains of western Canada. *Sedimentology* 40, 431–48.

—— 1995: Evolution of saline lakes in western Canada. In Campbell, Campbell, C. Lemmen, D.S. and Vance, R.E., editors, *Climate, landscape, and vegetation change in the Canadian prairie provinces*, Edmonton: University of Alberta, p. 55–64.

—— 1996: Lithostratigraphy and mineralogy of Oro Lake, Saskatchewan. *The Canadian Association of Geographers Annual Meeting*, Program and Abstracts, 130–31.

Last, W.M. and **Schweyen, T.H.** 1983: Sedimentology and geochemistry of saline lakes of the Great Plains. *Hydrobiologia* 105, 245–63.

Last, W.M. and **Slezak, L.A.** 1986: Paleohydrology, sedimentology, and geochemistry of two deep saline lakes from southern Saskatchewan. *Géographie Phys. Quat.* 11, 5–15.

Last, W.M., Teller, J.T. and **Forester, R.M.** 1994: Paleohydrology and paleochemistry of Lake Manitoba, Canada: the isotope and ostracod records. *Journal of Paleolimnology* 12, 269–82.

Leavitt, P.R., Vinebrook, R.D., Hall, R.I., Wilson, S.E., Smol, J.P., Vance, R.E., and **Last, W.M.** 1998: Multiproxy record of prairie lake response to climatic change and human activity: Clearwater Lake, Saskatchewan. In Lemmen, D.S. and Vance, R.E., editors, Holocene climate and environmental change in the Palliser Triangle, southern Canadian prairies, *Geological Survey of Canada Bulletin*, in press.

Lemmen, D.S., editor 1996: *Landscapes of the Palliser Triangle*. Guidebook for the Canadian Geomorphological Research Group Field Trip, Canadian Association of Geographers 1996 Annual Meeting, Saskatoon, Saskatchewan.

Lemmen, D.S., Dyke, L.D. and **Edlund, S.A.** 1993: The Geological Survey of Canada's Integrated Research and Monitoring Area (IRMA) projects: a contribution to Canadian global change research. *Journal of Paleolimnology* 9, 77–83.

Line, J.M., ter Braak, C.J.F. and **Birks, H.J.B.** 1994: WACALIB version 3.3 – a computer program to reconstruct environmental variables from fossil assemblages by weighted averaging and to derive sample-specific errors of prediction. *Journal of Paleolimnology* 10, 147–52.

MacDonald, G.M. 1989: Postglacial paleoecology of the subalpine forest – grassland ecotone of southwestern Alberta: new insights on vegetation and climate change in the Canadian Rocky Mountains and adjacent foothills. *Palaeogeography, Palaeoclimatology, Palaeoecology* 73, 155–73.

Marquart, R.G. 1986: μPDSM: Mainframe search/match on an IBM PC. *Powder Diffraction* 1, 34–36.

Martin, A.C. and **Barkley, W.D.** 1961: *Seed identification manual*. Berkeley: University of California Press, 220 pp.

Montgomery, F.H. 1977: *Seeds and fruits of plants of eastern Canada and northeastern United States*. Toronto: University of Toronto Press, 232 pp.

Morrow, D.W. 1982: Diagenesis I. Dolomite – part I: the chemistry of dolomitization and dolomite precipitation. *Geoscience Canada* 9, 5–13.

Morrow, D.W. and **Ricketts, B.D.**1986: Chemical controls on the precipitation of mineral analogues of dolomite: the sulfate enigma. *Geology* 14, 408–14.

—— 1988: Experimental investigation of sulfate inhibition of dolomite and its mineral analogues. In Shukla, V. and Baker, P.A., editors, *Sedimentology and geochemistry of dolostones*, SEPM Special Publication 43, 25–38.

Moss, E.H. 1983: *The flora of Alberta (second edition)*. Toronto: University of Toronto Press.

Mott, R.J. 1973: *Palynological studies in central Saskatchewan. Pollen stratigraphy from lake sediment sequences*. Geological Survey of Canada, Paper 72–49.

Müller, G. 1967: *Methods in sedimentary petrography.* New York: Hafner Press, 298 pp.

Müller, G., Irion, G. and **Förstner, U.** 1972: Formation and diagenesis of inorganic Ca-Mg carbonates in the lacustrine environment. *Naturwissenschaften* 59, 159–64.

Radle, N., Keister, C.M., and **Battarbee, R.W.** 1989: Diatom, pollen, and geochemical evidence for the paleosalinity of Medicine Lake, S. Dakota, during the Late Wisconsin and early Holocene. *Journal of Paleolimnology* 2, 159–72.

Risser, P.G., Birney, E.C., Blocker, H.D., May, S.W., Parton, W.J. and Weins, J.A. 1981: *The true prairie ecosystem*. Stroudsburg, Pennsylvania: Hutchinson Ross Publishing Company, 557 pp.

Ritchie, J.C. and **De Vries, B.** 1964: Contributions to the Holocene paleoecology of west central Canada. A late-glacial deposit on the Missouri Coteau. *Canadian Journal of Botany* 42, 677–92.

Ritchie, J.C. and **Harrison, S.P.** 1993: Vegetation, lake levels, and climate in western Canada during the Holocene. In Wright, H.E., Jr., Kutzbach, J.E., Webb, T., III, Ruddiman, W.F., Street-Perrott, F.A. and Bartlein, P.J., editors, *Global climates since the last glacial maximum,* Minneapolis: University of Minnesota Press, 401–14.

Ritchie, J.C. and **MacDonald, G.M.** 1986: The patterns of post-glacial spread of white spruce. *Journal of Biogeography* 13, 527–40.

Romanek, C.S., Grossman, E.L. and **Morse, J.W.** 1992: Carbon isotopic fractionation in synthetic aragonite and calcite: effects of temperature and precipitation rate. *Geochimica et Cosmochima Acta* 56, 419–30.

Rozkowski, A. 1969: *Chemistry of ground and surface waters in the Moose Mountain area, southern Saskatchewan*. Geological Survey of Canada Paper 67–9.

Sack, L.A. and **Last, W.M.** 1994: Lithostratigraphy and recent sedimentation history of Little Manitou Lake, Saskatchewan, Canada. *Journal of Paleolimnology* 10, 199–212.

Schultz, L.G. 1964: *Quantitative interpretation of mineralogical composition from X-ray and chemical data for the Pierre Shale*. United States Geological Survey Professional Paper 391-C.

Schweger, C.E. and **Hickman, M.** 1989: Holocene paleohydrology of central Alberta: testing the general-circulation-model climate simulations. *Canadian Journal of Earth Sciences* 26, 1826–33.

Schweyen, T.H., 1984. *Sedimentology of Waldsea Lake, Saskatchewan.* MSc thesis, University of Manitoba, Winnipeg, Canada.

Schweyen, T.H. and **Last, W.M.** 1983: Sedimentology and paleohydrology of Waldsea Lake, Saskatchewan. *Canadian Plains Proceedings* 11, 45–59.

Shang, Y. and **Last, W. M.** 1998: Mineralogy, lithostratigraphy and inferred geochemical history of North Ingebright Lake, Saskatchewan, Canada. In Lemmen, D.S. and Vance, R.E., editors, *Holocene climate and environmental change in the Palliser Triangle, southern Canadian prairies, Geological Survey of Canada Bulletin*, in press.

Slezak, L.A. 1989: *Sedimentology of Freefight Lake, Saskatchewan.* MSc thesis, University of Manitoba, Winnipeg, Canada.

Smol, J.P. 1985: The ratio of diatom frustules to chrysophycean statospores: a useful paleolimnological index. *Hydrobiologia* 123, 199–208.

Syvitski, J.P.M, LeBlanc, K.W.G. and **Asprey, K.W.** 1991: Interlaboratory, interinstrument calibration experiment. In Syvitski, J.P.M., editor, *Principles, methods, and application of particle size analysis,* Cambridge: Cambridge University Press, 174–93.

Tarutani, T., Clayton, R.N. and **Mayeda, T.K.** 1969: The effect of polymorphism and magnesium substitution on oxygen isotope fractionation between calcium carbonate and water. *Geochimica et Cosmochimica Acta* 33, 987–96.

Teller, J.T. and **Last, W.M.** 1981: Late Quaternary history of Lake Manitoba, Canada. *Quaternary Research* 16, 97–116.

—— 1982: Pedogenic zones in postglacial sediment of Lake Manitoba, Canada. *Earth Surface Processes and Landforms* 7, 367–79.

ter Braak, C. J. F. 1990: *Update Notes: CANOCO version 3.10.* The Netherlands: Agricultural Mathematics Group, 35 pp.

Valero-Garcés, B.L., Kelts, K. and **Ito, E.** 1995: Oxygen and carbon isotope trends and sedimentological evolution of a meromictic and saline lacustrine system: the Holocene Medicine Lake basin, North American Great Plains, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 117, 253–78.

Vance, R.E. 1997: The Geological Survey of Canada's Palliser Triangle Global Change Project: a multidisciplinary geolimnological approach to predicting potential global change impacts on the northern Great Plains. *Journal of Paleolimnology* 17, 3–8.

Vance, R.E. and **Last, W.M.** 1994: Paleolimnology and global change on the Canadian prairies. In Current Research 1994-B, Geological Survey of Canada, Ottawa, 49–58.

—— 1996a: Stop 31: Clearwater Lake. In Lemmen, D.S., editor, *Landscapes of the Palliser Triangle*, Guidebook for the Canadian Geomorphological Research Group Field Trip, Canadian Association of Geographers 1996 Annual Meeting, Saskatoon, Saskatchewan, 78–79.

—— 1996b: Stop 4: Oro Lake. In Lemmen, D.S., editor, *Landscapes of the Palliser Triangle*, Guidebook for the Canadian Geomorphological Research Group Field Trip, Canadian Association of Geographers 1996 Annual Meeting, Saskatoon, Saskatchewan, 26–27.

Vance, R.E., Beaudoin, A.B. and **Luckman, B. H.** 1995: The paleoecological record of 6ka BP climate in the Canadian prairie provinces. *Géographie Phys. Quat.* 49, 81–98.

Van Stempvoort, D.R., Edwards, T.W.D., Evans, M.S. and **Last, W.M.** 1993: Paleohydrology and paleoclimate records in a saline prairie lake core: mineral, isotope and organic indicators. *Journal of Paleolimnology* 8, 135–47.

Warren, J.W. 1989: *Evaporite Sedimentology.* Englewood Cliffs, NJ: Prentice Hall, 285 pp.

Wilson, S.E., Cumming, B.F. and **Smol, J.P.** 1996: Assessing the reliability of salinity inference models from diatom assemblages: An examination of a 219 lake data set from western North America. *Canadian Journal of Fisheries and Aquatic Sciences* 53, 1580–94.

Winter, T.C. 1992: A physiographic and climatic framework for hydrologic studies of wetlands. In Roberts, R.D., and Bothwell, M.L., editors, *Aquatic ecosytems in semi-arid regions: implications for resource management,* N.H.R.I. Symposium Series 7, Environment Canada, Saskatoon, Saskatchewan, 127–48.

—— 1995: Hydrological processes and the water budget of lakes. In Lerman, A., Imboden, D.M. and Gat, J.R., editors, *Physics and chemistry of lakes.* New York: Springer Verlag, 37–62.

Xia, J. Haskell, B.J., Engstrom, D.R. and **Ito, E.** 1997: Holocene climate reconstructions from tandem trace-element and stable-isotope composition of ostracodes from Coldwater Lake, North Dakota, USA. *Journal of Paleolimnology* 17, 85–100.

Yansa, C.H. 1995: An early postglacial record of vegetation change in southern Saskatchewan, Canada, MSc thesis, University of Saskatchewan.