A multi-proxy lithostratigraphic record of Late Glacial and Holocene climate variability from Piper Lake, Nova Scotia

Ian S. Spooner, Ian MacDonald, Brandon Beierle, and A.J. Timothy Jull

Abstract: A multi-proxy lithostratigraphic record from Piper Lake, Nova Scotia reveals environmental variability during the Late Glacial and Holocene. Piper Lake is a small, shallow (3 m), closed dystrophic basin located in the eastern Nova Scotia Highlands. The site was deglaciated about 14.5 cal (calibrated) ka BP and elevated loss on ignition values and relatively low carbon/nitrogen (C/N) isotope ratios indicate the establishment of a productive aquatic environment consistent with Allerød warming. The Late Glacial Lake record is punctuated by two thin, very fine-grained clay layers that are correlative to the Killarney and Younger Dryas (YD) oscillations; they were deposited when perennial ice covered the lake. The post-YD lithostratigraphy indicates the rapid establishment of an increasingly productive and stable landscape. This trend is reversed three times during the Holocene by minerogenic units. A complex 25 cm thick diamicton unique to Piper Lake was deposited ca. 10.8–10.3 cal ka BP by slumping that was associated with periglacial slope processes and (or) lake level changes; a direct correlation to early Holocene (Preboreal) cooling appears unlikely. Two thin minerogenic units deposited at ca. 8.1 and ca. 4.9 cal ka BP were likely the result of regional cooling and are broadly correlative with events noted in the GISP2 (Greenland Ice Sheet Project 2) ice-core record. The Holocene lithostratigraphic record from Piper Lake may be a consequence of unique limnological factors. Alternatively, the strong lithostratigraphic response may be the result of the absence of a strong and persistent regional climate mechanism (North Atlantic oscillation?), which if present might have obscured the impact of hemispheric or larger-scale climate forcing.

Résumé : Une séquence lithostratigraphique à indicateurs multiples du lac Piper, en Nouvelle-Écosse, révèle une variabilité environnementale au cours du Glaciaire tardif et de l'Holocène. Le lac Piper est un bassin dystrophe, fermé et peu profond, situé dans l'est des hautes-terres de la Nouvelle-Écosse. Le site a été déglacé il y a environ 14,5 ka cal. avant le présent et des valeurs élevées de perte par calcination et des rapports relativement faibles des isotopes carbone/azote indiquent l'établissement d'un environnement aquatique productif concordant avec le réchauffement Allerød. Deux minces couches d'argile à grain très fin présentent une corrélation avec les oscillations de Killarney et du Dryas récent qui entrecoupent la succession stratigraphique du lac glaciaire tardif. Ces couches se sont déposées alors qu'une glace pérenne recouvrait le lac. La lithostratigraphie post-Dryas récent indique l'établissement rapide d'un paysage de plus en plus productif et stable. Cette tendance a été renversée trois fois au cours de l'Holocène par des unités minérogénétiques. Un diamicton complexe d'une épaisseur de 25 cm, unique au lac Piper, a été déposé vers 10,8 – 10,3 ka cal. avant le présent par des processus de pente périglaciaire associés à des effondrements et (ou) des changements dans le niveau du lac; une corrélation directe au refroidissement à l'Holocène précoce (Préboréal) semble peu probable. Deux minces unités minérogénétiques déposées vers 8,1 et 4,9 ka cal. avant le présent résultent probablement d'un refroidissement régional et elles correspondent en gros aux événements notés dans la séquence de carotte glaciaire GISP2. La séquence lithostratigraphique de l'Holocène du lac Piper peut être une conséquence de facteurs limnologiques uniques. Par contre, la forte réponse lithostratigraphique peut résulter de l'absence d'un fort mécanisme climatique régional persistant (oscillation nord-atlantique?) lequel, s'il est présent, pourrait avoir masqué l'impact d'un forçage climatique à une échelle hémisphérique ou plus grande encore.

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Introduction

In this paper, we present a paleolimnological record from a small, headwater lake in the Antigonish Highlands of central Nova Scotia. We employ a multi-proxy lithostratigraphic approach aimed at resolving centennial-scale Holocene environmental change. Recent studies using similar techniques have shown that early Holocene events had a substantial impact in the region; however, these records are of limited extent (e.g., Kurek et al. 2004, Spooner et al. 2002). Biostratigraphic proxies exhibit relatively little change during the middle and late Holocene; indeed, few high-resolution (decadal–centennial scale) studies focusing on this period have been carried out.

Of particular interest is the resolution of short-lived centennial-scale climate oscillations, such as the Preboreal, 8.2 ka, and 4.6 ka events, that have been documented in the Greenland ice core (Greenland Ice Sheet Project 2 (GISP2)), the North Atlantic, and western Europe but are not as well resolved in North America (Alley et al. 1997; Anderson and Lewis 1992; Anderson and Macpherson 1994; von Grafenstein et al. 1998; Klitgaard-Kristensen et al. 1998; Kurek et al. 2004; McDermott et al. 2001; Tinner and Lotter 2001; Yu and Eicher 1998). A number of studies in central Nova Scotia and Newfoundland have focused on fossil pollen, midge, and diatom-based reconstructions of past climate variations and appear to be particularly sensitive to high-magnitude variations that occurred during the Late Glacial and early Holocene (Anderson and Lewis 1992; Jetté and Mott 1995; Kurek et al. 2004; Mott 1994). Other biostratigraphic records from Nova Scotia and Newfoundland suggest that a prolonged period of post-Younger Dryas (YD) – early Holocene cooling (ca. 9.5 ka BP) took place in Nova Scotia and Newfoundland (Anderson and Lewis 1992; Anderson and Macpherson 1994; Grant 1994; MacDonald and Spooner 2001), which has been attributed to eastward drainage of glacial Lake Agassiz (Fisher et al. 2002). This cooling period is roughly correlative to the Preboreal event recognized in high-resolution climate records from the North Atlantic region (Björck et al. 1997; Hald and Hagen 1998). Lithostratigraphic paleoenvironmental records in particular are rare in Atlantic Canada; the one comparative record of environmental change from the immediate region suggests that a climate oscillation took place at about 8400 cal (calibrated) years BP, although the exact nature of the change that took place is not well understood (Spooner et al. 2002).

Background

Timing of deglaciation is variable across Nova Scotia and a date of ca. 13.5 cal ka BP has been proposed for the Antigonish Highlands, with a slightly earlier timing along the southern coast (Stea and Mott 1998; Spooner et al. 2002). During the final stages of deglaciation, remnant ice caps may have been present in highland areas (Stea and Mott 1989; King 1994, 1996; Miller 1995). At this time open, herb-dominated tundra-like communities developed and were rapidly colonized by willow and shrub birch (Mott 1994). Lakes during this time contained primarily pioneer diatom taxa and stenothermic chironomids (Rawlence and Senior 1988; Walker et al. 1991). By 13 cal ka BP, poplar–aspen and spruce woodlands had begun to develop in lowland areas, whereas tundra persisted in the highlands (Mott 1991, 1994; Mott and Stea 1994; Miller 1995; Spooner et al. 2002). Lake diversity increased, and temperate diatoms and chironomids became abundant (Rawlence and Senior 1988; Walker et al. 1991).

The YD-equivalent regional cooling is hypothesized to have resulted in rejuvenation of local ice caps (Stea and Mott 1998). The formation of glacial lakes and buried organic deposits resulting from landscape instability associated with this regional cooling is well documented (Mott et al. 1986; Stea and Mott 1989, 1998; King 1994; Mott and Stea 1994; Stea et al. 1998). During YD cooling, trees declined, herbs and shrubs increased, and many sites returned to open, tundra communities (Mott 1994). Lakes experienced decreased productivity, a resurgence of pioneer species, and in some cases an influx of mineral sediments (Mott 1994). Stea and Mott (1998) have documented a YD-equivalent mineral sediment oscillation in four lakes in the immediate region, including Piper Lake.

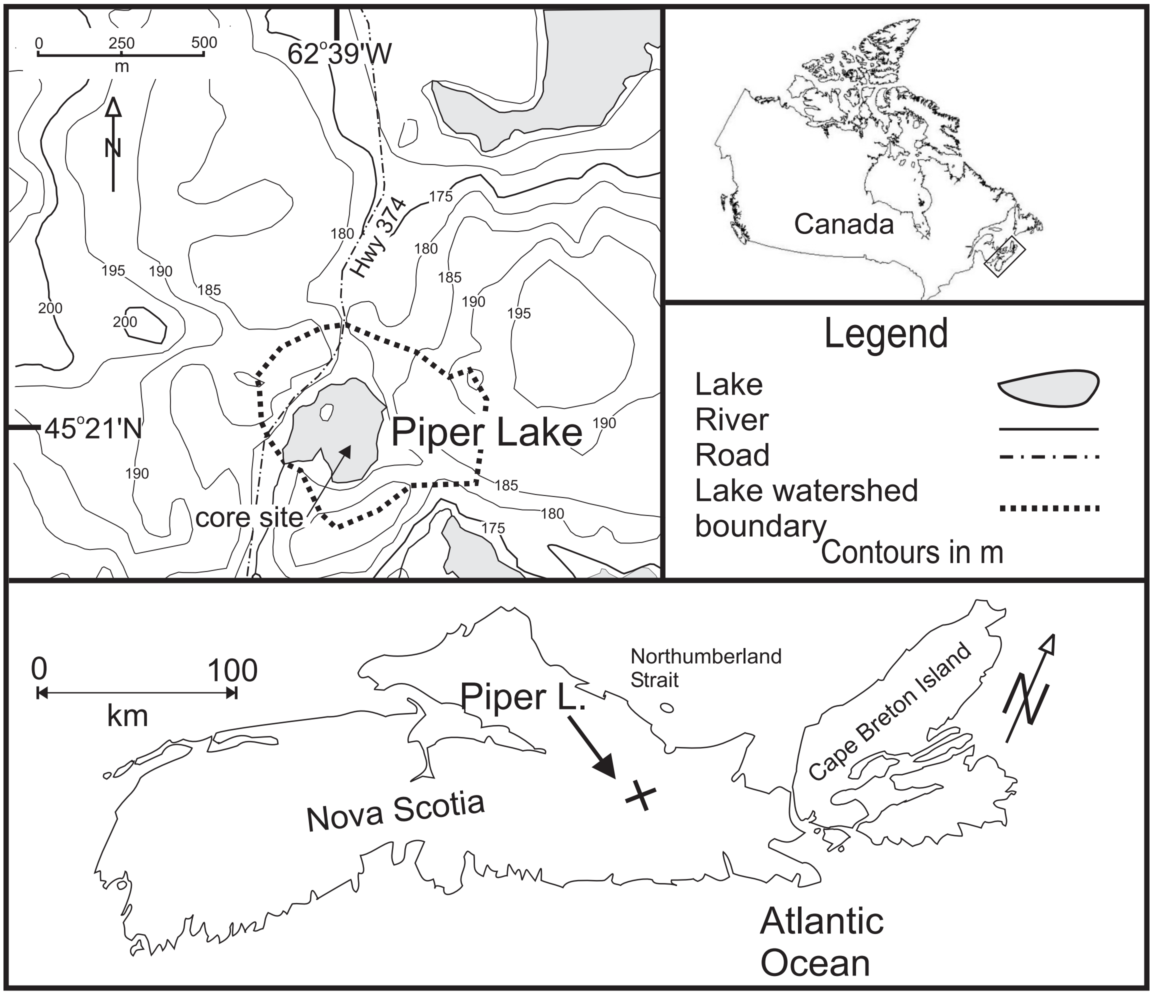
The post-YD Holocene climate in central Nova Scotia has been characterized as a time of relative quiescence with increasing warmth and dryness; however, records from northern Nova Scotia and Newfoundland suggest a period of post-YD cooling (11.2–9.5 cal ka BP; Anderson and Lewis 1992; Anderson and Macpherson 1994; Grant 1994; MacDonald 2001). Spooner et al. (2002) suggest that a lithostratigraphic oscillation (8600–8400 cal years BP) at Taylor Lake, Nova Scotia is most likely the consequence of cooling associated with a hypothesized disruption in North Atlantic thermohaline circulation at 8.2 cal ka BP (Barber et al. 1999).

The most detailed pollen study in the region was carried out at Pye Lake, ~50 km south of Piper Lake and indicates that climate following the YD became warmer and remained so through the middle Holocene (about 7.0–5.0 cal ka BP). Modern climate became established in the region by about 4.0 cal ka BP (Jetté and Mott 1995). Of note is the sharp decline in hemlock that took place at about 7500 cal ka BP. Jetté and Mott (1995) have attributed this to warm, dry conditions and drought in some localities. Subtle increases in spruce, alder, and herbs during the last 1000 years may indicate cooler and more moist conditions consistent with the onset of Little Ice Age cooling (Jetté and Mott 1995).

Study site

Piper Lake (45°21′N, 62°39′W; 179 m above sea level; Fig. 1) is a shallow, distrophic organic lake located in the relatively flat southern uplands of the Antigonish Highlands, Pictou County, Nova Scotia. It has a catchment area of 28 ha, a surface area of 6 ha, a maximum depth of 3.0 m, and is slightly acidic (pH 6.8). It occupies a glacially scoured depression within Carboniferous Horton Group sandstones and siltstones. Relief in the catchment area is subdued (<20 m), with slopes ranging from 1° in wetland areas to a maximum of 9° in three very limited areas. The wetlands are found around the perimeter of the lake and are dominated by small black spruce and alder fens, whereas the higher ground within the watershed has mixed hardwood stands because of the better drainage. The wetlands together with a shallow (<1.5 m maximum depth) sub-basin south of the main lake body con-

Fig. 1. Location of study site. Piper Lake is a small, shallow headwater lake situated near the apex of the Antigonish Highlands in eastern Nova Scotia. Other sites discussed in text are also indicated. Contour interval is 5 m.



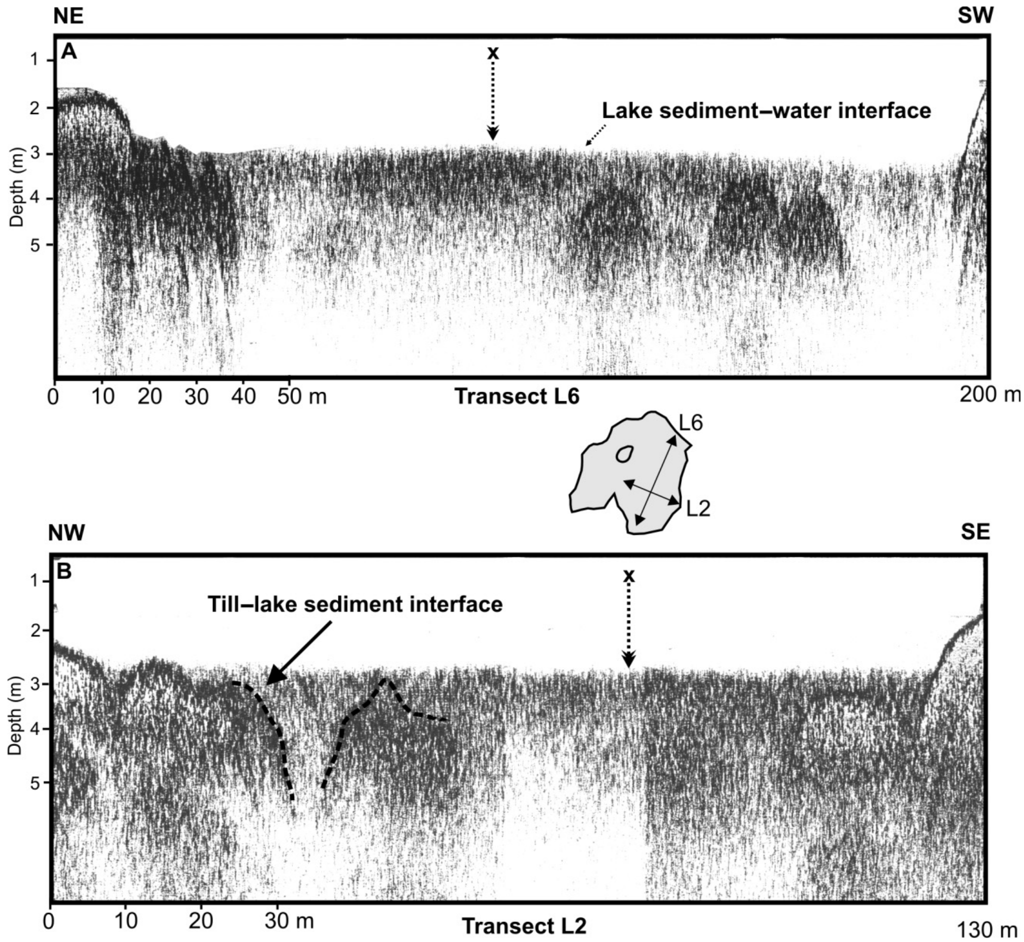
tains about 50% of the aquatic habitat at Piper Lake (Fig. 2). Sonar profiling of Piper Lake indicates that the central basin is flat bottomed with a maximum depth of 3 m (Fig. 2; Nova Scotia Department of Fisheries 1995). The till–lake sediment contact is complex, exhibits relief of > 2 m, and commonly occurs < 2 m below the sediment–water interface at the coring site (Fig. 2).

Method

Piper Lake was cored through lake ice using a portable percussion coring system (Gilbert and Glew 1985; Reasoner 1993). Three 7.5 cm diameter cores were extracted from the lake bed when subsurface penetration ceased, and the average length of the consolidated lake sediment obtained was about 165 cm. A basin-central sampling location was chosen with the aid of sonar profiling and a Nova Scotia Department of Fisheries bathymetry map. Sonar profiling was accomplished using a King 1570 echo sounder (50 khz), which is capable of penetrating lake sediment and providing information on the stratigraphy of post-glacial basin fill. Cores were frozen and transported to Acadia University, Wolfville, Nova Scotia where they were split with a high-speed diamond rock saw, photographed, and analyzed.

Carbon/nitrogen (C/N) ratio data were collected at the Isotope Science Laboratory, University of Calgary, Calgary, Alberta for the organic portion of the core using a Finnigan Mat "TRACERMAT" that includes a Carlo Erba NA 1500 elemental analyzer interfaced to a magnetic-sector mass analyzer. These data were collected to determine the origin of sedimentary organic matter. Allochthonous lake sediments generally produce relatively high C/N ratios, whereas low C/N ratios are usually associated with lake eutrophication (Huttunen and Meriläinen 1983). Duplicate analyses of samples indicated 1% error. Magnetic susceptibility was measured at 2 cm intervals using a KT-9 Kappameter in continuous scan mode. Two passes were made on the same core, with results correlated to determine their reproducibility. Percentage moisture content was determined by measuring the mass of the moist sample, drying the sample at 60 °C, and then measuring the mass of the oven-dried sample. Loss on ignition (LOI) analyses were carried out at 550 °C (total organic carbon) for 4 h following the procedures of Dean (1974) and Heiri et al. (2000). LOI was carried out on samples of known volume and cross-sectional area, facilitating the calculation of sediment influx rates (g cm^{-2} year^{-1}). Duplicate analyses of samples indicate error margins of between 1% (for LOI <5%) and 2.2% (for LOI >15%) Grey-scale values

Fig. 2. Interpretation of site morphology and sonar records from Piper Lake. The "x" on the sonar records denotes the core location. The dashed line is the interpreted contact between post-glacial lake sediment and underlying lodgement till. Note the hummocky appearance of this contact. Vertical exaggeration is × 10.



(0–255) were obtained from digital photographs of the core following procedures similar to those outlined by Christensen and Björck (2001). Grey-scale values indicate both the relative clastic and iron sulphide content of the lake sediment. The contrasting shades of the core were enhanced prior to being photographed by allowing the cleaned core face to oxidize for 24 h (Christensen and Björck 2001). Sediment colors were described using Munsell colour names and notations.

Analyses of the coarse clastic grains in lake sediment can assist in the interpretation of the complex relationship between water-column productivity and landscape stability (Levesque et al. 1994; Spooner 1998; Beierle et al. 2001). The small fetch (maximum of 250 m), muted relief, and moderate depth of Piper Lake limit the common processes that are capable of transporting coarse sediment (>0.5 mm disk diameter) into the center of the lake basin to inter-basin debris flow, sediment reworking owing to fluctuating lake levels, aeolian transport onto ice, and possibly ice rafting of littoral sediments (Luckman 1975; Cole 1983). For the latter two processes, the clasts (ice-rafted debris, IRD) melt through or out of the ice in the central part of the lake basin, accelerated by decreased albedo caused by their darker colour relative to the ice. A relative assessment of coarse sediment grain-size analysis was accomplished on lacustrine sediment by cutting the remainder of the core (after macrofossil analyses) into 5 cm portions which were subsequently impregnated with low viscosity embedding media (Spurr Epoxy ©) and thin-sectioned. These thin sections were then scanned into bitmaps using a Polaroid SprintScan 35® plus slide scanner. The area of individual grains was determined using the image-analysis program Imagetool®. Clast area was then converted to equivalent disk diameter (D0; Francus 1998) and grain counts of D0 values > 0.5 mm were recorded for specific stratigraphic intervals. The data serve as a useful indication of relative changes in the abundance of coarse-grained clasts referred to in the text as IRD. Inorganic grain-size distribution was determined using a Coulter Laser Particle Sizer Analysette 22-E. Samples were obtained every centimetre and the data were plotted as surface plots that allow qualitative interpretation of the characteristics of the entire particle size distribution (Beierle et al. 2001).

Chronological control for the core was established from

Table 1. Accelerator mass spectrometry radiocarbon dating results from Piper Lake.

Depth	Age (^{14}C years BP)	Calendar date	Calendar range (2σ)	Laboratory No.	Material	$\delta^{13}C$
45 cm	3180±85	3384 BP	3630–3174 BP	AA-38574	Needle fragment	–27.4
57 cm	4195±76	4745 BP	4869–4452 BP	AA-38575	Leaf fragments	–25.0
114.5 cm	7654±70	8411 BP	8590–8348 BP	AA-38576	Twig fragment	–28.5
132 cm	8906±81	10050 BP	10211–9699 BP	AA-38577	Alder twig	–29.5
145 cm	9812±76	11200 BP	11337–11121 BP	AA-38678	Sedge leaves	–25.6
150 cm	10 440±110	12504 BP	12901–11773 BP	AA-38581	Twig fragments	–27.8
160 cm	12 490±160	14748 BP	15603–14109 BP	AA-38579	Bark fragments	–21.0
165 cm	146 20±300	17503 BP	18388–16674 BP	AA-38580	Moss fragments	–32.5

Note: Calibration was accomplished using CALIB 4.3 (Stuiver et al. 1998). All samples were collected from core PL 2000-1 except sample AA-38581, which was collected from core PL-98. Depth indicates down-core sample depth. AA, NSF-Arizona AMS Laboratory, Tucson Arizona.

eight accelerator mass spectrometry (AMS) dates (Table 1). All AMS ^{14}C measurements were carried out at the NSF-Arizona AMS Laboratory.

Results and interpretation

Age and depth relationships

Seven dates were obtained on core PL 2000-1 (Fig. 3, Table 1); five have been positively identified as terrestrial in origin. One date was acquired from core PL-98 (150 cm, AA38581; Table 1). This core was obtained within 3 m of PL 2000-1 and exhibited identical stratigraphy. PL-98 contained twig fragments in a horizon that was barren in PL 2000-1. Calibrations were made using CALIB version 4.3 (Stuiver and Reimer 1993, Stuiver et al. 1998). The general trend of the age–depth plot and the absence of any major discrepancies suggest that the stratigraphic record was undisturbed by the coring process. However, both a sedge leaf sample (145 cm) and a moss sample (165 cm) were potentially aquatic in origin and therefore could have derived their carbon, in part, from lake waters, thus increasing the potential for the hard water effect and an erroneously old date (MacDonald et al. 1991). This appears to be unlikely for the sedge leaf sample because its position on the age–depth curve (14 748 cal years BP; see Fig. 3, Table 1) is consistent with other dates obtained in close proximity, and the relatively high $\delta^{13}C$ value of –21 suggests a terrestrial source for the carbon in the sample (Fry and Sherr 1984). The basal age of 17 503 cal years BP (14 620 ± 30^{0} ^{14}C years BP; $\delta^{13}C$ value of –32.5) is problematic because most researchers indicate that deglaciation of the region took place significantly later than this age suggests (see Discussion).

Core lithostratigraphy: description and interpretation

Piper Lake core PL 2000-1, chosen for detailed study, was longest (175 cm) and exhibited a complex stratigraphy (Fig. 4). The core was visually subdivided after it had been allowed to air-dry for 24 h (Fig. 4). Clastic sediment dominates the base of the core and is composed primarily of till and nascent lake-stage silt and clay. Organic lake sediment is prevalent throughout the core and was characteristically very dark brown to black in colour (7.5YR 2.5/1) with high water content, high total organic carbon (TOC), and low magnetic susceptibility. Clastic oscillations (COs) were identified on the basis of change with respect to bounding sediments. Each oscillation displayed specific attributes, which will be discussed later in the text. Boundaries were determined visually by noting points of maximum change in observed and measured parameters.

Till, post-glacial lacustrine silt and clay (unit 1: 175–162 cm, ca. >14750 cal years BP)

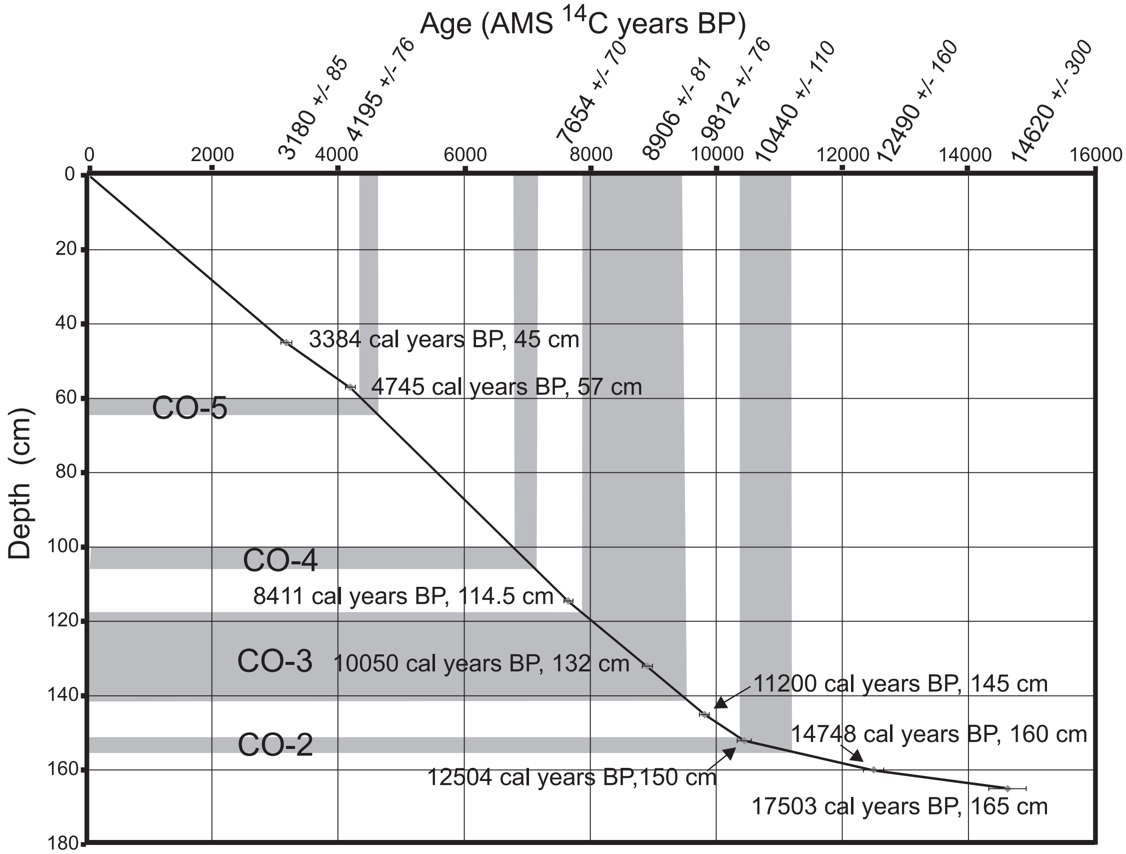
The base of the core consists of dark grey (7.5YR 4/1) clay rich till (175–168 cm, Fig. 4) with occasional angular, striated cobbles. The till is in diffuse contact with a massive, dark brown (7.5YR 3/2), fine sand to clay layer (168–162 cm; Fig. 4) deposited as the scoured basin filled with water. The latter contains fragments of moss that are concentrated in discreet bands, from which a date of 17 503 cal years BP was obtained (Table 1). Unit 1 is characterized by low moisture content and high magnetic susceptibility, with low LOI values. Grain counts of clasts with diameters > 0.5 mm were high at the base of the core and, as with grain size, decrease markedly towards the top of the unit (Fig. 4).

Organic lake sediment

Organic lake sediment at the base of the core (160 cm) is dominantly brown (7.5YR 4/3), contains occasional pebbles and is laminated in places. Terrestrial macrofossils are common. Low C/N values at this point (10.4, 10.6) indicate that the source of the organic sediment was primarily aquatic. From 152–143 cm (Fig. 4) the sediment becomes black (7.5YR 2.5/1), is highly organic, and contains very few visible minerogenic clasts. Occasional coarse sand and pebble clasts were noted and become more common towards the top of the core. Maximum grain size decreased, approaching very fine sand, and two higher C/N values from this zone indicate an increased organic input.

Organic lake sediment in the upper portion of the core (118 cm–top) is dark reddish brown (5YR 2.5/2) and is characterized by relatively high LOI values and moisture content; a trend of decreasing magnetic susceptibility is also noted. C/N values increase from 12.6 at 117 cm to 14.5 at 63 cm, possibly indicating an increased influx of terrestrial vegetation into the lake basin. These data indicate stabilization of the landscape, decreased clastic influx, and an increase in both lake and terrestrial productivity. Sediment influx rates and coarse sediment grain counts are both low. The clastic component of the sediment is poorly sorted and relatively coarse. Highest LOI values (48%) are reached at

Fig. 3. Age vs. depth graph. An inflection point is noted which corresponds to a point of change on the moisture content plot (point A). All material dated was from core PL 2000-1, with the exception of the date at 150 cm, which was from core PL-98.



the 22 cm. Subtle changes in lithostratigraphic proxies from 22 cm–present indicate that the last 1700 years was characterized by reduced autochthonous productivity and a reduction in the contribution of terrestrial vegetation to the organic lake sediment.

Clastic oscillations

CO-1 (162–160 cm, ca. 14 900 – 14 750 cal years BP; Fig. 4)

CO-1 consists of a thin, brown (7.5YR 4/3) clay layer that exhibits sharp bounding contacts and is internally homogeneous. Although moss fragments are common in this clay layer, the low LOI values indicate that little organic sedimentation was taking place.

CO-2 (156–152 cm, ca. 13 100 – 12 400 cal years BP; Figs. 4, 5)

CO-2 consists of a fine-grained minerogenic layer that exhibits abrupt upper and lower contacts and is internally homogeneous. The sediment is brown (7.5YR 4/4) in colour but becomes yellowish red (5YR 4/6) in the top 1 cm. CO-2 contained no macrofossils. CO-2 exhibits low moisture content, low TOC, and a significant increase in magnetic susceptibility. Grain size also decreased and is similar to values found in CO-1 (Fig. 4). Sand grain counts were low within this unit. The single C/N value measured in this zone dropped to 8.0, indicating a relative increase in the aquatic component of the lake sediment. The reduced sedimentation rates and sharp upper and lower contacts inherent in CO-2 are interpreted as indicating either an abrupt decrease in lake productivity and (or) an increase in clastic sediment flux.

CO-3 (143–118 cm, ca. 11 000 – 9000 cal years BP; Figs. 4, 5)

The transition between underlying organic lake sediment and CO-3 occurs over a distance of about 1 cm. CO-3 consists of a complex, 25 cm thick minerogenic layer. The lower 9 cm of the sediment is fine to coarse brown (7.5YR 4/3) diamicton; clasts up to 3 cm in long-axis length were recovered. The diamicton grades gradually into a dark brown (7.5YR 3/3) minerogenic layer that contains much sand and fine pebbles; organic matter is also present and a date of 10 050 cal years BP was obtained on a twig at 132 cm (Table 1). Large pebbles become absent above 134 cm and the sediment becomes somewhat darker coloured (7.5YR 3/2). The lowermost C/N value (10.1) indicates a relative decrease in terrestrial input, however values increase though the remainder of this zone. The contact with overlying organic lake sediment is diffuse. These sediments are also characterized by low LOI, and the highest magnetic susceptibility and sediment influx values in the core.

CO-4 (106–98 cm, ca. 8100 cal years BP; Fig. 4) and CO-5 (65–60 cm, ca. 4900 cal years BP; Fig. 4)

Both CO-4 and CO-5 are characterized as dark brown (7.5YR 3/2) minerogenic beds that exhibit diffuse upper and lower contacts and a subtle colour shift with respect to bounding

Fig. 4. Piper Lake lithostratigraphy. Lithostratigraphic zones (Z1) were arrived at by visually determining points of maximum change in observed and measured parameters. Grain counts were computed for select sites. Counts of zero are indicated by an asterix (*).

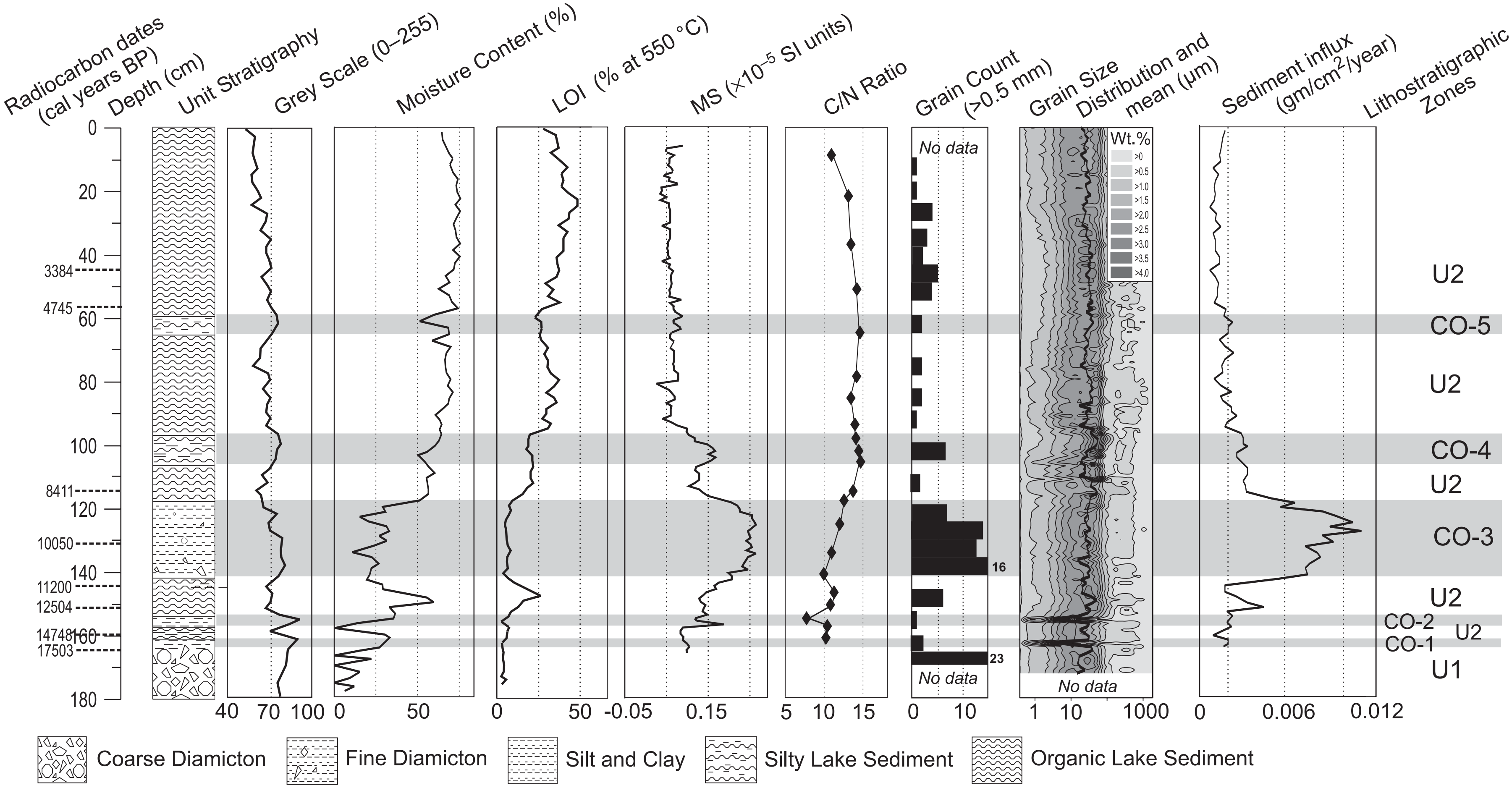
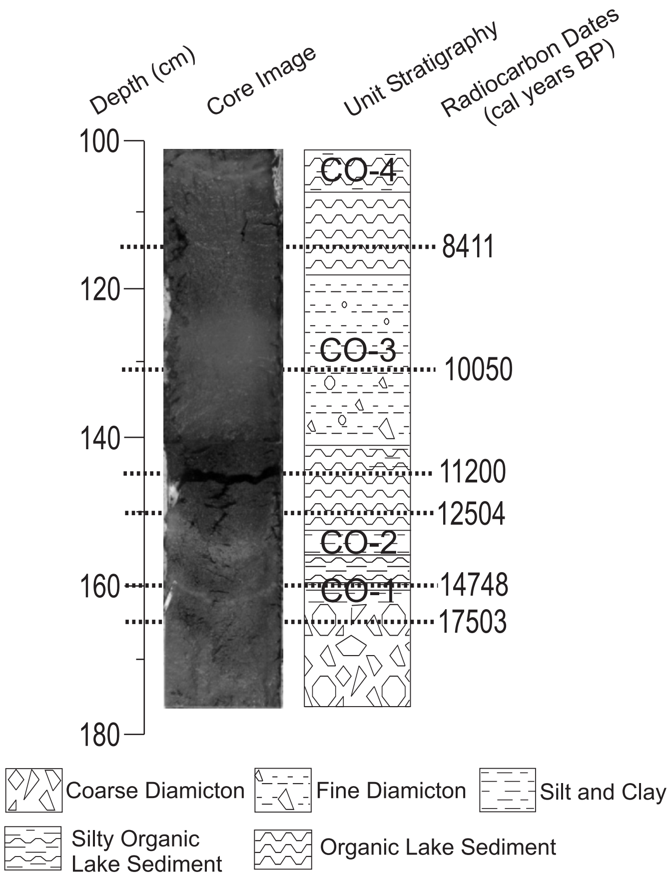


Fig. 5. Sedimentology of lower core PL 2000-1. Core width is 7 cm. CO-2 was deposited during the Younger Dryas.



organic lake sediment. No coarse-grained clastics or macrofossils were evident within either minerogenic layer. CO-4 is characterized by increased magnetic susceptibility, as well as lower LOI and moisture content values relative to bounding zones. Image analysis of the bounding sediment and CO-4 indicate that CO-4 is slightly sandier and mean grain size (D0) has increased slightly. CO-5 displays similar attributes with more subtle transitions in the lithostratigraphic proxies between the bounding sediment and CO-5. Two C/N values (14.6, 14.3) obtained within these oscillations demonstrate a continued increase in the influence of terrestrial vegetation on the organic component. The decrease in LOI and the slight increase in mean grain size observed for both CO-4 and CO-5 can be attributed to a relative increase in clastic sedimentation during this time.

Discussion

Deglaciation of the Piper Lake site probably took place shortly before 14 750 cal years BP (12 500 ^{14}C years BP). This date is somewhat problematic because Stea et al. (1998) indicated that the region was still ice covered at this time. However, other lake records in the region contain similar basal dates. Mayle et al. (1993) obtained a basal age of 14 850 cal years BP on a twig sample at Chase Pond, Cape Breton, Nova Scotia. Mott et al. (1986) have reported deglaciation ages of 13 400 cal years BP at Amaguadees, Central Cape Breton Island, and Stea and Mott (1989) have obtained at date of 13 450 cal years BP at Collins Pond, Nova Scotia (located about 150 km from Piper Lake). Basal ages for nearby sites are somewhat younger than those obtained at Piper Lake. Stea and Mott (1998) recorded a date of 13 000 cal years BP at Piper Lake. The sample was obtained 13 cm below the base of a horizon interpreted to be YD-equivalent. Pye Lake was deglaciated sometime prior to 12 800 cal years BP, although herb and shrub communities were already established by that time (Jetté and Mott 1995). Indian Lake became ice free shortly before 11 450 cal years BP (Stea and Mott 1998), and Taylor Lake was ice free in advance of 10 600 cal years BP (Spooner et al. 2002). The variance in these ages is to be expected given the general paucity of terrestrial organic material in shallow lake sediment (Spooner 1998) and the lack of precise age control on the glacial–postglacial contact in lake cores. Alternatively, this variability might be a function of the deglaciation complexity proposed by Stea and Mott (1998). It is possible that deglaciation and lake development may have occurred at some sites whereas stagnant ice might have persisted at other locales. A digital elevation model (DEM) of the region does indicate that hummocky moraine and scoured bedrock occur in close proximity, an indication perhaps of variability in the timing and nature of deglaciation in the region.

The lithostratigraphic record from Piper Lake indicates that following deglaciation (U1, Fig. 4) at least five periods of environmental change occurred. These clastic-rich beds (CO-1 to CO-5; Fig. 4) are all characterized by different physical properties; however, all are bounded by organic lake sediment that changes little in character up the core (Fig. 4). Oscillations CO-1 and CO-2 display very similar lithostratigraphic properties. In particular, they both exhibit extremely sharp upper and lower boundaries and very low LOI values, and both are very thin and composed primarily of clay. Similar sediment has been observed in other lake-sediment records and most commonly has been interpreted to be periglacial silts and clays deposited during the initial stages of lake-basin development following deglaciation (Stea and Mott 1989; Spooner 1998). However, CO-1 is roughly coeval with the timing of the Killarney oscillation (ca. 13 000 cal years BP; Levesque et al. 1993) and CO-2 was deposited during the early stages of YD cooling (Stea and Mott 1998; see Fig. 3). Both CO-1 and CO-2 differ from most other Killarney and YD-equivalent oscillations in that they do not contain the fine- to medium-grained sand that has been attributed to a combination of landscape denudation, shoreline instability, and inflow focussing as a result of ice and snow melt (Levesque et al. 1994). However, some analogues do exist. Both CO-1 and CO-2 are similar to YD-equivalent oscillations noted at Taylor Lake (Spooner et al. 2002) and Canoran Lake (Martin 2003). In the latter example, sediment cores recovered from the deepest locations in the lake basin contained a thin YD-equivalent clastic oscillation and very low counts of IRD.

For both CO-1 and CO-2, the extremely sharp upper and lower contacts are an indication of a rapid shift in sedimentation style brought on by a prominent change in the processes active in the lake basin. The very fine-grained nature of the sediment is most likely an indication that these changes took place during a period of perennial ice cover under which the site experienced a rapid decline in productivity and clastic sediment transfer processes were suppressed. The lack of IRD in both of these oscillations may be a consequence of (*i*) persistent regional snow cover that would limit aeolian sediment transfer and (*ii*) perennial lake ice cover that would limit the transferral of clastic sediment to the lake basin. Persistent snow cover is possible; pollen profiles from Pye Lake and Taylor Lake indicate that pioneering species still dominated at this time (Jetté and Mott 1995, Spooner et al. 2002). Of note is the difference in thickness of YD-equivalent CO-2 (2 cm) and the thickness of the YD unit (15 cm; U2, Fig. 4) obtained from Piper Lake by Stea and Mott (1998). Martin (2003) has noted that significant changes in the thickness and sedimentology of clastic oscillations occur within lake basins, often over relatively short distances. As well, CO-2 terminates at about 12 400 cal years BP, whereas YD cooling is broadly defined as ending about 11 500 cal years BP (Alley 2000). There is no indication in the core of sediment deformation or missing sediment that might have resulted in a hiatus. More likely, CO-2 represents a lake threshold response; amelioration of local conditions and enhanced lake productivity at Piper Lake may have preceded the termination of YD cooling (Björck et al. 2002).

Although the rapid transition between organic lake sediment (ca. 12.5–11 cal ka BP) and CO-3 (ca. 11–10 cal ka BP) suggests that an abrupt transition in environmental conditions took place in the early Holocene, no lithostratigraphic analogue exists in other lakes records from the region (Mott 1994, Spooner et al. 2002; Stea and Mott 1998). It is unlikely that the clastic oscillation was deposited by active ice because no other records from the region indicate that active ice existed at this time. This unit may have been formed as a consequence of a drop in lake levels and sediment reworking. Because Piper Lake is essentially a closed basin, it is possible that lake levels fluctuated during the period of record in response to climatically controlled changes in the water balance. Various biostratigraphic records indicate that accelerated warming and drying took place at about 10 500 cal years BP, coincident with the deposition of CO-3 (Jetté and Mott 1995; McCarthy et al. 1995). A similar oscillation in an organic lake-sediment core from central Nova Scotia has been reported by MacLeod (1999) and was created by lake ice and wave-induced sediment reworking as a consequence of lake-level lowering for a hydroelectric development.

Alternatively, sonar profiles of the till–lake sediment boundary (Fig. 2) and airphoto and DEM analysis of the region indicate hummocky terrain that is typically formed as a consequence of buried ice meltout. Although the emplacement of this ice must predate the deposition of CO-1 and CO-2, it is possible that periglacial process on land persisted for some time during the early Holocene and may have continued into water. Mott (1994) indicates that remnant ice may have endured until the early Holocene despite a regional warming and drying trend. However, the gradual transition at the top of CO-3 and gradually increasing C/N ratios suggest that the sediment was not deposited instantaneously; as well, internal grading is indicative of changing depositional processes through that period.

Both CO-4 (ca. 8100 cal years BP) and CO-5 (ca. 4900 cal years BP) mark short-lived events that were likely climatically induced. Though the lithostratigraphic proxies display similar trends for both oscillations, trends for CO-5 are more subtle. In the absence of topographic or other allochthonous initiation mechanisms, these two minerogenic beds were probably deposited during short-lived periods of environmental change—most likely changes in trophic states associated with regional climate cooling (Kurek et al. 2004; Spooner et al. 2002; Björck et al. 2001).

CO-4 (ca. 8100 cal years BP) is roughly correlative with the so-called 8.2 ka (calibrated age) cooling event, which is thought to have been triggered by a massive outflow of freshwater from the Hudson Strait (Barber et al. 1999) and is consistent with records from western Europe and the North Sea (von Grafenstein et al. 1998; Klitgaard-Kristensen et al. 1998; Tinner and Lotter 2001). However, it is considerably younger than the age proposed for the freshwater pulse (ca. 8550 cal years BP; von Grafenstein et al. 1998) and the age reported by Spooner et al. (2002) at for a similar oscillation at Taylor Lake (8400 cal years BP). These discrepancies may be the result of dating errors and variability in sedimentation rates, and are similar in magnitude to those reported by Stea and Mott (1998) in their study of the timing of the YD in Nova Scotia. More high-resolution records will no doubt lead to better constraints on the timing of this event. To our knowledge, CO-5 (ca. 4900 cal years BP) has not been recognized in any other records in the region. This oscillation may be correlative with the 4.6 ka (calibrated age) event, however the correlation is tentative owing to its subtle lithostratigraphic expression in the Piper Lake record.

Of interest is the lack of corroborating evidence of CO-4 and CO-5 equivalent cooling in most paleolimnological records in Atlantic Canada. Oickle and Spooner (2004) have shown that subtle changes in lake morphometry may result in differences in lake thermal response and productivity. It is possible that the morphometry and location of both Piper Lake and Taylor Lake resulted in an increased sensitivity to air-temperature change. Alternatively, Yu and Eicher (1998) have noted that lithostratigraphic paleolimnological studies using high-resolution sampling techniques and mulitproxy analyses are scarce. Those regional studies that do exist indicate that the sedimentology of minerogenic oscillations, regardless of their thickness, can be highly variable, and this is a topic that requires more study (Donner 1995; Spooner 1998; Martin 2003). Absence of the recognition of these short-lived cooling events in pollen records does not indicate that there was no vegetative response to cooling. The stratigraphic sampling resolution employed in previous studies over this time interval may not have been high enough to reveal such a subtle sedimentological or biological response.

The importance of the Piper Lake record lies in its resolution of four distinct lithostratigraphic oscillations (CO-1, CO-2, CO-4, CO-5) that can be correlated to recognized hemispheric climate change events. A compelling question is whether conditions specific to the site were responsible for the development of these oscillations or whether the region itself was (is?) particularly sensitive to climate change. Certainly the work of Jetté and Mott (1995), Levesque et al. (1993, 1997), Mott (1985, 1991,1994), Mott et al. (1986), Stea and Mott (1989, 1998), and many others has demonstrated a very strong regional response to Late Glacial events (Killarney, YD) in northeastern North America. Levesque et al. (1997) have shown that this strong response may be due, in part, to steep north–south thermal gradients linked to the close proximity of the Laurentide Ice Sheet. A strong response in the Holocene is more difficult to explain; however, a potential mechanism exists. Hurrell and Van Loon (1997) have noted that this region has not been significantly affected by variations in climate associated with the North Atlantic oscillation (NAO) because it lies at a neutral point with respect to the deviation of the NAO index (an index of decadal-scale North Atlantic precipitation and temperature variation). Thus, sites in the Nova Scotia may not be subject to decadal-scale variations in precipitation and temperature that could introduce noise into records of larger-scale climate forcing, such as those that may have resulted in the formation of CO-4 and CO-5.

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