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Stable oxygen and hydrogen isotopes in sub-Arctic lake waters from northern Sweden

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SUMMARY

Lakes in sub-Arctic regions have the potential of retaining many different aspects of water isotope composition in their sediments which can be used for palaeoclimate reconstruction. It is therefore important to understand the modern isotope hydrology of these lakes. Here we discuss the significance of variations in water isotope composition of a series of lakes located in north-west Swedish Lapland. Climate in this region is forced by changes in the North Atlantic which renders it an interesting area for climate reconstructions. We compare $\delta^{18}O_{lake}$ and $\delta^{2}H_{lake}$ collected between 2001 and 2006 and show that lakes in this sub-Arctic region are currently mainly recharged by shallow groundwater and precipitation which undergoes little subsequent evaporation, and that the $\delta^{18}O$ and $\delta^{2}H$ composition of input to the majority of the lakes varies on a seasonal basis between winter precipitation (spring thaw) and summer precipitation. Seasonal variations in the isotopic composition of the lake waters are larger in lakes with short residence times (<6 months), which react faster to seasonal changes in the precipitation, compared to lakes with longer residence times (>6 months), which retain an isotopic signal closer to that of annual mean precipitation. Lake waters also show a range of isotope values between sites due to catchment elevation and timing of snowmelt. The lake water data collected in this study was supported by isotope data from lake waters, streams and ground waters from1995 to 2000 reported in other studies.

Introduction

Reconstruction of past changes in lake water isotopic composition ($\delta^{18}O_{lake}$ and $\delta^{2}H_{lake}$), that in turn reflect changes in the isotopic composition of precipitation ($\delta^{18}O_{p}$ and $\delta^{2}H_{p}$), increase our understanding of the forcing mechanisms behind climate change (Darling, 2004). This is because shifts in $\delta^{18}O_{p}$ and $\delta^{2}H_{p}$ reflect changes in regional precipitation patterns that follow upon changes in atmospheric and oceanic circulation dynamics.

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) and diatom silica (Shemesh et al., 2001; Rosqvist et al., 2004; Jones et al., 2004) which accumulate in lake sediments. While we know on the broad scale that northern Scandinavia is strongly influenced by atmospheric and oceanic circulation changes over the North Atlantic (Luterbacher et al., 2002; Uvo, 2003; Rosqvist et al., 2007), a thorough understanding of local climate and hydrological factors is necessary before inferences about past changes in $\delta^{18}O_{lake}$ ($\delta^{2}H_{lake}$) and $\delta^{18}O_{p}$ ($\delta^{2}H_{p}$) based on proxy data can be made. This is because different types of lakes have been shown to perform isotopically differently to the same forcing (Darling, 2004; Roberts et al., 2008). For any given lake, $\delta^{18}O_{lake}$ and $\delta^{2}H_{lake}$ will depend on the hydrological balance between inputs (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 (c.f. Leng and Anderson, 2003; Leng and Marshall, 2004).

Knowledge of the isotopic composition of precipitation is important in stable isotope studies. The distribution of $\delta^{18}O$ and $\delta^{2}H$ in precipitation is in general more negative with increasing distance from the equator, from coastal regions (continental effect) and increasing elevation (altitude effect) (Rozanski et al., 1993). The correlation between $\delta^{18}O$ and $\delta^{2}H$ in precipitation on a global scale is well understood (Craig, 1961) and is known as the Global Meteoric Water Line (GMWL), and has been defined as: $\delta^{2}H = 8\,\delta^{18}O + 10‰$ (Rozanski et al., 1993). However, the stable isotope composition of rainfall in a region may represent different meteoric conditions and will better be described by a Local Meteoric

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Water Line (LMWL) which may have a slightly different slope and intercept compared to the GMWL (Ingraham, 1998). At mid- and high-latitudes there is a seasonal variation along the LMWL driven by temperature, with the isotopic composition of winter precipitation being more depleted compared to the precipitation received during the summer. Comparison of $\delta^{18}O_{lake}$ and $\delta^{2}H_{lake}$ to the LMWL can tell us about a lake's hydrological setting. Lake waters plotting on the LMWL are isotopically the same as the precipitation for that region. Whereas lake waters that plot below the LMWL along a different gradient, on a Local Evaporation Line (LEL), have undergone evaporation where the isotopic composition of the residual lake water becomes progressively more enriched (Clark and Fritz, 1997). Based on a seasonal conceptual model Gibson et al. (2008) suggest that for high-latitude regions with highly seasonal climates the slope of the LEL will approach the slope of the LMWL. Displacement of the individual lake water along the LEL provides information on water balance (evaporation/inflow ratio) of the lake (Gibson et al., 2005, 2008), the further the lake water plots along the LEL the more evaporated is the water. The intersection of the LEL with the LMWL corresponds to the average isotopic composition of the water entering the lake. Water in non evaporated lakes ($\delta^{18}O_{lake}$ and $\delta^{2}H_{lake}$ plot along the LMWL) which have residence times more than a year tend to naturally average and integrate seasonal signals (Edwards et al., 2004), and their water usually has an isotopic signal close to that of weighted average annual precipitation. However in shorter residence time lakes, the isotopic composition of the water may lie on the LMWL but will be dependent on other physical properties such as lake to catchment ratio and catchment elevation. For example, seasonal variation in the isotopic composition of precipitation input is likely to be more significant in short residence time lakes, as these lakes tend to have isotopic values which are constantly displaced by later precipitation (Leng et al., 2005). In fact, some lakes behave like upland streams with $\delta^{18}O_{lake}$ similar to individual rainfalls reflecting very short residence times (Tyler et al., 2007).

Here we discuss the significance of variations in water isotope composition from a series of sub-Arctic lakes located in north-west Swedish Lapland. Lake water samples were collected between 2001 and 2006. Previous studies of isotope hydrology in northern Sweden have presented sampling of (usually) single lakes (Shemesh et al., 2001; Hammarlund et al., 2002; Rosqvist et al., 2004, 2007) or are focussed on modern isotopic composition in precipitation and rivers (Burgman et al., 1987). This is the first systematic study of isotope lake hydrology in this region that presents data collected over a period of 6 years. We compare $\delta^{18}O_{lake}$ and $\delta^{2}H_{lake}$ from several lakes in the same region sharing common input water controlled by the isotopic composition of regional precipitation but having different hydrological sensitivities. We show the importance of characterizing the local hydrological setting before interpreting lacustrine isotope records. This is essential if relevant paleoclimate information is to be derived from such data sets, especially when the aim is ultimately to look at past changes in atmospheric circulation.

Site descriptions

The water data presented in this study were collected from lakes and streams in the north-west Swedish Lapland close to Abisko (Fig. 1), between 68°08′N to 68°30′N and 18°08′E to 19°07′E. The study area lies between 35 and 70 km from the North Atlantic. It is characterized by the Scandinavian mountain range with the highest mountains reaching *c.* 1500 m a.s.l. and the large Lake Torneträsk (322 km², 341 m a.s.l.). Local tree line is up to 600–700 m a.s.l. and largely comprises mountain birch, the upper limit of Scots pine reaches *c.* 450 m a.s.l. (Barnekow, 1999, 2000).

The climate in this region is influenced by the west-wind circulation from the Atlantic and the orographic effect caused by the Scandinavia mountain range. Atlantic derived precipitation decreases along a steep west-east gradient from the Norwegian coast to the Abisko area (Table 1). Mean annual precipitation in

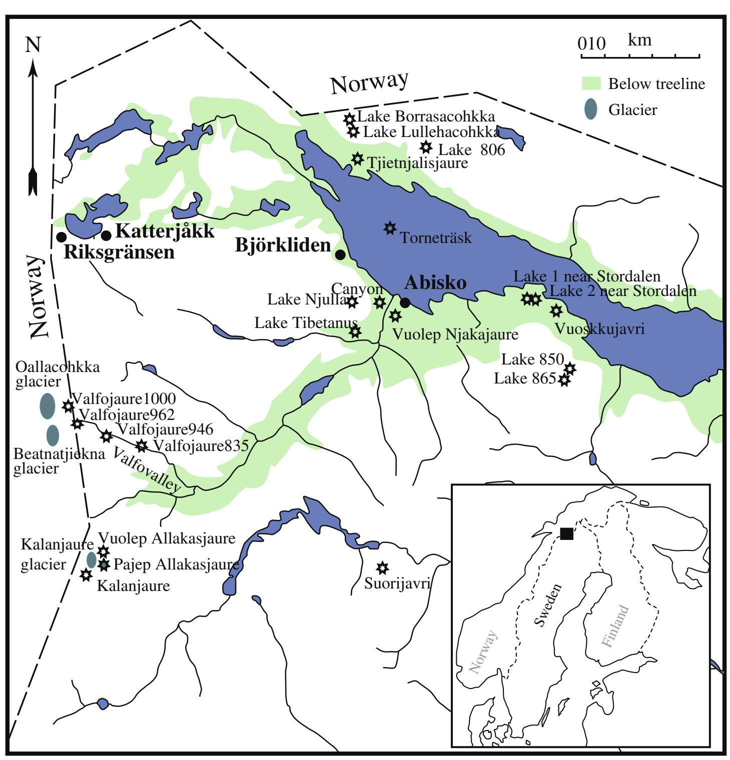

Fig. 1. Map showing the location of the sampled lakes in northern Swedish Lapland and locations of the metrological stations mentioned in the text.

Table 1
Metrological data from stations along a west-east transect through the study area (Alexandersson et al., 1991). Winter includes October–April and summer May–September. Mean annual $\delta^{18}O_p$ and δ^2H_p values for the metrological stations were calculated using the online isotopes in precipitation calculator (OIPC, http://www.waterisotopes.org).

	Riksgränsen	Katterjåkk	Björkliden	Abisko	Torneträsk
Altitude (m a.s.l.)	508	500	360	388	393
Location	68°25′N 18°08′E	68°25′N 18°10′E	68°23′N 18°41′E	68°21′N 18°49′E	68°13′N 19°43′E
Distance from Abisko (km)	30 NW	28.5 NW	8 NW	–	38 E
Monitoring period prec.	1961–1973	1969–1990	1961–1969	1961–1990	1971–1990
Monitoring period temperature	1961–1972	1972–1990	No data	1966–1990	1971–1990
Mean annual prec. (mm)	1001	848	652	304	472
Mean winter prec. (mm)	568	500	329	147	207
Mean summer prec. (mm)	433	347	322	157	263
Mean annual temperature (°C)	−1.4	−1.7	No data	−0.8	−1.0
Mean winter temperature (°C)	−7.1	−7.4	No data	−6.8	−7.4
Mean summer temperature (°C)	6.5	6.5	No data	7.6	7.9
Mean January temperature (°C)	−11.4	−11.9	No data	−11.8	−12.7
Mean July temperature (°C)	10.1	10.5	No data	11.1	11.8
Mean annual $\delta^{18}O_p$ (‰)	−14.6	−14.6	−14.4	−14.4	−14.4
Mean annual δ^2H_p (‰)	−108	−108	−106	−107	−107

Riksgränsen is *c.* 1000 mm (1961–1973; Alexandersson et al., 1991) whereas the Abisko Scientific Research Station, located in the rain shadow of the mountains, yields a mean annual precipitation of only *c.* 300 mm (1913–2006; 388 m a.s.l.; Abisko Scientific Research Station, 2007). Approximately 40–55% of the annual precipitation falls during the summer months (June–September) (Fig. 2). Due to the variable topography in the area there is large spatial variation in precipitation and runoff (Jutman, 1995). At Katterjåkk, which is located 1.5 km from Riksgränsen, 40% of the total winter precipitation is received when westerly winds prevail and more than 20% when north-westerly winds occur (Eriksson, 1987). Although the majority of the annual precipitation is derived from the Atlantic Ocean, there are periods when air masses from the north and north-east penetrate south as a result of more southerly position of the polar front (Rosqvist et al., 2007). Precipitation is also derived from a south-easterly flow of air masses from the Baltic Sea (Uvo, 2003). These air masses have different isotopic signatures, the north and north-easterly being more depleted.

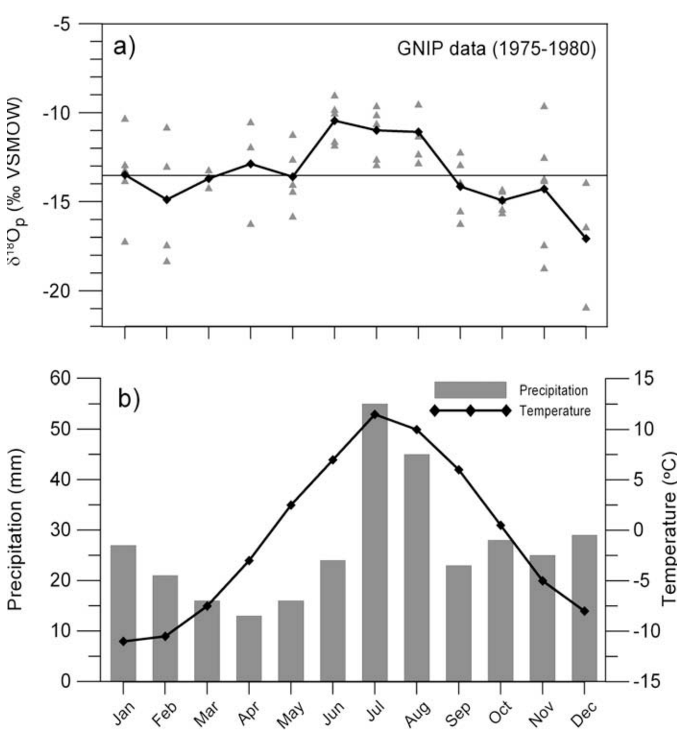

Fig. 2. (a) Monthly average values of $\delta^{18}O$ in precipitation for 1975–1980 (IAEA/WHO GNIP-data, Abisko). Solid horizontal line indicates the weighted annual average for $\delta^{18}O$ of precipitation (−13.4‰). Grey triangles refer to individual monthly sample and black diamonds to monthly means. (b) Monthly mean values of temperature and amount of precipitation for the period 2002–2006 at the Abisko Scientific research Station (ANS, 388 m a.s.l.).

Mean annual temperature (1961–1990) at the Abisko Scientific Research Station is −0.8 °C, and average values for January and July are −11.9 °C and +11.0 °C, respectively. Riksgränsen has a mean annual temperature of −1.4 °C and average values for January and July are −11.4 and +10.1 °C, respectively (1961–1972; Alexandersson et al., 1991). Slightly lower summer temperatures are likely in the higher elevations of the study area. Lakes are generally ice-free from late May or late June to early or mid-October (3–4 months), depending on elevation and local climate. At high elevations snow patches remain in the catchments throughout the entire summer.

Records of isotopic composition in precipitation from northern Scandinavia are limited but $\delta^{18}O_p$ data are available from the meteorological station in Abisko for the years 1975–1980 (IAEA/WMO, GNIP database 2007). The long-term monthly weighted average annual $\delta^{18}O$ of precipitation is −13.4‰ for this period (Fig. 2). The average seasonal amplitude of $\delta^{18}O_p$ is 7‰ with average winter precipitation (October–May) having a more depleted $\delta^{18}O$ value (−14.4‰) than average summer precipitation (June–September) (−11.7‰).

The lakes that were sampled represent a south-west to north-east transect from the Norwegian boarder to Abisko (Fig. 1) and the mean catchment elevation varies between 415 and 1270 m a.s.l. (Table 2). The majority of the lakes sampled are open, through-flow lakes. They have different surface areas and depths and the mean residence time (MRT) for the water in the lakes range from 2 weeks up to 4 years, estimated from a simple water balance where MRT is defined as V/Q, where V is the lake volume calculated from maximum depth/2 and lake surface area whereas Q is the flow rate of the water calculated from annual runoff (Jutman, 1995) and catchment area. Periods of fast snowmelt or high precipitation will probably cause higher through flow rates and decreasing residence times of the lake waters.

Methods

Water samples for isotope analysis were collected from 16 lakes and streams between 2001 and 2006, with various sampling frequency (Table 3). The dataset comprises 77 samples. Three of the lakes were sampled on two occasions during the summer of

Table 2
Physical characteristics of the study lakes. Annual runoff is from Jutman (1995). Mean catchment elevation was calculated from hypsometric curves based on the 50 × 50 m^2 digital elevation model (DEM) from the National Land Survey of Sweden (Lantmäteriverket). Residence times were estimated from catchment area, lake volume and annual runoff. Lakes are arranged in ascending order of their mean residence time.

Lake		Lake elevation (m a.s.l.)	Maximum water depth (m)	Surface area (km^2)	Catchment area (km^2)	Mean catchment elevation (m a.s.l.)	Lake: catchment ratio	Mean residence times (week)	Annual runoff (mm)
Lake Tibetanus	(68°20′N,18°42′E)	560	3.9	0.0085	1.1	715	1:129	2	550
Lake Njulla	(68°22′N,18°42′E)	999	4.2	0.01	0.78	1050	1:80	2.5	550
Valfojaure 835	(68°15′N,18°15′E)	835	19	0.25	43.0	1030	1:172	3	850
Kalanjaure	(68°08′N,18°07′E)	996	–	0.12	10.3	1270	1:86	4	1000
Vuolep Allakasjaure	(68°10′N,18°10′E)	995	11	0.4	22.7	1220	1:57	5	1000
Valfojaure 946	(68°16′N,18°11′E)	946	27	0.4	25.9	1115	1:65	13	850
Lake Borrasačohkka	(68°30′N,18°44′E)	815	5	0.08	0.6	1075	1:8	24	700
Lake 865	(68°17′N,19°07′E)	865	11	0.1	2.0	910	1:20	29	350
Lake 850	(68°18′N,19°07′E)	850	8	0.02	0.3	865	1:18	32	350
Vuoskkujávri	(68°20′N,19°06′E)	348	18	0.68	11.3	680	1:17	51	400
Vuolep Njakajaure	(68°20′N,18°47′E)	409	14	0.13	0.6	415	1:5	225	500

2001 and six lakes on two occasions during the summer of 2006, in early July and late August/early September to measure the isotopic variation of the lake water during one open water season. These lake waters were also sampled at different depths. Most of the lake water samples were taken from the deepest part of the lake using a boat and Ruttner sampler; a few samples were taken from the littoral zone although all the lakes are well mixed. Samples collected in April and May was taken through the ice. All samples for isotope analysis were collected in acid-washed airtight polyethylene bottles. Unfiltered samples from sites sampled in 2002 were subsequently treated with zinc turnings at 500 °C to generate hydrogen for D/H analysis and thereafter this method was superseded by Cr reduction (all other samples). Samples were equilibrated with CO_2 using an ISOPREP 18 device for $^{18}O/^{16}O$ analysis. Mass spectrometry was performed on a VG SIRA (Zn δ^2H and $\delta^{18}O$) and Micromass IsoPrime (Cr δ^2H) in conjunction with laboratory standards calibrated against NBS standards at the NERC Isotope Geosciences Laboratory (NIGL), UK. Analytical errors are estimated as 0.05‰ for $\delta^{18}O$, 2‰ for δ^2H (Zn) and 1‰ for δ^2H (Cr).

The ^{18}O and δ^2H data collected in this study was supported by isotope data from lake waters, streams and groundwater from 1995 to 2000 (58 samples) reported in Shemesh et al. (2001), Hammarlund et al. (2002) and Rosqvist et al., 2004 (Table 4).

All stable isotope results are expressed as 'δ' values, representing deviations in per mil (‰) from VSMOW standards for oxygen and deuterium, such that $\delta_{sample} = 1000[(R_{sample}/R_{VSMOW}) - 1]$, where R is the $^{18}O/^{16}O$ or $^2H/^1H$, ratio in sample and standard.

Results and discussion

GMWL and LMWL

The $\delta^{18}O$ and δ^2H composition of surface waters from the Abisko region from 1995 to 2006 (Fig. 3, Tables 3 and 4) show that the isotopic composition for almost all waters lie on or close to the GMWL with the exception of Vuolep Njakajaure and Lake 1 near Stordalen. The spread of the oxygen isotope composition in the waters falls mainly between the weighted average winter $\delta^{18}O_p$ (−14.4‰) and weighted average summer $\delta^{18}O_p$ (−11.7‰) calculated from Abisko precipitation data (GNIP, 1975–1980). Thus these lakes mostly preserve the isotopic signal of seasonally changing meteoric water. The seasonality in the data suggests that groundwater and lake waters generally have short residence times and that the seasonality in the isotopic composition of precipitation is preserved in the surface water composition. Since few of the lakes appear to be affected by evaporation, the LMWL for the region was determined from a linear regression of the lake water and stream water $\delta^{18}O$ and δ^2H data (excluding data from Vuolep Njakajaure and Lake 1 near Stordalen, Fig. 3). The equation of the LMWL is $\delta^2H = 7.2\delta^{18}O + 0.3$ ($r^2 = 0.84$), with a slope close to the GMWL. The water from Vuolep Njakajaure and Lake 1 near Stordalen plots below the LMWL, on a LEL indicating that water in these lakes has been affected by evaporation.

Lake water depth profiles of $\delta^{18}O$ and δ^2H (Table 3) show no isotopic stratification in any of the lakes at the time of sampling, and no isotopic enrichment of the surface water as a response to evaporation, indicating that these lakes have a well mixed water column, both shortly after the spring snowmelt and in late summer. Lake, outflow and inflow waters are all similar.

Overall, lake waters plot close to the GMWL, suggesting that the lake waters are insignificantly affected by isotopic enrichment due to evaporation (Fig. 3). This does not mean that there is no evaporation fractionation occurring at the surface of these lakes; rather, that the evaporation has not caused any significant change of the $^{18}O_{lake}$, probably because most of the lakes have short residence times and long periods of ice cover. Both these features act to minimize the effect of evaporation. During dry and warm summers, δ^1O_{lake} will probably change slightly through evaporation. In fact, there is some indication of this in the isotopic composition of lake water sampled in 2006. These data plot on a LEL with a slope of 4.8 ($r^2 = 0.97$) (Fig. 3), and intersect the LMWL at $\delta^{18}O = -13.5$‰, a value close to the mean weighted annual $\delta^{18}O_p$ (−13.4‰). This indicates that some lake waters (although they mostly fall within the overall scatter around the GMWL) during the summer 2006 were possibly affected by evaporation (the extreme end members of the LEL being Vuolep Njakajaure and Lake 1 near Stordalen). July and August 2006 were very dry in Abisko with only 54 mm of rainfall compared to the long-term July and August average of 93 mm (Abisko Scientific Research Station 1913–2006). Further intraseasonal sampling is needed to test this observation.

The studied lakes have residence times that range from a few weeks up to several years (Table 2). Currently, most lakes in the Abisko region receive sufficient inputs of recharge water to diminish the effects of summer evaporation. An exception is Vuolep Njakajaure which gave a $\delta^{18}O_{lake}$ value of around −9‰ in 2006, this lake was most significantly affected by evaporation because of its low lake/catchment ratio, a comparatively long residence time (>4 years) and the relatively low catchment mean elevation (415 m a.s.l.). The lake water is probably at or heading towards hydrologic and isotopic steady state, where the isotope composition of in- and out-flowing fluxes are equal and the isotopic composition of lake water, precipitation and river water will achieve

Table 3
Isotopic data from water samples collected in the Abisko area from 2001–2006, analysed at NERC Isotope Geosciences Laboratory, British Geological Survey, Nottingham, UK. d = deuterium excess value ($\delta^2H - 8\ \delta^{18}O$), analytical error ~1.5.

Site and water depth	$\delta^{18}O$ (‰)	δ^2H (‰)	d
July 2001			
Lake Borrasačohkka 2 m	−12.18	−85.80	11.6
Lake Lullehačorru 1 m	−11.72	−81.71	12.0
Lake 806 surface	−12.46	−88.11	11.6
September 2001			
Lake Borrasačohkka 2 m	−11.46	−82.21	9.4
Lake Lullehačorru 1 m	−11.60	−82.83	9.9
Lake 806 surface	−12.12	−87.09	9.9
April 2002			
Lake Tibetanus	−14.0	−102.8	9.3
Valfojaure835	−12.8	−92.2	10.3
Valfojaure946	−13.3	−95.5	10.7
August 2002			
Stream below Beatnatjiekna	−12.4	−83.2	15.9
Glacier mw from Oallačohkka	−12.4	−81.7	17.7
Valfojaure1000	−12.4	−84.9	14.7
Valfojaure1000 inlet1	−12.7	−86.9	14.8
Valfojaure1000 inlet2	−12.6	−85.3	15.3
Valfojaure1000 inlet3	−12.6	−86.5	14.1
Valfojaure962 inlet1	−12.4	−85	14.3
Valfojaure962 inlet2	−12.0	−81.6	14.7
Valfojaure962	−12.6	−87.1	13.5
Valfojaure946 inlet1	−11.8	−82.1	12.5
Valfojaure946 inlet2	−12.2	−84.8	12.6
Valfojaure946	−12.8	−89.9	12.3
Valfojaure835 inlet1	−12.7	−89	12.3
Valfojaure835	−12.7	−88.6	13.1
Valfojaure835 outlet	−12.7	−88.3	13.5
Valfojaure835 inlet2	−12.1	−85.5	11.2
Valfojaure835 inlet3	−12.0	−83	12.6
May 2003			
Valfojaure835 outlet	−13.9	−96.6	14.6
Vuolep Allakasjaure outlet	−13.9	−97.5	13.7
September 2005			
Tjietnjalisjaure surface	−12.3	−88.2	10.2
Tjietnjalisjaure inlet2	−12.8	−90.3	12.1
Tjietnjalisjaure inlet4	−12.7	−89.1	12.5
Tjietnjalisjaure inlet5	−12.3	−87.2	11.2
Tjietnjalisjaure inlet8	−12.2	−88.1	9.5
Tjietnjalisjaure outlet2	−12.7	−90.3	11.3
July 2006			
Vuoskkujavri 2 m	−13.05	−94.89	9.5
Vuoskkujavri 6 m	−13.08	−95.44	9.2
Vuoskkujavri 12 m	−13.11	−95.29	9.6
Vuolep Allakasjaure 0 m	−14.39	−103.03	12.1
Vuolep Allakasjaure 5 m	−14.44	−102.82	12.7
Vuolep Allakasjaure in	−14.45	−102.77	12.9
Vuolep Allakasjaure outlet	−14.36	−101.37	13.5
Kalanjaure 1 m	−14.63	−103.72	13.0
Kalanjaure 6 m	−14.51	−101.74	14.5
Lake 850 surface	−13.26	−96.48	9.6
Lake 850 3 m	−13.24	−96.53	9.4
Lake 850 7 m	−13.34	−96.58	10.1
Lake 850 inlet	−13.47	−96.18	11.6
Lake 850 outlet	−13.09	−94.84	9.9
Lake 865 3 m	−13.69	−98.35	11.2
Lake 865 inlet	−13.99	−99.31	12.6
Lake 865 outlet	−13.67	−98.77	10.6
Lake Njulla 1 m	−13.97	−99.92	11.8
Lake Njulla outlet	−13.97	−100.27	11.5
Vuolep Njakajaure 1 m	−9.63	−80.40	−3.4
Lake Tibetanus surface	−13.80	−100.2	11.7
August 2006			
Vuoskkujavri 1 m	−12.58	−92.56	8.0
Vuoskkujavri 6 m	−12.54	−92.50	7.8
Vuoskkujavri 12 m	−12.56	−92.48	8.0
Vuolep Allakasjaure 0 m	−13.64	−97.00	12.1
Vuolep Allakasjaure 4 m	−13.61	−97.27	11.6
Vuolep Allakasjaure 9 m	−13.64	−96.37	12.8
Vuolep Allakasjaure inlet	−13.63	−96.05	13.0
Vuolep Allakasjaure outlet	−13.64	−96.88	12.3
Kalanjaure 1 m	−13.64	−97.02	11.9
Kalanjaure 6 m	−13.62	−98.04	11.2
Lake 850 0 m	−12.29	−90.59	7.7
Lake 850 2 m	−12.35	−91.52	7.3
Lake 850 8 m	−12.35	−91.57	7.2
Lake 850 outlet	−11.88	−89.10	6.0
Lake 865 4 m	−13.12	−94.38	10.6
Lake 865 inlet	−13.52	−96.92	11.2
Lake 865 outlet	−13.00	−93.94	10.1
Lake Njulla 1 m	−12.56	−91.60	8.9
Lake Njulla inlet	−13.14	−95.05	10.0
Lake Njulla outlet	−12.51	−92.26	7.8
Vuolep Njakajaure 2 m	−9.13	−76.71	−3.7
Lake Tibetanus surface	−13.78	−100.6	8.5

an equilibrium (Gat, 1996). It is difficult to move a lake from steady state without a change in climate regime, so this lake will be interesting to monitor in the years to come.

Input water

Precipitation, groundwater and lake water interactions in northern Scandinavia have not been previously investigated in any detail, and little is known about the extent of input of groundwater versus direct precipitation to the lake and catchment runoff into lakes. In general the $\delta^{18}O$ of groundwater represents the local mean weighted annual composition of precipitation ($\delta^{18}O_p$) (Clark and Fritz, 1997) and for the Abisko area this value is −13.5‰. Groundwater fed lakes is generally more seasonally stable than lakes fed predominantly by precipitation and/or 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 were the water residence times are relatively short (Vitvar and Balderer, 1997). For example, in northern Finland aquifers show distinct seasonal cycles where the highest $\delta^{18}O$ values were measured in the autumn or mid-winter and the lowest values usually in the spring after snowmelt (Kortelainen and Karhu, 2004). Similarly, the rivers in northern Sweden are strongly affected by the amount of snow accumulation during winter and the snowmelt in spring resulting in a marked decrease in $\delta^{18}O$ followed by a slow recovery to almost constant values over the rest of the year (Burgman et al., 1987). In northern Scandinavia, the lakes seem to receive input waters with isotopic signatures which comprise a combined signal derived from waters that include different proportion of groundwater, meltwater inflow, runoff and direct precipitation, but without high resolution interseasonal monitoring we cannot understand the proportions for each individual lake.

Seasonal changes of $\delta^{18}O$ in lake waters

Shifts in the $\delta^{18}O_{lake}$ along the LMWL indicate seasonal variations in precipitation inputs to the lake as winter precipitation in general is isotopically depleted compared to summer precipitation (Clark and Fritz, 1997). Fig. 2 shows this for the Abisko region. Deuterium excess (d-excess = $\delta^2H - 8\ \delta^{18}O$, Dansgaard, 1964) although a complicated phenomena can provide information about changes in precipitation seasonality or moisture source in the region (Saulnier-Talbot et al., 2007) in relation to the GMWL, which has d-excess of +10‰. Generally, local precipitation has a seasonal cycle with d-excess values usually higher for winter precipitation than for summer precipitation (Gat, 1996; Froehlich et al., 2002), a pattern that is seen in UK rainfall and ascribed to evaporation of precipitation in the summer months (Darling and Talbot, 2003). The

Table 4
Published isotopic data from water samples collected in the Abisko area from 1995–2000. d = deuterium excess value ($\delta^2H - 8\,\delta^{18}O$). Isotopic data from 1995 to 1998 is reported in Hammarlund et al. (2002) (analytical uncertainties are ±0.2‰ for $\delta^{18}O$ and ±2‰ for δ^2H), data from 1999 in Shemesh et al. (2001) (0.06‰ for $\delta^{18}O$ and 0.86‰ for δ^2H) and data from 2000 in Rosqvist et al. (2004) (0.06‰ for $\delta^{18}O$ and 0.57‰ for δ^2H).

Site and water depth	Date	$\delta^{18}O$ (‰)	δ^2H (‰)	d
Tibetanus surface	3 July 1995	−14.42	−106.0	9.4
Tibetanus inflow (gw)	3 July 1995	−14.96	−107.3	12.4
Tibetanus surface	5 July 1995	−14.47	−107.9	7.9
Tibetanus under ice	11 March 1996	−15.03	−107.0	13.2
Tibetanus surface	19 March 1997	−13.97	−98.9	13.0
Tibetanus surface	29 June 1997	−14.36	−102.1	12.8
Tibetanus 2.5 m	29 June 1997	−14.48	−101.5	14.3
Tibetanus inflow (gw)	29 June 1997	−14.76	−102.5	15.6
Tibetanus surface	5 September 1997	−13.51	−96.2	11.9
Tibetanus under ice	16 March 1998	−14.95	−105.4	14.2
Tibetanus inflow (gw)	18 March 1998	−15.65	−106.5	18.7
Tibetanus surface	7 October 1998	−14.20	−100.1	13.5
Tibetanus inflow (gw)	7 October 1998	−14.34	−101.6	13.1
Lake 850	April 1999	−13.9	−101.8	9.4
Lake 850	April 1999	−13.8	−99.9	10.5
Lake 865	April 1999	−13.9	−102.5	8.7
Lake 865	April 1999	−13.9	−100.0	11.2
Vuoskkujavri	April 1999	−13.9	−103.3	7.9
Vuoskkujavri	April 1999	−13.9	−100.9	10.3
Vuoskkujavri	April 1999	−13.9	−103.3	7.9
Vuoskkujavri	April 1999	−13.9	−101.3	9.9
Lake 850 surface	August 1999	−13.0	−97.8	6.2
Lake 850 surface	August 1999	−12.8	−94.8	7.6
Lake 850 1 m	August 1999	−13.0	−98.4	5.6
Lake 850 2 m	August 1999	−13.1	−98.7	6.1
Lake 850 4 m	August 1999	−13.1	−98.7	6.1
Lake 850 6 m	August 1999	−13.0	−97.9	6.1
Lake 850 8 m	August 1999	−13.1	−98.7	6.1
Lake 850 inlet1	August 1999	−13.7	−100.3	9.3
Lake 850 inlet2	August 1999	−12.7	−93.3	8.3
Lake 850 outlet	August 1999	−12.9	−98.4	4.8
Lake 865 outlet	August 1999	−13.4	−99.3	7.9
Lake 865 surface	August 1999	−13.2	−96.2	9.4
Lake Njulla	August 1999	−12.8	−94.5	7.9
Abisko canyon	August 1999	−13.1	−93.2	11.6
Stream to the canyon	August 1999	−12.9	−96.5	6.7
Stream to the canyon	August 1999	−12.7	−92.4	9.2
Vuoskkujavri	August 1999	−13.4	−96.7	10.5
Vuoskkujavri inlet1	August 1999	−13.6	−97.7	11.1
Vuoskkujavri inlet2	August 1999	−13.4	−96.7	10.5
Suorijarvi inlet	August 1999	−13.6	−102.3	6.5
Suorijarvi surface	August 1999	−13.8	−103.1	7.3
Torneträsk	August 1999	−12.7	−92.6	9
Torneträsk	August 1999	−12.8	−93.0	9.4
Lake 1 near Stordalen	August 1999	−10.0	−81.3	−1.3
Lake 2 near Stordalen	August 1999	−13.5	−99.9	8.1
Vuolep Allakasjaure surface	6 September 2000	−13.98	−98.41	13.4
Vuolep Allakasjaure 1 m	6 September 2000	−13.81	−99.27	11.2
Vuolep Allakasjaure 3 m	6 September 2000	−13.88	−99.14	11.9
Vuolep Allakasjaure 5 m	6 September 2000	−13.89	−98.98	12.1
Vuolep Allakasjaure 7 m	6 September 2000	−13.97	−98.90	12.9
Vuolep Allakasjaure 9 m	6 September 2000	−13.93	−98.94	12.5
Vuolep Allakasjaure inlet	6 September 2000	−13.78	−98.28	12.0
Vuolep Allakasjaure outlet	6 September 2000	−13.75	−98.70	11.3
Pajep Allakasjaure surface	6 September 2000	−14.02	−99.27	12.9
Pajep Allakasjaure surface	6 September 2000	−13.98	−99.79	12.0
Pajep Allakasjaure inlet	6 September 2000	−14.06	−100.40	12.1
Pajep Allakasjaure outlet	6 September 2000	−14.08	−99.83	12.8

lakes sampled in the Abisko region generally have higher d-excess values in early summer after snowmelt compared with values in late summer (Table 3) indicating that these lakes indeed respond to seasonal variations in the precipitation.

During the summer of 2006 a decrease in d-excess values and enrichment in the $\delta^{18}O_{lake}$ values (0.5–1.4‰) from early summer (July) to late summer (August) occurred (Figs. 4–6). Such an enrichment of $\delta^{18}O_{lake}$ during summer months is expected because the effect of the isotopically depleted winter precipitation entering the lakes as meltwater in spring and summer is gradually counteracted by the input of enriched $\delta^{18}O$ summer precipitation as the lakes are replenished by rainfall and groundwater inflow. Any evaporation of the lake water would amplify this trend. The same pattern is seen in lakes in a similar setting in northern Canada (Saulnier-Talbot et al., 2007), where the winter lake water $\delta^{18}O$ is derived from the previous summers lake water which is preserved under the ice during winter when inflow and outflow cease. The residence time and the amount of winter precipitation will determine if the lake will preserve a signal from the previous summer.

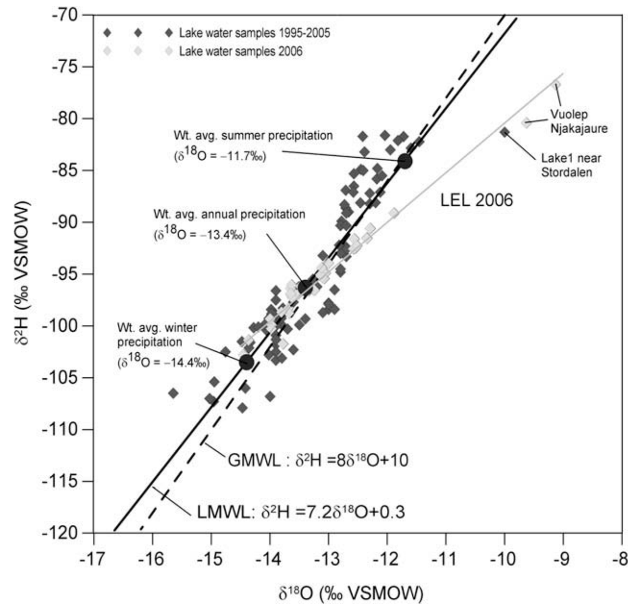

Fig. 3. Plot of $\delta^{18}O$ versus δ^2H values from lake waters, streams and groundwater springs in the Abisko area for 1995–2006 (this study; Shemesh et al., 2001; Hammarlund et al., 2002; Rosqvist et al., 2004), shown in relation to the Global Meteoric Water Line (GMWL, Craig, 1961). The Local Meteoric Water Line (LMWL) is determined from a linear regression of all the lake and stream water data excluding Vuolep Njakajaure and Lake1 near Stordalen. The weighted average precipitation values (annual, summer and winter) for Abisko are plotted on the LMWL using $\delta^{18}O$ values from the GNIP data base (1975–1980 (IAEA/WHO). The isotopic composition of the lake waters sampled in 2006 plots on a Local Evaporation Line (LEL).

The amplitude of the seasonal isotope variation is affected by the lake water residence time. For example, Lake Njulla experiences a significant seasonal variation (July–August) (∼1.4‰) (Fig. 5), while the larger and deeper Vuoskkujavri has low seasonal $\delta^{18}O_{lake}$ amplitude (∼0.5‰). The rapid response to seasonal variation in Lake Njulla is due to its relatively short residence time (2.5 weeks). The $\delta^{18}O_{lake}$ of this lake is therefore more responsive to the influence of snowmelt and summer precipitation compared to Vuoskkujavri which has a longer residence time (51 weeks). Vuoskkujavri has a greater storage capacity which buffers seasonal variation (e.g. effect of snowmelt) and will retain an isotopic signal closer to that of the annual mean precipitation. In contrast, rapid climate and hydrological change between seasons means that in fast through-flow lakes, equilibrium conditions are not archived on an annual scale (Gibson, 2002). Most of the lakes in this area are therefore isotopically non steady state lakes with the possible exception of Vuolep Njakajaure and Lake 1 near Stordalen. Another interesting lake is Lake Tibetanus, this groundwater dominated lake (Hammarlund et al., 2002) has a mean residence time of around 2 weeks, but displays only minor $\delta^{18}O_{lake}$ fluctuations around a mean of −13.8‰ (Fig. 5). This is because groundwater feeds the lake and damps the seasonal variation and produces $\delta^{18}O_{lake}$ which closely reflects the weighted average annual $\delta^{18}O_p$.

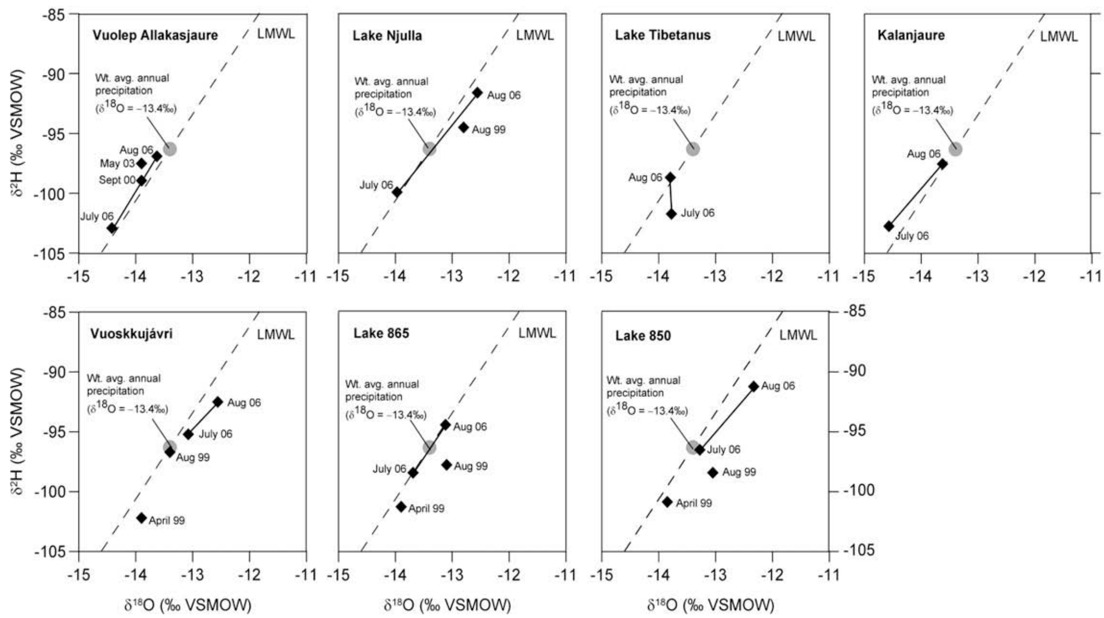

Fig. 4. Plot of $\delta^{18}O_{lake}$ versus $\delta^{2}H_{lake}$ for individual lakes sampled in July and August 2006, shown in relation to the Local Meteoric Water Line (LMWL), samples from previous years are also plotted (Shemesh et al., 2001; Rosqvist et al., 2004). Shifts along the LMWL indicate isotopic variations in precipitation inputs to the lake.

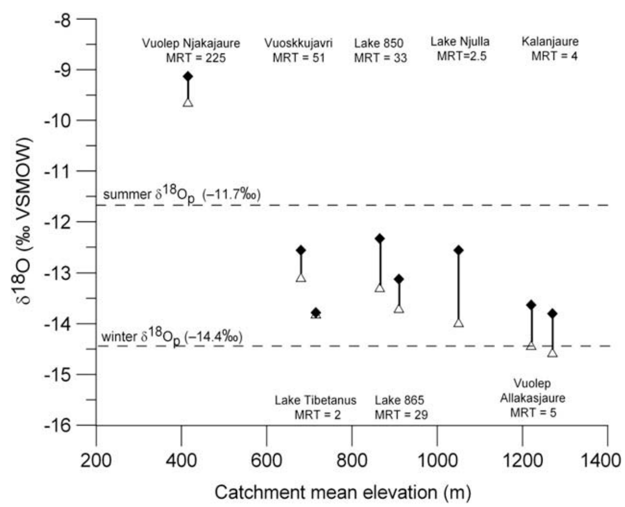

Fig. 5. Seasonal changes in $\delta^{18}O_{lake}$ for lakes sampled in 2006. Open triangles refer to July samples and diamonds to August samples. MRT is the lake mean residence time (number of weeks) and dashed horizontal lines indicate isotopic composition of the Abisko weighted average summer and winter precipitation for 1975–1980 (IAEA/WHO GNIP-data, Abisko). Analytical error is estimated as 0.05‰ for $\delta^{18}O_{lake}$.

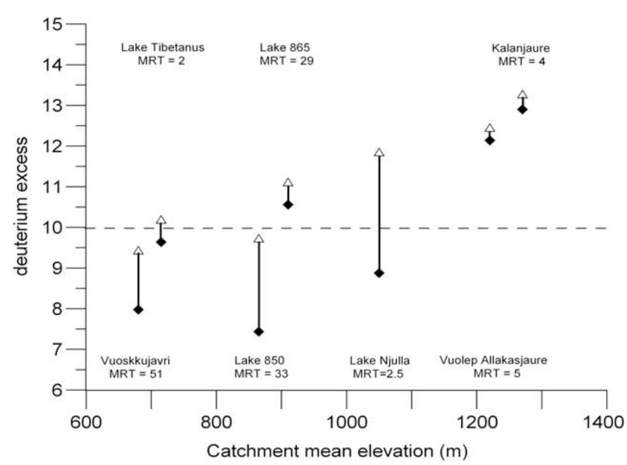

Fig. 6. Seasonal changes in lake water deuterium excess (d-excess = $\delta^{2}H - 8\ \delta^{18}O$, Dansgaard, 1964) for lakes sampled in 2006. Vuolep Njakajaure is not included in the figure because of negative d-excess values. Open triangles refer to July samples and diamonds to August/September samples. MRT is the lake mean residence time (number of weeks) and the dashed horizontal line indicates the GMWL with a d-excess of 10‰. Analytical error is estimated as 1.6 for d-excess.

Input from snowmelt can have a significant role in determining lake water $\delta^{18}O$ (Edwards et al., 2004), especially in this type of sub-Arctic or alpine environment when almost half of the annual precipitation falls as snow. The timing and the effect of the meltwater on the $\delta^{18}O_{lake}$ depend on catchment elevation, the amount of snow in the catchment and the lake to catchment ratio. The relationship between mean catchment elevation, $\delta^{18}O_{lake}$ and d-excess values in July are obvious (Table 3, Figs. 5 and 6). Lakes with higher catchment elevation have more depleted $\delta^{18}O_{lake}$ (Fig. 5) and higher d-excess (Fig. 6). At higher altitudes where the average temperature is lower, precipitation will be isotopically depleted (Rozanski et al., 1993) as a result from the cooling of air masses as they gain altitude (altitude effect). The depletion on the $\delta^{18}O_p$ from the altitude effect range from −0.15‰ to −0.5‰ per 100 m rise in altitude (Clark and Fritz, 1997). Additionally, higher catchment elevation results in later snowmelt in spring/summer because of lower temperatures, extending the period during which isotopically depleted snowmelt can influence the isotopic composition of the lake water. These together will result in depleted $\delta^{18}O_{lake}$ values. This is clearly seen in Vuolep Allakasjaure and Kalanjaure where higher catchment elevation (>1200 m a.s.l.) and the more westerly location (Fig. 1, Table 2) results in lower mean annual temperatures and higher mean annual precipitation compared to the other sites. Indeed these two alpine lakes display an almost identical lake water isotopic composition and also have the most negative $\delta^{18}O_{lake}$ values in July (−14.4‰ and −14.6‰) as well as in August (−14.0‰ and −14.2‰) (Fig. 5). This is likely due to a combination of altitude effect and late snowmelt. Furthermore, in these lakes the d-excess values in August (Fig. 6) were close to the d-excess values in July, likely due to snow contributing to the lake water over the entire

Table 5
The characteristics of a lake, which determines whether the $\delta^{18}O_{lake}$ signal reflects seasonal specific or annual $\delta^{18}O_p$, or changes in water budget (evaporation/inflow ratio).

Residence time	<6 month	>6 month or groundwater fed	>1 year
Evaporation	No	No	Yes
$\delta^{18}O_{lake}$ signal	Seasonality winter or summer $\delta^{18}O_p$	Mean annual $\delta^{18}O_p$	Evaporation/inflow ratio (E/I)

summer, especially Vuolep Allakasjaure which receives meltwater from the Kalanvare glacier. However, the similarity of the $\delta^{18}O_{lake}$ signals in these two lakes suggests that the effect of glacier meltwater input to Vuolep Allakasjaure is insignificant. This is either because the amount of glacial meltwater reaching the lake is insignificant or because the glacier meltwater has a similar $\delta^{18}O$ signature as the meltwater from the snow. In fact, meltwater from Kalanvare glacier sampled in September 2000 had a $\delta^{18}O$ value of −14.3‰ (Rosqvist et al., 2004) which is similar to the mean winter $\delta^{18}O_p$ (−14.4‰). Despite the fact that these lakes are fast throughflow lakes, the $\delta^{18}O$ of these lake waters will never attain a summer $\delta^{18}O_p$ value. Instead the lake water composition consists of a mix of winter and summer $\delta^{18}O_p$. Under this type of catchment conditions the potential effect of evaporation will be very small.

Lake 865 and Lake 850 are located only 350 m apart (Fig. 1) and are affected by the same climate conditions, have similar lake depth (11 and 8 m), mean catchment elevation (910 and 865 m a.s.l.) and mean residence times (29 and 32 weeks) (Table 2). These are the two most easterly lakes in this study. As expected, $\delta^{18}O_{lake}$ values for both lakes are higher compared to $\delta^{18}O_{lake}$ values in Vuolep Allakasjaure and Kalanjaure in July due to their lower altitude and less abundant winter precipitation. In August the lake water in Lake 850 shows a slightly larger alteration (0.95‰) towards enriched $\delta^{18}O$ values compared to Lake 865 (0.6‰) (Fig. 5). The semi-permanent snowfields on the north facing slope in the catchment of Lake 865 will contribute isotopically depleted meltwater to the lake all through the summer, damping the effect of more enriched summer precipitation. The catchment of Lake 850 is often completely snow free in July and lake surface water inlets are impossible to identify later in the summer season. As a result the lake water isotopic composition is closer to summer mean $\delta^{18}O_p$ with a slight effect of evaporative enrichment (Fig. 4).

Water samples collected from the Valfovalley hydrological system in August 2002 (Fig. 1, Table 3) show a consistent pattern in the $\delta^{18}O$. The lake system, consisting of 6 alpine lakes interconnected by streams, extends over 10 km and spans a catchment elevation of 850–1480 m a.s.l. The upper lakes receive direct input of glacier meltwater from glaciers on Mount Oallačohkka and Mount Beatnatjiekna on the Norwegian side of the boarder. The mean residence times for Valfojaure 946 and Valfojaure 835 are 12 and 3 weeks, respectively. $\delta^{18}O$ values in the system range from −12.7‰ to −11.8‰, i.e. a total variation of less than 1‰, which is close to the weighted average summer $\delta^{18}O_p$ (−11.7‰). However, the high *d*-excess values in these lakes (12.3–15.3) and the fact that the meltwater $\delta^{18}O$ from the Mount Oallačohkka glacier (−12.4‰) is similar to the $\delta^{18}O$ in these lakes indicates that the isotopic hydrology is mainly controlled by winter precipitation, suggesting that the winter $\delta^{18}O_p$ in this area is less depleted compared to the more easterly located Abisko.

Implications for palaeoclimate research

Our results show that lakes in northern Sweden have the potential of retaining many different aspects of water isotope composition which can be used for palaeoclimate reconstructions. These reconstructions can be divided into three types depending on lake characteristics: (i) changes in seasonally specific $\delta^{18}O_p$, (ii) changes in annual $\delta^{18}O_p$ in lakes with no evaporation and (iii) changes in evaporation to inflow ratio (E/I) in lakes were lake water plots on a LEL, reflecting lake water balance (Table 5). The seasonal variations in the isotopic composition of the lake waters are larger in smaller lakes with short residence times (<6 months), as they respond faster on seasonal changes in the precipitation, compared to larger lakes with longer residence times (>6 months) that retain a signal close to mean annual $\delta^{18}O_p$.

The oxygen isotope signals recorded in lake sediment materials may represent different $\delta^{18}O_{lake}$ signals depending on the lake specific hydrology and the timing of the production/precipitation/blooming. Early blooming diatoms in small lakes, for example, may actually be capturing winter $\delta^{18}O_p$ through the early summer thaw, while summer carbonates may capture mean weighted annual $\delta^{18}O_p$ if forming in bicarbonate rich groundwater fed lakes. Therefore interpretation of the $\delta^{18}O$ record requires a detailed knowledge of the processes that control and modify the lake water signal, and this must be determined for each individual lake system. A possible explanation of changing $\delta^{18}O_{lake}$ over time is a change in the seasonal distribution of precipitation. For example a trend towards more depleted values in the $\delta^{18}O$ sediment record might suggest a progressively increasing contribution of winter precipitation. More snow accumulation in the catchments would lengthen the period during which depleted snowmelt influence the $\delta^{18}O_{lake}$ in spring and summer, potentially even influencing the isotopic composition of late mineral precipitates production.

Conclusions

Based on results from $\delta^{18}O$ and $\delta^{2}H$ analyses of surface water collected from lakes and streams in northern Swedish Lapland we conclude that lakes in this sub-Arctic region are currently mainly recharged by meteoric water which undergoes little subsequent evaporation (with the exception of 2 of the 21 lakes). Overall, our results reveal that northern Scandinavian lake waters have a range of $\delta^{18}O$ values (−9.1‰ to −14.6‰), these signatures are forced by local hydrology and different seasonal parameters. In general the seasonal variation of $\delta^{18}O_{lake}$ is determined by the input of isotopically depleted snowmelt (winter precipitation) to the lakes in the early summer and the influence of relatively enriched summer precipitation and groundwater (in some lakes) and the effect of evaporation later in the summer. Lake waters also show a range of $\delta^{18}O$ between sites due to catchment elevation and different residence times, their waters contain different proportions of summer and winter precipitation. As expected the seasonal variations in the isotopic composition of the lake waters are larger in lakes with short residence times, like Lake Njulla, which react faster to seasonal change in precipitation, compared to lakes like Vuoskkujavri, with longer residence times, which retain an isotopic signal closer to that of annual mean precipitation. A short residence time (<6 months) means that $\delta^{18}O_{lake}$ is unlikely to reflect the isotopic composition of annual mean precipitation, except in Lake Tibetanus which are fed by regional groundwater. Slow melting of large amounts of snow extends the period during which depleted snowmelt influences $\delta^{18}O_{lake}$.

By collating data collected over the 11 years by three different teams, we show that most of the lakes sampled in northern Sweden are not currently evaporating and as such retain different aspects of $\delta^{18}O_p$ (seasonal or annual). Therefore we want to emphasize the importance of understanding the specific seasonality of each lake's hydrology as this must be considered before lake sediment archives of $\delta^{18}O$ (from carbonates, diatoms, etc.) can be used for palaeoclimate reconstructions. Further sampling and analysis of isotopic data from precipitation, groundwater and lake water are needed to improve the understanding of the modern and past relationship between climate changes, $\delta^{18}O_p$ and $\delta^{18}O_{lake}$.

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