Holocene climate trend, variability, and shift documented by lacustrine stable isotope record in the northeastern United States

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ABSTRACT

Earlier studies indicated that the general pattern of the Holocene climate in the northeastern United States changed from cool and dry (11.6–8.2 ka; 1 ka = 1000 cal yr BP) to warm and wet (8.2–5.4 ka) to warm and dry (5.4–3 ka) to cool and wet (after 3 ka). A new ~35-year resolution stable isotope record of endogenic carbonate from a sediment core for Lake Grinnell in northern New Jersey provides a chance to examine the Holocene climate variations of the region in a finer detail. After the Younger Dryas cold climate reversal, the δ18O fluctuated around a constant value of ~7.4‰ until 5.8 ka, thereafter shifted to a steadily decreasing trend to the most recent value of ~2.8‰. Responding to this shift, the widely observed hemlock decline in the northeastern USA occurred about ~350–500 (±143.5) years later. Detrended δ18O and δ13C records show a clear covariance at 910-year periodicity. The amplitudes of centennial-scale δ18O variations became much smaller after 4.7 ka. At the same time, the dominant frequency of these variations changed from 330 to 500 years. We suggest that a non-linear response of atmospheric circulation to the gradual decrease in insolation is responsible for the shift in the climate trend at 5.8 ka as indicated by the deceasing δ18O values. A dominant frequency shift in solar forcing and the decreased seasonal contrast of insolation might have caused the change in climate variability at 4.7 ka through modulating ocean and atmosphere circulations.

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1. Introduction

Stable isotopes from lacustrine carbonate can be a useful tool for reconstructing long-term climate changes and investigating possible causes and mechanisms of these climate changes. In the northeastern United States, many efforts have been made to reconstruct the Holocene climate history from lake sediments, including lithologically derived lake-level reconstructions (e.g. Dwyer et al., 1996; Mullins, 1998; Newby et al., 2000; Shuman et al., 2004; Dieffenbacher-Krall and Nurse, 2005), pollen-inferred vegetation changes (e.g. Davis, 1969; Watts, 1979; Davis et al., 1980; Suter, 1985; Maenza-Gmelch, 1997; Shuman et al., 2001), as well as stable isotope records (e.g. Anderson et al., 1997; Huang et al., 2002; Kirby et al., 2002; Ellis et al., 2004; Hou et al., 2006, 2007). This Holocene climate history includes a cool and dry period at 11.6–8.2 ka (1 ka = 1000 cal year BP), a warm and wet climate at 8.2–5.4 ka, a warm and dry interval at 5.4–3 ka, and a cool and wet condition during the past 3 ka (Shuman et al., 2004). Superimposed on this general trend, a series of millennial-, centennial- and multidecadal-scale climate oscillations have been reported (e.g. Kirby et al., 2002; Cronin et al., 2005; Viau et al., 2006; Li et al., 2007; Springer et al., 2008). In addition, the mid-Holocene was a period of large change. For example, many lakes in eastern North America experienced a low lake-level from 5 to 3 ka (e.g. Dwyer et al., 1996; Yu et al., 1997; Newby et al., 2000; Lavoie and Richard, 2000; Almquist et al., 2001; Mullins and Halfman, 2001; Shuman et al., 2001, 2004; Dieffenbacher-Krall and Nurse, 2005), which has been attributed to broad-scale changes in atmospheric circulation (e.g. Yu et al., 1997; Kirby et al., 2002). In response to the dry climate, an abrupt decline of the hemlock (Tsuga canadensis) occurred around 5.5 ka, which has been documented from pollen data across eastern North America (e.g. Allison et al., 1986; Foster and Zebryk, 1993; Fuller, 1998; Bennett and Fuller, 2002; Foster et al., 2006) and the decline has recently been suggested to be caused by increased aridity (e.g. Yu et al., 1997; Shuman et al., 2004; Foster et al., 2006). However, there are only a few stable isotope records with temporal resolution sufficient for reconstructing the multidecadal- to centennial-scale Holocene climate history (e.g. Kirby et al., 2002) and examining the possible mechanisms of climate changes.

In this paper, we present an ~35-year-resolution calcite stable isotope record of a marl sediment core collected from a small,
hydrologically open hardwater lake in northern New Jersey, USA. The objectives of this study are to reconstruct a detailed climate history covering the last ~12,000 years with a special focus on the nature of the proposed mid-Holocene climate shift and to identify and understand the possible forcing factors and how the climate system responded to those forcings.

2. Study region and site

Lake Grinnell (41°06'N, 74°38'W, 170 m above sea level) is located in Sussex County in northern New Jersey (Fig. 1A). The lake sits in a limestone terrain and is situated in the Kittatinny Valley of the Ridge and Valley physiographic province, which is underlain by northeast-trending folded and faulted blocks of dolostone, limestone, slate and siltstone (Witte and Monteverde, 2006). The Laurentide Ice Sheet retreated from this area around 14,000 years ago (Witte, 2001). The regional climate has been strongly affected by the moisture from the Gulf of Mexico and tropical Atlantic Ocean (Lawrence et al., 1982; Peixoto and Oort, 1983). Based on the instrumental data (1895–2006) from New Jersey Climate Division 1 (northern New Jersey) (http://www1.ncdc.noaa.gov/pub/data/cirs), the regional mean annual temperature is 10.4 °C with an average summer (June–August) temperature of 21.5 °C and an average winter (December–February) temperature of −1.1 °C (Fig. 2A). The mean annual precipitation is 1177 mm with almost uniform monthly distribution (Fig. 2A). The average of monthly weighted mean precipitation $\delta^{18}O$ values (collected from 5 stations in the Global Network of Isotopes in Precipitation (GNIP), including Ste. Agathe, Quebec; Ottawa, Ontario; Simcoe, Ontario; Coshocton, Ohio; and Chicago, Illinois (http://isohis.iaea.org/userupdate/GNIPMonthly.xls)) in eastern North America (see locations in Fig. 1A) shows a high correlation with the mean monthly temperature in northern New Jersey (Fig. 2A).

Lake Grinnell is a small hardwater lake of ~0.2 km$^2$ in surface area (Fig. 1B) and ~7 km$^2$ in catchment area. It has two basins separated by a ridge where water is <3 m deep. The lake has a short mean residence time of ~80 days and is mostly recharged by shallow groundwater and perhaps precipitation on the catchment basin (Laura Nicholson, personal communication). The water depth in the basins was raised about 2.4 m artificially by a dam to form the present impoundment area, with a maximum water depth of about 10.4 m. The date of the first dam construction is not well known except that it occurred before 1913 (Valgenti et al., 1950). Lake Grinnell is the headwater lake of a branch of Wallkill River, which...
monitoring around White Lake from fall 1992 through spring 1994 showed that water table responded quickly to precipitation events, suggesting local recharge of shallow aquifer located in the limestone (Nicholson, 1995). Given the high permeability of limestone aquifers and hardwater composition of the lake, the hydrologic budget of the lake likely includes significant groundwater component as well as surface inflow.

3. Methods

Water samples at 50 cm water depth from different locations in Lake Grinnell (Fig. 1B) were collected mostly from 2006 to 2008 (Table 1). Several domestic wells around the lake were sampled in 2006 and 2008 by first pumping out about 1.5 L of water before collecting in 500 ml bottles. One shallow well (Travis' well) was sampled in both years. Samples for hydrogen and oxygen isotopes were sealed in 20 ml high density polyethylene plastic bottles with scintillation caps right after collection. Samples for dissolved inorganic carbon (DIC) were filtered through a 0.45-micron filter paper and 'poisoned' by adding two drops of CuSO4 solution to kill organisms and sealed in Amber DIC glass bottles. All water and DIC samples were sent to the University of Arizona Isotope Lab for δDwater and δ18Owater and δ13CDIC analyses. Isotope values are reported in conventional δ notation relative to VSMOW (Vienna Standard Mean Ocean Water) for water δD and δ18O, and VPDB (Vienna Pee Dee Belemnite) for δ13C. The analytical precisions are 0.09‰ for δ18O, 0.9‰ for δD, and 0.3‰ for δ13C.

The 900-cm long sediment core (GL05-1; Fig. 1C) was taken on 30 January 2005 from near the center of the lake (see Fig. 1B for coring location) at 660 cm water depth with a Livingstone-Wright piston corer of 5 cm inside diameter. Sub-samples were taken directly from the core at every 2 cm, and then homogenized for loss-on-ignition (LOI), stable isotope and pollen analyses. LOI analysis was carried out to estimate organic matter content based on weight loss from combustion at 550 °C about 1 h and carbonate content at 1000 °C about 1 h (Dean, 1974). Terrestrial plant macrofossils and charcoal were picked from seven intervals and dated using an accelerator mass spectrometry (AMS) at the University of California, Irvine (Table 2). All the 14C dates were calibrated to calendar ages using CALIB 5.01 program based on the IntCal04 dataset (Reimer et al., 2004). The age model is derived from a fourth-polynomial curve with a fixed surface age (-55 cal year BP = 2005 AD; Fig. 3A).

Samples for calcite stable isotope analyses were air-dried at room temperature. Plant macrofossils, mollusk and ostracode shells, and other macroscopic fragments were removed and discarded under a stereo-microscope. Each of the 334 calcite samples was air-dried at room temperature. Plant macrofossils, mollusk and ostracode shells, and other macroscopic fragments were removed and discarded under a stereo-microscope. Each of the 334 calcite samples

Table 1

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Collection date (month/year)</th>
<th>δD (‰, VSMOW)</th>
<th>δ18O (‰, VSMOW)</th>
<th>δ13C of DIC (‰, VPDB)</th>
<th>Sampling location</th>
</tr>
</thead>
<tbody>
<tr>
<td>020623C</td>
<td>06/2002</td>
<td>-31</td>
<td>-5.2</td>
<td></td>
<td>Boat deck</td>
</tr>
<tr>
<td>060211D</td>
<td>02/2006</td>
<td>-51</td>
<td>-7.9</td>
<td></td>
<td>Boat deck</td>
</tr>
<tr>
<td>060618a</td>
<td>06/2006</td>
<td>-48</td>
<td>-6.8</td>
<td></td>
<td>Boat deck</td>
</tr>
<tr>
<td>060618b</td>
<td>06/2006</td>
<td>-44</td>
<td>-6.9</td>
<td></td>
<td>Boat deck</td>
</tr>
<tr>
<td>060101a</td>
<td>10/2006</td>
<td>-43</td>
<td>-7.3</td>
<td></td>
<td>Lake outlet</td>
</tr>
<tr>
<td>12302006</td>
<td>12/2006</td>
<td>-50</td>
<td>-7.7</td>
<td></td>
<td>Lake outlet</td>
</tr>
<tr>
<td>080517a</td>
<td>05/2008</td>
<td>-45</td>
<td>-7.0</td>
<td>-8.5</td>
<td>Lake outlet</td>
</tr>
<tr>
<td>080517b</td>
<td>05/2008</td>
<td>-47</td>
<td>-7.0</td>
<td>-8.5</td>
<td>Lake outlet</td>
</tr>
<tr>
<td>080908c</td>
<td>09/2008</td>
<td>-43</td>
<td>-7.3</td>
<td>-7.1</td>
<td>Boat deck</td>
</tr>
<tr>
<td>080908d</td>
<td>09/2008</td>
<td>-41</td>
<td>-5.9</td>
<td>-7.1</td>
<td>Lake outlet</td>
</tr>
<tr>
<td>060618c</td>
<td>06/2006</td>
<td>-53</td>
<td>-8.1</td>
<td>-8.1</td>
<td>Travis' well (1.5 m)</td>
</tr>
<tr>
<td>060618d</td>
<td>06/2006</td>
<td>-51</td>
<td>-8.2</td>
<td>-8.2</td>
<td>Naby's well (1.5 m)</td>
</tr>
<tr>
<td>080517c</td>
<td>05/2008</td>
<td>-53</td>
<td>-8.2</td>
<td>-8.2</td>
<td>Sincaglia's well (30 m)</td>
</tr>
<tr>
<td>080508a</td>
<td>09/2008</td>
<td>-53</td>
<td>-12.0</td>
<td>-12.0</td>
<td>Travis' well (1.5 m)</td>
</tr>
</tbody>
</table>

a The analytical precisions are 0.09‰ for δ18O, 0.9‰ for δD, and 0.3‰ for δ13C as reported from the University of Arizona Isotope Lab.

b See Fig. 1B for sampling location.
was analyzed for oxygen and carbon isotopes in the Stable Isotope Laboratory at the University of Minnesota using a Finnigan MAT 252 isotope ratio mass spectrometer coupled to a Kiel II carbonate preparation device. All results are reported in standard δ notation relative to VPDB. The analytical precision is ±0.06‰ for both δ18O and δ13C. To evaluate the regional vegetation response to climate changes, pollen analysis was done on 0.7 cm² sub-samples at every 8-cm interval (Zhao et al., in press). The samples were prepared using a modified standard acetolysis procedure (see details in Zhao et al., in press).

The computer program RAMPFIT (Mudelsee, 2000) was used to analyze raw δ18O data to objectively detect the shift in the long-term trend. In order to visualize and evaluate the magnitude of centennial-scale variability of the δ18O time series, we obtained the deviations of raw δ18O values from the RAMPFIT trend, filtered the raw δ18O time series after applying a 200–500 year band-pass filter using the program AnalySeries (version 2.0.4), and calculated the rates of changes (expressed as 100 × $\Delta$δ18O/$\Delta$t) between each adjacent pair of raw measurements. The raw δ18O and δ13C data were fitted and detrended with a 3rd order-polynomial curve using the program AutoSignal (version 1.0) to remove the long-term trend. In order to evaluate the potential solar forcing on the centennial-scale climate variability and to highlight these centennial-scale variations recorded in δ18O time series, we also performed spectral analysis using the program REDFIT (Schulz and Mudelsee, 2002) using a Hanning window and band-pass filters in the detrended δ18O and raw residual 14C time series (Reimer et al., 2004). The program AnalySeries was also used to perform cross-spectral analysis of the detrended δ18O and δ13C time series in order to evaluate their coherence, and to apply a band-pass filter (at 910 ± 80 year periodicity) to show their corresponding changes at millennial-scale.

### Table 2

<table>
<thead>
<tr>
<th>Depth (cm, below lake surface)</th>
<th>AMS lab IDa</th>
<th>14C date ± error (yr BP)</th>
<th>Median age (cal BP ± 2σ range)b</th>
<th>Materials dated</th>
</tr>
</thead>
<tbody>
<tr>
<td>720–724</td>
<td>194B2</td>
<td>1330 ± 30</td>
<td>1265.5 ± 36.5</td>
<td>2 Larix needles, 1 Cyperaceae seed, plant macrofossils and charcoals</td>
</tr>
<tr>
<td>762–764</td>
<td>16918</td>
<td>1430 ± 15</td>
<td>1324 ± 24</td>
<td>7 woody scales from Abies or Picea, plant macrofossils</td>
</tr>
<tr>
<td>898–902</td>
<td>19483</td>
<td>3495 ± 15</td>
<td>3769 ± 63</td>
<td>plant macrofossils and charcoal</td>
</tr>
<tr>
<td>1022–1024</td>
<td>16915</td>
<td>5340 ± 20</td>
<td>6082.5 ± 80.5</td>
<td>1 Scirpus seed, 1 woody scale, plant macrofossils and charcoal</td>
</tr>
<tr>
<td>1150–1154</td>
<td>16914</td>
<td>7625 ± 30</td>
<td>8417 ± 41</td>
<td>1 woody scale, plant macrofossils and charcoal</td>
</tr>
<tr>
<td>1242–1244</td>
<td>16919</td>
<td>9475 ± 20</td>
<td>10715.5 ± 59.5</td>
<td>1 Pinus seed wing, 4 Pinus needles, 5 pieces Pinus seed fragments and plant macrofossils</td>
</tr>
<tr>
<td>1278–1282</td>
<td>16921</td>
<td>9870 ± 45</td>
<td>11283 ± 80</td>
<td>8 Pinus needles, plant macrofossils</td>
</tr>
</tbody>
</table>

*a* Sample lab ID for Keck-Carbon Cycle AMS facility at the University of California, Irvine.

*b* Calibrated ages for radiocarbon dates were obtained based on the IntCal04 dataset using CALIB 5.01 (Reimer et al., 2004).

![Graphs and figures](image)

Fig. 3. Results from core GL05-1 at Lake Grinnell, New Jersey. (A) Age model based on seven calibrated 14C dates using a fourth-polynomial curve. (B) Organic matter content (%), dotted curve and carbonate content (%), solid curve from LOI analysis. (C) δ18O values from calcite. (D) δ13C values from calcite.

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4. Results

The water isotope values of the lake and surrounding wells are within the range of amount-weighted isotope values in annual precipitation from the five GNIP stations in eastern North America (Fig. 1A) (http://isohis.iaea.org/userupdate/GNIPMonthly.xls). Also, these water samples scatter around the global meteoric water line (GMWL, Fig. 2B). The lake water DIC δ13C values are higher (average −7.6‰) than the groundwater δ13C value of −12.5‰ (Table 1).

The 900-cm long core has about 230 cm clay sediment at the bottom (1560–1330 cm below the lake surface). The rest of the core is mostly marl (1330–670 cm), except for 10 cm of organic-rich gyttja at the very top of the core (670–660 cm; Fig. 1C). This study focuses only on the marl section. The marl contains >90% carbonate and ~6% organic matter (Fig. 3B). The seven calibrated 14C dates from the core GL05-1 are all in order. The chronology indicates the marl section covers the last 12500 calibrated years (Fig. 3A). The accumulation rate of the marl is almost constant throughout the core (~0.54 mm/year based on the age model.

The δ18O values range from −8.9‰ to −6.7‰, with centennial-scale fluctuations (Fig. 3C). A major negative excursion at 12.5–11.3 ka reached a minimum of −8.9‰. After 11.3 ka, the δ18O values fluctuate around a constant value of −7.4±0.4‰ until 5.8 ka, thereafter steadily decreasing to most recent value of −8.2‰ (Fig. 4A), which suggests an equilibrium precipitation of calcite from present-day lake water. The high δ18O values of about −6.7 to −6.8‰ occur at 11.1, 10.2, 9.5, 6.5–5.7 and 4.8 ka. The δ13C values were steady around 12.5 ka to the present, disclosing a 200–500 year band-pass filter also show ~0.2 to 0.3‰ variations before and ~0.1‰ variations after 4.7 ka (Fig. 4C). The Δ(δ18O)/Δt results indicate highest rates of changes around 4.7 ka (Fig. 4D).

In addition, spectral analysis of the detrended δ18O data shows a 500-year periodicity above 95% confidence level (CL) after 4.7 ka (Fig. 5C) and a 330-year periodicity above 95% CL before 4.7 ka (Fig. 5D).

The δ13C values range from −9.0‰ to −4.2‰ (Fig. 3D). A major positive excursion between 12.5 and 11.3 ka reached −6‰ at ~11.7 ka. The δ13C values were steady around −7.4‰ between 11.3 and 9 ka and thereafter steadily increased to −4.5‰ at 8.5 ka (Fig. 3D). From 8.2 to 5.5 ka, δ13C values averaged −6.0±0.4‰. From 5.5 to 4.7 ka, δ13C increases from −6 to −5‰ with little fluctuation, followed by a decreasing trend to −5.5‰ at 3.5 ka. From 3 ka to the present, δ13C values remained nearly constant at −5.0±1‰. The cross-spectral analysis of the detrended δ18O and δ13C values (Fig. 6A, B) shows highest coherency at ~0.8 (Fig. 6C) and in-phase (~0°) relationship (Fig. 6D) at the 910-year periodicity (although both curves have a very weak correlation, with an R value of ~0.05), which is also supported by the covariance of the smoothed detrended δ18O and δ13C time series after applying a 910±80 year band-pass filter (Fig. 6E).

5. Discussion

5.1. Interpretation of calcite stable isotopes at Lake Grinnell

Changes in calcite δ18O values obtained at Lake Grinnell likely reflect changes in climate conditions in northern New Jersey, rather...

![Fig. 4. Time series analyses of calcite δ18O time series from Lake Grinnell, New Jersey. (A) Raw δ18O values with RAMPFIT line to show change in the long-term trend. (B) δ18O deviations from long-term fitted RAMPFIT line (detrended record). (C) Smoothed curve of raw δ18O time series after a 200–500 year band-pass filter to show low-frequency components. (D) The rate of change (100 × Δ(δ18O)/Δt) based on raw δ18O time series.](image)
than a local hydrogeological behavior for the following reasons: (1) the lake has a very short-residence time of about 80 days. Thus, the lake water would be quickly replaced by inflowing ground and surface water. (2) Groundwater table monitoring around White Lake (Sussex County), 500 m south of Lake Grinnell within the same valley, from fall 1992 to spring 1994 showed that the water table responded quickly to precipitation events (Nicholson, 1995). (3) Chara encrustations (see details below), the calcite materials that we used for \( \delta^{18}O \) analyses, mostly precipitate from late spring to summer. Thus they only capture the summer water and climatic signal. (4) Our measurements of summer lake water show higher \( \delta^{18}O \) values when comparing to well water and winter lake water (Fig. 2B, Table 1), suggesting that precipitation contributes significantly to the lake water during summers. Therefore, we argue that calcite \( \delta^{18}O \) values at Lake Grinnell likely reflect the summer conditions in the region.

The fine-grained calcite that we analyzed is comprised of mostly disintegrated Chara encrustations as well as some endogenic calcite. Chara uses dissolved CO\(_2\) from the lake water for photosynthesis, which is facilitated significantly by Chara calcification (McConnaughey, 1991; McConnaughey and Falk, 1991), inducing calcite precipitation as a result. The \( \delta^{18}O \) values of Chara encrustations have been documented to be close to the expected equilibrium values (Coletta et al., 2001; Ito, 2002). Our stable isotope measurements of Lake Grinnell water are consistent with prior results. At our study region, the average warm season (May–September) temperature between 2006 and 2008 is about 21°C, with a maximum mean monthly temperature of \( \sim 27^\circ C \) and a minimum temperature of \( \sim 15^\circ C \). The temperature of Chara photosynthetic activity is probably close to the average warm season temperature, considering the high heat capacity of water assuming that the groundwater inflow mixes quickly with the lake water. For an average temperature of 21°C and average warm season \( \delta^{18}O_{\text{water}} \) values of \( -6.6^\circ\text{m} \) (VSMOW) (see Table 1), the calculated \( \delta^{18}O_{\text{calcite}} \) at equilibrium is \( -8.1^\circ\text{m} \) (VPDB) using the fractionation equation published in Leng and Marshall (2004) as modified from Kim and O’Neil (1997). The uppermost \( \delta^{18}O_{\text{calcite}} \) value from the core is \( -8.2^\circ\text{m} \) (VPDB), which is very close to the calculated value of \( -8.1^\circ\text{m} \).

The \( \delta^{18}O \) values of calcite in isotopic equilibrium reflect isotopic composition and temperature of the lake water during calcite precipitation. The water in Lake Grinnell is derived from groundwater and precipitation, with groundwater table responding quickly to precipitation events suggesting a relatively shallow aquifer and local recharge (see details in Section 2). The \( \delta^{18}O \) values of meteoric precipitation are controlled by moisture sources and air temperature. Evaporative enrichment can also increase the lake water \( \delta^{18}O \) values. The lake groundwater isotope values and regional precipitation isotopic compositions fall on the GMWL (Fig. 2B), suggesting that this short-residence time lake has a minimal evaporation (Leng and Marshall, 2004 and references therein). Therefore, it is unlikely that the changes in precipitation/evaporation ratio control the \( \delta^{18}O_{\text{calcite}} \) signal. Changes in the moisture source of precipitation could also influence the \( \delta^{18}O \) values of lake water. For example, Burnett et al. (2004) investigated the moisture sources of winter precipitation in New York state and concluded that the lake-effect precipitation had a much lower \( \delta^{18}O \) values (\( -17.9^\circ\text{m} \)) than the moisture originating in the Gulf of Mexico and in the nearby Atlantic Ocean (\( -8.2^\circ\text{m} \)). However, the averaged \( \delta^{18}O \) values of annual precipitation in northern New Jersey is about
–8‰ (Bowen and Wilkinson, 2002). The shortest distance from the Laurentide Great Lakes (Lake Erie) to Lake Grinnell is ~300 km and they are separated by the Allegheny Mountains which could act as a sufficient orographic barrier against any lake-effect precipitation from reaching northern New Jersey. Thus the major moisture source in northern New Jersey is more likely from the Gulf of Mexico and tropical Atlantic Ocean (Lawrence et al., 1982; Peixoto and Oort, 1983). Changes in lake water temperature of this small, short-residence-time lake likely follow the air temperature. Hence changes in air temperature are the most likely cause of changes in δ18O values of Chara encrustations at Lake Grinnell. As the average of monthly weighted mean precipitation δ18O values in eastern North America show a close correlation with the monthly air temperature in northern New Jersey (Fig. 2A). The annual weighted mean precipitation δ18O values and air temperatures (http://isohis.iaea.org/userupdate/GNP2001yearly.xls) from the five GNP stations (Fig. 1A) are also correlated (δ18Oprecipitation = 0.7 × Tair − 14.9; R² = 0.70) for the period 1962–2001. Considering both the positive relationship between meteoric water δ18O and air temperature at 0.7 ‰ per °C in eastern North America, and the negative relationship with water temperature during calcite precipitation at −0.24 ‰ per °C (Craig, 1965), we can use a simple carbonate δ18O-air temperature relation of 0.46 ‰ per °C as a first approximation to estimate air temperature changes.

The δ13C values of Chara encrustation depend mainly on local factors, particularly the δ13C values of DIC in lake waters (Ito, 2002). The δ13C of lake water DIC could be influenced by the atmospheric-lake exchange of CO2, the DIC isotope composition of groundwater inflow, the decomposed or respired organic matter input into the lake water, and especially the photosynthetic activity by Chara (McKenzie, 1985; Cole et al., 1994; Ito, 2002; Leng and Marshall, 2004). The δ13C values were unlikely influenced by interaction with atmospheric CO2 because of the lake’s small size, short water residence time and the relatively constant atmospheric CO2 concentration during the Holocene (with a range of about 25 ppmv; Ingersoll et al., 1999). The groundwater DIC δ13C value of −12‰ (Table 1) reflects about equal contributions from both limestone bedrock (~0‰) and C3 plants in this region (from about −32 to −20‰; Leng and Marshall, 2004 and references therein). The decomposed and respired organic matter could decrease the DIC δ13C value of lake water by releasing 13C from organic matter. Chara will increase the δ13C of the DIC by preferentially incorporating 13C in photosynthesis (McConnaughey et al., 1997). The lake water DIC δ13C values range from −7 to −8.5‰ (Table 1), which are higher than that of groundwater DIC, likely to be the result of extensive photosynthetic activity by Chara that covers the lake floor.

5.2. Holocene climate history inferred from calcite isotope records

The negative excursion of ~1.5‰ in δ18O at 12.5–11.3 ka represents a 3–4 °C cooling in air temperature, corresponding to the Younger Dryas (YD) cold event (e.g. Cwynar and Levesque, 1995; Shuman et al., 2002; Yu, 2007). The 3–4 °C temperature decrease during the YD inferred at Lake Grinnell is comparable to other temperature records in the northeastern United States, including a similar cooling inferred from pollen records in New Jersey and Connecticut (Peteet et al., 1993), the ~5 °C cooling derived from δ18O values in carbonate at White Lake (Warren County), New Jersey (Yu, 2007), the 5.6 °C cooling at Blood Pond in Massachusetts (Hou et al., 2007), and the more than 5 °C cooling at Crooked Pond in Massachusetts (Huang et al., 2002) estimated from δD values in palmitic acid. This cold period was locally
characterized by increased abundance of boreal taxa such as Pinus, Picea, Betula, and Alnus (Zhao et al., in press). The positive excursion of $\sim 3{\%}_{\text{w}}$ in $\delta^{13}C$ at YD may reflect the decreased decomposition and respiration rate of organic matter in the cold climate. The timing of the YD event recorded at Lake Grinnell is 400–500 years younger than other regional records (e.g., Shuman et al., 2002), probably due to the age extrapolation in our age model from the oldest dating control in the early Holocene.

From 11.3 to 5.8 ka, $\delta^{18}O$ values remain relatively high around $-7{\%}_{\text{w}}$. For the period before 9 ka, other climate reconstructions from eastern North America show a cool but gradually increasing temperature (Suter, 1985; Anderson et al., 1997; Huang et al., 2002; Viau et al., 2006) and a dry climate (e.g., Dwyer et al., 1996; Newby et al., 2000; Lavoie and Richard, 2000; Almquist et al., 2001; Shuman et al., 2001, 2004; Dieffenbacher-Krall and Nurse, 2005).

The cool temperatures would have caused lower precipitation $\delta^{18}O$ values to Lake Grinnell region, but the dry climate might cause higher lake water $\delta^{18}O$ values due to longer water residence time leading to stronger evaporation effect than present. This interval was also characterized by a vegetation transition from a pine forest to an oak forest, and a decline of hemlock (Zhao et al., in press), consistent with reduced moisture availability. The low $\delta^{13}C$ values at 11.3 ka could be caused by the less limestone contributions to the groundwater DIC induced by reduced groundwater recharge under a dry climate.

A warm early to mid-Holocene climate indicated by the stable $\delta^{18}O$ values from 9 to 5.8 ka is in agreement with many studies in the northeastern United States and around the North Atlantic region (e.g., Stuiver, 1970; Davis et al., 1980; Suter, 1985; Huang et al., 2002; Ellis et al., 2004; McFadden et al., 2004; Hou et al., 2006). At Crooked Pond in Massachusetts, $\delta^{13}C$ values indicated an annual temperature rise by 3–7 °C between 8.2 and 5.4 ka (Huang et al., 2002). Similar magnitude temperature rise was also reconstructed from the pollen-inferred January mean temperature at Winneconnet Pond in Massachusetts (Suter, 1985). At Lake Grinnell, this period was also characterized by the stabilization of oak forest, with a high hemlock population (Zhao et al., in press). The relatively high and stable $\delta^{13}C$ values could reflect the high photosynthesis activity by *Chara*.

Starting at 5.8 ka, $\delta^{18}O$ values shifted from a stable state to a gradual and steady decreasing trend reflecting a $-2{\circ}C$ cooling since the mid-Holocene if this shift was controlled by temperature alone. A dry climate between 5 and 3 ka was reported from most lake sediment records in eastern North America (Dwyer et al., 1996; Mullins, 1998; Yu et al., 1997; Newby et al., 2000; Lavoie and Richard, 2000; Almquist et al., 2001; Mullins and Halfman, 2001; Shuman et al., 2001, 2004; Dieffenbacher-Krall and Nurse, 2005). This dry interval corresponds with a period of low lake water $\delta^{18}O$ values due to longer water residence time leading to stronger evaporation effect than present. This interval was also characterized by a vegetation transition from a pine forest to an oak forest, and a decline of hemlock (Zhao et al., in press), consistent with reduced moisture availability. The low $\delta^{13}C$ values at 11.3 ka could be caused by the less limestone contributions to the groundwater DIC induced by reduced groundwater recharge under a dry climate.

5.3. Climate shift in the mid-Holocene

The $\delta^{18}O$ value shift at mid-Holocene seems to be a regional signal and has been revealed by other $\delta^{18}O$ records from lake sediment carbonate in eastern North America (Fig. 7). In New York, the $\delta^{18}O$ values from Queechy Lake (Stuiver, 1970) and Green Lake (Kirby et al., 2002) show a negative shift around 6–5 ka. In southern Ontario, the $\delta^{18}O$ values from Crawford Lake (Yu et al., 1997) and Inglesby Lake (Edwards and Fritz, 1988) also show a negative shift around 5 ka. Similar negative shift was recorded in the $\delta^{18}O$ values from Pretty Lake in Indiana (Stuiver, 1968).

Although the rate and specific timing of the shift are different from site to site, all these shifts seem to imply a consistent climate transition in eastern North America during the mid-Holocene. In addition, a mid-Holocene decrease in $\delta^{18}O$ was found from Crooked Pond in Massachusetts (Huang et al., 2002). We suggest that the transition is likely caused by the same climate driver and occurred at same time considering the poor dating and coarse resolution for some of these published records.

The $\delta^{18}O$ deviations to the RAMPFIT output (Fig. 4B), the smoothed time series by the band-pass filter (Fig. 4C), and spectral analysis results of $\delta^{18}O$ data (Fig. 5C, D) indicate that the isotopic record of Lake Grinnell had higher frequency and larger amplitude variations during the early to mid-Holocene (before 4.7 ka) than the late Holocene (after 4.7 ka; Figs. 4, 5). The change in the amplitude and/or frequency of climate variability during the Holocene has been reported in paleoclimate archives in North America. For example, temperature reconstructions from pollen data in eastern North America showed that the amplitude of millennial-scale oscillations has been decreasing during the Holocene (Viau et al., 2006). Similarly, a clear change in variability (both amplitude and frequency) at 5 ka was shown in the sea surface temperature (SST) data obtained from the subpolar North Atlantic (Berner et al., 2008). Decreased amplitude of millennial-scale oscillations since the mid-Holocene in the western Canada was also suggested by a model of climate fluctuations extended back to 15 ka based on a 4000 year paleoclimatic record in conjunction with the Milankovitch cycle (Campbell et al., 1998). In addition, the ratio of $\delta^{18}O$ changes (Fig. 4D) indicated that the most rapid climate variations happened in the mid-Holocene around 4.7 ka.

The mid-Holocene hemlock decline was thought to be a rapid vegetation response to climate change (Yu et al., 1997; Shuman et al., 2004; Foster et al., 2006). Close examination of the difference in the timing of the shift in $\delta^{18}O$ and the decline in hemlock pollen (Zhao et al., in press) at Lake Grinnell (both pollen and isotope data from the same samples) (Fig. 8), indicates there is a delay between the hemlock decline and the of climate change. The climate shift occurred at 5.8 ka (at 1014 cm below the lake surface), while the hemlock decline started between 5.45 and 5.3 ka (at 988–996 cm below the lake surface). We have radiocarbon ages from above and below the hemlock decline. The dating error for both ages yields a maximum error of $\pm$ 143.5 year (also see Table 2). The age difference between the $\delta^{18}O$ shift and hemlock decline, based on our age-depth model, indicates the hemlock decline lagged $\sim$350–500 ($\pm$143.5 year) behind the climate shift. Similar age difference is obtained if we use linear interpolation to construct the age model. If the mid-Holocene hemlock decline was...
really caused by climate change, our data suggest that the decline took ~350–500 years to respond. The ~350–500 year lag is similar to the suggestion from other sites that the hemlock decline occurred >300 years after climate change (Shuman et al., 2009).  

5.4. Possible causes of Holocene climate trend, variability and shift  

Changes in insolation have been previously considered as driving long-term climate trends in the northeastern United States (Kirby et al., 2002) (Fig. 9A, B, and C). The presence of the remaining Laurentide Ice Sheet (LIS; Fig. 9A) also affects the climate system. For example, the cool and dry early Holocene was probably caused by the effect of cold glacial anticyclone circulation from the remaining LIS (COHMAP, 1988). The glacial anticyclone creates an anomalous north-northeasterly flow on the south-east of the ice sheet (Kutzbach et al., 1998; Webb et al., 1998). Thus, cloudiness and precipitation south of the ice sheet were suppressed (Felzer et al., 1996; Webb et al., 1998). This mechanism would be responsible for the cool and dry early Holocene condition in the northeastern United States (Shuman and Donnelly, 2006). An analysis of spatial and temporal expression of early Holocene climate around the North Atlantic region shows that locations close to LIS experienced maximum Holocene temperatures that lagged peak summer insolation about 1000–3000 years (Kaplan and Wolfe, 2006). In addition, reconstructed SSTs along the northwestern Atlantic slope waters show a cool early Holocene close to the LIS but a warm early Holocene farther south (Sachs, 2007). After the retreat of LIS, insolation became the dominant control (Fig. 9A).  

The mid-Holocene climate shift at 5.8 ka recorded at Lake Grinnell (Fig. 9D) likely reflects a non-linear response to the gradual decrease of insolation within the climate system. General circulation models (GCM) have shown that subtle changes in insolation, aided with the positive feedbacks between atmosphere, ocean, sea ice, and vegetation, could produce significant climate responses (e.g. Foley et al., 1994; Ganopolski et al., 1998; Claussen et al., 1999; Kerwin et al., 1999). As summarized by Steig (1999) and Sandweiss et al. (1999), the mid-Holocene climate shift has been documented in various paleoclimate records and in cultural changes at global scale. In addition, the fastest rate of annual insolation decrease (Fig. 9C) could promote a more unstable climate during the mid-Holocene through rapidly changing thermal gradient at different latitudes. The insolation decrease could cause a significant atmosphere circulation change in eastern North America. Several mechanisms of the atmosphere circulation change have been proposed for the decreased precipitation δ18O values observed in eastern North America (Fig. 7), including the more frequent intrusions of dry Pacific air mass with low δ18O values (Yu et al., 1997), the increased cold season precipitation (Denniston et al., 1999; Shuman and Donnelly, 2006), and the higher altitude of the moisture condensation elevated by the cooler and drier western air (Smith and Hollander, 1999). The contribution of Pacific air mass is unlikely the cause of the δ18O shift at Lake Grinnell because it can bring very little moisture even into the continental interior region like Iowa (Simkins, 1995). In the case of the latter two mechanisms, the average air temperature would be lower at Lake Grinnell region and thus decrease the δ18O values.  

The frequency of centennial-scale climate oscillations recorded in Lake Grinnell sediment is possibly caused by a change in dominant frequency of solar variations. The results of spectral analysis show that the 330-year periodicity was significant before 4.7 ka, but the 500-year periodicity was significant after 4.7 ka in both δ18O and Δ14C time series (Fig. 5). In order to better show the relationship of these two curves at the dominant frequencies, we smoothed both time series after applying a 330 (±30)-year band-pass filter before 4.7 ka and a 500 (±50)-year filter after 4.7 ka on both the detrended δ18O and raw residual Δ14C time series. Both filtered curves seem to show corresponding variations, especially after 6 ka (Fig. 9E). The poor correlation in some intervals during the early Holocene may be
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caused by the relatively larger dating errors at Lake Grinnell. Small solar output variations would be amplified in the climate system through several mechanisms, including the changes of ozone production and stratospheric temperature structure caused by changes in ultraviolet part of solar spectrum (Haigh, 1994; Shindell et al., 1999), the changes in cosmic ray intensity on cloud formation and precipitation (van Geel et al., 1999), the changes in global-scale atmosphere circulation (Rind and Overpeck, 1994), and the changes of sea ice and the North Atlantic deep water formation (Chapman and Shackleton, 2000; Bond et al., 2001). Solar-forced centennial-scale climate oscillations have also been widely hypothesized in North America, for example, in the northern Great Plains (Yu and Ito, 1999), in the midcontinent (Tian et al., 2006), in the southwestern United States (Asmerom et al., 2007), and in the sub-arctic Alaska (Hu et al., 2003).

The solar forcing alone may not fully explain the mid-Holocene shift in centennial-scale climate oscillations as the amplitude of solar variations does not match the amplitude of the climate variability (Fig. 9E). The observed amplitude shift of the centennial-scale climate oscillations was possibly caused by seasonality changes in insolation. For example, the contrast between summer and winter insolation was larger during the early to mid-Holocene than during the late Holocene (Fig. 9E). The strong seasonality could cause intense internal oscillations in the oceanic and possibly atmospheric circulations during the early to mid-Holocene, as shown in the North Atlantic region (Schulz and Paul, 2002; Berner et al., 2008) and in western Canada (Campbell et al., 1998). In addition, the most rapid decrease in annual insolation during the mid-Holocene might further facilitate climate oscillations (Fig. 9C).

Consequently, the early to mid-Holocene centennial-scale climate variability was not only at higher frequency but also at higher amplitude than the late Holocene, especially with the highest amplitude during the mid-Holocene.

6. Conclusions and implication

1. The isotopic record at Lake Grinnell provides a detailed climate history and landscape changes from the late-glacial to Holocene for the northeastern United States. The negative excursion of ~1.5 ‰ in δ18O at ~11.3–11.9 ka represents a regional expression of the Younger Dryas cold event. The δ18O data show a high and stable mean value of about ~7.4‰ during the early to mid-Holocene from 11.3 to 5.8 ka and a negative shift to a steady decreasing trend from ~7.4‰ to ~8.2‰ after 5.8 ka. The carbon–oxygen isotope covariance at 910-year periodicity likely reflects the controls of air temperature on both δ18O and δ13C values in temperate lakes at millennial timescale.

2. The general Holocene climate trend recorded at Lake Grinnell was dominantly controlled by the insolation. The remaining Laurentide Ice Sheet would be responsible for the cool and dry climate from 11.3 to 9 ka in eastern North America, as also shown in other regional records. The 5.8 ka climate shift in the long-term trend recorded at Lake Grinnell likely indicates a non-linear response of atmosphere circulation to a gradual decline of insolation during the mid-Holocene, as also documented in other regional isotope records. Responding to this climate shift, the hemlock declined in the northeastern United States, but lagged ~350–500 (+143.5) years behind the climate shift.

3. The mid-Holocene shift in climate variability at centennial-scale at 4.7 ka was detected not only in the periodicity shift from 330 years to 500 years, but also in the decrease in the amplitude. The increased periodicity at 4.7 ka was likely controlled by the dominant frequency shift in variations of solar output. The reduced amplitude at 4.7 ka was possibly caused by the decrease in seasonal insolation contrast and accompanying weakening ocean and atmosphere circulations. The fast decline in annual insolation would also facilitate the amplitude of centennial-scale climate variability during the mid-Holocene.

4. Our record from Lake Grinnell supports the notion that a gradual change in climate forcing could trigger an shift in the climate system. Also, changes in mean climate states are likely associated with changes in climate variability, in terms of both magnitude and frequency. This implies that a continued global warming may induce changes in climate variability (including climate extremes), especially if the climate system crosses a threshold into a new mean state.

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