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Rapid Communication

Quantitative paleotemperature estimates from $\delta^{18}O$ of chironomid head capsules preserved in arctic lake sediments

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Abstract

A paleoenvironmental perspective of temperature change is paramount to understanding the significance of recent warming in the Arctic. Late Quaternary sediments from many arctic lakes provide environmental archives with decadal resolution, but reconstructions are hampered by the relative insensitivity of many traditional proxies to temperature. Here, we show that the $\delta^{18}O$ of head capsules of chironomid larvae are equilibrated with the $\delta^{18}O$ of lakewaters in which they live. In suitable lakes, lakewater $\delta^{18}O$ is controlled by the $\delta^{18}O$ of local precipitation, which is strongly correlated to mean annual air temperature (MAT). From this correlation, chironomid $\delta^{18}O$ can be used to examine past changes in MAT. We illustrate the potential of this novel approach to paleothermometry with examples from two arctic lakes that reveal strong regional paleoclimatic gradients in the early Holocene.

Introduction

The Arctic is warming, with increases in mean annual temperatures (MAT) of 2–5 °C recorded in most regions since the 1960s (Serreze et al. 2000). The significance of this climate change can be better evaluated by comparison to long-term changes inferred from the geological record (Overpeck et al. 1997). For the terrestrial realm, only lake sediments have the spatial density and temporal resolution necessary to capture the full range of arctic climate variability (PARCS 1999). A primary factor limiting the use of arctic lake sediments has been the dearth of quantitative paleotemperature proxies. In particular, the high-latitude terrestrial pollen record lacks sensitivity to temperature because of sparse local pollen production and overprinting by wind-blown exotic taxa (Gajewski et al. 1995).

In order to better quantify paleoclimate on late Quaternary timescales, a number of studies have focused on materials that may reliably record lakewater $\delta^{18}O$. This is because the $\delta^{18}O$ of precipitation is highly correlated to MAT at high latitudes

(Dansgaard 1964; Rozanski and Araguás-Araguás 1993). For suitable lakes with short residence times (i.e., catchment : lake area $\leq 20:1$), lakewaters are completely replaced every spring during snowmelt, minimizing evaporative effects on δ^{18}O. Lakewater δ^{18}O has been inferred from the δ^{18}O of authigenic calcite (Anderson et al. 2001) and aquatic cellulose (e.g., Wolfe et al. 2001) in high-latitude lake sediments. However, some lakes in the Arctic (and in other regions) are acidic and calcite is not present. Furthermore, cellulose of unequivocal aquatic origin is rarely continuously preserved (Sauer et al. 2001).

Advances in mass spectrometry (Koziet 1997; Kornexl et al. 1999) have allowed us to investigate an alternative medium to reconstruct lakewater δ^{18}O: the remains of chironomid (Diptera: Chironomidae) larvae that are ubiquitous in the sediments of mid- and high-latitude lakes. Aquatic animals acquire oxygen for biosynthesis primarily from the water in which they live, and to a lesser extent from diet (Schimmelmann et al. 1987). Stable isotopic signatures of chitin from aquatic invertebrates therefore reflect the environment in which they live (Schimmelmann and Deniro 1986). Chironomids are holometabolous insects with larval stages that are obligately aquatic. Most chironomid larvae are benthic and feed on combinations of detritus, bacteria, algae, and other invertebrates. Their head capsules, molted by the third and fourth larval instars, preserve well under a wide range of limnological conditions (Walker 2001) and can be used for AMS ^{14}C dating (Fallu et al. 2004).

Materials and methods

For calibration purposes, chironomid head capsules were manually isolated from the modern sediments of four lakes (Figure 1) spanning a wide climate and latitudinal gradient: Green Pond, MA, USA (GP: 42°59′N, 72°51′W); Levi Pond, Vermont, USA (LP: 44°27′N, 72°23′W); Qipisarqo Lake, southwestern Greenland (QL: 61°01′N, 47°75′W); and Fog Lake, Baffin Island, Canada (FL: 67°11′N, 63°15′W). For subsequent paleoclimatic assessments, additional material was extracted from well-dated cores from the two arctic sites: QL (Kaplan et al. 2002) and FL (Wolfe et al. 2000). In every case, 1 cm slices of sediment were

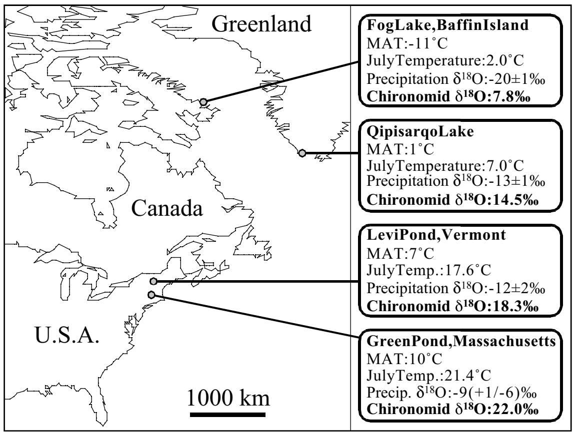

Figure 1. Location of study sites and modern environmental data for each of the calibration lakes.

processed, corresponding to between 20 and 100 years of deposition, thus minimizing interannual and within-population variability. Approximately 100 μg of chironomid head capsules (representing 300–700 head capsules) from each sample was required for the measurement of δ^{18}O.

Measurements of chironomid head capsule δ^{18}O were made using an on-line pyrolysis thermochemical reactor (Finnigan ThermoQuest TC-EA) coupled to a continuous flow isotope ratio mass spectrometer (Finnigan MAT Delta Plus XL). Duplicate analyses of separate chironomid picks from GP surface sediments, FL core 59–60 cm, and QL core 137–138 cm yielded 2σ uncertainties of ± 0.12‰, ± 0.14‰, and ± 0.17‰, respectively. Thus, we conservatively estimate our down-core chironomid δ^{18}O time-series to have 2σ uncertainties of ± 0.2‰. Diagenetic alteration of fossil chitin appears minimal since O_2 yields from modern and fossil samples are directly comparable. Furthermore, microstructures of fossil head capsules are in all cases sufficiently preserved to allow taxonomic differentiation. The integrity of chitin δ^{18}O was also evaluated by soaking standard chitin (Sigma C-9752; $\delta^{18}O = 24.98‰ \pm 0.30$, $n = 3$) in ^{18}O-enriched water (+48‰) at neutral pH for 3 days at 25 °C, after which no significant change in chitin δ^{18}O could be detected (25.43‰ ± 0.93, $n = 3$). Additional tests indicate that no change in δ^{18}O is induced by processing with either HCl, KOH, or deionized water ($p = 0.5$, $n = 3$).

Standards of known $\delta^{18}O$ (NBS N-1, NBS-18, NBS-19, and an internal calcite standard) were analyzed with each run of chironomid head capsules (measured *versus* expected $r^2 \geq 0.98$). All resulting $\delta^{18}O$ values are expressed relative to Standard Mean Ocean Water.

The $\delta^{18}O$ measured from modern head capsules were plotted against $\delta^{18}O$ of precipitation and projected MAT for each site (Figure 2). The $\delta^{18}O$ of precipitation was interpolated from the International Atomic Energy Agency's Global Network of Isotopes in Precipitation database (IAEA-GNIP 1995). We have one-time measurements of $\delta^{18}O$ from FL and GP (Autumn). FL $\delta^{18}O$ (−19.7‰) is close to the predicted value (−20‰; IAEA-GNIP 1995) and, coupled with δ^2H (−143‰), yields a deuterium excess of 10.4, which is indistinguishable from the meteoric water line and confirms that evaporative influences are minimal. GP was sampled during drought (Summer) and suffers from isotopic enrichment ($\delta^{18}O = -0.8$‰, $\delta^2H = -21$‰, deuterium excess = −14.5). IAEA-GNIP (1995) data predict a $\delta^{18}O$ of −9‰ for GP, which we have used in Figure 2a, although with a larger uncertainty (+6‰, −1‰) than the other calibration lakes. Because our surface samples span at least 10 years, the impacts of episodic drought should not significantly influence whole-assemblage chironomid $\delta^{18}O$. An uncertainty of ± 2‰ is used for LP, which is not close to a GNIP reporting station, and ± 1‰ is the estimated uncertainty for the other two lakes (QL is close to a GNIP station, whereas FL has measured $\delta^{18}O$).

For down-core analyses at QL, chironomid head capsules were picked from an additional 20 samples spanning the Holocene. The core's chronology is based on 11 Accelerator Mass Spectrometer (AMS) ^{14}C dates that have been calibrated to calendar years (Stuiver et al. 1998). The lake is small (1 km diameter), 9 m deep, and situated near the coast at 7 m asl. Additional site details are given by Kaplan et al. (2002). At FL, four additional samples were analyzed from early and late Holocene sediments, as well as two samples from a lower organic unit dated to late oxygen isotope stage 5 (OIS 5; ~100 ka BP). The core's chronology is constrained by 18 AMS ^{14}C dates as well as luminescence age estimates (thermal and infrared stimulation) for sediments beyond the range of ^{14}C (Wolfe et al. 2000). This lake is 170 m long, 140 m wide, 9.5 m deep, and is situated 460 m asl on a highly weathered upland (Steig et al. 1998). During Wisconsinan glaciation, the lake was apparently deeply frozen, but not glacially eroded, resulting in a stratigraphy of OIS 5 organic sediments directly overlain by Holocene facies, with little or no intervening sedimentation.

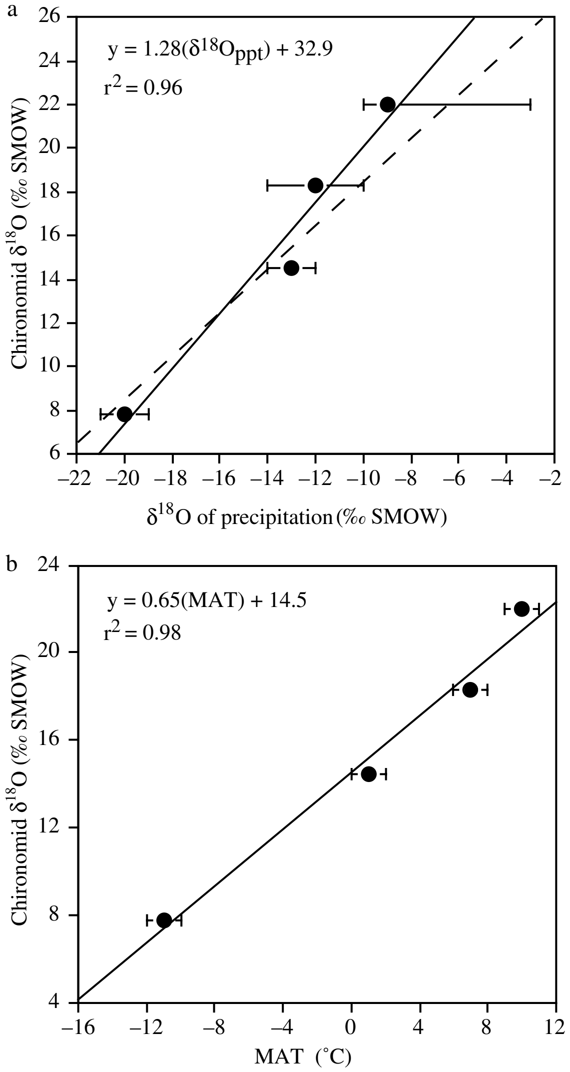

Figure 2. Calibration of modern chironomid $\delta^{18}O$ to precipitation $\delta^{18}O$ (a) where the constant fractionation line (dashed line) displays slightly lower slope than the regression line (solid black) from our data. Assuming constant fractionation between chironomid head capsules and water, we obtain a fractionation factor $\{\alpha = [\delta^{18}O$ chironomid head capsule + $1000/\delta^{18}O\ (H_2O) + 1000]\}$ of 1.028. Calibration of modern chironomid $\delta^{18}O$ to MAT (b).

For both QL and FL, inferred MAT from chironomid head capsule $\delta^{18}O$ was independently assessed by summer water temperature (SWT) reconstructions using the 39 lake chironomid faunal assemblage transfer function of Walker et al. (1997), augmented by 29 additional sites from Baffin Island (D. Francis, unpublished data). This model uses weighted-averaging regression and calibration with inverse deshrinking and tolerance down-weighting (e.g., Lotter et al. 1999). In the expanded model, measured *versus* inferred SWTs have $r^2 = 0.88$ ($n = 68$), and the root-mean-squared error of prediction is ± 2.2 °C.

Results

Our modern calibration data (Figure 2) span over 20 °C of MAT and 11‰ of precipitation $\delta^{18}O$. Chironomid $\delta^{18}O$ is highly correlated with both the expected $\delta^{18}O$ of precipitation ($r^2 = 0.96$) and MAT ($r^2 = 0.98$). We note that a line of slope 1.0, required to produce a unique solution for chironomid biological fractionation factor (BFF) over the encompassed climate gradient, can be fitted within uncertainties of lakewater $\delta^{18}O$ (dashed line on Figure 2a). This line gives a BFF of 1.028, assuming constant fractionation between chironomid head capsules and water, illustrating that the head capsules are 28.5‰ more positive than the precipitation at each site, which is indistinguishable from the BFF between aquatic moss cellulose and lakewater $\delta^{18}O$ (28.3‰; Sauer et al. 2001). The slope of the regression against current MAT (0.65; Figure 2b) is also similar to that observed between precipitation $\delta^{18}O$ and temperature (0.69; Dansgaard 1964). These data demonstrate that modern chironomid $\delta^{18}O$ faithfully tracks the $\delta^{18}O$ of precipitation in watersheds of small lakes. Given the analytical uncertainty associated with chironomid head capsule $\delta^{18}O$ (± 0.2‰), the new proxy can be used to estimate MAT with an average uncertainty of ± 1.0 °C. The true uncertainty of the paleotemperature estimates likely incorporates additional error terms which, when coupled with the analytical uncertainty, produce a composite error in the order of ± 2 °C.

Chironomid $\delta^{18}O$ decreases through the Holocene at QL (Figure 3a). The observed 2.5‰ shift in $\delta^{18}O$ corresponds to a 3 °C MAT decline. The isotope-based MAT reconstruction is supported by SWT estimates from changes in chironomid faunal assemblages, which indicate a progressive 2 °C decrease in Holocene SWT. At FL, chironomid $\delta^{18}O$ values are comparable between late OIS 5 and mid-Holocene samples, but decline sharply during the early Holocene, corresponding to a MAT decrease of 5 °C (Figure 3b). Chironomid transfer function SWT estimates suggest that OIS 5 lake waters were 2 °C warmer than the mid-Holocene, and 5 °C warmer than the early Holocene. Thus, for both lakes the directions of change in MAT from chironomid $\delta^{18}O$ are well supported by faunal assemblage transfer function results.

Discussion

The results from QL indicate that peak Holocene water and air temperatures were reached before 7 ka BP. Coastal West Greenland is strongly influenced by the West Greenland Current (WGC), a branch of North Atlantic Drift that carries warm, salty water along the West Greenland coast and maintains a strong E–W thermal gradient across the Labrador Sea. The WGC was established by 10 ka BP, and was strongest between about 9 and 5 ka BP (Dyke et al. 1996). Paleoclimate reconstructions from other sites in the northern North Atlantic sector also reflect maximum summer temperatures in the earliest Holocene, driven in part by peak summer insolation (Funder and Weidick 1991; Bennike et al. 2001). These observations are entirely consistent with our chironomid-based reconstructions of both MAT and SWT. Mid-Holocene sediments from QL indicate substantial variability of chironomid $\delta^{18}O$ values between 3.5 and 6 ka BP (Figure 3a), a feature also present in other North Atlantic records (e.g., Bradley (2000)). Whether these inferred mid-Holocene temperature oscillations at decadal to centennial scales reflect noise in the proxies or real events deserves future scrutiny through analyses at higher stratigraphic resolution.

At FL, chironomid-based proxies indicate that OIS 5 was as warm (MAT) or warmer (SWT) than any time in the Holocene (Figure 3b). The abundance of shrub pollen (alder and birch) in these sediments supports the hypothesis of

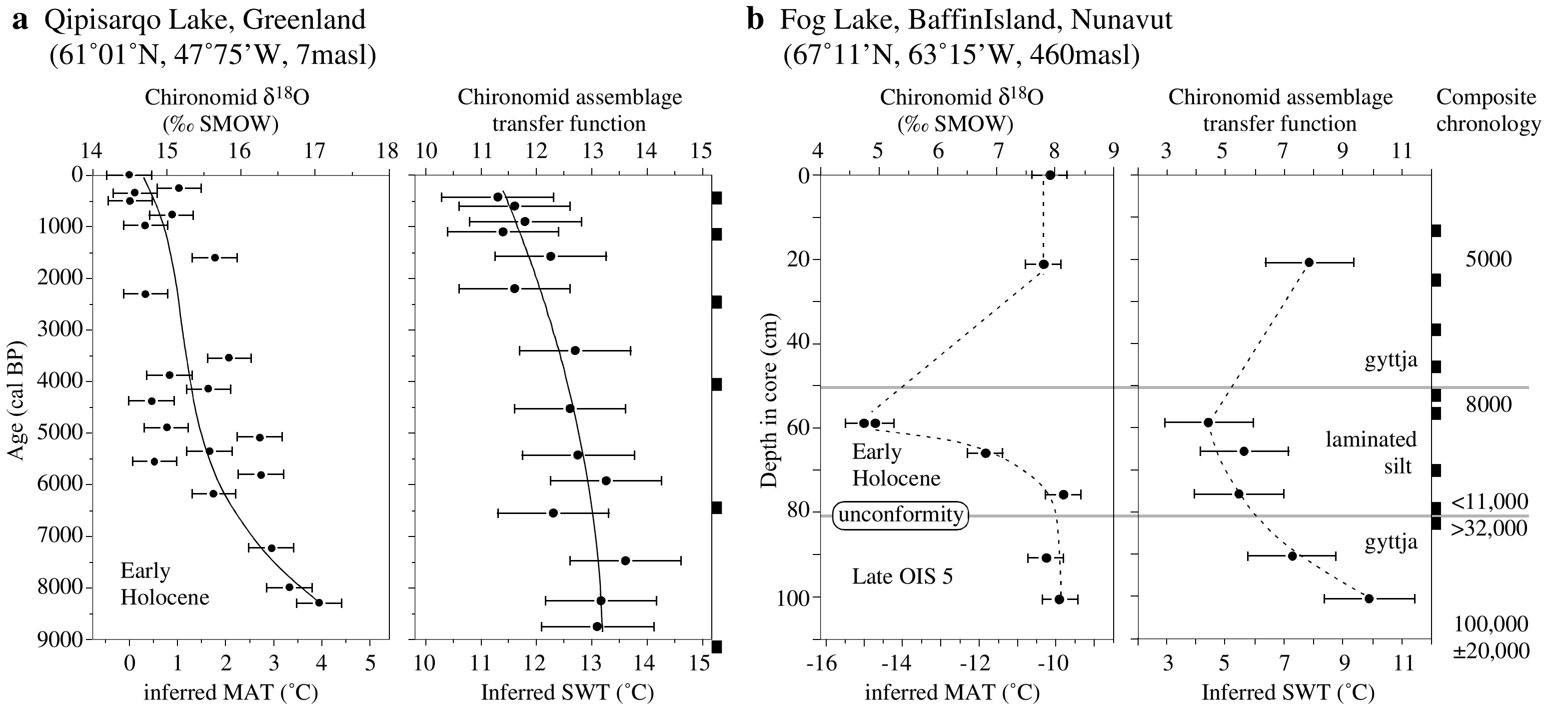

Figure 3. Down-core reconstructions of MAT from chironomid $\delta^{18}O$ and SWT from the chironomid faunal assemblage transfer function for Qipisarqo Lake (a) and Fog Lake (b). Down-core trends in (a) are fitted polynomial interpolations between the data points. Black squares indicate the positions of radiocarbon AMS dates.

greater summer warmth (Wolfe et al. 2000), in agreement with peak insolation values of the last 100 ka (Berger and Loutre 1991). However, reworking of OIS 5 pollen is evident in the early Holocene sediments of FL, until now precluding any realistic constraints on early Holocene paleotemperatures. Both chironomid proxies illustrate that marked cooling occurred during this interval.

When viewed comparatively for the early Holocene (~8–9 ka BP), the chironomid data from QL and FL show trends in opposite directions on either side of Baffin Bay. In contrast, similarities observed between temperature reconstructions from the various Greenland ice cores imply that they record a regionally integrated high-latitude climate signal (Cuffey 2000). Pronounced differences in the early Holocene thermal histories of our two sites, situated only 400 km apart, demonstrate that the lake record has greater sensitivity to local climate history, thus fine-tuning our understanding of spatial heterogeneity in paleoclimatic trends. For instance, QL is strongly influenced by the warm WGC, whereas FL is influenced by the Baffin Current, which was modulated in the early Holocene by glacial melt originating from the Canadian Archipelago and throughput from transpolar drift between Ellesmere Island and northwestern Greenland. These cold ocean currents were especially active during the earliest Holocene due to influence of delayed deglaciation of the Innuitian (England et al. 2000) and northeast Laurentide (Dyke et al. 1996) ice sheets.

Detailed comparisons of our isotopic and faunal proxies may also provide additional paleoclimatic insights. Where the two reconstructions diverge (e.g., 65–80 cm in FL), it may eventually be possible to decipher large-scale changes in ocean–atmosphere circulation that account for source variability of precipitation ^{18}O, a similar approach to that adopted by the ice core community (Shuman et al. 1995; Cuffey 2000). Evaporative losses are generally negligible from lakes at high latitudes (Gibson and Edwards 2002) and are therefore less likely to have significantly influenced the $\delta^{18}O$ of lake waters at the two high-latitude sites we used to reconstruct paleo-temperatures. Finally, we concede that several details of the biochemical processes associated with the offset between chironomid $\delta^{18}O$ and the $\delta^{18}O$ of waters they inhabit are incompletely understood and will improve with the development of the method. Detailed studies of marine invertebrate chitin show that the $\delta^{18}O$ of chitin is invariably enriched relative to water, with a magnitude similar to the fractionation we have shown in our modern samples (Figure 2a). Controlled *in vitro* experiments capitalizing on increased instrument sensitivity towards $\delta^{18}O$ in organic matrices will allow more rigorous estimates of BFF to be established, as well as potential species-specific vital effects.

Conclusion

The $\delta^{18}O$ of chironomid head capsules provides a new quantitative paleoclimate proxy where few had been available previously. Our results demonstrate that chironomid $\delta^{18}O$ is equilibrated with the $\delta^{18}O$ of lakewaters in which larvae live. By selecting appropriate lakes with short residence times, down-core changes in chironomid $\delta^{18}O$ provide a proxy for the evolution of precipitation $\delta^{18}O$ and MAT. In sediments from lakes on Greenland and Baffin Island, chironomid $\delta^{18}O$-based MAT reconstructions compare favorably with those of SWT based on faunal assemblages. These techniques appear equally viable at millennial (QL) and glacial/interglacial (FL) timescales. Chironomid $\delta^{18}O$ measurements provide the first quantitative MAT reconstructions from arctic lake sediments and circumvent problems inherent to the pollen record, including delayed plant immigration to island settings (e.g., Greenland, Iceland, and the Canadian Arctic Archipelago). Chironomids colonize aquatic habitats rapidly, thus providing a robust paleoclimate proxy that is largely independent of catchment edaphic factors. Because 300–700 head capsules are presently required for each measurement, the largest obstacle to enhancing stratigraphic resolution is the effort required to manually isolate sufficient material for $\delta^{18}O$ analysis. However, we envisage that advances in instrument sensitivity will reduce mass requirements to less than 100 specimens in the future. The ubiquity of chironomid head capsules in lake sediments worldwide implies great potential for their use as a medium for reconstructing lakewater $\delta^{18}O$ and, in many regions, direct applications to quantitative paleothermometry.

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