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ORIGINAL PAPER

Middle to late Holocene chironomid-inferred July temperatures for the central Northwest Territories, Canada

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Abstract We analyzed subfossil chironomids, sediment organic matter and sediment particle size data from a 1.11-m-long freeze core collected from Carleton Lake (*unofficial name*), located approximately 120 km north of the modern treeline. This well-dated core spans the last *ca.* 6,500 years. Two chironomid transfer functions were applied to infer mean July air temperatures. Our results indicated that the chironomid-inferred temperatures from this lake sediment record did not pass a significance test, suggesting that other factors in addition to temperature may have been important in structuring the chironomid community through time. Although not statistically significant, the chironomid-inferred temperatures from this site do follow a familiar pattern, with highest inferred temperatures occurring during the Holocene Thermal Maximum (~6–4 cal kyr BP), followed by a long-term cooling trend, which is reversed during the last 600 years. The largest change in the chironomid assemblage, which occurred between *ca.* 4,600 and 3,900 cal yr BP is possibly related to the well-documented northward advance and subsequent retreat of treeline in this region.

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Introduction

In recent decades northwestern Canada has experienced some of the most rapid warming in the Northern Hemisphere (~1.5–2.0 °C/decade from 1981 to 2001; Comiso 2003) and climate models predict continued temperature increases in this region for at least the next century (Arctic Climate Impact Assessment 2005; Intergovernmental Panel on Climate Change 2007). This rapid warming is already impacting northern environments and communities (Krupnik and Jolly 2002; Arctic Climate Impact Assessment 2005) and the predicted future rise in temperature is anticipated to further alter this region (Arctic Climate Impact Assessment 2005; Stephenson et al. 2011). A striking feature of global climate change has been its spatial heterogeneity and temporal variation, highlighting the need for local records of climate variability (Kaufman et al. 2004; Arctic Climate Impact Assessment 2005). Despite the importance of climate change to the social and economic development of the arctic and subarctic, long-term (centennial timescales and longer) climate dynamics of these regions remain poorly understood. For example, in Canada's vast Northwest Territories (NT), only two instrumental climate records extend into the nineteenth century and systematic collection of meteorological data largely began only in the 1940s and 1950s (Environment Canada 2012). The poor spatial coverage and short length of the instrumental records have made it difficult to properly assess regional climate dynamics and evaluate the role of decadal- and longer-scale climate drivers that have been shown to affect the NT (Pisaric et al. 2009). Furthermore, long-term climate data are required to provide context for the observed rapid warming that is presently occurring across much of the NT.

When instrumental data are absent, proxy records offer the best method for assessing long-term climate trends. Previous proxy-based paleoclimate research in the central NT has documented a northward movement of the treeline between *ca.* 6,000 and *ca.* 3,500 cal yr BP in response to warmer-than-present temperatures (Moser and MacDonald 1990; MacDonald et al. 1993; Pienitz et al. 1999). Diatom analysis in this region has also been used to examine recent warming patterns, with an increase in planktonic taxa over the last century suggesting longer ice-free seasons and enhanced thermal stratification of northern lakes (Rühland et al. 2003; Rühland and Smol 2005). Although these earlier studies have recorded mid-Holocene climate variability and provided valuable insights into twentieth-century warming, long-term climate inferences for the region have been largely qualitative, with a few exceptions (Edwards et al. 1996; MacDonald et al. 2009).

High-resolution, quantitative proxy records of mid- to late Holocene climate variability in subarctic and arctic North America have generally reported highly variable summer temperatures, particularly over the last 2,000 years. For example, in southern Alaska, Clegg et al. (2010) reported climatic cooling events of up to 1 °C occurring at 4,000, 3,300, 1,800–1,300, 600, and 250 cal yr BP, related to periods of minima in solar irradiance. The timing of these cooling events over the last 2,000 years is consistent with other lake sediment proxy records in both Alaska (Hu et al. 2001; 2003) and the Canadian Arctic Archipelago (Thomas et al. 2011) and also with glacier expansion records along the Pacific Alaskan and British Columbian coasts (Reyes et al. 2006). The consistency of the timing of these climate events from southern British Columbia to Baffin Island in the Canadian Arctic Archipelago suggests broad-scale synchroneity in climate forcing during the late Holocene over much of northern North America.

The chitinous remains of non-biting midges (Order Diptera, Family Chironomidae) are abundant and well preserved in lake sediments and can provide robust, quantitative estimates of historical mean July temperature (Porinchu and MacDonald 2003; Walker and Cwynar 2006; Larocque-Tobler et al. 2009; Clegg et al. 2010). In the past decade, there has been widespread use of chironomids as paleoenvironmental indicators in high-latitude regions of North America (Francis et al. 2006; Kurek et al. 2009; Clegg et al. 2010; Irvine et al. 2012; Medeiros et al. 2012), in large part as a consequence of development of training sets that relate modern chironomid assemblages to environmental variables (Barley et al. 2006; Porinchu et al. 2009; Medeiros and Quinlan 2011). The objective of this research is to use paleolimnological techniques,

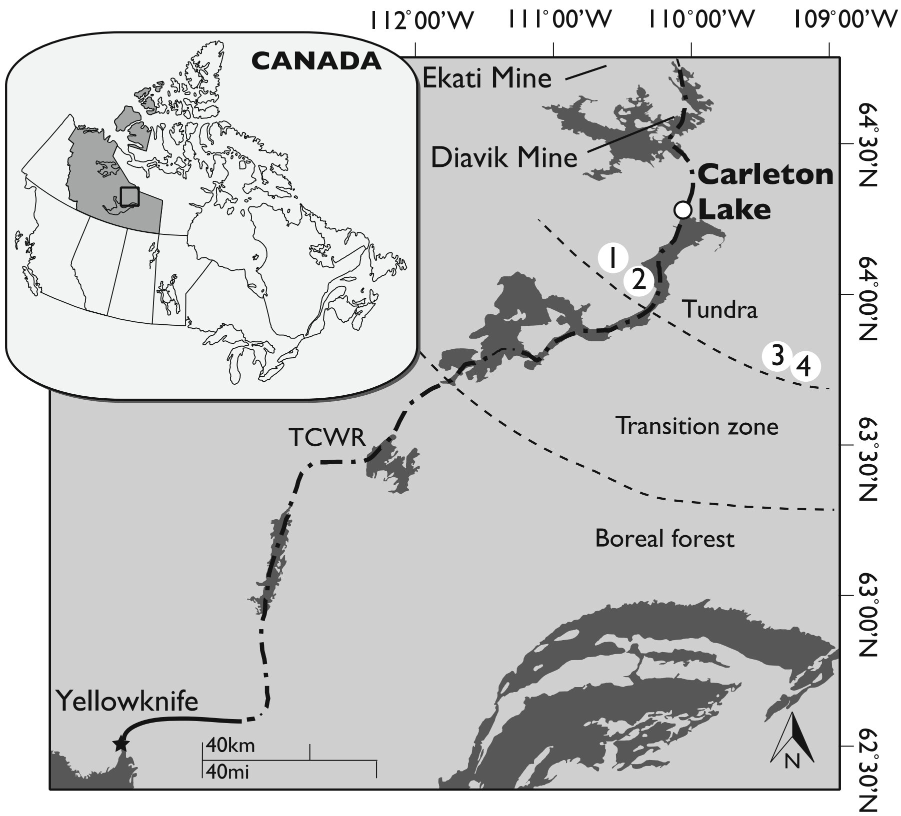

Fig. 1 Map showing location of Carleton Lake in relation to the city of Yellowknife, the Diavik and Ekati diamond mines, and other nearby paleolimnological records in the central Northwest Territories (*1*) McMaster Lake (MacDonald et al. 1993), (*2*) Queen's Lake (Pienitz et al. 1999), (*3*) Lake S41 (MacDonald et al. 2009), and (*4*) Toronto Lake (MacDonald et al. 1993). The route of the Tibbitt to Contwoyto Winter Road (TCWR) which links the remote diamond mines to Yellowknife is shown as a dashed line, as are the approximate transition from Boreal forest to tundra. *Inset map* shows the location of the Northwest Territories (*dark grey*) within Canada and the study region highlighted in a *black square*

including chironomid analysis, to infer quantitatively mid- to late Holocene July air temperatures in the central NT. We focus specifically on an environmental reconstruction from Carleton Lake (*unofficial name*; also informally designated as P-49 by the Tibbitt to Contwoyto Winter Road (TCWR) Joint Venture) in the central NT (Fig. 1). This site is located along the economically important TCWR, a critical supply route to diamond mines in the NT and an area where there is considerable concern about the impacts of future temperature change. These reconstructions are compared with published climate proxy records from three other regional lake sediment records to better elucidate climate-related environmental changes during the mid-to late Holocene.

Study site

Carleton Lake (64°15′26″N, 110°06′03″W; elevation 420 m asl) has a surface area of 29.8 ha and is the first lake to the north of the 49th land portage along the TCWR in the central NT, approximately 17 km south of Lac de Gras (Fig. 1). Detailed bathymetry of Carleton Lake is not available, however the lake is shallow (maximum depth <5 m) and polymictic. The lake is located approximately 120 km north of the treeline, at the transition between the Taiga Shield and Southern arctic ecozones, where the landscape is characterized by continuous permafrost cover (Natural Resources Canada 2009). The catchment of Carleton Lake is underlain by Archean granitic and gneissic bedrock of the Slave Geological Province. These rocks consist of amphibolite-grade paragneiss to quartz biotite schist (Stubley 1990; Davis et al. 1996). No carbonates were mapped in the area, limiting the possibility of hard-water effect on the radiocarbon dates.

The study region currently experiences a subarctic climate, with annual precipitation in the region ranging from 201 to 400 mm (Natural Resources Canada 2009) and mean daily air temperatures ranging between −34 and −25 °C in January and between 6 and 20 °C in July (Natural Resources Canada 2009). The closest weather station to the study site is located near the Ekati diamond mine airport (55 km north; elevation 469 m asl) which reported a mean July air temperature of 14.9 °C in 2011, when our fieldwork was conducted. Climate records from the Ekati station extend back to only 1998. Thus, to estimate mean July temperature for our study region over multi-decadal time scales, which are more comparable to our chironomid temperature inferences, we used a linear regression model that relates mean July temperatures for Ekati to Yellowknife, ~300 km to the southwest. Our resulting model,

$$T_E = -8.74 + 1.28 \times T_Y; \quad \left(r^2 = 0.83, p < 0.001, n = 10\right)$$

where T_E is the mean July temperature at Ekati and T_Y is the mean July temperature in Yellowknife, provides an estimated average mean July temperature of 12.5 °C for the Ekati airport weather station. Accounting for a lapse rate of −0.65 °C per 100 m elevation, we estimate that mean July temperature at our study region between 1942 and 2011 was approximately 12.8 °C.

Materials and methods

Sediment sampling, laboratory analysis and chronology

A 111-cm sediment core was retrieved from a water depth of 1.5 m at Carleton Lake in March 2010 using a single-faced freeze corer (methods described in Galloway et al. 2010; Macumber et al. 2012). The freeze corer was filled with a dry ice and isopropyl alcohol slurry and lowered through a hole augured through the ice. The corer was slowly pushed into the lake-bed and left in situ for 15 min, during which time the lake sediments froze to the metal surface of the coring device. The sediment was carefully retrieved by winching the corer back to the surface. The frozen sediment slabs were then detached from the corer, kept frozen, and shipped to Carleton University, Ottawa, Ontario, Canada, for further analysis.

Seven bulk sediment samples were used for accelerator mass spectrometry (AMS) ^{14}C dating at the ^{14}CHRONO Centre in Belfast (Table 1). All samples underwent a standard hydrochloric acid wash to remove any potential carbonate material. Radiocarbon ages were calibrated using CALIB software version 6.1.0 (Stuiver and Reimer 1993) and the IntCal09 calibration curve (Reimer et al. 2009) with the exception of the uppermost date (UBA-18472), which was younger than AD 1950 and was therefore calibrated in CALIBomb (Reimer et al. 2004) with the NH_zone1.14c dataset (Hua and Barbetti 2004). An age-depth model was constructed using the Bayesian age-depth modeling software Bacon (Blaauw and Christen 2011). The age modeling procedure is similar to that outlined in Blaauw and Christen (2005), but more numerous and shorter sections are used to generate a more flexible chronology. A mean accumulation rate of 60 year/cm, was entered a priori in Bacon, based on a summary of accumulation rates by Crann (2013).

The upper 100 cm of the Carleton Lake core was sub-sampled at 0.5-cm intervals using a custom-designed sledge freeze-core microtome (Macumber et al. 2011) for organic content, particle size, and microfossil analyses. Because of the limited amount of material in the upper 10 cm of the core, organic content was analyzed only below 10 cm core depth. Total organic and carbonate content of the sediment was estimated on 1 cm^3 sub-samples by loss-on-ignition (LOI) at 550 and 950 °C for 4 and 2 h, respectively (Heiri et al. 2001). Particle-size distributions were determined for every sample interval using a Beckman Coulter LS 13 320 Laser Diffraction-Particle Size Analyzer with a Universal Liquid Module (Beckman Coulter 2003). Utilizing a protocol modified from van Hengstum et al. (2007) and Murray (2002), 30 % H_2O_2 was added to sub-samples in a warm water bath to digest organic matter. Grain size statistical parameters were calculated from the LS 13 320 results in conjunction with the software package Gradistat v8.0 (Blott and Pye

Table 1 Radiocarbon dates for the Carleton Lake single-faced core reported in years BP except UBA-18472, which is reported as fraction modern carbon as it is younger than 1950 AD

Sample ID	Lab ID	Depth (cm)	^{14}C age (BP ± 1σ)	Calibrated age (cal BP ± 2σ)
P49-1B-F1_0cm	UBA-18472	0.0 ± 0.5	1.0264 ± 0.0035	1956 AD ± 1
P49-1B-F1_10cm	UBA-17934	10.0 ± 0.5	1046 ± 24	954 ± 29
P49-1B-F1_19.5cm	UBA-17347	19.5 ± 0.5	1925 ± 25	1874 ± 52
P49-1B-F1_40cm	UBA-17935	40.0 ± 0.5	2762 ± 35	2863 ± 83
P49-1B-F1_64.5cm	UBA-17348	64.5 ± 0.5	3675 ± 24	3939 ± 13
				4022 ± 65
P49-1B-F1_80cm	UBA-17936	80.0 ± 0.5	4635 ± 32	5319 ± 15
				5418 ± 47
P49-1B-F1_100cm	UBA-17349	100.0 ± 0.5	5663 ± 26	6448 ± 49

All dates use IntCal09 (Reimer et al. 2009) and Calib version 6.1.0 (Stuiver and Reimer 1993) for calibration, except UBA-18472, which was calibrated using the NH_zone1.14c dataset (Hua and Barbetti 2004) and CALIBomb (Reimer et al. 2004)

2001). Stratigraphic profiles of both LOI and the particle size analysis were plotted using C2 v. 1.6.7 (Juggins 2010).

Chironomids were isolated from sediment sub-samples (ranging in wet weight from 0.53 to 2.14 g), following Walker (2001). The sub-samples were sieved using a 100-μm mesh size, following hot KOH treatment (performed on all sub-samples below 10 cm in the Carleton Lake core), rinsed with distilled water, and stored in vials. A modification of the Walker (2001) method was necessary for sub-samples obtained from the top 10 cm of the Carleton Lake core, where the volume of sample material was limited. Samples from the upper 10 cm of the core were not treated with KOH, as this treatment damages other microfossil remains that will be used in a parallel study. A Bogorov counting tray and dissecting microscope were used to pick fossil midge remains, which were then mounted onto microscope slides using Permount®. Chironomid head capsules were identified using a compound microscope under 200× and 400× magnification, following Brooks et al. (2007) and Walker (2007). A minimum of 50 head capsules was identified for each sample, except for one, in which 48 head capsules were identified (Heiri and Lotter 2001; Quinlan and Smol 2001). Statistical analysis and temperature inferences were based on the square-root transformed relative abundance data. Chironomid relative abundance was plotted using C2 v.1.6.7 (Juggins 2010). Statistically significant stratigraphic assemblage zones were identified using constrained incremental sum of squares (CONISS) clustering compared against a broken stick model using the rioja package in R (Juggins 2012).

Historical mean July air temperatures at Carleton Lake were inferred using the Barley et al. (2006) northwest North American and the Porinchu et al. (2009) central Canadian Arctic transfer functions. The Barley et al. (2006) transfer function is based on a training set of environmental variables and midge assemblages from 145 lakes in British Columbia, Alaska, Yukon, the Northwest Territories, and the Canadian Arctic Archipelago. The Porinchu et al. (2009) transfer function is based on a training set of 88 lakes in the central Canadian Arctic, spanning the boreal forest in the south to mid-Arctic tundra in the north. Both transfer function models for temperature are based on a weighted averaging-partial least squares (WA-PLS) two-component model (Barley et al. 2006, $r^2_{boot} = 0.82$, $RMSEP_{boot} = 1.46$ °C; Porinchu et al. 2009, $r^2_{jack} = 0.77$, $RMSEP_{jack} = 1.03$ °C). Chironomidae that could not be identified below sub-family or tribe, or that were absent from the training sets, were not included in the temperature reconstruction. Results of the transfer functions were plotted and a LOESS regression (span = 0.18) was used to smooth the historical temperature data. Squared chord distance (SCD) measurements were used to identify no-analogue situations between our sub-fossil chironomid assemblages and those in the Barley et al. (2006) and the Porinchu et al. (2009) training sets. Core samples that had SCD values greater than 5 % of SCD values among training set lakes were identified as having poor modern analogues (Simpson and Oksanen 2009). The significance of the chironomid-inferred paleotemperature reconstruction was further tested using the randomization method proposed by Telford and Birks (2011) using the R package palaeoSig (Telford 2011). If the reconstructed temperature explained more variability in chironomid assemblages than 95 % of the randomly generated reconstructions, the reconstruction was considered significant (Telford and Birks 2011).

To examine temporal trends in chironomid community assemblages preserved in the Carleton Lake sediment core, multivariate analysis was conducted using the vegan package (Oksanen et al. 2012) in R (R Core Team 2012). A detrended correspondence analysis (DCA) indicated that chironomid species turnover throughout the sediment core was <2 standard deviations (axis 1 length = 1.3 standard deviations), therefore a principal component analysis (PCA) was chosen to evaluate trends in chironomid assemblages over time (ter Braak and Šmilauer 2002). To place our study in a regional context, our results were then compared to other regional records of paleoenvironmental change from three nearby lakes: Lake S41 (MacDonald et al. 2009), Queen's Lake and Toronto Lake (Pienitz et al. 1999) (Fig. 1). A regional composite graph of paleoenvironmental change was created by digitizing treeline and temperature indicators from these lakes using DigitizeIt v. 1.5 to facilitate comparison between records. The chironomid-inferred mean July air temperatures from Carleton Lake based on both the Barley et al. (2006) and the Porinchu et al. (2009) training sets have been deposited at the National Oceanic and Atmospheric Administration (NOAA) National Climatic Data Center and are available to download at http://www.ncdc.noaa.gov/data-access/paleoclimatology-data/datasets/insect.

Results

Chronology and stratigraphy

Age-depth modeling, based on the seven radiocarbon dates, indicated that the 111-cm Carleton Lake sediment core spanned the last *ca.* 6,500 cal yr (Fig. 2). Radiocarbon dates are reported in Table 1 as calibrated radiocarbon years before present (cal yr BP), defined as AD 1950. The uppermost date (UBA-18472), however, is presented as fraction modern because it is younger than AD 1950.

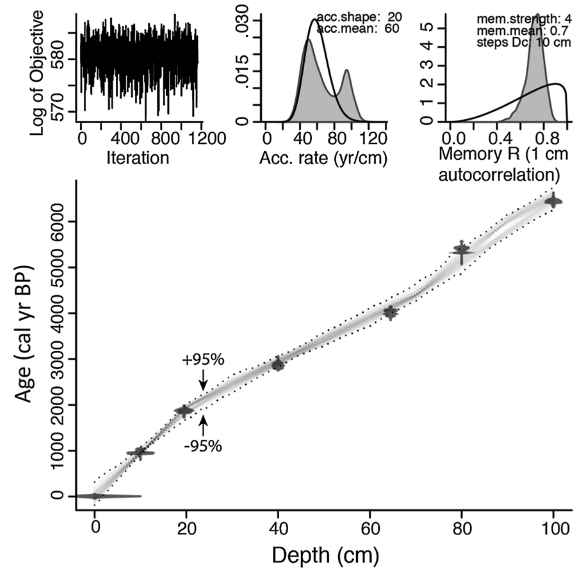

Fig. 2 Bayesian age-depth model for Carleton Lake. On the *top panel*, the *leftmost plot* shows that the Markov Chain Monte Carlo runs were stable (1,200 iterations), the *middle plot shows* the prior (*curve*) and posterior (*filled distribution curve*) distribution for the accumulation rate (year/cm), and the *rightmost plots* show the prior (*curve*) and posterior (*filled distribution curve*) for the dependence of accumulation rate between sections. The *major plot* shows the age distributions of calibrated ^{14}C dates and the grey-scale age-depth model indicates precisely dated sections of the chronology in *darker grey*, whereas *lighter grey areas* indicate less precise sections

Freeze coring successfully captured the sediment–water interface, which was flat and showed a distinct change from water to sediment. The entire core was massive and composed of light brown (2.5Y 3/2) sandy silt with a gradual transition to darker brown (2.5Y 4/3) sandy silt towards the base of the core (80–60 cm). Loss-on-ignition analysis revealed that the sediment core was mostly comprised of inorganic material (mean = 84.4 %, SD = 5.3 %), with a mean organic content of 14.2 % (SD = 5.2 %), and minimal carbonate content (mean = 1.4 %, SD = 0.28; Fig. 3). The organic matter concentration was fairly constant, except for the interval 71–61 cm (*ca.* 4,600–4,000 cal yr BP), where it achieved a maximum of 33 % (mean = 22.1 %, SD = 4.4 %). Two peaks of organic content occurred during this interval, one at 68 cm (*ca.* 4,300 cal yr BP), and the second at 62 cm (*ca.* 4,000 cal yr BP). Particle size analysis revealed that the sediment was dominated by fine silt

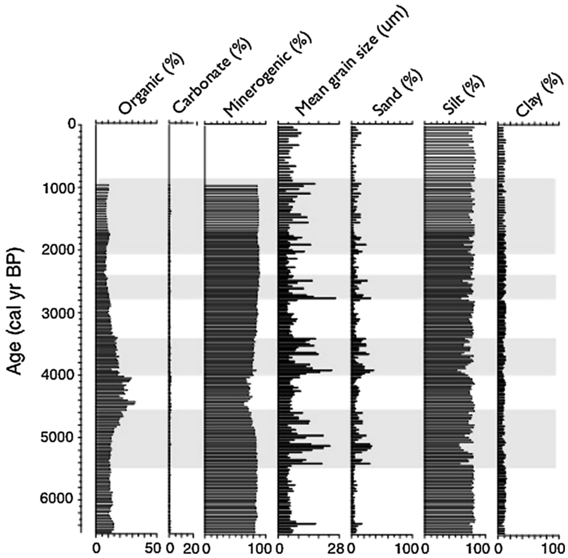

Fig. 3 Comparison of the percent organic, carbonate, and minerogenic content of the sediment core along with the mean particle size and the percent sand, silt, and clay of each sample. The *grey boxes* highlight regions of the core that are punctuated with greater input of larger size particles (sand), suggesting more pronounced, high-energy, erosion during these periods

(mean = 11.2 μm, SD = 3.8 μm), but overall was composed of silt (mean = 77.3 %, SD = 5.8 %), with some sand (mean = 11.1 %, SD = 7.2 %) and clay (mean = 11.7 %, SD = 1.9 %) content. Clay content remained fairly consistent throughout (range 6.4–16 %), whereas intervals of low and relatively high sand content punctuated the core.

Chironomids and temperature inferences

Twenty-nine chironomid taxa were observed in 50 sediment sub-samples analyzed from the sediment–water interface to a depth of 90 cm in the sediment core (Figs. 4, 5; Table 2). Three statistically significant chironomid zones were identified based on comparison of the CONISS results to a broken stick model. The basal zone, Zone 1, spanned 90 to 71 cm (*ca.* 6,000–4,600 cal yr BP) and was dominated by *Corynocera ambigua* type Zetterstedt and *Sergentia* Kieffer, which made up 7.9–70.8 % and 1.8–19.1 % of the assemblage, respectively. The transition to Zone 2 (71–61 cm; *ca.* 4,600–4,000 cal yr BP) was marked by a dramatic decline in the relative abundance of *C. ambigua* type, which falls from ~50 % near the top of Zone 1 to ~10 % in Zone 2. Zone 2 was also characterized by chironomid taxa associated with northern boreal lakes, including *Tanytarsus* van der Wulp, *Psectrocladius* Kieffer, *Polypedilum* Kieffer, *Procladius* Skuse, *Cladotanytarsus* Kieffer and *Cricotopus/Orthocladius* van der Wulp (Walker and MacDonald 1995),

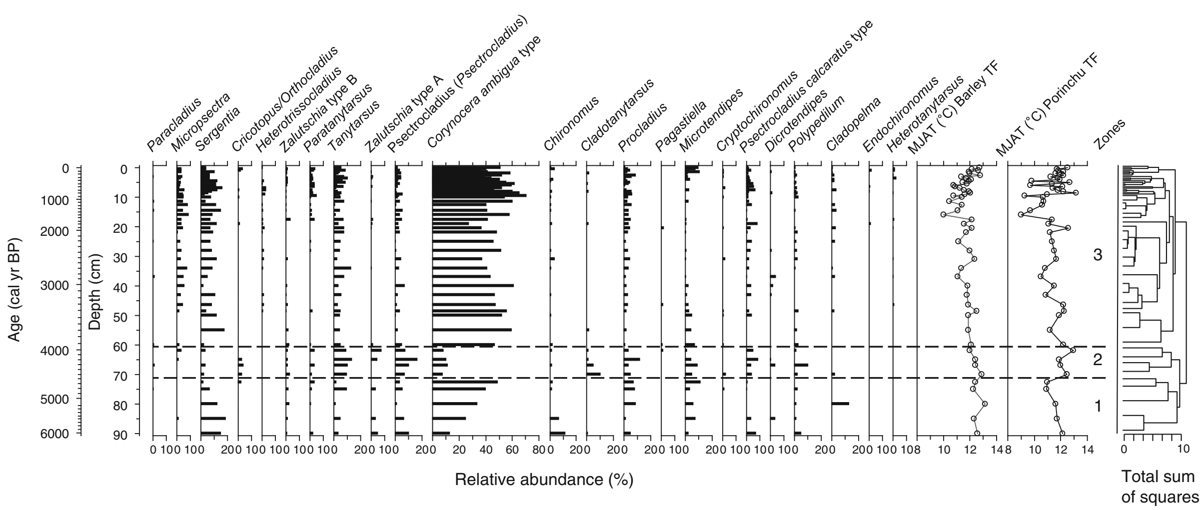

Fig. 4 Chironomid relative abundance and inferred mean July air temperature (MJAT) for Carleton Lake versus sediment core depth over the last *ca.* 6,000 years. Chironomid taxa are ordered left to right from coldest to warmest temperature optima based on the Barley et al. (2006) training set. The inferred MJAT for the two-component WAPLS models described in Barley et al. (2006) and Porinchu et al. (2009) are shown following the relative abundance data. Chironomids not present in more than one sample were omitted from the graph. Significant changes in the chironomid assemblage as determined using constrained incremental sum of squares (CONISS) are shown as stratigraphic zones

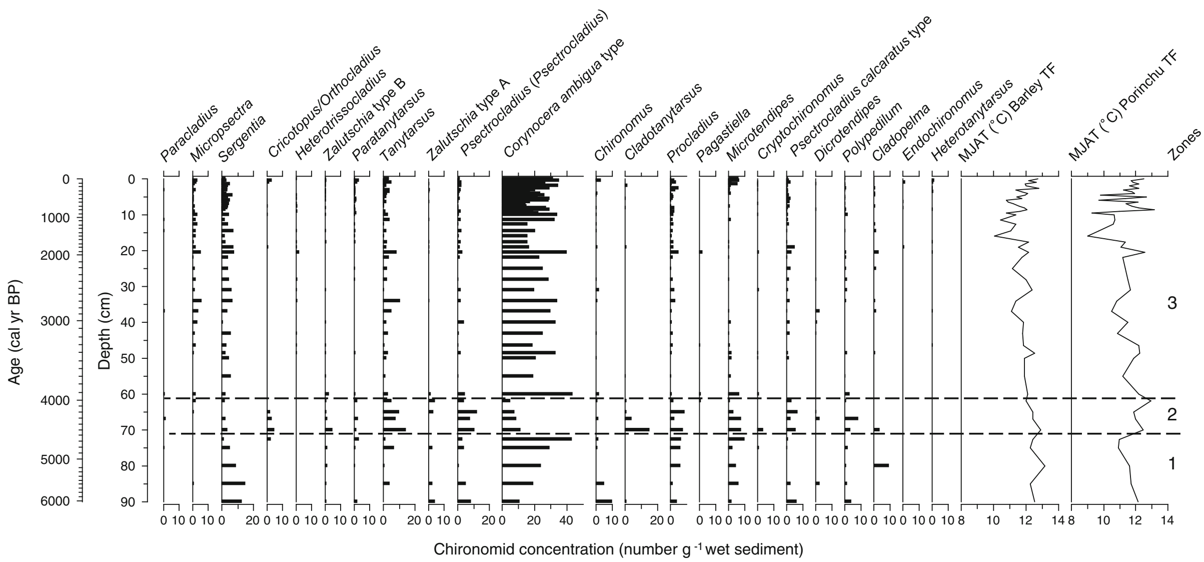

Fig. 5 Chironomid concentrations (number of head capsules per gram of wet sediment) and inferred mean July air temperature (MJAT) for Carleton Lake versus sediment core depth over the last *ca.* 6,000 years. Chironomid taxa are ordered left to right from coldest to warmest temperature optima based on the Barley et al. (2006) training set. The inferred MJAT for the two-component WAPLS models described in Barley et al. (2006) and Porinchu et al. (2009) are shown following the concentration data. Stratigraphic zones based on the chironomid relative abundance data are also shown

which all achieved their greatest relative abundance values.

A return to dominance of *C. ambigua* type and *Sergentia* and a decline of taxa associated with lakes south of treeline occurred in Zone 3 (61–0 cm; *ca.* 4,000 cal yr BP to present). *Corynocera ambigua* type frequently attained relative abundances of >50 %, whereas *Sergentia* was the second most dominant taxon, reaching ~20 %. *Micropsectra* also reached its greatest relative abundance (8.7 %) in the middle part of Zone 3, but declined toward the top of the core.

Principal component analysis demonstrated that axes 1 and 2 explained 28.4 % and 10.5 % of the variation in the chironomid assemblages preserved in the Carleton Lake sediment core, respectively (Fig. 6). Chironomid taxa positively associated with PCA axis 1 included taxa commonly associated with higher temperatures and lakes south of treeline (e.g. *Tanytarsus*, *Psectrocladius*, *Polypedilum*, *Procladius*, *Cladotanytarsus*, *Microtendipes* Kieffer and *Cricotopus/Orthocladius*; Walker and MacDonald 1995), whereas *Corynocera ambigua* type and *Microspectra* Kieffer were negatively related to PCA axis 1. The second principal component largely divided the assemblage based on the presence of *Paratanytarsus* Thienemann and Bause and *Sergentia*.

Analogue comparison showed that the down-core chironomid assemblages had overall poor modern analogues in the Barley et al. (2006) training set, but stronger analogues to the Porinchu et al. (2009) training set (Fig. 7). This is likely because of the dominance of *C. ambigua* type in the record, which often exceeded >50 % of the total chironomid abundance. In comparison, only ~5 % of lakes in the Barley et al. (2006) training set had relative abundance of *C. ambigua* type >25 % whereas approximately 20 % of lakes in the Porinchu et al. (2009) training set had relative abundance of *C. ambigua* type >25 %. The significance test proposed by Telford and Birks (2011) revealed that our reconstructed temperature values, using both transfer functions, did not explain a significant proportion of the variability in the chironomid assemblage through time (Fig. 7).

Both the Barley et al. (2006) and the Porinchu et al. (2009) transfer functions yielded broadly similar patterns of temperature change over the last 6,000 years, although the magnitude of inferred temperature change was greater using the Porinchu et al. (2009) model compared to the Barley et al. (2006) model (Fig. 8). The difference in the magnitude and variability of the temperature inferences is driven largely by the range of temperature estimates

Table 2 The number of occurrences, Hill's N2 diversity index, and the minimum (Min), Maximum (Max), mean and median chironomid relative abundances for the Carleton Lake sediment core

Name	Number of occurrences	Hill's N2	Min	Max	Mean	Median	Temp inference
Chironomus	21	10	0	12	1	0	Incl
Cladopelma	16	8	0	13	1	0	Incl
Cladotanytarsus	10	5	0	11	1	0	Incl
Corynocera ambigua type	50	45	8	71	45	47	Incl
Cricotopus/Orthocladius	9	8	0	4	0	0	Incl
Cryptochironomus	13	11	0	3	0	0	Incl
Dicrotendipes	9	7	0	4	0	0	Incl
Endochironomus	4	3	0	2	0	0	Incl
Heterotanytarsus	10	7	0	2	0	0	Incl
Heterotrissocladius	18	14	0	3	1	0	Incl
Micropsectra	41	31	0	9	3	3	Incl
Microtendipes	33	20	0	11	3	2	Incl
Pagastiella	4	4	0	2	0	0	Incl
Paracladius	8	7	0	2	0	0	Incl
Paratanytarsus	33	26	0	5	2	1	Incl
Polypedilum	29	18	0	11	1	1	Incl
Procladius	46	34	0	12	4	3	Incl
Psectrocladius (*Psectrocladius*)	47	30	0	17	4	3	Incl
Psectrocladius calcaratus type	44	30	0	9	3	2	Incl
Sergentia	48	38	0	19	8	7	Incl
Tanytarsus	46	34	0	14	5	5	Incl
Zalutschia Type A	15	8	0	8	1	0	Incl
Zalutschia type B	29	22	0	4	1	1	Incl
Chironomini indeterminable	38	30	0	4	1	1	Excl
Glyptotendipes	1	1	0	1	0	0	Excl
Orthocladiinae indeterminable	48	36	0	8	3	3	Excl
Paracricotopus	1	1	0	2	0	0	Excl
Psectrocladius (*Mseopsectrocladius*)	1	1	0	3	0	0	Excl
Pseudochironomus	1	1	0	1	0	0	Excl
Stempellinella-Zavrelia	1	1	0	2	0	0	Excl
Tanytarsini indeterminable	49	42	0	17	7	7	Excl
Indeterminable	48	34	0	14	5	5	Excl

The Temp inference column indicates if the chironomid group was included (Incl) or excluded (Excl) from the temperature inference model

during colder periods. Whereas both models estimate a maximum temperature of 13.1 °C over the last 6,000 years, the minimum temperature estimated by the Barley et al. (2006) is 10.2 °C, compared to 9.0 °C based on the Porinchu et al. (2009) model. Overall, both temperature reconstructions showed a general cooling trend from approximately 6,000 cal yr BP, although the temperature estimates based on the Porinchu et al. (2009) transfer function have more pronounced variability. The coldest inferred temperatures over the last 6,000 years occurred between approximately 1,550 and 900 cal yr BP. Inferred temperatures then began to increase, peaking at *ca.* 800 cal yr BP, after which the inferred temperatures declined again to *ca.* 600 cal yr BP. After the *ca.* 600 cal yr BP minimum, inferred temperatures subsequently rose to 12.5 °C at the top of the core for both models.

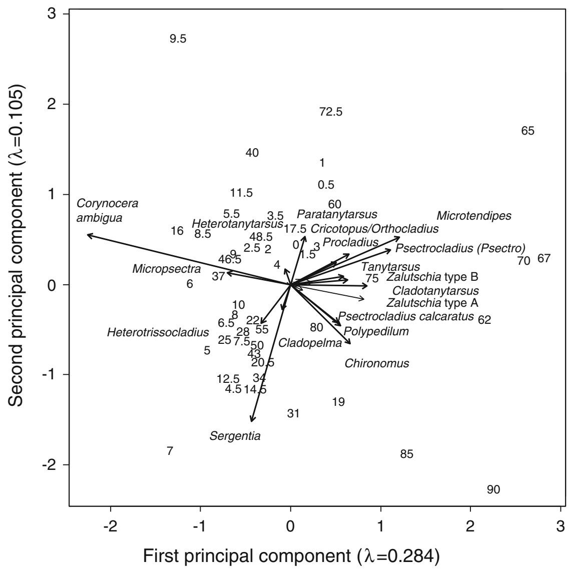

Fig. 6 Axis 1 and 2 of a principal component analysis (PCA) of the chironomid relative abundance data. PCA sample scores are shown as *open circles* and the corresponding sediment depth of the sample in cm. Axis 1 explained 28.4 % of the variation in the chironomid assemblage and largely a division of taxa associated with lakes in forested (e.g. *Tanytarsus*, *Psectrocladius*, *Polypedilum*, *Procladius*, *Cladotanytarsus*, *Microtendipes* and *Cricotopus/Orthocladius*) versus tundra (e.g. *Corynocera ambigua* type and *Microspectra*) environments (Walker and MacDonald 1995). The second principal component explained 10.5 % of the variation in the chironomid assemblage

Discussion

The largest shift in the chironomid community of Carleton Lake occurred *ca.* 4,600–4,000 cal yr BP, when *C. ambigua* type and *Sergentia*, taxa more closely associated with tundra lakes (Walker and MacDonald 1995; Porinchu et al. 2009), were replaced by *Tanytarsus*, *Psectrocladius*, *Polypedilum*, *Procladius*, *Cladotanytarsus*, and *Cricotopus/Orthocladius*, which are more common in boreal environments (Walker and MacDonald 1995; Larocque et al. 2006; MacDonald et al. 2009). This assemblage shift suggests that the catchment of Carleton Lake may have become forested at this time, although analysis of the pollen, stomata, and the macrofossil record are needed to test this hypothesis. The timing of this putative transition from tundra to forest environment and then back to tundra conforms to the reported regional expansion and subsequent contraction of *Picea* forest near our study area in the central NT (MacDonald et al. 1993; Pienitz et al. 1999; Huang et al. 2004; Fig. 9). It has been shown that shifts in chironomid communities can be driven by a number of factors in addition to temperature, including nutrient enrichment and hypolimnetic oxygen concentration (Brodersen and Quinlan 2006). It may be that the movement of the treeline into the catchment of the lake altered the productivity of the system. The increased organic content and the particle size changes recorded *ca.* 4,600–4,000 cal yr BP suggest a period of higher productivity and/or decreased terrestrial erosion, which is consistent with a landscape stabilized by the establishment of forest vegetation. Around 4,000 cal yr BP, a concomitant rise in sand content and reduced organic matter content were observed, likely reflecting increased erosion and/or a reduction in lake productivity. At that time, the regional treeline had begun to recede southward, as demonstrated by a reduction in *Picea* pollen and diatom-inferred DOC in central NT lakes (MacDonald et al. 1993; Pienitz et al. 1999).

Our chironomid-inferred mean July temperature estimates suggest that the well-documented regional treeline movement during the mid-Holocene was associated with a modest (~1 °C) change in mean July temperature. Around 6,000 cal yr BP, *Picea* pollen percentages (MacDonald et al. 1993) and dissolved organic carbon (DOC; Pienitz et al. 1999) were low, suggesting that the boreal treeline was south of the study region (Fig. 9). Chironomid-inferred July temperatures at that time (*ca.* 6,000 cal yr BP) were ~12 °C based on both transfer function models. At the height of regional forest expansion, between *ca.* 5,500 and 3,500 cal yr BP (Moser and MacDonald 1990; MacDonald et al. 1993; Pienitz et al. 1999; Huang et al. 2004), inferred mean July temperatures reached a maximum of ~13 °C for both transfer function models, although the timing of the maximum inferred temperature varied, occurring at about 5,000 cal yr BP for the Barley et al. (2006) model and 4,000 cal yr BP for the Porinchu et al. (2009) model. Although the change in inferred July temperature associated with treeline movement is within the error of the Barley et al. (2006; RMSEP = 1.46 °C) and the Porinchu et al. (2009; RMSEP = 1.03 °C) models, the range of inferred temperatures is similar in magnitude to the modern relationship between mean July temperature and the northern treeline (12.5–10 °C; MacDonald et al. 2008).

The temperature reconstructions at Carleton Lake failed the significance test proposed by Telford and Birks (2011), indicating that the Holocene temperature

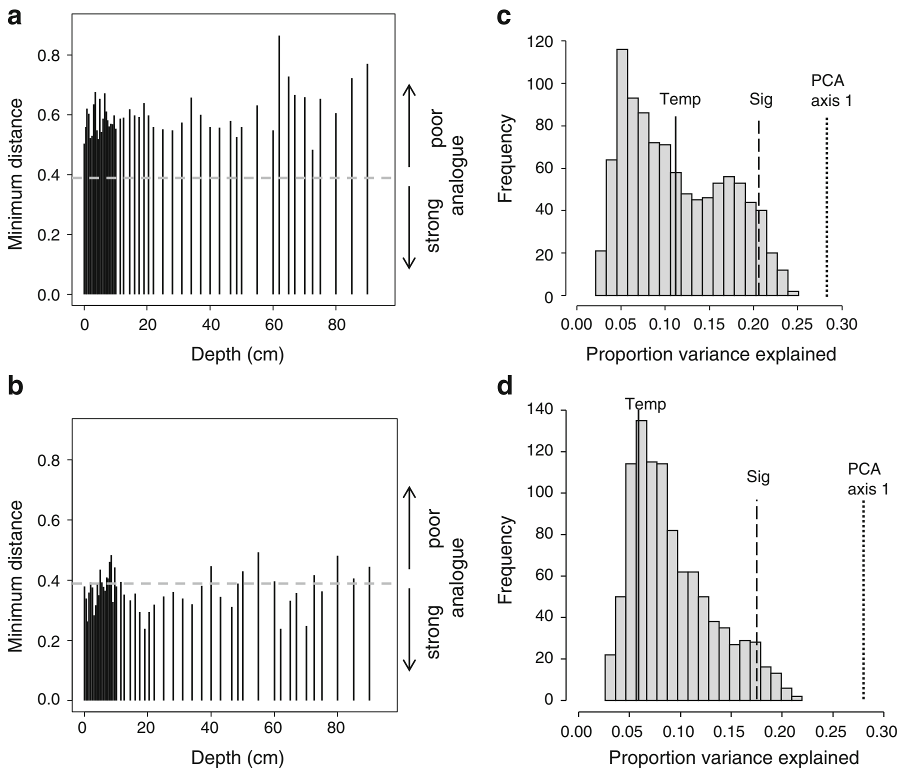

Fig. 7 Analog comparison based on squared chord distance (SCD) as the dissimilarity measure for **a** the Barley et al. (2006) training set and **b** the Porinchu et al. (2009) training set. Samples with minimum dissimilarity coefficient (minimum distance) lower than 5 % of SCD values among the Barley et al. (2006) training set lakes (*grey line*) were considered good analogs, whereas samples greater than 5 % of SCD values among training set lakes were identified as having poor modern analogues in the training sets. Reconstruction significance tests for our inferred temperature (Temp; *solid line*) values based on **c** the Barley et al. (2006) transfer function and **d** the Porinchu et al. (2009) transfer function. Reconstructions are considered significant (Sig; *dashed line*) if they explain more variation in the subfossil assemblage than 95 % of the reconstructed variables based on randomly generated transfer function models (Telford and Birks 2011). For comparison the variance explained by PCA axis 1 (*dotted line*) is also shown

reconstructions over the last 6,000 years are not statistically significant. We have, however, included a discussion of some of the observed, although non-significant, climate variability because we believe this discussion will be of value for potential future research on Holocene climate variability, such as a meta-analysis. Late Holocene climate variability was modest in our record. The largest change in temperature was an inferred cooling event that occurred between *ca.* 1,500 and 900 cal yr BP, when estimated July temperatures reached a minimum in the record of 10.2 or 9.0 °C for the Barley et al. (2006) and the Porinchu et al. (2009) transfer functions, respectively. This cooling event is >2 standard deviations from the mean Holocene temperature of the record. This cold period is consistent with the timing of First Millennial Cooling (Reyes et al. 2006; Clegg et al. 2010; Thomas et al. 2011) observed in other high-latitude North American proxy records. This cold event was briefly interrupted by a *ca.* 200-year relative warming event centered around 800 cal yr BP, coeval with the Medieval Climate Anomaly, when, based on the Porinchu et al. (2009) transfer function inferences, summer air temperatures reached 13.1 °C, the highest inferred temperature for the entire record. Following this warm event, temperatures then declined between 700 and 350 cal yr BP. Our temperature reconstructions suggest that in the central NT the First Millennial Cooling event was colder and of longer duration (~500 years longer) than the Little Ice Age. The consistent timing of First Millennial Cooling across British Columbia (Reyes et al. 2006), Alaska (Hu et al.

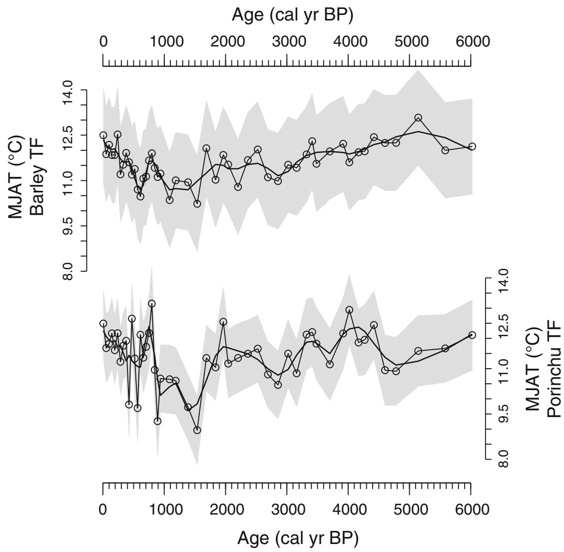

Fig. 8 Chironomid-inferred mean July air temperatures over the last *ca.* 6,000 years from Carleton Lake based on the Barley et al. (2006; Barley TF) and the Porinchu et al. (2009; Porinchu TF) transfer functions. Inferred temperature values are shown as *open circles* and smoothed using a LOESS smoother (span = 0.18; *thick black line*). Sample-specific error is indicated by *grey shading*

2001; Reyes et al. 2006; Clegg et al. 2010), the Canadian Arctic Archipelago (Thomas et al. 2011), and in the central Canadian subarctic (this study), suggests that this late Holocene cold event was a widespread climate anomaly.

The chironomid-inferred temperature record from Carleton Lake is broadly similar with interpretations based on a shorter (~2,000 cal yr BP) chironomid study from nearby Lake S41 (MacDonald et al. 2009). Both chironomid-based temperature reconstructions show warming during the Medieval Climate Anomaly (*ca.* 1,000–700 cal yr BP; Mann et al. 2009), followed by cooler temperatures during the Little Ice Age (*ca.* 550–250 cal yr BP; Mann et al. 2009). It is plausible that differences between inferred temperatures from these two sites are driven by local, lake-specific reactions to larger-scale radiative forcing. For example, the greatest discrepancy between the chironomid-inferred temperature of Carleton Lake and that of Lake S41 is that the Carleton Lake record shows warming following the Little Ice Age, which is not evident in the MacDonald et al. (2009) reconstruction from Lake S41. MacDonald et al. (2009) suggest that warmer nineteenth- and twentieth-century temperatures may have led to increased productivity (estimated using LOI and biogenic silica) and duration of stratification in Lake S41. These changes may have resulted in low-oxygen conditions that confounded their chironomid-inferred air temperatures in this section of their record (MacDonald et al. 2009). Quinlan et al. (2012) have shown that the thermal regime of a lake can influence chironomid-based environmental inferences, lending support to our suggestion that the discrepancy between the Carleton Lake and S41 temperature reconstructions for the twentieth century may be a consequence of greater stratification in Lake S41. Because the Carleton Lake core was taken from a shallower location (1.5 m water depth), the potential influence of thermal stratification and low oxygen in the hypolimnion were minimized.

Our chironomid-based temperature reconstruction is consistent with the timing of reported climate events in the central NT and broader western North American high-latitude patterns, such as the Holocene Thermal Maximum, First Millennial Cooling (centered around 1,400 cal yr BP), the Medieval Climate Anomaly (*ca.* 1,000–700 cal yr BP; Mann et al. 2009), the Little Ice Age (*ca.* 550–250 cal yr BP; Mann et al. 2009), and twentieth-century warming. Further, the magnitude of the temperature inferences is consistent with both historical treeline movement and the instrumental record (modern mean July temperature: chironomid-inferred ~12.5 °C based on both the Barley et al. (2006) and the Porinchu et al. (2009) transfer functions; estimated instrumental average between 1942 and 2011 ~12.8 °C). Despite the general agreement between our temperature record and other regional and western North American climate reconstructions, our subfossil chironomid assemblages had no strong modern analogues in the Barley et al. (2006) surface sample training set, and our temperature reconstructions failed the significance test recommended by Telford and Birks (2011). Furthermore, in line with other Holocene chironomid-based temperature reconstructions (MacDonald et al. 2009; Clegg et al. 2010; Thomas et al. 2011), the magnitude of temperature variation is largely within the error envelope of the transfer function models. Our reported temperature reconstruction from Carleton Lake therefore needs to be considered with a degree of caution.

As temperature was not a significant driver of changes in the chironomid assemblage through time, other factors may have been important controls on the structure of chironomid communities in Carleton

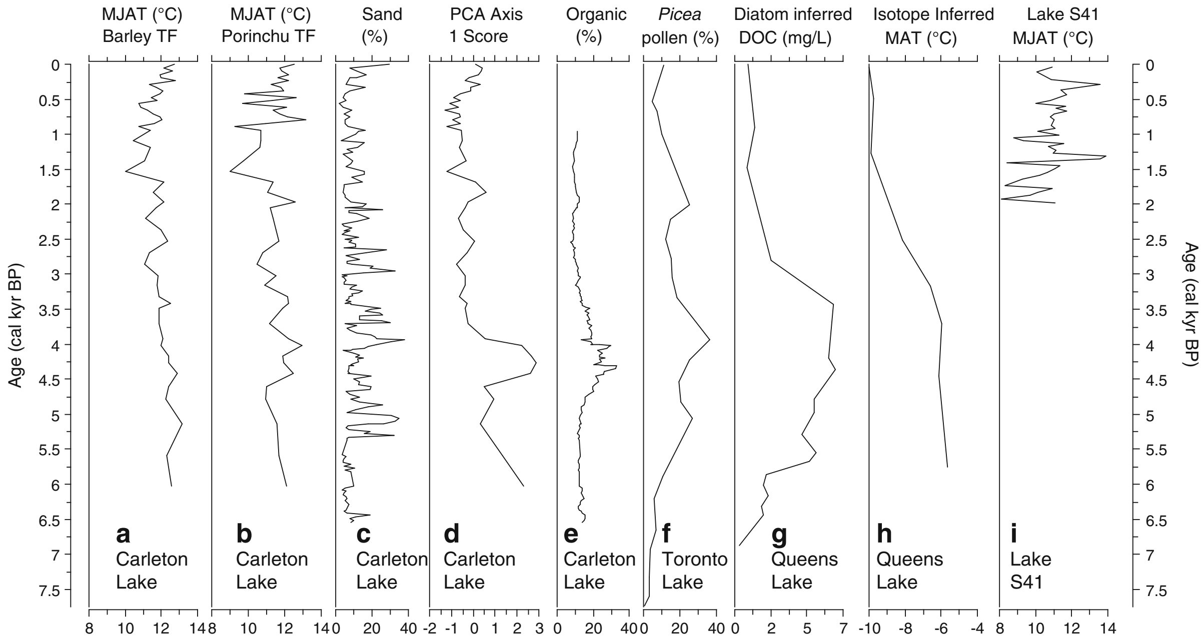

Fig. 9 Summary diagram of paleoenvironmental proxies investigated in the Carleton Lake sediment core and selected records from the area surrounding Carleton Lake. Chironomid-inferred mean July air temperatures from Carleton Lake based on (**a**) the Barley et al. (2006) and (**b**) the Porinchu et al. (2009) transfer functions. (**c**) Percent sand in Carleton Lake sediment samples. (**d**) Principal component analysis axis 1 sample scores based on the Carleton Lake chironomid relative abundance data. (**e**) Percent organic content of the Carleton Lake sediment samples based on loss-on-ignition. (**f**) Relative abundance of *Picea* (spruce) pollen in a lake sediment record from nearby Toronto Lake (MacDonald et al. 1993). (**g**) Diatom-inferred dissolved organic carbon (DOC) from nearby Queen's Lake (Pienitz et al. 1999). (**h**) Isotope-inferred ($\delta^{18}O$) mean annual temperature (MAT) from Queen's Lake (Edwards et al. 1996; Pienitz et al. 1999). (**i**) Chironomid-inferred mean July air temperatures over the last *ca.* 2,000 years from nearby Lake S41 (MacDonald et al. 2009)

Lake. Between *ca.* 4,600 and 4,000 cal yr BP there is a substantial change in the chironomid assemblage to taxa that are known to be associated with the presence of trees (Fig. 8). The fact that the major change in our chironomid record may be a response to treeline movement and occurs over a small temperature gradient may explain why our temperature reconstruction failed the Telford and Birks (2011) significance test.

Inferred July temperature from the uppermost sediment sample was ~12.5 °C based on both transfer function models, suggesting that treeline could once again advance north of the study site. The current magnitude of warming is comparable to that inferred for the Holocene Thermal Maximum. Inferred July temperatures during the Holocene Thermal Maximum averaged ~12.6 °C, similar to the ~12.5 °C inferred at the top of our sediment record. Given that modern mean July air temperatures are relatively high in our study area, based both on our chironomid-based inferences (~12.5 °C) and the average of the instrumental record calculated in this study (~12.8 °C), it seems probable that the present-day forest limit may reflect a time lag in the response of the vegetation to warming temperatures, rather than to unfavourable temperature conditions at our study site. MacDonald et al. (1993) reported that the transition from tundra to forest-tundra that occurred in this region *ca.* 5,000–4,000 years ago, took only 150 years. It is therefore plausible that if the modern warming trend continues, within the next century the terrestrial landscape in the study region will transition, once again, from tundra to forest.

Conclusions

The lack of statistically significant chironomid temperature reconstructions at our study site may indicate that changes in the chironomid assemblages through time at Carleton Lake were primarily driven by factors other than air temperature. At Carleton Lake the largest shift in the chironomid assemblage occurred during a time period of known treeline advance and subsequent retreat in the region. This suggests that overall, the presence or absence of trees may have had a greater effect on the chironomid community than temperature alone. The lack of a statistically significant temperature trend indicates that these temperature reconstructions need to be interpreted with caution. Our quantitative, inferred mean July air temperature reconstructions at the millennial-scale, however, do agree with other records of Holocene climate and show a trend of warmer than present temperatures during the Holocene Thermal Maximum (*ca.* 6,000–4,000 cal yr BP), followed by a general cooling trend that has reversed in approximately the last 600 years. Our results further suggest that warming did occur during the period of regional treeline advance (centered around *ca.* 4,500 cal yr BP), highlighting the connection between summer temperatures and the position of the northern treeline.

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