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Holocene climatic change and the development of the lake-effect snowbelt in Michigan, USA

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ABSTRACT

Lake-effect snow is an important constraint on ecological and socio-economic systems near the North American Great Lakes. Little is known about the Holocene history of lake-effect snowbelts, and it is difficult to decipher how lake-effect snowfall abundance affected ecosystem development. We conducted oxygen-isotope analysis of calcite in lake-sediment cores from northern Lower Michigan to infer Holocene climatic variation and assess snowbelt development. The two lakes experience the same synoptic-scale climatic systems, but only one of them (Huffman Lake) receives a significant amount of lake-effect snow. A 177-cm difference in annual snowfall causes groundwater inflow at Huffman Lake to be ^{18}O-depleted by 2.3‰ relative to O'Brien Lake. To assess when the lake-effect snowbelt became established, we compared calcite-δ^{18}O profiles of the last 11,500 years from these two sites. The chronologies are based on accelerator-mass-spectrometry ^{14}C ages of 11 and 17 terrestrial-plant samples from Huffman and O'Brien lakes, respectively. The values of δ^{18}O are low at both sites from 11,500 to 9500 cal yr BP when the Laurentide Ice Sheet (LIS) exerted a dominant control over the regional climate and provided periodic pulses of meltwater to the Great Lakes basin. Carbonate δ^{18}O increases by 2.6‰ at O'Brien Lake and by 1.4‰ at Huffman Lake between 9500 and 7000 cal yr BP, suggesting a regional decline in the proportion of runoff derived from winter precipitation. The Great Lakes snowbelt probably developed between 9500 and 5500 cal yr BP as inferred from the progressive ^{18}O-depletion at Huffman Lake relative to O'Brien Lake, with the largest increase of lake-effect snow around 7000 cal yr BP. Lake-effect snow became possible at this time because of increasing contact between the Great Lakes and frigid arctic air. These changes resulted from enhanced westerly flow over the Great Lakes as the LIS collapsed, and from rapidly rising Great Lakes levels during the Nipissing Transgression. The δ^{18}O difference between Huffman and O'Brien lakes declines after 5500 cal yr BP, probably because of a northward shift of the polar vortex that brought increasing winter precipitation to the entire region. However, δ^{18}O remains depleted at Huffman Lake relative to O'Brien Lake because of the continued production of lake-effect snow.

1. Introduction

Snowfall abundance is a prominent factor constraining ecological and socio-economic systems in the heavily populated region surrounding the North American Great Lakes. Because of their enormous thermal capacity and continental climatic setting, the Great Lakes generate copious lake-effect snowfall along their leeward shoreline. As a result, a strong gradient in annual snowfall exists between a snowbelt located within ~80 km of the shoreline, and non-snowbelt areas further inland. Mean annual snowfall is up to three times more abundant in the snowbelt than in nearby non-snowbelt areas (Fig. 1; Norton and Bolsenga, 1993; Scott and Huff, 1996). Intense storms can imperil human life and property, and restrict communication and transportation systems (Schmidlin and Kosarik, 1999; Kunkel et al., 2002; Kristovich and Spinar, 2005). However, lake-effect snow can also provide important societal benefits, including positive impacts on water resources and economic activity. Where lake-effect snowfall is abundant, snowmelt dominates runoff and groundwater recharge, and is a major

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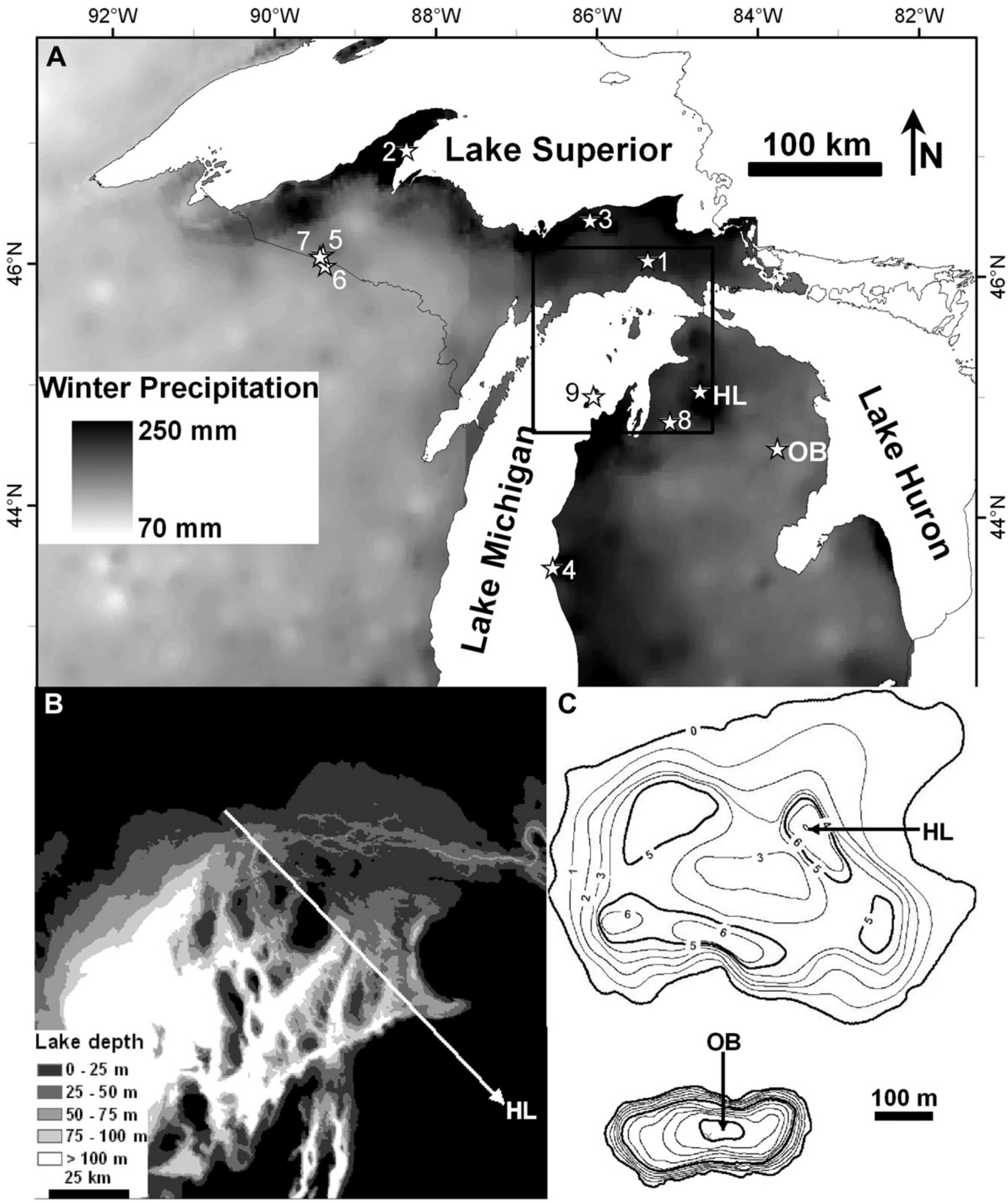

Fig. 1. (A) Map showing mean December–February precipitation (1961–1990 normals interpolated from National Weather Service Cooperative stations) in the upper Great Lakes region (Daly and Taylor, 2000). Black areas receive abundant lake-effect precipitation. Locations of study sites are represented with stars: HL – Huffman Lake, OB – O'Brien Lake. Numbers mark locations of additional lakes and peatlands discussed in text: 1 – Nelson Lake (Delcourt et al., 2002), 2 – Mud Lake (Booth et al., 2002), 3 – South Rhody Peatland (Booth et al., 2004), 4 – Silver Lake (Fisher et al., 2007), 5 – Glimmerglass Lake (Brugam et al., 2004), 6 – Lake O'Pines, 7 – Jay Lake (Ewing and Nater, 2002), 8 – Green Lake (Lawrenz, 1975; Davis, 2000), and 9-Tamarack Lake (Davis, 2000). The black box is presented in detail in (B), a simplified bathymetric map of northern Lake Michigan. Black areas denote the current shoreline including islands. Grey shades denote lake depth. The white arrow illustrates the fetch crossed by northwesterly air masses approaching Huffman Lake (C) Bathymetric maps of Huffman Lake (HL, left), and O'Brien Lake (OB, right). Dark lines are 5 m contours, and arrows denote coring locations.

constituent of streamflow (Schmidlin, 1993; Gat et al., 1994; Stottlemyer and Toczydlowski, 2006). The ecological impacts of lake-effect snowfall are also striking. For example, the distribution of forest community types near the Great Lakes corresponds to snowfall abundance. Mesic tree species (e.g., *Acer saccharum*) dominate forest communities on all soil types in the snowbelt, but are restricted to fine-textured soils outside the snowbelt (Henne et al., 2007). Winter soil temperatures and soil development follow a similar pattern. In the snowbelt where deep snowpacks typically persist from December through April, soils seldom freeze and soil development (i.e., podzolization) is more intense (Schaetzl, 2002; Schaetzl et al., 2005).

Several authors have suggested the abundance of lake-effect snow changed during the Holocene, and invoked snowbelt development as a factor contributing to vegetational (e.g., mesic tree expansion) and hydrological (e.g., rising water levels) transitions (e.g., Booth et al., 2002; Calcote, 2003; Brugam et al., 2004). However, such linkages remain difficult to validate because little is known about snowbelt history or the factors controlling snowbelt development. Recent studies suggested lake-effect snowfall increased during the late Holocene. In a review of pollen records from sites near the Great Lakes shoreline, Davis et al. (2000) attributed the expansion of mesic trees during the last 2000–4000 years to a major increase in lake-effect snow. However, because the analysis did not include comparison to adjacent non-snowbelt sites, or a vegetation-independent proxy for snowfall, it was difficult to isolate the impact of lake-effect snow from broader changes in regional climate. Delcourt et al. (2002) used calcite $\delta^{18}O$ as a proxy for lake-effect snow at Nelson Lake in Upper Michigan (Fig. 1) to infer an abrupt snowfall increase between 2950 and 2600 cal yr BP. Unfortunately, inconsistent calcite preservation made a full-Holocene $\delta^{18}O$ record impossible at this site, and isotopic data are available only from 6500 to 2600 cal yr BP. Thus it remains uncertain if snowbelts existed throughout or developed during the Holocene, or how changing lake-effect snowfall affected Holocene ecosystem dynamics.

In addition to assessing ecological impacts, an understanding of snowbelt history can help resolve the relationship between large-scale climatic forcings and snowfall abundance in the Great Lakes region. Lake-effect snow results from the interplay of forcing factors operating at local (e.g., boundary layer climatology), regional (e.g., Great Lakes water levels and ice cover), and continental (e.g., trajectory of cyclones) scales. Thus it is difficult to anticipate future changes on the basis of historical meteorological records alone, and the results of recent studies are contradictory. For example, an analysis of modern meteorological data attributed an observed lake-effect snowfall increase during recent decades to climatic warming (Burnett et al., 2003). In contrast, an approach combining global circulation model output with meteorological records anticipated a long-term decline in lake-effect snow in response to greenhouse forcing (Kunkel et al., 2002). Elucidating the Holocene history of lake-effect snowfall provides the only means to directly evaluate the impact of conditions that have not existed during the modern era (e.g., lower Great Lakes levels, Croley and Lewis, 2006) on lake-effect snow. Such data are necessary to develop regional models capable of accurately anticipating the impacts of climatic change in a region where the Great Lakes dominate regional climate.

In this paper we present lake-sediment isotopic records from two lakes in northern Lower Michigan. These two lakes were selected to provide a contrast in the abundance of lake-effect snow in order to assess the Holocene development of the lake-effect snowbelt. Huffman Lake (45°07′55″N, 84°46′42″W, 350 m asl) is located inside the snowbelt and receives abundant lake-effect snow. Nearby O'Brien Lake (44°38′29″N, 83°52′51″W, 274 m asl) is outside the snowbelt where lake-effect snowfall is negligible (Fig. 1). Because the oxygen-isotope composition of lake water is influenced by moisture sources, carbonate-$\delta^{18}O$ records from both sites should reflect past variations in synoptic-scale climate (Yu et al., 1997; Kirby et al., 2001). In contrast, only the record from the lake within the snowbelt (Huffman Lake) should be significantly impacted by lake-effect snow (Hu et al., 1997; Burnett et al., 2003). Thus we compare the $\delta^{18}O$ records from these two sites to infer the development of the lake-effect snowbelt during the Holocene. Our results provide the first full-Holocene $\delta^{18}O$ record from the lake-effect snowbelt. Comparison of adjacent snowbelt and non-snowbelt records enables placement of the numerous paleoecological and paleolimnological reconstructions from the region in the context of snowbelt development.

2. Study sites

Small inland lakes are abundant in the upper Great Lakes region. However, few provide a complete post-glacial record containing calcareous sediments for isotopic analysis. We chose Huffman Lake and O'Brien Lake for their locations relative to the Great Lakes snowbelt and for their calcite-rich sediment. Precipitation is similar between these sites during spring and summer, but Huffman Lake (snowbelt) receives more precipitation during fall and winter because of lake-effect snow and rain (Table 1). At Huffman Lake, annual snowfall averages 335 cm (1961–1990 interpolated normal of freshly fallen snow), whereas O'Brien Lake, 89 km away, receives an average of 161 cm of snow (Daly and Taylor, 2000). Lowland forests typical of groundwater-fed wetlands (i.e., *Thuja occidentalis* swamps) adjoin both lakes. Uplands near Huffman Lake support mesic northern hardwood forests dominated by sugar maple (*A. saccharum*). In contrast, pine-oak forest communities with abundant jack pine (*Pinus banksiana*), red pine (*Pinus resinosa*), and northern pin oak (*Quercus ellipsoidalis*) surround O'Brien Lake.

Both sites are hydrologically open basins on sandy glacial outwash. Huffman Lake occupies a narrow outwash channel, and O'Brien Lake is situated in a large outwash plain. Although neither lake has a surface inlet, groundwater seepage is apparent along the shoreline at both sites, and outlet streams flow continuously. Despite these similarities in hydrological setting, the two lakes differ greatly in surface areas and water volumes at present. Huffman Lake has a surface area of 50 Ha with three deep basins (maximum depths: 7 m) and a volume of 13.4×10^5 m^3. O'Brien Lake is much smaller (4 Ha) with a maximum depth of 10 m and a volume of 2.7×10^5 m^3. Huffman Lake remains well mixed during the summer due to a combination of shallow water and a large fetch. O'Brien Lake is thermally stratified during summer (7 °C temperature gradient in August 2004) and has an anoxic hypolimnion. Dissolved oxygen declines from 8.2 mg/l above 6 m of water to <1 mg/l below 8 m.

3. Materials and methods

To compare the hydrological impact of lake-effect snow, we estimated monthly runoff (i.e., precipitation in excess of evapotranspiration) for both sites using mean monthly temperature, mean monthly precipitation, and soil water-holding capacity as inputs to a water-balance model (Table 1; Willmott et al., 1985; Gavin and Hu, 2005). Monthly estimates were summed and presented using three-month seasons (e.g., December–February for winter). We obtained climatic data from interpolated weather-station normals of 1961–1990 (Daly and Taylor, 2000). Soil available water capacity was estimated from digital county-level soils maps (http://soildatamart.nrcs.usda.gov).

To help constrain paleoclimatic interpretations of sediment isotope data, we collected lake-surface water and groundwater for $\delta^{18}O$ and δD analyses at both sites in May and August 2004.

Table 1
Mean seasonal precipitation (Daly and Taylor, 2000), and water balance for Huffman Lake (HL) and O'Brien Lake (OB) (estimated from climatic data following: Willmott et al., 1985; Gavin and Hu, 2005). Runoff is amount generated during that period, not necessarily amount running off.

		Winter	Spring	Summer	Fall	Annual
Snow (mm)	HL	2342	558	0	483	3383
	OB	1113	350	0	148	1611
Precipitation (mm)	HL	169	194	245	283	891
	OB	121	170	236	204	731
PET (mm)	HL	0	107	350	115	572
	OB	0	104	350	116	570
AET (mm)	HL	0	107	333	115	555
	OB	0	104	321	116	541
Runoff (mm)	HL	169	88	0	78	336
	OB	117	74	0	0	105

Groundwater was sampled from a spring that emerges from the steep shoreline at O'Brien Lake and through a hand pump ~20 m from the shoreline at Huffman Lake. At O'Brien Lake, we also sampled water below the thermocline at a depth of 9 m. Hydrogen and oxygen isotopic composition of these waters was measured on a Finnigan Delta S gas-source isotope ratio mass spectrometer (IRMS).

Two sediment cores were collected from each lake. The cores from each site were stratigraphically overlapping and were visually correlated using lithologic transitions. At Huffman Lake, the upper 68 cm was captured with a polycarbonate piston corer that preserved the sediment–water interface. At O'Brien Lake, the upper 30 cm were obtained with a freeze corer. At both lakes, the remaining sediment was extracted using a square-rod piston corer (Wright et al., 1984).

Sediment samples from select levels were washed through a 150-μm sieve to concentrate terrestrial-plant macrofossils for AMS ^{14}C dating at the Lawrence Livermore National Laboratory, Center for Accelerator Mass Spectrometry (CAMS, Table 3). Sediment from Huffman Lake and the lower portion of the O'Brien Lake core contain few terrestrial macrofossils. Therefore, we concentrated Pinaceae pollen from 2-cm sections of one half of the core using an acid/base protocol and 20 μm sieve cloth (modified from Brown et al., 1989). Radiocarbon ages were converted to calibrated years using Calib 5.0.2 (http://radiocarbon.pa.qub.ac.uk/calib/). Calibrated 2σ age ranges were input to a mixed effects regression model to estimate an age-depth relationship (Heegaard et al., 2005) using the software package R 2.5 (Fig. 2).

We sampled the Huffman Lake core at 10-cm increments for loss-on-ignition (LOI) analysis, and for the analysis of oxygen and carbon isotopes. The O'Brien Lake core was sampled for the same analyses at ~200-year resolution on the basis of the age-depth model. The percentages of organic matter and carbonate in sediment were estimated from LOI at 550 °C and 1000 °C, respectively. The carbonate is almost exclusively calcite on the basis of X-ray-diffraction analysis of bulk sediment from each site at a 1000-year resolution with a Scintag theta–theta diffractometer. Bulk samples for isotopic analysis were sieved through an 80-μm mesh cloth to remove mollusk shells, ostracode shells, and large fragments of *Chara* encrustations. The residue was captured on Whatman microfiber filter paper, air-dried, and ground to powder. Because the percentage of carbonate was low and organic carbon high, samples from O'Brien Lake were treated with a 6% solution of sodium hypochlorite to remove organic carbon and concentrate calcite prior to the sieving step. Isotope analyses were performed on a Finnigan MAT 253 triple collector IRMS coupled to a Finnigan MAT Kiel IV carbonate-preparation device. The analytical precision of measurements was $\leq 0.1‰$ for $\delta^{18}O$ and 0.03‰ for $\delta^{13}C$.

To compare results from the two lakes, we smoothed the $\delta^{18}O$ profiles from each lake using a kernel smoother with an 800-year window in S-Plus 7.0. The O'Brien Lake $\delta^{18}O$ profile was resampled to the same temporal resolution as the Huffman Lake record using linear interpolation. We then subtracted Huffman Lake $\delta^{18}O$ from the resampled O'Brien Lake $\delta^{18}O$ to estimate Holocene variation in $\delta^{18}O$ difference between the sites.

4. Results

4.1. Sediment chronology

Most of the ^{14}C ages are in stratigraphic order at both of our study sites (Table 3). However, four of the 15 ^{14}C ages from Huffman Lake produced reversals in the age-depth model. The identities are unclear for the plant remains used to obtain three of these dates, and the fourth date was measured on concentrated pollen from a section of the core with a low abundance of large pollen types (e.g., *Pinus, Picea*). Thus, our final Huffman Lake age-depth model is grounded in 11 ^{14}C dates, whereas the O'Brien Lake age-depth model includes all 17 ^{14}C dates (Fig. 2). Palynological evidence from Huffman Lake and O'Brien Lake support our chronologies. Well-documented regional changes in the abundance of certain taxa (e.g., the *Tsuga* expansion about 6800 cal yr BP, and subsequent decline about 5400 cal yr BP) are synchronous between our two sites and between these two sites and others within the same region (Calcote, 2003; Henne, 2006). On the basis of our age-depth models, the 10-cm sampling resolution at Huffman Lake corresponds to a mean temporal resolution of 193 ± 76 years, and a similar sampling resolution was obtained at O'Brien Lake (190 ± 89 years).

4.2. Sediment lithology and composition

Calcite-rich sediments began accumulating over the basal deposit of coarse sand at Huffman Lake about 11,500 cal yr BP.

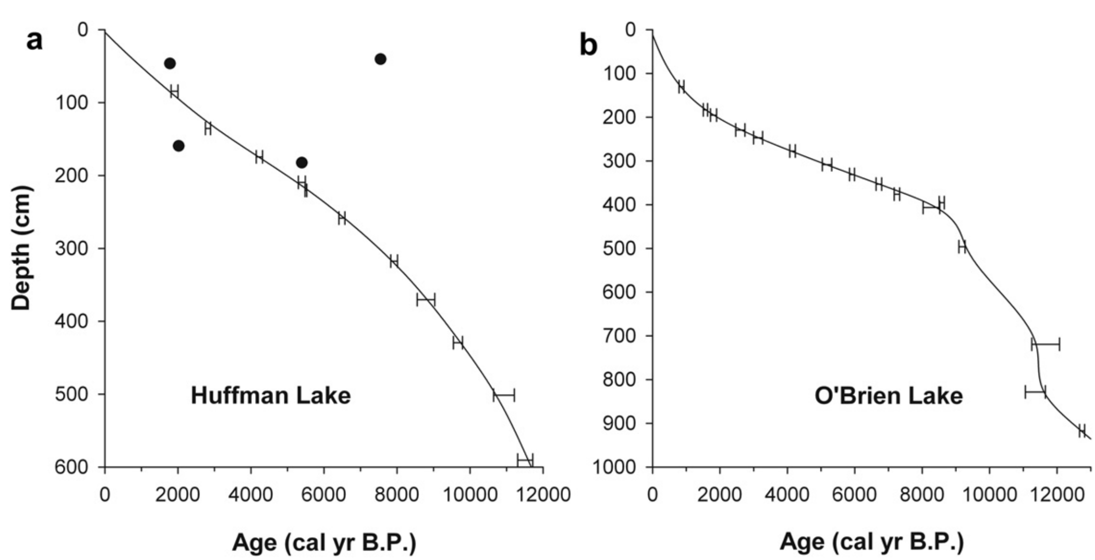

Fig. 2. Age-depth curves with the 2σ cal yr BP ranges of AMS ^{14}C dates for Huffman Lake (a) and O'Brien Lake (b). Four black circles represent dates excluded from the Huffman Lake age-depth model.

These sediments are typified by oxidized light grey marl with low organic matter content (i.e., <5%; Fig. 3). Sediment of ~6800–0 cal yr B.P. has higher organic content. A band of brownish marl with elevated organic matter (>10%) exists from about 6800 to 6300 cal yr BP. Sediment from 6300 to 5900 cal yr BP is light grey and low in organics. A second band of brown marl exists from 5900 to 5300 cal yr BP. The abundance of organic matter remains relatively high (>7%) and sediment darker in color until 2600 cal yr BP. Sediment of 2600–0 cal yr BP cm is low in organics (4–8%) and light grey in color.

The basal sediments of O'Brien Lake are primarily *Chara* encrustations. These sediments are overtopped by ~50 cm of peat consisting of aquatic moss and indicative of a rich fen (lithology not shown). All sediments younger than ~13,000 cal yr BP were black (i.e., anoxic) when extruded from the coring device. Organic matter content displays marked fluctuations throughout the record. Calcite abundance ranges from ~13 to 50% and generally increases upcore with a major rise after 8000 cal yr BP (Fig. 3). Fine-scale laminations are evident after about 13,000 cal yr BP, but cease above 11,700 cal yr BP. Gyttja with bands of *Chara* encrustations and moss exists between 11,700 and 10,500 cal yr BP. Sediment younger than 10,500 cal yr BP is dark gyttja. After about 9300 cal yr BP, the sedimentation rate declines (Fig. 2), and the abundance of terrestrial-plant macrofossils (e.g., *Pinus* needles and bud scales) increases. The gyttja of ~5300–0 cal yr BP is noticeably lighter in color, but otherwise is uniform brown with occasional lighter bands.

4.3. Water and sediment isotopic composition

The isotopic composition of groundwater inputs and lake water is consistently different between Huffman Lake and O'Brien Lake. Groundwater entering Huffman Lake has a δ^{18}O value 2.3‰ lower than that entering O'Brien Lake (Table 2). The ratio of *D* to ^{18}O in these waters also differs. Deuterium excess ($d = \delta D - 8\cdot\delta^{18}O$; Dansgaard, 1964) of groundwater is 13.7 at Huffman Lake and 11.2 at O'Brien Lake. Surface-water δ^{18}O at Huffman Lake was 1.0‰ lower than that at O'Brien Lake in May 2004. However, by August, Huffman Lake was 0.5‰ higher. During this period, surface-water δ^{18}O increased by 1.6‰ at Huffman Lake and by only 0.1‰ at O'Brien Lake. The δ^{18}O of water below the thermocline at O'Brien Lake was similar to that of groundwater in August 2004.

Table 2
Comparison of Huffman Lake and O'Brien Lake basin morphometries and lake-water isotopic composition. Isotopic data is from samples collected in 2004 and is relative to VSMOW.

Site	Huffman lake	O'Brien lake
Surface area	51.7 Ha	4.7 Ha
Volume	13.4×10^5 m^3	2.7×10^5 m^3
Surface area/volume	0.38 m^{-1}	0.17 m^{-1}
Groundwater δ^{18}O	−12.8‰	−10.5‰
Groundwater *d* excess	13.7	11.2
Surface δ^{18}O May	−10.9‰	−9.9‰
Surface δ^{18}O Aug	−9.3‰	−9.8‰
Bottom δ^{18}O Aug	NA	−10.3‰

The calcite-δ^{18}O records from Huffman and O'Brien lakes show similar general trends during the last 11,500 years (Fig. 3). From 11,500–9500 cal yr BP, δ^{18}O values are low, with minima of −11.3‰ and −12.0‰ at Huffman and O'Brien lakes, respectively. A transition to high δ^{18}O occurs at both sites by 8500 cal yr BP, with a peak of −9.5‰ at Huffman Lake and −9.2‰ at O'Brien Lake about 6900 cal yr BP. A major middle-Holocene decline brings δ^{18}O to low values at both sites by 4000 cal yr BP. δ^{18}O remains approximately −11.5‰ at Huffman Lake, and −10.9‰ at O'Brien Lake for the remainder of the record.

Despite these overall similarities, the δ^{18}O records display centennial- to millennial-scale differences between the two sites. Prior to 9500 cal BP, δ^{18}O increases gradually from −11.3 to −10.8‰ at Huffman Lake but fluctuates around −11.5‰ at O'Brien Lake (Fig. 3). The lowest δ^{18}O values of the Holocene occur during this interval at O'Brien Lake and after 5000 cal yr BP at Huffman Lake. A marked δ^{18}O increase occurs between 9500 and 8500 cal yr BP at both lakes (1.0‰ at Huffman Lake and 1.5‰ at O'Brien Lake). At O'Brien Lake, this positive trend continues until 6900 cal yr BP, and δ^{18}O remains high (i.e., above −10‰) until 6000 cal yr BP. After 6000 cal yr BP, δ^{18}O begins to decrease, reaching about −10.9‰ by 4000 cal yr BP and fluctuating around that value subsequently. At

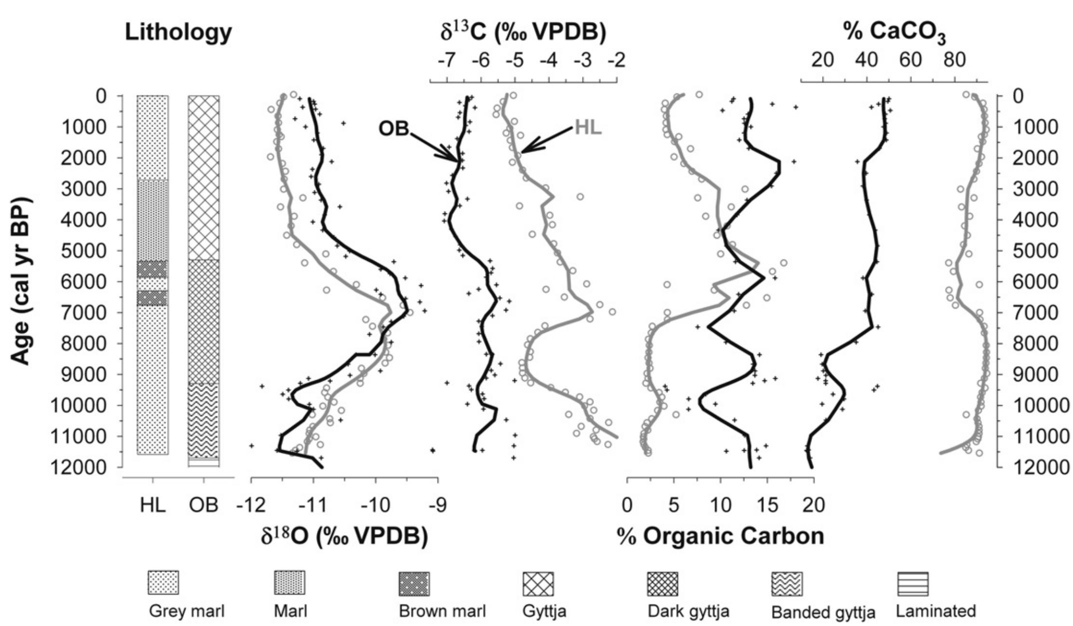

Fig. 3. Sediment lithology, oxygen and carbon isotope compositions, and organic and carbonate contents. Grey curves: Huffman Lake; black curves: O'Brien Lake. Raw data are represented by circles (Huffman Lake) and crosses (O'Brien Lake). Bold curves represent an 800-year moving average.

Table 3
Radiocarbon dates from Huffman and O'Brien Lakes. Dates excluded from age-depth model are indicated by. *

Depth (cm)	Material dated	CAMS number	^{14}C yr BP	Cal 2σ range
Huffman lake				
40–41*	Plant remains	119772	6680 ± 70	7439–7658
46–47*	Pollen	113553	1835 ± 40	1694–1873
84–85	Pollen	111177	1950 ± 45	1812–2001
135–136	Pollen	111178	2725 ± 45	2752–2894
159–160*	Plant remains	119773	2045 ± 35	1924–2116
174–175	Pollen	113554	3830 ± 35	4146–4319
182–183*	Plant remains	98373	4680 ± 35	5316–5475
209–210	Pollen	111179	4670 ± 60	5299–5489
220–221	Wood	108698	4870 ± 40	5483–5529
258–259	*Pinus* bud scales	113555	5715 ± 45	6409–6574
317–318	Pollen	113556	7095 ± 50	7830–8012
370–371	Pollen	119774	7979 ± 100	8553–9035
429–430	Charcoal	98372	8710 ± 40	9547–9792
501–502	Pollen	119775	9590 ± 120	10,645–11,215
590–591	*Picea* needle	98371	10,015 ± 45	11,304–11,717
O'Brien lake				
130–131	Leaf fragment	111174	950 ± 35	788–930
183–184	*Pinus* bud scales	108694	1660 ± 40	1507–1633
195–196	Leaf fragment	119776	1890 ± 40	1720–1901
229–230	*Pinus* bud scales	119777	2510 ± 35	2468–2739
246–248	*Pinus* bud scales	111175	2965 ± 40	2997–3264
277–278	*Pinus* bud scales	113551	3760 ± 35	4069–4237
308–309	*Pinus* bud scales	108695	4525 ± 45	5039–5316
330–332	*Pinus* bud scales	119778	5140 ± 50	5843–5991
352–354	*Pinus* bud scales	111176	5885 ± 40	6630–6797
375–377	*Pinus* bud scales	119779	6345 ± 35	7171–7330
395–396	*Pinus resinosa* needle	108696	7810 ± 40	8505–8656
406–407	Pollen	113552	7490 ± 130	8028–8522
496–497	*Pinus* bud scales	108697	8250 ± 40	9089–9328
496–497	*P. resinosa* needle	107896	8195 ± 40	9026–9273
719–720	*Picea* seed wing	119780	10090 ± 120	11,253–12,073
828–829	Charcoal	119781	9820 ± 90	11,060–11,653
916–918	Wood	107895	10,680 ± 40	12,669–12,822

Huffman Lake, the $\delta^{18}O$ increase does not continue after 8500 cal yr BP, although $\delta^{18}O$ values remain high between 8500 and 7000 cal yr BP. High $\delta^{18}O$ values about 7000 cal yr BP are followed by a marked decline until 4500 cal yr BP. Thus the transitions to the maximal $\delta^{18}O$ values of the middle Holocene and to the low $\delta^{18}O$ values of the late Holocene occur ~1000 years earlier at Huffman Lake. As a result, the $\delta^{18}O$ difference between the two sites decreases by 2.0‰ between 9500 and 5500 cal yr BP (Fig. 4). After 4500 cal yr BP, the $\delta^{18}O$ difference between the two sites remains about −0.6‰.

The highest calcite-$\delta^{13}C$ values at Huffman Lake (i.e., +2.0–−3.0‰) occur between 11,500 and 9500 cal yr BP (Fig. 3). A 2‰ decline in $\delta^{13}C$ between about 9800 and 8600 cal yr BP precedes an increase of similar magnitude between 8600 and 6900 cal yr BP. $\delta^{13}C$ generally decreases for the remainder of the record. At O'Brien Lake, $\delta^{13}C$ fluctuates around −5.8‰ before 6900 cal yr BP and declines by 1.3‰ between 6900 and 4000 cal yr BP. After 4000 cal yr BP, $\delta^{13}C$ generally increases upcore, but remains around −6.6‰. The Huffman Lake isotope records reveal a striking reversal in the relationship between $\delta^{18}O$ and $\delta^{13}C$, with negative covariance (Cov = −0.55) before 7000 cal yr BP but positive covariance after (Cov = 0.47; Fig. 5). At O'Brien Lake, $\delta^{13}C$ and $\delta^{18}O$ exhibit positive covariance after 7000 cal yr BP (Cov = 0.26) and no strong relationship before this time.

5. Discussion

5.1. Influence of lake-effect snow on water isotopic composition: rationale of interpreting down-core data

With the exception of snowfall abundance (i.e., lake-effect snow at Huffman Lake), the climatic determinants of $\delta^{18}O$ do not differ between Huffman and O'Brien lakes. For example, although the isotopic composition of precipitation varies with atmospheric temperature (Dansgaard, 1964), this effect should be nearly identical at the two sites because there is little difference in seasonal or annual temperature (Scott and Huff, 1996). Likewise, although the source (e.g., trajectory of cyclones) and seasonality of precipitation control the $\delta^{18}O$ of input waters (Machavaram and Krishnamurthy, 1994; Kirby et al., 2001; Burnett et al., 2004), Huffman Lake and O'Brien Lake experience the same synoptic-scale weather systems (Scott and Huff, 1996). Therefore, the disparity in lake-water isotopic composition between these nearby sites must have resulted from mesoscale variation in precipitation source and abundance (i.e., snowbelt vs. non-snowbelt), basin-specific processes (e.g., differences in exposure to evaporation), or a combination thereof.

The threefold difference in annual snowfall can best explain the 2.3‰ difference in the $\delta^{18}O$ of groundwater entering Huffman and O'Brien lakes. Lake-effect snow can affect groundwater $\delta^{18}O$ in a number of ways. First, lake-effect snow derives from vapor transported across the continent in frigid air masses, and recycled from the Great Lakes. These vapor sources are depleted in ^{18}O relative to moisture carried by more southerly air masses from the Gulf of Mexico, the principal source of precipitation in the upper Great Lakes region (Machavaram and Krishnamurthy, 1994; Isard et al., 2000). Furthermore, snow in general has lower $\delta^{18}O$ than rain because the abundance of ^{18}O in precipitation declines with air temperature (Dansgaard, 1964). As a result, the $\delta^{18}O$ of precipitation and runoff is lower in snowbelt areas than outside the snowbelt (Gat et al., 1994; Burnett et al., 2004). High *D* excess in groundwater at Huffman Lake provides further evidence for the impact of lake-effect snow. Deuterium excess of snowbelt precipitation and runoff is higher because vapor from the Great Lakes is enriched in *D* relative to vapor from the Gulf of Mexico (Gat et al., 1994; Machavaram and Krishnamurthy, 1995).

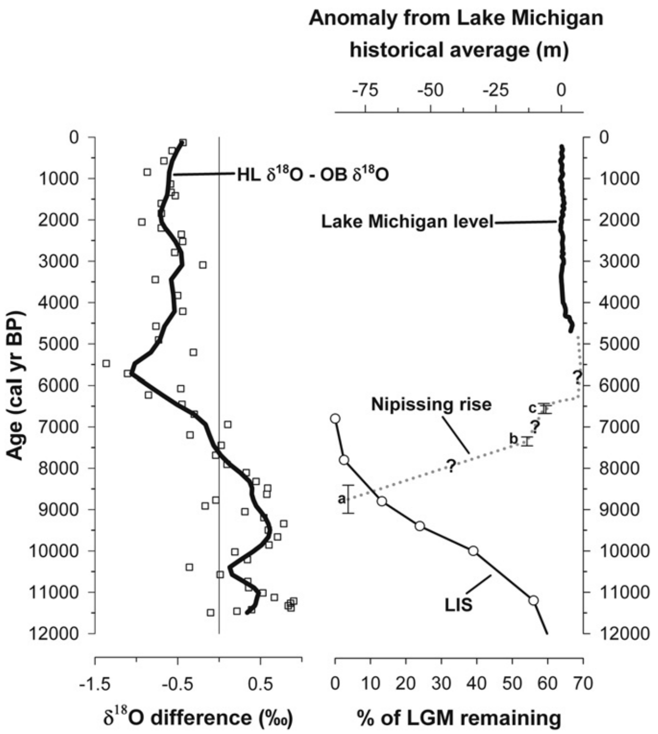

Fig. 4. $\delta^{18}O$ difference between Huffman Lake and O'Brien Lake, changes in Lake Michigan level during the Holocene, and percentage of LIS volume remaining after the last glacial maximum (redrawn from estimates in Licciardi et al., 1998). Rising waters of the Nipissing Transgression are estimated from (a) Lewis et al., 2007b; (b) Hunter et al., 2006; and (c) Fisher et al., 2007. Lake Michigan level after 4700 cal yr BP, adapted from Baedke et al. (2004).

The isotopic signal of lake-effect snow is retained in lake water through groundwater inflow at Huffman Lake. In some temperate continental regions, ^{18}O-depleted precipitation from the cool season supplies most groundwater recharge (Abbott et al., 2000; O'Driscoll et al., 2005). Snow in particular contributes disproportionately to groundwater because evapotranspiration is negligible in winter and snowpacks promote soil conditions conducive to groundwater recharge (Maule et al., 1994). At both Huffman and O'Brien lakes, surplus moisture from winter precipitation runs off to groundwater in spring, water deficits (i.e., PET > AET) develop during summer, and soils return to field capacity as PET declines during fall. However, because of lake-effect snow and rain, annual runoff is three times more abundant near Huffman Lake than O'Brien Lake (Table 1). Most of this extra precipitation enters the soil during winter and spring thaws because soils in the snowbelt (e.g., at Huffman Lake) seldom freeze (Isard and Schaetzl, 1998; Stottlemyer and Toczydlowski, 1999; Schaetzl, 2002). Within the snowbelt, abundant fall precipitation brings soils to field capacity by October, and November precipitation also runs off to groundwater. In contrast, outside the snowbelt at O'Brien Lake, soil-water deficits remain until December. Although heavy summer rain falling on the sandy soils near our study sites may also contribute to groundwater, summer recharge is diluted by the much greater abundance of low-$\delta^{18}O$ winter runoff, particularly in the snowbelt (Table 1).

The isotopic composition of lake water is preserved in calcite precipitated from Huffman and O'Brien lakes. Lake-water $\delta^{18}O$ reflects not only the distinct isotopic values of input groundwater, but also basin-specific differences in evaporation. Under present conditions, evaporation has a larger effect on the isotopic composition of water at Huffman Lake because it occupies a large, shallow basin with a high surface area to volume ratio (Fig. 2). Consequently, in late summer, Huffman Lake has higher $\delta^{18}O$ than O'Brien Lake. However, turnover time at Huffman Lake is short enough that by spring, the influx of ^{18}O-depleted snowmelt and groundwater results in lower lake-water $\delta^{18}O$ at Huffman Lake. Thus, because authigenic precipitation of carbonates by phytoplankton typically peaks during spring and early summer in temperate hardwater lakes (Thompson et al., 1997), the short lake-water residence times of these hydrologically open basins limits the importance of evaporative enrichment for the isotopic composition of sedimentary calcite.

5.2. Dominant effects of the Laurentide ice on the early-Holocene climate

Comparison of the Huffman Lake and O'Brien Lake isotopic records is first possible after 11,500 cal yr BP when sedimentation began at Huffman Lake. Low $\delta^{18}O$ values at both lakes around this time are consistent with evidence for a continental climate with cold winters in the Great Lakes region. The Laurentide Ice Sheet (LIS) was near the southern edge of the Lake Superior basin at 11,500 cal yr BP, ~200 km north of Huffman and O'Brien lakes (Lowell et al., 1999). High-pressure anticyclones that formed over the ice sheet dominated regional climate by blocking meridional transport of subtropical heat and moisture (COHMAP members, 1988; Kutzbach et al., 1998; Krist and Schaetzl, 2001). Thus air masses bearing high-$\delta^{18}O$ moisture from the Gulf of Mexico were less common than at present. Cool annual temperatures probably also contributed to the low $\delta^{18}O$ values. Pollen transfer functions indicate a highly seasonal climate, with a mean annual temperature 4–5 °C and an annual precipitation ~200 mm below the modern averages (Davis et al., 2000). Although $\delta^{18}O$ increased after 11,500 at Huffman and O'Brien lakes, the increases were small prior to 9500 cal yr BP, probably because the LIS continued to dominate atmospheric circulation and the sources of precipitation near the Great Lakes during the early Holocene.

Low water levels in the ancestral Great Lakes probably prevented snowbelt development during the early Holocene. Because of the combined influence of isostatically depressed outlets and the cool/dry regional climate, the ancestral Great Lakes may have been >100 m below the 20th-century average at the onset of our isotopic records (Colman et al., 1994; Lewis et al., 1994). The lakes remained on average far below modern levels until ~7000 cal yr BP; at the final lowstand estimated between 9000 and 8300 cal yr BP, the lakes may have been >75 m lower (Fig. 4; Lewis et al., 2007a,b). At such low levels, the Great Lakes held less thermal energy and provided little fetch upwind from our study area to moderate the continental climate and produce lake-effect snowfall. In particular, there was probably little open water in the relatively shallow northern sections of Lake Michigan (Fig. 1b; Clark et al., 2007). Thus the most productive wind directions for the development of lake-effect storms (i.e., NW – NNW; Braham, 1983; Niziol et al., 1995; Liu and Moore, 2004) would not have crossed open water upwind from Huffman Lake. However, lake-effect snowfall may have been possible to the lee of deeper sub-basins. At present, northwesterly air masses approaching Huffman Lake cross approximately 83 km of Lake Michigan. If Lake Michigan was 25, 50, or 75 m lower (i.e., within the range of change since the last lowstand; Fig. 4; Fisher et al., 2007; Lewis et al., 2007b), the fetch would be reduced 28%, 54%, and 100% respectively (Fig. 1b). The fetch crossed by more northerly air masses would be similarly reduced. Although isostatic rebound also constrains the position of Holocene shorelines, Lake Michigan bathymetry alone provides a reasonable approximation of the impact of changing lake levels on fetch (Clark et al., 2007).

The major $\delta^{18}O$ increases at Huffman and O'Brien lakes after 9500 cal yr BP coincide with the rapid retreat of the LIS (Figs. 3 and

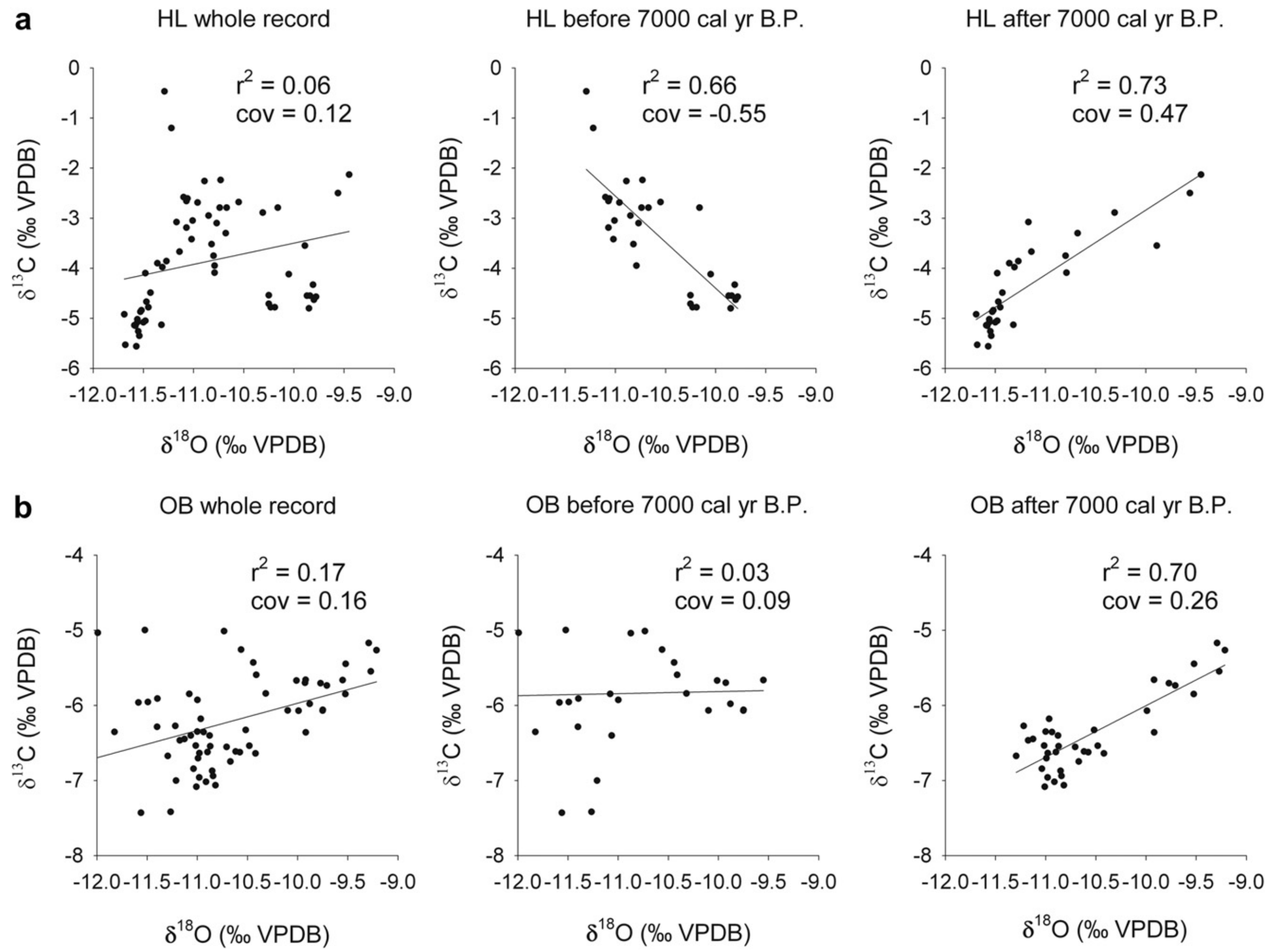

Fig. 5. Cross plots comparing $\delta^{18}O$ versus $\delta^{13}C$ of individual samples from (a) Huffman Lake (HL) and (b) O'Brien Lake (OB). Figures on left show all measurements; middle, before 7000 cal yr BP; and right, after 7000 cal yr BP. Covariance reverses at Huffman Lake and increase at O'Brien Lake after 7000 cal yr BP.

4; Dyke, 2004; Carlson et al., 2008). Recession of the LIS probably allowed northward penetration of warm air masses bearing ^{18}O-enriched moisture from the Gulf of Mexico during summer (Krishnamurthy et al., 1995; Shuman et al., 2002). Thus increasing $\delta^{18}O$ at our sites probably reflects increasing summer temperatures and a growing proportion of annual precipitation falling during summer. This interpretation is supported by evidence for low winter but increasing summer precipitation from other sites in the upper Midwest. For example, falling water tables in small lakes and wetlands in Upper Michigan after 9500 cal yr BP (Brugam et al., 1998; Booth et al., 2002, 2004) indicate reduced groundwater recharge from winter precipitation (Almendinger, 1990; Filby et al., 2002). In areas west of the Great Lakes, an eastward expansion of prairie vegetation around 9000 cal yr BP (e.g., Nelson et al., 2006) suggests declining winter soil–water recharge. This expansion is characterized by the increasing pollen abundance of *Ambrosia* sp. (ragweed), which requires ample growing-season precipitation (Grimm, 2001).

5.3. Middle- and late-Holocene climatic variation across the region

The $\delta^{18}O$ records from Huffman and O'Brien lakes continue to display overall similarities in temporal trends through the middle and late Holocene (Fig. 3). These broad similarities indicate regional climatic changes affected areas both within and outside today's snowbelt. High $\delta^{18}O$ values at Huffman and O'Brien lakes between about 8500 and 7000 cal yr BP probably resulted from a regional transition to warmer and drier conditions evident at both sites after 9500 cal yr BP. It is likely that the lowest regional abundance of winter precipitation during the Holocene occurred during this interval. Coincident $\delta^{18}O$ maxima at Huffman and O'Brien lakes may indicate the driest regional conditions occurred about 7000 cal yr BP. The combined isotopic records also provide evidence for a regional shift to moister winters and perhaps cooler conditions between 7000 and 4000 cal yr BP. Although the timing and cause of the shift differ in relation to mesoscale climate (i.e., snowbelt development, see following section), the $\delta^{18}O$ decline of 2.0‰ at Huffman Lake and 1.8‰ at O'Brien Lake suggests that both sites experienced a major increase in winter precipitation by 4000 cal yr BP.

Our interpretations of climatic shifts during the middle- and late-Holocene are consistent with evidence from other sites across the region. High and increasing $\delta^{18}O$ values at our study sites between 8500 and 7000 cal yr BP coincide with low water levels at small lakes and wetlands in both snowbelt and non-snowbelt areas. At Mud Lake in the Lake Superior snowbelt, dry conditions after 8600 cal yr BP caused a shift from a lake to a wetland environment that lasted until about 6600 cal yr BP (Booth et al., 2002). Similar transitions occur at nearby South Rhody Peatland and in the Lake Michigan snowbelt at Silver Lake (Booth et al., 2004; Fisher et al., 2007). Lake-level reconstructions outside the snowbelt in Upper Michigan from Glimmerglass Lake, Crooked Lake, and Lake O'Pines also indicate that the lowest water levels of the Holocene occurred during this interval (Brugam et al., 1998, 2004; Ewing and Nater, 2002). Furthermore, pollen-inferred vegetational changes support our δ^{18}O-based inference of increasing moisture after 7000 cal yr BP. For example, increasing moisture availability favored the rapid expansion of hemlock (*Tsuga canadensis*) near Huffman and O'Brien lakes (Henne, 2006), and across the entire region between 7000 and 5500 cal yr BP when δ^{18}O declines (Davis et al., 1986; Brugam and Johnson, 1997; Booth et al., 2002). Mesic forests were well established across the region by 4000 years cal yr BP when low δ^{18}O at Huffman Lake and O'Brien Lake indicate winter precipitation was abundant, and water levels in small lakes and wetlands reached their current levels (Davis et al., 2000; Jackson and Booth, 2002; Booth et al., 2004; Brugam et al., 2004).

Although carbonate-δ^{13}C values are affected by numerous climatic and non-climatic factors, the δ^{13}C record from Huffman Lake is congruent with our δ^{18}O-based climatic inferences. Climatic changes that lengthen the growing season or favor primary productivity increase δ^{13}C, whereas a shorter growing season or less productivity lowers δ^{13}C (Drummond et al., 1995; Leng and Marshall, 2004). Given this interpretation, rising δ^{13}C at Huffman Lake between 9500 and 7000 cal yr BP reflects the regional trend toward warmer and drier conditions with shorter winters. Likewise, declining δ^{13}C and the covariance between δ^{13}C and δ^{18}O after 7000 cal yr BP are consistent with the onset of a shift to cooler and moister conditions with a shorter growing season. Declining δ^{13}C at O'Brien Lake after 7000 cal yr BP also supports this interpretation. However, we are unable to explain the early-Holocene δ^{13}C changes at Huffman Lake with the same mechanisms.

The isotope records from Huffman and O'Brien lakes, in conjunction with other paleorecords from the region, suggest synoptic-scale climatic changes across the entire region during the middle- and late-Holocene. Dry regional conditions and high δ^{18}O between 8500 and 7000 cal yr BP probably mean the frequency of incursions by southerly air masses during summer continued to increase. Because the modal position of the winter vortex remained south of the upper Great Lakes region at this time, dry arctic air dominated the winter climate (Kirby et al., 2002). After 7000 cal yr BP the trend toward increasing incursions of southerly air masses reversed, and more moderate summer conditions with lower temperatures and reduced evaporation began to develop (Davis et al., 2000; Calcote, 2003). By 4000 cal yr BP, northward contraction of the modal position of the winter vortex (Kirby et al., 2002) allowed winter precipitation to approach its modern abundance. When the winter vortex has a more northerly position, winter storms originating to the southwest of the study area cross the Great Lakes region more frequently. Such storms entrain moist air masses from the Gulf of Mexico and produce copious snowfall and winter runoff in snowbelt and non-snowbelt areas alike (Rodionov, 1994; Isard et al., 2000).

5.4. Initiation of the lake-effect snowbelt

Lake-effect snowstorms require (1) a favorable wind direction, (2) a broad expanse of open water, and (3) a strong contrast between atmospheric and Great Lakes water temperatures. Glacially-dominated atmospheric circulation patterns and low Great Lakes water levels made the coincidence of such conditions near Huffman Lake unlikely during the early Holocene. The $\sim 1.5‰$ decline in the δ^{18}O difference between Huffman and O'Brien lakes between 9500 and 5500 cal yr BP suggests that snowbelts began to develop within this period. However, the largest increase in lake-effect snowfall probably occurred after 7000 cal yr BP when Huffman Lake δ^{18}O shifts to low late-Holocene values. The crossover of the two δ^{18}O records during this interval, with Huffman Lake becoming more isotopically depleted, is also consistent with the hypothesis that snowbelts became established around this time (Fig. 4).

The increase of lake-effect snow was made possible by two key events between 9500 and 7000 cal yr BP: the disintegration of the LIS and contemporaneous rise of the Nipissing Great Lakes. Atmospheric circulation patterns became increasingly favorable for snowbelts with the disintegration of the LIS (Dyke, 2004; Carlson et al., 2008). As the ice sheet dissipated, the glacial anticyclone ceased, and westerly flow over the Great Lakes strengthened (COHMAP members, 1988; Shuman et al., 2002). The modal position of the polar-front jet stream south of our study area also favored lake-effect snowfall (Kirby et al., 2002). Today, under such conditions (i.e., an amplified wave-train structure with a deep trough over eastern North America), the jet stream steers cyclones originating northwest of the study region over the Great Lakes (Isard et al., 2000). These storms carry little precipitation to the region, but their passage creates synoptic-scale conditions conducive to lake-effect snow (Rodionov, 1994; Liu and Moore, 2004; Notaro et al., 2006).

The dramatic rise in Great Lakes levels of the Nipissing Transgression was also key to snowbelt development. By 8300 cal yr BP, water levels in the Upper Great Lakes were rising due to a combination of isostatic rebound of the North Bay outlet and climatic factors (Lewis et al., 2007a, 2008). The most rapid lake-level rise probably occurred after 7000 cal yr BP when regional moisture availability began to increase (Booth et al., 2002). Evidence preserved in the sediments and near-shore basins of the three Upper Great Lakes (i.e., Michigan, Huron, and Superior) is indicative of the expanding fetch upwind of Huffman Lake because the Upper Great Lakes became confluent as lake levels rose (Larson and Schaetzl, 2001). About 7400 cal yr BP, Lake Huron flooded a forest 13 m below the mean modern lake level (Hunter et al., 2006). Lake Superior and Lake Michigan flooded near-shore basins close to the mean modern level of Lake Michigan about 6600 cal yr BP, and the lakes reached the Nipissing high, above the modern elevation by 6300 cal yr BP (Fig. 4; Booth et al., 2002; Fisher et al., 2007). Summer water temperatures in Lake Michigan were warm as the Nipissing Great Lakes rose (Odegaard et al., 2003), and temperature-transfer functions from fossil pollen indicate high seasonality (Davis et al., 2000; Calcote, 2003). Thus by 7000 cal yr BP, it is likely that environmental conditions suitable for snowbelts had become established in the upper Great Lakes region.

The $2.0‰$ δ^{18}O decrease at Huffman Lake between 7000 and 4500 cal yr BP coincides with the Great Lakes attaining their modern elevation. The contemporaneous hemlock expansion at Huffman Lake and O'Brien Lake, and across the upper Great Lakes region, indicates that the rapid rise in Great Lakes water levels at this time was probably driven by increasing moisture across the entire region (Davis et al., 1986; Brugam and Johnson, 1997; Booth et al., 2002; Henne, 2006). We inferred that a major increase in lake-effect snowfall resulted from a positive feedback to this moisture increase. That is, increasing runoff drove the rise in Great Lakes water levels, which in turn provided a broader fetch to feed lake-effect storms. Interestingly, this time period also corresponds with a major pollen-inferred January temperature increase at nearby Green and Tamarack lakes (i.e., 5 °C between $\sim$7000 and 5000 cal yr BP; Davis et al., 2000). Such a temperature increase near the Great Lakes shoreline is consistent with a rise in Great Lakes levels. The expanding fetch not only favored production of low δ^{18}O lake-effect snow, but also increasingly moderated winter temperatures downwind.

Changes consistent with increasing snowfall are evident at other snowbelt sites during this interval. At Mud Lake in the Lake Superior snowbelt (Fig. 1), the water table rose rapidly after 7200 cal yr BP, with an end to middle-Holocene dry conditions inferred by 6600 cal yr BP (Booth et al., 2002). Silver Lake in the Lake Michigan snowbelt also transitioned from a wetland to a lake at this time

(Fisher et al., 2007). Because these lakes are situated near the Great Lakes shoreline, the hydrological changes may be due in part to backfilling by the rising Great Lakes. However, increasing moisture around 6800 cal yr BP also occurred at South Rhody peatland further inland in the Lake Superior snowbelt, suggesting that increasing lake-effect snowfall was also important (Booth et al., 2004). Evidence of increasing winter precipitation outside the snowbelt is more subtle until about 5500 cal yr BP. For example, lake level remained low between 8500 and 5500 cal yr BP at Glimmerglass Lake in Upper Michigan (Fig. 1), but rose 3 m between 5500 and 4000 cal yr BP (Brugam et al., 2004). Rising water levels and intensification of podzolization after 6000 cal yr BP at nearby Lake O'Pines and Jay Lake also suggest a regional increase in winter precipitation (Schaetzl and Isard, 1996; Ewing and Nater, 2002; Schaetzl, 2002). These changes coincide with the shift to late-Holocene δ^{18}O values at O'Brien Lake. Thus, decreasing Huffman Lake δ^{18}O after about 5500 cal yr BP may relate to a region-wide increase in winter precipitation, not lake-effect snow.

A major increase in lake-effect snowfall may have also contributed to the δ^{13}C decline after 7000 cal yr BP at Huffman Lake. Lake-effect snowpacks facilitate soil development and the leaching of dissolved C (Schaetzl, 2002; Stottlemyer and Toczydlowski, 2006). Because terrestrial C has lower δ^{13}C than aquatic C, an influx of dissolved terrestrial C may have lowered Huffman Lake δ^{13}C as spodosols developed in the catchment (Leng and Marshall, 2004). Declining δ^{13}C also coincides with a sharp transition to brown sediments with high organic content about 6800 cal yr BP (Fig. 3), which may be of terrestrial origin.

Because we rely on direct temporal comparison of δ^{18}O profiles, chronological errors could compromise our interpretation of snowbelt establishment. In particular, the Huffman Lake chronology includes several AMS ^{14}C dates of pollen and charcoal, substrates that could conceivably be redeposited, and therefore produce ^{14}C dates that are older than the surrounding sediment. Such dates could potentially create the appearance of an earlier transition from high to low δ^{18}O values at Huffman Lake. However, our age-depth models are well constrained during the snowbelt-establishment period. Six AMS ^{14}C dates fall between 9500 and 5500 cal yr BP at Huffman Lake (Table 3), and this section of the age-depth model is nearly linear, suggesting that all dates are reasonable when they are considered together. Similarly, the O'Brien Lake chronology is well constrained with eight dates during the snowbelt development interval. Our use of pollen extracts for dating is supported by the calibrated age ranges of two pollen dates that overlap with the age ranges of adjoining terrestrial macrofossils (Table 3). Equally important, our chronologies are validated by the temporal correspondence of the well-described *Tsuga* expansion between Huffman and O'Brien lakes as well as between these lakes and other paleorecords from the Upper Great Lakes region (Henne, 2006).

Divergent δ^{18}O trends at Huffman and O'Brien lakes can alternatively be ascribed to basin-specific differences in isotopic responses to climatic change. The declining δ^{18}O difference between Huffman and O'Brien lakes from 9500 to 5500 cal yr BP may have resulted from reduced evaporation related to regional climatic changes, not lake-effect snow. However, if evaporative enrichment constrains the δ^{18}O difference, the difference should fluctuate with regional climate, with increases during periods of increasing aridity and temperature. Instead, between 9500 and 7000 cal yr BP, when water levels fell in small lakes and wetlands, and regional temperatures rose (Davis et al., 2000; Booth et al., 2002; Ewing and Nater, 2002; Calcote, 2003), the δ^{18}O difference between Huffman and O'Brien lakes decreases. Huffman Lake δ^{18}O even crosses over to become more negative than O'Brien Lake δ^{18}O during this interval. Conversely, cooler and moister conditions should cause the δ^{18}O difference to decline. However, the largest regional moisture increase occurs after 5500 cal yr BP, when the δ^{18}O difference stops decreasing. Although additional δ^{18}O records from paired sites within and outside the snowbelt are ultimately needed to verify our interpretations, the impact of snowbelt development on δ^{18}O appears to far outweigh the importance of changes in evaporative enrichment related to regional climate.

Overall our results are inconsistent with a hypothesized increase in lake-effect snowfall during the last 2000–4000 years (Davis et al., 2000; Delcourt et al., 2002). Huffman Lake δ^{18}O remains similar to the present during this interval (i.e., $<0.5‰$ variation since 4500 cal yr BP; Fig. 3), and the δ^{18}O difference between Huffman Lake and O'Brien Lake remains $\sim -0.6‰$ (Fig. 4). Furthermore, it is likely that environmental conditions were favorable for lake-effect snowfall throughout the late Holocene. The mean level of Lake Michigan has not fallen below 2 m of the 20th-century average since at least 4700 cal yr BP, and probably since the Nipissing highstand (Fig. 4; Baedke et al., 2004; Fisher et al., 2007). Pollen-inferred January temperature changes little after 4500 cal yr BP, whereas July temperature declines slightly (i.e., <1 °C; Davis et al., 2000; Calcote, 2003). Thus, during the late Holocene, as at present, frigid air masses crossed a broad expanse of open water that held residual summer warmth, and lake-effect storms probably developed. We suggest that observed increases in mesic pollen types at snowbelt lakes relate instead to the interaction of cooling regional summer temperatures with a snowfall gradient in existence since the middle Holocene. Testing this hypothesis will require comparisons of pollen profiles from proximal study sites within and outside of the lake-effect snowbelt.

6. Conclusions

Our data demonstrate that lake-effect snowbelts are not a recent phenomenon; nor have they been present since regional deglaciation. Snowbelts developed in the Upper Great Lakes region between 9500 and 5500 cal yr BP, and persisted through the remainder of the Holocene. Prior to this time, glacially-dominated circulation patterns and low water levels in the Great Lakes basin made abundant lake-effect snow unlikely. The largest increase in lake-effect snowfall probably occurred after 7000 cal yr BP when Huffman Lake δ^{18}O shifts from high middle-Holocene to low late-Holocene values. This major increase in lake-effect snow coincides with the Great Lakes approaching their current water level. The combination of cold middle-Holocene winters and the expanded breadth of Lake Michigan made abundant lake-effect snowfall possible at Huffman Lake. After 5500 cal yr BP, the abundance of snowfall remained high in the snowbelt, as winter precipitation increased throughout the region.

During recent decades, the abundance of lake-effect snow varied with atmospheric temperatures, the frequency of arctic air mass incursions over the Great Lakes, and Great Lakes water temperatures (Braham and Dungey, 1984; Burnett et al., 2003; Liu and Moore, 2004). The future abundance of lake-effect snow will depend not only upon winter temperatures, but also on changes in Great Lakes water levels, as evidenced by our Holocene records. Therefore, water policy decisions (e.g., diversion of water from the Great Lakes) and climatically-driven changes to the Great Lakes water balance can impact lake-effect snowfall abundance. Major reduction in the Great Lakes levels, such as a return to terminal Great Lakes (Croley and Lewis, 2006), could eliminate a dominant source of winter precipitation in much of the region surrounding the Great Lakes.

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