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Short Paper

Rapid response of forested vegetation to multiple climatic oscillations during the last deglaciation in the northeastern United States

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Abstract

Isotopic and pollen results from a marl lake (White Lake) in the Mid-Atlantic region of USA indicate the coupling of climate and vegetation changes. Oxygen isotopes of calcite from this site show multiple oscillations at millennial and centennial scales, including the Younger Dryas with 3‰ negative shifts in $\delta^{18}O$ at 12.4–11.4 ka (1 ka = 1000 cal yr BP) and three cold events of magnitude 1–2‰ shifts during the Bølling–Allerød warm period (BOA) at 14.3–12.4 ka. Pollen data from the same core show nearly synchronous, close correspondence with isotope-inferred climate shifts, indicating rapid forest response to deglacial climate oscillations in southern New England. A plateau-like BOA is similar to other records around the North Atlantic Ocean.

Keywords: Bølling–Allerød period; Climate change; Fossil pollen; Lacustrine carbonate; Late glacial; Northeastern United States; Stable isotopes; Younger Dryas

Introduction

Abundant evidence indicates large and abrupt climatic oscillations occurred during the last glacial–interglacial transition around the North Atlantic region (e.g., Dansgaard et al., 1993; Levesque et al., 1993; Hughen et al., 1996; Von Grafenstein et al., 1999). However, most terrestrial records from North America show a truncated Bølling–Allerød period (Peteet et al., 1990; Grimm and Jacobson, 2004), which prevents comparison with Greenland and European records and detection of possible cross-Atlantic spatial patterns. Documenting and delineating the possible spatial patterns of these climate events will provide useful insights into potential mechanisms of regional climate variability. Also, there has been a long-lasting debate on the climatic interpretation of regional pollen sequences first proposed by Deevey (1939) in southern New England (Davis, 1969; Webb, 1986). Comparisons of independent paleoclimate and pollen records have shown the importance of climate for shaping regional vegetation history (e.g., Huang et al., 2002; Shuman et al., 2004), but temporally-detailed data from a region that is not near tree line are needed to evaluate the length of biotic lags in forested regions following abrupt climate changes during the last deglaciation (Williams et al., 2002).

Here I present a high-resolution pollen and calcite stable-isotope record from a marl sediment core collected from a hardwater lake in the Mid-Atlantic region. The objectives of this study are (1) to provide a detailed and complete climate record from the onset of Bølling warming to the early Holocene, (2) to evaluate the rapidity of forested vegetation response to deglacial climatic oscillations in southern New England using combined pollen and stable-isotope data from the same sediment core; and (3) to document and understand the spatial gradient of deglacial climate changes by comparing the new record with other records around the North Atlantic Ocean. The results presented here provide a detailed climate history for the late glacial period in North America and convincing evidence supporting the initial climatic interpretation of pollen sequences in southern New England by Deevey (1939). Also, the changes in pollen are synchronous with isotope-inferred climate shifts, indicating rapid response of forested vegetation to climate changes.

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Study site

White Lake is located in northwestern New Jersey (41°00′N, 74°55′W; 138 m asl; Fig. 1). It is a small hardwater lake of 0.26 km^2 in surface area and ~2 km ^{2} in catchment area. The lake is situated in a glaciated limestone valley (Cotter et al., 1986) and is primarily recharged by groundwater. It has a tiny ephemeral inlet at its northeast end and a single outlet at its south end (Fig. 1B). A marl bench, a band of unconsolidated calcareous deposits, occurs around most parts of the lake in shallow water (Fig. 1B). The lake's maximum water depth is about 13 m.

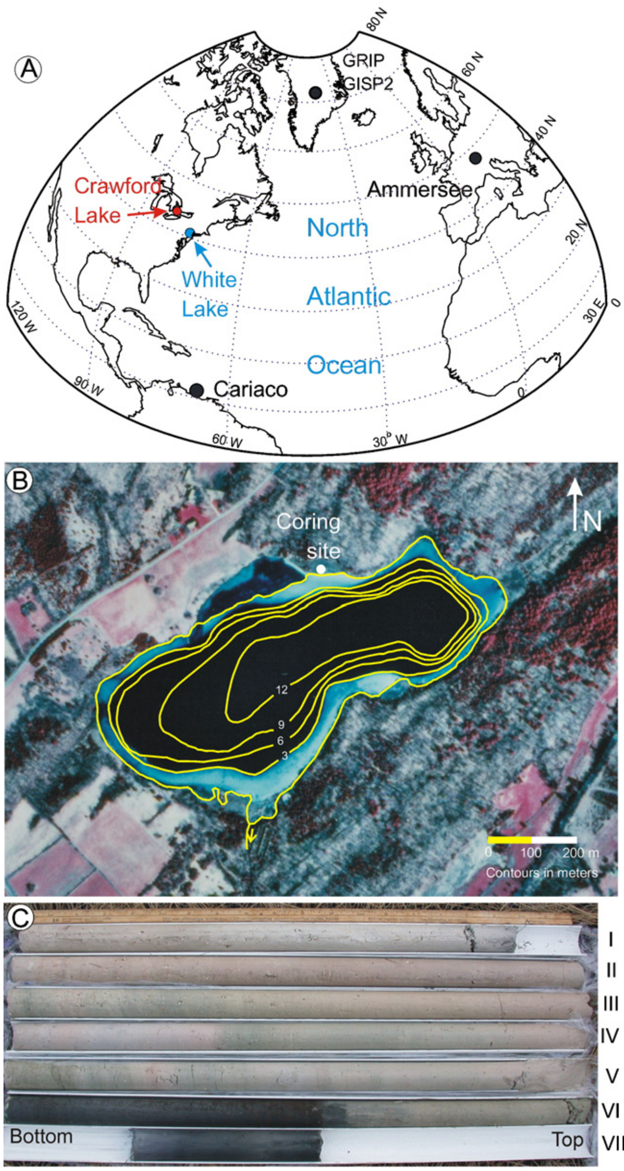

Figure 1. Location, study site and study core. (A) Location map of paleoclimate sites around the North Atlantic Ocean, including White Lake (study site) and Crawford Lake in North America, GRIP/GISP2 ice cores from Summit Greenland, Ammersee in central Europe, and Cariaco basin in tropical Atlantic. (B) Air photo, bathymetry map of White Lake and coring location. (C) Photo of sediment sections of core WL02-1 at White Lake. Sections I: 15–110 cm (top 10–15 cm peat discarded); II: 110–210 cm; III: 210–310 cm; IV: 310–410 cm; V: 410–510 cm (possible contaminated 410–420 cm); VI: 510–610; and VII: 610–640 cm.

Water samples collected in the summers of 2002 and 2003 had an averaged pH of 9.1, electrical conductivity of 576 μS/cm, δ^{18}O of −4.0‰ and δD of −31‰ (relative to VSMOW).

Methods

The sediment core (WL02-1) was taken from an old marl bench on the northern edge of White Lake (Fig. 1B) with a Livingstone–Wright piston corer of 5 cm in diameter on 16 February 2002. Two other cores were taken from open water for Holocene climate studies (Li et al., 2006). This study focuses on analysis of the lowest marl section at 555–220 cm. Loss-on-ignition analysis was carried out at 1–5 cm intervals to estimate organic matter content after combustion at 500°C and carbonate content at 1000°C. Terrestrial plant macrofossils were picked from seven samples and dated using accelerator mass spectrometry at Beta Analytic, Inc (Miami, Florida), and the age model was based on linear interpolation of calibrated ages of five accepted dates (Table 1; Fig. 2A).

Pollen analysis was done on 0.7 cm^3 subsamples at 43 intervals using a modified acetolysis procedure (Fægri et al., 1989). Pollen sum for each sample was at least 300 terrestrial pollen grains. Ordination analysis was carried out on pollen assemblages to facilitate the comparison of vegetation shifts with the isotope-derived climate variation from White Lake. I used the percentages of the 19 pollen types that reached a value of at least 2% in any one sample for principal component analysis (PCA) using the CANOCO program (Ter Braak, 1988).

For stable isotope analysis, precipitated calcite samples were taken at 1–5 cm intervals and were air dried at room temperature. Macroscopic plant remains, mollusc and ostracode shells and fragments were picked under a microscope and discarded. Each of 154 calcite samples was analyzed for oxygen and carbon isotopes at the Stable Isotope Laboratory at the University of Minnesota using a Finnigan 252 isotope ratio mass spectrometer coupled to a Kiel II carbonate preparation device. The results were presented as conventional delta (δ) notation, which is defined as $[(R_{sample} - R_{standard})/R_{standard}] \times 1000$ (where R is the absolute ratio of ^{18}O/^{16}O or ^{13}C/^{12}C, and Vienna-PDB [Peedee belemnite] is the standard for carbonates). The analytical precision is ±0.06‰ for both δ^{18}O and δ^{13}C.

Results

The 630-cm long core shows a lithologic sequence from clay (640–552 cm) to marl (552–10 cm), capped by 10 cm of wetland peat (Fig. 1C). The accumulation rate of marl was almost constant during the analyzed section (Fig. 2A). The marl contains >90% carbonate (Fig. 2B), with ~5% organic matter and silicate. The δ^{18}O values above 552 cm register the isotopic composition of authigenic calcite, ranging from −7.9 to −4.1‰ (Fig. 2C). The δ^{18}O values show large amplitude shifts, especially in the lower half. A major negative excursion at 450–390 cm reaches a minimum of −7.9‰ but was interrupted by a peak of up to −4.7‰ at 418–410 cm. δ^{13}C values range from −5.6 to 0.1‰, with the same positive spike as δ^{18}O (Fig. 2D). δ^{18}O and δ^{13}C generally show strong co-variance.

Table 1
AMS radiocarbon dates and calibrated ages for White Lake, New Jersey

Core depth (cm)	Beta lab #	Material dated	^{14}C age ± 1 SD (BP)	$\delta^{13}C$ (‰)	95% cal age range (2σ)	Age used (cal. yr BP)
281–283	179977	Charred macrofossils	9320 ± 40	−25.1	10403–10608	10505
370–372	179978	*Larix* needles, charred scales	9360 ± 40	−25.7	10494–10697	*
390–392	179979	*Picea* and *Larix* needles, charcoal	9960 ± 60	−29.1	11239–11639	11439
464–465	182730	Wood, twigs	10660 ± 120	−29.6	12347–12881	12614
464–465	180552	Charcoal (25 mg)	3070 ± 40	−25.2	3206–3378	*
499–501	179981	*Picea* needles, wood	11660 ± 80	−26.7	13330–13705	13517
545–547	179982	*Picea* needles, charred twigs	12270 ± 80	−26.8	13925–14635	14280

Note. The dates were converted into calendar years by using IntCal04 calibration dataset (Reimer et al., 2004).
* Dates were rejected and not used in the age model.

The $\delta^{18}O$ profile records the Bølling warming at 14.3 ka and the Younger Dryas event (YD) at 12.4–11.4 ka (Fig. 3A). Also, several minor oscillations are apparent during the Bølling–Allerød warm period (BOA) at 14.3–12.4 ka (Figs. 3A and 4A). The summary pollen diagram (Fig. 3E) shows clear vegetation changes, corresponding with the classic pollen sequence of Deevey (1939) for southern New England, including succession from *Picea* (spruce) zone (Deevey's regional pollen zones A-2 to A-4), through *Pinus* (pine) zone (B), to *Quercus* (oak) zone (C). *Picea* pollen abundance shows a two-step decline: from 60 to 20% at the onset of Bølling warming (Deevey's zones A-2 to A-3) and from 20 to 0% after ~11.4 ka at the beginning of the Holocene (from A-4 to B), corresponding to increases in *Pinus*, *Quercus* and *Ostrya/Carpinus* in both cases. This two-step decline pattern is similar to pollen record at Tannersville Bog in eastern Pennsylvania (Watts, 1979). A prominent peak of *Alnus* and low abundance of *Quercus* and *Ostrya/Carpinus* (zone A-4) correspond with the YD interval. Three PCA axes capture 74% of the total variance in pollen data (Figs. 3B–D). PCA-1 mostly shows a large vegetation shift from *Picea* and *Alnus* to *Quercus* and other hardwood trees at ~11.4 ka, corresponding with the warming at the beginning of the Holocene, while PCA-2 predominantly reflects cooling during the YD and possibly minor oscillations during the BOA. PCA-3 emphasizes the *Picea* decline during the Bølling warming trend.

Discussion

Deglacial climate oscillations

The oxygen-isotope records from White Lake document the classic Bølling–Allerød–Younger Dryas–Holocene climate sequence (Fig. 3A). In high- and mid-latitude regions, the isotopic signal of the source water generally prevails over the temperature effect (Rozanski et al., 1992). If we consider both the positive relationship between meteoric water $\delta^{18}O$ and air temperature at 0.6‰/C (Dansgaard, 1964), and the negative relationship with water temperature during calcite precipitation at −0.24‰/C (Friedman and O'Neil, 1977), and attribute the isotopic shifts to temperature changes, we can use a simple carbonate $\delta^{18}O$–air temperature relation of 0.36‰ per °C as a first

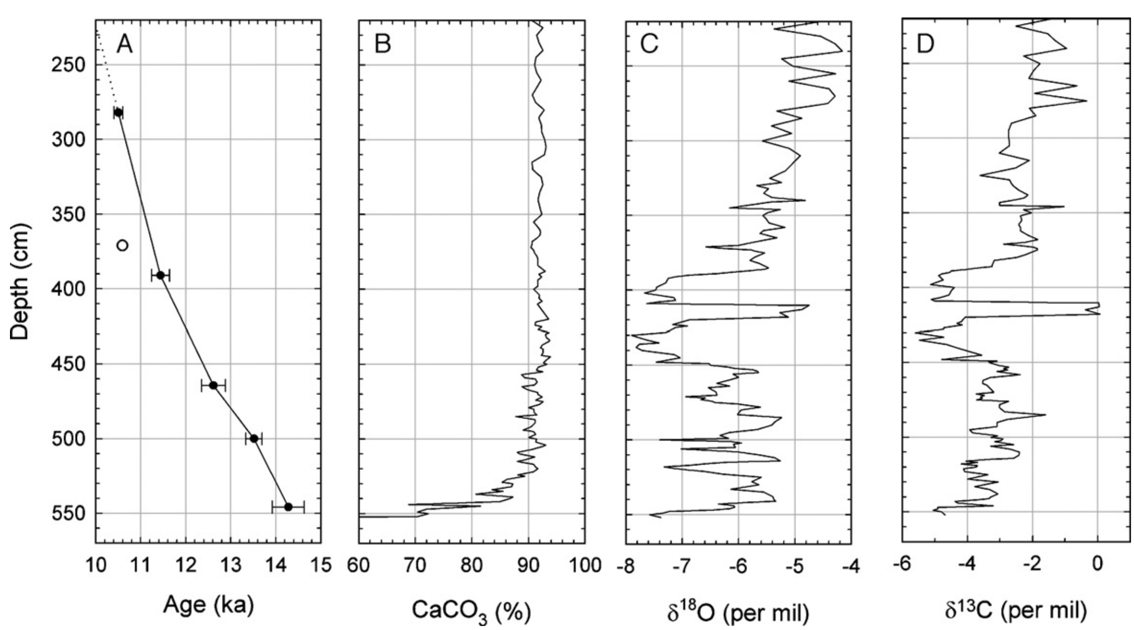

Figure 2. Results from core WL02-1 at White Lake. (A) Age-depth model. Error bars are 2σ range (95% probability). Open circle is a rejected date. (B) Carbonate percentage. (C) Oxygen isotopes from precipitated calcite. (D) Carbon isotopes from precipitated calcite.

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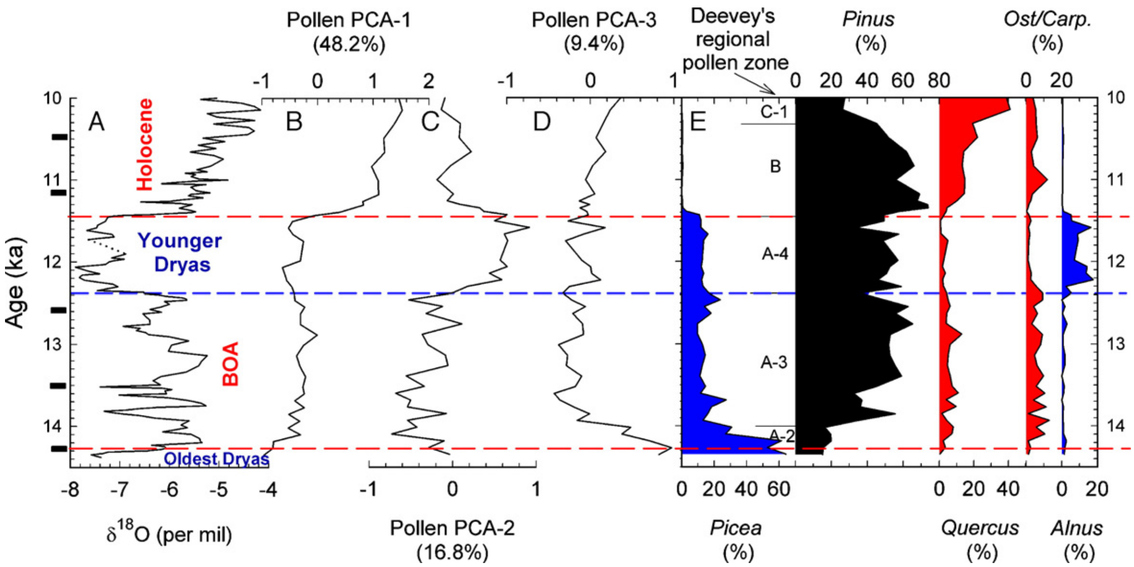

Figure 3. Oxygen isotopic and pollen data from the White Lake core. (A) Oxygen isotope record. Likely contaminated data points during the Younger Dryas were removed. Climate zones are marked. BOA = Bølling–Allerød warm period. (B–D) Scores of the first three PCA axes, representing ~70% of total variance in pollen data (numbers in brackets). (E) Selected pollen percentages: *Picea, Pinus, Quercus, Ostrya/Carpinus* (Ost/Carp) and *Alnus*. Deevey's (1939) classic pollen zones are also shown.

approximation to estimate temperature changes. The major climate shifts of ~2‰ in δ^{18}O at onsets of Bølling warming, the YD and the Holocene would then represent 5°C shifts in temperatures. During the BOA warm period, three cold events with negative excursions of ~1‰ in δ^{18}O would represent ~3°C cooling. These events include the intra-Bølling cold period (IBCP), the Older Dryas (OD), and the intra-Allerød cold period (IACP). All these major and minor oscillations are evident in other high-resolution paleoclimate records around the North Atlantic (Fig. 4).

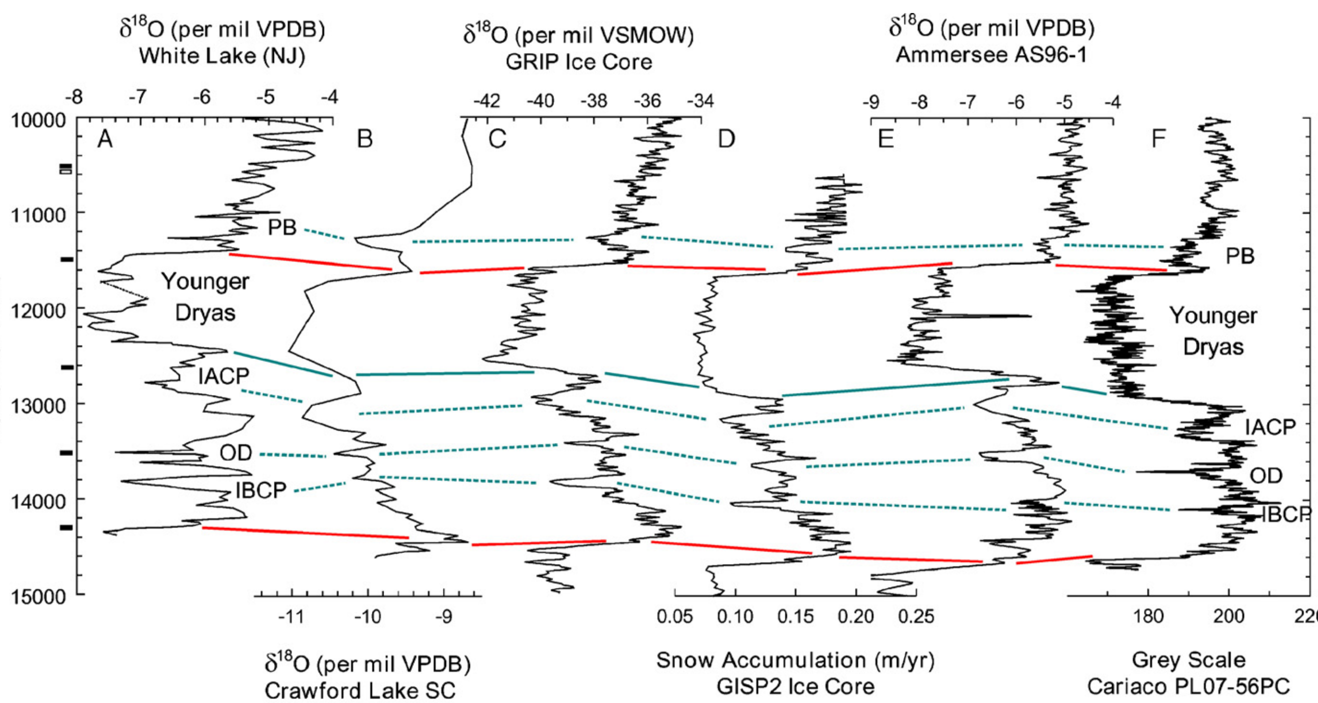

Figure 4. Correlation of paleorecords during the last deglaciation around the North Atlantic. (A) δ^{18}O of authigenic calcite from White Lake; (B) δ^{18}O of lacustrine carbonates at Crawford Lake (Yu and Eicher, 1998, 2001); (C) δ^{18}O of ice-core GRIP (Dansgaard et al., 1993); (D) Snow accumulation rates of ice-core GISP2 (Alley et al., 1993); (E) δ^{18}O of lacustrine carbonates at Ammersee, south Germany (Von Grafenstein et al., 1999); (F) Grey scale of core PL07-56PC at Cariaco basin off Venezuela (Hughen et al., 1996). Solid correlation lines indicate major climatic shifts, whereas dashed lines indicate minor cold climatic events. Three century-scale cold events during the Bølling–Allerød warm period are the IBCP (intra-Bølling cold period), OD (Older Dryas), and IACP (intra-Allerød cold period). PB, Preboreal Oscillation.

The $^{13}C/^{12}C$ ratio of authigenic calcite depends mainly on local factors, particularly through changes in $\delta^{13}C$ of dissolved inorganic carbon (DIC) of lake water. Factors affecting the ratio include exchange rates between water and atmospheric CO_2, decomposition of organic matter, and biological productivity. Considering the small size of White Lake and the short time period under consideration, the most dominant influence is likely aquatic productivity, which is also controlled by climate. This is confirmed by the strong covariance between $\delta^{18}O$ and $\delta^{13}C$ (Figs. 2C and D). As proposed by Drummond et al. (1995), in temperate lakes the strong C–O isotopic covariance is induced by positive controls of air temperature on both $\delta^{18}O$ and $\delta^{13}C$ values. High air temperatures cause high $\delta^{18}O$ values of inflowing meteoric water and high lake productivity, inducing high $\delta^{13}C$ values.

The chronology at White Lake places the major climate transitions consistently 300–400 yr younger than the timing indicated from Greenland and elsewhere, although the chronology for these events is by no means consistent from site to site (Fig. 4). The onset of Bølling warming is dated at 14.3 ka from White Lake, compared to 14.6 ka at most other sites (Fig. 4). The same is true for the ages of the YD. I do not believe that the age difference is caused by dating error, as all the dates were from terrestrial plant macrofossils that are consistent with pollen stratigraphy. Delayed response at White Lake is a possibility (Fig. 4).

The large spikes of $\delta^{18}O$ and $\delta^{13}C$ in the middle of the YD are perplexing. The fact that this loose 10-cm interval of sediments is located at the top of one core segment at 410–420 cm (see Fig. 1C) suggests that contamination during field coring might be responsible. Even if the main part of spikes are contaminated, the YD was still not uniformly cold (Fig. 3A), as also been documented elsewhere (Fig. 4E; Von Grafenstein et al., 1999; Cwynar and Spear, 2001; Ebbeson and Hald, 2004). The possible contamination would not affect the main conclusions reached in this study.

Rapid responses of forested vegetation to climate changes

Individual key pollen taxa and PCA scores indicate rapid response of forested vegetation to major climate oscillations. Pollen responses to the YD cooling and the Holocene warming were simultaneous with isotopic shifts within the sampling resolution (30–100 yr) for both pollen and isotope analysis (Fig. 3). The YD period clearly corresponds with a peak of *Alnus* pollen (Mayle et al., 1993) and decreased *Quercus* and *Ostrya/Carpinus* pollen (Fig. 3E). On the other hand, vegetation response to the Bølling warming appears to be delayed for a couple of hundred years. $\delta^{18}O$ reached Bølling peak values before 14.2 ka (550–545 cm), while the *Picea* decline and *Pinus* increase occurred at 14.0 ka (540–535 cm). This lag response is consistent with other records from around the North Atlantic (Williams et al., 2002). However, the responses at White Lake were in a forested region rather than near tree line as reported in other records, implying that forests can also respond rapidly to climate changes within the life span of these tree species. Also, the very rapid and clear response of *Alnus* to the YD is likely owing to that it is a fast-growing shrub rather than a long-lived tree. The pollen sampling resolution is not high enough to resolve the possible vegetation response to minor century-scale oscillations during the BOA, but several small peaks of PCA-2 scores and variations in *Picea* and *Pinus* are suggestive of detectable forest responses to minor climate oscillations (Fig. 3C).

It is clear that the pollen sequence first described by Deevey (1939) in southern New England closely corresponds with climate change as indicated by $\delta^{18}O$ shifts during the last deglaciation. The Bølling warming was probably responsible for the first decline of *Picea* (*Picea* to *Pinus* transition; A2 to A3), despite a possible short time lag. If that is the case, the earlier transition from Cyperaceae to *Picea* (A1 to A2) documented in regional pollen diagrams (e.g., Davis, 1969; Shuman et al., 2004) was not in response to the Bølling warming. A4 is clearly associated with the YD interval, as proposed by Peteet et al. (1990) and Shuman et al. (2004). The onset of the Holocene corresponds with the final disappearance of *Picea*. *Picea* also disappeared from southern Ontario at the onset of the Holocene, despite little change in the pollen signal during the YD (Yu and Eicher, 1998).

Cross-Atlantic connection and forcing mechanism implications

Major millennial-scale climate shifts and three century-scale cold events during the Bølling–Allerød warm period have been clearly documented at several high-resolution paleoclimatic records around the North Atlantic Ocean (Fig. 1). In addition to the $\delta^{18}O$ record (Fig. 4C; Dansgaard et al., 1993) and snow-accumulation record (Fig. 4D; Alley et al., 1993) from Greenland, which both represent changes in local climate, similar oscillations in chemistry of the GISP2 ice core (e.g., Ca, Cl, K, Mg, Na) and the derived Polar Circulation Index suggest shifts in large-scale atmospheric-circulation patterns (Mayewski et al., 1997). These multiple cold events have also been documented in terrestrial $\delta^{18}O$ records at Ammersee, southern Germany (Fig. 4E; Von Grafenstein et al., 1999), suggesting changes in mid-European air temperatures. In the Cariaco basin of the tropical Atlantic Ocean, grey-scale measurements of varved marine sediments, thought to be related to upwelling and trade-wind strength, also show three oscillations during the BOA warm period (Fig. 4F; Hughen et al., 1996).

The general trend during the BOA warm period is different among various records. At Crawford Lake, the $\delta^{18}O$ values declined more than 2‰ during the BOA warm period (Fig. 4B; Yu and Eicher, 1998), and similar declining trends occurred in Greenland $\delta^{18}O$ and snow accumulation (Figs. 4C and D). However, the records from White Lake, Ammersee and Cariaco show a plateau-like Bølling–Allerød warm period (Figs. 4A, E and F). The atmospheric CH_4 records from GRIP and GISP2, which mostly reflected wetland area in tropical regions at that time, also show a plateau-like or even increasing trend during the BOA (Chappellaz et al., 1993; Brook et al., 2000). These trans-Atlantic similarities and differences hint at the existence of a strong spatial gradient in climatic changes between low and high latitudes. Von Grafenstein et al. (1999)

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identified a climatic asymmetry between Greenland and Europe on the basis of their reconstructed meteoric precipitation $\delta^{18}O$ values derived from ostracode shells. The downward-trending and plateau-like patterns could be controlled by shifts in polar front at the time (Ruddiman and McIntyre, 1981). The polar front may have been anchored in North America between White Lake and Crawford Lake and shifted back and forth during the last deglaciation. For example, the polar front might have acted as a boundary of two different climate regions/controls during the BOA period: maintaining long-term mean climate state in the southeast of the boundary, but cooling trend in the northwest. These isotopic shifts might also reflect change in atmospheric circulation and associated changes in moisture sources. In any case, three minor cold climate events during the BOA appeared to be able to penetrate the boundary and affect both low and high latitude regions in the same manner. If this notion can be confirmed by additional records, it implies a much steeper climate gradient in eastern North America than previously thought (Levesque et al., 1997).

Conclusions

(1) The isotopic record at White Lake provides a detailed climate history during the last deglaciation for northeastern North America, revealing the onset of the Bølling warming at 14.3 ka, the Younger Dryas at 12.4–11.4 ka and three century-scale oscillations during the Bølling–Allerød warm period. These oscillations are comparable with other records around the North Atlantic but are consistently younger by a few hundred years.

(2) Pollen analysis documents rapid forest response to late-glacial climate changes, indicated by the isotopic record from the same core, with almost no time lags and unequivocally confirms that the pollen sequence as first identified by Deevey (1939) for southern New England closely corresponds with climate changes.

(3) The trans-Atlantic pattern of the Bølling–Allerød period suggests that a steep climate gradient was present along eastern North America, possibly in response to the orientation of the polar front at the time.

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