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Quantifying groundwater dependence of a sub-polar lake cluster in Finland using an isotope mass balance approach

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Abstract

A stable isotope study of 67 kettle lakes and ponds situated on an esker aquifer (90 km^2) in northern Finland was carried out in the summer of 2013 to determine the role of groundwater inflow in groundwater-dependent lakes. Distinct seasonal fluctuations in the $\delta^{18}O$ and δ^2H values of lakes are the result of seasonal ice cover prohibiting evaporation during the winter. An isotope mass balance approach was used to calculate the inflow-to-evaporation ratios (I_{TOT}/E) of all 67 lakes during the summer of 2013 when the isotopic compositions of the lakes were approaching a steady-state. The normalised relative humidity needed in this approach came from assuming a terminal lake situation for one of the lakes showing the highest isotope enrichment. Since evaporation rates were derived independently of any mass balance considerations, it was possible to determine the total inflow (I_{TOT}) and mean turnover time (MTT) of the lakes. Furthermore, the groundwater seepage rates of those lakes revealing no visible surface inflow were calculated. Here, a quantitative measure was introduced for the dependence of a lake on groundwater (G index) that is defined as the percentage contribution of groundwater inflow to the total inflow of water to the given lake. The G index values of the lakes studied ranged from 27.8–95.0 %, revealing large differences in groundwater dependency among the lakes. This study shows the effectiveness of applying an isotope mass balance approach to quantify the groundwater reliance of lakes situated in a relatively small area with similar climatic conditions.

1 Introduction

The characterisation of groundwater dependent ecosystems (GDEs) is a requirement of the Groundwater Directive (EC, 2006). These systems are often complex and their hydrology and contact with aquifers are not well established. Lakes can be dependent on groundwater directly or indirectly, and this dependence can vary over time (Kløve et al., 2011). Understanding groundwater and lake water interaction is important not

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only for water resource management (Showstack, 2004), but also for understanding the ecology and eutrophication of lakes, since groundwater may be a key element in the lake nutrient balance (Ala-aho et al., 2013; Belanger et al., 1985; Brock et al., 1982; Kidmose et al., 2013). Furthermore, the vulnerability of lakes to pollution can be controlled by their dependency on groundwater (Kløve et al., 2011). Methods such as seepage meters (Ala-aho et al., 2013; Rosenberry et al., 2008, 2013), environmental tracers (e.g. Dinçer, 1968; Shaw et al., 2013; Yehdegho et al., 1997; Zuber, 1983) and numerical modelling (e.g. Krabbenhoft et al., 1990; Stichler et al., 2008; Winter and Carr, 1980) can be used to determine the groundwater reliance of lakes.

Heavy stable isotopes of water (^{18}O, ^{2}H) can be considered as ideal tracers for studying the hydrological cycle (e.g. Clark and Fritz, 1997). Fractionation of isotopes of water is the very factor enabling their use in hydrological studies, as it governs the changes in isotopic abundances within the water cycle (Gat, 2010). At a global scale, the ^{2}H and ^{18}O isotope composition of meteoric waters cluster along the line called the global meteoric water line (GMWL), first determined by Craig (1961): $\delta^2\text{H} = 8 \cdot \delta^{18}\text{O} + 10$. Locally, this linear relationship may have a slightly different form (local meteoric water line – LMWL). Evaporation from an open water body fractionates isotopes so that the remaining liquid phase is enriched in both ^{2}H and ^{18}O in proportion with their effective fractionation factors accompanying this process. Consequently, the isotopic composition of the evaporating water body evolves in the δ-space along the line known as the local evaporation line (LEL), whose slope is significantly smaller than that characterising the local or global meteoric water lines. The position of the isotopic composition of lake water along this line is strongly related to the water balance of the lake (e.g. Gat, 1996; Gibson and Edwards, 2002; Rozanski et al., 2001).

The methodology of isotope-aided studies of the water balance of lakes has been thoroughly discussed in a number of review papers and textbooks (e.g. Froehlich et al., 2005; Gat and Bowser, 1991; Gat, 1995; Gonfiantini, 1986; Rozanski et al., 2001). Although several authors have applied isotope techniques in studying lakes in cold climates (Gibson and Edwards, 2002; Gibson, 2002; Gibson et al., 1993; Jonsson et al.,

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2009; Turner et al., 2010; Yi et al., 2008), mostly in Canada and northern Sweden, these studies were generally focused on lakes spread over large areas.

The central aim of this study was to quantify the groundwater dependence of 67 kettle lakes and ponds situated across a relatively small area (90 km^2) of the Rokua esker aquifer occupying a large glaciofluvial deposit in northern Finland. To quantify the extent of the interaction between the aquifer and the lakes, a dedicated isotope study was launched in 2013. This was part of comprehensive investigations (2010–2012) aimed at understanding the hydrology of an esker aquifer area where some of the kettle lakes and ponds have suffered from water level decline or eutrophication. Since the seasonal isotopic behaviour of the selected lakes in the study area was already fairly well understood based on the data collected from 2010 to 2012, it was decided to conduct a large-scale one-off survey of the isotopic composition of all 67 lakes on the esker in order to quantify their dependence on groundwater. Ala-aho et al. (2013), who studied 11 lakes on the esker, show that the water levels in closed-basin seepage lakes have more fluctuations than the drainage lakes, which have more stable water levels. On the other hand, the drainage lakes are more trophic. Their study also shows that subsurface flow can transport phosphate to lakes. Therefore, it was important to quantify the groundwater dependence of all lakes on the esker and propose an appropriate index reflecting this dependence. The large-scale field campaign conducted in July and August 2013 comprised the sampling of water in all 67 lakes for isotope analyses, combined with continuous temperature measurements and aerial thermal imaging of the lakes.

2 The study area

The Rokua esker aquifer area, situated in northern Finland, was formed during the last deglaciation period 9000 to 12 000 years ago (Tikkanen, 2002). As ice retreated, a long ridge formation, consisting mainly of fine and medium sand (Pajunen, 1995), was shaped. Ancient sea banks surrounding the esker show that the esker was originally an

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island that gradually rose from the sea (Aartolahti, 1973). Today, the highest elevation of the esker is 100 m above the surrounding low-lying peatlands and the layer thickness of sand ranges from 30 m to more than 100 m above the bedrock. Sea banks, dunes and kettle holes form a rolling and geologically unique terrain (Aartolahti, 1973). Kettle holes were formed when ice blocks were buried in the ground and, as they melted, left depressions in the landscape. The ground surface of the esker is mainly lichen-covered pine forests. Hydrologically, the Rokua esker is an unconfined aquifer, one of the largest in Finland, and it has two regional groundwater mounds (Rossi et al., 2014). The recharge area of the aquifer is 90 km^2 and the discharge zones are situated in the surrounding peatlands, which partially confine the aquifer (Rossi et al., 2012).

Kettle holes – long and narrow depressions – give Rokua esker its distinct character. The sizes of these kettle holes vary. They can be 1 to 80 m deep, between 10 m and 1.5 km long, and 0.4 km wide (Aartolahti, 1973). Most of the kettle holes are now dry, but due to the influence of groundwater in the past, peat has accumulated at the bottom of them, creating kettle hole mires (Pajunen, 1995). However, the alternating topography of the area is reflected in the existence of approximately 90 lakes or ponds, referred to as kettle lakes or ponds. Peat started to accumulate in the border regions of the lakes more than 8000 years ago, so most of the kettle lakes and ponds are partly paludified (Pajunen, 1995). Nevertheless, the majority of the lakes and ponds are characterised by their crystal clear water, which is reflected in the number of holiday homes and hotels on the lakeshores. The lakes are widely used for different recreational activities, such as swimming, fishing and scuba diving (Anttila and Heikkinen, 2007). The uniqueness of the glaciofluvial formation of Rokua, in which the actions of ice, water and wind can be seen, has been recognised in many ways. Some of the Rokua esker is protected by Natura 2000 and by the Finnish nature reserve network. Rokua was recently chosen to be part of the UNESCO GeoPark Network and is currently the northernmost region in this network.

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3 Materials and methods

3.1 Hydrological measurements and thermal imaging

During 2010–2012, 11 lakes, 13 piezometers and 11 streams were sampled in the study area four times per year to analyse the stable isotope composition of water, nutrients, water quality parameters (T, pH, E.C., O_2) and geochemical parameters (silica, major cations and anions) (Fig. 1). During the field campaign conducted in July and August 2013, a total of 67 lakes and ponds were surveyed for the same parameters and thermal images of lakes taken from the air in a helicopter using a FLIR thermal camera. In addition, composite precipitation samples were mainly collected once a month during the open water season at the station on the esker during 2010–2013. Precipitation samples for winter were collected once a year before snowmelt by taking a uniform sample of the whole snowpack depth.

Water quality parameters were analysed in the field using WTW Multi 3430 or Multi 350i meters for oxygen, EC and pH. Samples of lake water were collected with a Limnos sampler, approximately 1 m below the water surface and 1 m above the bottom of the lake. If the depth of the lake was less than 2 m, only one sample was taken and if it was more than 20 m, samples were taken from the middle as well. Depending on their shape and size, the lakes had between 1 and 4 sampling locations. Stream samples were collected by submerging a bottle in water, facing upstream. Piezometers were pumped for at least 10 min prior to taking groundwater samples or until the colour of the water was clear. The samples were collected one metre below the water table. All sampling bottles (HDPE) were rinsed with the sampled water prior to filling. Samples for isotope analyses were stored in the dark at a cold temperature (4 °C ± 2 °C).

The isotopic composition of water samples was analysed using CRDS technology with a Picarro L2120-i analyser. Samples with visible colour or suspended matter were filtered (pore size 25 µm) prior to analysis. The measured $^{2}H/^{1}H$ and $^{18}O/^{16}O$ isotope ratios are reported as relative deviations from the VSMOW standard. Typical uncertainty of the reported $\delta^{18}O$ and $\delta^{2}H$ values are ±0.1 and ±1.0 ‰, respectively.

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Lake water temperature was measured continuously during the ice-free period in 2013 for two lakes on the esker: Ahveroinen (3.3 ha) and Saarinen (15.3 ha). Hobo loggers (pendant temperature data logger UA-001-08 and conductivity logger U24-001, accuracy 0.1 °C) were installed 50–100 cm below the lake surface. In addition, the surface water temperature for lake Oulujärvi (92 800 ha) was obtained from the database of the Finnish Environmental Institute (2013). Lake Oulujärvi is located east, next to the study site, 1 km from the easternmost lake studied. Thermal imaging of the lakes was conducted on 5 August 2013 using a Flir Thermacam P-60 thermal camera. This camera had 320 × 240 pixel sensor resolution and an opening of 24°. It covered the electromagnetic spectrum from 7.5 to 13 µm. The imaging was taken by helicopter 150 m above the lakes. The image data were correlated to the predominant weather conditions (temperature and relative humidity) with data from the FMI Pelso weather station measured every 10 min.

Depth profiling was undertaken in the lakes for which no depth contour lines are available (National Land Survey of Finland, 2010a). It was carried out with a portable depth-sounding radar (resolution 0.1 m) or with a measuring cable and GPS system. Typically, two profiles from the shore to the deepest point were defined. The number of measurement points differed between the lakes depending on their size. In total, 52 lakes were surveyed.

3.2 Lake volumes

The volumes of the lakes were determined in an ArcGIS environment using depth profiling measurements, contour lines and border lines. The water surface levels of the lakes were estimated using elevation levels presented in the basic map (National Land Survey of Finland, 2010a). Lake morphology was mostly interpolated using spline that results in a smooth surface passing by all the input points (ESRI, 2014). The Tension method with 0.1 weight and 3 input points was used to calculate the values for the interpolated cells. Interpolation rasters were extracted by surface water areas. The

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mean depths of these new rasters were multiplied by the water surface areas in order to calculate the volumes of the lakes.

3.3 Evaporation from lakes

Evaporation was calculated individually for all the lakes surveyed using a mass transfer approach (Rosenberry et al., 2007; Dingman, 2008). The following expression was used:

$$E = K_\mathrm{E} \cdot v_a \cdot (e_\mathrm{s} - e_\mathrm{a}), \tag{1}$$

where:

- K_E is the mass-transfer coefficient in $\mathrm{m\,km^{-1}\,kPa^{-1}}$ describing the impact of turbulent eddies of the wind on vertical transport of water vapour from the lake with area A_L ($\mathrm{km^2}$), $K_\mathrm{E} = 1.69 \times 10^{-5} \cdot A_L^{-0.05}$ (Harbeck, 1962)
- v_a is wind speed ($\mathrm{m\,s^{-1}}$) at 2 m height
- e_s is the saturation vapour pressure in kPa at surface water temperature T_s (°C), $e_\mathrm{s} = 0.611 \cdot \exp(\frac{17.3 \cdot T_\mathrm{s}}{T_\mathrm{s}+237.3})$
- e_a is the vapour pressure in the air in kPa, $e_\mathrm{a} = h \cdot 0.611 \cdot \exp(\frac{17.3 \cdot T_\mathrm{a}}{T_\mathrm{a}+237.3})$, where h is relative humidity and T_a is air temperature (°C).

Wind speed measured at 10 m height was adjusted to the corresponding speed at 2 m height using the power law profile (Justus and Mikhail, 1976):

$$v_z = v_\mathrm{r}\left(\frac{z}{z_\mathrm{r}}\right)^n, \tag{2}$$

where v_r is the measured wind speed at the reference height z_r (10 m), z is the height for which speed is adjusted (2 m) and n is an empirical constant (0.15 can be used for neutral stability cases).

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The meteorological parameters necessary for the calculations (relative humidity, wind speed and air temperature) were obtained from the meteorological station 5502 (Vaala-Pelso) of the Finnish Meteorological Institute (2014), located approximately 10 km from the site. A range for the lakes' water temperature was evaluated using continuous temperature measurements from one of the studied lakes, Ahveroinen (see Sect. 4.1), and a standard deviation of lake water temperature values determined from thermal images. Using the derived temperature range, a range for evaporation from all the lakes was calculated. The adopted method relies only on temperature difference measured on one day, but since all the lakes are in a relatively small area with almost identical weather conditions, it is highly probable that the seasonal behaviour of the lake water temperature is similar between the lakes.

3.4 Isotope mass balance

When considering isotope and mass balances under hydrologic and isotope steady-state conditions, the following approximate expression describing the isotope enrichment of an evaporating lake can be derived (e.g. Gat and Bowser, 1991; Rozanski et al., 2001):

$$\Delta\delta = \delta_{LS} - \delta_{IT} \cong \frac{\delta_A - \delta_{IT} + \varepsilon/h_N}{1 + \frac{I_{TOT}}{E}\frac{1-h_N}{h_N}} \tag{3}$$

where:

$\Delta\delta$ evaporative enrichment

I_{TOT} total inflow to the lake (surface and underground fluxes plus rainfall)

E evaporation rate

h_N relative humidity over the lake, normalised to the temperature of the lake surface

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δ_A isotopic composition of atmospheric moisture over the lake (‰)

δ_{LS} measured steady-state isotopic composition of the lake (‰)

δ_{IT} isotopic composition of the total inflow to the lake (precipitation, surface and underground inflow) (‰)

ε the total effective isotope fractionation ($\varepsilon^* + \Delta\varepsilon$) where

ε^* equilibrium isotope enrichment $(1-1/\alpha_{L/V}) \times 10^3$. $\alpha_{L/V}$ stands for equilibrium isotope fractionation between liquid and gaseous phase (‰)

$\Delta\varepsilon$ kinetic isotope enrichment ($\Delta^{18}\varepsilon = C_k^{18}(1-h_N)$; $\Delta^2\varepsilon = C_k^2(1-h_N)$) where C_k^{18} and C_k^2 stand for kinetic enrichment parameters.

If it can be further assumed that the isotopic composition of atmospheric water vapour is in isotopic equilibrium with the total inflow, then Eq. (3) can be further simplified:

$$\Delta\delta = \delta_{LS} - \delta_{IT} \cong \frac{(1-h_N)\varepsilon^* + \Delta\varepsilon}{\left[h_N + (1-h_N)\frac{I_{TOT}}{E}\right]} \tag{4}$$

where ε^* and $\Delta\varepsilon$ are defined as in Eq. (3).

From Eq. (4) the expression for the total inflow-to-evaporation ratio characterising the hydrological balance of the given lake can be derived:

$$\frac{I_{TOT}}{E} \cong \frac{\varepsilon^* + C_k}{\Delta\delta} - \frac{h_N}{1-h_N}. \tag{5}$$

If $I_{TOT}/E = 1$ (terminal lake situation), Eq. (4) can be further simplified to the following formula:

$$\Delta\delta_T = \delta_{LS} - \delta_{IT} \cong (1-h_N)\varepsilon^* + \Delta\varepsilon = (1-h_N)(\varepsilon^* + C_k). \tag{6}$$

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From Eq. (6) the following expression for relative humidity normalised to the temperature of the evaporating surface can be derived:

$$h_N = 1 - \frac{\Delta\delta_T}{\varepsilon^* + C_k} \tag{7}$$

where $\Delta\delta_T$ is the heavy isotope enrichment observed in the terminal lake.

4 Results and discussion

4.1 Climate and lake water temperature data

The climate is strongly seasonal in the Rokua esker study area. The long-term monthly mean values of surface air temperature vary from −10.9 °C (January) to 13.2 °C (June) for the period between 1959 and 2013 (Fig. 2). The amount of monthly precipitation varies from 29 mm (February and April) to 79 mm (August) for the same period. The warmest months of the year are June, July and August. The long-term (1970–2013) monthly mean relative humidity of air varies from 61 % (May) to 91 % (November).

The seasonal temperature patterns of the monitored lakes were very similar, despite significant differences in lake size (Fig. 3). Thus the surface water temperature of lake Ahveroinen 1 (mean 18.60 °C) during the period from 1 June 2013 to 31 August 2013 was used as a basis for estimating the water temperature of other lakes. Thermal images collected on 5 August 2013 yielded comparable surface water temperatures that ranged from 19.5 to 24.6 °C, with a mean of 21.3 °C and a standard deviation of 0.87 °C. Combining the results of continuous temperature measurements and thermal images, an estimate of the mean surface water temperature of all lakes was derived to be 18.60 °C ± 0.87 °C. This temperature was used in the isotope mass balance calculations.

4.2 Local isotopic compositions

An overview of the isotopic composition of different types of water in the study area is presented on the δ^2H-δ^{18}O space in Fig. 4. It comprises precipitation data collected during the period 18 March 2010–29 October 2013 at the station located on the esker, the mean isotopic compositions of the selected lakes, streams and groundwater monitored during the period 2010–2013 (Table 1), as well as the isotopic compositions of 67 lakes surveyed in July and August 2013 (Table 2).

The local meteoric water line (LMWL) of Rokua ($\delta^2\text{H} = 7.77\,\delta^{18}\text{O} + 9.55$) was defined using the δ values of precipitation samples collected during the years 2010–2013 (Fig. 5). The local evaporation line (LEL), $\delta^2\text{H} = 5.09\,\delta^{18}\text{O} - 28.19$, is the best fit line of the δ values representing lake water data for the year 2013. The intersect of the LMWL and LEL lines yields the estimate of the weighted annual mean δ^{18}O and δ^2H values of precipitation (−14.1 and −100 ‰, respectively). Slightly elevated mean δ^{18}O and δ^2H values of local groundwater (−13.1 and −95 ‰), combined with reduced deuterium excess ($d = \delta^2\text{H} - 8 \cdot \delta^{18}\text{O} = 4.8$ ‰) when compared to precipitation (12.8 ‰), indicate the presence of an evaporation signal in the local groundwater. Furthermore, some of the winter precipitation is most probably returned to the atmosphere via sublimation and does not contribute to groundwater recharge.

Although the majority of groundwater samples cluster near the LMWL-LEL intersect, there are some data points in the δ^{18}O-δ^2H plot that clearly indicate the contribution of (evaporated) lake water to groundwater. This influence is illustrated in Fig. 5, showing the isotopic composition of lake Ahveroinen 1 and adjacent groundwater. The mean δ^{18}O and δ^2H values of piezometers MEA 2010 and MEA 1907 situated on the south-eastern and north-western sides of the lake were −13.4 and −97 ‰ and −10.7 and −83 ‰ respectively, clearly indicating a substantial (ca. 55 %) contribution of lake water to groundwater at the north-western side of the lake. A smaller contribution (ca. 10 %) of lake water to groundwater can be seen on the eastern and western sides of the lake where the mean δ^{18}O and δ^2H values of groundwater were −12.9 and −93 ‰

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respectively. The main direction of groundwater flow is therefore from south-east to north-west, which coincides with the results from seepage measurements conducted by Ala-aho et al. (2013). The direction of groundwater flow can also be noted from the difference in the mean isotopic composition of the lake water between points 2 and 3 of −8.7 and −73 ‰ and −8.5 and −72 ‰, respectively. The difference in isotopic compositions between points 2 and 3 was greatest during the winter, e.g. in March 2011 the difference in $\delta^{18}O$ and $\delta^{2}H$ between these points was −1.1 and −4 ‰ respectively.

4.3 Temporal variations in the isotopic composition of the selected lakes

The seasonal variability in the isotopic composition of lake water is illustrated in Fig. 6, showing the changes of $\delta^{18}O$ in lake Ahveroinen 1 at two different depths (1 and 4 m). $\delta^{18}O$ of lake Ahveroinen 1 reveals distinct seasonal fluctuations with peak-to-peak amplitude in the order of 1 ‰. After the disappearance of ice cover during the spring (April–May), the lake starts to evaporate, which results in a gradual heavy isotope enrichment of water, approaching steady-state value sometime in September–October. Freezing of the lake in the late autumn stops the evaporation flux. The systematic fall in $\delta^{18}O$ value during the ice-cover period stems from the gradual dilution of lake water with groundwater seeping into the lake. Figure 6 shows that the lake is well mixed throughout the year. Thus the declining parts of the $\delta^{18}O$ curve in Fig. 6 can be used to assess the intensity of groundwater inflow, when the volume of the lake is known and the isotopic composition of groundwater is constant. As a rough approximation, the observed reduction of $\delta^{18}O$ of the lake water by ca. 1 ‰ over the six-month period is a result of the admixture of a certain volume of groundwater with a specific isotopic signature ($\delta^{18}O_{GW} = -13.4$ ‰). It appears that the required groundwater inflow is in the order of 150 $m^3\,day^{-1}$, which is close to 137 $m^3\,day^{-1}$ produced by the total inflow-to-evaporation ratio for this lake derived from isotope mass balance considerations (see Sect. 4.4).

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4.4 Quantifying groundwater dependence of the studied lakes

The isotopic composition of 67 lakes (July–August 2013 sampling campaign) covers a wide range of δ values: from −5.6 to −12.7 ‰ and from −57 to −93 ‰ for δ^{18}O and δ^{2}H respectively. This large variability reflects a wide spectrum of the heavy isotope enrichment of the lakes studied. Since Rokua lakes are situated in a unique climatic setting, their steady-state isotope enrichment is primarily controlled by their water balance, which in turn can be characterised by the total inflow-to-evaporation ratio (cf. Eq. 3).

Equation (4) was used to calculate the total inflow-to-evaporation ratio (I_{TOT}/E) for each lake using normalised relative humidity, h_N, derived from Eq. (6) that applies to terminal lakes. Lake Kissalampi showing the highest isotope enrichment of all the sampled lakes (Table 2) and as it did not reveal any surface outflow, can reasonably be considered to be a terminal lake. The isotope enrichment for lake Kissalampi ($\Delta\delta_T$) was calculated using the isotopic composition of this lake measured in August 2013 (δ_{LS} = −5.6 ‰) and the δ_{IT} value represented by the weighted mean isotopic signature of the local precipitation (the intersect of LMWL-LEL – cf. Fig. 4). The h_N value of 64.6 % was derived from Eq. (6) using measured ^{18}O data and the adopted value for the kinetic enrichment parameter C_k (14.2 ‰; Gonfiantini, 1986). Since the air and lake temperatures were available for this period, normalised relative humidity could be also calculated using the field data, i.e. the relative humidity measured by a nearby meteorological station, and the mean air and lake water temperatures. The resulting normalised relative humidity is approximately 63.1 %, a similar value to that derived from Eq. (5). Introducing the assumption that the total inflow to each lake also contains a groundwater component (50 %) changes the best estimate of h_N based on Eq. (6) from 64.6 to 66.6 %. A calculation of the I_{TOT}/E ratio for each lake was undertaken for the period 1 June–31 August 2013 using the relative humidity value 64.6 % and δ^{18}O value of the total inflow −14.1 ‰. The I_{TOT}/E ratios obtained range from 1 (assumed

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for lake Kissalampi) to 14.8 for lake Siirasjärvi 2 and are functions of the measured isotope enrichment of lakes (Fig. 7).

Since the evaporation rates from the studied lakes were derived from Eq. (1) independently of any mass balance considerations, the total inflow to each lake could also be calculated from the assessed I_{TOT}/E ratios. Knowing the volume and total inflow, the mean turnover time of water (MTT) in each lake could be quantified as the ratio of lake volume to total inflow. The calculated MTT values are listed in Table 2. They range from 0.5 months for the Pasko pond ($V = 2 \times 10^3\,m^3$, mean depth 0.2 m, maximum depth 0.7 m) to 9.1 years for lake Saarijärvi 2 ($V = 2.47 \times 10^6\,m^3$, mean depth 11.8 m, maximum depth 26 m), with the mean in the order of 18 months. Lake Saarijärvi 2 is the deepest of all the lakes surveyed. As expected, the calculated MTT values correlate well with the mean depth of the studied lakes, expressed as the volume-to-surface area ratio (Fig. 8). The link between MTT and the I_{TOT}/E ratio is much weaker (Fig. 9); lakes with higher I_{TOT}/E ratios tend to have shorter mean turnover times.

Knowledge of the total inflow to each of the surveyed lakes (I_{TOT}) derived from isotope mass balance considerations enables an evaluation of the groundwater inflow to each individual lake by subtracting the precipitation and surface water inflow (if it exists and its value is known). Such an evaluation was undertaken for the lakes listed in Table 2, which do not reveal visible surface inflows. The resulting groundwater seepage rates (I_{GW}) vary from $3\,m^3\,day^{-1}$ for Kissalampi pond to $5556\,m^3\,day^{-1}$ for lake Rokuanjärvi 1.

The dependence of a lake on groundwater input can be quantified through an index defined as the percentage contribution of groundwater inflow to the total inflow of water to the given lake. Figure 10 summarises the values of this index (G index) obtained for the lakes surveyed during the July–August 2013 sampling campaign for which no visible surface inflow could be identified. The lowest G value (27.8 %) was obtained for Kissalampi pond, which is comparable to a terminal lake. The highest G values were derived for lakes Kiiskeroinen (92.9 %), Levä-Soppinen (93.0 %) and Siirasjärvi 2 (95.0 %). Interestingly, these lakes are characterised by a high degree of eutrophication

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induced by high loads of phosphorus brought to the lakes with groundwater (Ala-aho et al., 2013).

4.5 Uncertainty assessment

The above methodology for quantifying elements of water balance in the lakes studied introduces some uncertainties linked to the assumptions made and parameters used in the evaluation process. Although no formal uncertainty evaluation was carried out for the parameters obtained for each individual lake, sensitivity tests were performed to derive the range of uncertainties associated with the quantities being evaluated, such as mean turnover time, total inflow-to-evaporation ratio and the G index.

One of the major sources of uncertainty of MTT, I_{TOT}/E and G values derived for each individual lake is the uncertainty with regard to lake water temperature. As discussed in Sect. 4.1, the temperature of the studied lakes was estimated to be 18.6 °C ± 0.87 °C. This uncertainty of lake water temperature leads to uncertainty as regards the evaporation flux, which in turn influences the MTT and G values derived for each lake. The mean turnover time increases by ca. 16 % when the temperature of the lake is reduced by 0.87 °C, and decreases by approximately 13 % when the temperature increases by the same amount. A similar situation occurs with the G index. Its sensitivity to changes in lake water temperature, via changes in the evaporation flux, depends on the actual value of this index. As seen in the lower panel in Fig. 11, the relative change of G resulting from the change in lake water temperature by ±0.87 °C varies from ca. 37 % for $G = 27.8$ % to approximately 0.8 % for $G = 95.0$ %.

The uncertainty of the I_{TOT}/E ratio derived for each lake using Eq. (4) depends in the first instance on the uncertainty of the relative humidity of the atmosphere normalised to the temperature of the lake surface. The upper panel in Fig. 11 illustrates the impact of changing relative humidity on the I_{TOT}/E ratio across the whole range of this parameter obtained for the lakes surveyed. The relative change in the I_{TOT}/E ratio resulting from the change in relative humidity by ±2 % varies from 15–16 % for $I_{TOT}/E = 1$ to approximately 1 % for highest recorded values of this ratio (ca. 15).

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5 Conclusions

The Rokua esker, with its numerous lakes located across a relatively small area, provided a unique opportunity to explore the possibilities offered by environmental isotope techniques in quantifying the water balances of lakes and their dependency on groundwater in a sub-polar climatic setting. The quantification of groundwater seepages to lakes using conventional methods is notoriously difficult and associated with considerable uncertainty. The study presented here demonstrates the power of the isotope mass balance approach for resolving this issue. It appears that a stable isotope analysis of lake water samples, collected at the right time and supplemented with field observations, may lead to a quantitative assessment of the water balance of a large number of lakes located in a similar climatic setting.

The specific behaviour of lakes located in sub-polar regions, with their seasonal ice cover extending over several months, offers another opportunity for quantifying groundwater seepage. As shown in this study, observations of seasonal changes in the stable isotope composition of lake water, in particular during the ice-cover period, combined with a survey of groundwater isotope composition in the vicinity of the lakes studied, allow the quantification of groundwater fluxes to those lakes.

The G index of groundwater dependency of a lake proposed in this study, and defined as a percentage contribution of groundwater inflow to the total inflow of water to the given lake, appears to be a straightforward, quantitative measure of this dependency. Whereas in this study the G index was derived only for lakes without a visible surface inflow, it can also be applied in cases where an inflow of this kind is present and quantified by other means.

Acknowledgements. This work is funded by the 7th framework project GENESIS (226536), Renlund Foundation, Maa- ja vesitekniikan tuki r.y, and the Academy of Finland, AKVA-programme. KR's contribution was partly supported through the statutory funds of the AGH university of Science and Technology (project no. 11.11.220.01). We would like to thank Kirsti Korkka-Niemi and Anne Rautio from University of Helsinki for their assistance with the thermal imaging.

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Table 1. Mean isotope composition of selected lakes, streams and groundwater, sampled four times per year during the period 2010–2012.

	δ^2H (‰)	δ^{18}O (‰)	d-excess (‰)	Mean amplitude of δ^{18}O seasonal signal (‰)
Lakes:				
1. Ahveroinen 1	−72	−8.5	−4.0	1.2
2. Rokuanjärvi	−72	−8.7	−2.4	1.3
3. Jaakonjärvi	−64	−6.7	−10.4	0.8
4. Kolmonen 2	−67	−7.1	−10.2	0.9
5. Loukkojärvi	−63	−6.6	−10.2	0.9
6. Saarijärvi 2	−62	−6.8	−7.6	0.9
7. Saarinen	−72	−8.5	−4.0	3.3
8. Salminen	−71	−8.4	−3.8	0.5
9. Soppinen	−63	−6.8	−8.6	0.9
10. Tulijärvi	−86	−11.2	3.6	1.3
11. Vaulujärvi	−68	−7.7	−6.4	0.5
Streams:				
1. Heinäjoki	−93	−12.7	8.6	0.5
2. Hieto-oja	−89	−11.8	5.4	1.2
3. Kangasoja	−94	−12.9	9.2	0.7
4. Lianoja	−83	−10.7	2.6	2.2
5. Lohioja	−93	−12.9	10.2	0.8
6. Matokanava	−95	−13.1	9.8	0.9
7. Päiväkanava	−94	−13.0	10.0	0.9
8. Rokuanoja	−83	−11.0	5.0	3.1
9. Siirasoja	−95	−13.1	9.8	0.7
10. Soppisenoja	−90	−12.1	6.8	2.2
11. Valkiaisoja	−93	−12.8	9.4	1.0
Groundwater:				
1. MEA 106	−96	−13.3	10.4	0.6
2. MEA 206	−95	−13.1	9.8	0.8
3. MEA506	−93	−12.9	10.2	0.5
4. MEA 706	−93	−12.9	10.2	0.7
5. MEA 1106	−95	−13.1	9.8	0.8
6. MEA 1807	−93	−12.9	10.2	1.0
7. MEA 1907	−83	−10.7	2.6	1.0
8. MEA 2010	−97	−13.4	10.2	0.5
9. ROK 1	−96	−13.2	9.6	0.9
10. Siirasoja 1 harju	−95	−13.1	9.8	0.8
11. Siirasoja 1 rinne	−94	−13.1	10.8	1.0
12. Siirasoja 1 hiekka	−97	−13.4	10.2	0.5
13. Siirasoja 1 turve	−94	−13.	10.0	1.2

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Table 2. Characteristics and isotope data for 67 lakes sampled during the July–August 2013 field survey.

Lake No.	Name	Volume (10^3 m^3)	Surface area (ha)	Mean depth (m)	E^a (mm)	$\delta^{18}O$ (‰)	$\Delta\delta^{18}O$ (‰)	I_{TOT}/E	I^b_{GW} (m^3 day^{-1})	MTT (months)	G index[b] (%)
1	Heinälampi	1764	0.22	0.8	286	−11.0	3.1	5.85	n.d.	1.4	n.d.
2	Holma	23 778	1.89	1.3	257	−11.1	3.0	6.23	n.d.	2.4	n.d.
3	Koivujärvi 1	39 237	2.30	1.7	254	−11.4	2.7	7.02	n.d.	2.9	n.d.
4	Koivujärvi 2	76 884	2.93	2.6	251	−11.6	2.5	7.84	n.d.	4.0	n.d.
5	Pitkäjärvi 1	36 103	1.22	3.0	262	−6.5	7.6	1.34	19.7	25.4	42.5
6	Nurkkajärvi	269 134	3.93	6.8	247	−8.8	5.3	2.67	196.5	31.0	69.5
7	Luontolampi	19 995	0.49	4.1	275	−7.3	6.8	1.70	14.1	26.3	56.8
8	Ahveroinen 1	119 000	3.32	3.6	249	−8.3	5.8	2.33	137.3	18.5	65.3
9	Lianjärvi	113 678	15.13	0.8	231	−10.3	3.8	4.51	n.d.	2.2	n.d.
10	Syväjärvi 1	398 463	11.52	3.5	234	−11.2	2.9	6.38	n.d.	6.9	n.d.
11	Soppinen	81 630	6.04	1.4	242	−6.9	7.2	1.52	109.0	11.0	45.1
12	Salminen	681 860	25.31	2.7	225	−8.3	5.8	2.34	897.0	15.3	61.8
13	Saarinen	981 506	15.32	6.4	231	−7.8	6.3	2.01	437.9	41.3	56.6
14	Kivi-Ahveroinen	216 018	5.57	3.9	243	−9.0	5.1	2.91	305.5	16.5	71.4
15	Irvi-Ahveroinen	40 867	1.07	3.8	264	−7.5	6.6	1.80	31.6	24.2	57.5
16	Loukkojärvi	126 450	2.85	4.4	251	−6.8	7.3	1.45	50.2	36.5	44.5
17	Ylimmäinen	94 313	8.93	1.1	237	−9.7	4.4	3.58	629.8	3.7	76.3
18	Hietajärvi	223 118	7.30	3.1	240	−7.1	7.0	1.59	141.7	24.1	46.9
19	Saarijärvi 2	2 467 169	20.94	11.8	228	−6.7	7.4	1.43	279.1	108.9	37.8
20	Syväjärvi 2	1 863 446	31.95	5.8	223	−10.6	3.5	5.11	n.d.	15.4	n.d.
21	Pasko	2005	0.82	0.2	268	−10.6	3.5	5.08	n.d.	0.5	n.d.
22	Kuikkalampi	11 776	0.61	1.9	271	−7.2	6.9	1.66	16.5	12.8	55.1
23	Soppisenlampi	25 645	0.59	4.4	272	−7.3	6.8	1.72	16.9	28.1	56.8
24	Kirvesjärvi	654 618	13.47	4.9	233	−11.4	2.7	7.24	n.d.	8.7	n.d.
25	Tulijärvi Kirvesniemi	589 617	24.81	2.4	226	−10.5	3.6	4.80	n.d.	6.6	n.d.
26	Jaakonjärvi 2	5497	0.46	1.2	275	−7.4	6.7	1.76	14.2	7.4	58.3
27	Jaakonjärvi 3	13 594	0.66	2.1	271	−7.6	6.5	1.87	21.7	12.3	60.1
28	Maitolampi 2	44 812	2.04	2.2	256	−7.5	6.6	1.83	58.8	14.1	56.8
29	Kotalampi	25 980	2.60	1.0	253	−11.2	2.9	6.54	409.3	1.8	87.8
30	Rokuanjärvi 1	4 364 781	164.59	2.7	205	−8.5	5.6	2.50	5555.9	15.5	60.6
31	Tervatienlampi	21 970	0.63	3.5	271	−7.9	6.2	2.06	24.4	18.7	63.9
32	Valkiaislampi	32 073	0.70	4.6	270	−8.8	5.3	2.72	40.5	18.7	72.4
33	Ankkalampi	103 700	3.86	2.7	248	−8.6	5.5	2.52	177.0	12.9	67.6
34	Saarilampi 1	12 044	0.45	2.7	276	−10.5	3.6	4.86	55.9	6.0	84.9
35	Kiiskeroinen	6517	0.63	1.0	271	−12.1	2.0	10.43	180.7	1.1	92.9
36	Jaakonjärvi 1	93 614	3.53	2.7	249	−6.6	7.5	1.36	52.6	23.5	40.4
37	Vaulujärvi	432 813	8.97	4.8	237	−7.6	6.5	1.85	231.8	32.9	54.1
38	Levä-Soppinen	34 000	2.31	1.5	254	−12.3	1.8	11.30	669.5	1.5	93.0
39	Anttilanjärvi	60 230	1.06	5.7	264	−7.4	6.7	1.76	30.3	36.8	56.6
40	Hautajärvi 1	189 222	2.56	7.4	253	−7.7	6.4	1.93	79.7	45.4	58.7

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Table 2. Continued.

Lake No.	Name	Volume (10^3 m^3)	Surface area (ha)	Mean depth (m)	E^a (mm)	$\delta^{18}O$ (‰)	$\Delta\delta^{18}O$ (‰)	I_{TOT}/E	I^b_{GW} (m^3 day^{-1})	MTT (months)	G index[b] (%)
41	Lepikonjärvi	191 118	2.98	6.4	251	−7.8	6.3	1.97	94.6	39.0	59.2
42	Kolmonen 1	35 700	0.68	5.3	270	−7.1	7.0	1.60	17.0	36.6	53.4
43	Kolmonen 2	21 890	0.56	3.9	273	−6.8	7.3	1.48	12.2	29.2	49.9
44	Kolmonen 3	21 630	0.54	4.0	273	−7.3	6.8	1.69	15.1	26.3	56.2
45	Hätäjärvi	35 217	1.73	2.0	258	−6.5	7.6	1.34	26.8	17.7	41.4
46	Kissalampi	800	0.36	0.2	279	−5.6	8.5	1.00	3.0	2.4	27.8
47	Valkiajärvi	582 662	8.36	7.0	238	−8.5	5.6	2.46	348.7	35.7	65.5
48	Hautajärvi 2	445 413	14.60	3.1	232	−7.4	6.7	1.78	332.8	22.2	50.9
49	Keskimmäinen	428 515	13.07	3.3	233	−10.8	3.3	5.36	n.d.	7.9	n.d.
50	Siirasjärvi 2	23 126	0.56	4.1	273	−12.7	1.4	14.85	233.8	3.1	95.0
51	Siirasjärvi 1	9701	0.34	2.9	280	−10.4	3.7	4.73	41.4	6.5	84.7
52	Telkkälampi	7284	0.25	2.9	284	−7.3	6.8	1.71	7.7	17.9	58.3
53	Maitolampi 1	48 892	1.56	3.1	259	−6.4	7.7	1.28	21.9	28.4	39.1
54	Taka-Salminen	364 757	7.31	5.0	240	−9.6	4.5	3.55	n.d.	17.6	n.d.
55	Etu-Salminen	196 453	5.83	3.4	243	−8.1	6.0	2.15	n.d.	19.4	n.d.
56	Pikku-Salminen	39 856	1.32	3.0	261	−7.6	6.5	1.88	n.d.	18.5	n.d.
57	Kylmäjärvi	394 443	7.89	5.0	239	−8.5	5.6	2.44	327.5	25.7	65.4
58	Kourujärvi 1	68 023	1.83	3.7	257	−8.5	5.6	2.50	87.6	17.4	68.6
59	Kourujärvi 2	31 136	1.91	1.6	256	−7.6	6.5	1.90	59.1	10.0	58.5
60	Huttunen	54 588	1.49	3.7	260	−7.5	6.6	1.82	43.8	23.2	57.2
61	Saarijärvi 1	13 590	1.36	1.0	261	−6.5	7.6	1.33	21.5	8.6	41.8
62	Pitkäjärvi 2	451 093	7.93	5.7	239	−8.1	6.0	2.15	268.2	33.3	60.6
63	Pyöräinen	76 077	3.76	2.0	248	−7.9	6.2	2.06	n.d.	11.9	n.d.
64	Likainen	33 763	8.28	0.4	238	−10.9	3.2	5.78	n.d.	0.9	n.d.
65	Nimisjärvi	1 840 396	167.53	1.1	205	−8.7	5.4	2.59	n.d.	6.2	n.d.
66	Ahveroinen 2	670 515	16.30	4.1	230	−8.0	6.1	2.09	n.d.	25.7	n.d.
67	Tervalampi	29 707	0.79	3.7	268	−8.0	6.1	2.11	31.3	19.9	64.3

[a] Calculated for the period 1 June–31 August 2013.
[b] Calculated for lakes with no visible surface inflow.
n.d. – not determined.

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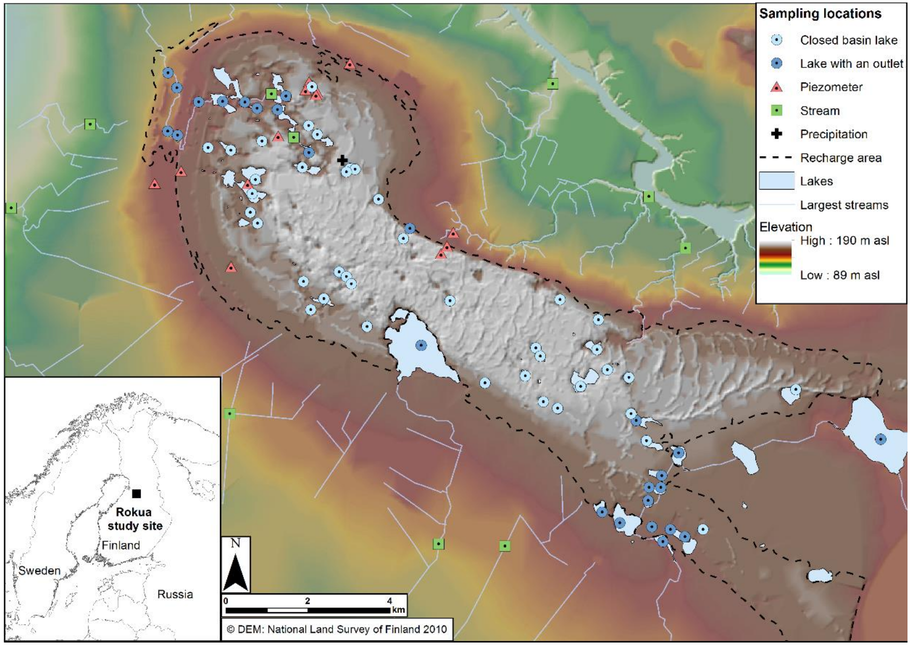

Figure 1. The study site of Rokua esker aquifer area. Digital elevation model by the National Land Survey of Finland (2010b).

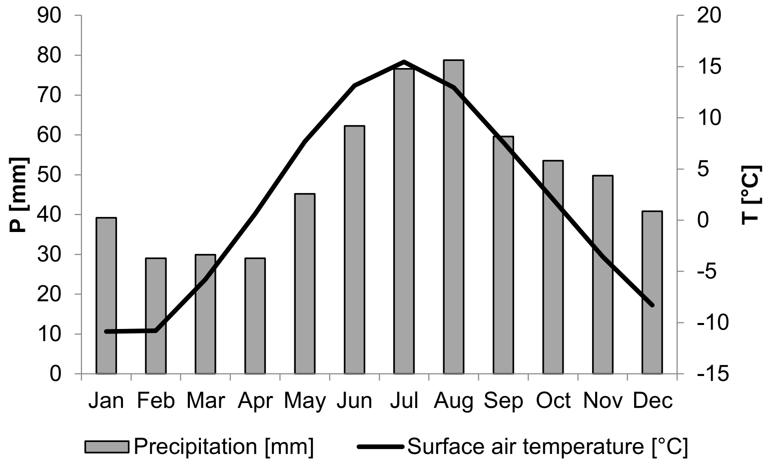

Figure 2. The long-term (1959–2013) monthly mean values of surface air temperature and the amount of precipitation recorded at the station located 10 km south-west of the study site (Finnish Meteorological Institute, 2014).

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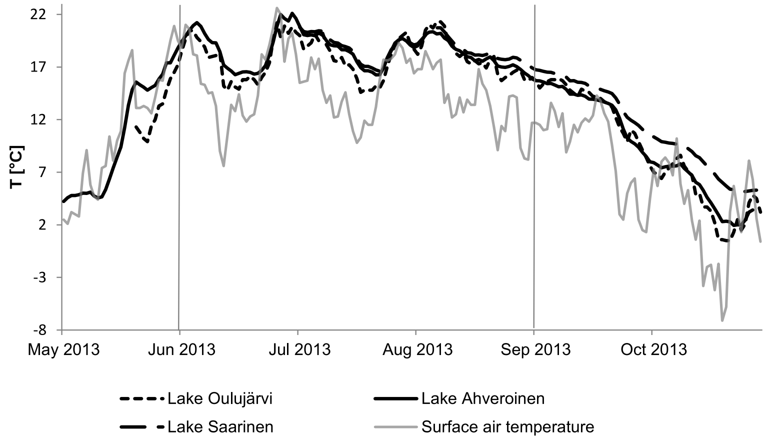

Figure 3. Daily mean surface water temperatures of lakes Oulujärvi (92 800 ha) (Finnish Environmental Institute, 2013), Ahveroinen 1 (3.3 ha) and Saarinen (15.3 ha) during the summer of 2013, compared with the surface air temperature data for the same period. Lake Oulujärvi is located in the east, next to the study site, 1 km from the easternmost lake studied. Vertical lines mark the period used in the calculations of evaporation and isotope mass balance.

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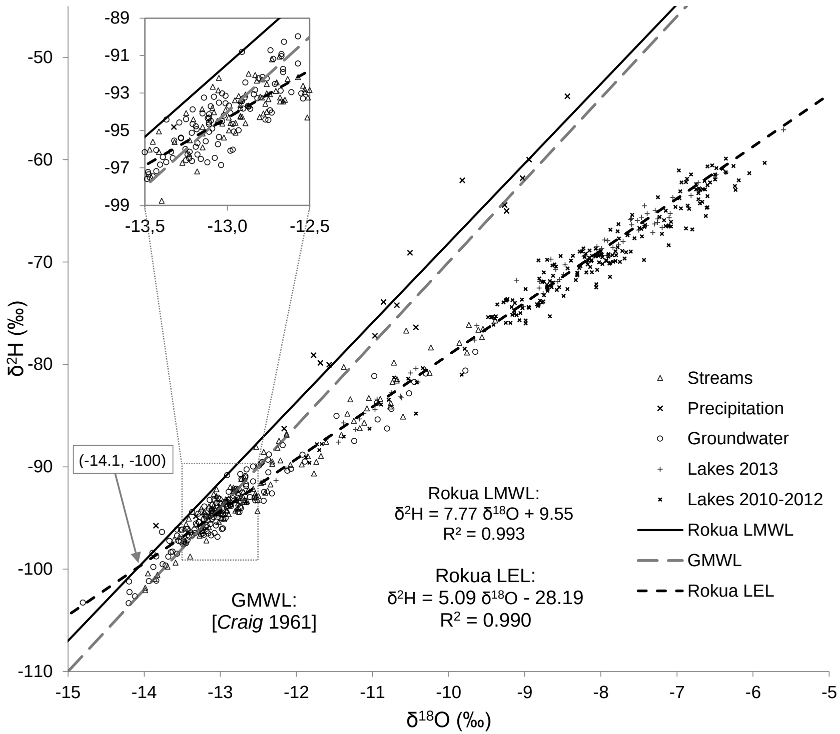

Figure 4. δ^2H–δ^{18}O relationship for different appearances of surface water (lakes, streams) and groundwater in the study area, investigated within the scope of this study. The Rokua evaporation line (local evaporation line – LEL) was defined as the best fit line of the data points representing lakes sampled during the July–August 2013 campaign.

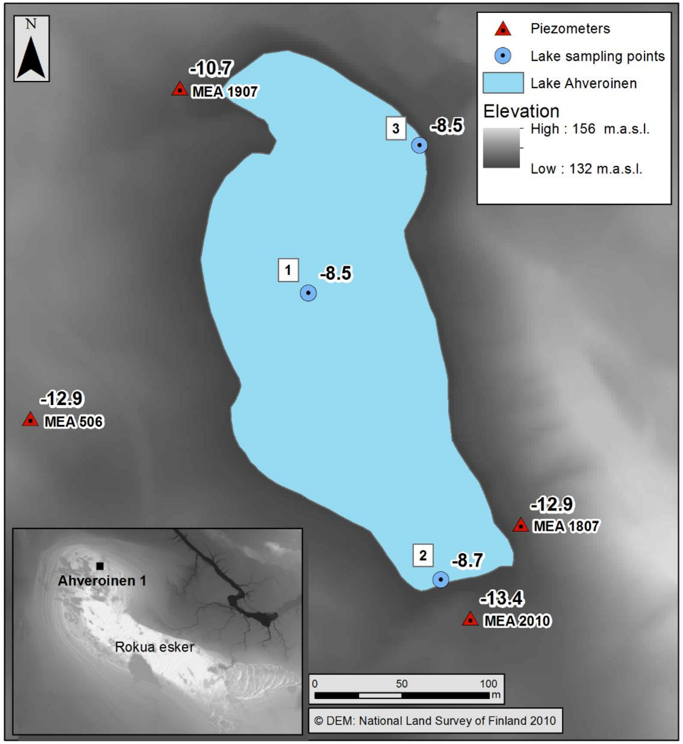

Figure 5. Mean δ^{18}O values (‰) of lake Ahveroinen and adjacent groundwater. The mean δ^{18}O value for site 1 is 8.5 ‰ at both sampled depths (1 and 4 m).

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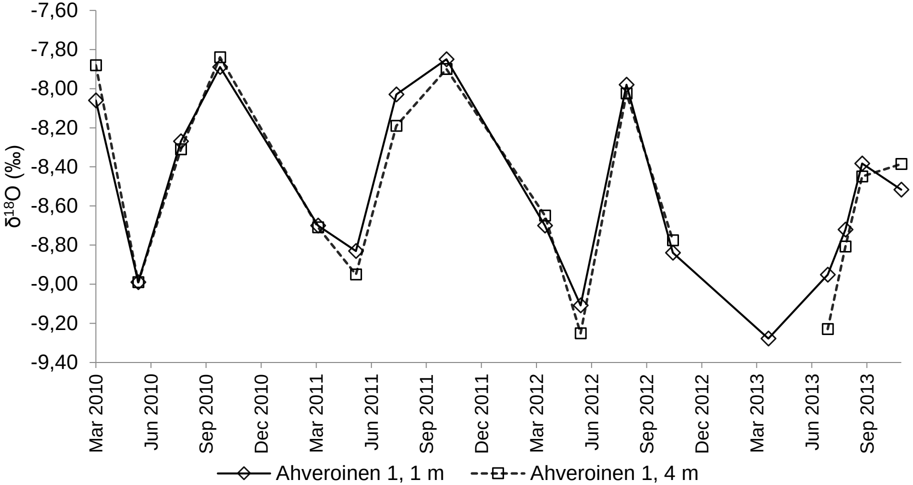

Figure 6. Seasonal variations of δ^{18}O in lake Ahveroinen 1, observed at 1 and 4 m depths. Maximum depth of the lake is 4.8 m.

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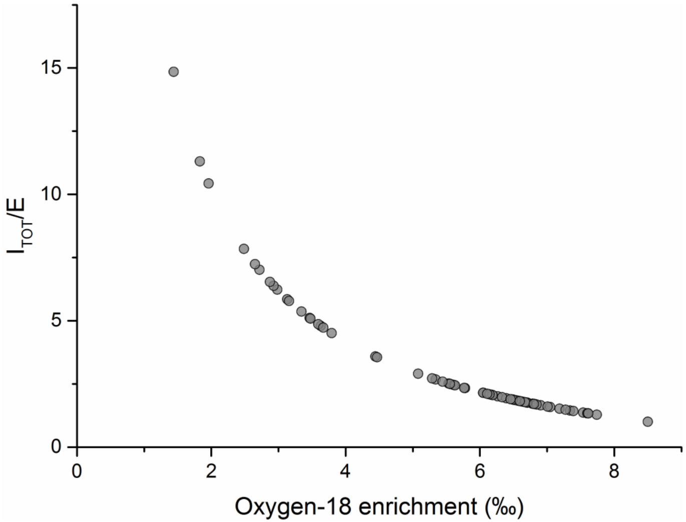

Figure 7. The ratio of total inflow to evaporation (I_{TOT}/E) as a function of the measured ^{18}O isotope enrichment ($\Delta\delta^{18}O$) for 67 Rokua lakes surveyed during the July–August 2013 sampling campaign. The calculations were performed for normalised relative humidity derived from terminal lake expression (see text for details).

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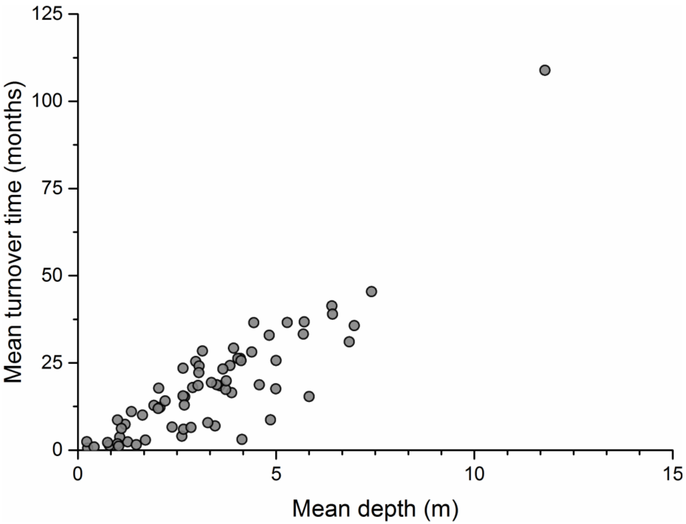

Figure 8. Mean turnover time of water (MTT) in the Rokua lakes surveyed in July–August 2013, as a function of mean water depth.

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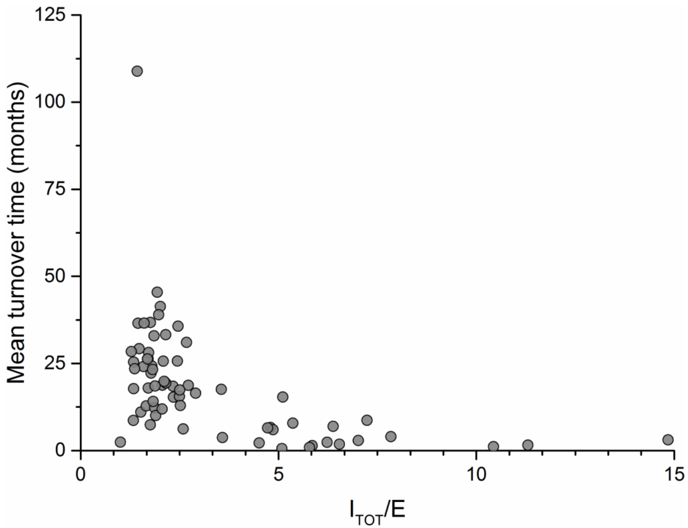

Figure 9. Mean turnover time of water (MTT) in the Rokua lakes surveyed in July–August 2013, plotted as a function of the I_{TOT}/E ratio for those lakes.

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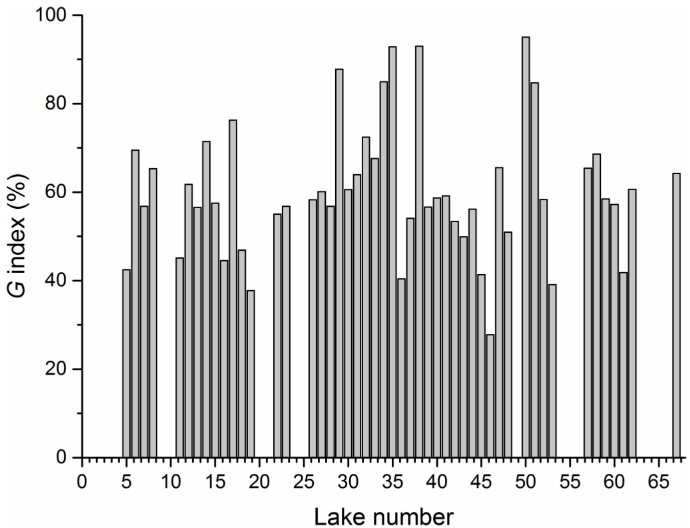

Figure 10. The *G* index quantifying the groundwater dependency of lakes in the Rokua study area. The index is defined as the percentage contribution of groundwater inflow to the total inflow of water to the given lake.

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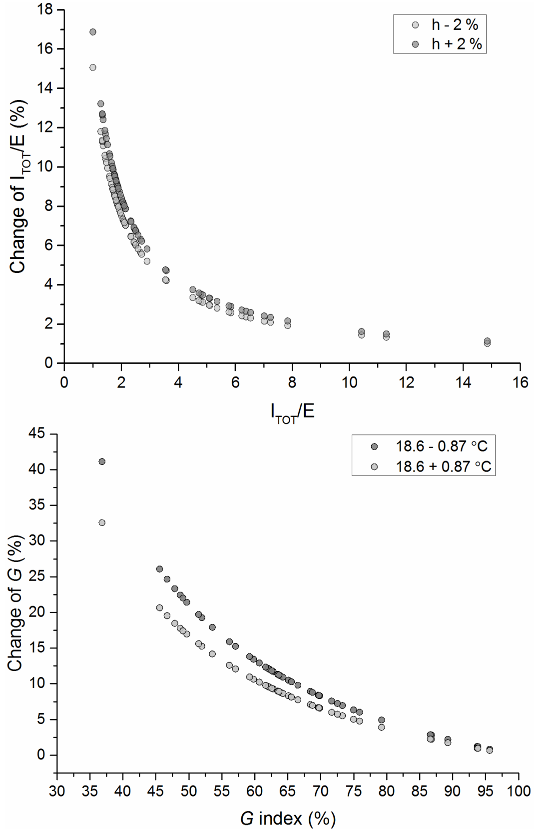

Figure 11. Top panel: change of the total inflow-to-evaporation ratio (in %) as a function of I_{TOT}/E value, for two different values of relative humidity: increased and decreased by 2 % with respect to the value assumed in the calculations (64.6 %). Bottom panel: change in the G index (in %), as a function of the value of this parameter, for two different temperatures: increased and decreased by 0.87 °C with respect to the temperature assumed in the calculations (18.6 °C). See text for details.

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