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# Late Holocene climate change in central Sweden inferred from lacustrine stable isotope data

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**ABSTRACT:** Stable isotopes ( $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$ ) of lacustrine carbonates (*Chara* spp. algae and *Pisidium* spp. molluscs) from a lake sedimentary sequence in central Sweden were analysed to infer changes in lake hydrology and climate during the late Holocene. Results from analysis of lake water isotopes ( $\delta^{18}\text{O}$  and  $\delta^2\text{H}$ ) show that Lake Blektjärnen water isotope composition is responsive to the balance between evaporation and input water (*E/I* ratio). A high *E/I* ratio results from a dry and probably warmer climate, decreasing the relative importance of precipitation input. Under such conditions evaporation and atmospheric equilibration probably enrich lake water in  $^{18}\text{O}$  and  $^{13}\text{C}$ , respectively, which is reflected in the isotopic composition of the carbonates in the lake. From the relatively positive *Chara*  $\delta^{18}\text{O}$  values we infer that conditions were dry and warm between 4400 and 4000 cal. a BP, whereas more negative values indicate that conditions were wetter and probably cooler between 4000 and 3000 cal. a BP. A drier climate is inferred from more positive values between 2500 and 1000 cal. a BP. However, a successive depletion after ca. 1750 cal. a BP, also detected in several other  $\delta^{18}\text{O}$  records (carbonate and diatom), suggest increasingly wetter conditions in Scandinavia after that time, which is probably related to increased strength of the zonal flow. Copyright © 2010 John Wiley & Sons, Ltd.



**KEYWORDS:** stable isotopes; lake hydrology; humidity; late Holocene; central Sweden.

## Introduction

During the Holocene (the present interglacial that commenced ca. 11 650 cal. a BP; Walker *et al.*, 2009) periods of marked climate change have been detected in palaeoclimatic archives in northern Europe. Climate has changed on millennial (e.g. Marchal *et al.*, 2002; Mayewski *et al.*, 2004; Moberg *et al.*, 2005) as well as on centennial to annual scales (e.g. Grudd *et al.*, 2002; Hammarlund *et al.*, 2002; Rosqvist *et al.*, 2007; Grudd, 2008). Results show that the climate was maritime, i.e. warmer and wetter, in Fennoscandia between ca. 10 000 and 8000 cal. a BP (Hammarlund *et al.*, 2003; Seppä *et al.*, 2005), and that between ca. 8000 and 5000 cal. a BP mean annual temperatures were at least 1°C higher than today (Heikkilä and Seppä, 2003; Seppä *et al.*, 2005; Antonsson and Seppä, 2007). A change in climate mode between ca. 5000 and 4000 cal. a BP that led to a more unstable colder and humid climate has been detected in lake and glacier records (Nesje *et al.*, 2001, 2008; Seppä and Birks, 2001; Hammarlund *et al.*, 2003; Bjune *et al.*, 2005; Jessen *et al.*, 2005; Seppä *et al.*, 2005; Antonsson *et al.*, 2008; de Jong *et al.*, 2009; Ojala *et al.*, 2008). Climate

variability in Fennoscandia is related to changes in the North Atlantic ocean and atmospheric circulation, although the forcing is still not understood (Wanner *et al.*, 2008).

This study aims to reconstruct lake water  $\delta^{18}\text{O}$  variation as a means of inferring changes in moisture supply in an area where such information is scarce. Other studies in northern Sweden have shown that variations in stable isotopes in lacustrine sediments reflect changes in precipitation and temperature that can be related to shifts in atmospheric circulation patterns (Hammarlund *et al.*, 2002; Rosqvist *et al.*, 2004, 2007). Here we focus on the late Holocene between 4400 cal. a BP and the present. We use oxygen ( $\delta^{18}\text{O}$ ) and carbon ( $\delta^{13}\text{C}$ ) isotopes of *Chara* encrustations and shells of *Pisidium* molluscs obtained from Lake Blektjärnen, a hard-water lake located east of the Scandes Mountains in central Sweden, to reconstruct past changes in evaporation/inflow ratio (*E/I*). Charophytes are photosynthesising green algae that precipitate calcite mainly between June and August (cf. Mörner and Wallin, 1977), whereas *Pisidium* are bottom-dwelling filter-feeding bivalves mainly active during the ice-free season (late May to early November) (Kilgour and Mackie, 1991) with life spans of approximately 1–4 years (Holopainen and Hanski, 1986; Killeen *et al.*, 2004). Isotope analyses of both *Chara* and *Pisidium* thus provide us with the opportunity to identify whether changes in the *E/I* ratio occurred mainly during summer (*Chara*) or during a longer season (*Pisidium*).

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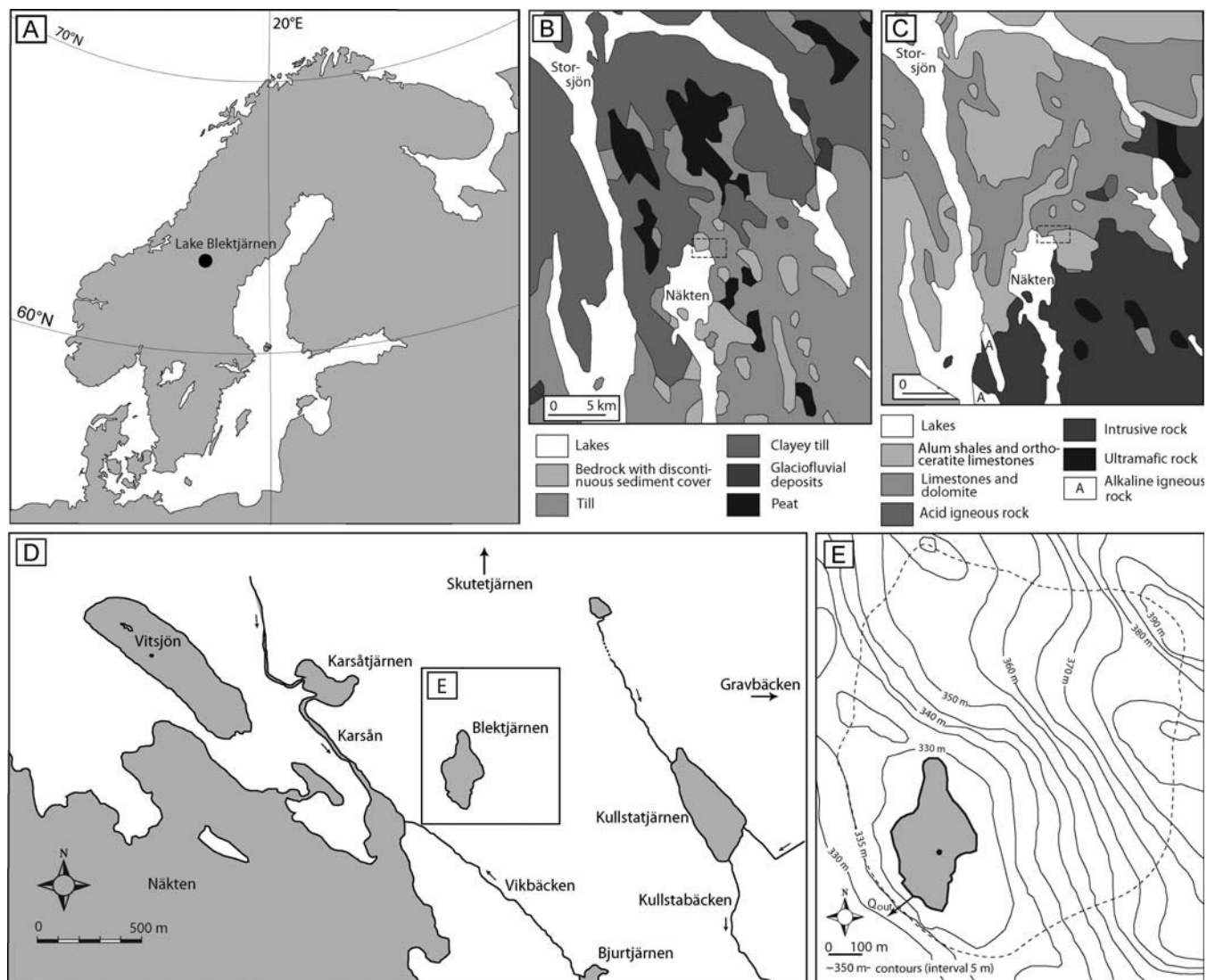
## Site description

Lake Blektjärnen (62° 1' N, 14° 39' E; 330 m above sea level (a.s.l.)) is a small hard-water lake located in central Sweden. It is situated ~200 m north-northeast of the northern shore of Lake Näkten (Fig. 1(A–D)), to which it was connected during the last deglaciation and early Holocene (Lundqvist, 1969). Lake Blektjärnen has a surface area of approximately 0.05 km<sup>2</sup> and a catchment area of ~0.45 km<sup>2</sup> (Fig. 1(E)) and is 10 m deep (mean water depth ~4 m). The lake has an estimated water volume of 156 000 m<sup>3</sup>. Lake water residence time has been calculated to >1 year based on the modern lake volume and the amount of mean annual effective precipitation (precipitation–evaporation) falling in the catchment. At present the only outflow is via a pipe placed ~50 cm below the lowest shoreline (Fig. 1(E)); maximum measured flow rates from this pipe were 1.5–2 L s<sup>-1</sup>.

The vegetation in the catchment is woodland dominated by birch, willow and pine. The beach of Lake Blektjärnen is composed of calcareous gyttja, which indicates that lake level has been higher in the past. The bedrock around Lake Blektjärnen is composed of Cambro-Silurian sedimentary rocks, mostly alum shale and limestone (Norling, 2002), which

extend to the north and northwest, while igneous rocks predominate in the south (Fig. 1(C)). The catchment is covered in a layer of Quaternary sediments (sandy till and clayey till) ~0.5 m thick (Lundqvist, 1969) (Fig. 1(B)). The region was ice free by ca. 10 000 cal. a BP (Lundqvist, 2002) and is characterised by several calcareous lakes. During the deglaciation ice-dammed lakes formed, which were subsequently drained. Lake Näkten and Lake Blektjärnen were part of such an ice-dammed lake which had a paleoshoreline up to 365 m a.s.l. (Lundqvist, 1969), 35 m above the present level. The base of the Lake Blektjärnen sediment sequence is composed of clayey sand, which is overlain by clayey silt and clay, and calcareous gyttja. A date of 7850 ± 60 <sup>14</sup>C a BP obtained on macrofossils (*Betula* twig) at ~620–625 cm depth (8729 ± 249 cal. a BP) (S. Andersson, unpublished) provides a minimum age for the end of the ice lake stage and the start of lake infilling.

Mean annual precipitation (1961–1990) measured at the nearest meteorological station, Östersund, is 564 mm a<sup>-1</sup>, with maximum monthly precipitation in July (86 mm), while the lowest occurs in February (29 mm). Mean annual temperature is +3°C and average temperatures for January and July are –9°C and +14°C, respectively (1961–1990) (Alexandersson and Eggertsson Karlström, 2001), and evaporation amounts to ~400 mm a<sup>-1</sup> (Bringfelt and Forsman, 2004). The climate in the



**Figure 1** (A) Map of Scandinavia with the location of Lake Blektjärnen. (B) Quaternary deposits in the area modified after Lundqvist (1969) and (C) bedrock map modified after Gorbatshev *et al.* (1997), Karis and Strömberg (1998) and Sturkell *et al.* (1998). Field area is indicated by a box (dotted line). (D) Lake Blektjärnen and other surface waters where water was sampled. (E) Lake Blektjärnen and its catchment (dotted line). The coring site is marked by a black dot and Q<sub>out</sub> denotes the location of the modern pipe

study area is more continental, i.e. drier and cooler, compared to the climate at the lake sites located in the Scandes Mountains, where other palaeoclimatic reconstructions have been conducted (Rosqvist *et al.*, 2004; St Amour, 2009). Lake Blektjärnen was chosen as it has a small catchment and is effectively a closed lake despite the small modern outflow. Therefore, water loss is predominantly controlled by evaporation and the lake water isotope composition is probably responsive to the balance between evaporation and input water. Because of this, changes in humidity can potentially produce significant changes in water  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values (Leng and Marshall, 2004). We assume that lower humidity, or more arid climatic conditions, leads to higher (less negative)  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values that reflect  $^{18}\text{O}$  and  $^{13}\text{C}$  enrichment due to increased evaporation and atmospheric equilibration. In contrast, increased humidity or wetter climatic conditions result in lower (more negative)  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values that reflect less  $^{18}\text{O}$  and  $^{13}\text{C}$  enrichment because of reduced evaporation and atmospheric equilibration (Li and Ku, 1997). Changes in humidity often lead to covariance between  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  isotopes in lacustrine carbonate (Talbot, 1990), which can also be evident on a seasonal basis. For example, during the summer months  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  in lake water becomes enriched due to increased evaporation and productivity (Turner *et al.*, 1982) but this relationship can be lost during the winter months due to lake water recharge by precipitation. However, non-evaporating lakes can also have higher  $\delta^{18}\text{O}$  due to addition of enriched  $^{18}\text{O}$  from summer precipitation (Tyler *et al.*, 2007), but the amount of  $\delta^{18}\text{O}$  change will be significantly less.

## Materials and methods

### Fieldwork

Fieldwork was conducted in 2005, 2006 and 2008. Long sediment cores were extracted from the lake bed from the frozen lake surface at a water depth of 400 cm using a Russian corer (Jowsey, 1965) (100 cm length, diameter 5 cm), with overlaps of 50 cm in April 2006 (Fig. 1(D)). The cores were placed in PVC tubes and wrapped in plastic, before being transferred to cold-room storage until required for analysis. This study focused on the uppermost 100 cm of the 266 cm long sediment sequence. To obtain enough material for all planned analyses six parallel cores were obtained between 400 and 500 cm below the water surface.

Surface water samples were collected from lakes and a stream in the proximity of Lake Blektjärnen (Fig. 1(D)). Properties for each lake were calculated based on estimated catchment and lake area and mean water depth (Table 1). Modern water samples were taken in May and September in 2008 in order to detect seasonal  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  variation of water due to

**Table 1** Estimated properties for lakes Blektjärnen, Bjurtjärnen, Karsätjärnen, Kullstatjärnen, Skutetjärnen and Vitsjön

Name of lake	Estimated catchment area (km <sup>2</sup> )	Estimated lake volume (m <sup>3</sup> )	Inflow and outflow
Blektjärnen	0.56	160 000	Modern outflow
Bjurtjärnen	0.94	44 000	Yes
Karsätjärnen	0.7	63 000	Yes
Kullstatjärnen	1	360 000	Yes
Skutetjärnen	0.3	80 000	Yes
Vitsjön	0.63	440 000	No

**Table 2**  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  analytical results of water samples collected in Lake Blektjärnen and surrounding lakes and streams. Samples were analysed at the Department of Geology and Geochemistry Stockholm University, and at NIGL, Nottingham, UK (marked in bold). See Fig. 1(D) for location of sites

Location of water samples	Date of sampling (day/month/year)	$\delta^{18}\text{O}$ (‰)	$\delta^2\text{H}$ (‰)
Blektjärnen (lake)	27/7/2005	-9.95	-80.78
	25/10/2005	-9.82	-79.97
	4/4/2006	-10.23	-82.89
	7/9/2006	-9.7	-80.52
	<b>23/5/2008</b>	<b>-11.07</b>	<b>-86.3</b>
	<b>15/7/2008</b>	<b>-10.41</b>	<b>-82.1</b>
Blektjärnen (outlet)	<b>16/9/2008</b>	<b>-10.31</b>	<b>-83.6</b>
	27/7/2005	-9.97	-80.59
	25/10/2005	-9.97	-81.29
	7/9/2006	-9.7	-81.24
Blektjärnen (outlet)	<b>23/5/2008</b>	<b>-11.13</b>	<b>-86.4</b>
	<b>16/9/2008</b>	<b>-10.38</b>	<b>-83.3</b>
Snow (in the catchment of Blektjärnen)	4/4/2006	-17.05	-130.04
Vitsjön (lake)	<b>4/4/2006</b>	<b>-8.7</b>	<b>-73.9</b>
	<b>23/5/2008</b>	<b>-10.22</b>	<b>-81.4</b>
	<b>16/9/2008</b>	<b>-7.79</b>	<b>-71.8</b>
Bjurtjärnen (lake)	<b>23/5/2008</b>	<b>-12.99</b>	<b>-95.9</b>
	<b>16/9/2008</b>	<b>-12.07</b>	<b>-90.5</b>
Vikbäcken (outflow)	<b>23/5/2008</b>	<b>-12.92</b>	<b>-95.5</b>
	<b>16/9/2008</b>	<b>-12.28</b>	<b>-92.6</b>
Kullstatjärnen (lake)	<b>23/5/2008</b>	<b>-12.61</b>	<b>-93.9</b>
	<b>16/9/2008</b>	<b>-11.6</b>	<b>-89.2</b>
Kullstabäcken (outflow)	<b>23/5/2008</b>	<b>-12.56</b>	<b>-93.4</b>
	<b>16/9/2008</b>	<b>-11.42</b>	<b>-89.0</b>
Skutetjärnen (lake)	<b>23/5/2008</b>	<b>-13.05</b>	<b>-95.9</b>
	<b>16/9/2008</b>	<b>-11.62</b>	<b>-88.5</b>
Karsätjärnen (lake)	<b>23/5/2008</b>	<b>-13.33</b>	<b>-97.3</b>
	<b>16/9/2008</b>	<b>-12.19</b>	<b>-90.6</b>
Karsån (outlet)	<b>23/5/2008</b>	<b>-13.38</b>	<b>-98.3</b>
	<b>16/9/2008</b>	<b>-12.15</b>	<b>-90.8</b>
Gravbäcken (stream)	<b>23/5/2008</b>	<b>-13.54</b>	<b>-97.9</b>
	<b>16/9/2008</b>	<b>-13.34</b>	<b>-97.1</b>

precipitation and evaporation effects (Fig. 1(D) and Table 2). At Lake Blektjärnen water samples were taken from the lake and its modern outlet in April, May, July, September and October (2005, 2006, 2008) (Fig. 1(D) and Table 2).

### Laboratory work

The lithostratigraphy of the parallel sediment cores was described and cores were correlated with each other based on specific marker horizons (i.e. organic-rich layers). One core was contiguously sampled (in 1 cm segments) for terrestrial plant macrofossils, *Chara* spp. and *Pisidium* spp. shells (sample set I), another for tephra analysis (sample set II) and a third for measurements of total organic carbon and total carbon content (%TOC and %TC, respectively) (sample set III).

The isotope analysis of the water samples was undertaken at the Department of Geology and Geochemistry, Stockholm University, and at the NERC Isotope Geosciences Laboratory (NIGL), British Geological Survey. At Stockholm the water samples were analysed for  $^{18}\text{O}/^{16}\text{O}$  ratios using the  $\text{CO}_2\text{-H}_2\text{O}$  technique with a Gasbench II coupled to a Finnigan Delta plus mass spectrometer, whereas the  $^2\text{H}/^1\text{H}$  ratios were measured using the Cr oxidation method with an H-device coupled to a Finnigan MAT 252 mass spectrometer. At NIGL, the waters were equilibrated with  $\text{CO}_2$  using an ISOPREP 18

device for analysis of  $^{18}\text{O}/^{16}\text{O}$  ratios with mass spectrometry performed on a VG SIRA. For hydrogen isotope analysis, an online Cr reduction method was used with a EuroPyrOH-3110 system coupled to a Micromass IsoPrimemass spectrometer. Isotopic ratios ( $^{18}\text{O}/^{16}\text{O}$  and  $^2\text{H}/^1\text{H}$ ) are expressed as delta units,  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  (‰, parts per mil), and defined in relation to the international standard, VSMOW (Vienna Standard Mean Ocean Water). The precision of measurements is  $\pm 0.1\text{‰}$  for  $\delta^{18}\text{O}$  and  $\pm 1\text{‰}$  for  $\delta^2\text{H}$  for samples analysed at both Stockholm University and NIGL.

#### Sample set I

The first set of samples was gently wet-sieved through a 125  $\mu\text{m}$  sieve under running water. Terrestrial plant macrofossils, *Chara* spp. encrustations and *Pisidium* spp. shells were extracted by hand and examined under a binocular microscope. *Betula* seeds, catkin bracts and *Pinus* needles were selected for  $^{14}\text{C}$  measurements, cleaned by extra rinsing with distilled water, dried, weighed and sent for analysis at the accelerator mass spectrometry (AMS) unit at the Tandem Laboratory, Uppsala.

*Chara* encrustations (5–10 tubular encrustations,  $\sim 1\text{--}2\text{ mm}$  in size) were cleaned in distilled water, dried in plastic beakers and ground to a fine powder, whereas the *Pisidium* shells (two single shells per sample) were gently crushed and  $\text{H}_2\text{O}_2$  was added for 30 min to remove organic material. Methanol was added and samples were placed in an ultrasonic bath for 30 s, filtered and dried at  $50^\circ\text{C}$  overnight. The crushed *Chara* spp. and *Pisidium* spp. were weighed ( $\sim 100\ \mu\text{g}$ ) and placed in glass vials, and stable isotope measurements ( $^{18}\text{O}/^{16}\text{O}$  and  $^{13}\text{C}/^{12}\text{C}$  ratios) were conducted on a Kiel device coupled to a Finnegan MAT 252 mass spectrometer, at the Department of Geology and Geochemistry, Stockholm University. The isotope ratios are reported in conventional  $\delta$  notation (in deviations per mil) relative to the V-PDB standard. The analytical precision is  $\pm 0.1\text{‰}$  for  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  based on duplicate sample analysis and long-term precision of the laboratory standard.

#### Sample set II

The identification of cryptotephra followed the method outlined by Pilcher *et al.* (1995). Contiguous samples of 5–10 cm (sample set II) were combusted at  $550^\circ\text{C}$  for 4 h and then treated with 10% HCl. The residue was washed through a 25  $\mu\text{m}$  sieve and mounted in Canada balsam for microscopy. To determine the concentration and distribution of tephra particles more precisely, those samples in which tephra shards were detected were resampled in 1 or 2 cm slices. Samples chosen for geochemical analyses of tephra were treated in the same way but without the initial combustion. The tephra sample was prepared for electron microprobe analysis using wavelength dispersive spectrometry (WDS). The samples were analysed with a Cameca SX1000 electron microprobe equipped with five vertical WD spectrometers. Eleven major elements were measured with a counting time of 10 s. An accelerating voltage of 10 kV, a beam strength of 10 nA (3 nA for  $\text{SiO}_2$  and  $\text{Na}_2\text{O}$ ) and 5  $\mu\text{m}$  beam diameter were used.

#### Sample set III

The third set of samples was prepared for measurements of total organic carbon (TOC) and total carbon content (TC). Visible shells and roots were removed and samples were freeze-dried.

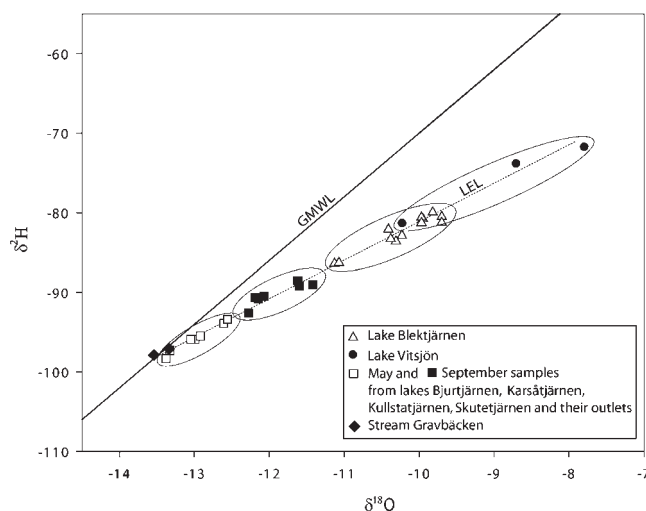
The samples were ground and combusted in pure oxygen, with the evolved  $\text{CO}_2$  content analysed by infrared radiation in an Eltra CS 500 carbon analyser, with an accuracy of  $\pm 1\%$ . The machine measures the carbon content in percentage dry weight; TOC at  $550^\circ\text{C}$  and TC at  $950^\circ\text{C}$ . The TOC content was calculated as a percentage using the molar weight of carbon of 12 and an estimated organic matter composition of  $\text{CH}_2\text{O}$ , with a molar weight of 30. The difference between %TOC and %TC was used to calculate the  $\text{CaCO}_3$  content ( $\text{CaCO}_3$ ), with a molar weight of 100, as a percentage in the sediments. The residue (i.e. non-organic compounds) was calculated as residue =  $100 - (\text{OC} \times 30/12) + (\text{CaCO}_3 \times 100/12)$  (Fig. 3).

## Results

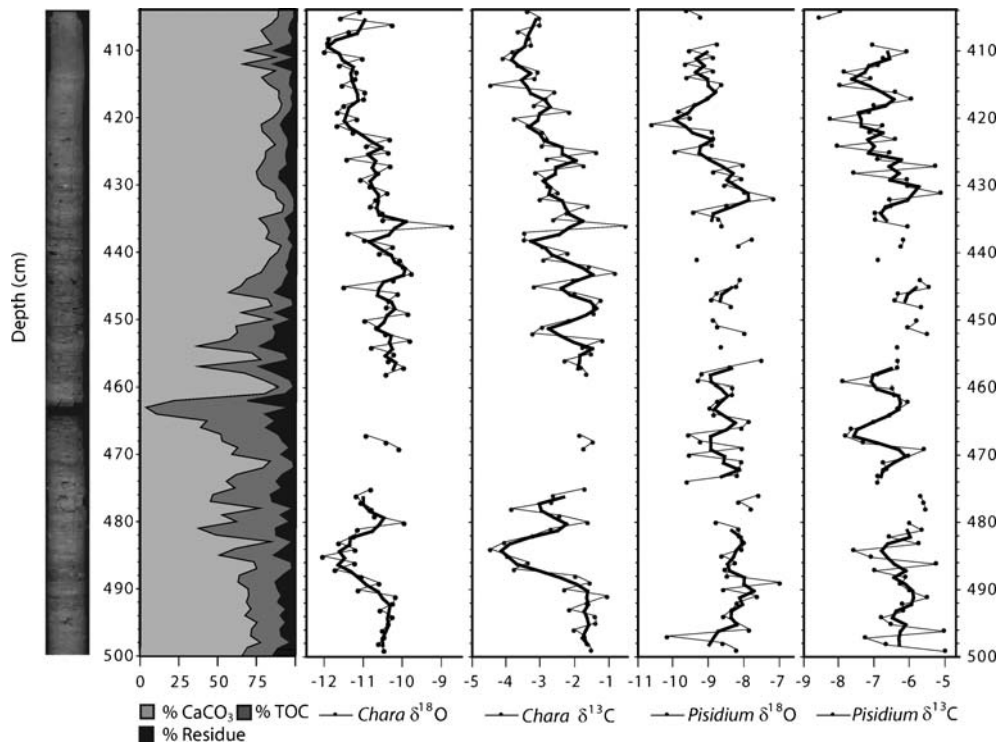
### Isotope composition of lake and stream waters in the study area

Samples of water collected from the stream Gravbäcken had  $\delta^{18}\text{O}$  values close to  $-13.5\text{‰}$  and  $\delta^2\text{H}$  values of  $-98\text{‰}$  for both May and September in 2008 and fall on the Global Meteoric Water Line (GMWL) (Fig. 2 and Table 2). Water sampled in lakes Bjurtjärnen, Karsåtjärnen, Kullstatjärnen and Skutetjärnen in May 2008 had  $\delta^{18}\text{O}$  values between  $-12.5\text{‰}$  and  $-13.5\text{‰}$  and  $\delta^2\text{H}$  values between  $-94\text{‰}$  and  $-97\text{‰}$ , which are close to the GMWL (Fig. 2). However, the  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  values of water collected in these lakes in September 2008 were slightly enriched, with  $\delta^{18}\text{O}$  values between  $-11.5\text{‰}$  and  $-12\text{‰}$  and  $\delta^2\text{H}$  values between  $-88.5\text{‰}$  and  $-90.5\text{‰}$  (Fig. 2 and Table 2).

The isotopic composition of the water samples from Lake Blektjärnen collected in April, July, September and October 2005, 2006 and 2008 has  $\delta^{18}\text{O}$  values between  $-9.7\text{‰}$  and  $-10.2\text{‰}$ , and  $\delta^2\text{H}$  values between  $-80\text{‰}$  and  $-83.6\text{‰}$  (Fig. 2 and Table 2). However, water samples from May 2008, which were taken in the lake and from its outlet, were relatively depleted, with  $\delta^{18}\text{O}$  values of  $-11\text{‰}$  and  $\delta^2\text{H}$  values of  $-86\text{‰}$ . Together, the water samples from Lake



**Figure 2**  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  of sampled lakes and their outlets. Samples taken in May are marked by open squares and samples taken in September by filled squares. Lakes Bjurtjärnen, Karsåtjärnen, Kullstatjärnen and Skutetjärnen and their outlets Vikbäcken, Karsån and Kullstabäcken (open squares), and the stream Gravbäcken (diamonds) plot on or close to the global meteoric water line (GMWL). Water samples from lakes Blektjärnen (triangles) and Vitsjön (filled circles) display enriched values and plot on a local evaporation line (LEL) (dotted line). The GMWL is shown as a black line:  $\delta^2\text{H} = 8\ \delta^{18}\text{O} + 10$  (Craig, 1965)



**Figure 3** Lithostratigraphy and sediment carbon content as percentage (organic carbon, carbonate carbon and residue), *Chara*  $\delta^{18}\text{O}$ , *Chara*  $\delta^{13}\text{C}$ , *Pisidium*  $\delta^{18}\text{O}$  and *Pisidium*  $\delta^{13}\text{C}$  vs. depth (cm) presented as raw data (thin line and dots) and three-point running mean (thick line)

Blektjärnen and its outlet plot on a Local Evaporation Line (LEL) which intersects the GMWL at a  $\delta^{18}\text{O}$  value of  $-13.5\text{‰}$  and a  $\delta^2\text{H}$  value of  $-98\text{‰}$  (Fig. 2), and these values are assumed to be the mean weighed  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  of precipitation for the region.

Measured  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  in water sampled from Lake Vitsjön in May 2008 has similar values to those from Lake Blektjärnen. However, the April 2006 and September 2008 samples were more enriched, indicating that evaporation was more predominant compared to Lake Blektjärnen, probably because of the great surface area and no direct outflow of Lake Vitsjön (Fig. 2).

### Lithostratigraphy

The sediments of Lake Blektjärnen between 500 and 404 cm depth from the lake surface, representing 4400 cal. a BP to the present, are described here. The sediments comprise partly laminated calcareous gyttja, which varies in colour (*Munsell Soil Color Chart*, 2000) from yellowish white (2.5Y 7/2) to dark brown (2.5Y 3/2) (Fig. 3). The laminae are more pronounced between  $\sim 470$  and  $430$  cm depth, whereas no laminae are found between  $\sim 415$  and  $404$  cm depth. A distinct dark-brown organic layer, composed of  $\sim 80\%$  organic carbon, is seen between  $\sim 463$  and  $462$  cm depth. Small air voids found throughout the sediment sequence ( $\sim 1$  mm in size) could be signs of bioturbation by bottom-dwelling and borrowing molluscs (Håkanson and Jansson, 1983).

The  $\text{CaCO}_3$  decreases from  $\sim 75\%$  to  $\sim 60\%$  between 500 and 470 cm depth, followed by a sharp decrease to  $\sim 4\%$  in the organic-rich layer at 463 and 462 cm depth (Fig. 3). Between 460 and 404 cm depth the values fluctuates around  $\sim 80\%$ . The TOC increases from  $\sim 20\%$  at the bottom of the section to a maximum of  $\sim 90\%$  at 463–462 cm depth. A sharp decrease follows and the values are relatively stable around 10% to the top of the core. The residue is  $<10\%$  for most of the record except between 466 and 454 cm depth, where it increases to  $\sim 20\%$  (Fig. 3).

### Isotope analyses

The *Pisidium* shells and *Chara* encrustations recovered from the sediments indicate generally good preservation conditions. However in the interval between 475 and 460 cm, *Chara* encrustations are not present and there are fewer *Pisidium* shells. *Chara*  $\delta^{18}\text{O}$  values generally vary between  $-10\text{‰}$  and  $-12\text{‰}$  (Fig. 3), the lowest values occurring between  $\sim 490$  and  $480$  cm depth, whereas values around  $-10\text{‰}$  to  $-11\text{‰}$  occur between 480 cm and 422 cm depth. Low  $\delta^{18}\text{O}$  values of  $\sim -11\text{‰}$  and  $-12\text{‰}$  follow, before a trend towards relatively higher values from 409 cm depth to the surface. *Chara*  $\delta^{13}\text{C}$  vary between  $-1.5\text{‰}$  and  $-4\text{‰}$  and follow a similar trend to that seen in the  $\delta^{18}\text{O}$  record. The lowest  $\delta^{13}\text{C}$  values ( $\sim -4\text{‰}$ ) occur between  $\sim 490$  and  $480$  cm depth, above which values are slightly higher (between 458 and 436 cm depth). Above 436 cm depth values become gradually lower until the uppermost part (above 410 cm depth) which is characterised by a trend towards higher values (Fig. 3).

The record of *Pisidium*  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  is less complete compared to that for *Chara*, because the shells are less abundant (Fig. 3). *Pisidium*  $\delta^{18}\text{O}$  values are relatively stable between 500 and 456 cm, fluctuating between  $\sim -8\text{‰}$  and  $-9\text{‰}$ . A gradual depletion to  $-10\text{‰}$  is visible between 432 and 420 cm. The  $\delta^{18}\text{O}$  values become slightly enriched in the upper 15 cm of the record. *Pisidium*  $\delta^{13}\text{C}$  values fluctuate around  $-6\text{‰}$  between 500 and 480 cm, and around  $-7\text{‰}$  between 474 and 456 cm depth. Around 430 cm depth  $\delta^{13}\text{C}$  values become gradually depleted. The uppermost 12 cm show fluctuations between  $-7\text{‰}$  and  $-8\text{‰}$  (Fig. 3).

### Chronology

The chronology is constrained by 11  $^{14}\text{C}$  dates on terrestrial plant macrofossils (Table 3) and geochemically identified tephra shards at 411–412 cm depth (Table 4) from the Askja

**Table 3** AMS radiocarbon dates for Lake Blektjärnen

Lab. no.	Depth (cm)	Material	<sup>14</sup> C age
Ua-33388	421–423	<i>Betula</i> catkin bracts and seed	985 ± 30
Ua-34567	428–430	<i>Betula</i> catkin bracts and seed	665 ± 40
Ua-34568	434–436	<i>Betula</i> catkin bracts and seeds	1835 ± 40
Ua-33389	446–448	<i>Betula</i> catkin bracts	2245 ± 30
Ua-33390	451–453	<i>Betula</i> catkin bracts and <i>Pinus</i> needles	2310 ± 30
Ua-33391	463–465	<i>Betula</i> catkin bracts and <i>Pinus</i> needle	2485 ± 30
Ua-34569	472–473	<i>Betula</i> catkin bracts and <i>Pinus</i> needle	2725 ± 40
Ua-34570	482–483	<i>Betula</i> catkin bracts	3005 ± 40
Ua-33392	486–488	<i>Betula</i> catkin bracts and <i>Pinus</i> needle	3450 ± 45
Ua-34571	493–495	<i>Betula</i> catkin bracts and seed	3310 ± 40
Ua-33393	499–500	<i>Betula</i> catkin bract and <i>Pinus</i> needle	3935 ± 55

1875 eruption (Mohn, 1878). This was the only tephra in the sequence that provided enough material for identification. All tephra shards were confined to one sample, with a concentration of ~15 shards cm<sup>-3</sup> wet sediment. This indicates that reworking, bioturbation or downward migration at this level was limited. It was only possible to analyse six shards due to the low concentration of tephra. Two of these were outliers, possibly with an origin in the Katla volcanic system (Table 4). The age model was constructed with the IntCal04 atmospheric calibration curve (Reimer *et al.*, 2004) and the Bayesian software Bpeat (Blaauw and Christen, 2005; Blaauw *et al.*, 2007), using the following prior assumptions: (i) accumulation was approximately linear within sections (in order to allow for many changes in accumulation rate, we applied this to 11 sections, or the number of dated levels minus one); (ii) the accumulation rate probably varied between 10 and 50 a cm<sup>-1</sup> (AlpaMean 30, AlphaStd 15, Epsilon 2); (iii) hiatus between sections was considered unlikely; and (iv) prior outlier probabilities for all <sup>14</sup>C dates were 5%. The calendar age of the Askja 1875 tephra was set at 75 ± 1 cal. a BP. More than 1 billion iterations were run to obtain the posterior age model and its uncertainty (Fig. 4).

## Discussion

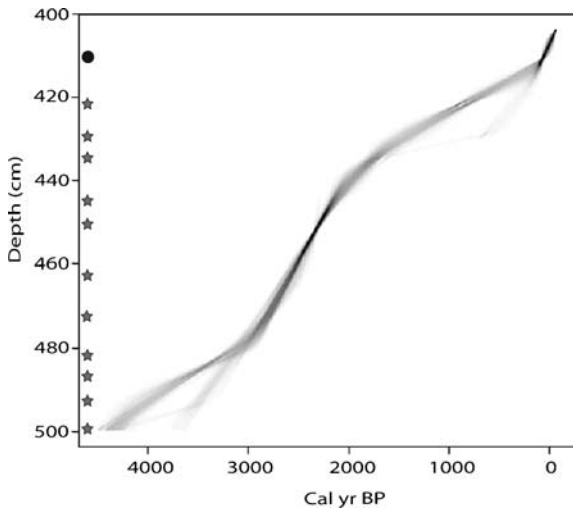
### Modern lake hydrology

The lake is infilled by groundwater seepage and direct precipitation, direct outflow is small although some water is lost from the outflow pipe and there is likely to be some

groundwater seepage. The water samples from Lake Blektjärnen show relatively stable δ<sup>18</sup>O values of around -10‰ from April to October. The lower δ<sup>18</sup>O value of ~-11‰ from May is probably due to the influence of isotopically <sup>18</sup>O-depleted snowmelt during the spring thaw (Fig. 2). The lake water δ<sup>18</sup>O is ~3‰ enriched when compared to the groundwater and does not lie on the GMWL (Fig. 2). In contrast, the stream Gravbäcken has δ<sup>18</sup>O values comparable to groundwater and lies on the GMWL at the intercept with the LEL. Gravbäcken water is therefore likely to be derived mostly from groundwater, and its position at the point of interception of the GMWL and LEL suggests it has values close to the mean weighed precipitation (Fig. 2). δ<sup>18</sup>O values for the other lakes in the study area that have both inflows and outflows (Table 1) were close to the GMWL in May 2008, which indicates that they are probably continually flushed. However, all the lakes are influenced by summer evaporation as they have more positive δ<sup>18</sup>O values in September and lie on the LEL (Fig. 2). The difference between δ<sup>18</sup>O and δ<sup>2</sup>H in the water of Lake Blektjärnen, as compared to the other sampled lakes and streams, shows that it is more strongly affected by evaporation and is not flushed by winter fresh water as the other lakes seem to be (except Lake Vitsjön). The most likely reason for this difference is the larger volume of Lake Blektjärnen (Table 1). A comparison between δ<sup>18</sup>O values in Lake Blektjärnen and δ<sup>18</sup>O in precipitation, which has been regularly monitored (1975–1980) at Bredkälén (400 m a.s.l.; 63° 50' N, 15° 18' E; ~130 km north-northeast of Lake Blektjärnen) and Forshult (192 m a.s.l.; 60° 05' N, 13° 47' E; ~430 km to the south) show that lake water δ<sup>18</sup>O is more stable throughout the ice-free season compared to δ<sup>18</sup>O in precipitation (Fig. 5), which suggests a relatively long residence time for the lake water, at

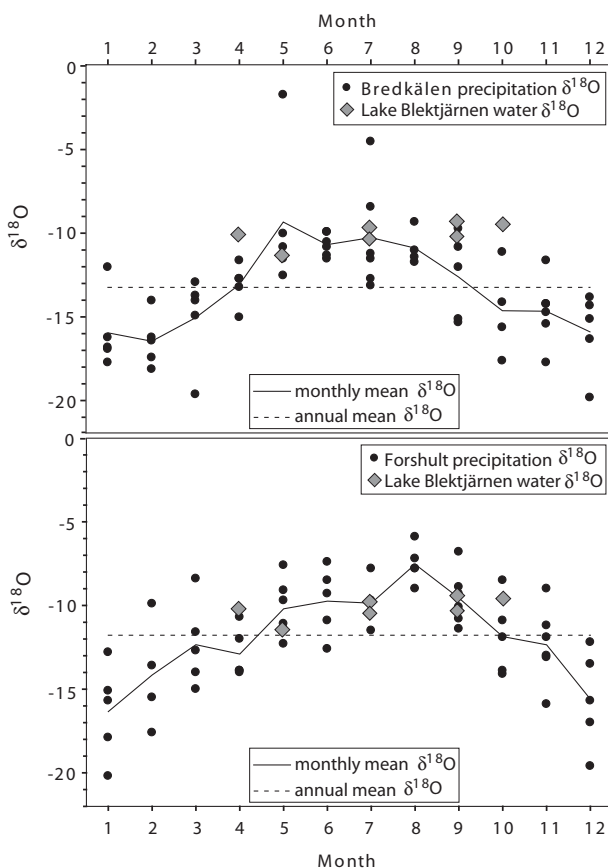
**Table 4** Major oxide percentages in tephra from Blektjärnen compared with analyses of the Askja-1875 tephra detected at other sites in Sweden and Norway (Bergman *et al.*, 2004; Pilcher *et al.*, 2005; Borgmark and Wastegård, 2008)

	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	FeO <sub>tot</sub>	MnO	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O	P <sub>2</sub> O <sub>5</sub>	Cl	Total	Correlation
Blektjärnen, 411 cm	73.93	0.67	12.28	3.07	0.10	0.65	2.54	3.31	2.31	0.13	0.06	99.05	Askja-1875
	71.13	0.70	12.06	3.43	0.22	0.65	2.52	3.56	2.30	0.13	0.19	96.90	Askja-1875
	72.75	0.92	12.37	3.56	0.02	0.70	2.69	3.27	2.28	0.20	0.04	98.80	Askja-1875
	71.54	0.84	12.11	3.34	0.09	0.72	2.61	3.22	2.41	0.16	0.09	97.13	Askja-1875
Mean (n = 10)	72.34	0.79	12.21	3.35	0.11	0.68	2.59	3.34	2.33	0.15	0.09	97.97	
1σ	1.26	0.12	0.14	0.21	0.08	0.04	0.08	0.11	0.06	0.03	0.07		
	69.65	0.34	13.27	3.47	0.18	0.21	1.45	4.21	2.47	0.03	0.19	95.48	Outlier (Katla?)
	70.38	0.28	13.32	3.85	0.15	0.20	1.35	4.57	3.02	0.01	0.20	97.33	Outlier (Katla?)
Klocka Bog, mean (n = 16)	72.19	0.80	12.33	3.65	0.09	0.76	2.61	3.11	2.41	–	–	97.94	
1σ	0.98	0.06	0.21	0.39	0.03	0.10	0.29	0.30	0.15	–	–		
Fågelmossen, mean (n = 10)	70.81	0.77	12.32	3.35	0.09	0.68	2.43	3.85	2.42	0.14	–	96.86	
1σ	1.19	0.04	0.13	0.27	0.02	0.05	0.19	0.35	0.12	0.02	–		
Lofoten, mean (n = 15)	72.70	0.83	12.71	3.46	–	0.70	2.51	3.71	2.34	–	–	99.02	
1σ	1.15	0.06	0.20	0.33	–	0.09	0.29	0.15	0.10	–	–		



**Figure 4** Grey-scale graph of the computed age model. Dark-grey areas correspond to sections where the chronology is well constrained, while light-grey sections indicate higher uncertainty. The age model is constrained by 11  $^{14}\text{C}$  dates (indicated by stars) and the Askja-1875 tephra (indicated by a black dot) and was constructed with the IntCal04 calibration curve (Reimer *et al.*, 2004) and the Bayesian software Bpeat (Blaauw and Christen, 2005; Blaauw *et al.*, 2007) using prior assumptions (see text)

least more than 1 year. Thus the modern lake water data for Lake Blektjärnen show that the  $\delta^{18}\text{O}$  values of lake water are mainly controlled by the balance between evaporation and input of precipitation, which is expressed as the evaporation to input ratio ( $E/I$ ).

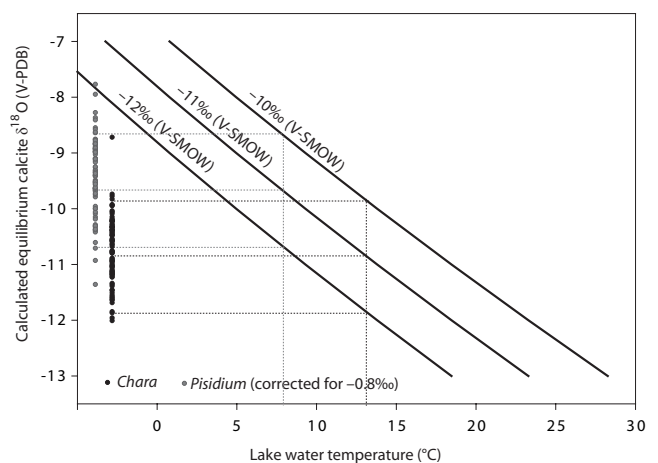


**Figure 5** Precipitation  $\delta^{18}\text{O}$  values measured at Bredkälén and Forshult (1975–1980) compared to the lake water samples of Lake Blektjärnen (diamonds) (presented in Table 2). The data for Bredkälén and Forshult were obtained from the GNIP/WMO database (2008)

## Past changes in *Chara* and *Pisidium* $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$

$\delta^{18}\text{O}$  values of *Chara* and *Pisidium* depend on the temperature and isotopic composition of the lake water. An increase in water temperature results in lower  $\delta^{18}\text{O}$  due to mineral–water fractionation of  $\sim -0.24\text{‰ } ^\circ\text{C}^{-1}$  (Craig, 1965). This effect is counteracted by contrasting isotopic change in meteoric water due to changes in condensation temperature, which is  $\sim +0.6\text{‰ } ^\circ\text{C}^{-1}$  (Dansgaard, 1964), assuming the Dansgaard relationship for northwestern Europe. Therefore, if temperature increases by  $1^\circ\text{C}$  the  $\delta^{18}\text{O}$  of the carbonates change by  $+0.36\text{‰}$  ( $-0.24\text{‰} + 0.6\text{‰}$ ) (Leng and Marshall, 2004). Calculated equilibrium calcite values (in V-PDB) using mean air temperatures as estimates for water temperatures (Livingstone and Lotter, 1998) and lake water  $\delta^{18}\text{O}$  compositions of  $-10$  to  $-12\text{‰}$  (in V-SMOW) (Fig. 6) show that, despite a constant lake water  $\delta^{18}\text{O}$  value, relatively enriched equilibrium calcite can be yielded if it is formed when temperatures are lower. As expected, relatively depleted equilibrium calcite will be precipitated if calcite is precipitated when temperatures are relatively higher. The isotopic composition of Lake Blektjärnen water is determined by the isotopic composition and amount of input water and evaporation (as described above). Shifts in the isotopic composition of the precipitation reflect changes in regional temperatures and precipitation patterns that follow upon changes in atmospheric circulation dynamics. The change of the isotopic composition of lake water due to evaporation can be very significant, often swamping any per mil change in  $\delta^{18}\text{O}$  due to temperature.

All these effects need to be considered when comparing *Chara* and *Pisidium*  $\delta^{18}\text{O}$  records because of the difference in the actual length of their respective growth season. *Pisidium*  $\delta^{18}\text{O}$  values are probably influenced by lower lake temperatures and more depleted (spring–autumn) lake water. This is because of their longer active season that lasts from late May until early November when the mean air temperature is ca  $8^\circ\text{C}$ ,



**Figure 6** y-Axis indicates calculated equilibrium calcite  $\delta^{18}\text{O}$  values (V-PDB) depending on the lake water temperature (x-axis) and isotopic composition of the lake water (V-SMOW, bold lines). Mean air temperatures,  $\sim 8^\circ\text{C}$  during late May to early November and  $\sim 13^\circ\text{C}$  for June–July, are used as estimates for water temperatures. The measured *Chara* (black circles) and *Pisidium* (grey circles)  $\delta^{18}\text{O}$  values are indicated to the left in the graph. A  $\delta^{18}\text{O}$  offset of  $\sim -0.8\text{‰}$  for aragonite in *Pisidium* shells is removed in the figure, since it is independent of temperature (von Grafenstein *et al.*, 1999; Kim *et al.*, 2007). Equilibrium calcite values were calculated using  $T^\circ\text{C} = 13.8 - 4.58(\delta\text{c} - \delta\text{w}) + 0.08(\delta\text{c} - \delta\text{w})^2$  (Leng and Marshall, 2004), where  $\delta\text{c}$  = carbonate  $\delta^{18}\text{O}$  (V-PDB), and  $\delta\text{w}$  = water  $\delta^{18}\text{O}$  (V-SMOW)



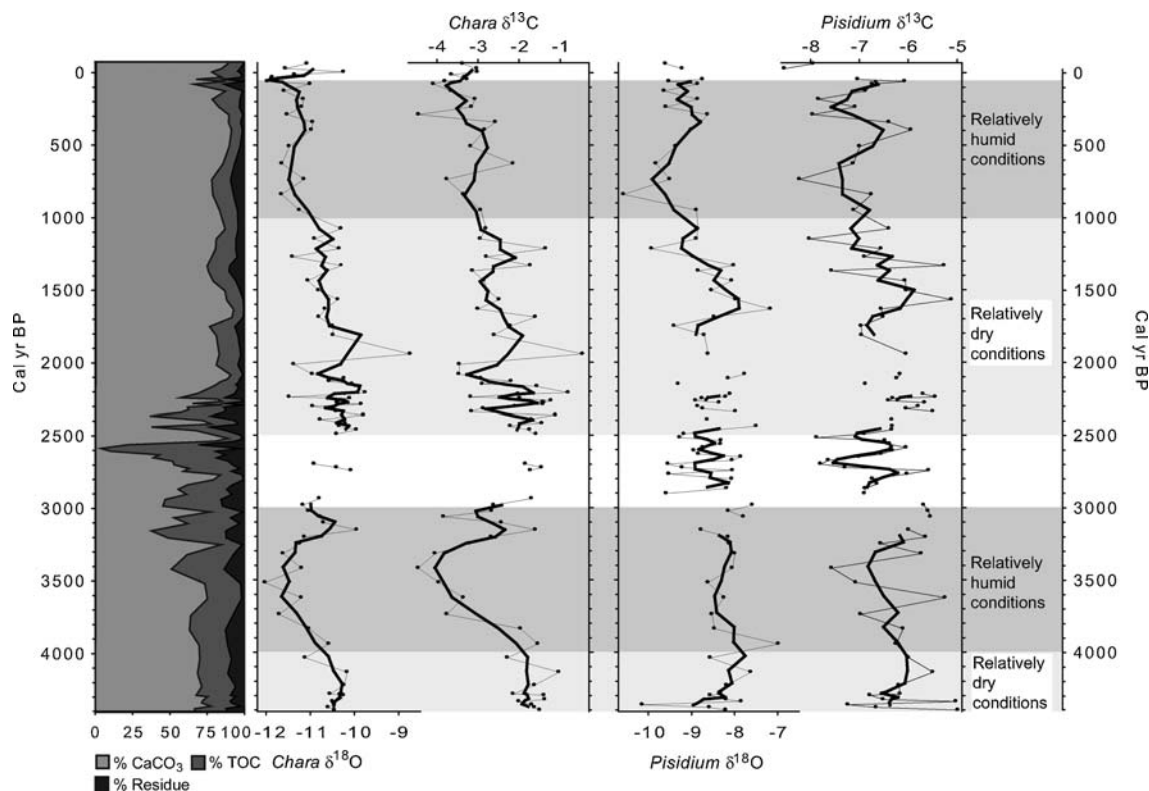
whereas *Chara* mainly photosynthesise during June–July when the mean air temperature is  $\sim 13^{\circ}\text{C}$  (Fig. 6) (cf. Hammarlund *et al.*, 2002). This difference of  $5^{\circ}\text{C}$  corresponds to a change of  $\sim 1.2\%$  in  $\delta^{18}\text{O}$  and probably explains some of the difference between the *Chara* and *Pisidium*  $\delta^{18}\text{O}$  records. The isotopic composition of the lake water will also be more negative early in the season due to the influence of  $^{18}\text{O}$ -depleted snowmelt, which would further deplete the  $\delta^{18}\text{O}$  of *Pisidium*.

The cycling of carbon in lakes is complex and involves several steps where fractionation can occur, which complicates the interpretation of carbon isotopic signals. Dissolved inorganic carbon (DIC) in a lake contains a number of carbon species (Kelts and Hsü, 1978) and the  $\delta^{13}\text{C}$  in the DIC pool can vary by several per mil during a year due to contribution from sources including soil-derived  $\text{CO}_2$  ( $\delta^{13}\text{C} \sim -25\%$ ; Hoefs, 2004), dissolution of local bedrock ( $\delta^{13}\text{C} \sim -2$  to  $+2\%$ ; Sturkell *et al.*, 1998), aquatic production (discrimination against  $^{13}\text{C}$  by photosynthesising plants; Hollander and McKenzie, 1991), deposition and breakdown of organic material (deposition removes carbon from the system and breakdown releases  $^{12}\text{C}$ -enriched  $\text{CO}_2$ ; e.g. Dettman *et al.*, 1999; Teranes *et al.*, 1999; Myrbo and Shapley, 2006). During the summer months when lakes are actively evaporating, the lake water also exchanges  $\text{CO}_2$  with the isotopically heavier atmospheric  $\text{CO}_2$ , resulting in  $\delta^{13}\text{C}$  enrichment in DIC when residence times are longer (Li and Ku, 1997). The covariance often seen between  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  in lacustrine carbonates is commonly attributed to evaporation and atmospheric exchange, which often occur concurrently.

*Chara* algae convert bicarbonate,  $\text{HCO}_3^-$ , into two  $\text{CO}_2$  molecules by proton pumping (McConnaughey, 1991). Kinetic fractionation occurs as the lighter molecule is preferentially used for photosynthesis and the heavier is used to precipitate calcite (McConnaughey, 1991; McConnaughey and Falk,

1991; Hammarlund and Lemdahl, 1994; Hammarlund *et al.*, 1997). von Grafenstein *et al.* (2000) suggested a kinetic effect on *Chara*  $\delta^{13}\text{C}$  of  $\sim 4\%$ , compared to the  $\delta^{13}\text{C}$  composition of DIC. For *Pisidium*, however, it has been suggested that they build their shells in or close to isotopic equilibrium with DIC in water (Stuiver, 1968; Fritz and Poplawski, 1974; McConnaughey *et al.*, 1997; von Grafenstein *et al.*, 1999, 2000). The recorded difference between mean  $\delta^{13}\text{C}$  values in *Chara* ( $\sim -2.5\%$ ) and *Pisidium* ( $\sim -6.5\%$ ) from Lake Blektjärnen of  $\sim 4\%$  is consistent with the previously suggested offsets for kinetic effects in *Chara*, assuming that *Pisidium* precipitates  $\delta^{13}\text{C}$  in equilibrium with DIC (Fritz and Poplawski, 1974; Hammarlund *et al.*, 1997; von Grafenstein *et al.*, 1999, 2000).

Overall, the Lake Blektjärnen isotope record shows similar  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  trends (Figs 3 and 7). We argue that this is because increased evaporation and enhanced vapour exchange led to preferential loss of lighter isotopes, which resulted in elevated values of  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  during periods with relatively drier climate (Talbot, 1990; Li and Ku, 1997). The correlation coefficient for covariance is large ( $r^2 = 0.54$ ) between  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values in *Chara*, indicating that past changes in the isotopic composition of Lake Blektjärnen in the summer is controlled to a large degree by variation in the  $E/I$  ratio. If so, more negative  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values would imply more humid conditions. The *Pisidium* records display small covariance ( $r^2 = 0.22$ ) between  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values, which might be explained to some extent by their relatively longer active season and different habitat. A long season has implications for the  $\delta^{13}\text{C}$  compositions as *Pisidium* experiences variable  $\delta^{13}\text{C}$  values due to seasonally changing rates of productivity and decomposition in the lake. Also the *Pisidium*  $\delta^{13}\text{C}$  composition is likely to be more readily affected by oxidation of organic matter at the sediment–water interface because of their preferred habitat (and burrowing) in deeper



**Figure 7** (A) Comparison of sediment carbon content (% TOC,  $\text{CaCO}_3$  and residue) and  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  in *Chara* and *Pisidium* raw data (dots and thin line) and three-point running mean (thick black line) vs. calendar years BP. Four intervals of humidity changes inferred from Lake Blektjärnen records are indicated: dark grey denotes relatively high humidity between ca. 4000–3000 and 1000–50 cal. a BP, and light grey indicates relatively dry conditions between ca. 4400–4000 and 2500–1000 cal. a BP

sediments where decomposition processes often elevate the concentration of  $^{13}\text{C}$ -depleted  $\text{CO}_2$ .

## Palaeoclimatic significance

### Prior to 4000 cal. a BP

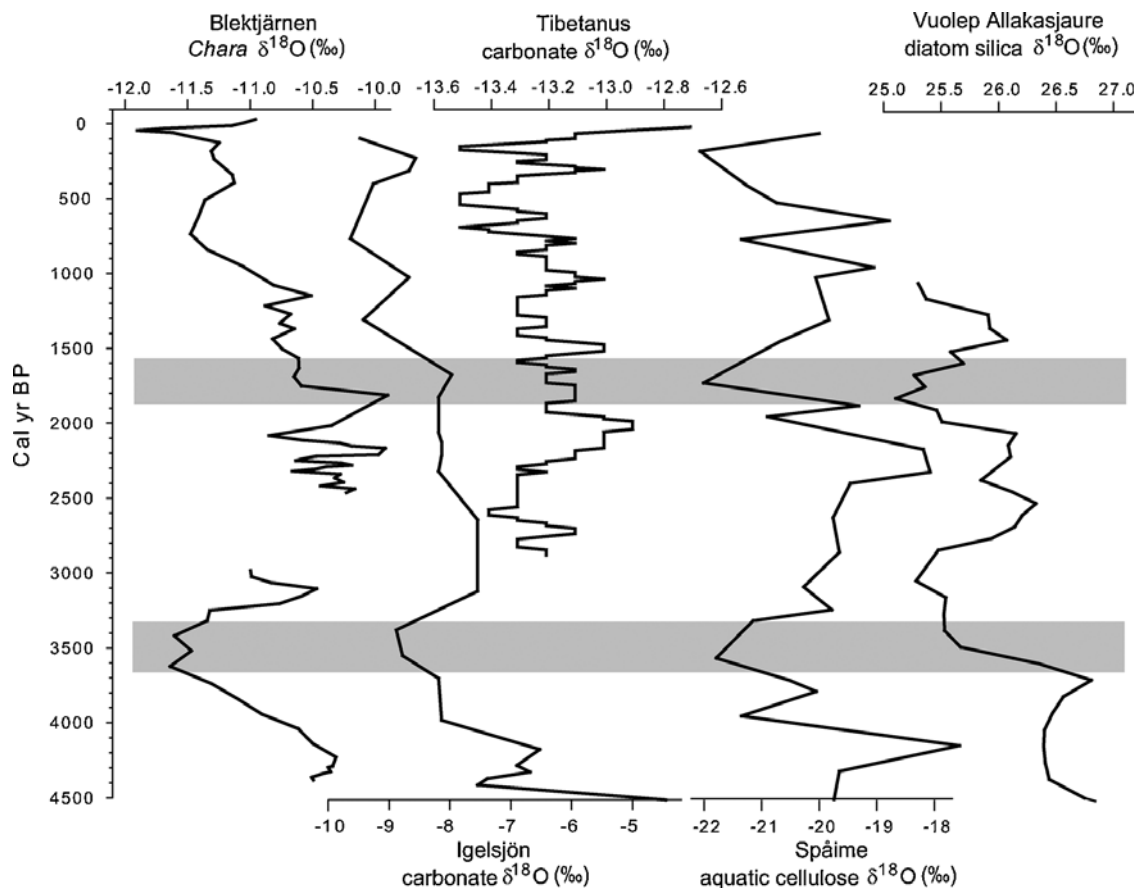
*Chara*  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values are relatively high between ca. 4400 and 4000 cal. a BP. Given the strong evaporation dependence of Lake Blektjärnen we suggest that the *E/I* ratio (the balance between evaporation and input water) was relatively high due either to low precipitation or higher temperatures, or a combination of these factors (Fig. 7). Pollen-based temperature reconstructions show that climate was warmer in Scandinavia from ca. 8000 to ca. 4000 cal. a BP (Seppä *et al.*, 2005; Antonsson and Seppä, 2007). A warmer climate could have favoured the aquatic productivity illustrated by the relatively high TOC content (Fig. 7).

### Between ca. 4000 and 3000 cal. a BP

Isotope minima in *Chara*  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  occur at ca. 3500 cal. a BP; the negative trend begins around 4000 cal. a BP (Fig. 7). We

suggest that precipitation increased, suppressing evaporation (which increases  $\delta^{18}\text{O}$ ) and atmospheric equilibration (which increases  $\delta^{13}\text{C}$ ) through input of  $^{18}\text{O}$ -depleted rainwater and shortened lake water residence time (lower *E/I* ratio). Somewhat cooler summer conditions ( $<0.5^\circ\text{C}$ ) have been inferred based on chironomids (Velle *et al.*, 2005) and pollen records (Seppä *et al.*, 2009) at this time. Lower temperatures probably changed the isotopic composition of precipitation towards more negative  $\delta^{18}\text{O}$  values; however, temperature alone could not have caused the isotopic shift of 1‰ in *Chara*  $\delta^{18}\text{O}$  because a lowering of as much as  $3^\circ\text{C}$  would be required to account for the  $\delta^{18}\text{O}$  trend (assuming  $+0.36\text{‰ }^\circ\text{C}^{-1}$ ; Leng and Marshall, 2004).

Isotope minima have also been recorded at ca. 3500 cal. a BP in the  $\delta^{18}\text{O}$  carbonate record from Lake Igelsjön (Hammarlund *et al.*, 2003; Jessen *et al.*, 2005), in the  $\delta^{18}\text{O}$  cellulose record from Lake Spåime (located only  $\sim 60\text{ km}$  west of Lake Blektjärnen) (St Amour, 2009) and in the  $\delta^{18}\text{O}$  diatom record from Lake Vuolep Allakasjaure, northern Sweden (Rosqvist *et al.*, 2004) (Fig. 8). Lake Spåime and Lake Vuolep Allakasjaure are both high-altitude through-flow lakes and the isotopic composition of the lake waters are therefore mainly influenced by the amount and seasonal distribution of rainfall and the resulting lower isotopic composition of the precipitation (Jonsson *et al.*, 2009; St Amour, 2009; Jonsson *et al.*, 2010). Geochemical and vegetation data from Lake Igelsjön (Hammarlund *et al.*, 2003; Jessen *et al.*, 2005) further support the



**Figure 8** Comparison of *Chara*  $\delta^{18}\text{O}$  three-point running mean values with previously published  $\delta^{18}\text{O}$  carbonate record from Lake Igelsjön (Hammarlund *et al.*, 2003) with revised chronology (Jessen *et al.*, 2005), carbonate  $\delta^{18}\text{O}$  values from Lake Tibetanus (five-point running mean) (Rosqvist *et al.*, 2007), aquatic cellulose  $\delta^{18}\text{O}$  from Lake Spåime (St Amour, 2009) and diatom  $\delta^{18}\text{O}$  from Lake Vuolep Allakasjaure (Rosqvist *et al.*, 2004). Lake Blektjärnen and Lake Igelsjön are evaporative and thus reflect changes in *E/I* ratio. Lake Tibetanus is mainly groundwater fed and reflects changes in isotopic composition of mean annual precipitation, whereas the isotopic composition of the lake waters in Lake Spåime and Lake Vuolep Allakasjaure mainly reflect the amount and seasonal distribution of precipitation. A diatom  $\delta^{18}\text{O}$  value of  $+25\text{‰}$  indicates larger contribution of relatively  $^{18}\text{O}$ -depleted precipitation, whereas a value of  $+27\text{‰}$  indicates larger contribution of relatively  $^{18}\text{O}$ -enriched precipitation (Jonsson *et al.*, 2010). The marked areas indicate depleted  $\delta^{18}\text{O}$  values inferred from these lakes indicating wetter/cooler conditions around 3500 cal. a BP, and the onset of the successive depletion trend in  $\delta^{18}\text{O}$  values around 1750 cal. a BP inferred from the records from Lake Blektjärnen. Note the different scales

inference that there was a major shift towards wetter and/or cooler climate at this time in southern Sweden and evidence for a glacier advance is reflected in the geochemical record in Lake Vuolep Allakasjaure (Rosqvist *et al.*, 2004). The similar isotope response of the proxy records in these environments suggests that the change was significant and regional. Today increased humidity in Scandinavia results from an intensified westerly air flow, and we therefore suggest that zonal flow was intensified and was the dominating atmospheric circulation pattern during the period.

Unfortunately, there are periods with no *Chara* encrustations, and therefore the isotopic record is incomplete between ca. 3000 and 2500 cal. a BP (Fig. 7). Growing or preservation conditions must have changed, possibly due to a climate shift. Increased aquatic productivity due to warmer summers could have decreased light penetration, which in turn inhibited photosynthesis in *Chara*. The increased aquatic productivity could also have resulted in increased decomposition, which dissolved *Chara* calcite through elevated CO<sub>2</sub> levels (Dean, 1999).

### Between ca. 2500 and 1000 cal. a BP

Values for *Chara* δ<sup>18</sup>O and δ<sup>13</sup>C are relatively enriched between ca. 2500 until 1000 cal. a BP, indicating that the *E/I* ratio increased compared to the previous period, either due to decreased input of winter/spring precipitation, higher temperatures and a relatively larger contribution of more positive summer precipitation, or a combination of these factors (Fig. 7). Again, a change in δ<sup>18</sup>O forced by temperature alone seems unlikely as an increase of several degrees would be required. Reconstructions based on chironomids and pollen reveal relatively stable temperatures (Velle *et al.*, 2005; Seppä *et al.*, 2009). Therefore we argue that the winter/spring and summer climate must have been drier and the lake water residence time longer to allow evaporation and atmospheric equilibration to enrich the lake water in δ<sup>18</sup>O and δ<sup>13</sup>C respectively. Increased evaporation during this time period can also be inferred from the Lake Igelsjön isotope data (Fig. 8) (Hammarlund *et al.*, 2003). Drier summers are probably associated with a dominance of an anticyclonic circulation pattern (Seppä *et al.*, 2009) and we suggest that this forced the higher isotope values through this period.

From 1750 cal. a BP onwards we note that both the *Chara* and *Pisidium* isotope values (for *Pisidium* until 700 cal. a BP only) become progressively more negative (Fig. 7). A negative trend at this time has also been detected in isotope records from northern Sweden (Fig. 8) (Shemesh *et al.*, 2001; Rosqvist *et al.*, 2007; Jonsson *et al.*, 2010), again showing a common response to atmospheric forcing. This suggests that climate became wetter even at these relatively more continental sites (i.e. at Lake Blektjärnen, Lake Tibetanus and Lake 850 (Shemesh *et al.*, 2001) not included in Fig. 8). The steadily decreasing δ<sup>18</sup>O values in Lake Igelsjön from ca 3000 cal. a BP onwards (Hammarlund *et al.*, 2003) suggest that south Sweden possibly was influenced by increased strength of zonal flow already from this time.

### After ca. 1000 cal. a BP

The *E/I* ratio in Lake Blektjärnen remained low after ca. 1000 cal. a BP, suppressing δ<sup>18</sup>O and δ<sup>13</sup>C in the lake water. A decrease in the δ<sup>18</sup>O of precipitation is expected as temperatures were decreasing at this time (to ~0.5°C lower

compared to 1961–1990 average temperature; Moberg *et al.*, 2005) and compared to the Holocene mean (Velle *et al.*, 2005). However, this change is too small to account for the relatively large shift detected in *Chara* δ<sup>18</sup>O. Therefore it is likely that climate was generally humid especially prior to ca. 500 cal. a BP and during the last 100 a. Humid conditions during the last 1000 a were also inferred from the Lake Igelsjön data (Hammarlund *et al.*, 2003). The isotope records from Lake Tibetanus and Lake Spåime also show a negative trend over the past 1000 a, again indicating a common response to regional shifts in circulation (Fig. 8). Wetter conditions in northern Europe, caused by a persistent positive North Atlantic Oscillation (NAO) mode, have been reconstructed for the period 1000–500 cal. a BP (Trouet *et al.*, 2009).

The *Chara* δ<sup>18</sup>O values are slightly enriched after ca. 500 cal. a BP, until ca. 100 cal. a BP, suggesting that climate was drier (Fig. 8). Enriched carbonate δ<sup>18</sup>O values are also seen at this time in the Lake Igelsjön (Hammarlund *et al.*, 2003) and Lake Tibetanus (Rosqvist *et al.*, 2007) records (Fig. 8). This suggests that climate was both dry (our data) and cool (Moberg *et al.*, 2005; Velle *et al.*, 2005). Drier conditions in northern Europe during this period were also inferred from reconstructed weaker NAO conditions (Trouet *et al.*, 2009).

The most negative value in the Lake Blektjärnen record occurs at AD ca. 1900, after which values increase again. A similar isotope minimum occurred at Lake Tibetanus at this time (Rosqvist *et al.*, 2007), indicating that the isotope hydrology of these lakes responded to the large climate shifts that occurred at the end of the Little Ice Age in northern Fennoscandia (Fig. 8).

## Conclusions

Stable isotopes on *Chara* calcite and *Pisidium* shells in Lake Blektjärnen in central Sweden are used to reconstruct past changes in *E/I* ratio over the last ca. 4400 a. Modern lake water isotope data show that Lake Blektjärnen has a relatively high *E/I* ratio, which is probably the result of relatively high summer temperatures, enriched summer precipitation and relatively small contribution of depleted winter/spring precipitation. The relatively high *Chara* δ<sup>18</sup>O and δ<sup>13</sup>C values between 4400 and 4000 cal. a BP suggest that conditions may have been similar to today. From the low *Chara* δ<sup>18</sup>O and δ<sup>13</sup>C values between 4000 and 3000 cal. a BP, especially around 3500 cal. a BP, we infer that precipitation increased and temperatures decreased. Relatively enriched *Chara* δ<sup>18</sup>O and δ<sup>13</sup>C values between 2500 and 1000 cal. a BP indicate that climate became drier and warmer, probably as a result of dominance of an anticyclonic atmospheric circulation. However, a depletion trend starts after ca. 1750 cal. a BP, which has also been detected in several other δ<sup>18</sup>O records (carbonate and diatom) in Sweden, suggesting increasingly wetter conditions. Our data show that climate was wetter and probably cool between ca. 1000 and 500 cal. a BP and at the end of the 19th century, i.e. at the end of the Little Ice Age. Between ca. 500 and 100 cal. a BP summers were probably both dry and cool. Our results show the potential of using lacustrine stable isotope data to reconstruct past changes in humidity during summer (*Chara*) and during the ice-free season (*Pisidium*) at a continental site. Such information is crucial for the identification of past shifts in atmospheric circulation patterns.

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