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# Stable isotope ratios in swale sequences of Lake Superior as indicators of climate and lake level fluctuations during the Late Holocene

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## Abstract

Beach ridges along the coastline of Lake Superior provide a long-term and detailed record of lake level fluctuations for the past 4000 cal BP. Although climate change has been invoked to explain these fluctuations, its role is still in debate. Here, we reconstruct water balance by employing peat samples collected from swale deposits present between beach ridge sequences at two locations along the coastline of Lake Superior. Carbon isotope ratios for *Sphagnum* remains from these peat deposits are used as a proxy for water balance because the presence or absence of water films on *Sphagnum* controls the overall isotope discrimination effects. Consequently, increased average water content in *Sphagnum* produces elevated  $\delta^{13}$ C values. Two maxima of *Sphagnum*  $\delta^{13}$ C values interpreted to reflect wetter conditions prevailed from 3400 to 2400 cal BP and from about 1900 to 1400 cal BP. There are two relatively short drier periods as inferred from low *Sphagnum*  $\delta^{13}$ C values one is centered at about 2300 cal BP, and one begins at 1400 cal BP. A good covariance was found between *Sphagnum*  $\delta^{13}$ C values and reconstructed lake-levels for Lake Michigan in which elevated carbon isotope values correlate well with higher lake levels. Based on this covariance, we conclude that climate exerts a strong influence on lake levels in Lake Superior for the past 4000 cal BP.

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# 1. Introduction

Climatic fluctuations in the Great Lake region can directly affect lake levels, potentially posing serious challenges to the ecosystems, the water quality, and the economy of the entire region (Keddy and Reznicek, 1986; Wilcox, 1989). The Great lakes have experienced short-term lake level variations lasting hours to months that have shown the potential impact that these variations could have on the region, if they were long lived. While storms and seasonal changes in local hydrology cause these short-term variations, long-term changes in the lake levels are the result of climate change. The direct linkage between climate and lake-level fluctuations was identified through the comparison of gauge records and regional water balance inferred from tree rings for the last century (Cook et al., 1992). Because gauge records only span the last 140 years, a more complete source of information is needed to evaluate long-term climate changes and their influence on the Great Lakes. Holocene beach ridges are common along many Great Lakes coasts. These relict coastal features formed in response to millennial to decadal scale fluctuations in lake level during the Holocene (Thompson and Baedke, 1997) and are abandoned away from the modern coast in response to strandplain progradation and isostatic rebound. Thus they offer an opportunity to study millennial-scale changes in climate and lake level fluctuations.

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The chronology of beach-ridge development is based on radiocarbon ages obtained from basal deposits of wetland depressions (swales) located between beach ridges (e.g., Johnston et al., 2000, 2001, 2004). Studies on these abandoned beach ridges follow a sedimentological approach that includes a determination of the elevation of basal foreshore deposits within a given ridge. This datum is related to lake level when each ridge formed (Thompson, 1992). Elevations of foreshore deposits from abandoned beach ridge include, however, two components: local hydrological balance that controls lake levels and long-term isostatic uplift that tilts the basin. Baedke and Thompson (2000) showed that isostatic rebound can be extracted from relative lakelevel hydrographs by incrementally subtracting rates of rebound from each relative hydrograph and comparing residuals. The result is long-term lake-level change at the outlet, which is likely related to changes in water balance. Using this approach, reconstructed lake levels indicate numerous fluctuations with variable magnitude for Lake Michigan and its hydrologic twin, Lake Huron, during the last 4700 years (Thompson et al., 1991; Thompson, 1992; Baedke and Thompson, 2000).

The mechanism triggering these water-level fluctuations is still in debate. Whereas some researchers have suggested that hydrological responses to climatic variations caused these fluctuations (e.g., Larsen, 1985; Fraser et al., 1990; Thompson and Baedke, 1997), others have invoked alternative mechanisms, including isostatic rebound, outlet incision, and outlet switching (e.g., Larsen, 1985; Larsen and Schaetzl, 2001; Booth et al., 2002). The uncertainty about the mechanism(s) responsible for lake-water fluctuations come from the difficulty of correlating reconstructed lake levels with paleoclimatic data. Our study, therefore, strives to fill this gap by inferring climate changes during the past 4000 years from the study of peat deposits accumulated on the shorelines of Lake Superior, and by comparing reconstructed climate events with a chronologically consistent lake levels obtained from sedimentological studies in that area. We employ two complementary beach ridge/swale sequences that cover the interval from 4000 to 900 cal BP.

#### 2. Study area and methods

The first site is the Tahquamenon Bay embayment located in Chippewa County, Michigan, along the southeastern coast of Lake Superior (Fig. 1). This 6km-long and 2.5-km-wide strand plain contains about 80 beach ridges separated by wetland-filled swales that parallel the shoreline. The chronology for the evolution of these ridges shows that ridge formation took place approximately every 28 years (Johnston et al., 2001, 2004). The second site is the Grand Traverse Bay embayment (Fig. 2) on the Keweenaw Peninsula, Houghton County, Michigan, which is a 4-km-long, 2.4-km-wide strand plain located along the Lake Superior shoreline. The average timing for beach ridge



Fig. 1. Map of Tahquamenon Bay embayment. The embayment contains approximately 80 beach ridges. Lines indicate beach-ridge crests and dots indicate the core location.



Fig. 2. Study area of Grand Traverse Bay. This eastward sloping embayment contains 70 shore-parallel beach ridges. The peat samples were taken from the cores depicted by black dots.

development at this site is approximately every 36 years (Johnston et al., 2000). The first two swales of the sequence are about 20 m wide, with a pH of 4.35, and exhibit a 30 cm thick organic-rich deposit. The aquatic vegetation of the swales can be quite diverse, including aquatic or emergent grasses, sedges, herbs, and shrubs. The studied sites exhibit low relief (<2 m) beach ridges and swales that have thick (up to 2 m) accumulations of acidic organic matter with bog-like vegetation in its first and second swales from the modern shoreline. Mosses (*Sphagnum* spp) and sedges (*Carex* spp) are the dominant plants in the swales. Other species include black spruce (*Picea mariana*), paper birch (*Betula papyrifera*), and tamarach (*Larix laricina*).

In the present study, we employ conventional radiocarbon dates obtained from basal peat deposits to develop a chronological framework for the swale sequences. The samples were collected from the deepest point in the swales with the aid of a 10.2 cm piston corer and a 7.5 cm diameter hand-auger (Johnston et al., 2000, 2001, 2004). The radiocarbon dating of these bulk peat samples show that the beach ridges in Tahquamenon Bay formed approximately from 4000 to 2000 cal BP and the beach ridges at Grand Traverse Bay recorded lakelevel fluctuations from about 4100 to 900 cal BP. The ages were calibrated to calendar years before 1950 using the University of Washington's Quaternary Isotope Lab Radiocarbon Calibration Program, CALIB version 4.3 (Stuiver and Reimer, 1993; Stuiver et al., 1998). The ages obtained from the two studied sites show considerable variability when plotted against distance from the modern shoreline (Fig. 3a and b). The variability observed in radiocarbon ages is caused by the use bulk peat samples for radiocarbon dating. Errors in the radiocarbon dating of bulk peat may result either from contamination with older carbon from underlying soils or from errors related to sampling methods in which the lower part of the peat (the true basal peat) is pushed aside during drilling (Törnqvist et al., 1998). Another potential source of error in dating may be caused by post-depositional mobility of certain organic fractions (e.g., humin, humic and fulvic acids) exhibiting



Fig. 3. Graphs of calibrated radiocarbon ages of basal peat deposits versus distance from modern shoreline for (a) Grand Traverse Bay embayment and (b) Tahquamenon Bay embayment. The radiocarbon ages of samples were calibrated to correct for variations in atmospheric <sup>14</sup>C through time using the Radiocarbon calibration program CALIB version 4.3. (Stuiver and Reimer, 1993; Stuiver et al., 1998). A regression line was used to calculate the approximate age of each beach ridge at the two studied sites.

significant differences in radiocarbon contents within a single peat sample (Shore et al., 1995).

To better constrain the ages and to implement a chronology that follows a progradational system, a bestfit line was calculated for each sequence. The correlation coefficient between the radiocarbon ages and the distance from modern shoreline is statistically significant for each site, i.e., 0.64 for Grand Traverse (Fig. 3a) and 0.70 for Tahquamenon Bay (Fig. 3b). The age of the individual peat samples is obtained from this best-fit line. The standard error in individual ages as reported in this study is based on the uncertainty associated with the linear regression model.

The peat samples collected from the deepest point of swales were used for the isotopic study. To maximize the signal-to-noise ratio, we concentrated our efforts on *Sphagnum* because it not only accounts for a significant proportion of vegetation in peat bogs, and its physiology is well understood, but also because it contains a large amount of complex lignin-like compounds that are more resistant to decay (Williams et al., 1998).

Retrieved peat deposits were freeze-dried, and Sphagnum macrofossils were manually picked out, air-dried, powdered, homogenized, and washed with alkaline solutions to remove soluble organic acids. Homogenization of macrofossils ensures that the measured isotopic composition is representative of the isotopic signature of the most abundant type of mosses, which minimizes inter-specific effects. The dried samples ( $\sim 1 \text{ mg}$ ) were then combusted in a Costech elemental analyzer fitted to a Finnigan MAT Delta Plus XL mass spectrometer housed in the Department of Geological and Atmospheric Sciences at Iowa State University. All samples were analyzed for carbon and nitrogen isotopic compositions ( $\delta^{13}$ C and  $\delta^{15}$ N, respectively) as well as total nitrogen (N) and carbon (C) concentrations. Precision for  $\delta^{13}$ C and  $\delta^{15}$ N analyses based on duplicate analysis is better than  $\pm 0.1\%$ , and they are expressed relative to the international standards VPDB and atmospheric nitrogen (AIR), respectively.

#### 3. Results and discussion

### 3.1. Carbon isotopes in Sphagnum

The  $\delta^{13}$ C values in Grand Traverse Bay range between -26% and -30.5% (Fig. 4a). The highest  $\delta^{13}$ C values occur in peat deposits of ridges located 800–400 m from the shoreline and the lowest  $\delta^{13}$ C values were found in peats located about 1200 m from the shoreline. In Tahquamenon Bay,  $\delta^{13}$ C values vary from -27% to -29% with the highest  $\delta^{13}$ C values measured from peat deposits of the ridges located 900–1400 m inland from the modern shoreline (Fig. 4b). The variability found in the carbon isotopic composition of



Fig. 4. (a)Values of  $\delta^{13}$ C and  $\delta^{15}$ N plotted as a function of distance from the modern shoreline and age (in cal BP) in the Grand Traverse Bay embayment. (b)Values of  $\delta^{13}$ C and  $\delta^{15}$ N plotted as a function of distance from the modern shoreline and age (in cal BP) in the Tahquamenon Bay embayment.

*Sphagnum* remains from the two sites falls in the range described in the literature (Rice and Giles, 1996; Hornibrook et al., 2000; Ménot and Burns, 2001).

Variations in  $\delta^{13}$ C values in mosses typically result from changes in the amount of moisture on the plants and from shifts in the concentration and isotopic composition of atmospheric CO<sub>2</sub> (Ménot and Burns, 2001). The gametophytes of mosses are morphologically similar to higher vascular plants and they fix CO<sub>2</sub> directly. As a result, the mathematical treatment of carbon isotope discrimination in mosses is similar to that of the vascular plants (White et al., 1994; Ménot and Burns, 2001). Fractionation of carbon isotopes during vascular plant photosynthesis is described by the relation (Farquhar et al., 1982):

$$\delta^{13}C_{\text{plant}} = \delta^{13}C_{\text{air}} - a - (b - a)p_i/p_a,\tag{1}$$

where  $\delta^{13}C_{plant}$  and  $\delta^{13}C_{air}$  are the carbon isotopic composition (relative to VPDB) of plant tissues and atmospheric CO<sub>2</sub>, respectively. The terms  $p_i$  and  $p_a$ represent the intercellular and atmospheric partial pressures of  $CO_2$ . The constant *a* is the enrichment factor associated with the different diffusivities of <sup>12</sup>CO<sub>2</sub> and  ${}^{13}CO_2$ , and the constant b expresses the fractionation associated with carboxylation. Environmental factors dictate plant  $p_i/p_a$  ratios, thereby playing a major role in determining the carbon isotopic composition of plants. Unlike vascular plants that regulate carbon and water exchange by controlling stomatal apertures on leaf surfaces, mosses lack stomata or an epidermis with an impermeable cuticle. Consequently, mosses are unable to regulate the uptake of atmospheric  $CO_2$  (Rice and Giles, 1996).

In mosses, CO<sub>2</sub> passively diffuses from the atmosphere through the cell walls and into the chloroplast, where photosynthesis occurs. In the absence of a water film and when hyaline cells are less full, metabolic activity limits carbon fixation, which produces higher  $p_i$  $p_{\rm a}$  ratios. In contrast, the presence of a water film that coats a moss plant prevents the direct passage of CO<sub>2</sub> to the chloroplast, which decreases cellular CO<sub>2</sub> concentrations and  $p_{i}/p_{a}$  ratios because the diffusivity of CO<sub>2</sub> is four orders of magnitude slower in water than in air (Ménot and Burns, 2001). The degree to which water films on moss plants affect  $\delta^{13}C$  values has been investigated in the laboratory (e.g. Rice and Giles, 1996). These studies indicate that there is a preferential incorporation of <sup>12</sup>C relative to <sup>13</sup>C when no water films exist. Even though the presence of water films enhances the discrimination against <sup>13</sup>C as a result of the faster diffusivity of <sup>12</sup>CO<sub>2</sub> relative to that of <sup>13</sup>CO<sub>2</sub>, much slower diffusion of CO2 through water film limits the total amount of CO<sub>2</sub> that reaches the chloroplast. Photosynthesis in the presence of water film is thus CO<sub>2</sub> limited, resulting in reduced discrimination against <sup>13</sup>C. Consequently, the overall effect of increased average water content in Sphagnum produces an increase in  $\delta^{13}$ C values. This effect can then be used in paleoclimatic reconstructions because an increase in precipitation or moisture in the air results in formation of water films on mosses, causing an increase in  $\delta^{13}$ C values.

The concentration and carbon isotopic composition of CO<sub>2</sub> in the atmosphere can also affect the  $\delta^{13}$ C values recorded in plants as seen in Eq. (1). To quantify the magnitude of this effect, Ménot and Burns (2001) investigated variations in  $\delta^{13}$ C values of moss plants as a function of varying CO<sub>2</sub> concentrations along an altitude transect in the Swiss Alps. Their results suggest a linear increase of 0.9–1.1‰/km in moss  $\delta^{13}$ C values. This finding indicates a decrease in  $\delta^{13}$ C values of about 1‰ per 36.5 ppmv increase in atmospheric CO<sub>2</sub>. Atmospheric CO<sub>2</sub> concentrations have fluctuated little during the past 4000 years i.e., from a low of about 275 ppmv at about 4000 cal BP to a high of about 282 ppmv around 1200 cal BP (Indermühle et al., 1999). This change in CO<sub>2</sub> concentrations only accounts for about 0.19‰ of the 4.5‰ variation measured for the peat samples. Moreover, Indermühle et al. (1999) found that the  $\delta^{13}$ C values of CO<sub>2</sub> in the atmosphere have fluctuated by about 0.1‰ over the time period in question, thus causing a negligible effect on the carbon-isotopic composition of mosses. Thus, the 4.5‰ variability in moss  $\delta^{13}$ C values for the peat deposits is likely to result from physiological constraints.

We argue that changes in regional water balance produced the measured variability in  $\delta^{13}$ C values by affecting the level of water cover on mosses growing in swales around Lake Superior. Consequently, higher  $\delta^{13}$ C values are interpreted to reflect higher humidity levels. If this interpretation is correct, variability in humidity levels as inferred from  $\delta^{13}$ C values would likely reflect late spring to early summer conditions given that mosses in the studied area primarily grow during this time of the year (McNeil and Waddington, 2003). Based on this correspondence, we obtain a fairly continuous time-series for peat deposits of the region by combining the  $\delta^{13}$ C values of *Sphagnum* remains from the two studied sites (Fig. 5). We interpret that wetter late spring-early summer conditions prevailed from 3400-2300 cal BP and from 1900-1400 cal BP as inferred from elevated  $\delta^{13}$ C values of peat samples. In contrast, we interpret that drier conditions existed around 2400-2200 cal BP and from about 1400 to 1000 cal BP as indicated by lower  $\delta^{13}$ C values.

A possible complication to our interpretations may come from interspecies variation, which results from the different rate of exposure of photosynthetic cells at the leaf surface. This process affects the rate at which  $CO_2$ 



Fig. 5. Variation of  $\delta^{13}$ C values of *Sphagnum* remains from the two studied sites over time (in cal BP).

can diffuse through and into these cells (Rice and Giles, 1996). Experimental studies on the fractionation of carbon isotopes by mosses have shown that variations in leaf morphology have little impact on the resistance to CO<sub>2</sub> uptake and carbon isotope discrimination compared to the variation in surface water films (Rice and Giles, 1996). This conclusion was confirmed with field studies that showed a stronger effect of environmental conditions on the extent of fractionation of carbon isotopes. For example, Rice (2000) studied three Sphagnum species grown in a greenhouse and in a natural wetland in North Carolina to evaluate the extent of isotopic discrimination resulting from the influence of micro-site, seasonality, and genetic variations. This author found that although genetics can potentially determine the extent of carbon isotope discrimination as deduced from  $\delta^{13}$ C values of species in a common greenhouse, environmental variability in different micro-sites is by far the dominant effect, obscuring speciesspecific effects. Based on these results, it is likely that interspecies variations play a minor role in the observed variability of *Sphagnum*  $\delta^{13}$ C values through time in the Lake Superior region. Consequently, we can assume that the observed variation in *Sphagnum*  $\delta^{13}$ C values is due to environmental causes, i.e., differences in water exposure levels in the micro-sites (i.e., hummocks, hollows, and carpets) and seasonal variations rather than from genetic causes.

Available  $\delta^{13}$ C data for mosses growing in modern settings experiencing different environmental conditions show a maximum variability of 8‰ (Rundel et al., 1979; Proctor et al., 1992). This variability decreases, however, to about 1.5% within a single bog (e.g., Proctor et al., 1992; Rice, 2000; Ménot and Burns, 2001). For instance, Proctor et al. (1992) measured the isotopic composition of mosses present in a bog in East Devon, UK. These authors found that  $\delta^{13}$ C values of mosses growing in different areas of the bog show an average value of -28.33‰ and a standard deviation of 1.35‰. Similar variability, although with different average values, have been reported for other bogs (Teeri, 1981; Rice, 2000). Within a single site, the range of moss  $\delta^{13}$ C values likely results from the fact that hollows and hummocks experience different levels of water exposure (Ménot and Burns, 2001; Rice, 2000). The only exception is S. cuspidatum (a submerged species), which appears to utilize recycled carbon and, therefore, exhibits lower  $\delta^{13}$ C values than those shown by co-occurring mosses (Proctor et al., 1992; Ménot and Burns, 2001). Our  $\delta^{13}$ C data show a 4.5% increase in  $\delta^{13}$ C values for peat deposits formed from 3400-2300 cal BP and from 1900–1400 cal BP relative to peats formed around 2400–2200 cal BP. We argue that this offset is unlikely to result from inter-specific effects because these effects appear to have a negligible effect on the extent of carbon isotope enrichment in mosses (Rice and Giles, 1996;

Rice, 2000). The observed 4.5% shift is unlikely to be associated with noise produced by the natural variability of  $\delta^{13}$ C values shown by mosses growing in different micro-habitats within a region because available data suggests typical ~1.5% variability associated with micro-habitats (Teeri, 1981, Proctor et al., 1992; Rice, 2000). We argue, therefore, that the 4.5% shift in  $\delta^{13}$ C values of the studied peat samples is likely caused by changes in humidity levels occurring through time.

Another factor affecting the carbon isotopic composition of organic remains is diagenesis. Although organic matter abundance typically decreases during early diagenesis, studies indicate that organic matter can retain its original  $\delta^{13}$ C signature for long (million year) time periods (Meyers, 1994; Schelske and Hodell, 1995). Some studies (e.g., Spiker and Hacker, 1984; Harvey et al., 1995) show, however, a decline in  $\delta^{13}$ C values during organic matter degradation. This decline results from the preferential degradation of compounds enriched in <sup>13</sup>C (e.g., carbohydrates and proteins), which are more susceptible to microbial degradation compared to <sup>13</sup>C-depleted compounds such as lipids. The selective loss of carbohydrates and proteins during microbial degradation leads to a decrease in the  $\delta^{13}$ C value of residual organic matter.

Two lines of evidence indicate that diagenesis likely had a negligible effect on the original carbon isotopic signatures of the studied samples. First,  $\delta^{13}$ C and  $\delta^{15}$ N values and C/N ratios of the studied Sphagnum remains fall in the range of typical values reported in literature for modern Sphagnum plants (Rice and Giles, 1996; Hornibrook et al., 2000; Ménot and Burns, 2001; Kohzu et al., 2003). Second, diagenesis affects primarily nitrogen-bearing compounds, thus producing an increase in C/N ratios and a decrease in  $\delta^{13}$ C values (Herczeg, 1988; Meyers, 1997). This trend is absent in the studied samples, suggesting a minor effect of diagenesis on the carbon isotopic composition of the studied Sphagnum macro-fossils. Additionally, indirect evidence for the negligible effect of diagensesis comes from the nitrogen-isotopic composition of organic matter. A gradual enrichment of  $\delta^{13}C$  occurs under oxic conditions during diagenesis, while the opposite effect occurs under anoxic conditions (Libes and Deuser, 1988; Lehmann et al., 2002). Differences in type, timing, and degree of microbial activity and preferential degradation appear to explain the divergent trends in  $\delta^{13}$ C values with diagenesis. Based on these observations, an inverse correlation should be expected between Sphagnum remains  $\delta^{13}$ C and  $\delta^{15}$ N values, if diagenesis played a significant role on the isotopic composition of the studied peat deposits given that anoxic conditions likely existed during peat formation. No significant correlation (r = 0.05) was found between  $\delta^{13}$ C and  $\delta^{15}$ N values (Fig. 6), indicating that diagenesis did not significantly alter the original isotopic composition of the Sphagnum remains found in the studied peat



Fig. 6. Cross plot of  $\delta^{13}$ C and  $\delta^{15}$ N values for *Sphagnum* remains from the two studied sites, showing a poor correspondence between these two isotopic compositions.

samples. Additional support for the minimal role of diagenesis on the  $\delta^{13}$ C and  $\delta^{15}$ N values of Sphagnum remains comes from C/N ratios. As discussed above, diagenesis induces an increase in C/N ratios which is accompanied by an increase in  $\delta^{13}$ C values under oxic conditions and a decrease in  $\delta^{13}$ C values under anoxic conditions. Our data show a poor (r = 0.02) correlation between C/N ratios and  $\delta^{13}$ C values in the studied samples (Fig. 7), suggesting that there was no common factor affecting the two variables. It is possible that degradation may have occurred under anoxic conditions followed by oxic/sub-oxic conditions, or vice versa, as a result of seasonal fluctuations of the water table or changes in climate. Under this scenario of fluctuating oxygen levels, little or no correlation would likely result between  $\delta^{13}$ C and  $\delta^{15}$ N values and between C/N ratios and  $\delta^{13}$ C values. We argue, however, that the similarity in  $\delta^{13}$ C values and in C/N ratios between modern mosses and our collected samples suggest that diagenesis likely produced a minimal effect on the original carbon isotopic signature. Moreover, it is difficult to envision a diagenetic scenario that could produce a 4.5% shift in  $\delta^{13}$ C values without affecting C/N ratios. We conclude, therefore, that the observed 4.5‰ fluctuations in the observed time series of  $\delta^{13}$ C values (Fig. 5) are likely related to changes in humidity levels over time. On the other hand, the small-scale fluctuations (<1.5%) can be produced by mosses growing in different micro-habitats, diagenesis, or climate changes.

# 3.2. Nitrogen isotopes in Sphagnum

No discernable trend in  $\delta^{15}$ N values is observed with time at the studied sites (Fig. 8). In Grand Traverse Bay,



Fig. 7. Plot of  $\delta^{15}$ N values against C/N ratios of *Sphagnum* remains from the two studied sites, showing no correlation between these two parameters.



Fig. 8. Change in  $\delta^{15}$ N values of *Sphagnum* remains from the two studied sites over time (in cal BP).

 $\delta^{15}$ N values vary from -2.5% to about +2.0%(Fig. 4a), exhibiting elevated  $\delta^{13}$ C values in peat deposits located 900, 1300 and 1800 m landward from the modern shoreline. The  $\delta^{15}$ N values in Tahquamenon Bay vary from -2.5% to +2% (Fig. 4b), exhibiting the highest  $\delta^{15}$ N values in peat deposits of ridges located 900–1400 m and 800 m inland from the modern shoreline.

Interpretations of  $\delta^{15}$ N data are difficult due to the dynamics of plants and of nitrogen biogeochemical cycling in swales. Nitrogen (N) can enter the swale waters in several forms. For example, N carried in surface water may be in dissolved or particulate

(suspended sediment) form, and may exist both as organic and inorganic compounds. Organic N compounds are either dissolved or in suspension, whereas inorganic N-forms are primarily in solution as nitrate  $(NO_3^-)$  or ammonium  $(NH_4^+)$ . A significant amount of  $NO_3^-$  in swales comes from allochthonous sources, including groundwater, surface water, and precipitation.

Sphagnum takes up  $NH_4^+$  and  $NO_3^-$  from the surrounding water or through accumulation of dry deposition. Although changes in *Sphagnum*  $\delta^{15}$ N values can be attributed to patterns of  $NH_4^+$  and  $NO_3^$ utilization from the dissolved inorganic N pool, Sphagnum species growing in the hummocks or the hollows of the peat bog differ in their relative  $NO_3^-$  and  $NH_4^+$  uptake. The hummock species have higher cation exchange capabilities and therefore have higher  $NH_4^+$ uptake compared to species growing in the hollows (Jauhiainen et al., 1998). Consequently, two species growing at the same time and under similar climatic conditions may show potentially different  $\delta^{15}N$  values due to differences in their relative uptake of  $NO_3^-$  and  $NH_4^+$  which are related to their position in the bog. This difference occurs despite the significantly higher rate of NH<sub>4</sub><sup>+</sup> uptake in different Sphagnum species relative to that of  $NO_3^-$  (Jauhiainen et al., 1998). In our study it is not possible to identify the Sphagnum remains on species level. Therefore, the results of the nitrogen isotopic study are not well constrained to give meaningful interpretation. Unlike many vascular plants, however, Sphagnum mosses posses a short lag phase in activation of nitrate reductase, thereby making these plants capable of assimilating even short  $NO_3^-$  pulses (Woodin et al., 1985; Rudolph et al., 1993). This effect probably accounts for the different patterns of  $NO_3^-$  and  $NH_4^+$ uptake within a single swale.

In addition to differences in NO<sub>3</sub><sup>-</sup> and NH<sub>4</sub><sup>+</sup> uptake, variability in *Sphagnum*  $\delta^{15}$ N values can result from differences in  $\delta^{15}$ N values of NH<sub>4</sub><sup>+</sup> and NO<sub>3</sub><sup>-</sup>. For example, a study in the Grand Traverse Bay area along Lake Michigan shows that the average  $\delta^{15}$ N of NO<sub>3</sub><sup>-</sup> is about 9‰ lower than that of NH<sub>4</sub><sup>+</sup> (McCusker et al., 1999). This isotopic difference results from the isotopic discrimination associated with microbial nitrification (i.e., the oxidation of NH<sub>4</sub><sup>+</sup> to NO<sub>3</sub><sup>-</sup>) in aerobic or oxygenating conditions. Under the low Eh, pH, and oxygen levels typical of submerged conditions in the study sites, nitrification rates are probably low. Consequently, NH<sub>4</sub><sup>+</sup> is the most likely dissolved N-species.

Another factor complicating the interpretation of  $\delta^{15}$ N data is the potential addition of NO<sub>3</sub><sup>-</sup> and NH<sub>4</sub><sup>+</sup> through the atmosphere as wet or dry precipitation. The sum of relative percentages of NO<sub>3</sub><sup>-</sup> and NH<sub>4</sub><sup>+</sup>, their  $\delta^{15}$ N values, and their annual variation are not well understood, further complicating the interpretation of nitrogen isotopes in terms of palaeoclimatic variations. Because of the complexities associated with changing

N-sources (i.e.,  $NO_3^-$  and  $NH_4^+$ ), their isotopic composition, moss physiology, and inter-species effects, it is difficult to constrain the precise process responsible for the measured variability in  $\delta^{15}N$  values with the available data.

## 3.3. Palaeoclimatic implications

Interpretation of the carbon isotope data indicates that wetter conditions prevailed from 3400 to 2400 cal BP and from about 1900 to 1400 cal BP. There are two relatively short drier periods around 2400 to 2200 cal BP and from 1400 cal BP to the most recent peat deposit.

The palaeoclimate record of this region indicate an increase in effective moisture during the past 3200 years as inferred from increased abundance of mesic taxa, enhanced wetlands, and elevated water levels in small lakes of the Great Lakes region (e.g., Brubaker, 1975; Webb et al., 1983; Futyma and Miller, 1986; Winkler et al., 1986; Brugam and Johnson, 1997). Using a modern analog method for pollen, Davis et al. (2000) reconstructed the Holocene climate in Western Great Lakes National Parks and Lakeshore, concluding that the Great Lakes region experienced a steep increase in annual precipitation from 4000 to 2000 cal BP. Although this regional paleoclimate reconstruction does not show the dry period recorded in our  $\delta^{13}$ C data from 2400 to 2200 cal BP, reconstructed lake levels for Lake Michigan do show low lake levels for this time interval (Thompson et al., 1991; Thompson, 1992; Baedke and Thompson, 2000). A comparison of the time-series of  $\delta^{13}$ C values with Michigan Lake levels (Fig. 9) shows a good correspondence. The two maxima in  $\delta^{13}$ C values coincide with two high water levels (Algoma high stand from about 3400 to 2300 cal BP. and an unnamed high stand between 1900 and 1400 cal BP). In addition to the dry event centered at about 2300 cal BP, our  $\delta^{13}$ C data also show a dry event starting from about 1400 cal BP. This latter event also correlates well with declining lake levels in Lake Michigan.

Although our  $\delta^{13}$ C record indicates that the two major lake level declines are probably related to a decrease in water balance, reconstructions based on pollen and plant macrofossils from a coastal lake in Upper Michigan (Booth et al., 2002) suggest no clear evidence for effective moisture decrease. This reconstruction does show that the high water phase (Algoma phase) is related to effective moisture increase. Consistent with this finding, Woods and Davis (1989), Brubaker (1975), and Davis et al., (2000) also reported a humid phase during this time period based on the pollen studies in Northern Michigan sites. Booth et al. (2002) postulate that the absence of evidence for effective moisture decreases after the two major lake-level high stands suggest other processes unrelated to climate.



Fig. 9. Comparison of the combined carbon isotopic data for the two studied sites with regional lake-level hydrograph reconstructed for Lake Michigan (after Baedke and Thompson, 2000) relative to International Great Lakes Datum (IGLD) for 1985. Dotted line represents the average lake level between 1819 and 1999.

They hypothesize that these large-magnitude water-level declines were most probably a result of outlet incision. In contrast, our  $\delta^{13}$ C record reveals that the two major lake level declines are accompanied by a sharp decrease in  $\delta^{13}$ C values best interpreted to result from reduced precipitation. However, additional data is needed to support this hypothesis given the uncertainty associated with the obtained radiocarbon dates.

# 4. Conclusions

We use  $\delta^{13}$ C values of *Sphagnum* remains from peat deposits located along a sequence of beach ridges of Lake Superior to reconstruct changes in regional water balance. Based on modern physiological constraints, we interpret that elevated  $\delta^{13}$ C values correspond to wetter conditions, while lower  $\delta^{13}$ C values reflect drier conditions. Although diagenesis could be a confounding factor in our interpretations, the similarity in isotope and C/N ratios between the studied macro-fossils and modern mosses, along with the absence of discernable trends in C/N ratios and  $\delta^{13}$ C values suggest little diagenetic effect on the original carbon isotope signature of *Sphagnum* remains. Therefore, it can be used to reconstruct the climate of the region. Our results indicate, however, that it is difficult to interpret  $\delta^{15}N$  data of peat deposits because of the number of variables affecting N-fractionation.

We found two maxima of *Sphagnum*  $\delta^{13}$ C values in peat deposits developed from 3400 to 2400 cal BP and from 1900 to 1400 cal BP. We also identified two minima of *Sphagnum*  $\delta^{13}$ C values: one centered around 2300 cal BP and one beginning at 1400 cal BP. These maxima and minima in  $\delta^{13}$ C values correlate well with lake-level high stand and low stand reconstructed for Lake Michigan. Although errors related to radiocarbon dating are confounding factors in our interpretations, the close correspondence between lake levels and *Sphagnum*  $\delta^{13}$ C values lead us to conclude that regional water balance has played a major role in controlling water levels in the Great Lakes region.

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