

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

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12

13 **Abstract**

14 A stable isotope study of 67 kettle lakes and ponds situated on an esker aquifer (90 km²) in northern
15 Finland was carried out to determine the role and extent of groundwater inflow in groundwater-
16 dependent lakes. Distinct seasonal fluctuations in the $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of lakes are the result of
17 seasonal ice cover prohibiting evaporation during the winter. An iterative isotope mass balance
18 approach was used to calculate the inflow-to-evaporation ratios (I_{TOT}/E) of all 67 lakes during the
19 summer of 2013 when the isotopic compositions of the lakes were approaching a steady-state. The
20 balance calculations were carried out independently for ²H and ¹⁸O data. Since evaporation rates were
21 derived independently of any mass balance considerations, it was possible to determine the total
22 inflow (I_{TOT}) and mean turnover time (MTT) of the lakes. Furthermore, the groundwater seepage rates
23 to all studied lakes were calculated. A quantitative measure was introduced for the dependence of a
24 lake on groundwater (G index) that is defined as the percentage contribution of groundwater inflow
25 to the total inflow of water to the given lake. The G index values of the lakes studied ranged from ca.
26 40 % to 98 %, revealing generally large groundwater dependency among the studied lakes. This study
27 shows the effectiveness of applying an isotope mass balance approach to quantify the groundwater
28 reliance of lakes situated in a relatively small area with similar climatic conditions.

29

30 **Key words**

31 Stable isotopes of water, groundwater-surface water interaction, lake hydrology, mean turnover time
32 (MTT), inflow-to-evaporation ratio (I_{TOT}/E), thermal imaging

33 1. Introduction

34 The characterisation of groundwater dependent ecosystems (GDEs) is a requirement of the
35 Groundwater Directive (EC, 2006). These systems are often complex and their hydrology and contact
36 with aquifers are not well established. Lakes can be dependent on groundwater directly or indirectly,
37 and this dependence can vary over time (Kløve et al., 2011). Understanding groundwater and lake
38 water interaction is important not only for water resource management (Showstack, 2004), but also
39 for understanding the ecology and eutrophication of lakes, since groundwater may be a key element
40 in the lake nutrient balance (Ala-aho et al., 2013; Belanger et al., 1985; Brock et al., 1982; Kidmose
41 et al., 2013). Furthermore, the vulnerability of lakes to pollution can be controlled by their
42 dependency on groundwater (Kløve et al., 2011). Methods such as seepage meters (Ala-aho et al.,
43 2013; Rosenberry et al., 2008, 2013), environmental tracers (e.g. Dinçer, 1968; Shaw et al., 2013;
44 Stets et al., 2010; Yehdegho et al., 1997; Zuber, 1983) and numerical modelling (e.g. Krabbenhoft et
45 al., 1990; Stichler et al., 2008; Winter and Carr, 1980) can be used to determine the groundwater
46 reliance of lakes.

47 Heavy stable isotopes of water (^{18}O , ^2H) can be considered as ideal tracers for studying the
48 hydrological cycle (e.g. Clark and Fritz, 1997). Fractionation of isotopes of water is the very factor
49 enabling their use in hydrological studies, as it governs the changes in isotopic abundances within the
50 water cycle (Gat, 2010). At a global scale, the ^2H and ^{18}O isotope composition of meteoric waters
51 cluster along the line called the global meteoric water line (GMWL), first determined by Craig (1961):
52 $\delta^2\text{H} = 8 \cdot \delta^{18}\text{O} + 10$. Locally, this linear relationship may have a slightly different form (local
53 meteoric water line - LMWL). Evaporation from an open water body fractionates isotopes so that the
54 remaining liquid phase is enriched in both ^2H and ^{18}O in proportion with their effective fractionation
55 factors accompanying this process. Consequently, the isotopic composition of the evaporating water
56 body evolves in the δ -space along the line known as the local evaporation line (LEL), whose slope is
57 significantly smaller than that characterising the local or global meteoric water lines. The position of
58 the isotopic composition of lake water along this line is strongly related to the water balance of the
59 lake (e.g. Gat, 1996; Gibson and Edwards, 2002; Rozanski et al., 2001).

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

64 2002; Gibson et al., 1993; Jonsson et al., 2009; Turner et al., 2010; Yi et al., 2008), mostly in Canada
65 and northern Sweden, these studies were generally focused on lakes spread over large areas.

66 The central aim of this study was to quantify the groundwater dependence of 67 kettle lakes and ponds
67 situated across a relatively small area (90 km²) of the Rokua esker aquifer occupying a large
68 glaciofluvial deposit in northern Finland. To quantify the extent of the interaction between the aquifer
69 and the lakes, a dedicated isotope study was launched in 2013. This was part of comprehensive
70 investigations (2010-2012) aimed at understanding the hydrology of an esker aquifer area where some
71 of the kettle lakes and ponds have suffered from water level decline or eutrophication. Since the
72 seasonal isotopic behaviour of the selected lakes in the study area was already fairly well understood
73 based on the data collected from 2010 to 2012, it was decided to conduct a large-scale one-time survey
74 of the isotopic composition of all 67 lakes on the Rokua esker in order to quantify their dependence
75 on groundwater. Ala-aho et al. (2013), who studied 11 lakes on the esker, showed that the water levels
76 in closed-basin seepage lakes have more fluctuations than the drainage lakes, which have more stable
77 water levels. On the other hand, the drainage lakes are more productive. Their study also showed that
78 subsurface flow can transport phosphate to lakes. Therefore, it was important to quantify the
79 groundwater dependence of all lakes on the esker and propose an appropriate index reflecting this
80 dependence. The large-scale field campaign conducted in July and August 2013 comprised the
81 sampling of water in all 67 lakes for isotope analyses, combined with continuous temperature
82 measurements and aerial thermal imaging of the lakes.

83

84 **2. The study area**

85 The Rokua esker aquifer area, situated in Northern Finland, was formed during the last deglaciation
86 period 9,000 to 12,000 years ago (Tikkanen, 2002). As ice retreated, a long ridge formation,
87 consisting mainly of fine and medium sand (Pajunen, 1995), was shaped. Ancient sea banks
88 surrounding the esker show that the esker was originally an island that gradually rose from the sea
89 (Aartolahti, 1973). Today, the highest elevation of the esker is 100 m above the surrounding low-
90 lying peatlands and the layer thickness of sand ranges from 30 m to more than 100 m above the
91 bedrock. Sea banks, dunes and kettle holes form a rolling and geologically unique terrain (Aartolahti,
92 1973). Kettle holes were formed when ice blocks were buried in the ground and, as they melted, left
93 depressions in the landscape. The ground surface of the esker is mainly lichen-covered pine forests.
94 Hydrologically, the Rokua esker is an unconfined aquifer, one of the largest in Finland, and it has

95 two regional groundwater mounds (Rossi et al., 2014). The recharge area of the aquifer is 90 km² and
96 the discharge zones are situated in the surrounding peatlands, which partially confine the aquifer
97 (Rossi et al., 2012).

98 Kettle holes – long and narrow depressions – give Rokua esker its distinct character. The sizes of
99 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
100 (Aartolahti, 1973). Most of the kettle holes are now dry, but due to the influence of groundwater in
101 the past, peat has accumulated at the bottom of them, creating kettle hole mires (Pajunen, 1995).
102 However, the alternating topography of the area is reflected in the existence of approximately 90
103 lakes or ponds, referred to as kettle lakes or ponds. Peat started to accumulate in the border regions
104 of the lakes more than 8,000 years ago, so most of the kettle lakes and ponds are partly paludified
105 (Pajunen, 1995). Nevertheless, the majority of the lakes and ponds are characterised by their crystal
106 clear water, which attract people; number of holiday homes and hotels are located on the lakeshores.
107 The lakes are nowadays widely used for different recreational activities, such as swimming, fishing
108 and scuba diving (Anttila and Heikkinen, 2007). The uniqueness of the glaciofluvial formation of
109 Rokua, in which the actions of ice, water and wind can be seen, has been recognised in many ways.
110 Some of the Rokua esker is protected by Natura 2000 and by the Finnish nature reserve network.
111 Rokua was recently chosen to be part of the UNESCO GeoPark Network and is currently the
112 northernmost region in this network.

113

114 **3. Materials and methods**

115 **3.1. Hydrological measurements and thermal imaging**

116 During 2010-2012, 11 lakes, 13 piezometers and 11 streams were sampled in the study area four times
117 per year to analyse the stable isotopic composition of water, nutrients, water quality parameters (T,
118 pH, E.C., O₂) and geochemical parameters (silica, major cations and anions) (Fig. 1). During the field
119 campaign conducted in July and August 2013, a total of 67 lakes and ponds were surveyed for the
120 same parameters and thermal images of lakes taken from the air in a helicopter using a FLIR thermal
121 camera. In addition, composite precipitation samples were mainly collected once a month during the
122 open water season at the station on the esker during 2010-2013. Precipitation samples for winter
123 were collected once a year before snowmelt by taking a uniform sample of the whole snowpack depth.
124 Water quality parameters were analysed in the field using WTW Multi 3430 or Multi 350i meters for
125 temperature, oxygen, EC and pH. Samples of lake water were collected with a Limnos sampler,

126 approximately 1 m below the water surface and 1 m above the bottom of the lake. If the depth of the
127 lake was less than 2 m, only one sample from the depth of 1 m was taken and if it was more than 20
128 m, samples were taken from the middle of the water profile as well. Depending on their shape and
129 size, the lakes had between 1 and 4 sampling locations. Stream samples were collected by submerging
130 a bottle in water, facing upstream. Piezometers were pumped for at least 10 minutes prior to taking
131 groundwater samples or until the colour of the water was clear. The samples were collected one metre
132 below the water table. All sampling bottles (HDPE) were rinsed with the sampled water prior to
133 filling. Samples for isotope analyses were stored in the dark at a reduced temperature ($4\text{ }^{\circ}\text{C} \pm 2\text{ }^{\circ}\text{C}$).

134 The isotopic composition of water samples was analysed using CRDS technology with a Picarro
135 L2120-i analyser. Samples with visible colour or suspended matter were filtered (pore size $25\text{ }\mu\text{m}$)
136 prior to analysis. The measured $^2\text{H}/^1\text{H}$ and $^{18}\text{O}/^{16}\text{O}$ isotope ratios are reported as relative deviations
137 from the VSMOW standard. Typical uncertainty of the reported $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values are $\pm 0.1\text{ }‰$ and
138 $\pm 1.0\text{ }‰$ respectively.

139 Lake water temperature was measured continuously during the ice-free period in 2013 for two lakes
140 on the Rokua esker: Ahveroinen (3.3 ha) and Saarinen (15.3 ha). Hobo loggers (pendant temperature
141 data logger UA-001-08 and conductivity logger U24-001, accuracy $0.1\text{ }^{\circ}\text{C}$) were installed 50 cm
142 below the lake surface. In addition, the surface water temperature for lake Oulujärvi (92800 ha) was
143 obtained from the database of the Finnish Environmental Institute (2013). Lake Oulujärvi is located
144 east, next to the study site, 1 km from the easternmost lake studied. Thermal imaging of the lakes was
145 conducted on 5 August 2013 using a Flir Thermacam P-60 thermal camera. This camera had $320 \times$
146 240 pixel sensor resolution and an opening of 24 ° . It covered the electromagnetic spectrum from 7.5
147 to $13\text{ }\mu\text{m}$. The imaging was taken by helicopter 150 meters above the lakes. The image data were
148 correlated to the predominant weather conditions (temperature and relative humidity) with data from
149 the FMI Pelso weather station measured every 10 minutes.

150 Depth profiling was undertaken in the lakes for which no depth contour lines are available (National
151 Land Survey of Finland, 2010a). It was carried out with a portable depth-sounding radar (resolution
152 0.1 m) or with a measuring cable and GPS system. Typically, two profiles from the shore to the
153 deepest point were defined. The number of measurement points differed between the lakes depending
154 on their size. In total, 52 lakes were surveyed.

155

156

157 3.2. Lake volumes

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

166

167 3.3. Evaporation from lakes

168 Evaporation (E) was calculated individually for all the lakes surveyed using a mass transfer approach
169 (Rosenberry et al., 2007; Dingman, 2008). This method was chosen because it yields instantaneous
170 rates of evaporation and takes into account lake sizes (Harbeck, 1962). Our data enabled the
171 calculations of daily mean values of evaporation flux for all the lakes studied. Averaging the
172 parameters for periods of time longer than one day can lead to significant biases in the calculated E
173 values since on these time scales the vapour pressure differences and wind speed may correlate
174 (Jobson, 1972). The following expression was used to calculate the evaporation flux (Dingman,
175 2008):

$$176 E = K_E \cdot v_a \cdot (e_s - e_a), \quad (1)$$

177 where:

178 E is evaporation rate (mm s^{-1})

179 K_E is the mass-transfer coefficient in $\text{m km}^{-1} \text{kPa}^{-1}$ describing the impact of turbulent eddies of the
180 wind on vertical transport of water vapour from the lake with area A_L (km^2), $K_E = 1.69 \cdot 10^{-5} \cdot$
181 $A_L^{-0.05}$ (Harbeck, 1962)

182 v_a is wind speed (m s^{-1}) at 2 m height

183 e_s is the saturation vapour pressure in kPa at surface water temperature T_s ($^{\circ}\text{C}$), $e_s = 0.611 \cdot$
184 $\exp\left(\frac{17.3 \cdot T_s}{T_s + 237.3}\right)$

185 e_a is the vapour pressure in the air in kPa, $e_a = h \cdot 0.611 \cdot \exp\left(\frac{17.3 \cdot T_a}{T_a + 237.3}\right)$, where h is relative humidity
186 and T_a is air temperature ($^{\circ}\text{C}$).

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

$$189 \quad v_z = v_r \left(\frac{z}{z_r}\right)^\beta, \quad (2)$$

190 where v_r is the measured wind speed at the reference height z_r (10 m), z is the height for which speed
191 is adjusted (2 m) and β is the friction coefficient. The value of 0.15 for β , characteristic for grassland,
192 was employed in our study (0.1 characterizes oceans and lakes) (Bañuelos-Ruedas et al., 2010) since
193 the lakes are surrounded by forests lowering the wind speed.

194 The meteorological parameters necessary for the calculations (relative humidity, wind speed and air
195 temperature) were obtained from the meteorological station 5502 (Vaala-Pelso) of the Finnish
196 Meteorological Institute (2014), located approximately 10 km from the site. A probable range for the
197 lakes' surface water temperature was evaluated using continuous temperature measurements at 50 cm
198 depth from one of the studied lakes, Ahveroinen (see section 4.1), and a standard deviation of this
199 temperature determined from thermal images. Using the derived temperature range, a probable range
200 for evaporation rates from all the lakes was calculated. The adopted method relies only on temperature
201 difference measured on one day, but since all the lakes are in a relatively small area with almost
202 identical weather conditions, it is highly probable that the seasonal behaviour of the surface water
203 temperature of the studied lakes is similar.

204

205 **3.4. Isotope mass balance**

206 Instantaneous water and isotope balances for an evaporating surface water body can be formulated as
207 follows:

$$208 \quad \frac{dV}{dt} = I_{TOT} - E - O_{TOT} \quad (3)$$

$$209 \quad \delta_L \frac{dV}{dt} + V \frac{d\delta_L}{dt} = \delta_{IT} \cdot I_{TOT} - \delta_E \cdot E - \delta_{OT} \cdot O_{TOT} \quad (4)$$

210 where V stands for the volume of the surface water body, δ_L signifies its isotopic composition and
211 I_{TOT} , E and O_{TOT} represent the total inflow, evaporation and total outflow of water from the system,
212 respectively, whereas δ_{IT} , δ_E and δ_{OT} stand for their respective isotopic compositions, expressed in
213 ‰. As the total inflow may consist of several components (precipitation, underground and surface
214 inflows), each with its specific isotopic composition, δ_{IT} should be calculated as a flux-weighted mean
215 of the respective isotopic compositions of individual components. The total outflow may also consist

216 of surface and underground components. For well-mixed systems it is typically assumed that $\delta_{OT} =$
 217 δ_L .

218 The isotopic composition of the evaporation flux, δ_E , cannot be measured directly. However, it can
 219 be calculated using the expression derived from the linear resistance model describing isotope effects
 220 accompanying evaporation process (Craig and Gordon, 1965; Horita et al., 2008):

$$221 \quad \delta_E = \frac{(\delta_L/\alpha_{LV}) - h_N \cdot \delta_A - \varepsilon}{(1 - h_N) + \Delta\varepsilon \cdot 10^{-3}} \quad (5)$$

222 where ε is the total effective isotope fractionation, $\varepsilon = \varepsilon^* + \Delta\varepsilon$, where ε^* stands for equilibrium isotope
 223 enrichment ($\varepsilon^* = (1 - 1/\alpha_{LV}) \cdot 10^3$) expressed in ‰ and α_{LV} represents the equilibrium isotope
 224 fractionation factor between liquid and gaseous phases. The kinetic isotope enrichment $\Delta\varepsilon$, is defined
 225 as $\Delta\varepsilon = C_k(1 - h_N)$ where C_k stands for the kinetic enrichment parameter. δ_A represents the isotopic
 226 composition of atmospheric water vapour over the evaporating water body (‰) and h_N is the relative
 227 humidity of the local atmosphere, normalized to the temperature of evaporating water.

228 When the evaporating water body is in hydrologic and isotopic steady-state ($dV/dt=0$ and $d\delta_L/dt=0$,
 229 respectively), the following approximate expression describing the isotope enrichment of the
 230 evaporating water body can be derived from eqs. 3-5 (see e.g. Gat and Bowser, 1991):

$$231 \quad \Delta\delta = \delta_{LS} - \delta_{IT} \cong \frac{\delta_A - \delta_{IT} + \varepsilon/h_N}{1 + \frac{I_{TOT} \cdot 1 - h_N}{E \cdot h_N}} \quad (6)$$

232 where $\Delta\delta$ stands for the evaporative enrichment and δ_{LS} is the steady-state isotopic composition of
 233 the studied system. The following expression describing the ratio of the total inflow to the evaporation
 234 rate can be derived from eq. (6):

$$235 \quad \frac{I_{TOT}}{E} = \frac{\delta_A - \delta_{LS} + \varepsilon/h_N}{(\delta_{LS} - \delta_{IT}) \frac{1 - h_N}{h_N}} \quad (7)$$

236 If the evaporation flux E can be assessed independently, as is the case of this study, the total inflow
 237 of water to the given lake can be calculated from eq. (7). From this, the groundwater component of
 238 this inflow can be further inferred, provided that the other components of this inflow are known or
 239 can be independently assessed.

240 To quantify the water balance of a lake with the aid of eq. (7) the knowledge of the isotopic
 241 composition of atmospheric moisture interacting with the lake, δ_A , is required. Field measurements
 242 of this parameter become feasible only recently thanks to advancements of CRDS technology.

243 However, the combined studies of the isotopic composition of atmospheric water vapour and
244 precipitation performed in moderate climates (Jacob and Sonntag, 1991; Schoch-Fischer et al., 1984)
245 have shown that, on monthly basis, the isotopic composition of precipitation is generally in isotopic
246 equilibrium with the local atmospheric moisture at ground-level temperature. This is particularly true
247 for summer season. Therefore, the value of δ_A can be derived from the isotopic composition of local
248 precipitation, δ_P , which is also required to quantify the isotopic composition of the total inflow to the
249 studied lake, δ_T , appearing in eq. (7). The following relation can be used to calculate δ_A (in ‰):

$$250 \quad \delta_A = \frac{1}{\alpha_{LV}} (\delta_P + 10^3) - 10^3 \quad (8)$$

251 Equation (7) is valid under two basic assumptions: (i) the evaporating water body is in hydrologic
252 and isotopic steady-state, and (ii) the water body is isotopically homogeneous. Natural surface water
253 systems, such as lakes, typically operate close to their hydrologic and isotopic steady-states attained
254 in the course of their long history. Their steady-state characteristics are defined by local climate,
255 morphological setting and prevailing hydrological regime. Such systems usually exhibit seasonal
256 fluctuations of varying amplitude, caused by seasonal fluctuations of local climate (surface air
257 temperature, relative humidity, precipitation amount), superimposed on long-term trends. Seasonal
258 ice-cover, typical for mid and high latitudes, may also contribute to the seasonal fluctuations of the
259 steady-state characteristics of such systems. The gradual attainment of the isotopic steady-state,
260 which is characterized by an exponential function describing the temporal evolution of δ_L , can be
261 observed only for surface water systems which are artificially created, such as dredging lakes
262 resulting from exploitation of gravel deposits (Zimmerman, 1979). The time constant characterizing
263 the dynamics of this process is mainly controlled by the mean turnover time of water in this system,
264 defined as the ratio of its volume to the total inflow, its hydrological balance (I_{TOT}/E ratio) and the
265 normalized relative humidity (e.g. Gonfiantini, 1986; Zimmerman, 1979).

266

267 **4. Results and discussion**

268 **4.1. Climate and lake water temperature data**

269 The climate is strongly seasonal in the Rokua esker study area. The long-term monthly mean values
270 of surface air temperature vary from -10.9 °C (January) to 13.2 °C (June) for the period between 1959
271 and 2013 (Fig. 2). The amount of monthly precipitation varies from 29 mm (February and April) to
272 79 mm (August) for the same period. The warmest months of the year are June, July and August. The

273 long-term (1970-2013) monthly mean relative humidity of air varies from 61 % (May) to 91 %
274 (November).

275 The seasonal temperature patterns of the monitored lakes were very similar, despite significant
276 differences in lake size (Fig. 3). Thus the surface water temperature of lake Ahveroinen 1 (mean value
277 of 19.1 °C) during the period from 1 June 2013 to 31 August 2013 was used as a basis for estimating
278 the water temperature of other lakes. Thermal images collected on 5 August 2013 yielded comparable
279 surface water temperatures that ranged from 19.5 °C to 24.6 °C, with a mean of 21.3 °C and a standard
280 deviation of 0.87 °C. Combining the results of continuous temperature measurements and thermal
281 images, an estimate of the mean surface water temperature of all lakes was derived to be 19.10 °C ±
282 0.87 °C. This temperature was used in the isotope mass balance calculations.

283

284 **4.2. Local isotopic compositions**

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

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

300 Although majority of groundwater samples cluster near the LMWL-LEL intersect, there are some
301 data points lying along the LEL in the $\delta^{18}\text{O}$ - $\delta^2\text{H}$ plot indicate the contribution of (evaporated) lake
302 water to groundwater. However, the interconnection between the lakes via groundwater is likely to
303 be minor since lakes probably have groundwater table maxima between them as they are situated in

304 deep holes in the landscape. Nevertheless, the existence of deeper flow paths from the upper elevation
305 lakes to the lower ones cannot be excluded. Based on the estimation of elevation differences and
306 distances between the lakes, we identified the lakes which do not have surface inflows but which
307 might receive some groundwater input originating from upper elevation lakes. These are lakes No. 6,
308 14, 20, 34, 38, 41, 47, 50 and 59. The sensitivity study presented in section 4.5 considers probable
309 changes of the isotopic composition of the total inflow to each lake caused by the presence of
310 evaporated lake water component.

311 The influence of evaporated lake water seeping to groundwater is illustrated in Fig. 5, showing the
312 isotopic composition of lake Ahveroinen 1 and adjacent groundwater. The mean $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values
313 of piezometers MEA 2010 and MEA 1907 situated on the south-eastern and north-western sides of
314 the lake were -13.4‰ and -97‰ and -10.7‰ and -83‰ respectively, clearly indicating a
315 substantial (ca. 55 %) contribution of lake water to groundwater at the north-western side of the lake.
316 A smaller contribution (ca. 10 %) of lake water to groundwater can be seen on the eastern and western
317 sides of the lake where the mean $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of groundwater were -12.9‰ and -93‰
318 respectively. The main direction of groundwater flow is therefore from south-east to north-west,
319 which coincides with the results from seepage measurements conducted by Ala-aho et al. (2013). The
320 direction of groundwater flow can also be noted from the difference in the mean isotopic composition
321 of the lake water between points 2 and 3 of -8.7‰ and -73‰ and -8.5‰ and -72‰ respectively.
322 The difference in isotopic compositions between points 2 and 3 was greatest during the winter: in
323 March 2011 the difference in $\delta^{18}\text{O}$ and $\delta^2\text{H}$ between these points was -1.1‰ and -4‰ respectively.

324

325 **4.3. Temporal variations in the isotopic composition of lake water**

326 The seasonal variability in the isotopic composition of the lakes studied is illustrated in Fig. 6,
327 showing changes of $\delta^{18}\text{O}$ in lake Ahveroinen 1 at two depths (1 and 4 meters). $\delta^{18}\text{O}$ of lake
328 Ahveroinen 1 reveals distinct seasonal fluctuations with peak-to-peak amplitude in the order of 1 ‰.
329 This lake does not have any surface inflows or outflows. After the disappearance of ice cover during
330 the spring (April-May), the lake starts to evaporate, which results in its gradual enrichment in heavy
331 isotopes, approaching the steady-state value sometime in September-October. Freezing of the lake in
332 late autumn stops the evaporation flux. Systematic decline of $\delta^{18}\text{O}$ during ice-cover period seen in
333 Fig. 6 stems from gradual dilution of lake water with groundwater seeping into the lake. Figure 6
334 shows that the lake is well mixed throughout the year.

335 The declining parts of the $\delta^{18}\text{O}$ curve in Fig. 6 can be used to assess the intensity of groundwater
 336 inflow during ice-cover period, if the volume of the studied lake is known and the isotopic
 337 composition of groundwater is constant. The isotope balance of such lake system (eq. 4) can be then
 338 expressed as follows:

$$339 \quad V \frac{d\delta_L}{dt} = \delta_{IT} \cdot I_{TOT} - \delta_{OT} \cdot O_{TOT} \quad (9)$$

340 Since $I_{TOT} = I_{GW}$, $\delta_{IT} = \delta_{GW}$, $O_{TOT} = O_{GW} = I_{GW}$ and $\delta_{OT} = \delta_L$, eq. (9) becomes:

$$341 \quad \frac{d\delta_L}{dt} = \frac{I_{GW}}{V} (\delta_{GW} - \delta_L) \quad (10)$$

342 The solution of this differential equation reads as follows:

$$343 \quad \delta_L = (\delta_{Lo} - \delta_{GW}) \cdot e^{-k \cdot t} + \delta_{GW} \quad (11)$$

344 where $k = I_{GW}/V$ and δ_{Lo} is the isotopic composition of the lake at the beginning of ice-cover period
 345 ($\delta^{18}\text{O}_{Lo} = -8.5 \text{ ‰}$ for lake Ahveroinen 1). First derivative of eq. (11) at $t = 0$ is:

$$346 \quad \left(\frac{d\delta_L}{dt} \right)_{t=0} = \frac{I_{GW}}{V} (\delta_{Lo} - \delta_{GW}) \quad (12)$$

347 Equation (12) allows to calculate the mean flux of groundwater to the lake during ice-cover period.
 348 The observed reduction of $\delta^{18}\text{O}$ of lake Ahveroinen 1 by ca. 1.2 ‰ over the six-month period (cf.
 349 Fig. 6, Table 1) results from the continuous inflow of groundwater with specific isotopic signature
 350 ($\delta^{18}\text{O}_{GW} = -13.4 \text{ ‰}$). The groundwater seepage rate obtained from eq. (12) is ca. $160 \text{ m}^3 \text{ day}^{-1}$.
 351 Identical value was obtained using relevant ^2H data. It is worth noting that groundwater inflow to lake
 352 Ahveroinen 1, derived for the summer period of 2013 from isotope mass balance calculations (ca.
 353 $300 \text{ m}^3 \text{ day}^{-1}$) is almost two times higher, which suggests significant seasonal variations of
 354 groundwater inflow to Rokua lakes, with high groundwater fluxes during summer and low fluxes
 355 during winter.

356

357 **4.4. Quantifying groundwater dependence of the studied lakes**

358 Out of 67 Rokua lakes sampled during July-August 2013 field campaign, 50 lakes do not reveal any
 359 surface water inflow in the form of a stream or a creek. For the remaining 17 lakes surface inflows

360 were identified. For all but one lake those surface inflows could be linked to specific upstream lakes
361 which were sampled during the July-August 2013 campaign.

362 The isotopic composition of the lakes sampled covers a wide range of δ values: from -5.6 ‰ to -12.7
363 ‰ and from -57 ‰ to -93 ‰ for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ respectively. This large variability reflects a wide
364 spectrum of the heavy isotope enrichment of the lakes studied. Since Rokua lakes are situated in a
365 unique climatic region, their observed isotopic composition is primarily controlled by their water
366 balance, which in turn can be characterised by the total inflow-to-evaporation ratio.

367 Equation (7) was used to calculate the total inflow-to-evaporation ratios (I_{TOT}/E) for all studied lakes.
368 The isotope mass balance calculations were run for the period of June-August 2013 separately for
369 ^{18}O and ^2H . Since the evaporation rates from the studied lakes were derived from eq. (1),
370 independently of any mass balance considerations, the total inflow to each lake could also be
371 calculated from the assessed I_{TOT}/E ratios.

372 The values of the parameters occurring in eq. (7) were derived as follows:

373 (i) The mean isotopic composition of atmospheric moisture (δ_{A}) was calculated with the eq. (8)
374 using the available isotope data for local precipitation. The following mean δ_{A} values were used in
375 isotope mass balance calculations: $\delta^{18}\text{O}_{\text{A}} = -20.4$ ‰ and $\delta^2\text{H}_{\text{A}} = -149$ ‰.

376 (ii) The mean relative humidity, normalized to lake water temperature (h_{N}) was calculated using the
377 mean surface air temperature ($+14.9$ °C), the mean surface water temperature of the lakes ($+19.1$ °C)
378 and the mean relative air-based humidity calculated on the basis of daily mean values available from
379 meteorological station in the area (79.2 %). The resulting mean h_{N} value was 60.2 %.

380 (iii) The total effective isotope fractionation (ϵ) was derived as the sum of the equilibrium and kinetic
381 isotope enrichments ($\epsilon = \epsilon^* + \Delta\epsilon$). The equilibrium isotope enrichment (ϵ^*) values for ^{18}O and ^2H ,
382 were calculated for the mean surface water temperature of $+19.1$ °C using the known temperature
383 dependence of the empirical equilibrium fractionation factors, α_{LV} (Horita and Wesolowski, 1994).
384 The values of kinetic enrichment parameters (C_{k}) used to calculate $\Delta\epsilon$, 14.2 ‰ for ^{18}O and 12.5 ‰
385 for ^2H , were adopted after Gonfiantini (1986). Those values were obtained in wind-tunnel
386 experiments (Vogt, 1976) and are widely used in lake studies.

387 (iv) Isotopic homogeneity of the studied lakes was addressed through multiple samplings of large
388 lakes (in both horizontal and vertical direction - cf. section 3.1). The range of the measured $\delta^{18}\text{O}$
389 values for the given lake was generally lower than one per mil. It was assumed that the average values

390 calculated on the basis of individual measurements performed in each lake represent sufficiently well
391 the studied systems.

392 (v) The isotopic composition of lake water obtained during the sampling campaign in July and
393 August 2013 for each studied lake was used in the isotope mass balance calculations. As discussed
394 in section 3.4 above, the isotopic compositions of the studied lakes fluctuate seasonally reaching their
395 steady-state values toward the end of ice-free period (September-October). Therefore, the isotopic
396 compositions of lake water samples collected during the late summer (July-August) may still deviate
397 slightly from the respective steady-state values, thus creating uncertainty in the assessed components
398 of the water (see section 4.5).

399 (vi) As the isotopic compositions of the surface and underground components of water inflow to each
400 studied lake were not measured directly, an iterative approach was adopted to calculate δ_T
401 individually for each lake. In the first step it was assumed that δ_T can be defined by the intercept of
402 the LMWL and LEL lines (Fig. 4) which represents the weighted annual mean $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values
403 of local precipitation (-14.1‰ and -100‰ respectively). With these δ_T values the underground
404 component of I_{TOT} for each lake was derived. The second step differed for the lakes without surface
405 inflow and with identified surface inflow from an upstream lake. For the lakes without surface inflow
406 the δ_T values were calculated individually as flux-weighted averages of the underground inflow
407 obtained from the first/previous step and precipitation input to the given lake, each with their
408 respective mean isotopic compositions representing the period of June-August 2013 (cf. section 4.2
409 above). For the lakes with identified surface inflow from an upstream lake, in the first instance the
410 total outflow of water from the upstream lake was calculated using appropriate mass balance equation.
411 Then, it was assumed that 25 % of the total outflow from the upstream lake flows to the downstream
412 lake as surface inflow carrying the isotopic composition characteristic for the upstream lake. The δ_T
413 values were calculated individually for each lake from this group of lakes as flux-weighted averages
414 of three components: precipitation, surface inflow and groundwater inflow. For both lake groups the
415 calculations were repeated until the change of δ_T in subsequent iteration step was in the order of the
416 analytical uncertainty of isotope measurements (0.1‰ for $\delta^{18}\text{O}$ and 1‰ for $\delta^2\text{H}$).

417 The calculated inflow-to-evaporation ratios of the studied lakes based on ^{18}O isotope mass balance
418 are shown in Fig. 7 as a function of isotope enrichment of lake water with respect to the isotopic
419 composition of the total inflow to the given lake. They vary in a wide range, from I_{TOT}/E values
420 between 2 and 3 and large ^{18}O isotope enrichments between approximately 6.5 and 8.5 ‰, indicating
421 evaporation dominated systems, to typical through-flow lakes characterized by I_{TOT}/E ratios higher

422 than 10 and moderate ^{18}O isotope enrichments of less than 2 ‰ (Table 2). Knowing the volume and
423 the total inflow, the mean turnover time of water (MTT) in each lake could be quantified as the ratio
424 of lake volume to the total inflow. The calculated MTT values range from approximately one week
425 for the Pasko pond ($V = 2 \times 10^3 \text{ m}^3$, mean depth 0.2 m, maximum depth 0.7 m) to approximately five
426 years for lake Saarijärvi 2 ($V = 2.47 \times 10^6 \text{ m}^3$, mean depth 11.8 m, maximum depth 26 m), with the
427 mean in the order of ten months (Table 2). Lake Saarijärvi 2 is the deepest of all the lakes surveyed.
428 As expected, the calculated MTT values correlate well with the mean depth of the studied lakes,
429 expressed as the volume-to-surface area ratio. However, the link between MTT and the I_{TOT}/E ratio
430 is much weaker; lakes with higher I_{TOT}/E ratios tend to have shorter mean turnover times.

431 The dependence of the studied lakes on groundwater can be quantified through an index (G index)
432 defined as the percentage contribution of groundwater inflow to the total inflow of water to the given
433 lake. Groundwater inflow was derived by subtracting the precipitation and surface water inflow (if
434 exists) from the total inflow. Such an evaluation was undertaken for all the lakes listed in Table 2.
435 Note that for the group of lakes with identified surface inflow from an upstream lake it was assumed
436 arbitrarily that this surface inflow is 25 % of the total outflow from the upstream lake (discharges of
437 surface inflows were not measured). The resulting groundwater seepage rates vary from less than 20
438 $\text{m}^3\text{day}^{-1}$ for Kissalampi pond to around $15 \times 10^3 \text{ m}^3\text{day}^{-1}$ for lake Nimisjärvi, the lake with the largest
439 surface area (167.5 ha) among all studied lakes.

440 Figure 8 summarises the values of G index obtained for the lakes surveyed during the July-August
441 2013 sampling campaign. The mean values of G index obtained from ^2H and ^{18}O balance are shown.
442 They vary from ca. 41 % to more than 95 %. The lowest value (40.7 %) was obtained for lake Etu-
443 Salminen. The highest G values were derived for lakes Kiiskeroinen (97.3 %) and Levä-Soppinen
444 (97.0 %). Interestingly, these lakes are characterised by a high degree of eutrophication induced by
445 high loads of phosphorus brought to the lakes with groundwater (Ala-aho et al. 2013). Although the
446 G index describes groundwater dependency of the studied lakes rather unambiguously, also lakes
447 with moderate G index values can suffer if groundwater table in the esker aquifer would decline as a
448 result of climate and/or land-use changes.

449 The isotope mass balance calculations for Rokua lakes were run independently for ^{18}O and ^2H data.
450 Consistent results were obtained with respect to three elements of lake water balance (I_{TOT}/E ratios,
451 MTT values and the G index) reported in Table 2. These quantities, derived independently from ^2H -
452 based and ^{18}O -based isotope mass balances are highly correlated ($R^2 = 0.9887$, 0.9963 and 0.9832 for

453 I_{TOT}/E ratios, MTT values and the G index, respectively). The total inflow-to-evaporation ratios
454 derived from 2H -based balance turned out to be ca. 5 % higher on the average than those derived from
455 ^{18}O -based balance. For the G index this difference is approximately 0.8 %. The MTT values were
456 ca. 4.3 % higher for ^{18}O -based balance.

457

458 **4.5. Uncertainty assessment**

459 The above methodology for quantifying elements of water balance in the lakes studied introduces
460 some uncertainties linked to the assumptions made and the uncertainties associated with the
461 parameters used in the evaluation process. Sensitivity tests were performed to derive the range of
462 uncertainties associated with the quantities being evaluated, such as mean turnover time, total inflow-
463 to-evaporation ratio and the G index. The sensitivity analysis was focussing on eq. (7). All variables
464 present in this equation were considered in this process. The results for ^{18}O -based calculations are
465 summarized in Table 3.

466 The uncertainty with regard to lake water temperature was probed assuming the temperature change
467 of ± 0.87 °C (cf. section 4.1). The uncertainty of lake water temperature leads to uncertainty as regards
468 the evaporation flux, which in turn influences the I_{TOT}/E , MTT and G values derived for each lake.
469 Also equilibrium isotope enrichment is a function of temperature. The mean turnover time increases
470 by ca. 15 % when the temperature of the lake is reduced by 0.87 °C, and decreases by approximately
471 12 % when the temperature increases by the same amount. The G index reveals lower sensitivity (2.9
472 and 3.6 %, respectively). The smallest changes were obtained for I_{TOT}/E ratios (0.8 and 0.9 %,
473 respectively).

474 The changes of relative humidity of the atmosphere normalised to the temperature of the lake surface
475 have an impact on the evaporation flux, control the I_{TOT}/E ratios through eq. (7) and determine the
476 actual value of kinetic isotope enrichment $\Delta\epsilon$. It was assumed in the calculations that normalized
477 relative humidity changes by ± 2 %. As seen in Table 3, the resulting changes of the derived quantities
478 are moderate, the mean turnover time being the most sensitive parameter.

479 It is apparent from Table 3 that among isotope parameters occurring in eq. (7), the isotopic
480 composition of lake water (δ_{LS}) and the isotopic composition of the total inflow (δ_{IT}), are the two
481 most important variables in the isotope mass balance calculations. An increase of $\delta^{18}O_{LS}$ by 0.5 ‰,
482 which may account for possible departures from the isotopic steady-state of the investigated lakes

483 (cf. section 4.4), leads to decrease of calculated I_{TOT}/E ratios on the average by 15.5 %, increase of
484 MTT values by 19.1 % and decrease of G index values by 5.8 %. An increase of $\delta^{18}O_{IT}$ by 0.5 ‰,
485 which may result from the contribution of an evaporated lake water originating from an upstream
486 lake to groundwater input, leads to substantial increase of I_{TOT}/E ratios (20.1 % on the average),
487 comparable decrease of MTT values (14.7 % on the average) and moderate increase of the G index
488 (4.3 % on the average). Variation of the ^{18}O isotopic composition of atmospheric water vapour by \pm
489 1.0 ‰ introduces comparable changes in the calculated elements of the water balance of the studied
490 lakes in the order of several per cent (Table 3).

491 Figure 9 shows the percentage changes of I_{TOT}/E ratios calculated for all studied lakes using eq. (7),
492 in response to the increase of δ_{LS} or δ_{IT} by 0.5 ‰. The sensitivity of the calculated I_{TOT}/E ratios to the
493 given increase of δ_{LS} or δ_{IT} raises sharply with increasing value of this parameter. This is particularly
494 true for through-flow systems characterized by high I_{TOT}/E ratios. Therefore, when isotope studies
495 aimed at quantifying water balance of such systems are planned, it is important to characterize these
496 two isotope quantities as good as reasonably possible.

497

498 **5. Conclusions**

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

508 The presented study has demonstrated that consistent results are obtained when the isotope mass
509 balance calculations are run independently for oxygen-18 and deuterium. This strengthens the
510 position of heavy stable isotopes of water as a unique tool for quantifying elements of water balance
511 of lakes, particularly for groundwater-dominated systems. Solving three equations simultaneously
512 (one water balance equation plus two isotope balance equations) can result in quantifying three key

513 balance-related parameters such as evaporation and groundwater inflow and outflow rates for the
514 studied lake system, which are difficult to quantify using conventional methods.

515 The specific behaviour of lakes located in sub-polar regions, with their seasonal ice cover extending
516 over several months, offers another opportunity for quantifying groundwater seepage during ice-
517 cover periods. As shown in this study, observations of seasonal changes in the stable isotopic
518 composition of lake water, in particular during the ice-cover period, combined with the survey of
519 isotopic composition of groundwater in the vicinity of the lakes studied, allows the quantification of
520 groundwater fluxes to this lake during winter. If such an approach is combined with the isotope mass
521 balance calculations performed for ice-free summer season, important information about the seasonal
522 variability of groundwater seepage to lakes located in sub-polar and polar regions can be obtained.

523 The *G* index of groundwater dependency of a lake proposed in this study, and defined as a percentage
524 contribution of groundwater inflow to the total inflow of water to the given lake, appears to be a
525 straightforward, quantitative measure of this dependency. The studied Rokua lakes appear to be
526 strongly dependent on groundwater; more than 40 % of water received by these lakes comes as
527 groundwater inflow. The quantitative evaluation of groundwater dependency of lakes via the *G* index
528 proposed in this study may assist lake restoration policies in areas where groundwater is a source of
529 nutrients to the studied lakes.

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536

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693 Table 1. Mean isotopic composition of selected lakes, streams and groundwater, sampled four times
 694 per year during the period 2010-2012.

	$\delta^2\text{H}$ (‰)	$\delta^{18}\text{O}$ (‰)	d-excess (‰)	Mean amplitude of $\delta^{18}\text{O}$ seasonal signal (‰)
Lakes:				
1. Ahveroinen 1	-72.4	-8.50	-4.42	1.2
2. Rokuanjärvi	-72.1	-8.75	-2.14	1.3
3. Jaakonjärvi	-64.1	-6.72	-10.34	0.8
4. Kolmonen 2	-66.7	-7.07	-10.14	0.9
5. Loukkojärvi	-63.0	-6.64	-9.88	0.9
6. Saarijärvi 2	-62.4	-6.81	-7.92	0.9
7. Saarinen	-72.3	-8.52	-4.14	3.3
8. Salminen	-71.4	-8.45	-3.80	0.5
9. Soppinen	-63.1	-6.84	-8.38	0.9
10. Tulijärvi	-85.6	-11.20	4.00	1.3
11. Vaulujärvi	-67.9	-7.68	-6.46	0.5
Streams:				
1. Heinäjoki	-93.0	-12.74	8.92	0.5
2. Hieto-oja	-88.9	-11.84	5.82	1.2
3. Kangasoja	-93.7	-12.90	9.50	0.7
4. Lianoja	-83.2	-10.67	2.16	2.2
5. Lohioja	-92.9	-12.93	10.54	0.8
6. Matokanava	-94.6	-13.06	9.88	0.9
7. Päiväkanava	-94.0	-13.01	10.08	0.9
8. Rokuanoja	-82.8	-10.95	4.80	3.1
9. Siirasoja	-94.6	-13.08	10.04	0.7
10. Soppisenoja	-89.9	-12.07	6.66	2.2
11. Valkiaisaja	-93.3	-12.80	9.10	1.0
Groundwater:				
1. MEA 106	-96.4	-13.32	10.16	0.6
2. MEA 206	-95.0	-13.12	9.96	0.8
3. MEA506	-93.5	-12.92	9.86	0.5
4. MEA 706	-93.3	-12.89	9.82	0.7
5. MEA 1106	-95.1	-13.05	9.30	0.8
6. MEA 1807	-93.3	-12.92	10.06	1.0
7. MEA 1907	-83.2	-10.65	2.00	1.0
8. MEA 2010	-97.3	-13.42	10.06	0.5
9. ROK 1	-95.5	-13.22	10.26	0.9
10. Siirasoja 1 esker	-94.5	-13.06	9.98	0.8
11. Siirasoja 1 slope	-94.3	-13.08	10.34	1.0
12. Siirasoja 1 sand	-97.0	-13.36	9.88	0.5
13. Siirasoja 1 peat	-93.8	-12.98	10.04	1.2

696 Table 2. The results of isotope mass balance calculations for 67 lakes of Rokua esker sampled during the July-August 2013 field survey.

Lake No.	Lake name	Volume (10 ³ m ³)	Surface area (ha)	Mean depth (m)	<i>E</i> ^(a) (mm)	δ ¹⁸ O (‰)	δ ² H (‰)	<i>I</i> _{GW} (m ³ /day)		<i>I</i> _{TOT} / <i>E</i>		MTT (mo.)		<i>G</i> index (%)	
								¹⁸ O	² H	¹⁸ O	² H	¹⁸ O	² H	¹⁸ O	² H
1	Heinälampi ^(b)	1764	0.22	0.8	286	-10.97	-84.4	72.2	78.5	14.4	15.4	0.6	0.5	72.5	73.4
2	Holma ^(b)	23778	1.89	1.3	257	-11.12	-84.8	575.1	607.6	15.9	16.4	0.9	0.9	68.4	70
3	Koivujärvi 1 ^(b)	39237	2.3	1.7	254	-11.38	-85.7	832.1	793.5	19.2	18.7	1	1.1	68.2	66.6
4	Koivujärvi 2 ^(b)	76884	2.93	2.6	251	-11.61	-87.3	1571.8	1610.8	20.9	21.4	1.5	1.5	94.2	94.1
5	Pitkäjärvi 1	36103	1.22	3	262	-6.50	-61.5	62.9	68.9	2.6	2.8	13.1	12.3	70.2	72.1
6	Nurkkajärvi	269134	3.93	6.8	247	-8.76	-73.1	413.1	442.9	4.7	5	17.6	16.6	82.7	83.7
7	Luontolampi	19995	0.49	4.1	275	-7.29	-64.9	35.2	36.7	3.2	3.3	14.2	13.8	76.6	77.4
8	Ahveroinen 1	119000	3.32	3.6	249	-8.32	-70.6	300.8	318.0	4.2	4.3	10.4	9.9	80.5	81.3
9	Lianjärvi ^(b)	113678	15.13	0.8	231	-10.31	-80.5	2402.1	2434.2	10.1	10.4	1	0.9	62.3	61.5
10	Syväjärvi 1 ^(b)	398463	11.52	3.5	234	-11.17	-85.3	3321.8	3525.4	12.8	13.6	3.5	3.3	88.4	88.5
11	Soppinen	81630	6.04	1.4	242	-6.91	-62.3	320.5	321.5	2.8	2.9	5.9	5.9	70.7	70.8
12	Salminen	681860	25.31	2.7	225	-8.33	-69.7	2015.4	2000.5	4.1	4.1	8.6	8.7	78.4	78.3
13	Saarinen	981506	15.32	6.4	231	-7.84	-67.5	1053.9	1079.0	3.6	3.7	23	22.6	75.8	76.2
14	Kivi-Ahveroinen	216018	5.57	3.9	243	-9.02	-74.4	631.6	675.6	5.1	5.4	9.3	8.8	83.8	84.7
15	Irvi-Ahveroinen	40867	1.07	3.8	264	-7.47	-65.6	77.4	79.4	3.3	3.4	13.2	13	76.8	77.3
16	Loukkojärvi	126450	2.85	4.4	251	-6.76	-63.0	151.3	166.9	2.7	2.9	19.3	18	70.7	72.7
17	Ylimmäinen	94313	8.93	1.1	237	-9.66	-77.6	1266.9	1351.6	6.3	6.7	2.1	2	86.6	87.3
18	Hietajärvi	223118	7.3	3.1	240	-7.06	-63.7	401.4	419.2	3	3	13	12.6	71.5	72.3
19	Saarijärvi 2	2467169	20.94	11.8	228	-6.71	-61.1	938.4	935.6	2.7	2.7	57.5	57.7	67.1	67
20	Syväjärvi 2 ^(b)	1863446	31.95	5.8	223	-10.63	-83.1	7761.1	8787.7	12.2	13.5	6.4	5.8	82.4	83.8
21	Pasko ^(b)	2005	0.82	0.2	268	-10.62	-83.3	191.2	218.4	11	12.2	0.3	0.2	73.3	75.6
22	Kuikkalampi	11776	0.61	1.9	271	-7.20	-66.6	42.2	50.3	3.1	3.5	6.9	6	75.8	78.9
23	Soppisenlampi	25645	0.59	4.4	272	-7.32	-67.1	42.0	49.7	3.2	3.6	15.2	13.4	76.6	79.5
24	Kirvesjärvi	654618	13.47	4.9	233	-11.45	-87.6	5260.9	6477.3	18.8	22.4	3.3	2.8	82.1	84.7
25	Tulijärvi ^(b)	589617	24.81	2.4	226	-10.47	-81.3	4895.4	4976.3	11.4	11.6	2.8	2.7	70.8	70.6
26	Jaakonjärvi 2	5497	0.46	1.2	275	-7.39	-65.3	34.6	35.6	3.2	3.3	4	3.9	77.3	77.8
27	Jaakonjärvi 3	13594	0.66	2.1	271	-7.60	-66.4	51.5	53.0	3.4	3.5	6.7	6.6	78.1	78.6
28	Maitolampi 2	44812	2.04	2.2	256	-7.52	-65.9	144.2	148.2	3.3	3.4	7.7	7.6	76.3	76.8

29	Kotalampi	25980	2.6	1	253	-11.23	-86.3	869.4	1015.4	13	15	0.9	0.8	93.8	94.7
30	Rokuanjärvi 1	4364781	164.59	2.7	205	-8.54	-70.1	12484.5	11837.8	4.4	4.2	8.8	9.2	77.6	76.6
31	Tervatienlampi	21970	0.63	3.5	271	-7.92	-69.3	55.1	61.5	3.7	4.1	10.4	9.5	80	81.7
32	Valkiaislampi	32073	0.7	4.6	270	-8.81	-72.5	83.5	84.8	4.8	4.9	10.6	10.4	84.4	84.6
33	Ankkalampi	103700	3.86	2.7	248	-8.57	-71.9	378.5	401.8	4.5	4.7	7.3	6.9	81.7	82.6
34	Saarilampi 1	12044	0.45	2.7	276	-10.51	-80.8	111.6	107.3	9	8.7	3.2	3.4	91.8	91.5
35	Kiiskeroinen	6517	0.63	1	271	-12.14	-90.2	487.0	512.5	26.9	28.2	0.4	0.4	97.2	97.4
36	Jaakonjärvi 1	93614	3.53	2.7	249	-6.56	-62.5	172.1	196.9	2.6	2.9	12.2	11.1	69	71.8
37	Vaulujärvi	432813	8.97	4.8	237	-7.57	-66.7	581.5	619.8	3.4	3.5	18.1	17.3	74.7	75.9
38	Levä-Soppinen	34000	2.31	1.5	254	-12.27	-91.4	1939.5	2351.8	31.2	37.7	0.6	0.5	97.5	97.9
39	Anttilanjärvi	60230	1.06	5.7	264	-7.40	-66.4	75.0	82.5	3.2	3.5	20	18.6	76.4	78.1
40	Hautajärvi 1	189222	2.56	7.4	253	-7.71	-68.0	189.9	208.9	3.5	3.8	25.1	23.3	77.2	78.8
41	Lepikonjärvi	191118	2.98	6.4	251	-7.77	-68.3	223.6	246.2	3.6	3.8	21.6	20	77.4	79
42	Kolmonen 1	35700	0.68	5.3	270	-7.09	-65.2	44.6	50.5	3	3.3	19.6	17.8	75	77.3
43	Kolmonen 2	21890	0.56	3.9	273	-6.82	-64.9	34.1	41.1	2.8	3.2	15.4	13.4	73.6	77
44	Kolmonen 3	21630	0.54	4	273	-7.26	-66.1	38.0	43.0	3.1	3.4	14.2	12.9	76.3	78.5
45	Hätäjärvi	35217	1.73	2	258	-6.50	-61.8	87.0	96.9	2.6	2.8	9.2	8.5	69.6	71.9
46	Kissalampi	800	0.36	0.2	279	-5.60	-57.1	14.8	16.5	2.1	2.3	1.2	1.1	65.4	67.9
47	Valkiajärvi	582662	8.36	7	238	-8.49	-71.2	758.2	788.5	4.4	4.5	20.2	19.5	80.5	81.1
48	Hautajärvi 2	445413	14.6	3.1	232	-7.43	-64.5	871.2	845.7	3.2	3.2	12.2	12.5	73.1	72.5
49	Keskimmäinen ^(b)	428515	13.07	3.3	233	-10.76	-82.8	3115.7	3123.8	10.9	11.1	3.9	3.8	86.3	85.3
50	Siirasjärvi 2	23126	0.56	4.1	273	-12.66	-93.5	985.8 ^(d)	1494.3 ^(d)	60.2 ^(d)	90.9 ^(d)	0.8 ^(d)	0.5 ^(d)	98.8 ^(d)	99.2 ^(d)
51	Siirasjärvi 1	9701	0.34	2.9	280	-10.43	-80.4	82.3	78.7	8.7	8.3	3.5	3.7	91.7	91.3
52	Telkkälampi	7284	0.25	2.9	284	-7.30	-66.3	19.0	21.4	3.2	3.5	9.7	8.8	77.5	79.5
53	Maitolampi 1	48892	1.56	3.1	259	-6.36	-61.2	75.2	84.7	2.5	2.7	14.6	13.4	68.8	71.3
54	Taka-Salminen ^(b)	364757	7.31	5	240	-9.63	-76.2	790.1	740.2	7.8	7.6	8	8.3	53.4	51.4
55	Etu-Salminen ^(b)	196453	5.83	3.4	243	-8.05	-68.5	316.6	330.4	5.1	5.2	8.1	8	40.1	41.4
56	Pikku-Salminen	39856	1.32	3	261	-7.61	-65.8	98.9	98.2	3.4	3.4	10.2	10.2	77.4	77.3
57	Kylmäjärvi	394443	7.89	5	239	-8.47	-70.3	713.0	702.4	4.3	4.3	14.5	14.7	80.5	80.2
58	Kourujärvi 1	68023	1.83	3.7	257	-8.55	-71.7	186.3	196.3	4.4	4.6	9.8	9.4	82.3	83
59	Kourujärvi 2	31136	1.91	1.6	256	-7.64	-67.3	141.7	152.5	3.4	3.6	5.5	5.2	77.1	78.4
60	Huttunen	54588	1.49	3.7	260	-7.50	-66.9	107.2	117.5	3.3	3.6	12.7	11.8	76.6	78.2

61	Saarijärvi 1	13590	1.36	1	261	-6.49	-61.4	69.3	75.7	2.6	2.7	4.5	4.2	69.9	71.7
62	Pitkälkäjärvi 2	451093	7.93	5.7	239	-8.05	-68.7	616.2	635.5	3.8	3.9	18.6	18.2	78	78.5
63	Pyöräinen ^(b)	76077	3.76	2	248	-7.91	-67.9	274.8	282.1	4.8	4.9	5.1	5	56.1	56.6
64	Likainen ^(b)	33763	8.28	0.4	238	-10.94	-83.4	2132.2	2020.9	13.3	13.1	0.4	0.4	74.9	72.1
65	Nimisjärvi ^{(b), (c)}	1840396	167.53	1.1	205	-8.66	-69.7	15971.5	14421.5	5.3	4.8	3.1	3.3	81.3	79.7
66	Ahveroinen 2 ^(b)	670515	16.3	4.1	230	-7.96	-67.8	1093.4	1083.3	4.4	4.4	12.3	12.2	61.4	60.6
67	Tervalampi	29707	0.79	3.7	268	-7.99	-68.7	70.3	73.5	3.8	3.9	11	10.7	80.1	80.8

697 (a) - Calculated for the period June 1 - August 31, 2013.

698 (b) - Lakes with identified surface inflow from an upstream lake.

699 (c) - Since the upstream lake for lake Nimisjärvi was not sampled, the mean isotopic composition of total inflows to lakes with identified surface water
700 inflows was used for isotope mass balance calculations of this particular lake.

701 (d) - As the isotopic composition of a lake becomes comparable with the isotopic composition of the total inflow, the isotope mass balance calculations
702 become very uncertain. Therefore, for lake Siirasjärvi 2 ($\Delta\delta^{18}\text{O} = 0.45\text{‰}$ and $\Delta\delta^2\text{H} = 1.6\text{‰}$) the values of I_{TOT}/E , MTT and G index reported in the
703 table have only indicative character.

704 Table 3. Sensitivity of selected elements of ^{18}O -based water balance of the studied lakes to changes
 705 of the parameters involved.
 706

Parameter	Parameter change	Mean change of the selected elements of water balance (%) ^(a) :		
		I_{TOT}/E	MTT	G index
1. Lake water temperature ($T_w = 19.1$ °C)	+0.87 °C	-0.8	-11.8	+2.9
	-0.87 °C	+0.9	+14.8	-3.6
2. Normalized relative humidity ($h_N = 60.7$ %)	+2.0 %	-1.5	+7.0	-1.9
	-2.0 %	+1.2	-6.1	+1.6
3. Isotopic composition of atmospheric water vapour ($\delta^{18}\text{O}_A = -20.4$ ‰)	+1.0 ‰	+7.4	-6.9	+2.3
	-1.0 ‰	-7.4	+8.0	-2.7
4. Isotopic composition of lake water ($\delta^{18}\text{O}_{\text{LS}}$)	+0.5 ‰	-15.5	+19.3	-5.8
5. Isotopic composition of the total inflow ($\delta^{18}\text{O}_{\text{IT}}$)	+0.5 ‰	+20.1	-14.7	+4.3

707 (a) The (+) and (-) signs signify an increase or reduction, respectively, of the derived quantity by the given
 708 percentage.
 709

710 **Figure captions**

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

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

716 Figure 3. Daily mean surface water temperatures of lakes Oulujärvi (92,800 ha) (Finnish
717 Environmental Institute, 2013), Ahveroinen 1 (3.3 ha) and Saarinen (15.3 ha) during the summer of
718 2013, compared with the surface air temperature data for the same period. Lake Oulujärvi is located
719 in the east, next to the study site, 1 km from the easternmost lake studied. Vertical lines mark the
720 period used in the calculations of evaporation and isotope mass balance.

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

725
726 Figure 5. Mean $\delta^{18}\text{O}$ values (‰) of lake Ahveroinen and adjacent groundwater. The mean $\delta^{18}\text{O}$ value
727 for site 1 is 8.5 ‰ at both sampled depths (1 m and 4 m).

728
729 Figure 6. Seasonal variations of $\delta^{18}\text{O}$ in lake Ahveroinen 1, observed at 1 and 4-meter depths.
730 Maximum depth of the lake is 4.8 m.

731 Figure 7. The ratio of total inflow to evaporation (I_{TOT}/E) as a function of the measured ^{18}O isotope
732 enrichment ($\Delta\delta^{18}\text{O}$) for Rokua lakes surveyed during the July-August 2013 sampling campaign.

733 Figure 8. The G index quantifying the groundwater dependency of lakes in the Rokua study area. The
734 index is defined as the percentage contribution of groundwater inflow to the total inflow of water to
735 the given lake. Shown are the mean G values obtained from independent isotope mass balance
736 calculations based on ^2H and ^{18}O data.

737 Figure 9. Changes of the total inflow-to-evaporation ratio (in %) as a function of I_{TOT}/E value, in
738 response to parameter change, calculated for the studied lakes on the Rokua esker. Two cases are
739 considered: (a) an increase of the measured $\delta^{18}\text{O}$ of lake water by 0.5 ‰, and (b) an increase of $\delta^{18}\text{O}$
740 of the total inflow to the given lake by 0.5 ‰. See text for details.

741