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Reconstructing past atmospheric circulation changes using oxygen isotopes in lake sediments from Sweden

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Abstract

Here we use lake sediment studies from Sweden to illustrate how Holocene-aged oxygen isotope records (from lakes located in different hydrological settings) can provide information about climate change. In particular changes in precipitation, atmospheric circulation and water balance. We highlight the importance of understanding the present and past lake hydrology, and the relationship between climate parameters and the oxygen isotopic composition of precipitation ($\delta^{18}O_p$) and lake waters ($\delta^{18}O_{lakewater}$) for interpretation of the oxygen isotopic record from the sediments ($\delta^{18}O$). Both precipitation reconstructions from northern Sweden and water balance reconstructions from south and central Sweden show that the atmospheric circulation changed from zonal to a more meridional air flow over the Holocene. Superimposed on this Holocene trend are $\delta^{18}O_p$ minima resembling intervals of the negative phase of the North Atlantic Oscillation (NAO), thus suggesting that the climate of northern Europe is strongly influenced by atmospheric and oceanic circulation changes over the North Atlantic.

1 Introduction

For interpretation of anthropogenic climate changes, an understanding of natural climate variability is important (e.g. Mayewski et al., 2004; IPCC, 2007; Jones et al., 2009). Globally distributed climate proxy records reconstructing different climate variables demonstrate that Holocene climate variations have, although weaker in amplitude than the dramatic shifts of the glacial cycle, been large enough to have significant effects on ecosystems and humans (e.g. Mayewski et al., 2004; Wanner et al., 2008). Proxy-based reconstructions can help to evaluate climate model responses and improve the understanding of important mechanisms and feedbacks (Fricke and O'Neil, 1999; Sturm et al., 2005; Schmidt et al., 2007; Jones et al., 2009). Due to the intimate link between oxygen isotopic composition of precipitation ($\delta^{18}O_p$) and climate parameters (Rozanski et al., 1993; Gat et al., 1996), it is expected that changes in $\delta^{18}O_p$ will be transferred to the isotopic composition of lake waters ($\delta^{18}O_{lakewater}$) and recorded in lake sediments $\delta^{18}O$ (Darling et al., 2004; Leng and Marshall, 2004; Jonsson et al., 2009b). Therefore, $\delta^{18}O$ records from lake sediments improve our understanding about Holocene climate change and variability.

The global distribution of the oxygen and hydrogen isotopic composition in precipitation ($\delta^{18}O_p$ and δ^2H_p) is in general more negative with increasing distance from the equator (rain-out effect), from coastal regions (continental effect) and increasing elevation (altitude effect) (Rozanski et al., 1993). At mid and high latitudes there is a seasonal variation in $\delta^{18}O_p$ and δ^2H_p driven by temperature, resulting in relatively depleted winter precipitation compared to summer precipitation. Factors other than temperature influence local $\delta^{18}O_p$ over time, for example at high altitudes a change in the position of boundaries between air masses play a predominate role in determine $\delta^{18}O_p$ (Edwards et al., 1996; Shemesh et al., 2001; Rosqvist et al., 2004, 2007).

Knowledge about the factors that have influenced $\delta^{18}O_{lakewater}$ is essential for the interpretation of the sediment $\delta^{18}O$ signal (Leng and Marshall, 2004). For any given lake, $\delta^{18}O_{lakewater}$ will depend on the hydrological balance between recharge water (groundwater, direct precipitation, surface and stream inflows) and outputs (groundwater loss, evaporation, surface and stream outflows) (Gibson et al., 1999; Leng et al., 2005). Therefore, lakes need to be identified that have the potential to accurately record specific aspects of climate change and environmental variations (cf. Leng and Anderson, 2003; Leng and Marshall, 2004). The climate of North Europe is strongly influenced by atmospheric and oceanic circulation changes over the North Atlantic (Luterbacher et al., 2002) and therefore lakes in Fennoscandia have the potential of retaining many different aspects of water isotope composition in their sediments which can be used for palaeoclimate reconstruction. This may be annual $\delta^{18}O_p$, seasonally specific $\delta^{18}O_p$ or changes in lake water budget determined by the evaporation to inflow ratio (E/I).

Here we illustrate how Holocene-aged $\delta^{18}O$ sediment records, from lakes located in different hydrological settings in Sweden, can provide information about different aspects of climate change, such as changes in precipitation pattern, atmospheric circulation and water balance. We highlight the importance of understanding the modern and past lake hydrology and its relationship with climate parameters in order to interpret the $\delta^{18}O$ sediment signal and provide a reliable reconstruction of climate change.

2 Modern climate

The climate in Sweden is strongly influenced by oceanic and atmospheric circulation over the North Atlantic and the Scandinavian mountain range (Fig. 1). Precipitation is closely related to the passage of cyclones that normally follow the westerly-easterly track across Scandinavia (Ångström, 1974). The highest amount of precipitation in Sweden is found along the west coast and in the Scandinavian mountain range (1961–1990; Alexandersson and Andersson, 2004) (Fig. 2). Precipitation distribution is strongly influenced by topography (Johansson and Chen, 2003). On the windward west side of the Scandes, forced lifting of approaching air masses over the mountain causes the release of rainfall and an increase in precipitation with elevation (orographic rain). The eastern leeward side of the Scandes is often drier with a more continental climate (Smith, 1979). Strong zonal air flow with strong westerly winds (positive North Atlantic Oscillation (NAO) index) and increased cyclonic frequency across the North Atlantic leads to high amounts of winter precipitation and higher winter air temperatures in Fennoscandia (Chen and Hellström, 1999; Chen, 2000; Jacobeit et al., 2001; Marshall et al., 2001). In contrast, cold and dry winters occur when westerly winds in the North Atlantic region are weak (negative NAO index) and meridional air flow brings cold polar air south over Fennoscandia. Additionally, during negative NAO winters southeast airflow could bring moisture to Sweden from the Baltic Sea (Uvo, 2003). Based on instrumental temperature data with information on cloud amount, meridional geostrophic wind and air pressure, Moberg et al. (2003) showed that low summer temperatures over Scandinavia are associated with dominance of cyclonic circulation (cool and wet conditions), and conversely, high summer temperatures with dominance of anticyclonic circulation (warm and dry conditions).

At mid to high latitudes there is a good correlation between annual surface temperature and the weighted oxygen isotope composition of precipitation, where higher temperatures correspond to higher $\delta^{18}O_p$ values (Dansgaard, 1964; Rozanski et al., 1993; Fricke and O'Neil, 1999). The present day relationship between air temperature (T in °C) and $\delta^{18}O_p$ for North Atlantic coastal stations is $\delta^{18}O_p$=0.69T−13.6 (the Dansgaard relationship; Dansgaard, 1964). Monthly $\delta^{18}O_p$ values from metrological stations in Sweden are available from the Global Network of Isotopes in Precipitation data set (GNIP, IAEA-WMO). The mean annual $\delta^{18}O_p$ values (1975–1980) from 17 GNIP stations have been used by Burgman et al. (1987) to infer isotopic isolines over Sweden (Fig. 2). The annual mean $\delta^{18}O_p$ over Sweden decreases with increasing distance from the ocean and with increasing latitude/altitude due to continental and altitude effects. Monthly $\delta^{18}O_p$ values from a few individual GNIP stations in Sweden are also shown in Fig. 2, together with monthly mean temperature and precipitation from nearby metrological stations (Table 1) (1961–1990; Alexandersson and Eggertsson Karlström, 2001). Seasonal variations in temperature are responsible for the seasonal variation observed in monthly mean $\delta^{18}O_p$ with higher values in summer months than in winter months. The monthly mean $\delta^{18}O_p$ values show that the highest $\delta^{18}O_p$ values and smallest amplitude are found on the west coast while the lowest $\delta^{18}O_p$ values and largest amplitudes are observed in the continental areas in the northern part of the country. An exception is the north westernmost station in Abisko, separated from the North Atlantic by relatively low mountains (Burgman et al., 1987), and therefore show low seasonal $\delta^{18}O_p$ amplitude.

In addition to regional condensation temperature, $\delta^{18}O_p$ is a function of the conditions at the vapor region (generally the ocean) and the air-mass trajectory (vapor transport history). For example, evaporation from the oceans at higher latitudes can modify the $\delta^{18}O$ in air masses moving in from lower latitudes by introducing local moisture with relatively high $\delta^{18}O_p$ values (Fricke and O'Neil, 1999).

3 The oxygen isotope composition of lacustrine materials

The stable isotope composition of carbonates has been used in paleoclimatic studies for more than 50 years (Urey et al., 1948; McCrea, 1950; Epstein et al., 1953). Carbonate materials in lake sediments are composed of authigenic and biogenic components (Kelts and Hsü, 1978). Authigenic carbonates precipitate when the concentration of $CaCO_3$ content reaches supersaturation which commonly occurs as a result of photosynthetic removal of CO_2 in the water column by algae or photosynthesis around plant organics causing calcite to precipitate such as commonly occurs on *Chara*. In most mid and high latitude regions authigenic carbonates are precipitated mainly in the summer months during periods of maximum phytoplankton productivity (Leng et al., 1999; Hammarlund et al., 2003; Leng and Marshall, 2004). Biogenic carbonates are derived from crustaceans and molluscs, such as ostracods and gastropods. These have different life spans and precipitate calcite during different parts of the year. A seasonal record can (for example) be obtained from ostracods as they moult several times during a year and analysis of individual shells can provide information on the seasonal range in $\delta^{18}O$ variation (von Grafenstein et al., 1999). Biogenic calcites often have species specific offsets in their $\delta^{18}O$ compared to calcite precipitated in equilibrium (McConnaughey, 1989a, b). For example, von Grafenstein et al. (1999) found that some ostracod species had vital (kinetic) $\delta^{18}O$ offsets between ca +0.7 to +2.2‰. Also, *Pisidium* sp. was found to precipitate aragonite with an offset of +0.8‰, apparently independent of temperature (Kim et al., 2007; von Grafenstein et al., 1999) relative to the value expected at equilibrium. Encrustations from *Chara* algae can precipitate close to equilibrium (Hammarlund et al., 2003), however during certain conditions, such as when calcite precipitate rapidly, non-equilibrium $\delta^{18}O$ ratios have been reported (Fronval et al., 1995; Andrews et al., 2004).

The use of carbonate material in Swedish lakes are restricted to regions with calcareous bedrock which mainly occur in Gotland (southeast Sweden), Scania (south Sweden), the area around Lake Storsjön (central Sweden) and scattered areas in the

Scandes Mountain chain. In lakes where carbonate are absent or poorly preserved, diatoms (Shemesh et al., 2001; Jones et al., 2004; Rosqvist et al., 2004; Schiff et al., 2009) and aquatic cellulose (Edwards et al., 1996, 2004; Wolfe et al., 2000, 2007; St. Amour, 2009) have shown an increasing potential reflecting past isotope hydrological variations.

Diatoms are photosynthetic algae that form frustules composed of opalline or biogenic silica ($SiO_2{\cdot}nH_2O$) and are found in almost all lakes wherever the macronutrients of Si, N and P are sufficient to sustain productivity (Leng and Barker, 2006). In high latitude lakes diatom blooms occur both immediately after the ice break up, when light conditions in the water columns improves and nutrients become available, and later during the summer season when the environmental conditions are less stressful for the algae (Lotter and Bigler, 2000; Catalan et al., 2002; Forsström et al., 2005, 2006). There is no evidence that changes in species composition would influence the oxygen isotopic composition of diatom silica ($\delta^{18}O_{diatom}$) in lake sediments (Shemesh et al., 2001; Jones et al., 2004; Schiff et al., 2009). Although diatom $\delta^{18}O$ might not be affected by any species-specific difference in fractionation (e.g. Shemesh et al., 1995; Schiff et al., 2009), species bloom in different habitats that might differ in temperature and $\delta^{18}O_{lakewater}$. Therefore, changes for example in the planktonic:benthic ratio through time may drive part of the $\delta^{18}O_{diatom}$ signal. Another factor that may affect the $\delta^{18}O_{diatom}$ signal is the silica maturation process within the diatom during sedimentation. There is some indication that maturation leads to enrichment of $\delta^{18}O_{diatom}$ after deposition (Schmidt et al., 1997, 2001; Brandriss et al., 1998; Moschen et al., 2006). This secondary isotope exchange is not fully understood but has been related to a difference of 2.5 in the $\delta^{18}O_{diatom}$ between living diatoms and surface sediment diatoms in Lake Holzmar, Germany (Moschen et al., 2006). However, there is currently no evidence for this effect on $\delta^{18}O_{diatom}$ from high altitude lakes (Schiff et al., 2009; Jonsson et al., 2009b).

Cellulose is a structural component found in cell walls of terrestrial and aquatic vascular plants, as well as in most algae (Wolfe et al., 2001). The aquatic cellulose productivity takes place during the ice-free season (spring to early autumn) in high latitude regions. The isotopic composition in cellulose ($\delta^{18}O_{cellulose}$) will, as long as the sediment cellulose is aquatic in origin, incorporate a $\delta^{18}O_{lakewater}$ signal reflecting the isotope hydrology during this period. The oxygen isotope fractionation between cellulose and lake water (ca. 1.025–1.030) is believed to be independent of water temperature, plant species, the photosynthetic mode and the oxygen isotopic composition of CO_2 (Epstein, 1977; Yakir and DeNiro, 1990; Wolfe et al., 2001; Sauer et al., 2001; St. Amour, 2009).

To interpret $\delta^{18}O$ records a detailed knowledge of the processes that control and modify the signal is required (Leng et al., 2005). The main factors (processes) that influence the sediment $\delta^{18}O$ (both diatom and carbonate) proxy signal are lake water $\delta^{18}O$ and lake water temperature. However, also non-climatic effects such as long-term lake development during the Holocene can have an influence on the $\delta^{18}O$ signal. For example, vegetation succession, soil development and hydrological changes will affect the input of water, the nutrient concentration and pH in the lake (Engstrom et al., 2000), which in turn might affect the lake productivity (timing and community) and precipitation of minerals. Also, the lake volume might gradually decrease as the basin is filled with sediments and the residence time will, especially in shallow lakes, change towards shorter times if assuming the same runoff. A 50% loss in volume would reduce the residence time by half. This might change the lake water $\delta^{18}O$ and lead to the sedimentary $\delta^{18}O$ record changing from interannual to an annual signal.

All the stable isotope ratios discussed here are expressed as "δ" values, representing deviations in per mil (‰) from a standard, such that $\delta_{sample}=1000[(R_{sample}/R_{VSMOW})-1]$, where R is the $^{18}O/^{16}O$ or $^{2}H/^{1}H$ ratio in sample and standard. For lake water oxygen and deuterium, diatom and cellulose oxygen the data are presented on the VSMOW scale, while carbonate data are presented relative to VPDB (Leng et al., 2005).

3.1 The oxygen isotope composition of lake waters

The δ^{18}O and δ^{2}H composition of lake waters in Sweden (Figs. 2 and 3) show that the isotopic signatures are forced by local hydrology and different climate parameters. Due to lower temperatures and rain-out effect relatively lower $\delta^{18}O_{lakewater}$ and $\delta^{2}H_{lakewater}$ are found in lakes located at high altitude and/or latitude. The correlation between modern δ^{18}O and δ^{2}H in precipitation on a global scale is known as the Global Meteoric Water Line (GMWL) (Craig, 1961; Rozanski, 1993). Comparison of $\delta^{18}O_{lakewater}$ and $\delta^{2}H_{lakewater}$ for a particular lake to the GMWL can provide information about a lake's hydrological setting. Lakes that plot close to GMWL, like Lake 850 (Shemesh et al., 2001), Vuolep Allakasjaure (Rosqvist et al., 2004) and Lake Tibetanus (Hammarlund et al., 2002; Rosqvist et al., 2007), indicate that the lake water is isotopically the same as the precipitation for that region (Fig. 3). Shifts along the GMWL indicate seasonal variations in precipitation inputs (Clark and Fritz, 1997). Seasonal $\delta^{18}O_{lakewater}$ variation in Sweden is strongly determined by snow accumulation during winter, snowmelt in spring and early summer and enrichment during the ice free season due to evaporation and influence of isotopically enriched precipitation. The inputs from snowmelt have a significant role in determining lake water δ^{18}O composition, especially in northern Sweden where almost half of the annual precipitation falls as snow. Due to catchment elevation and different residence times between sites these lakes contain different proportions of summer and winter precipitation and show a range of $\delta^{18}O_{lakewater}$ values between −14.9‰ and −7.7‰. Because of short residence times, long ice cover and low air temperatures the effect of evaporation is insignificant in these high altitude/latitude lakes. The seasonal variations in the isotopic composition of the non-evaporative lakes are larger in smaller lakes with short residence time, as they respond faster to seasonal changes in precipitation, compared to larger lakes with longer residence times, which retain an isotopic signal closer to that of annual mean precipitation (Jonsson et al., 2009a). Groundwater fed lakes like Lake Tibetanus are (despite having short residence times) generally more seasonally stable, as groundwater in general represents the local mean weighted annual $\delta^{18}O_p$ (Clark and Fritz, 1997), compared to lakes fed predominantly by surface waters (Gat, 1996). Although it is possible that seasonal groundwater variation occurs, this is only likely where the unsaturated zones are thin (Darling, 2004) and in catchments where the water residence times are relatively short (Vitvar and Balderer, 1997).

Lake waters sampled from more continental sites or sites in the rain shadow of the Scandes, for example the continental Jämtland (Andersson et al., 2009), on Gotland (Rosqvist et al., 2009) and in south central Sweden (Lake Igelsjön; Hammarlund et al., 2003) have $\delta^{18}O$ that plot below the GMWL on Local Evaporation Lines (LEL), with slopes between 7 and 4. This is as a result of lake waters that have undergone evaporation, where the isotopic compositions of the residual lake water have become progressively more enriched in ^{18}O (Clark and Fritz, 1997). The slope of the LEL depends primarily on relative humidity together with wind speed and surface temperature. Lower relative humidity values will generate LEL with lower slopes (Clark and Fritz, 1997). Displacement of the individual lake water along the LEL provides information on water balance of the lake (E/I) (Gibson et al., 2005, 2008), the further the lake water plots along the LEL the more evaporated water. Evaporation can affect $\delta^{18}O_{\text{lakewater}}$ by several per mille (Fig. 3). The intersection of the LEL with the GMWL corresponds to the average isotopic composition of the recharge water entering the lake. In lakes affected by evaporation dry periods results in high $\delta^{18}O_{\text{lakewater}}$ (increased E/I ratio) and more humid periods in low $\delta^{18}O_{\text{lakewater}}$ (decreased E/I ratio).

3.2 Oxygen isotopes to derive lake water temperatures

The oxygen isotopic composition of the diatom silica and carbonate is controlled by the oxygen isotope composition of the lake water within which they formed at a given temperature. The temperature dependent fractionation between calcite and lake water has a well established negative temperature coefficient of ca -0.25‰/°C (Craig, 1965); while the fractionation between diatom silica and lake water is more controversial with

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published estimates ranging from −0.5‰/°C to −0.2‰/°C (Juillet-Leclerc and Labeyrie, 1987; Shemesh et al., 1992; Brandriss et al., 1998; Moschen et al., 2005). These negative responses to changes in lake water temperature will have an opposing effect on the sediment $\delta^{18}O$ record and damp the effect of changes in $\delta^{18}O_p$ caused by air temperature changes according to the Dansgaard relationship (increases in air temperature lead to increasing $\delta^{18}O_p$). Recent analyses of $\delta^{18}O_{diatom}$ from surface sediments across Europe (Tyler et al., 2008) and Alaska (Schiff et al., 2007, 2009) indicate that the effect of temperature may be less than previously estimated, and suggests an almost constant fractionation value in the range of 35.5–42.8‰ between $\delta^{18}O_{lakewater}$ and $\delta^{18}O_{diatom}$. In addition, studies of $\delta^{18}O_{diatom}$ and $\delta^{18}O_{carbonate}$ from northern Sweden (Shemesh et al., 2001; Hammarlund et al., 2002; Rosqvist et al., 2004, 2007) and the Kola Peninsula (Jones et al., 2004) show that the lake water temperature effect is not the main factor responsible for the variations in the $\delta^{18}O$ records.

The proxy signals recorded in authigenic carbonates, biogenic carbonates, diatom silica and aquatic cellulose may represent different $\delta^{18}O_{lakewater}$ signals depending on the lake specific hydrology and the timing of the productivity/blooming/precipitation. In lakes with strong seasonal isotope signals, the proxy records might provide $\delta^{18}O$ signatures weighted by different seasonal biological productivities and mineral precipitation (Leng and Barker, 2006). For example, if the diatom silica production mainly occurs early in the summer season, the isotopic signal preserved would reflect an early season $\delta^{18}O_{lakewater}$, possibly influenced by isotopically depleted snowmelt. In contrast, carbonates formed mainly in the summer months, during periods of maximum algal productivity for example, would be dominated by the composition of summer precipitation and any evaporative enrichment. Therefore, a comparison of different $\delta^{18}O$ records may provide additional information about seasonality (i.e. spring to summer) (Leng and Barker, 2006). In addition, as $\delta^{18}O_{cellulose}$ are thought to incorporate a lake water $\delta^{18}O$ signal independent of water temperature, a combination of $\delta^{18}O_{cellulose}$ and $\delta^{18}O_{diatom}$ and/or $\delta^{18}O_{carbonate}$ might yield additional information about lake water temperatures (Klisch et al., 2007) assuming similar timings and lake habitats for the production of the cellulose by biological productivity and the precipitation of the diatom/carbonate.

4 Climate implications of oxygen isotope data from lakes in Sweden

Results from studies of oxygen isotopes in lacustrine sediments from Sweden (Table 2) can in general be divided into two types of reconstructions:

1. changes in $\delta^{18}O_p$ in lakes with no evaporation, reflecting past changes in amount and seasonal distribution of precipitation, influences of different air masses and temperature; and

2. changes in water balance (E/I) in lakes where lake water $\delta^{18}O$ plots on a LEL, reflecting moisture availability and temperature.

Past changes in $\delta^{18}O_p$ in northern Scandinavia have been reconstructed using $\delta^{18}O$ from carbonates (authigenic calcite, ostracodes and gastropods) (Hammarlund et al., 2002; Rosqvist et al., 2007), diatom silica (Shemesh et al., 2001; Rosqvist et al., 2004; Jones et al., 2004; Jonsson et al., 2009b) and aquatic cellulose (St. Amour, 2009). The lakes used to reconstruct changes in $\delta^{18}O_p$ have similarities in that all are through-flow high altitude/latitude or groundwater fed lakes with insignificant evaporation effects. Changes in the local $\delta^{18}O_p$ depend on variations of ambient air temperatures and modifications in atmospheric circulation which lead to a shift of the moisture source, a change in the vapor transport efficiency or to changes in winter to summer precipitation distribution. In lakes with short residence time the isotope record can reflect seasonal aspects of $\delta^{18}O_p$ e.g. amount of winter precipitation (Jonsson et al., 2009b).

Reconstructions from lakes sensitive to evaporation provide information about changes in water balance (E/I) (Hammarlund et al., 2003; Andersson et al., 2009). Decreasing $\delta^{18}O$ in lake waters suggest a more positive water balance caused by increased humidity, whereas increasing $\delta^{18}O$ indicate a negative water balance caused by increased evaporation. Changes in water balance can be inferred in the absence of any major changes in $\delta^{18}O_p$. The residence time of lake water can determine if seasonal or annual changes in water budget can be inferred. For example, Lake Igelsjön in south central Sweden has an estimated residence time of a few weeks, providing information about water balance changes during summers (Hammarlund et al., 2003). In contrast, the residence time of Lake Blektjärnen in central Sweden is >1 year, which results in a partially inherited annual water balance signal (Andersson et al., 2009).

4.1 Holocene long-term trend

Numerous paleoclimate studies, a few including oxygen isotopes, have been undertaken in northern Fennoscandia, especially in the Abisko area (Fig. 1). A comparison of carbonate and diatom silica $\delta^{18}O$ records from two small non-evaporative lakes in Abisko show similar decreasing trends over the Holocene (Fig. 4). Both the $\delta^{18}O_{carbonate}$ record from the groundwater fed Lake Tibetanus (Hammarlund et al., 2002) and the $\delta^{18}O_{diatom}$ record from the open through flow Lake 850 (Shemesh et al., 2001) are assumed to reflect annual mean $\delta^{18}O_p$. These records show high $\delta^{18}O$ in the early Holocene (ca. 10 000–8000 cal yr BP) and lower values in the mid- (ca. 8000–4000 cal yr BP) and late Holocene (ca. 4000–100 cal yr BP). A similar depletion trend is also shown in a speleothem $\delta^{18}O$ record (SG93) from Mo I Rana in northern Norway (Fig. 4) (Lauritzen and Lundberg, 1999). The long-term Holocene trend in the records follows the pattern of orbital forcing of summer insolation at 65° N, with considerably higher insolation values during early Holocene than today (Berger and Loutre, 1991), and therefore Holocene decreasing summer temperatures would be expected and consequently lower $\delta^{18}O_p$. However, quantitative reconstructions of Fennoscandia summer and annual mean temperatures based on biological proxies from lake sediments (Barnekow, 1999; Seppä and Birks, 2001; Hammarlund et al., 2002; Seppä and Birks, 2002; Bjune et al., 2005; Seppä et al., 2005; Bigler et al., 2006) suggest a Holocene climate with a cool early Holocene, a mid-Holocene temperature maximum and decreasing temperature in late Holocene. Seppä and Birks (2002) suggest that the effect of high summer solar radiation on summer temperatures in Fennoscandia in the early Holocene was subdued by a cooling effect of a stronger than present zonal circulation (Seppä and Birks, 2001), with enhanced Atlantic air flow across the Scandes mountains, and as a result a cool and moist climate.

By combining pollen inferred mean July temperatures with the $\delta^{18}O_p$ record from Lake Tibetanus, Hammarlund et al. (2002) show that the $\delta^{18}O_p$ was about 2‰ higher in the early Holocene than would be predicted by the modern $\delta^{18}O_p$ – temperature relation. The authors suggest that the high zonal index, which was caused by an enhanced sea-level air-pressure gradient between the North Atlantic and the Eurasian continent, could have been responsible for this deviation from the Dansgaard – relationship as a result of changes in the efficiency of moisture transport over the Scandes. A similar explanation was suggested by Edwards et al. (1996) for parts of North America where high $\delta^{18}O$ values also characterized the early Holocene. The successively decreasing $\delta^{18}O$ values would then be a response to changes from a maritime climate with a strong zonal index towards a more continental climate with a meridional circulation and weaker westerlies with increasing isotopic distillation of moisture due to deepening of the precipitation and isotope shadows on the east side of the Scandes. The modern $\delta^{18}O_p$ – temperature relation appears to have been established in Fennoscandia ca. 6000–4000 cal yr BP, probably as a result of weaker zonal atmospheric circulation in response to lower summer insolation (Hammarlund et al., 2002; St. Amour, 2009).

Shemesh et al. (2001) argue that because a change in the relative contribution of different air masses to the local $\delta^{18}O_p$ would influence $\delta^{18}O_{lakewater}$, it is possible to detect such changes in lacustrine $\delta^{18}O$ records. The maritime air mass from the west-southwest carry precipitation with isotopic values between −8 and −10‰, whereas the precipitation in the more negative Arctic polar continental air mass from the north, northeast has values between −14 to −17‰. Shemesh et al. (2001) suggest that the recorded 3.5‰ Holocene depletion trend reflects an enhanced influence of the isotopically more negative Arctic polar continental air mass. The mean sea-level pressure and surface-wind reconstruction in the early Holocene indeed shows enhanced transport of Atlantic moisture supply to northern Fennoscandia (zonal air flow) (Hebbeln et al., 1994; Hald and Aspeli, 1997). It is likely that when the high pressure system weakened after deglaciation, the Arctic polar continental air mass could penetrate more southward (meridional air flow), and increase the contribution of the Arctic moisture to northern Fennoscandia (Shemesh et al., 2001). Strong zonal circulation bringing relatively humid conditions during the early Holocene was also reported in southern Sweden from relatively low $\delta^{18}O$ values in Lake Igelsjön (Hammarlund et al., 2003) (Fig. 4). We conclude that both the records from northern Sweden (Lake 850 and Lake Tibetanus) and the southern Sweden record from Lake Igelsjön are consistent with a successive diminishing influence of maritime zonal air flow (strong westerlies) over Fennoscandia in favor of an increasing proportion of more continental meridional air flow (weak westerlies). This is supported by quantitative reconstructions from pollen indicating that the climate in Fennoscandia has become increasingly more continental over the last 7000 years mainly as an effect of winter cooling (Gisecke et al., 2008).

The record from Lake Igelsjön (Fig. 4) (Hammarlund et al., 2003; Jessen et al., 2005), together with lake level studies (Digerfeltd, 1988; Banrnekow, 2000; Korhola et al., 2005) and temperature reconstructions (Seppä and Birks; 2001, 2002; Seppä et al., 2005) show that conditions were relatively dry and summer temperatures high during the so called Holocene Thermal Maximum (HTM) between ca. 8000 and 4000 cal yr BP. This is also a period when many of the glaciers in Scandinavia melted away due to high summer temperatures and/or reduced winter precipitation (Nesje et al., 2005, 2008). These results are consistent with the explanation that zonal air flow decreased and was replaced by a strong summer high pressure system with a blocking anticyclonic situation as suggested by Antonsson et al. (2008). The late Holocene (4000 to 100 cal yr BP) is characterized by increasingly cold, moist and unstable climate (Seppä and Birks, 2001; Hammarlund et al., 2003; Seppä et al., 2005) suggesting reduced influence of the blocking anticyclones.

Changes in the ratio between summer to winter precipitation, cooling due to landuplift and changes in the moisture source area might have amplified the $\delta^{18}O$ signal, especially in the early Holocene. The land-uplift in Sweden has been estimated to be ca. 100 m since 9000 cal yr BP (Renberg and Segerström, 1981; Møller, 1987; Dahl and Nesje, 1996) and may account for a ca. −0.26‰ to −0.5‰ change in the $\delta^{18}O_p$ depending on the used lapse rate (Siegenthaler and Oeschger, 1980; Ingrahamn, 1998; Poage and Chamberlain, 2001). The impact on ocean $\delta^{18}O$ caused by melting of the global ice is approximately 0.8‰ since the last glacial to 10 000–8000 cal yr BP when modern values were archived (Fairbanks, 1989; Schrag et al., 1996, 2002).

4.2 Short-term changes during the last 5000 years

Superimposed on the circulation and summer insolation forced Holocene $\delta^{18}O_p$ trend are short-term changes. From the similarity of a $\delta^{18}O$ record derived from diatoms and a sedimentary proxy record for glacier fluctuations from a pro-glacial lake, Vuolep Allakasjaure, located in northwestern Sweden, Rosqvist et al. (2004) concluded that atmospheric circulation changes also occurred on a centennial-millennium time scale during the last 5000 years. The authors argue that precipitation with low $\delta^{18}O$ values dominated during a few hundred years around 4400, 3000, 2000 and 1200 cal yr BP (Fig. 5), simultaneously as the catchment glacier advanced. The mass balance of the present day glaciers in this area primarily responds to summer air temperatures but winter precipitation amounts is also important (Holmlund et al., 1996). The fact that the isotope depletion minima coincide with glacier advances indicates that changes in atmospheric circulation affected both the $\delta^{18}O_{\text{lakewater}}$ and the mass balance of the glacier. Probably weaker westerlies and a more southerly position of the polar front allowed colder and isotopically more depleted air to influence the region during the summers. However, because an increase in the amount of winter precipitation would in the same way result in decreasing $\delta^{18}O_{\text{lakewater}}$ values and glacier mass balance increase, alternatively the record might reveal changes in winter circulation (Jonsson et al., 2009b). Several isotopic minima have also been detected in a high resolution $\delta^{18}O_p$ record from Lake Tibetanus. These occurs around 2900, 1600, 1300, 500 and

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200 cal yr BP (Rosqvist et al., 2007). Interestingly the δ^{18}O minima in both these lakes occur at times of relative ice rafted debris (IRD) maxima in the North Atlantic (Bond et al., 1997, 2001) (Fig. 5). From the IRD maxima it has been inferred that temperatures decreased and sea ice extent increased in the area of the North Atlantic Current (Bond et al., 2001; Moros et al., 2006). Bütikofer (2007) suggested that the IRD maxima might resemble prolonged intervals of the negative phase of the contemporary NAO with below normal precipitations and cooler conditions in northern Europe. The fact that δ^{18}O changed over Fennoscandia at the same periods as the IRD changes reveal a significant terrestrial response to changes in the North Atlantic. Changes in atmospheric circulation during different NAO phases might then result in shift of moisture source location for precipitation in Fennoscandia as well as changes in the atmospheric fractionation temperature. This effect of NAO variability on $\delta^{18}O_p$ has been identified for precipitation in Greenland in a study using a new moisture source diagnostic (Sodemann et al., 2008).

Periods with relatively large amounts of IRD also correspond to periods when the winter precipitation gradient between southern and northern Norway was high due to a dominant southerly position of the westerlies (Bakke et al., 2008). Around 2000 cal yr BP there was a marked shift to wetter conditions in both southern and northern Norway indicating stronger effects of the westerlies (Bakke et al., 2008) in both regions. Over the last 2000 years (until ca. 100 cal yr BP) the $\delta^{18}O_{lakewater}$ records from south and central Sweden, Lake Igelsjön (Hammarlund et al., 2003) and Lake Blektjärnen (Andersson et al., 2009), also show increasing humidity (Figs. 4 and 5), probably as a response to stronger westerlies. Wet conditions is related to low summer evaporation and mainly interpreted as a summer signal. However changes in winter precipitation may have affected the E/I records through changed annual precipitation. Decreasing δ^{18}O values in the Lake 850 record might also indicate an increasing amount of winter precipitation in the Abisko area over the last 2000 years.

Preliminary results from ongoing work focussing on high resolution reconstructions covering the last millennia preliminary show that significant change in $\delta^{18}O_{lakewater}$ of both through flow and evaporative lakes have occurred also on the centennial-decadal time scale (Cunningham et al., 2009).

In order to retrieve a better understanding of the processes that affect the $\delta^{18}O_{lakewater}$ and how to interpret the $\delta^{18}O$ sediment signal, high resolution $\delta^{18}O_{diatom}$ records from two fast through flow high altitude lakes were compared with instrumental data from the last century obtained from nearby metrological stations (Jonsson et al., 2009b). The results from the $\delta^{18}O_{diatom}$ analyses from Vuolep Allakasjaure and Lake Spåime (Fig. 1) sediments show remarkably similar trends over the last 150 years with a decreasing trend (~3‰) from 1850 to the lowest values of the records around 1990, with a peak around 1980. After 1990 both records increase by ca 2.5‰. The results show that the isotopic hydrology in these two lakes, situated 650 km apart in the Swedish Scandes, responded to the same main forcing during the last 150 years and that $\delta^{18}O$ signatures reflect changes in the seasonality of precipitation over this period. The decreasing trend from 1900 to 1990 reflects an increase in winter precipitation associated with both zonal (positive winter NAO) and meridional (negative winter NAO) air flow patterns in these regions.

5 Summary and conclusions

Here we demonstrate how the oxygen isotope composition of lacustrine sedimentary materials from Sweden yield both long-term and short-term paleoclimate information. The interpretation of the $\delta^{18}O$ records requires a detailed knowledge of the processes that control the signal, and this must be determined for each individual lake. Thus, it is important to investigate the isotope hydrology in the modern lake environment to establish the relationship between both the measured signal and $\delta^{18}O_{lakewater}$. The amount of isotopic variation in lake water $\delta^{18}O$ is a combination of the original $\delta^{18}O_{lakewater}$, the amount and time of the snowmelt, the amount of seasonally specific precipitation and groundwater, any evaporation effects, and possibly most crucially, by lake water residence time.

Oxygen isotopes can be obtained from a large range of materials such as authigenic calcites, ostracod shells, diatom silica and aquatic cellulose. The signatures captured by these sedimentary materials may represent different $\delta^{18}O_{lakewater}$ signals depending on the lake specific hydrology and the timing of the /production/blooming/ precipitation. A combination of $\delta^{18}O$ records derived from different material can reflect seasonal aspects of $\delta^{18}O_p$ and might yield additional information about lake water temperatures.

Results so far from studies of oxygen isotopes in lacustrine sediments from Sweden can be divided into two types of reconstruction: (i) changes in annual or seasonal $\delta^{18}O_p$ in lakes with no evaporation, and (ii) changes in E/I water balance in lakes where evaporation is forcing the $\delta^{18}O_{lakewater}$ and the record therefore reflects catchment water balance. The proxy $\delta^{18}O_p$ records show that the air masses bringing precipitation to this region shifted between zonal (high $\delta^{18}O_p$) and meridional air (low $\delta^{18}O_p$) flow on different time scales over the past 10 000 years. We show that in non-evaporative lakes where $\delta^{18}O_{lakewater}$ is close to annual $\delta^{18}O_p$, diatom and carbonate sediment records show similar changes in the Holocene $\delta^{18}O_p$. The Holocene decreasing $\delta^{18}O_p$ trend in northern Sweden is likely forced by a change from a dominance of strong zonal westerly airflow in early Holocene to a more meridional flow pattern. By coupling independent pollen and $\delta^{18}O$ records the deviation from the Dansgaard $\delta^{18}O_p$ – temperature relation was found in northern Sweden during the early Holocene supporting the use of a multiproxy approach. The reconstructions based on sediments from the evaporative sites in south and central Sweden reflect warm and dry conditions during mid Holocene and a change towards increasingly wetter conditions which we associate with increased zonal air flow, which is to be expected from a more southerly position of the westerlies.

Periods of $\delta^{18}O_p$ minima in Scandinavia and IRD maxima in the North Atlantic are possibly connected to negative phase of the NAO with below normal temperatures and precipitation amounts in Sweden and relatively cooler conditions over the northeastern North Atlantic. The recorded $\delta^{18}O_p$ variations might then be a result from a combination of shift of moisture source location for the precipitation and changes in the atmospheric fractionation temperature. Study of high alpine lakes in the Swedish Scandes show that high amount of winter precipitation can be derived both during positive and negative NAO phases.

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Table 1. GNIP data stations and nearby metrological stations for metrological data. The $\delta^{18}O$ precipitation measurements were performed between 1975 and 1980.

GNIP station	Location	Elevation, (m a.s.l.)	Metrological station (1961–1991)
Bredkälen	63°53′ N, 15°18′ E	400	Hallhåxåsen
Forshult	60°10′ N, 13°46′ E	192	Forshult
Smedby	56°42′ N, 16°11′ E	20	Kalmar
Abisko	68°12′ N, 18°29′ E	392	Abisko
Rickleå	64°03′ N, 20°33′ E	10	Lövånger
Göteborg	57°42′ N, 11°58′ E	15	Göteborg

Table 2. Lakes and their characteristics.

	Vuolep Allakasjaure	Lake Tibetanus	Lake 850	Igelsjön	Blektjärnen
Study	Rosqvist et al., 2004	Hammarlund et al., 2002 Rosqvist et al., 2007	Shemesh et al., 2001	Hammarlund et al., 2003	Andersson et al., 2009
$\delta^{18}O$ proxy	Diatom	Carbonate	Diatom	Carbonate	Carbonate
Location	Northern Sweden	Northern Sweden	Northern Sweden	South Central Sweden	Central Sweden
Elevation m a.s.l.	995	560	850	111	330
Lake area (km^2)	0.3	0.085		0.0025	0.05
Max depth (m)	11	4	8.2	2.5	10
Catchment area (km^2)	22	0.15	0.5	0.32	0.45
pH		7.7	6.8	7.1	
Residence time	5 weeks	20–25 days	32 weeks	20 days	2 years
Time period (cal yr BP)	0–5000	0–10 500 (Hammarlund et al., 2002); 0–3000 (Rosqvist et al., 2007)	0–9500	0–10,700	0–4400
Annual mean precipitation (mm)	1000	300 (Abisko)	400	745	564
Annual mean air temp (°C)	−0.8 (Abisko)	−1.4	−0.8 (Abisko)	+5.9	+3
Ice cover (months)	Mid-October to early June	Mid-October to early June	Mid-October to late May		Early November to late May
$\delta^{18}O$ lake (‰)	−14.4 to −14.4	−13.5 to −15.0	−13.9 to −12.8	−8.2 to −11.7	−9.8 to −111
Evaporation effect	Insignificant	Insignificant	Insignificant	Yes	Yes

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Fig. 1. Map of Sweden, with numbers indicating location of lakes referred to in the text. 1) Lake Vuolep Allakasjaure (Rosqvist et al., 2004; Jonsson et al., 2009b), 2) Lake 850 (Shemesh et al., 2001), 3) Lake Tibetanus (Hammarlund et al., 2002; Rosqvist et al., 2007), 4) Lake Spåime (Jonsson et al., 2009b), 5) Lake Blektjärnen (Andersson et al., 2009), 6) Lake Igelsjön (Hammarlund et al., 2003) and 7) Lakes in Gotland (Rosqvist et al., 2009).

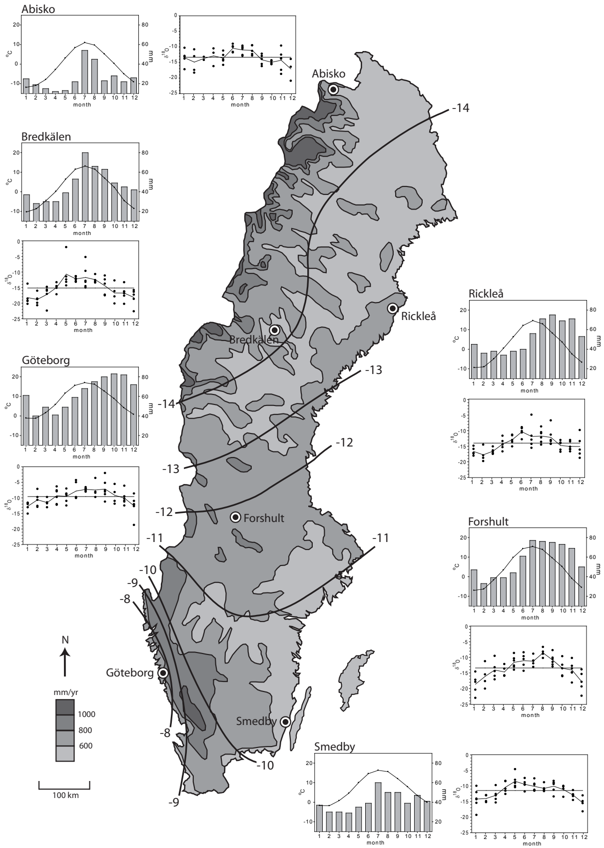

Fig. 2. Annual precipitation pattern over Sweden (Alexandersson and Andersson, 2004) and isolines of δ^{18}O in precipitation after Burgman et al. (1987). Selected GNIP stations with monthly mean average values of δ^{18}O in precipitation (1975–1980; IAEA/WHO) and monthly mean values of temperature and amount of precipitation from nearby metrological stations (1961–1990; Alexandersson and Eggertsson Karlström, 2001).

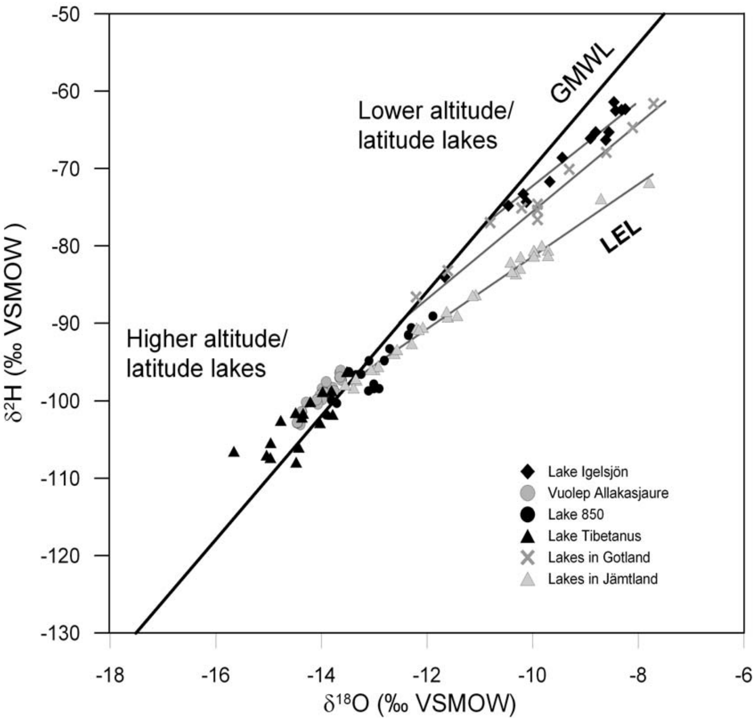

Fig. 3. $\delta^{18}O_{lakewater}$ versus $\delta^{2}H_{lakewater}$ for Vuolep Allakasjaure (Rosqvist et al., 2004; Jonsson et al., 2009a), Lake 850 (Shemesh et al., 2001), Lake Tibetanus (Hammarlund et al., 2002; Rosqvist et al., 2007), Lakes in Jämtland (Andersson et al., 2009), Lakes in Gotland (Rosqvist et al., unpublished), and Lake Igelsjön (Hammarlund et al., 2003); shown in relation to the Global Meteoric Water Line (GMWL: $\delta^{2}H=8\delta^{18}O+10‰$; Rozanski et al., 1993). Local Evaporation Lines (LEL) are determined from a linear regression of corresponding lake water data. Shifts along the GMWL indicate isotopic variations in precipitation input to the lake.

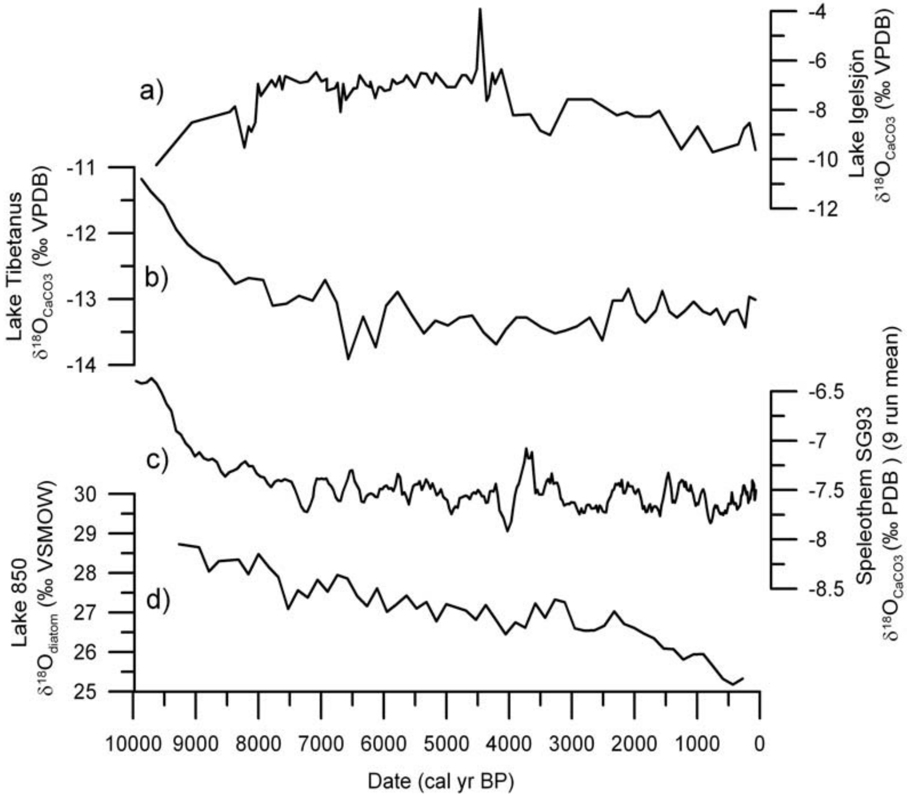

Fig. 4. $\delta^{18}O$ records from lake sediments in Sweden: **(a)** Lake Igelsjön (Hammarlund et al., 2003; Jessen et al., 2005), **(b)** Lake Tibetanus (Hammarlund et al., 2002), **(d)** Lake 850 (Shemesh et al., 2001) and, **(c)** a Speleothem $\delta^{18}O$ record from Norway (Lauritzen and Lundberg, 1999) plotted from 10 000 to 0 cal yr BP. Note the different $\delta^{18}O$ scales.

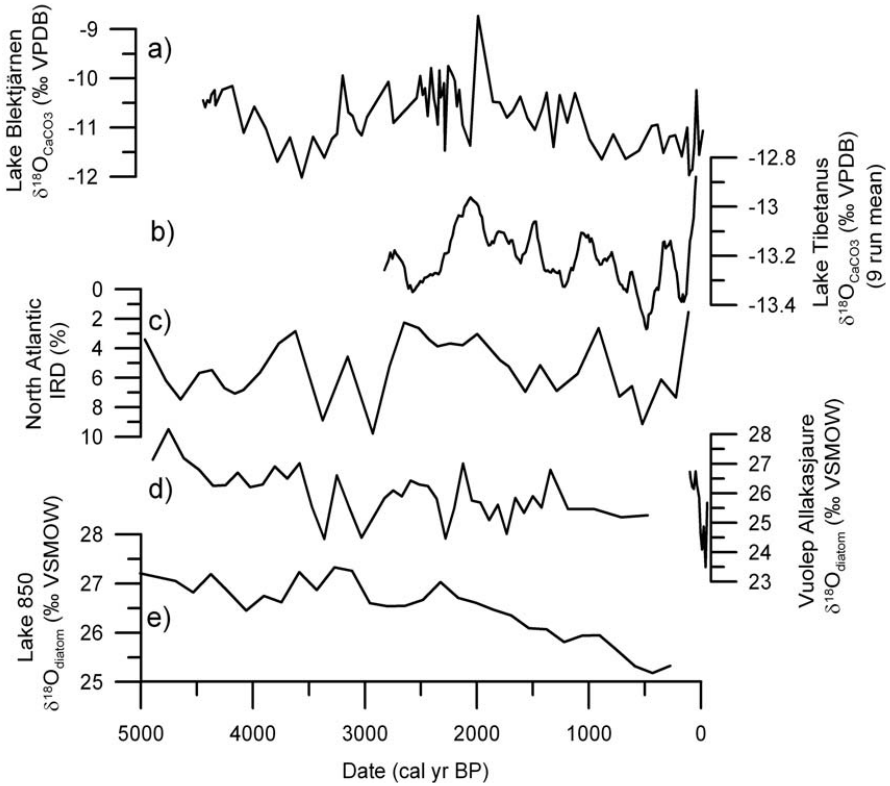

Fig. 5. δ^{18}O records from lake sediments in Sweden: **(a)** Lake Blektjärnen (Andersson et al., 2009), **(b)** Lake Tibetanus (Rosqvist et al., 2007), **(d)** Vuolep Allakasjaure (Rosqvist et al., 2004; Jonsson et al., 2009b), **(e)** Lake 850 (Shemesh et al., 2001), and **(c)** ice rafted debris (IRD) in the North Atlantic (Bond et al., 1997, 2001) plotted from 5000 to 0 cal yr BP. Note the different δ^{18}O scales and the inversed axis for IRD data.

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