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Paleoecological evidence for abrupt cold reversals during peak Holocene warmth on Baffin Island, Arctic Canada

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ABSTRACT

A continuous record of insect (Chironomidae) remains preserved in lake sediments is used to infer temperature changes at a small lake in Arctic Canada through the Holocene. Early Holocene summers at the study site were characterized by more thermophilous assemblages and warmer inferred temperatures than today, presumably in response to the positive anomaly in Northern Hemisphere summer insolation. Peak early Holocene warmth was interrupted by two cold reversals between 9.5 and 8 cal ka BP, during which multiple cold-stenothermous chironomid taxa appeared in the lake. The earlier reversal appears to correlate with widespread climate anomalies around 9.2 cal ka BP; the age of the younger reversal is equivocal but it may correlate with the 8.2 cal ka BP cold event documented elsewhere. Widespread, abrupt climate shifts in the early Holocene illustrate the susceptibility of the climate system to perturbations, even during periods of enhanced warmth in the Northern Hemisphere.

Introduction

Although the climate of the Holocene has been more stable than that of the preceding glacial period, there is mounting evidence for high-amplitude climate variations during the Holocene, particularly in the North Atlantic region (e.g., Mayewski et al., 2004; Alley and Ágústdóttir, 2005). Large Holocene climate changes imply undiscovered climate forcings and/or greater than expected sensitivity to known forcings (e.g., due to positive feedbacks in the climate system). Superimposed over known changes in solar insolation through the Holocene are a number of less understood climate forcings, including volcanic eruptions and solar variability, and perturbations such as freshwater outbursts to the North Atlantic Ocean (e.g., Schmidt et al., 2004). Estimates of climate sensitivity and feedbacks are a major source of uncertainty in projections of future climate change (Knutti et al., 2002; Stainforth et al., 2005; IPCC, 2007). Paleoclimate data provide our only empirical observations of long-term climatic responses to natural forcings, and therefore make important contributions to improving climate projections.

The eastern Canadian Arctic occupies a pivotal location for paleoclimate reconstruction: Baffin Island (Fig. 1) is 500 km west of Greenland; thus paleoclimate records from Baffin Island provide a regional context for Greenland records and clarify how climate events recorded in central Greenland ice cores affected nearby land masses. This high-latitude region, like the rest of the Arctic, may be especially sensitive to past and future changes in climate forcing. Increased radiative forcing is amplified in the Arctic by positive feedbacks, and the Arctic in turn plays an important role in Earth's energy budget (Holland and Bitz, 2003; Chapin et al., 2005; Serreze and Francis, 2006). Instrumental records indicate that the Arctic has warmed faster than the global average over the past century, and model simulations predict enhanced future warming at high northern latitudes (Serreze et al., 2000; ACIA, 2005; IPCC, 2007). Paleoclimate data provide a basis for constraining the magnitude of this amplified response to radiative forcing, and means for testing whether polar amplification is indeed a persistent feature of arctic climate.

This study reconstructs Holocene summer temperatures at Lake CF8 on Baffin Island (Fig. 1) using subfossil insect (chironomid or non-biting midge; Diptera: Chironomidae) assemblages as a proxy for summer air temperatures. Our results add to the small database of records that quantify temperatures throughout the Holocene in the North American Arctic (e.g., reviewed by Kaufman et al., 2004; Kaplan and Wolfe, 2006), and to our knowledge this is the first study in the eastern Canadian Arctic to reconstruct terrestrial temperatures at adequate temporal resolution to identify abrupt early Holocene events.

Materials and methods

Lake CF8 (informal name) is a small, shallow lake (0.3 km^2, $depth_{max}$ = 10 m) situated on an inter-fjord lowland on northeastern Baffin Island in Nunavut, Arctic Canada, at 70° 33.37'N, 68° 57.14'W (Fig. 1). The lake sits at 195 m asl and was overridden during the last

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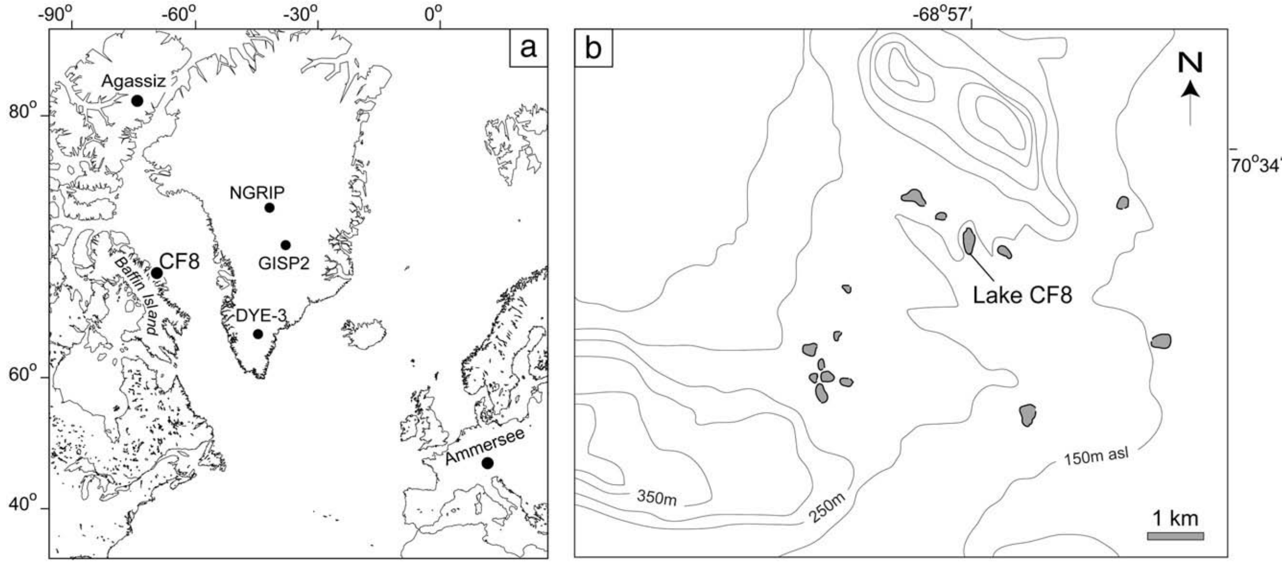

Figure 1. (a) Overview map of the North Atlantic region, showing locations of the study site (Lake CF8); the Agassiz Ice Cap; NGRIP, GISP2 and DYE-3 ice cores; and lake Ammersee; and (b) topographic map of the landscape surrounding Lake CF8 on northeastern Baffin Island.

glacial cycle by cold-based portions of the Laurentide ice sheet (Briner et al., 2005; 2007a). Deglaciation of the lowland was complete by ~12 cal ka BP (Briner et al., 2005; 2007b). The lake has no permanent inflow stream and presumably receives surface water mainly from snowmelt. At the nearby village of Clyde River, modern mean annual temperature is −12.8°C and mean July temperature is +4.4°C (Environment Canada, 2007).

A single-drive sediment core (02-CF8-01) containing 120 cm of Holocene gyttja underlain by sand was recovered from the deepest point in Lake CF8 using a percussion corer. Fifteen AMS ^{14}C ages (Table 1; Fig. 2) were obtained on aquatic bryophyte macrofossils, which are known to equilibrate with atmospheric CO_2 in similar crystalline terrains of Baffin Island (Wolfe et al., 2004). All ages were calibrated using Calib v 5.0.2 (Stuiver et al., 2005) and the IntCal04 calibration curve (Reimer et al., 2004). The age model used for this study is a polynomial fit to all but the two deepest ages, which are inverted (Fig. 2). The inverted ages make the age model relatively uncertain below 107.5-cm depth (10.5 cal ka BP), where two different macrofossils yielded statistically overlapping ages. 95% confidence intervals on the polynomial regression overlap with the 2σ calibrated age ranges of all but one of the ages used in the model (Fig. 2) and provide estimates of uncertainty for age assignments throughout the record.

Chironomid larvae are aquatic, and the larval head capsules are chitinous and generally well preserved in lake sediments. Chironomid assemblages are very sensitive to temperature change, and subfossil chironomids have been used to reconstruct both abrupt and low-amplitude temperature changes (e.g., Walker et al., 1991b; Brooks and Birks, 2000a,b; Cwynar and Spear, 2001; Larocque and Hall, 2003; Brooks and Birks, 2004; Heiri et al., 2004; Caseldine et al., 2006; Axford et al., 2007, 2008; Porinchu et al., 2007), provided that changes in other environmental factors (e.g., lake depth or chemistry) over

Table 1
Radiocarbon ages from core 02-CF8-01, Baffin Island

Lab number[a]	Depth (cm)	δ 13C	Fraction modern	14C age (14C yr BP)	Calibrated age (cal yr BP)[b]
AA60640	2	−24.8	0.9111±0.0040	748±35	695±35
AA60641	31	−26.6	0.6766±0.0052	3138±62	3360±190
AA60642	50	−29.1	0.5596±0.0028	4664±41	5440±130
CURL-8927	58	−29.2	0.4677±0.0009	6105±20	7025±125
CURL-8427	66.5	−24.7	0.3937±0.0009	7490±20	8295±85
CURL-8428	70.5	−21.7	0.3962±0.0009	7440±20	8260±70
AA60643	71	−24.2	0.3996±0.0034	7368±68	8185±155
CURL-8929	77	−24.0	0.3836±0.0012	7695±30	8480±60
CURL-8429	84	−24.6	0.3515±0.0008	8400±20	9410±75
AA60644	88	−29.0	0.3530±0.0022	8365±50	9375±115
CURL-8896	98	−27.8	0.3215±0.0008	9115±25	10300±75
CURL-6954	107	−30.5	0.3133±0.0017	9320±45	10490±185
CURL-8271	107.5	−21.4	0.3119±0.0011	9360±30	10590±85
CURL-8142	113	−16.2	0.2794±0.0009	10245±30	11965±140
AA60645	121	−26.5	0.3457±0.0037	8532±86	9505±195

[a] All samples are aquatic bryophyte macrofossils. Samples with AA-lab numbers were submitted to the Arizona Accelerator Mass Spectrometry Laboratory; samples with CURL-lab numbers were submitted to the University of Colorado INSTAAR Laboratory for AMS Radiocarbon Preparation and Research.

[b] Each calibrated age is the midpoint±1/2 the 2σ range calculated using CALIB 5.0.2 (Stuiver et al., 2005).

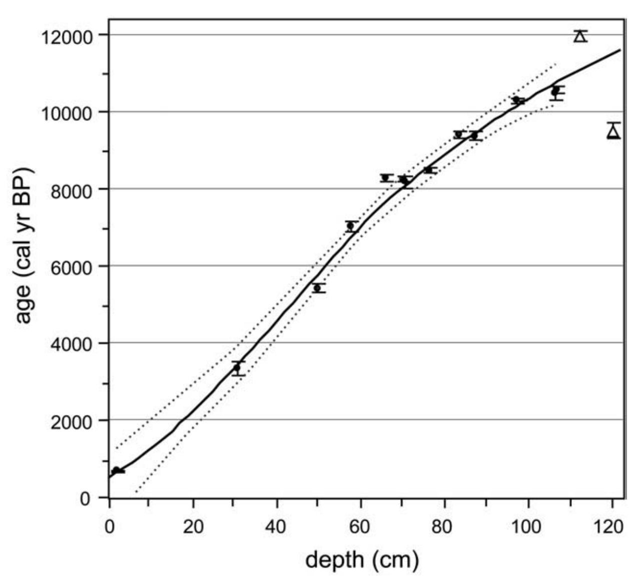

Figure 2. ^{14}C chronology from Lake CF8 (n=15; two ages shown as open triangles are excluded from polynomial fit). 2σ calibrated age ranges are shown for each age. Dashed gray lines are 95% confidence intervals.

time do not overwhelm the influence of temperature (e.g., Velle et al., 2005; Antonsson et al., 2006; Brooks, 2006). For this study, at least 50 whole head capsules per sample were extracted and prepared following standard methods (e.g., Walker, 2001). Forty-six samples were analyzed, with sampling concentrated in the early Holocene. July air temperature inferences were derived from the transfer function published by Francis et al. (2006), who added 29 calibration sites from Baffin Island to a training set of 39 sites across eastern Canada (Walker et al., 1991a; 1997). The weighted-averaging regression model uses square-root transformed species data and has a root mean squared error of prediction (RMSEP) of 1.5°C for mean July air temperatures. Paleotemperatures were calculated using the computer program C2 v 1.4.3 (Juggins, 2003). One of the 19 taxa identified in Lake CF8 sediments (*Paracladopelma*) is not represented in the training set and therefore was not used for quantitative temperature inferences. *Paracladopelma* has a maximum abundance of <2% in the subfossil samples; thus, >98% of the taxa in subfossil samples are present in the training set. Canonical correspondence analysis (CCA) was used to test for analogues to subfossil assemblages in the training set. CCA was conducted using the statistical software program R v 2.5.1; rare taxa were downweighted, lake area data square-root transformed, and lake depth data log-transformed.

Qualitative environmental proxies provide supplementary information about changing paleoenvironments. The organic carbon content of lake sediments is a function of aquatic and terrestrial productivity, and influx of inorganic materials. Percent loss-on-ignition (%LOI) is highly correlated with the total carbon content (%C) of sediments in Clyde Foreland lakes (Briner et al., 2006). The %LOI of sediments was measured at 550°C (Heiri et al., 2001) and is reported as weight-percent C of dry sediment. Biogenic silica ($BiSiO_2$) in lake sediments primarily comprises diatoms (e.g., Conley, 1988), and therefore provides a proxy for aquatic primary productivity, although it is also a function of the flux of other materials to the lake. $BiSiO_2$ analysis followed Mortlock and Froelich (1989), except for the use of 10% Na_2CO_3 solution for $BiSiO_2$ extraction. $BiSiO_2$ concentration was measured by spectrophotometry and converted to weight-percent SiO_2 of dry sediments.

Results

At the onset of postglacial lacustrine sedimentation at Lake CF8, the cold stenotherm *Oliveridia/Hydrobaenus* dominated the chironomid assemblage (Fig. 3 and Fig. 4). *Oliveridia/Hydrobaenus* was abruptly replaced by an assemblage dominated by the subtribe Tanytarsina (including *Tanytarsus lugens/Corynocera oliveri* type and *Micropsectra*) beginning ~11 cal ka BP, and inferred temperatures rose rapidly to surpass modern values by ~10.5 cal ka BP. Early Holocene assemblages are dominated by Tanytarsina along with Tanypodinae, *Sergentia*, *Chironomus*, and Orthocladiinae (including *Cricotopus/Orthocladius*, *Corynoneura/Thienemanniella*, and *Psectrocladius* among others). The presence of relatively thermophilous taxa (e.g., *Psectrocladius* and the Tanypodinae), combined with the absence of cold stenotherms (including cold-tolerant taxa that were present in both the late glacial and late Holocene) indicate relatively warm temperatures in the early Holocene. In the eastern Canada training set, for example, *Psectrocladium* has an optimum (13.7°C; tolerance 3.6°C; Francis et al., 2006) far above present-day mean July temperature at Lake CF8 (~4.4°C), and this genus is not found in modern sediments from the lake (personal observation; Thomas et al., 2008). Accordingly, inferred temperatures for most of the early Holocene are nearly 5°C (±1.5°C) higher than for the late Holocene. The %LOI is generally higher for this early Holocene warm period than for the middle and late Holocene, whereas %$BiSiO_2$ is lowest for the early Holocene (Fig. 3).

Early Holocene warmth was interrupted by two inferred cold reversals between 9.5 and 8 cal ka BP (Fig. 4). Lower inferred temperatures during these two periods reflect clear changes in species assemblages, including the appearance (albeit in small numbers) of the cold stenotherms *Pseudodiamesa* (optimum 5.9°C, tolerance 0.9°C) and *Abiskomyia* (optimum 6.2°C, tolerance 0.8°C; Francis et al., 2006), which were otherwise absent throughout the early Holocene after the

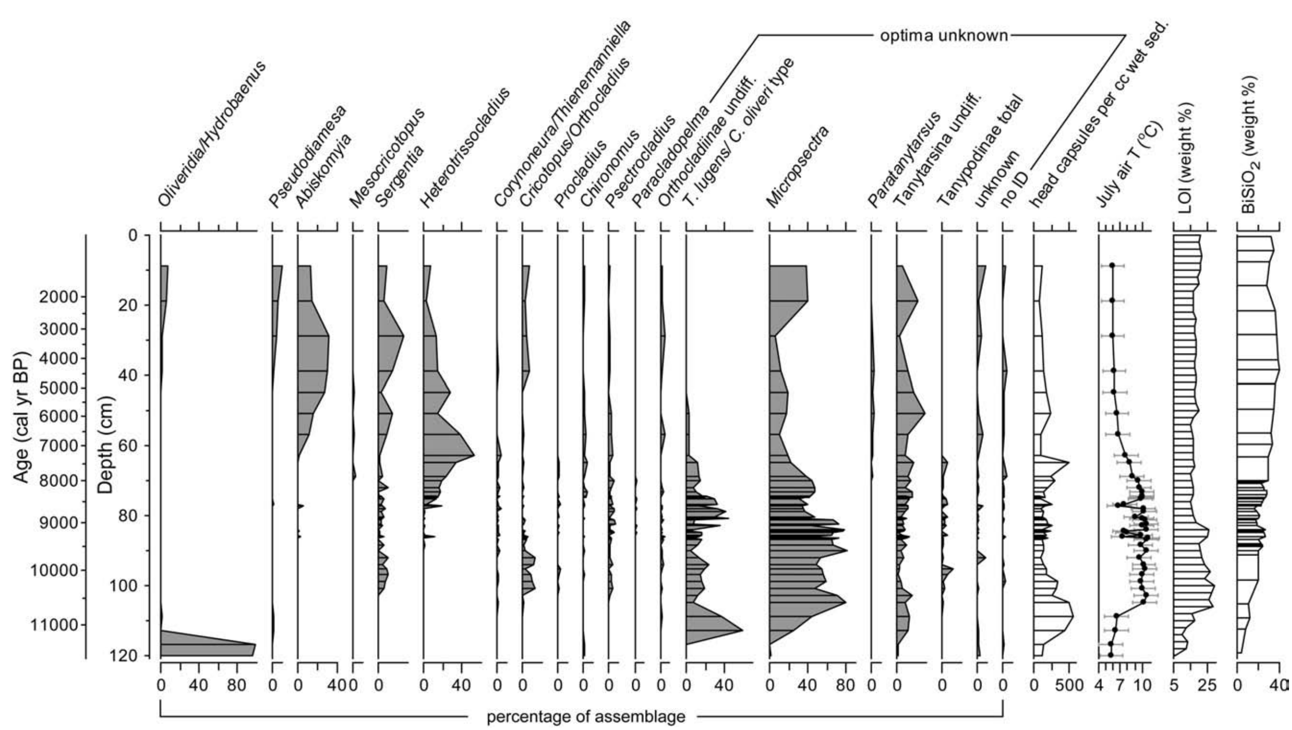

Figure 3. Figure showing percentages of all chironomid taxa, head capsule concentrations, chironomid-inferred July air temperatures, percent loss-on-ignition, and percent biogenic silica versus depth and age. Taxa on the left side of the diagram are ordered by temperature optimum from coldest to warmest (optima of remaining taxa are unknown). *Tanytarsus lugens/Corynocera oliveri*, *Micropsectra*, *Paratanytarsus*, and Tanytarsina undiff are lumped as "subtribe Tanytarsina" in the training set of Francis et al. (2006), and therefore also in temperature modeling. Error bars on temperature inferences are the model RMSEP.

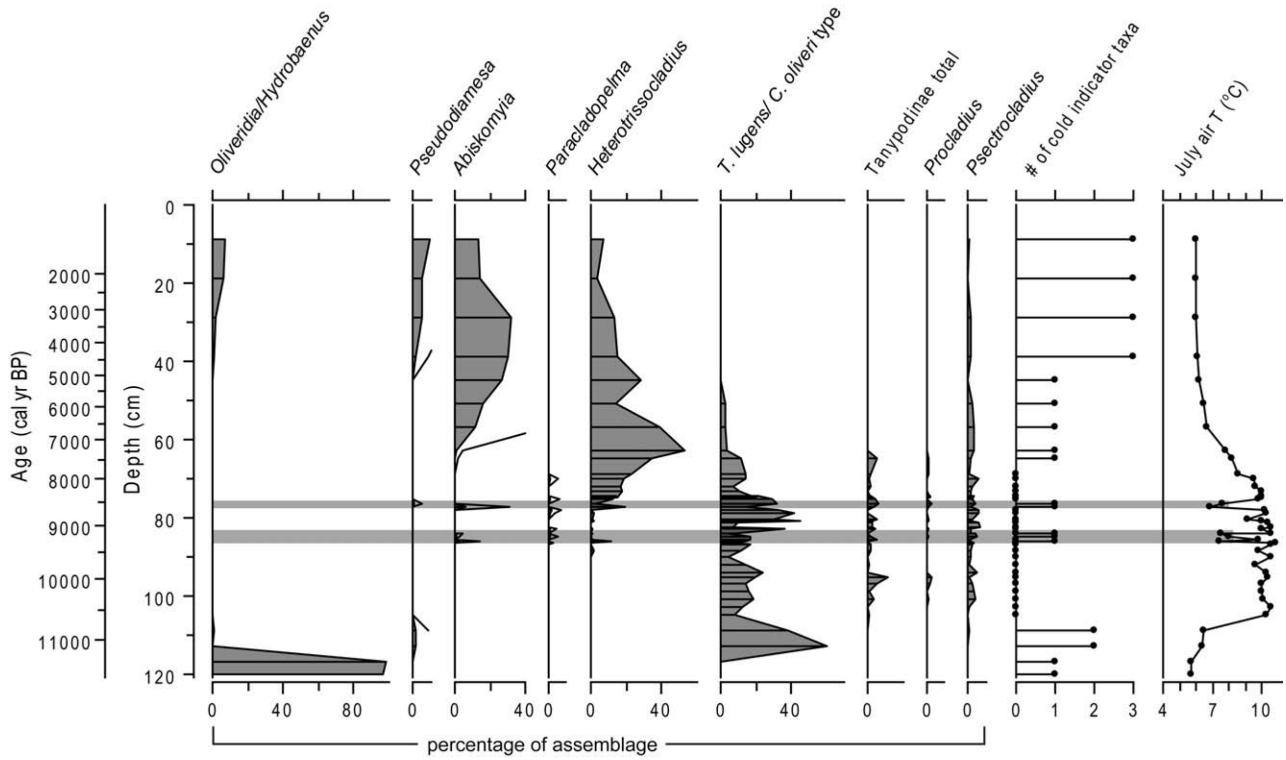

Figure 4. Percentages of selected chironomid taxa, number of cold indicator taxa present in each sample (indicator taxa are *Oliveridia/Hydrobaenus*, *Pseudodiamesa*, and *Abiskomyia*) and chironomid-inferred July air temperatures. Unfilled white curves on *Pseudodiamesa*, *Abiskomyia*, and *Paracladopelma* diagrams are 5× exaggerated. Horizontal gray bars highlight inferred cold reversals.

onset of warmth (Fig. 4). *Heterotrissocladius* (optimum 8.7°C, tolerance 3.3°C; Francis et al., 2006), which was rare during the early Holocene but abundant in the late Holocene, spiked in abundance during the cold events. *Paracladopelma* appeared during and immediately after the inferred reversals. The earlier reversal corresponds with a bryophyte-rich layer from 86–84 cm depth and correspondingly high %LOI (Fig. 3). The two temperature reversals were of similar amplitude, and both lasted 250 yr or less. The two cold periods were separated by several centuries of near-peak inferred warmth.

Following a return to warm conditions after the latter cold reversal, the site began to cool ~7.8 cal ka BP. Sedimentation rates decreased slightly in the late Holocene, a period characterized by the disappearance of *T. lugens*/*C. oliveri* type, and greater percentages of taxa with cold affinities, including *Abiskomyia*, *Oliveridia/Hydrobaenus*, and *Pseudodiamesa*. Tanypodinae disappeared ~7.5 cal ka BP and did not reappear. The cold stenotherms *Oliveridia/Hydrobaenus* and *Pseudodiamesa* became increasingly abundant after ~5 cal ka BP, with peak percentages in the last millennium, indicating intensified Neoglacial cooling.

Because the coldest site in the training set has a mean July air temperature of 5.0°C (Francis et al., 2006), the transfer-function calibration data do not contain adequate analogs for colder-than-present times when cold stenotherms were more abundant than today at our relatively cold study site (mean July air temperature 4.4°C). Therefore, the obvious chironomid assemblage changes in the late Holocene do not result in correspondingly lower temperature inferences (Fig. 3 and Fig. 4). For this reason, the amplitude of late Holocene temperature variation is presumably underestimated, and we restrict our discussion here to events of the early Holocene, when temperatures were warmer than present and well within the range of July air temperatures represented by the training set (5.0 to 19.0°C). Thomas et al. (2008) describe a temporally detailed chironomid record for the last two thousand years at this site, including rapid assemblage changes that occurred in the past century.

Discussion

Holocene thermal maximum

Chironomid-based reconstructions indicate that early Holocene summers at Lake CF8 and nearby Lake CF3 (22 km to the southeast; Briner et al., 2006) were nearly 5°C (±1.5°C) warmer than present, presumably in response to a positive anomaly in summer insolation forcing (Berger and Loutre, 1991; Fig. 5A). This magnitude of reconstructed early Holocene warmth far exceeds hemispheric averages, even for high-latitude sites. Kaufman et al. (2004) reviewed varied peak Holocene temperatures (mostly summer temperatures estimated from pollen) ranging from 1 to 3°C higher than present around the Western Hemisphere Arctic. Borehole measurements indicate that peak annually integrated Holocene temperatures on the central and southwestern Greenland Ice Sheet were ~2°C higher than present (Cuffey and Clow, 1997; Dahl-Jensen et al., 1998), suggesting either less summer warming over Greenland relative to Baffin Island, or greatly enhanced seasonality relative to present.

Like all proxy-based reconstructions, chironomid-inferred temperatures come with caveats: for example, if modern calibration datasets do not include appropriate analogues for past chironomid assemblages and past climate (e.g., Williams and Jackson, 2007), transfer functions will yield spurious temperature reconstructions. We used CCA to test whether our calibration dataset contains useful analogues for early Holocene assemblages at Lake CF8. In Figure 6, all of the downcore samples from Lake CF8 are plotted passively on a CCA biplot of surface-sediment samples from the calibration (training set) sites (Walker et al., 1997; Francis et al., 2006). The Holocene samples appear to have good analogues in the calibration dataset: Mid- to late Holocene samples have their closest analogues on modern-day Devon and Baffin islands. Samples from the Holocene Thermal Maximum plot in a cluster that overlaps with warmer calibration sites farther south in Atlantic Canada. The two late-glacial

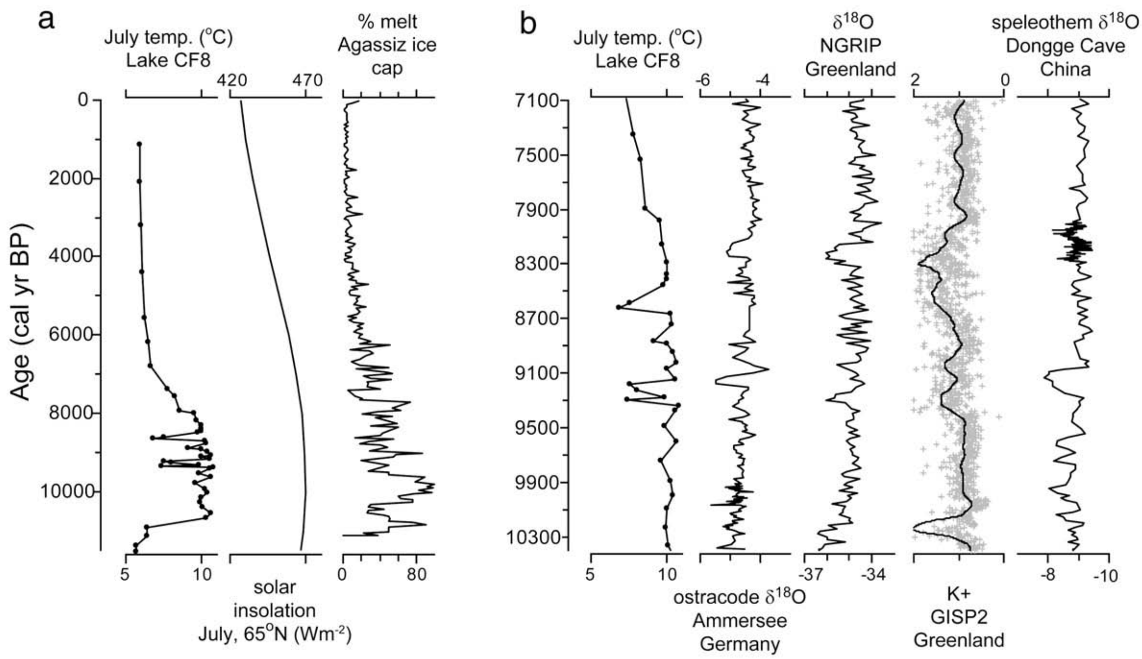

Figure 5. (a) Holocene July air temperatures at Lake CF8 compared with July insolation (Berger and Loutre, 1991) and melt layers in the Agassiz Ice Cap (Fisher and Koerner, 2003). (b) Early Holocene July air temperatures at Lake CF8 compared with Ammersee ostracode $\delta^{18}O$ (‰ PDB; von Grafenstein et al., 1999), NGRIP $\delta^{18}O$ (‰ VSMOW; NGRIP Project Members, 2004; GICC05 chronology from Rasmussen et al., 2006 and Vinther et al., 2006), GISP2 K+ ions (ppb; Mayewski et al., 1997), and Dongge Cave speleothem $\delta^{18}O$ (‰ VPDB; Dykoski et al., 2005). The World Data Center for Paleoclimatology in Boulder, Colorado, provided access to archived data.

samples have no analogues in the modern data set, but their strongly negative Axis 1 scores are consistent with very cold air and water temperatures.

Another potential problem with quantitative temperature reconstructions is that intercorrelations among environmental variables known to influence chironomid species distributions (e.g.,

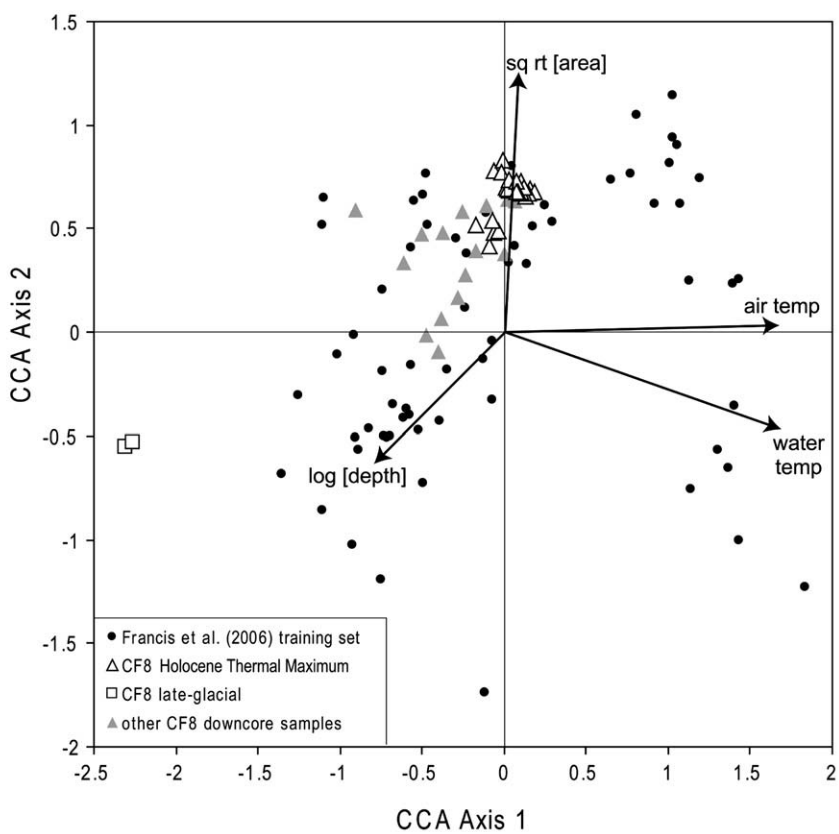

Figure 6. CCA site scores for chironomid assemblages in surface-sediment samples of the training set, shown as black circles (training set sites and samples are described by Walker et al., 1997, and Francis et al., 2006). Downcore samples from Lake CF8 are plotted passively and shown as open squares (late-glacial samples), open triangles (samples recording peak inferred temperatures of the early Holocene), and gray triangles (samples from the mid- to late Holocene and the early Holocene cold reversals). Vectors represent environmental data from the training set, including summer lake-water temperature, July air temperature, lake depth, and lake surface area.

intercorrelations among air and water temperatures, primary productivity, and food availability) may tend to amplify the effects of warming on chironomid assemblages. However, to the extent that such intercorrelations have persisted through time, this effect is implicitly factored into modern calibration data. An additional complication is that at some high-latitude sites, midge assemblage shifts on millennial time scales through the Holocene appear to primarily reflect the effects of post-glacial soil development and vegetation succession, and attendant changes in lake depth, water chemistry, and trophic status, rather than direct responses to temperature change (e.g., Rosen et al., 2003; Velle et al., 2005). In these cases, multi-proxy comparisons reveal divergent trends in chironomid-inferred temperatures versus other temperature-dependent proxies (e.g., pollen).

The timing and direction of chironomid-inferred temperature changes at Lake CF8 are consistent with results from several other studies in the region, such as inferred primary productivity in many northern Baffin Island lakes and chironomid-inferred temperatures at neighboring Lake CF3, peaked between 10 and 8 cal ka BP (Miller et al., 2005; Briner et al., 2006; Micheluti et al., in press). The first millennia of the Holocene also saw peak melt on the Agassiz Ice Cap on Ellesmere Island (Fig. 1 and Fig. 5; Fisher and Koerner, 2003). Knudsen et al. (2008) infer maximum Atlantic-water influence in northernmost Baffin Bay in the earliest part of the Holocene, with subsequent increased influence of sea ice after ~7.3 cal ka BP. However, the inferred timing of maximum Holocene warmth in Baffin Bay and the Arctic channels varies between sites and proxies, with some records suggesting that peak warmth within Baffin Bay occurred millennia after peak warmth at Lake CF8 (Dyke et al., 1996a,b; Levac et al., 2001). It has been proposed that early Holocene warming in parts of the North Atlantic region lagged summer insolation forcing due to the cooling influence of the residual Laurentide ice sheet (Kaufman et al., 2004; Kaplan and Wolfe, 2006), but results from Lake CF8 suggest that the terrestrial environment of northern Baffin Island responded rapidly and dramatically to insolation forcing.

Additional studies using quantitative temperature proxies should clarify whether chironomid assemblages are providing a realistic estimate of the magnitude and timing of early Holocene warmth at Lakes CF3 and CF8. Several factors may indeed have combined to cause exceptional warming over the eastern Canadian Arctic in response to enhanced early Holocene insolation forcing: for example, amplification of warming by changes in the duration of terrestrial snowcover (e.g., Chapin et al., 2005) and possibly sea-ice albedo feedbacks related to waning sea ice cover over Baffin Bay and the channels of the Arctic Archipelago (Dyke et al., 1996b).

Abrupt cold reversals in the early Holocene

Prior Holocene temperature reconstructions on Baffin Island (e.g., at Lake CF3, Fog Lake, and Brother of Fog Lake; Briner et al., 2006; Francis et al., 2006; Fréchette et al., 2006) were pioneering studies done at millennial-scale resolution and did not resolve sub-millennial climate shifts. A new finding from the more resolved Lake CF8 paleotemperature record is the pair of abrupt cold events that occurred during peak early Holocene warmth. Abundances of cold stenotherms were low during the reversals, but their presence at these times was unique to the early Holocene (Fig. 4). *Pseudodiamesa* and *Abiskomyia* are present in the surface sediments of only the coldest lakes in Canada and Scandinavia today (Olander et al., 1999; Francis et al., 2006). Notably, *Paracladopelma* also appears during the reversals. The temperature optimum of this genus in eastern Canada is unknown, but in west Greenland *Paracladopelma* is most abundant in the coldest oligotrophic lakes, and it appears only in the highest-elevation lakes in the European Alps (Lotter et al., 1997, 1999; Brodersen and Anderson, 2002).

Although our age model suggests that the latter cold event at Lake CF8 dates to ~8.5 to 8.6 cal ka BP, the age model has an uncertainty of ± 310 yr for this time period (Fig. 2), and correlation with the conspicuous cold reversal documented elsewhere ~8.2 cal ka BP (e.g., Alley et al., 1997; Alley and Ágústdóttir, 2005) is possible. The advance of a local ice cap and decreased lake productivity ~8 cal ka BP were recently reported from southern Baffin Island (Miller et al., 2005), and the abrupt decline of birch pollen in north-central Canada (Seppä et al., 2003) provides supporting evidence that Arctic Canada cooled ~8.2 cal ka BP. However, the age of the latter reversal at Lake CF8 is not tightly constrained and correlation with the "8.2 ka event" is tentative.

The earlier reversal dates to ~9.2 to 9.3 cal ka BP. We speculate that cooling at Lake CF8 at this time was part of a widespread climate perturbation that affected the North Atlantic region and parts of Asia. Prior paleoclimate studies have documented abrupt climate changes around the North Atlantic region and farther afield between 9.5 and 9.0 cal ka BP (Fig. 5B), including cooling over central Greenland and the DYE-3 site in southern Greenland (e.g., NGRIP, 2004; Vinther et al., 2006), changes in precipitation isotopes over the British Isles and central Europe (von Grafenstein et al., 1999; McDermott et al., 2001), and weakened monsoon precipitation inferred from speleothems in southern China (Dykoski et al., 2005) and Oman (Fleitmann et al., 2007). Fleitmann et al. (2008) review evidence for a climate perturbation that affected much of the Northern Hemisphere at this time, and they hypothesize that it may have been caused by a meltwater pulse into the North Atlantic. Perturbations in some paleoclimate proxies at this time show comparable amplitude to that of the 8.2 ka event, although at many sites the 8.2 ka event is of unique amplitude for the Holocene.

End moraines across Baffin Island record advances of Laurentide outlet glaciers and local mountain glaciers between ~9.5 and 8 cal ka BP (Cockburn substage *sensu lato*; e.g., Andrews and Ives, 1978; Dyke and Hooper, 2001; Miller et al., 2005). The cause of these glacier advances has been enigmatic, but their timing is broadly coincident with the timing of abrupt cold events documented at Lake CF8 and farther afield. If these cold reversals did indeed cause advances of the Laurentide ice sheet, cooling must have occurred over large parts of Canada.

The occurrence of two separate cold events of similar magnitude and duration indicates that the well-known 8.2 ka event, whether or not it affected Lake CF8, was not uniquely significant at this site. An event as catastrophic as the final draining of glacial lakes Agassiz and Ojibway—a postulated cause of the 8.2 ka event (e.g., Barber et al., 1999)—may not be required to trigger widespread perturbations of interglacial climate. Evidence from a growing number of sites for high-amplitude abrupt, widespread climate shifts in the early Holocene illustrates the sensitivity of the climate system to perturbations, even during periods of warmth and enhanced radiative forcing in the Northern Hemisphere.

Conclusions

Because enhanced summer insolation forcing in the early Holocene was long-lived and hemispherically symmetric, this time period provides a useful (albeit imperfect) analog for future greenhouse warming. Extreme early Holocene warmth on Baffin Island, if confirmed by future studies and additional proxies, may foretell strong future warming in the region. Climate feedbacks, along with the spatially heterogeneous nature of early Holocene warmth across the Arctic, present challenges for predicting the pattern of future warming and its impacts. Fortunately, a growing database of quantitative paleoclimate records is elucidating the pattern of early Holocene warmth across the Arctic and improving our understanding of the factors that modulate local and regional responses to radiative forcing.

This study adds to mounting evidence for widespread abrupt climate change ~9.2 cal ka BP, and for climatic variability during the Holocene in general. At some sites the amplitude of the "9.2 ka event" matches or exceeds that of the 8.2 ka event, indicating that major Holocene climate perturbations were not limited to 8.2 cal ka BP. Given lingering uncertainty regarding the probability of "climate surprises" in Earth's warm greenhouse future (NRC, 2002; IPCC, 2007), and about the potential climatic fallout of changing freshwater inputs to the North Atlantic Ocean (e.g., Curry and Mauritzen, 2005; Zickfeld et al., 2007), abrupt climate shifts in the early Holocene remain important targets for research. These events may especially shed light on the role of freshwater forcing on North Atlantic Ocean circulation, as outbursts of freshwater from the Laurentide ice sheet during this time are hypothesized triggers for abrupt early Holocene climate perturbations.

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