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ORIGINAL PAPER

Post-glacial climate change and its effect on a shallow dimictic lake in Nova Scotia, Canada

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Abstract A high-resolution, multi-proxy lake sediment record was used to establish the timing of Holocene environmental change in Canoran Lake, southwest Nova Scotia, Canada. Proxies include %C, $\delta^{15}N$, $\delta^{13}C$, HI, magnetic susceptibility, and pollen. Canoran Lake is a small, shallow (11 m) lake with two ephemeral inlets and an outlet. The site was deglaciated at ca. 15,300 cal (calibrated) year BP and elevated %C values indicate the establishment of a productive aquatic environment that is consistent with Allerød warming. The Allerød was interrupted by rapid air temperature cooling during the Younger Dryas (ca. 12,900–11,600 cal year BP). The Early Hypsithermal (ca. 11,600–8,500 cal year BP) was relatively warm and wet. A slight increase in clastic input occurred between 9,100 and 8,500 cal year BP but $\delta^{15}N$, $\delta^{13}C$, and HI values imply that the lithostratigraphic response may not be indicative of climate-induced change. The strong proxy response between 8,500 and 8,000 cal -year BP was likely due to cooling and drying coincident with the 8.2 k year event. The climate was relatively warm and dry during the Late Hypsithermal (ca. 8,000–3,500 cal year BP). None of the proxies' exhibit notable change during the 5,500 cal year BP hemlock decline, indicating that ecological change was likely due to a pathogen attack. Post-Hypsithermal (modern) climate was characterized by an increase in precipitation and a decrease in air temperatures from ca. 3,500 to 700 cal year BP (top of core).

Keywords Climate change · Paleolimnology · Nova Scotia · Stable isotopes · Multi-proxy · Hydrogen index · Limnology

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Introduction

A multi-proxy, chemostratigraphic approach was used to resolve centennial-scale, post-glacial paleoenvironmental change in southwestern Nova Scotia. This record is significant because few high-resolution paleoenvironmental studies have been carried out in the region, although many studies have noted that past environmental change had a significant ecological effect (e.g. Railton 1973; Wilson et al. 1993; McCarthy et al. 1995). Innovative proxies ($\delta^{15}N$, $\delta^{13}C$, and HI) are employed in an attempt to better constrain the nature of post-glacial environmental change and are used to resolve both moisture and temperature variability during the Holocene.

What is known about post-glacial climate change in Nova Scotia is largely based on biostratigraphic records, which will be briefly summarized. By 13,000 cal year BP, Nova Scotia was virtually ice-free (Mott 1994). Climatic warming during the Allerød and the establishment and migration of vegetation led to the stabilization of slopes and shorelines, decreased minerogenic input into lakes, and increased lacustrine productivity and sediment accumulation rates (Mott and Stea 1994).

Younger Dryas cooling (ca. 12,900–11,600 cal year BP, Mayle et al. 1993) resulted in the rejuvenation of local ice caps (Stea and Mott 1998). At that time, tree populations declined, herb and shrub populations increased, and many sites returned to open, tundra communities (Mott 1994). A Younger Dryas-equivalent, mineral sediment oscillation has been documented in many lakes throughout Nova Scotia (Stea and Mott 1998). Stea and Mott (1998) suggest that basin morphology and the distance of lakes from remnant ice and snow cover may account for the variability in lake response to Younger Dryas cooling. Spooner (1998) suggested that the remnant ice cover theory proposed by Stea and Mott (1998) may not have to be invoked in southwest Nova Scotia because change in these lake environments may be due to longer periods of seasonal lake ice cover and lower within-lake productivity. Whitney et al. (2005) and Mayle et al. (1993) indicate that temperatures in some lakes in Atlantic Canada may have dropped by as much as 6°C during the Younger Dryas.

Following Younger Dryas cooling, boreal forests were re-established and air temperatures were thought to be relatively cool or similar to present (Railton 1973). Anderson et al. (2007) report that regional cooling occurred at 11,200–10,900 cal year BP that was coincident with the Preboreal oscillation recognized in many places around the north Atlantic. At that time, sites in eastern New Brunswick, Prince Edward Island, and northern Nova Scotia show a lingering persistence of *Picea* and delay in arrival of *Pinus*, while in western and southwestern Newfoundland tundra vegetation persisted at high elevations (Anderson et al. 2007).

A 400 year cooling period (8,400–8,000 cal year BP, 8.2 ka cooling event) was likely related to a disruption in the North Atlantic thermohaline circulation (Barber et al. 1999; Anderson et al. 2007). Evidence for coincident cooling in northeastern North America has been mixed: loss on ignition data (proxy for organic content), diatom, and lithological change imply regional cooling around 8,400 cal year BP, but chironomids did not react strongly to this cool period (Spooner et al. 2002, 2005; Cwynar et al. 2003).

The time interval between ca. 8,000 and 3,000 cal year BP (Late Hypsithermal) was characterized by relatively warm and dry conditions (Jetté and Mott 1995; Gajewski et al. 2000; Williams et al. 2004). Hardwoods expanded and hemlock (*Tsuga canadensis*) arrived in Nova Scotia around 7,800 cal year BP (Jetté and Mott 1995). Vegetation distribution and composition became relatively stable from 7,800 to 5,800 cal year BP. However, the cause of a decline in hemlock at 5,500 cal year BP is unclear and both a pathogen attack (Davis 1981) and regional or hemispheric drought (Green 1987; Haas and McAndrews 2000) have been proposed.

The relatively warm, dry conditions of the Late Hypsithermal were followed by wetter and perhaps cooler conditions during the Post Hypsithermal, however, the timing of this transition is unclear (Railton 1973, 1975a, b). A transition from a relatively warm/dry climate to cool/wet, present-day climate has been assigned dates as early as 5,800 cal year BP (Railton 1975a) and as late as 2,000 cal year BP in offshore and onshore paleoenvironmental studies (McCarthy et al. 1995; Levac 2001).

In light of the research already undertaken in the region, this study was designed to determine if: (1) a suite of chemostratigraphic proxies is capable of tracking changes in the moisture and temperature state at the study site, and (2) a well-dated and high-resolution record would better define the presence, absence, and timing of regional climate transitions.

Study site

Canoran Lake (44°36′N, 64°34′W; Fig. 1) is situated in Lunenburg county, Nova Scotia, Canada, 33 km inland of the Atlantic ocean. The study site is located 100 m above sea level and it is in a topographically subdued area where watershed relief does not exceed 45 m (Fig. 1). The present-day climate of Nova Scotia is east-coast northern temperate (Ritchie 1987).

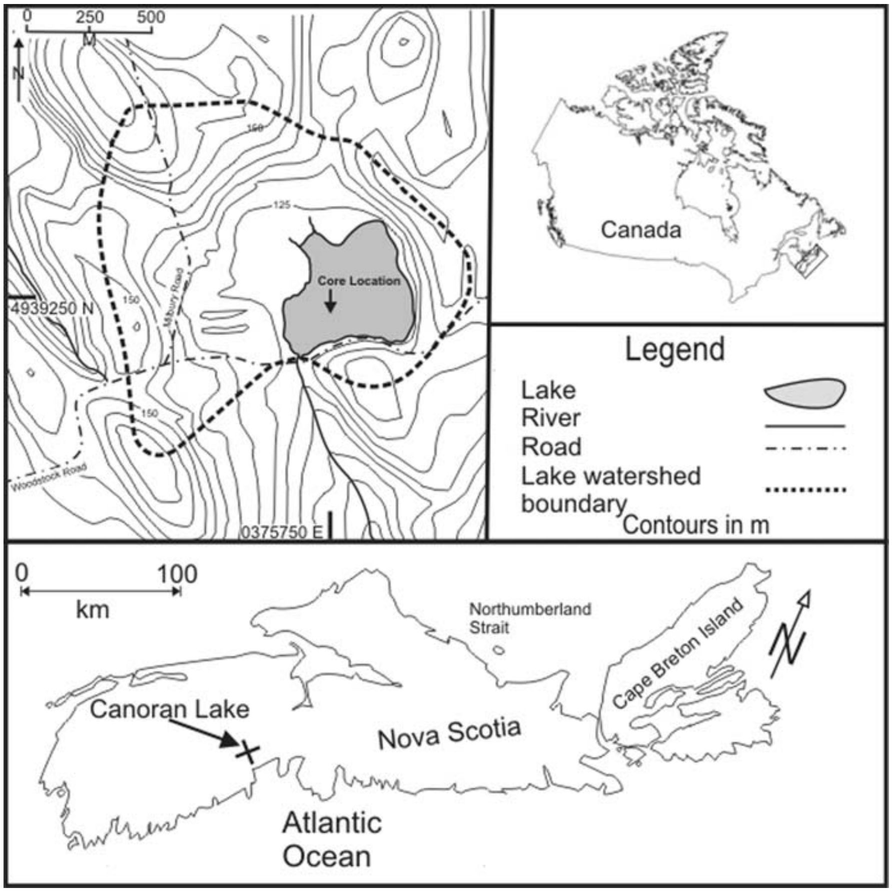

Fig. 1 Location of Canoran Lake, Lunenberg County, Nova Scotia, Canada. Canoran Lake has a relatively small watershed. It is considered to be a headwater lake. Also shown is the location of the percussion and gravity cores used for analysis (Core 5)

Canoran Lake is shallow (11 m maximum depth) and has a low relative depth (2%), resulting in high sensitivity to air temperature change. The lake has two ephemeral inlets to the northwest, which makes it highly sensitive to precipitation changes. A N–S trending sill separates two profundal zones and a discontinuous, largely unvegetated littoral zone. Lake level is controlled by a sill at the outlet.

Canoran Lake is currently oligotrophic (total phosphorus = 3 ± 0.2 μg/l; chlorophyll $a = 0.37 \pm 1.90$ mg/l; Secchi disk depth = 3.92 ± 0.39 m). The lake is dimictic, with a hypolimnion that is generally well oxygenated. The average pH is 5.2, alkalinity is 0 mg/l, and hardness is 0 mg/l.

Methods

A 50 kHz King 1570 echo sounder was used to quantify the bathymetry and sediment distribution in Canoran Lake and to select a core site. Canoran Lake was cored through the ice in March 2005 using a portable percussion coring system (Reasoner 1993). Five 7.5 cm diameter cores were obtained. Cores were frozen and later split with a diamond rock saw. All analyzes were carried out on Core 5 because this core was taken in the basin center and is most representative of within-lake paleoenvironmental change. Core 5 (~3.8 m) was separated into visually distinguishable stratigraphic units based on grain size, color, and composition. The top 47 cm of the paleoenvironmental record (c.a. 700 cal year BP to present) is not discussed in this paper, as percussion coring tends to disturb near surface sediment. A gravity core of this portion of the paleoenvironmental record was obtained and is interpreted and discussed in Lennox (2006).

The chronology of Core 5 was constrained by seven ^{14}C accelerator mass spectrometer (AMS) dates that were analyzed at the NSF-Arizona AMS facility, University of Arizona. The conversion to calendar years from radiocarbon years was accomplished using CALIB 4.2 (Stuiver and Reimer 1998). All AMS dates were performed on terrestrial material that was sub-sampled at 1 cm intervals (Table 1).

Magnetic susceptibility is largely dependent on the relative amount of terrestrial magnetic minerals (Nowaczyk 2001). Two sets of 24 magnetic susceptibility values were measured and averaged at a 1-cm resolution using a KT-9 Kappameter® susceptibility meter. To measure the experimental error of magnetic susceptibility values, seven sets of 24 magnetic susceptibility readings were measured and averaged every 5 cm.

Water content is a measure of sediment compaction and is used to identify shifts in sediment lithology and composition (Dean 1974). Water content was calculated as the percent weight of the wet sample lost as a result of drying (Dean 1974). Triplicates for estimation of error were taken every 20 cm.

Preparation and analysis of %C and %N were done in conjunction with the sample preparation for carbon and nitrogen stable isotopes. The samples were not acidified to remove carbonate material because previous analysis indicated that there was no carbonate in the lake sediments (LOI method; Dean 1974; Lennox 2006) and the geological units in the watershed are carbonate-poor. Thirty samples were processed at the University of Saskatchewan Stable Isotope Laboratory, Saskatoon, Saskatchewan and 63 samples were analyzed at Iso-Analytical Limited, Sandbach, Cheshire, England. EA-IRMS was used at both facilities. Stable isotope values were obtained using a Thermo Finnigan Flash 1112 EA coupled to a Thermo Finnigan Delta Plus XL through a Conflo III at the University of Saskatchewan Stable Isotope Laboratory and a Europa Scientific RoboPrep-CN elemental analyzer coupled with a Europa Scientific 20-20 IRMS at Iso-Analytical Ltd. Seven percent of the samples were analyzed 2–4 times to estimate error (Lennox 2006).

The analytical techniques and application of the pyrolysis-based Hydrogen index (HI) have been outlined by others (Peters 1986; Talbot and Livingston 1989). Analyzes were carried out at the Geological Society of Canada (GSC) Calgary organic geochemistry laboratory using an HP5890 Rock-Eval instrument. Duplicates for error analysis were analyzed for 11% of the samples.

Geochemical background

Our reconstruction of post-glacial paleoenvironmental change in Canoran Lake is predominantly based on %Carbon (%C), %Carbon/%Nitrogen (C/N weight) ratios, carbon stable isotopes ($\delta^{13}C$), nitrogen stable isotopes ($\delta^{15}N$), and hydrogen index (HI). In this section we briefly outline the major controls on each of these proxies that are relevant to the interpretation of the paleoenvironmental record.

Percent C (%C) values are influenced by changes in primary productivity, clastic input, and diagenesis (Meyers and Teranes 2001); these factors are largely dependent on the distance from the shoreline to the site of deposition (i.e. lake water levels; Talbot and Laerdal 2000). Percent loss on ignition (%LOI) data of cores from near-shore to basin center locations in Canoran Lake implies that %C values are higher in near-shore than basin center locations (Martin 2003). If the lake water level falls, then the contribution of C-rich terrestrial organic matter increases in Core 5

Table 1 Radiocarbon data[a] (Lennox et al. 2008)

Depth (cm)	Age (^{14}C year BP)	Calendar Date (2σ) [cal. year BP]	Calendar Range (2σ) [cal. year BP]	Material	δ^{13}C
47	985 ± 41	880	794–962	Leaves	−26.3
120	3,234 ± 89	3,480	3,262–3,687	Organics	−28.0
170	4,600 ± 39	5,270	5,068–5,466	Wood	−26.6
197	5.268 ± 48	6,060	5,930–6,182	Wood	−25.0
271	6,931 ± 54	7,800	7,668–7,924	Wood	−26.8
322	8,282 ± 48	9,280	9,126–9,434	Seed	−23.6
353	10,873 ± 69	12,870	12,793–12,942	Charcoal	−25.6

[a] All radiocarbon analyzes were performed at the University of Arizona (AA) AMS Laboratory, Tuscon, AZ and were not corrected. Calibrated ages are given as mean calendar years (Stuiver and Reimer 1998). Radiocarbon data are from a ~3.8 m core from Canoran Lake, Nova Scotia, Canada

because the distance of transport from terrestrial organic matter on the shorelines to Core 5 decreases.

The relative input of aquatic and terrestrial organic matter can be roughly distinguished by the characteristic %Carbon/%Nitrogen (C/N) ratios of aquatic and terrestrial sources. Algal matter is primarily composed of N-rich lipids and proteins (C/N $\leq$ 10), while terrestrial organic matter is primarily composed of C-rich lignin and cellulose (C/N $\geq$ 20; Meyers and Lallier-Vergès 1999; Talbot 2001). C/N ratios from 10 to 20 represent mixed aquatic and terrestrial matter.

Stable carbon isotope values (δ^{13}C) are a proxy for carbon cycling. δ^{13}C values are forced by the rate of dissolved inorganic carbon (DIC) uptake during photosynthesis (preferential uptake of ^{12}C) and the DIC source such as the atmosphere, bedrock carbon, terrestrial vegetation, etc. (Meyers and Teranes 2001). When (1) the rate of DIC uptake increases due to higher water temperatures or (2) the uptake of watershed bicarbonate ($\delta^{13}C = 1‰$) increases compared to the uptake of atmosphere CO_2 ($\delta^{13}C = -7‰$), δ^{13}C values ($\sim{}^{13}C/{}^{12}C$) increase (Talbot and Laerdal 2000, Meyers and Teranes 2001). DIC uptake is generally a minor control on δ^{13}C values of Canoran Lake sediments because the bedrock and surficial sediments in its watershed are carbonate and bicarbonate poor. Consequently, the DIC source for primary producers was generally constant and δ^{13}C trends are likely related to the rate of DIC uptake, which is influenced by air and epilimnetic water temperatures (Turney 1999; Meyers and Teranes 2001).

Stable nitrogen isotope values (δ^{15}N) are interpreted to be related to the abundance of littoral wetland communities (Talbot and Laerdal 2000). For example, an increase in the effective moisture and a coincident water table level increase favor littoral wetland communities because the littoral volume increases. An increase in littoral wetland communities results in an increase in the rate of dissolved inorganic nitrogen (DIN) uptake and more positive δ^{15}N values because microbial nitrification and denitrification processes are ten times greater in vegetated soil than non-vegetated soil (Reddy et al. 1989). Lake level lowering leads to a reduction in the littoral zone volume (Fig. 2) and terrestrial vegetation occupies the previously submerged sediment. This interpretation implies that cyanobacteria abundance

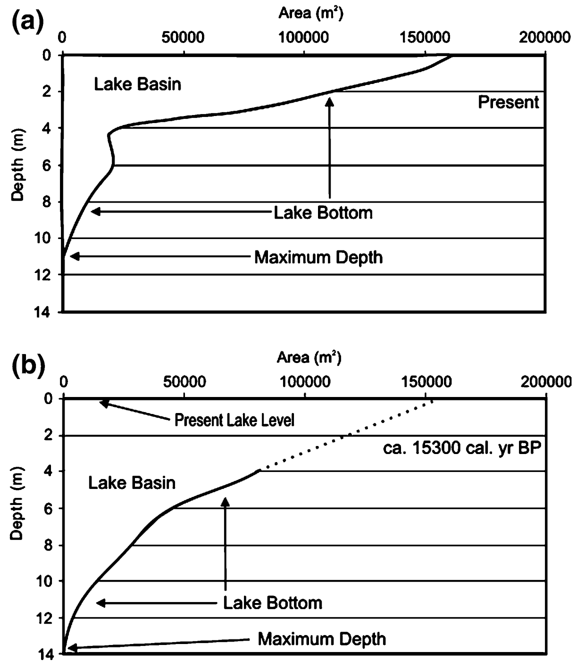

Fig. 2 Hypsographic curves for **a** present-day and **b** nascent bathymetry of Canoran Lake imply that an increase in lake water levels has resulted in an increase in littoral habitat since it's the lake's inception. These graphs were generated using sub-bottom profiling and echo sounder data. *Dotted line* in graph **b** indicates hypothetical lake bottom profile

had little influence on δ^{15}N values. This assumption is reasonable because cyanobacteria biomass is low in nutrient-poor lakes such as Canoran Lake (Downing et al. 2001). There is typically a surplus of DIN in freshwater lakes such as Canoran Lake and primary productivity is not predominantly controlled by variations in DIN input (Wetzel 1983).

Hydrogen index (HI) is dependent on two factors: (1) the type of organic matter, and (2) the degree of oxidative degradation of organic matter (Talbot and Livingston 1989). Algal organic matter is generally more hydrogen rich than herbaceous and woody matter (Talbot and Livingston 1989). Consequently, algal organic matter has high HI values. If the type of organic matter is consistent, warmer air temperatures result in: an increase in organic matter delivery to the hypolimnion; an increase in respiration by decomposers; and more positive HI values. Additionally, warmer air temperatures result in prolonged summer stratification and depleted hypolimnetic dissolved oxygen concentrations (more positive HI).

Results

Four distinct lithostratigraphic units were recognized visually. The basal unit (Unit 1; 381–370 cm) is a brown-yellow, well-indurated diamicton with rare angular pebble clasts. Unit 1 grades into a brown-black gyttja (370–364 cm; Unit 2). The contact between Unit 2 and the overlying clay gyttja (364–356 cm; Unit 3) is diffuse. Unit 3 is sharply overlain by a highly organic, brown-black gyttja with no visually identifiable minerogenic layers (324–47 cm; Unit 4).

Magnetic susceptibility is generally low in Core 5 (mean magnetic susceptibility = 0.00 SI units; $2\sigma = 0.02$ SI units). Relatively large amplitude magnetic susceptibility anomalies coincide with clastic-dominated stratigraphic intervals, including Unit 1 (diamicton) and Unit 3 (clay gyttja).

The mean water content for the entire lake sediment record is 79%, with values generally stabilizing and increasing up-core (Fig. 3). Water content changes with lithology: negative excursions of water content were recorded for clastic intervals that include Unit 1 (64%; diamicton), Unit 3 (56%; clay gyttja), and between ca. 9,100 and 8,400 cal year BP (71%). The analytical error of duplicate water content measurements is low (average $2\sigma = 2\%$; $N = 19$).

Percent C values were generally lower and more variable from ca. 13,000 to 8,400 cal year BP than from ca. 8,000 cal year BP to the top of the core. A short, negative excursion of %C values was recorded for Unit 3 and from ca. 9,100 to 8,400 cal year BP, which are minerogenic-rich intervals. A significant positive shift in %C values occurred between ca. 8,400 and 8,000 cal year BP. A linear regression of %C and %N intercepts the *y*-axis at 0.2%. Talbot (2001) has interpreted the y-intercept of similar records to reflect inorganically bound nitrogen derived from ammonium released by diagenesis or soil organic matter retained on clay minerals. In this study, the 2σ is as high as 0.4% and the experimental error is larger than the amount of inorganically bound nitrogen identified using methods outlined in Talbot (2001). Experimental error must be determined before applying an inorganic nitrogen correction. The inorganic nitrogen correction is not used in this study and the total nitrogen content likely resides in organic matter.

Mean C/N ratio for the entire lake sediment record is 11.1 ($2\sigma = 1.8$; $N = 62$). C/N ratios are variable from 5,300 to 3,500 cal year BP (range of C/N ratios = 9.0–13.8). The standard deviation of C/N ratios is low ($2\sigma = 1.0$; $N = 62$) from 3,500 cal year BP to the top of the core.

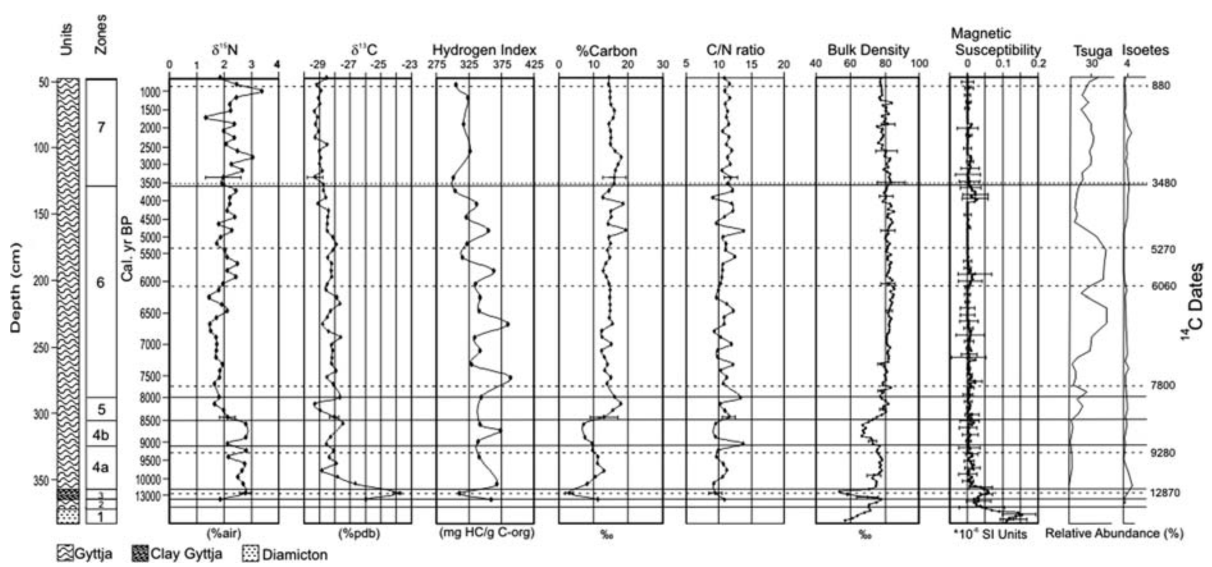

Fig. 3 Eight distinct zones were recognized based on the $\delta^{15}N$, $\delta^{13}C$, %C, C/N ratios, HI, water content, and magnetic susceptibility values in Core 5. Cross-correlation with previous pollen work on Canoran Lake (Railton 1973) was possible due to Unit 3 (Younger Dryas inorganic marker horizon). *Error bars* indicate 2σ

Nitrogen stable isotope values are generally high from ca. 13,000 to 8,400 cal year BP (Fig. 3). Between 8,400 and 8,000 cal year BP, $\delta^{15}N$ values gradually decrease. $\delta^{15}N$ values are stable and low from 8,000 to 3,500 cal year BP, with a low-amplitude, gradual, and positive excursion around 6,100 cal year BP. $\delta^{15}N$ values became more variable and positive from 3,500 cal year BP to the top of core.

The mean $\delta^{13}C$ value for the entire lake sediment record is −28.5‰ ($2\sigma = 0.2$‰), with values generally decreasing up-core (Fig. 2). There is an abrupt shift in $\delta^{13}C$ values from Unit 2 to Unit 3 of +2.1‰. Carbon stable isotope values gradually decrease above the Unit 3–Unit 4 transition and then remain constant and high from ca. 10,500 to 3,500 cal year BP except for a local minimum (−27.7‰) around 8,200 cal year BP. Up-core $\delta^{13}C$ values become progressively more negative from ca. 3,500 cal year BP—the top of the core. Based on typical $\delta^{13}C$ and C/N ratio values for aquatic and terrestrial organic matter, the organic matter in Core 5 is predominantly of aquatic origin (Fig. 4).

The mean HI value (332 mg HC/g TOC, $2\sigma = 56$ mg HC/g TOC) is low (Fig. 5). HI values are generally high from ca. 13,000 to 3,500 cal year BP and low from 3,500 cal. year BP to the top of the core. A ~50 mg HC/g TOC negative excursion of HI values occurs in Unit 3. All units contain organic matter derived from algal material as interpreted by a Van Krevelen diagram (Type II, Fig. 5).

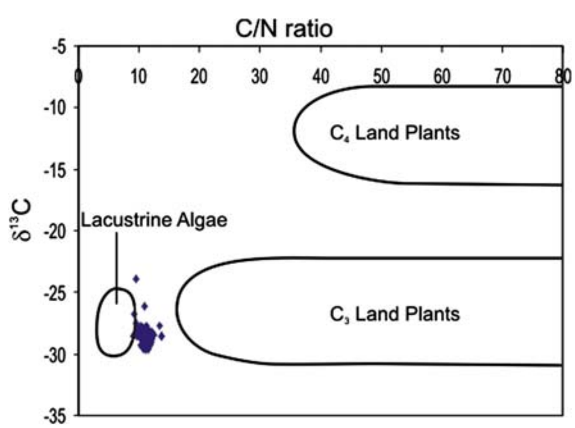

Fig. 4 Generalized $\delta^{13}C$ and C/N values of major sources of plant organic matter in lake sediments (Meyers and Lallier-Vergès 1994). *Plotted points* are from this study and indicate that organic matter in Canoran Lake sediment is predominantly from lacustrine algae with a minor contribution from C_3 land plants

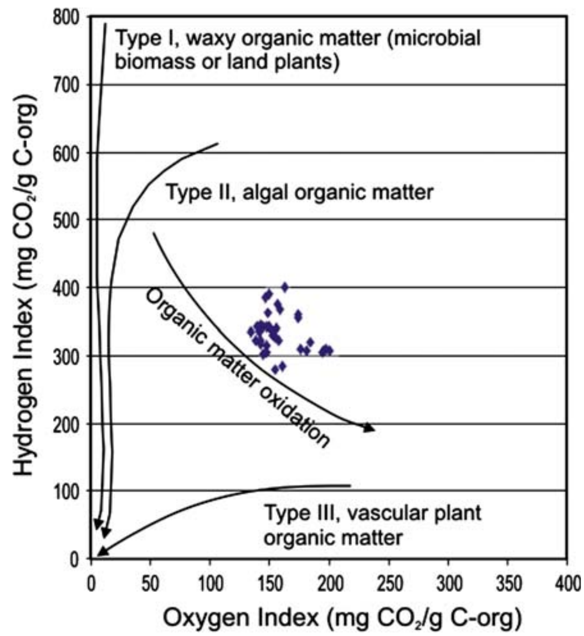

Fig. 5 Van Krevelen diagram of HI vs. OI for bulk organic matter in Core 5 (Meyers and Teranes 2001). *Plotted points* are from this study and indicate that the bulk organic matter of Canoran Lake is composed of highly oxidized, algal organic matter (Type II)

Interpretation

Zone 1 (diamicton, 381–370 cm, >15,300 cal year BP, late-glacial)

Zone 1 is a diamicton with rare angular pebble clasts that is interpreted to be glacial in origin. Railton (1973) surmised that the minimum age of Canoran Lake is 13,550 cal year BP (bulk date) and estimated that deglaciation of the lake basin occurred around ca. 15,300 cal year BP.

Zone 2 (gyttja; 370–364 cm, ca. 15,300–12,900 cal year BP, Allerød)

In Zone 2, gyttja was deposited in the newly formed lake basin around 15,300 cal year BP. Tundra vegetation was replaced by woodland vegetation in Zone 2 (Allerød climatic warming; Railton 1973; Mayle et al. 1993). Limited proxy data were collected for Zone 2 in this study. The response at ca.

13,000 cal year BP indicates that the climate was relatively warm (high $\delta^{13}C$, high HI) and possibly wet (high $\delta^{15}N$).

Zone 3 (364–356 cm, clay gyttja, ca.12,900–11,600 cal year BP, Younger Dryas)

Air temperatures cooled rapidly and pine-dominated (*Pinus* spp.) forests were replaced by a tundra-like vegetation assemblage during the Younger Dryas (Stea and Mott 1998). The deposition of clay to sandy gyttja (Unit 3) was likely due to: (1) the dilution of organic material by mineral sediment (high magnetic susceptibility), (2) a decrease in within-lake productivity (low %C), and/or (3) a decrease in terrestrial organic matter deposition (low C/N ratio; Fig. 3).

The anomalously high $\delta^{13}C$ value is indicative of a change in the DIC source (Fig. 3). Though permanent ice cover has been proposed for some lakes during the Younger Dryas, it is unconvincing as an explanation for high $\delta^{13}C$ because published $\delta^{13}C$ values of $-24‰$ and $\delta^{15}N \sim 2‰$ are relatively common in paleoclimate records of dimictic lakes in temperate and subtropical climates (e.g. Talbot and Livingston 1989; Brenner et al. 1999; Talbot 2001). The high $\delta^{13}C$ value is certainly not due to climatic warming based on the terrestrial vegetation response and buried peat deposits in Nova Scotia (Stea and Mott 1998).

The herb- and shrub-dominated vegetation cover and poor soil development during the Younger Dryas likely favoured silicate weathering (Turney 1999). The shift from a bicarbonate- to a silicate-derived DIC source may have resulted in an increase in the uptake of isotopically heavy bicarbonate by primary producers ($\delta^{13}C$ for silicate weathering $= +2.8‰$, $\delta^{13}C$ for dissolved atmospheric carbon $= -7‰$; Turney 1999; Meyers and Teranes 2001). A possible source for silicate-derived DIC was the clays being deposited within the lake at this time (Turney 1999).

A significant decrease in HI values may indicate an increase in mixing due to wind shear associated with katabatic winds originating from re-activated remnant ice in northern Nova Scotia or a decrease in the strength of summer stratification (Fig. 3; Stea and Mott 1998). Nitrogen stable isotope values had low variability before, during, and after deposition of Unit 3, implying that DIN cycling within the lake did not vary (constant wetland communities) and lake levels were relatively constant (Fig. 3). The proxy response is consistent with the Younger Dryas being a period of relatively cool and stormy weather that was coincident with a shift in terrestrial vegetation and within-lake primary productivity.

Zone 4a (gyttja, 356–324 cm, ca.11,600–9,100 cal year BP, Early Hypsithermal)

Ruddiman (2001) has indicated that summer insolation in the northern hemisphere was as much as 8% higher than present during the Early Hypsithermal. Various researchers have indicated that the Early Hypsithermal in Nova Scotia was characterized by present-day air temperatures (Railton 1975b; McCarthy et al. 1995), possibly high precipitation (Railton 1973; Martin 2005), and sea surface temperatures (SST) higher than present (Levac 2001).

As air temperatures ameliorated, boreal forests replaced transitional shrubs and herbs in Zone 4a (Early Hypsithermal) around 10,500 cal year BP (Railton 1973). Percent C values increase with time from ca. 11,600 to 10,500 cal year BP and are indicative of an increase in within-lake organic matter production and/or a decrease in minerogenic material. Percent C values stabilized after ca. 10,500 cal year BP (Fig. 3). $\delta^{13}C$ values decreased rapidly from the beginning of Zone 4a to ca. 10,500 cal year BP (Fig. 3). Isotopically heavy bicarbonate derived from silicate weathering was no longer a major DIC source after ca. 10,500 cal year BP, indicating a stable landscape (Turney 1999).

$\delta^{15}N$ values were high and did not change markedly during the Younger Dryas (Zone 3) to Early Hypsithermal (Zone 4) transition (Fig. 3). These data suggest that littoral wetland communities were well established and water table levels were high at this time. High HI values imply less wind-induced mixing and rapid and prolonged stratification, that may have led to oxygen depletion in the hypolimnion and the development of anaerobic conditions in the months before destratification.

Zone 4b (gyttja, 324–304 cm, ca. 9,100–8,500 cal year BP, Early Hypsithermal)

A major disruption of ocean thermohaline circulation resulted in the slowing of deep-water formation at ca. 8,800 cal year BP (Dyke and Prest 1987). Atmospheric circulation patterns reorganized and northeastern North America became relatively cooler and drier (Hu et al. 1999). Polar air mass incursions may have become more common and the mean position of the jet stream trough overlying northeastern North America may have shifted to a more southerly location (Hu et al. 1999).

The rate of DIC uptake was high and water temperatures during the ice-free season likely increased from ca. 9,100 to 8,500 cal year BP. Based on HI values, the hypolimnion was poorly oxygenated (Fig. 3). Carbon stable isotope (^{13}C) values and HI values of organic matter with C/N ratios <10 were similar to values recorded in Zone 4a (Fig. 3).

In Zone 4b, decreases in %C and water content values indicate that there was a subtle increase in clastic input from 9,100 to 8,500 cal year BP (Fig. 3). This increase in clastic deposition may have been due to a decrease in lake water levels, which would shorten the sediment transport distance from the shoreline to the basin center. Nitrogen stable isotope values of organic matter that is predominantly composed of aquatic organic matter (C/N ratios <10) are high and stable before and during the clastic oscillation, indicating lake water levels were likely similar to Zone 4a (Fig. 3). Aside from an increase in clastic input, the proxies used in this study imply that the paleoenvironmental conditions in Zone 4a were comparable to Zone 4b.

Zone 5 (gyttja, 304–286 cm, ca. 8,500–8,100 cal year BP, 8.2 k year event)

The 8.2 k year event was the largest-magnitude climate change event recorded in Greenland ice cores during the Holocene (Alley et al. 1997). Thermohaline circulation was disrupted between ca. 8,400 and 8,100 cal year BP (Barber et al. 1999) and the subsequent reorganization of atmospheric circulation resulted in relatively cool, dry, and windy conditions throughout continental North America (Stager and Mayewski 1997) and cool air temperatures in Nova Scotia (Spooner et al. 2005).

A rapid negative excursion of δ^{13}C values at ca. 8,400 cal year BP (Zone 5) is likely due to a decrease in the rate of DIC uptake as water temperatures cooled rapidly over a 300-year time interval (Fig. 3). The negative excursion of δ^{13}C values in Zone 5 was the most rapid and largest-magnitude excursion of δ^{13}C values in Core 5 during the Holocene. δ^{13}C values became more positive at ca. 8,100 cal year BP and were relatively stable for the remainder of the Holocene, indicating that the rate of DIC uptake increased rapidly at ca. 8,100 cal year BP due to rapidly ameliorating air temperatures.

Nitrogen stable isotope values decreased rapidly in Zone 5 by 0.8% (Fig. 3). Lake level lowering due to a decrease in effective moisture likely caused a reduction in wetland communities and the amount of N that wetland microbial populations fixed. Lake level lowering resulted in an increase in C-rich terrestrial organic matter (%C) and more positive C/N ratios (Fig. 3). The δ^{15}N values the lake level may not have returned to pre-Zone 5 levels until 2,600 cal year BP (Post Hypsithermal; Fig. 3).

The above proxy response indicates that there was a strong within-lake reaction at Canoran Lake to the 8.2 k year event cooling, and a decrease in effective moisture. The pollen record from Canoran Lake indicates that (1) the terrestrial vegetation did not react strongly to the 8.2 k year event, or (2) the resolution of the pollen record is too coarse to detect change (Fig. 3). Similarly, the sampling resolution for HI was too low to determine how the 8.2 k year event affected the mixing of Canoran Lake (Fig. 3).

Zone 6 (gyttja, 286–128 cm, 8,100–3,500 cal year BP, Late Hypsithermal)

From 8,100 to 3,500 cal year BP, summer insolation was high and eastern North America was relatively warm (Ruddiman 2001). Harrison (1989) concluded that Holocene lake levels in continental locations throughout eastern North America were at their lowest levels at 6,200 cal year BP and precipitation was low. A shift from boreal to Carolina Forest (8,000–3,500 cal year BP; Railton 1973) is coincident with the Zone 5 to Zone 6 transition.

At the Canoran Lake site, water temperatures were high (δ^{13}C) and hypolimnetic oxygen concentrations were low (high HI; Fig. 3). High HI values indicate a decrease in the duration of water column mixing, which likely led to an increase in the depth or duration of summer stratification, and a coincident decrease in hypolimnion dissolved oxygen (Fig. 3).

Low δ^{15}N values indicate a decrease in water table levels, shoreline wetlands, and microbial uptake of N (Fig. 3). The highest %C values in the post-glacial record are recorded during the Late Hypsithermal and are likely a consequence of low lake levels (Fig. 3). Railton (1973) suggested that there were minor alder (*Alnus* spp.) and Myrica (*Myrica* spp.) wetlands surrounding the lake at this time. Lake levels were low based on δ^{15}N values, %C values, and low quillwort counts (*Isoetes* spp.; Fig. 3; Railton 1973).

The 5,500 cal year BP hemlock decline observed both regionally and in the Canoran Lake pollen record has been alternately attributed to either a pathogen attack (Davis 1981) or drought (Haas et al. 1998; Haas and McAndrews 2000). There is little within-lake proxy evidence for drought at Canoran Lake. If drought did occur, lake levels would be expected to have remained low. Nitrogen stable isotope values were constant and there was not a rapid decrease in the lake level that was synchronous with the 5,500 cal year BP hemlock decline (Fig. 3). Clastic input was low based on magnetic susceptibility values. Percent C, %N, and HI values were relatively constant during the 5,500 cal year BP hemlock decline and within-lake environmental conditions do not appear to have been affected (Fig. 3). These findings suggest that drought did not cause the decline; the alternative hypothesis, pathogen attack, seems more likely (Davis 1981).

Zone 7 (128–47 cm, 3,500 cal. year BP—top of core, gyttja, Post Hypsithermal)

A decrease in summer insolation during the Post Hypsithermal resulted in summer temperatures ~1–2°C cooler than present (Ruddiman 2001). The transition to a wetter and relatively cooler climate was assigned dates from 5,100 to 1,800 cal year BP (Railton 1973, 1975a). Railton (1975b) assigned a poorly constrained date of 2,900 ^{14}C year BP (ca. 3,000 cal year BP) to a pollen assemblage change indicative of a transition to relatively cooler, moister conditions. Pollen records indicate that the Post Hypsithermal was characterized by relatively wet and cool conditions (increasing spruce; Railton 1973; Green 1987; Ogden 1987). Railton (1973) surmised that an increase in quillwort (*Isoetes*), an early colonizer in shallow water, was due to rising water tables (Fig. 3). A dinocyst reconstruction indicates that the sea surface temperatures (SSTs) and salinity on the Scotian Shelf did not change markedly during the Late Hypsithermal to Post Hypsithermal transition (Levac 2001).

Low δ^{13}C values in Zone 7 are likely due to a reduction in the rate of DIC uptake associated with lower water temperatures (Fig. 3). HI values display a negative trend which implies that the oxidative degradation of organic matter was high and the lake was well oxygenated, possibly due to lower air temperatures (Fig. 3).

High quillwort counts coincided with more positive nitrogen stable isotope values than at the beginning of the Late Hypsithermal (Zone 6; Fig. 3). At present, Canoran Lake has a large littoral zone; an increase in lake water level would have substantially increased shallow-water environments that would have been well suited to wetland formation and wetland microbial communities, thereby promoting an increase in uptake of DIN (Talbot and Laerdal 2000). δ^{15}N values are variable in Zone 7 (Fig. 3). This variability in δ^{15}N values may be indicative of fluctuating lake water levels. Alternately, low δ^{15}N values may correspond to time periods when nitrogen was abundant, but phosphorus limited phytoplankton growth (Meyers and Teranes 2001).

HI values suggest that increased water table levels may have resulted in more oxidation of terrestrial material by littoral wave action because the volume of shallow water environments and the distance from the shoreline to the basin center likely increased. Variable C/N ratios at the transition between the Late Hypsithermal (Zone 6) and Zone 7 may reflect an increase in terrestrial organic matter input (Fig. 3).

Discussion

The sediment record from Canoran Lake provides Holocene paleoclimate information at higher temporal resolution than was available from previous work in Nova Scotia. Climate change inferred from both pollen records (e.g. Railton 1973, 1975b; Ogden 1987) and lithostratigraphic records (Spooner et al. 2005) are in agreement with the interpretation of the proxy records in this study. A number of major climatic shifts recognized by Railton (1975b) and others are roughly synchronous with, and of the same magnitude and duration as paleoenvironmental changes recorded in the lake sediment record. Notable exceptions and/or time periods that have been poorly studied include the unnamed cool and dry climate interval from 9,100 to 8,500 cal year BP

(Zone 4a; Fig. 3), the 8.2 k year event (Zone 5; Fig. 3), and the Late Hypsithermal to Post Hypsithermal transition (Zone 6–8; Fig. 3).

An oscillation in clastic content accompanied the 8.2 k year event in two other lakes in Nova Scotia (Taylor Lake, Spooner et al. 2002, 2005; Piper Lake). Zone 4b in Canoran Lake appears to record this event. However, if linear sedimentation rates are assumed, then Zone 4b occurred between 9,100 and 8,500 cal year BP, which is well before the 8.2 k year event. Percent C and water content values indicate that a subtle increase in clastic input occurred between 9,280 and 7,800 cal year BP. Spooner et al. (2005) observed a clastic oscillation at the Piper Lake site between the Younger Dryas (Zone 3) and the 8.2 k year event (Zone 5), but the timing of the clastic oscillation observed by Spooner et al. (2005) at Piper Lake was earlier (10,800–10,300 cal year BP) than Zone 4b (9,100–8,500 cal year BP) in Canoran Lake. Spooner et al. (2005) suggested that the clastic oscillation in Piper Lake was due to lake level lowering. Given the apparent increase in clastic material, a similar process might be invoked for Zone 4b in Canoran Lake. However, δ^{15}N values imply that lake level lowering probably did not occur. Hu et al. (1999) proposed that the polar jet stream trough overlying Nova Scotia shifted to a more southern location around 8,900–8,500 cal year BP, resulting in a negative excursion of δ^{18}O and δ^{13}C values in a varved lake record in Minnesota, which is in agreement with both the timing and lake response in this study. However, δ^{13}C and HI values in Canoran Lake are constant and unaffected by the environmental change responsible for the increase in clastic input. The reasons for the shift in clastic input from Zone 4a to Zone 4b remain enigmatic, as the majority of environmental proxies indicate that conditions during Zone 4b were similar to those in Zone 4a.

Climate proxies in Zone 5 (8,500–8,100 cal year BP) are indicative of cooling that is coincident with the timing and duration of the 8.2 kyr event (e.g. Alley et al. 1997; Stager and Mayewski 1997; Spooner et al. 2005). Water temperatures rapidly decreased (inferred from δ^{13}C values) and the climate became increasingly arid (δ^{15}N) in southwest Nova Scotia. As the lake water level declined, the distance between the core location and the shoreline was reduced (relatively constant and low δ^{15}N values). Consequently, the amount of terrestrial input increased in Zone 5 and remained high in Zone 6 (C/N ratios; %C). This proxy record supports the assertion that a disruption in thermohaline circulation caused reorganization of atmospheric circulation (Stager and Mayewski 1997), and relatively cool air temperatures in Nova Scotia (Spooner et al. 2005). Cool and arid conditions in Zone 6 are likely associated with the southward displacement of the polar jet stream trough overlying Nova Scotia.

The transition from the Late Hypsithermal (Zone 6) to Post Hypsithermal (Zone 7) in the Maritimes is not well recognized or dated (typically 1–3 radiocarbon dates; Jetté and Mott 1995). In this study, the autochthonous proxy record indicates that this transition likely occurred at about 3,500 cal year BP (Fig. 3). However, the transition in the pollen record of Canoran Lake occurs significantly later (800 years) than the autochthonous proxy response (Railton 1973). This apparent lag is expected, as the life spans of phytoplankton are short compared to those of terrestrial vegetation. Additionally, terrestrial vegetation assemblages react to climate change more slowly than aquatic assemblages due to lags in dispersal (Webb 1986) and soil development (Pennington 1986). The duration of the lag in the pollen response to relatively cool and wet conditions in the Post Hypsithermal is in agreement with modeling (Solomon and Bartlein 1992) and pollen records (Campbell and McAndrews 1993) of climate transitions in other regions.

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