Palaeogeography, Palaeoclimatology, Palaeoecology 350–352 (2012) 171–179

Contents lists available at SciVerse ScienceDirect

Palaeogeography, Palaeoclimatology, Palaeoecology

journal homepage: www.elsevier.com/locate/palaeo

Multiple early Holocene climate oscillations at Silver Lake, New Jersey and their possible linkage with outburst floods

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ARTICLE INFO

Article history:
Received 12 October 2011
Received in revised form 21 June 2012
Accepted 30 June 2012
Available online 9 July 2012

Keywords:
Oxygen stable isotopes
Early Holocene
Lake level fluctuation
Proglacial Lake Agassiz
Outburst floods
Climate change

ABSTRACT

Episodic discharges from the Laurentide Ice Sheet are considered to be one of the major causes of climate oscillations during the glacial to interglacial transition. These outburst floods disrupted thermohaline circulation and might have caused cooling events in the North Atlantic region and beyond, including the Younger Dryas, Preboreal Oscillation, and 8.2-ka events. However, few terrestrial sedimentary records show regional climate changes directly linked to outburst floods during the early Holocene. Here we present lithologic and isotopic data from Silver Lake in northern New Jersey to document temperature and moisture changes during the early Holocene and to discuss possible linkage with outburst flooding events. The lithologic record shows multiple intervals with coupled decreases in carbonate and increases in organic matter during the early Holocene, suggesting lower lake levels and drier climate conditions. Simultaneous with these changes were decreases in carbonate $\delta^{18}O$ values, likely reflecting a decrease in air temperature during these dry periods. The apparent correlation of these sedimentary events at Silver Lake with documented outburst floods from the melting ice sheet suggests a possible causal connection, implying that the climate in the Mid-Atlantic region was extremely sensitive to perturbations of ocean circulation.

1. Introduction

Increasing evidence suggests that late glacial and early Holocene climates were highly variable in some parts of the world (Levesque et al., 1993; Peteet, 1995; Hoek and Bos, 2007). A classic deglacial climate sequence has been documented around the North Atlantic Ocean (Fig. 1A) (Dansgaard et al., 1993; Hughen et al., 1996; von Grafenstein et al., 1999; Johnsen et al., 2001; Yu and Eicher, 2001; Anderson et al., 2006). The sequence includes glacial conditions ending around 14.6 ka (1 ka = 1000 cal yr BP) with a period of warmth known as the Bølling–Allerød warm period (BOA). The Younger Dryas (YD) cooling period (12.9–11.6 ka) has been documented all over the North Atlantic Seaboard (Peteet, 1995). Warmer temperatures began at the start of the Holocene around 11.6 ka. A very sudden and short Preboreal Oscillation cooling (PBO) occurred in the early Holocene at 11.3 ka (Magny et al., 2007) with another cooling period known as the 8.2 ka event (Alley et al., 1997).

Sudden releases of freshwater from proglacial lakes into the North Atlantic Ocean have been linked to some of these deglacial climate events. For example, the final drainage of proglacial Lake Agassiz, resulting in release of 163,000 km^3 of freshwater into the North Atlantic, is believed to have produced the 8.2-ka cooling event (Alley et al., 1997; Clarke et al., 2003; Teller and Leverington, 2004; Rohling and Pälike, 2005; Flesche Kleiven et al., 2008). This final outburst flood was caused by the collapse of an ice dam that blocked the drainage route into the Hudson Strait during retreat of the Laurentide Ice Sheet (LIS; Teller, 2004). Similar outburst floods associated with ice sheet retreat and changing isostatic conditions occurred repeatedly between the PBO and the 8.2-ka event (Teller and Leverington, 2004). In a synthesis of records from the Northern Hemisphere, Fleitmann et al. (2008) showed a climate anomaly around 9.2 ka, suggesting that a small-scale release of freshwater (on the order of ~8100 km^3) from Lake Agassiz was responsible. Although the flood at 9.2 ka was much smaller than the flood that caused the 8.2-ka event, it was sufficiently powerful to disrupt Atlantic Meridional Overturning Circulation (MOC), causing widespread climate change. Furthermore, Hou et al. (2011) presented a review of paleoclimate records suggesting that other outburst floods from Lake Agassiz between the PBO and 8.2-ka event caused detectable climate changes globally.

Here we present lithologic and isotopic data from sediments at Silver Lake in northern New Jersey, located ~60 km from the Atlantic coast. In a sediment core from the surrounding wetlands of this lake, changes in sediment composition and carbonate stable isotopes indicate possible variations in temperature and moisture that may correlate with some of the smaller-magnitude outburst floods (as small as 3000 km^3) from Lake Agassiz between the PBO and the 8.2-ka event. This relationship is consistent with ocean circulation

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doi:10.1016/j.palaeo.2012.06.028

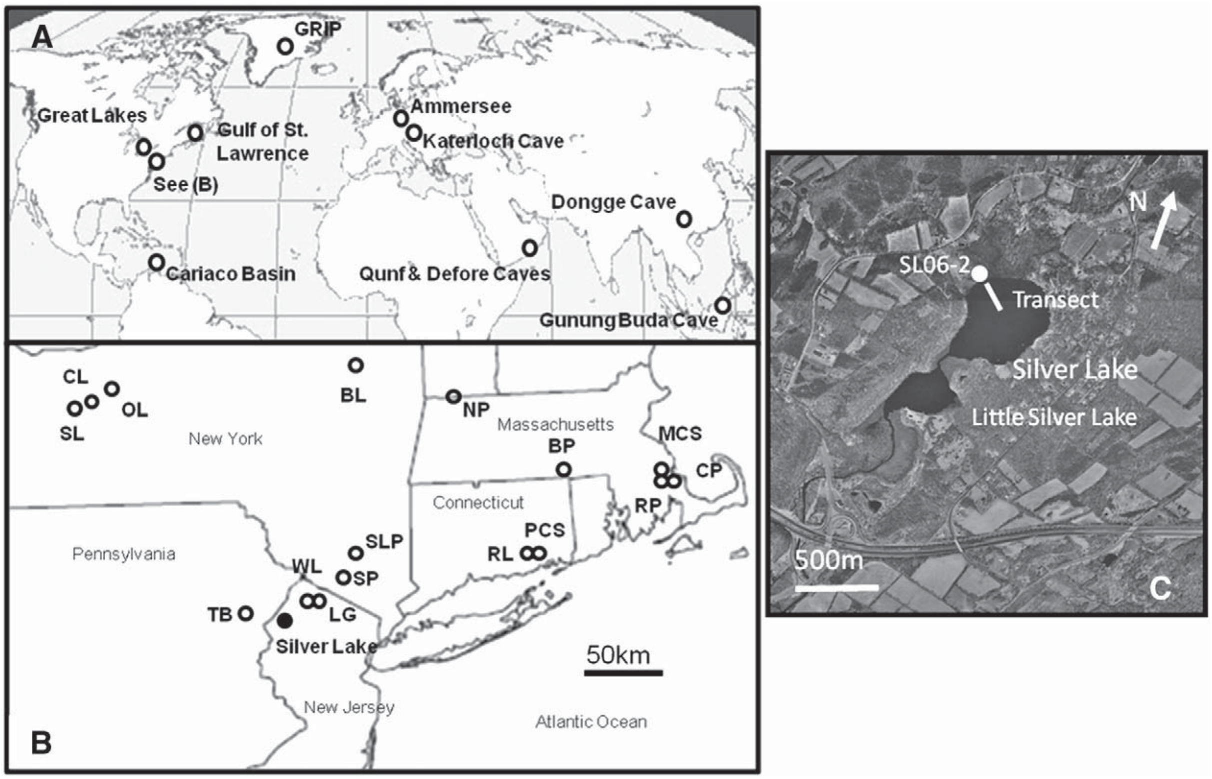

Fig. 1. (A) Location of paleoclimate sites that show deglacial climate sequences, GRIP (Vinther et al., 2006), Lake Ammersee (von Grafenstein et al., 1999), Katerloch Cave (Boch et al., 2009), Dongge Cave (Dykoski et al., 2005), Gunung Buda Cave (Partin et al., 2007), Qunf Cave (Fleitmann et al., 2007), Defore Cave (Fleitmann et al., 2007), Cariaco Basin (Hughen et al., 1996), Great Lakes region (Yu and Eicher, 1998, 2001), and Gulf of St. Lawrence (Anderson et al., 2006). (B) Locations of southern New England paleoclimate sites mentioned in this text. SL = Seneca Lake (Guiles Ellis et al., 2004); CL = Cayuga Lake (Mullins, 1998); OL = Owasco Lake (Dwyer et al., 1996); BL = Ballston Lake (Toney et al., 2003); NP = North Pond (Shuman et al., 2004); BP = Blood Pond (Hou et al., 2011); MCS = Makepeace Cedar Swamp (Newby et al., 2000); CP = Crooked Pond (Shuman et al., 2001); RP = Rocky Pond (Newby et al., 2009); PCS = Pequot Cedar Swamp (Newby et al., 2000); RL = Roger Lake (Davis, 1969); SLP = Sutherland Pond and SP = Spruce Pond (Maenza-Gmelch, 1997); TB = Tannersville Bog (Cai and Yu, 2011); WL = White Lake (Li et al., 2006 and Yu, 2007); LG = Lake Grinnell (Zhao et al., 2010a, 2010b); Silver Lake (this study). (C) Air photo of Silver Lake with location of coring site (white solid circle), SL06-2, in wetland at the northeast corner of the lake. Line leading southeast of coring site shows approximate location of the surface sediment transect in lake.

models showing that ~1600 km^3 of freshwater can disrupt ocean circulation and cause noticeable climate change (LeGrande et al., 2006). The objectives of this study were to generate high-resolution records of climate change using geochemical and sedimentological analyses of Silver Lake core samples; to reconstruct changes in moisture and temperature for this region during the early Holocene; and to document early Holocene climate oscillations and their potential relationship with meltwater outburst floods from proglacial lakes.

2. Study site and methods

Silver Lake is located in Warren County, New Jersey (40°56′12″N latitude; 74°56′12″W longitude) at ~130 m above sea level (Fig. 1). The underlying bedrock is largely Ordovician limestone and shale (New Jersey Geological Survey, 2007). The lake is in a small watershed with one surface inlet, and one outlet. Silver Lake is a glacial kettle lake that was formed from the melting of stagnant ice during the last deglaciation (New Jersey Fisheries Survey, 1950). A phragmites dominated wetland presently overlies lake marl deposits on the north shore of Silver Lake (New Jersey Fisheries Survey, 1950). Average water depth of the lake is ~4 m, with a maximum depth of ~10.5 m, and lake surface area is ~40 acres. Precipitation in this region has mean annual values of −9‰ in δ^{18}O and −58‰ in δD, with the highest values for both in August (δ^{18}O = −5.6‰; δD = −34‰) and lowest values in January (δ^{18}O = −14.6‰; δD = −102‰) (Bowen and Wilkinson, 2002; Bowen, 2010). Present lake water has O and H isotope compositions similar to those of average precipitation for this region (Fig. 2).

In February 2006, an 800-cm-long sediment core (SL06-2) was taken in the wetlands on the northwestern side of the lake, using a Livingstone-Wright piston corer (Fig. 1C). In August, 2007, surface sediments were taken from Silver Lake, using a plastic tube corer fitted with a piston, along a transect of water depths between 0.6 m and 6.7 m (Fig. 1C). The top 5 to 10 cm of the surface sediment samples were collected and homogenized in plastic bags. Loss on ignition (LOI) analyses were performed on dried sub-samples (0.7 ml) taken at 1-, 2-, or 4-cm intervals in core SL06-2 and on all surface sediment

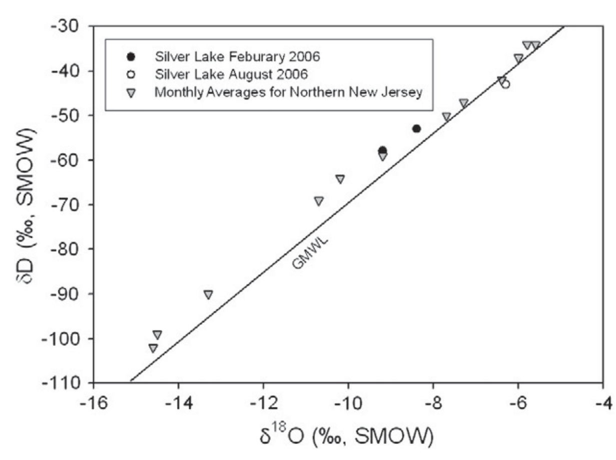

Fig. 2. Water isotopes (O and H) from surface water at Silver Lake, northern New Jersey. These δ^{18}O and δD values are similar to monthly averages for northern New Jersey precipitation (Bowen and Wilkinson, 2002; Bowen, 2010). GMWL denotes the Global Mean Water Line, and SMOW denotes standard mean ocean water.

samples. LOI samples were combusted for 2 h at 550 °C to estimate organic matter (OM) content, and then heated for 2 h at 1000 °C to estimate carbonate content (Dean, 1974).

Six intervals from core SL06-2 were selected for radiocarbon dating. Samples were sieved and terrestrial plant fragments larger than >250 μm were sent to the Keck AMS Laboratory, University of California at Irvine, for radiocarbon dating. Radiocarbon ages were calculated with the Calib 5.1.0 program using the INTCAL04 calibration data set (Reimer et al., 2004) (Table 1). An age-depth model was derived using linear interpolation (Fig. 3A) between four calibrated ages and two pollen-inferred ages. The time interval sampled by the complete core dates back to before 15.5 ka (Zelanko, 2008). This paper focuses on the early Holocene section of the record, between the PBO and the 8.2-ka cooling event.

A total of 45 subsamples from core SL06-2, from 660 to 260 cm depths, were taken for stable isotope analysis of biogenic carbonate (shells). Samples were soaked in water, subjected to a freeze–thaw cycle, and sieved through two mesh sizes, 500 μm and 125 μm. Mollusk shells of the genus *Valvata* spp. were selected from each sample. *Valvata* shells were chosen because optimal shell growth occurs within a few weeks to months during the summer (Jones et al., 2002). Therefore, their isotopic values represent the summer conditions of the water in which these animals lived (Tevesz et al., 1998). The shells were rinsed with water to remove sediment and organic matter, then dried at 50 °C in an oven. Approximately 6 shells from each sample were reacted with 100% phosphoric acid at 25 °C in order to evolve CO_2 gas (McCrea, 1950). The resulting CO_2 was cryogenically purified to remove water and then analyzed for $^{18}O/^{16}O$ and $^{13}C/^{12}C$ ratios, in dual-inlet mode, using the Finnigan MAT 252 mass spectrometer at Lehigh University.

For the stable isotope analyses of bulk sediment calcite, 69 sediment samples from 660 to 260 cm depths in core SL06-2 were taken at approximately 2 cm intervals. Samples were dried in a desiccator at room temperature. Shell and organic fragments were removed and discarded under a stereo-microscope. The cleaned inorganic carbonate samples were placed in Wheaton V-vials and capped with Kel-F discs and rubber septa. Anhydrous phosphoric acid (ρ = ~1.89 g/ml) was injected into evacuated vials for reaction at 90 °C in a MultiCarb reaction device and the CO_2 gas produced was cryogenically purified and then analyzed using a dual inlet GV Isoprime mass spectrometer at the University of Maryland Stable Isotope Laboratory.

The results for all isotope analysis are presented in conventional δ notation, defined as $\delta = [(R_{sample} - R_{standard})/R_{standard}] \times 1000$ (‰), where R is the absolute ratio of $^{18}O/^{16}O$ or $^{13}C/^{12}C$. In this paper, all O and C isotope data are expressed relative to the VPDB (Vienna PeeDee Belemnite) standard. For the analyses at Lehigh University, analytical precision, based on multiple analysis of the standard, was approximately 0.1‰ for both $\delta^{13}C$ and $\delta^{18}O$ (2σ). For the analyses at the University of Maryland, analytical precision, based on multiple analysis of the standard, was 0.04‰ for $\delta^{13}C$ and 0.1‰ for $\delta^{18}O$ (2σ).

3. Results

3.1. Chronology

Chronology for core SL06-2 was based on four calibrated AMS ^{14}C ages and two pollen-derived ages (Table 1, Fig. 3). The pollen-derived ages of 12.9 ka and 11.6 ka, representing the beginning and end of the YD, are assigned to the depths 716 cm and 680 cm, respectively (Shuman et al., 2002; Shuman et al., 2004; Yu, 2007). These pollen ages were assigned to these depths based on the presence of a peak in *Alnus* (alder) pollen that has been shown to be an indicator of the YD in the region (see Mayle et al., 1993; Peteet et al., 1993; Mayle and Cwynar, 1995; Yu, 2007; Zelanko, 2008). A large decline of *Picea* (spruce) is another indicator for the end of the YD in the northeast (Shuman et al., 2002). At Silver Lake, *Picea* disappeared at the same depth in the core where the *Alnus* peak subsided, lending another line of evidence for the age assignments (Fig. 3). Changes in sediment composition provide more evidence for the timing of the YD interval (Fig. 3). An abrupt decrease in carbonate content to ~10% during the same interval as the *Alnus* peak may be sedimentary evidence for the YD. Also, changes in sediment from 760 to 716 cm (from ~15.5 to 12.9 ka) may indicate the BOA warm period with increases in carbonate and decreases in silicate, suggesting that our chronology is consistent with regional climate sequence.

3.2. Sediment composition

Organic matter content in the surface sediments decreased from 47% at 0.6 m water depth to 19% at 6.7 m water depth, whereas carbonate increased from 6% to 35% over the same water depth transect (Fig. 4). This relationship between OM and carbonate suggests that the abundance of carbonate and OM may together be used as a relative water depth indicator for the core samples (Magny and Bégeot, 2004). Silicate content shows little variation at ~50%, this relatively high silicate content is due possibly to the recent residential development around Silver Lake.

In core SL06-2, sediments from before 15.5 ka (800–760 cm) show low carbonate (5–10%) and OM content (~5%), but high silicate (90%) (Fig. 3). Concentrations begin to fluctuate at ~15.5 ka (760 cm) and continue until 12.9 ka (716 cm), corresponding with the BOA warm period. Silicate and OM continue to fluctuate until 11.6 ka (680 cm) however carbonate remains constant during this interval at ~10%. During the period from 11.6 to 10.7 ka (655 cm), concentrations of OM, carbonate, and silicate fluctuated widely, whereas from 10.7 to 9.6 ka (550 cm), concentrations of OM, carbonate and silicate were constant at 5%, 90%, and 5% respectively (Figs. 3 and 5). Over all concentrations remained relatively constant from 9.6 ka to 8.0 ka (260 cm). However, multiple fluctuations involving increases in OM and silicate and decreases in carbonate occur, with the largest shift in sediment composition occurring between 8.3 and 8.15 ka (310–280 cm).

Table 1
Radiocarbon ages from Silver Lake, New Jersey.

UCIAMS code[a]	Depth (cm)	Material dated	Weight (mg)	^{14}C ages +/− error (yr BP)	Cal yr BP[b]	2σ range[b]
29768[c]	303–304	Bark	3.7	7440 ± 20	8260	8193–8330
29769[c]	539.5–540.5	Bark/wood, charcoal	0.8	862 ± 25	9554	9532–9630
34685[c]	645.5–646.5	Berry seeds	2.8	9260 ± 20	10,452	10,372–10,519
29770	673.5–674.5	Bark, charcoal	3.9	10,490 ± 35	12,511	12,362–12,676
34686	689.5–690.5	Bark, charcoal	1.4	9500 ± 80	10,818	10,572–11,107
29771[c]	745.5–746.5	Bark	0.7	12,890 ± 60	15,221	14,971–15,515

[a] Samples were analyzed at the University of California, Irvine Keck AMS lab.
[b] Calendar ages were calibrated with INTCAL04 calibration dataset Reimer et al. (2004).
[c] Calibrated ages used in age-depth model of core SL06-2.

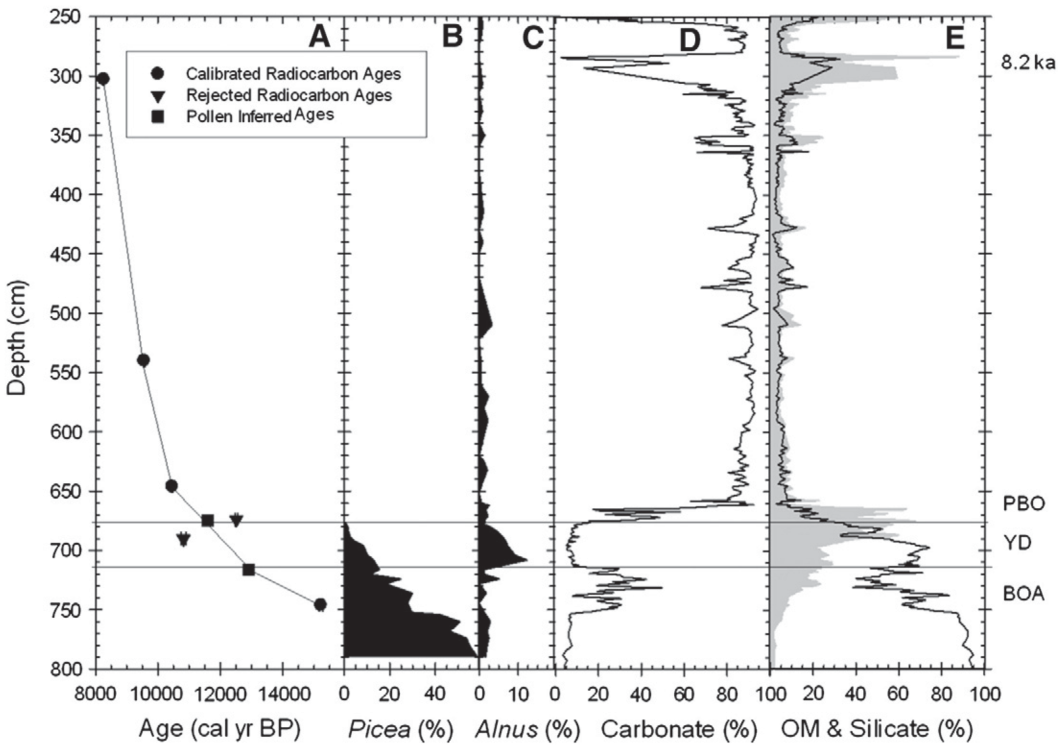

Fig. 3. Age model and sediment composition. (A) Age-depth model of core SL06-2 at Silver Lake, New Jersey. Age-depth model constructed from calibrated ^{14}C (circles) and pollen inferred ages (squares). Triangles are the rejected ^{14}C ages. (B) *Picea* percentage plotted against depth. The disappearance of *Picea* at 680 cm denotes the end of the Younger Dryas (YD). (C) *Alnus* percentage plotted against depth. The *Alnus* peak between 716 and 680 cm corresponds with the YD interval. Pollen ages are the ages for the YD chronozone for the region as in Shuman et al. (2002) and Yu (2007). (D) Carbonate, (E) OM (gray shading) and silicate (black line) percentages.

3.3. Oxygen and carbon stable isotopes

Stable isotopes from *Valvata* shells show a gradual increase of ~1‰ in both $\delta^{18}O$ and $\delta^{13}C$ from 10.9 to 9.6 ka (Fig. 5). The $\delta^{18}O_{shell}$ values then show a decline of ~2‰ until 8.4 ka, whereas $\delta^{13}C_{shell}$ values show a slightly increasing trend until 8.75 ka and then show a decrease of 1‰ until 8.4 ka. For the period from 8.4 ka to 8.0 ka, $\delta^{18}O_{shell}$ values fluctuate by 2‰, whereas $\delta^{13}C_{shell}$ values fluctuate by 3‰. Both $\delta^{18}O_{shell}$ and $\delta^{13}C_{shell}$ show multiple oscillations (up to 2‰) for the interval between 9.6 and 8.0 ka.

The $\delta^{18}O$ values for bulk carbonate follow a trend similar to that of the shells, increasing by approximately 1‰ for the period of 10.9 ka to 9.6 ka (Fig. 5). However, $\delta^{13}C_{carbonate}$ increases by more than 2.5‰ during the same interval. For the period from 9.6 ka to 8.0 ka,

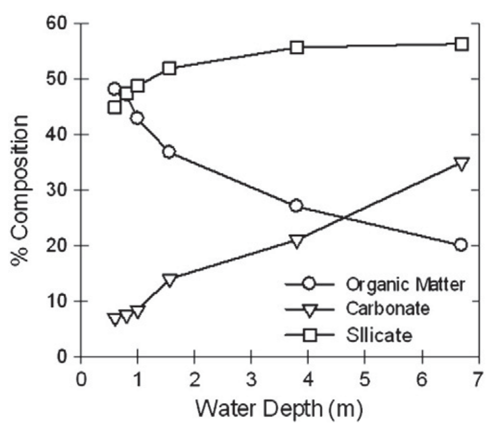

Fig. 4. Change in OM, carbonate, and silicate in surface sediments with water depth at Silver Lake. An increase in carbonate indicates higher lake levels, while an increase in organic matter indicates lower lake levels. High, steady silica content is most likely due to present day residential development.

$\delta^{18}O$ compositions of bulk carbonate show no overall trend, remaining at approximately −8.2‰ with only small fluctuations of up to 0.5‰, whereas $\delta^{13}C_{carbonate}$ varies from −3.5‰ to −7‰ with fluctuations of up to 2‰.

4. Discussion

4.1. Interpretations of sediment proxies

Changes in lake levels can provide information regarding the effective moisture of a region, whereas LOI data and sediment composition have been used to infer lake-level changes (e.g., Newby et al., 2000). Changes in sediment composition can be controlled by lake productivity, landscape erosion/mineral inputs, and water level (Shuman, 2003). If water level is the controlling factor of sediment composition then water depth can be estimated from the presence or absence of a particular sediment composition. In a lacustrine setting, OM and in particular peat often deposits on land or under very shallow water from wetland or abundant macrophyte vegetation in the littoral zone (Winkler et al., 1986). Whereas marl (unconsolidated carbonate sediments) can be deposited and preserved in open water either from *Chara*-facilitated carbonate precipitation or the lack of allogenic inputs of terrestrial organic and mineral materials to the benthic zone (Birks and Birks, 1980; Magny and Bégeot, 2004). Therefore, the presence or absence of a particular sediment composition can be indicative of lake level at the coring site. Using the established relationship between water depth and sediment composition, as seen from the surface sediment transect at Silver Lake (Fig. 4), we reconstruct lake-level change under the assumption that this relationship holds true throughout the record.

Superimposed on the water depth versus sediment composition relationship is the effects of primary productivity and mineral input in relation to sediment composition. An increase in temperatures would promote primary productivity, thereby increasing plant growth in the

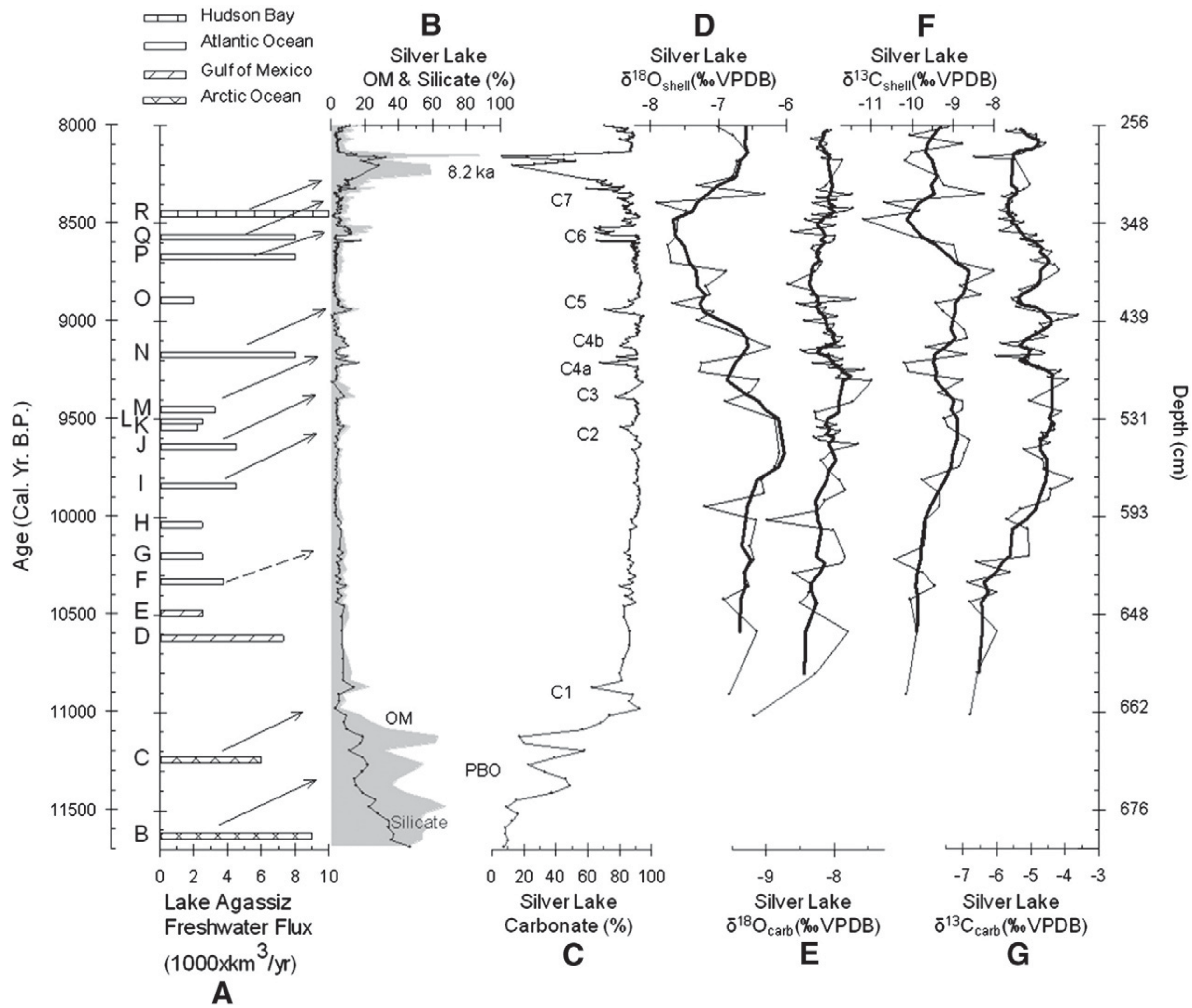

Fig. 5. Correlations of early Holocene climate changes at Silver Lake. (A) Lake Agassiz outburst floods (modified from Teller and Leverington, 2004). The 8.4-ka flood (labeled R) was 163,000 km^3/yr in magnitude (beyond the graph scale). Legend indicates the drainage routes of floods. (B) Percent OM and silicate, (C) carbonate, (D) $\delta^{18}O_{shell}$, (E) $\delta^{18}O_{carbonate}$, (F) $\delta^{13}C_{shell}$, and (G) $\delta^{13}C_{carbonate}$ from Silver Lake sediment core SL06-2. Thick lines on isotope graphs are 5-point averages. Arrows indicate the glacial outburst flooding events that may have indirectly caused climate changes and are recorded as change in sediment composition at Silver Lake. Floods are labeled according to Teller and Leverington (2004). The associated changes in lithology corresponding to the floods are labeled C1 through C7. The Preboreal Oscillation (PBO) and 8.2-ka event are noted.

littoral zone (increasing OM content) and increasing carbonate precipitation (increasing carbonate content) in the benthic zone (Anderson et al., 1997; Li et al., 2006). Before the Holocene, more siliceous materials were washed into the lake due to periglacial conditions and unstable and sparse vegetated landscape. However, since the start of the Holocene the deposition of silicate in freshwater lakes most likely indicates terrestrial erosion caused by dry or cold climates and reduced vegetation covers. A decrease or change in terrestrial vegetation may increase the proportion of silicate material in lake sediment. Diatoms can also contribute a small percentage to the total silicate content of lakes (Shemesh and Peteet, 1998).

The $\delta^{18}O$ values in freshwater carbonates can be used as a proxy for air temperature or hydrological changes. When authigenic carbonate precipitates in isotopic equilibrium with lake water, the $\delta^{18}O$ values are a function of the $\delta^{18}O$ and temperature of the water, with a relationship of −0.24‰ per °C (Friedman and O'Neil, 1977). At high and mid-latitudes, the $\delta^{18}O$ of water is primarily dependent on the atmospheric temperature effect, with a relationship of 0.6‰ per °C between air temperature and the $\delta^{18}O$ of precipitation (Rozanski et al., 1992). Combining these two factors, and assuming that water temperature closely follows air temperature, a coefficient of 0.36‰ per °C can be used to estimate a relationship between the $\delta^{18}O$ of carbonates and air temperature (Yu and Eicher, 1998). Increasing air temperature equates to increased $\delta^{18}O$ values in lake water and increased $\delta^{18}O$ values in carbonates that precipitate from that lake water.

The $\delta^{18}O$ values of open lakes also reflect the isotopic composition of the precipitation received by the lake (Leng and Marshall, 2004). Therefore, changes in the source of precipitation can contribute to changes in the $\delta^{18}O$ values of lake water and any carbonates that precipitate. Silver Lake sits east of the Appalachian Mountains, a natural orographic barrier from isotopically light lake-effect precipitation from the Great Lakes to the west. Thus, Silver Lake likely receives its moisture largely from the Atlantic Ocean (Lawrence et al., 1982; Peixoto and Oort, 1983). Average $\delta^{18}O$ values for moisture from the Atlantic Ocean (−8.2‰) are similar to the average annual precipitation in northern New Jersey (−8‰) (Bowen and Wilkinson, 2002; Bowen, 2010).

Changes in the precipitation-to-evaporation ratio also contribute to the change in $\delta^{18}O$ values of lake water, as determined by residence time, catchment area, and lake size (Leng and Marshall, 2004). Present water O and H isotope compositions at Silver Lake fall near the global meteoric water line (Fig. 2). Other small open lakes in the region, similar in various characteristics to Silver Lake, have been shown to be controlled by the temperature effect and record regional temperature changes in the $\delta^{18}O$ of carbonate during the Holocene (Yu, 2007; Zhao et al., 2010a). Therefore, for this study, we assume that air temperature controls Silver Lake $\delta^{18}O$ and that a change in $\delta^{18}O$ most likely reflects a change in regional air temperature.

The fractionation of carbon isotopes between carbonate and water is relatively insensitive to temperature (Drummond et al., 1995). Therefore, $\delta^{13}C$ values of freshwater carbonates are largely controlled by the dissolved inorganic carbon (DIC) budget of the lake water. The DIC $\delta^{13}C$ values are controlled by multiple local environmental factors within the lake, including biological productivity, decomposition of organic matter, exchange rates between water and atmospheric CO_2, and lake water residence time and associated evaporation effects (Talbot, 1990). However, changes in $\delta^{13}C$ values can be related to temperature changes when productivity is a major factor affecting lake DIC. An increase in temperature causes increases in productivity, which preferentially consumes the lighter ^{12}C isotope from the available carbon supply and in turn causes the residual carbon reservoir (lake DIC) to become enriched with the heavier isotope ^{13}C. Therefore an increase (or decrease) in $\delta^{18}O$ values due to changes in temperature can correlate with increases (or decreases) in $\delta^{13}C$ values due to temperature related changes in productivity (Drummond et al., 1995).

4.2. Climate during the early Holocene in the Silver Lake region

Numerous multi-proxy studies in the southern New England region have demonstrated warming after the YD into the early Holocene with dry conditions persisting until 9 ka. Warming from 11 ka to 10 ka at White lake, NJ is recorded by an increase in carbonate $\delta^{18}O$ (Yu, 2007). Lake Grinnell, NJ showed high temperatures from 11 to 5.8 ka, dry conditions until 9 ka and wetter climate until 5.4 ka (Zhao et al., 2010b). Tannersville Bog in eastern PA indicates dry conditions from 10 to 9 ka (Cai and Yu, 2011). Spruce and Sutherland Ponds in southern NY document a reversal in temperate forest development from 11.1 to 10.2 ka (Maenza-Gmelch, 1997); however, afterward white pine and oak took over indicating warming for the rest of the early Holocene.

There is overall warming from 11 to 5 ka at Ballston Lake, NY. A pine-dominated forest was taken over by hemlock with increases in herbs and shrubs, indicating warmer mean temperatures and/or dryer conditions during a longer growing period from 10.7 to 5.3 ka (Toney et al., 2003). Seneca Lake, NY shows an overall increase in temperature from the end of the YD to 7 ka, with a cool period from 9 to 8 ka from lake sediment composition (Guiles Ellis et al., 2004). Dwyer et al. (1996) indicate dry climate from low lake levels at Owasco Lake, NY, while near-by, Cayuga Lake, NY shows an overall increase in carbonate deposition at depths from 10 to 8.5 ka suggesting rising temperatures with one small reversal at ~9.2 ka (Mullins, 1998).

At Crooked Pond, MA (Shuman et al., 2001) and Makepeace Cedar Swamp, MA (Newby et al., 2000), LOI and pollen indicate rapid warming after the YD (Shuman et al., 2004). A replacement of spruce to white pine from 11.6 to 9 ka, indicating warming, was recorded along with decreased moisture availability in southern New England (Davis, 1969; Newby et al., 2000; Shuman et al., 2001). Newby et al. (2000) and Shuman et al. (2001) showed low lake levels from 11 to 8 ka with open water turning to peat at 10.6 ka at Makepeace Cedar Swamp and a max low from 10.2 to 9 ka at Crooked Pond. Warm and wet conditions were observed from 9 to 5.4 ka by Shuman (2003) and Shuman et al. (2001) in southeastern Massachusetts. Blood Pond in southern Connecticut shows overall increasing temperature from 11.5 to 8 ka from δD, with multiple reversals in temperature from 10 to 8 ka (Hou et al., 2011).

4.3. Potential climate changes recorded at Silver Lake

Despite multiple and complex factors that contribute to changes in sediment composition we assume that sediment deposition is mostly controlled by lake levels, which reflect effective moisture conditions and temperature. This interpretation is supported by our surface sediment results. Also, the $\delta^{18}O$ and $\delta^{13}C$ values of carbonates are largely indicative of air temperature and primary productivity, respectively. At Silver Lake, the onset of the Holocene was marked by a sharp increase in carbonate content and decrease in silicate (Fig. 5). From 11.6 ka to 11 ka the increase in carbonate and decrease in silicate suggests a deepening of the lake and/or the reduction of mineral inputs. The increase in carbonate and relatively steady concentration of OM during this interval may be indicative of increased primary productivity and warming. Based on the relationships between lake depth and carbonate (and OM) content at Silver Lake, as documented from our modern process study (Fig. 4), the substantial decreases in carbonate in two brief periods between 11.1 and 11.3 ka suggest lower lake levels and drier conditions during the PBO.

Carbonate slightly increased in concentration and OM and silicate slightly decreased during the period from 11 ka to 9.6 ka. This may indicate a slight increase in lake level and in effective moisture or more likely an increase in carbonate precipitation and temperature with one reversal at 10.9 ka (labeled C1 in Fig. 5C). A warming at Silver Lake correlates with warming seen in the data for nearby Lake Grinnell (Zhao et al., 2010a), White Lake (Yu, 2007) and others (Webb et al., 1993; Shuman and Donnelly, 2006; Newby et al., 2009). The $\delta^{18}O$ and $\delta^{13}C$ values of both bulk carbonate and *Valvata* shells show a general increase during this time, also indicative of warming and an increase in primary productivity (Fig. 5). Therefore, the increase in both carbonate content and $\delta^{18}O$ and $\delta^{13}C$ values, from the start of the Holocene to ~9.6 ka, seems to reflect a steady increase in temperature and primary productivity.

After 9.6 ka, carbonate, OM, and silicate contents remain at about 90%, 5%, and 5%, respectively. However, multiple fluctuations in sediment composition are recorded (labeled C2–C7 in Fig. 5). Decreases in carbonate with increases in OM may be consistent with abrupt periods of lower lake level and drier climate or cooler temperatures with decreasing primary productivity. An increase in silicate during intervals of increased OM may indicate greater terrestrial erosion due to dry conditions. During this period, all isotopes start to behave erratically suggesting an increase in climate instability. The $\delta^{18}O$ values of *Valvata* shells begin to decrease indicating a possible decrease in summer temperatures. Carbonate $\delta^{18}O$ values show a slight decline, also suggesting a cooling. However, overall $\delta^{13}C$ values slightly increase until ~8.7 ka and then decline until 8.4 ka.

For the period of 8.4 ka to 8 ka, the largest change in sediment composition from marl to peat may indicate a significant decrease in lake level. Also, the 20% increase in silicate suggests more erosion, possibly caused by drier climates. This interval at Silver Lake signifies the 8.2-ka event and suggests a drier climate during this cold event (cf. Alley et al., 1997). Low sampling resolution of isotope analyses during the 8.2 ka event makes it difficult to speculate further; however, shell isotope compositions show a range of values larger than those for bulk sediment calcite suggesting changes in summer temperatures.

4.4. Outburst floods and climate response

Outburst floods, and related freshwater pulses from Lake Agassiz that reduced the Atlantic MOC and disrupted the THC, have been accepted as a probable cause of some climate events (i.e., YD, PBO, 8.2-ka event) (Fisher et al., 2002; Clarke et al., 2003; Teller and Leverington, 2004; Carlson et al., 2007). The outburst flood that caused the 8.2-ka event is thought to have occurred at 8.4 ka (Clarke et al., 2003) (Fig. 5A). This 200-year time lag, between the time of the flood and the mid-point of the 8.2 ka event, may suggest that a disruption of the THC and MOC takes time to noticeably affect climate (Clark et al., 2002). There appears to be a consistent ~200-year lag-time between other recognized flood events (Teller and Leverington, 2004) and the mid-point of the recorded sediment composition changes at Silver Lake (Fig. 5). By assigning floods to changes in sediment composition and plotting the flood year versus the mid-point of the oscillation in sediment composition (Fig. 6), we see a strong correlation ($R^2 = 0.998$), suggesting that the changes in sediment composition were possibly caused by the associated floods' impact on climate. However, some of the floods have not been directly dated and their timing is only estimated using the nearby dated intervals. Nonetheless, the ~200-year lag time appears to hold true at Silver Lake, but only for outburst floods that drained into the Hudson Bay, the North Atlantic, or the Arctic Ocean (Fig. 5). Floods that run through these three routes ultimately drain into the North Atlantic Ocean, thereby affecting the ocean circulation in a similar manner. The floods that drained into the Gulf of Mexico (floods D and E) did not show corresponding changes in sediment composition at Silver Lake. A freshwater flood into the Gulf of Mexico may not disrupt ocean circulation on a scale large enough to affect North Atlantic climate. Also, the relationship appears to hold true only for floods with a discharge rate ≥ 3000 km^3/yr. Therefore, outburst floods with discharge rates ≥ 3000 km^3/yr directly into the North Atlantic Ocean via the Hudson Strait or by way of the Arctic Ocean appear to have disrupted ocean circulation sufficiently to cause changes in climate conditions at Silver Lake.

A few discrepancies in these relationships are worthy of mention. Flood F drained into the North Atlantic and had a discharge rate greater than 3000 km^3/yr, but shows no associated change in sediment composition at Silver Lake. This lack of change in sediment composition is most likely due to the ~10 cm sampling resolution where a change may have been otherwise recorded. Flood M correlates with a two-step change in sediment between 9.21 and 9.05 ka. This may be due not only to Flood M but also to Floods L and K. The combination of the two small floods, occurring very near each other in time, could have caused the first change (C4a), beginning at 9.21 ka, whereas Flood M could have caused the second change (C4b), beginning at 9.11 ka. Flood Q is not correlated to a change in sediment composition with a definitive beginning and end, but rather, correlated to a steady increase in organic matter and decrease in carbonate, which is then truncated by the substantial change in sediment composition associated with the 8.2-ka event. The 8.2-ka event could have obscured the rapid change in sediment composition associated with flood Q.

It has been suggested that the final outburst flood actually occurred as a two-step drawdown, with the complete drainage 200 years after it started (Teller, 2004). This would explain the vast majority of studies that report a change in climate at 8.2 ka, instead

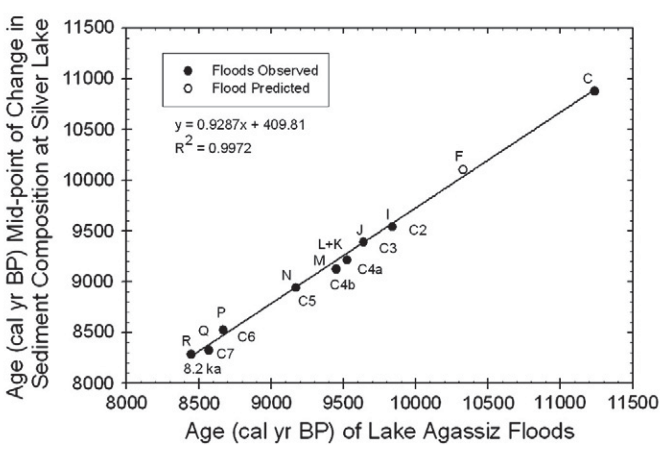

Fig. 6. Ages of floods from Lake Agassiz plotted against the mid-point ages of associated changes in carbonate at Silver Lake. Letters correspond to the floods and lithology changes as in Fig. 5. Flood F has no corresponding change in lithology at Silver Lake due possibly to a 10-cm-long sediment hiatus. However, the projected mid-point of where a change in lithology would occur during that interval is plotted and falls very close to the trend line of the other floods/lithology changes.

of 8.4 ka. If this is true then the 200 year lag relationship we have presented above is coincidental to our data. Also, models have shown that overturning in the North Atlantic responds immediately to a wide range of freshwater pulses (Wiersma et al., 2006). If we compare the changes in sediment composition at Silver Lake to the outburst floods with no lag time, then Floods C, L + K, M, N, O and Q may correlate to sediment changes PBO, C2, C3, C4a + b, C5, and C6, respectively (Figs. 5 and 7). The changes in sediment compositions, C2 and C5 occurred slightly before the recorded dates of floods L + K and O, which may be due to errors in age-depth models.

Changes in sediment composition at Silver Lake also correlate to other climate studies (Fig. 7). Fleitmann et al. (2008) recently showed evidence for widespread climate change caused by a meltwater pulse from Lake Agassiz (labeled N in Figs. 5 and 7). This flood, which released ~8100 km^3 in a single year, occurred between 9.25 and 9.09 ka (Teller and Leverington, 2004). The timing of changes documented with the various proxy data presented by Fleitmann et al. (2008) occurred during the period of 9.29 and 9.11 ka indicating that for this period global climate responded rapidly to the disruption of the North Atlantic THC and MOC. At Silver Lake a change in sediment is seen between 9.25 and 9.1 ka, potentially caused by the same flood, indicating the 9.2-ka event was recorded at Silver Lake.

Hou et al. (2011) took it one step further and linked other freshwater fluxes from Lake Agassiz to multiple climate change studies (Fig. 7). They propose cooling events at 10.6, 10.2, 9.5, 9.2, 8.8, and 8.4 ka. Not all of these climate changes correlate to specific outburst floods. This may be due to dating inaccuracies across various studies or it may indicate that the outburst floods were not the only culprits involved in these cooling periods. The variation in dates of climate changes and floods between the PBO and 8.2-ka event is most likely due to a combination of factors controlling global climate, with freshwater forcing and disruption of the THC, as only part of the whole.

5. Conclusions and implications

Silver Lake provides a record of changes in sediment composition during the transition from the late-glacial to early Holocene. High-resolution data from Silver Lake show evidence for the 8.2-ka event, the 9.2-ka event (Fleitmann et al., 2008; Hou et al., 2011), the PBO, the onset of Holocene, the YD and BOA. The multi-proxy data presented here demonstrate that both temperature and moisture varied significantly in the northern Mid-Atlantic region during the early Holocene. The beginning of the Holocene is recorded by increasing carbonate content, relatively stable OM, and decrease in silicate, indicating a rapid warming coming at the end of the YD. The PBO is distinguished by a decrease in carbonate and an increase in organic matter, indicating dry conditions during this known cooling period. Throughout the early Holocene, $\delta^{18}O$ and $\delta^{13}C$ data are consistent with overall warming whereas high carbonate and low OM are consistent with a stable deep lake or high primary productivity due to increasing temperatures. Multiple cool or dry events, indicated by decreases in carbonate and increases in organic matter, may correspond to glacial outburst floods from Lake Agassiz. At Silver Lake, the 8.2-ka event produced the largest change in sediment composition during the early Holocene.

It is critically important to understand the extent to which glacial meltwater affects the THC. As global temperatures rise and ice sheets melt, there will be increased fluxes of fresh glacial water into the oceans, and tracking the spatial and temporal changes in climate, due to past THC disruptions, will aid in predictions of future changes. Any large change in climate will undoubtedly affect future societies and their economies. Therefore, we should understand how, and to what extent, large-scale climate changes can affect highly populated areas (Kirby et al., 2001). Silver Lake, located in northern New Jersey, is near very densely populated areas such as New York, New Jersey, and eastern Pennsylvania. Time-series studies of Silver Lake, therefore, provide

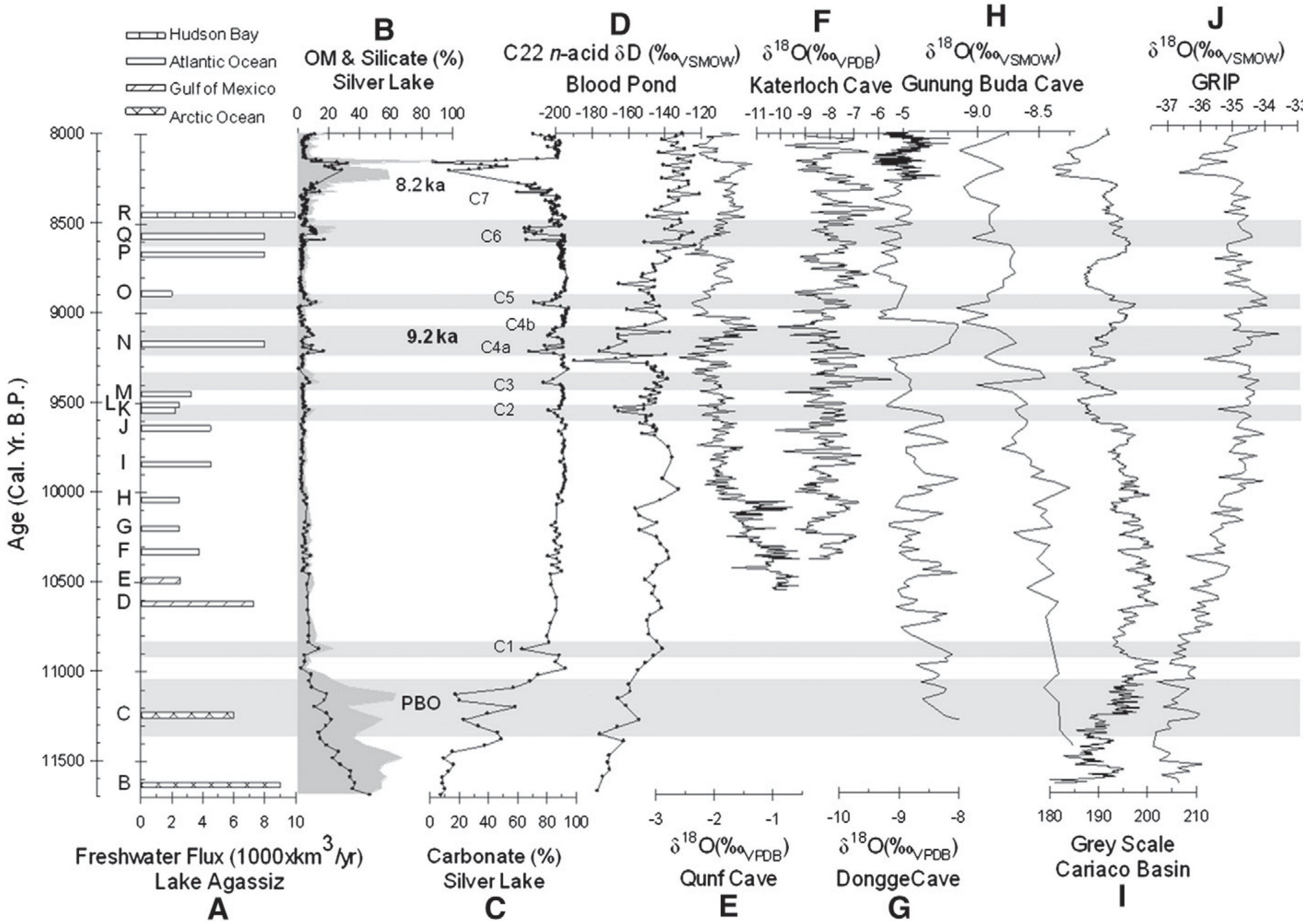

Fig. 7. Global correlations of early Holocene climate changes (A) Lake Agassiz outburst floods (modified from Teller and Leverington, 2004). The 8.2-ka flood was 163,000 km^3/yr in magnitude (beyond the graph). Key indicates the drainage routes of floods. (B) Percent OM (gray shading) and silicate (black line), (C) carbonate from Silver Lake. (D) δD Blood Pond (Hou et al., 2011); (E) δ^{18}O Qunf Cave (Fleitmann et al., 2008); (F) δ^{18}O Katerloch Cave (Boch et al., 2009); (G) δ^{18}O Dongge Cave (Dykoski et al., 2005); (H) δ^{18}O Gunnug Buda Cave (Partin et al., 2007); (I) Grey Scale Cariaco basin (Hughen et al., 1996); (J) δ^{18}O GRIP (Vinther et al., 2006).

useful insights into the sensitivity of THC to influxes of glacial melt water and the sensitivity of climate to perturbations of the THC. We observe that, in northern New Jersey, climate is susceptible to perturbation of the THC caused by flooding events as small as 3000 km^3/yr, suggesting that continued glacial melting may cause similar magnitude climate changes in this region in the future.

Acknowledgments

We thank Long Li and the EES 287 Spring 2006 class at Lehigh University for isotope analysis and coring assistance; Brendan Williams at the University of Maryland Stable Isotope Laboratory for stable isotope analysis; the W. M. Keck Carbon Cycle Accelerator Mass Spectrometry Laboratory at the University of California, Irvine, for radiocarbon dating analysis; the contributors of datasets used in this paper for data access; and two anonymous reviewers for helpful comments and suggestions. This research was supported by the American Chemical Society — Petroleum Research Fund and the National Science Foundation (both to Zicheng Yu).

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