Use of water isotopes and chemistry to infer the type and degree of exchange between groundwater and lakes in an esker complex of northeastern Ontario, Canada

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Abstract

Water chemistry and water isotopic composition were investigated in a set of 50 boreal lakes located at different elevations in an esker system near Timmins, Ontario, as well as in local streams, groundwater springs and information available from seasonal precipitation values. Insights gained from this study are essential to develop a more detailed understanding on the local hydrology as the region has widespread cottage development, mining activities and recreational fishing. Analyses focused on stable isotopic ratios of hydrogen and oxygen, as well as specific conductance as indicators of the position of a lake with respect to the influence of groundwater. Both isotopic composition and specific conductance distinguished higher elevation groundwater recharge lakes from lower elevation groundwater discharge lakes. Groundwater recharge lakes characterized by enriched isotopic values and low values of specific conductance are located above the hydraulic midline elevation of the study lakes. In contrast, groundwater discharge lakes were isotopically depleted and had higher values of specific conductance, and occurred below the hydraulic midline of the study lakes. An intermediate group of lakes was also defined (termed seepage lakes) and consisted of either recharge lakes that were alkaline, or discharge lakes that had no outlet. The seepage lakes group had intermediate isotopic and water chemistry characteristics compared to recharge and discharge lakes. A classification scheme for lakes was developed based on the specific conductivity, water isotopic composition, the presence of an outlet, and other characteristics to define three types of recharge lakes, and two types of discharge lakes.

Interannual (2013 and 2014) and summertime differences in water chemistry between early June and August revealed that upland groundwater recharge lakes showed evidence of evaporative drawdown, indicating sensitivity to short-term changes in climate, whereas the lower-elevation discharge lakes showed little variation between seasonal samples, and consequently would likely be affected only by hydroclimatological changes of greater duration and persistence.

Keywords

Water balance, landscape limnology, groundwater, hydrochemistry, lake-water chemistry, stable isotopes

1 Introduction

Nearly all surface waters, including lakes, interact with groundwater to some extent (Winter et *al*., 1998). The degree to which lakes interact with groundwater strongly influences lake-water chemistry and water balance as well as aquatic biota and biochemical cycles (Rosenberry et *al*., 2015; Lewandowski et *al*., 2015). These characteristics will influence the sensitivity and vulnerability of lakes to environmental stressors such as climate change, and other anthropogenic disturbances. Future climate change will likely affect lakes differently depending on their connectivity to groundwater, further increasing our need to better understand the relations between lake hydrology and lake water geochemistry. Further, understanding the climatic controls on water balance and adverse impacts to aquatic ecosystems is essential for informed ecosystem management and conservation practices (Winter et *al*., 1998).

Groundwater-lake water interactions are an often poorly-studied component of lake hydrology. However, recent advances in isotopic techniques and modelling approaches have enabled researchers the opportunity to better understand hydrological processes in lakes at a local to regional scale (Fleckenstein et *al*., 2010; Rosenberry et *al*., 2015). This has included studies using numerical simulations (e.g. Winter et *al*., 1998), as well as empirical studies at a range of different spatial and temporal scales. Studies at the regional scale (e.g. Gorham et *al*., 1983; Webster et *al*., 2000) emphasize differences in climate and surface geology as being the important drivers of changes in lake-water chemistry. By contrast, empirical studies performed at a local to regional scale (e.g. Webster et *al*., 1996; Quinlan et *al*., 2003; Martin and Soranno, 2006; Ala-aho et *al*., 2013) indicate the importance of landscape position and also lake morphology as being important in understanding lake-groundwater interactions. At the scale of an individual lake, a number of studies have shown that groundwater-lake water interactions can vary temporally according to changes in seasonality and longer term changes in hydroclimatic conditions (e.g. Kenoyer and Anderson, 1989; LaBaugh et *al*., 1997; Sebestyen and Schneider, 2001; Schuster et *al*., 2003; Arnoux et *al*., 2017b).

The vast majority of past empirical studies have relied on studying aspects of water chemistry (e.g. dissolved ions), as an indicator of hydrological connectivity between lakes, since such approaches can contain signals of important processes such as mineral weathering and dissolution (Bertrand et *al*., 2014). Chemical tracers are often referred to as non-conservative tracers because their composition changes as they react with catchment materials (Kendall and McDonnell, 1998). By contrast, other studies relied on the isotopic composition of water as a hydrological tracer (e.g. Turner et *al*., 2010; Isokangas et *al*., 2015). The latter are labelled to as conservative tracers because they are relatively conservative in reactions with catchment materials and retain their distinctive values until they mix with other water sources or they evaporate (Kendall and McDonnell, 1998). In addition to being a good indicator of source water, stable isotopes of water constitute an efficient and cost-effective means to quantify lake evaporation and water balance status of water bodies using the Craig-Gordon model (Craig and Gordon, 1965), given that the isotopic composition of lake-water and precipitation as well as air temperature and relative humidity are known, or can be estimated (Gibson and Edwards, 2002). The Craig-Gordon model is well established and has been used extensively to investigate the spatial and temporal variability in lake water balance in remote boreal regions of Canada (e.g. Wolfe et *al*., 2007; Bouchard et *al*., 2013; Tondu et *al*., 2013; Turner et *al*., 2014a; Turner et *al*., 2014b). However, precipitation and groundwater often display similar isotopic signatures as they both tend to retain their original isotopic composition because they undergo little to no evaporation (Gibson and Edwards, 2002; Gibson et *al*., 2008; Yi et *al*., 2008). This makes the distinction of the relative influence of groundwater and precipitation in lake water balance challenging. Given this, the combination of chemical and isotopic approaches has the ability to produce more reliable interpretations, especially if the two approaches converge on a mutually reinforcing interpretation. Chemical and isotopic tracers have been widely used together to investigate the connectivity between groundwater and lake water within a single lake (e.g. Labaugh et *al*., 1997; Schuster et *al*., 2003; Rautio and Korkka-Niemi, 2011) or for a cluster of a few selected lakes (e.g. Arnoux et *al*., 2017a; Arnoux et *al*., 2017b; Gurrieri and Furniss, 2004; Katz et *al*., 1997; Turner and Townley, 2006). Nonetheless, studies that have combined chemical and isotopic approaches to investigate the connectivity between groundwater and lake water for a large cluster of lakes at the scale of the landscape are lacking.

The main objective of this study is to examine the importance of landscape position on groundwater connectivity by scrutinising both water chemistry and isotopic composition of water in a boreal esker complex in northeastern Ontario. Investigating such processes in the context of esker hydrology is particularly relevant as eskers consist of porous and permeable materials that facilitate groundwater flows, are widespread in boreal regions (Ala-aho et *al*., 2015) and constitute one of the most common type of aquifers in boreal Canada (Cloutier et *al*., 2007). Results from the observations above, will be used to develop a lake typology of hydrological connectivity based landscape position and lake characteristics (Newton and Driscoll, 1990; Bertrand et *al*., 2014). The resultant typology will provide insights on lakes' sensitivity and vulnerability to environmental stressors such as climate change, for water management and conservation goals as the region is the site of cottage development, recreational fishing, mining activities and aggregate extraction (Cochrane, 2006). Additionally, the typology will be used for site-selection and interpretations of past hydrological changes from stratigraphic analysis of isotopic and geochemical indicators in sediment cores from lakes. Finally, we will evaluate the sensitivity of the proposed typology, by investigating if seasonal variations of lake-water characteristics are more readily detected in higher-elevation groundwater recharge lakes (*i.e.* lakes that receive the majority of their water from precipitation and feed the groundwater system) in comparison to lower elevation groundwater discharge lakes (*i.e.* lakes that receive the majority of their water from groundwater).

2 Study area

The study area is a portion of the Kettle Lake Esker between the southern shore of Frederick House Lake, and the northern shore of Night Hawk Lake, a region located approximately 35-km east of Timmins, Ontario, Canada (**Fig. 1a**). The Timmins region is characterized by a humid continental climate with a mean annual temperature of 1.8°C and average precipitation of 835 mm (Environment Canada, 2015). The area undergoes long cold winters and lakes are covered with ice from early November until early April. Summers are usually wet and mean air temperatures are 17.5°C in July. Many of the study lakes and streams are located within Kettle Lakes Provincial Park, as well as in lower elevation regions that have been moderately influenced by human activities (Cochrane, 2006). The study region is covered with well drained orthic humo-ferric podzols while surrounding clay plains are for the most part covered with moderately to poorly drained orthic gray luvisols and gleyed gray luvisols (OMNDM, 2006). Jack pine, poplar, black spruce, white birch, trembling aspen and balsam fir dominate the well-drained areas, with a dominance of spruce in poorly drained regions.

The eskers are the result of glaciofluvial, glaciolacustrine and related postglacial processes (Nadeau, 2011). The regional landscape is dominated by landforms and deposits of the Laurentide Ice Sheet during the last glacial maximum and deglaciation of the region was essentially complete approximately 10,000 years ago (Dyke, 2004). The ablation of the ice sheet was particularly dynamic and led to the formation of relatively large eskers composed of coarse grained glaciofluvial sediments in deposits oriented in a north-south direction and mantling the crystalline bedrock (Cloutier et *al*., 2007). The retreat of the ice sheet was accompanied by ponding of glacial meltwaters that led to the development of glacial Lake Ojibway that submerged most of the region (Roy et *al*., 2011) and the widespread deposition of glaciolacustrine clay (Nadeau, 2011), followed by the drainage of Lake Ojibway into Hudson Bay ~8,200 years ago (Roy et *al*., 2011). As Lake Ojibway levels dropped, wave action eroded the surface of the esker and redistributed some sand materials on the flanks on the esker, forming lateral littoral sand units that drape the glaciolacustrine clays (Cloutier et *al*., 2007) (**Fig. 1b**). The numerous kettle lakes on the esker formed once the glacial ice trapped in the outwash materials melted (Nadeau, 2011).

The esker stratigraphy ensures that its groundwater system is highly localized since the esker generally have a high hydraulic conductivity due to their coarse texture. The esker is surrounded by underlying bedrock and the adjacent fine-grained glaciolacustrine deposits, both characterized by very low hydrological conductivity (Stauffer and Wittchen, 1992). As a consequence, the esker can be conceptually partitioned into zones of unconfined aquifers in its center where coarse material is present at the surface and zones of confined aquifers at its edges when fine-grained sediment mantles the core of the esker (**Fig. 1b**) (Cloutier et *al*., 2007). Thus, the recharge of the esker will occur through infiltration of precipitation in the unconfined aquifer and discharge will take place on the esker flanks at the contact of the clay, where most groundwater springs emerge (Cloutier et *al*., 2007). Confined aquifers found on the surrounding clay plain are often covered by peatlands and shallow lakes fed by groundwater springs on the edges of the esker or by streams that drain the esker (Rossi et *al*., 2012).

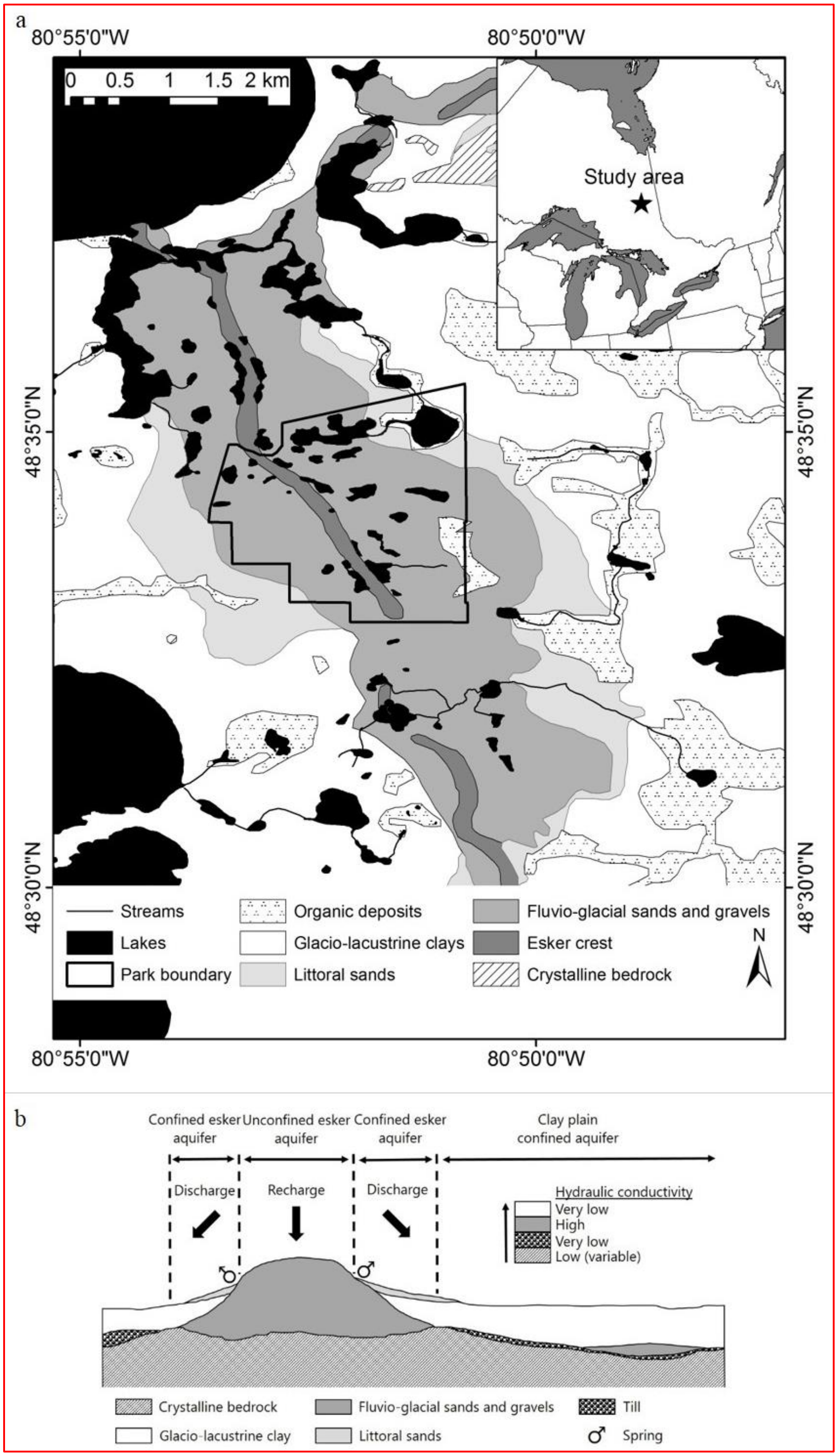

Fig. 1: Location of the study area in north east Ontario near Timmins, Ontario (Based on: Richard and McClenaghan, 2000) (a) and conceptual geological transverse section of an esker complex (b) (Modified from: Veilette et *al*., 2004)

3 Methods

3.1 Geomatic and lake morphometric data

Lakes and other geographic features were digitalized in ArcGIS 10.3 from Google Earth. Maximum lake depths Z_{max} were obtained from the Ontario Ministry of Natural Resources and Forestry (OMNRF, 2011). Digital Elevation Models (DEM) from the study area were obtained from Natural Resources Canada with a spatial resolution of 20 m (NRC, 2013). Lake elevation values were calculated as the mean elevation of the lake in ArcGIS from the available DEM, and lake watershed slopes were calculated from the DEM in ArcGIS in a 100 m buffer zone surrounding each lake. This approach was used because of the coarse spatial resolution of the DEM and the close proximity of the lakes made clear individual lake watershed delineation impossible. Since all lakes in the study area are kettles, which are characterized by steep slopes on their shore over a small distance, buffer zone of different widths were produced and the 100 m showed the best correlation with water tracers.

3.2 Field measurements and water sample collection

Water samples were collected during three field campaigns (7-14 June 2013, 16-23 June 2014, and 12-19 August 2014). 50 lakes were sampled, as well as a number of streams (lake outlets and lake inlets) and groundwater springs. The lakes were chosen to achieve a wide spatial coverage over this region encompassing a gradient of lake elevation as well as lake types. A few lakes located in the area of interest could not be sampled due to poor accessibility. Groundwater springs and streams flowing into the study lakes were also sampled to characterize the local groundwater-surface water interactions.

Temperature and specific conductance of the water (corrected to 25°C) at each site were measured with an YSI Salinity Conductivity Temperature meter (accuracy ±0.1 µS/cm and ±0.1°C). The pH was measured with an UP-5 Denver Instrument pH meter calibrated before usage (accuracy ±0.1). Water samples were collected using 1 L Nalgene bottles that were tripled-rinse with distilled water prior to use and again with sample water while sampling (Louiseize et *al*., 2014). To prevent cross-contamination, the tripled-rinsing of bottles with sampling water was carried away from the final sampling point in the lake or downstream of stream/spring sampling points. Bottles were completely filled in order to avoid headspace air and subsequently filtered within 3 hours of sampling.

Samples collected for water stable isotope, dissolved ion analysis and dissolved inorganic nitrogen were vacuum filtered with 0.45 µm Millipore Isopore filters. Vacuum filtering apparati were triple-rinse with distilled water and the lake water before and after use for each sample. The filtrates were then poured and placed into pre-cleaned 20ml plastic scintillation vials. Three replicates were collected for each sample. All vials were previously rinsed with filtered sample water that was discarded. Vials were completely filled to avoid any headspace, sealed with Parafilm™ to avoid evaporation, and kept cool and in the dark in a fridge prior to analysis.

Samples for total dissolved nitrogen (TDN), dissolved organic carbon (DOC) and dissolved inorganic carbon (DIC) analysis were filtered on-site using a glass filtration and Whatman GF/F glass fiber filters to avoid any *ex situ* carbon addition. Filters were pre-ashed at 550°C for two hours and wrapped with pre-ashed aluminum foil prior to utilization to eliminate any residual organic matter (Lamoureux and Lafrenière, 2014). The glass filtration apparatus was triple-rinsed with distilled water and lake water before and after use for each sample and was wrapped with new clean pre-ashed aluminum foil overnight. Filtrates were then poured and stored into pre-cleaned 45 ml amber EPA vials with Teflon lined septa with no headspace (Louiseize et *al*., 2014). Two replicates were collected for each sample. Vials were labelled and were kept cool and in the dark in a fridge prior to analysis.

3.3 Laboratory analysis

The stable isotope ratios of water ($\delta^{18}O$ and $\delta^{2}H$) were measured using a Los Gatos Research Liquid–Water Isotope Analyzer (LGR), which vaporizes injected sample and measures its absorbance relative to Vienna Standard Mean Ocean (‰ V-SMOW). All runs contain 6 replicate analyses and 3 standards produced by Los Gatos bracketing every 3 samples (e.g. LGR1A, $\delta^{18}O$=-19.50‰ and $\delta^{2}H$=-154.3‰; LGR2A, $\delta^{18}O$=-16.14‰ and $\delta^{2}H$=-123.6‰; LGR3A, $\delta^{18}O$=-13.10‰ and $\delta^{2}H$=-96.4‰; LGR4A, $\delta^{18}O$=-7.69‰ and $\delta^{2}H$=-51.0‰; LGR5A, $\delta^{18}O$=-2.80‰ and $\delta^{2}H$=-9.5‰). Sample reproducibility (1σ) based on repeated measurements of samples and standards was fixed at 0.25‰ for $\delta^{18}O$ and at 1.5‰ for $\delta^{2}H$.

Concentrations of inorganic ions (Ca, Mg, K, Na, Cl, SO_4) were measured by liquid ion chromatography with a Dionex ICS-3000. Detection limits were as follows: 0.5ppm for Ca, 0.01ppm for Mg, 0.2ppm for K, 0.3ppm for Na, 0.05ppm for Cl and 0.1ppm for SO_4). Concentrations of dissolved inorganic species (NO_2-NO_3 and NH_4) were measured by colorimetry using an Astoria Pacific FASPac II Flow Analyser (detection limits of 0.01ppm). Concentrations of DOC, DIC and TDN were measured by high temperature combustion and nondispersive infrared sensor and chemiluminescent detection using a

Shimadzu TOC-VPCH/TNM equipped with a high sensitivity catalyst system (detection limits of 0.08ppm for DOC and 0.015ppm for TN) (Louiseize et *al*., 2014). Total dissolved carbon (TDC) was calculated as the sum of DOC and DIC.

3.4 Water balance calculations

Monthly precipitation isotopic data are available from February, 1997 to November, 2010. Precipitation isotopic data were collected at Bonner Lake, about 100 km NNW of Timmins by the Canadian Network for Isotopes in Precipitation (CNIP) (Birks et *al*., 2010). The general water (Eq. 1) and isotope balance (Eq. 2) of a well-mixed lake may be written respectively as follow (Darling et *al*., 2005):

$$\frac{dV}{dt} = I - Q - E \qquad \text{Eq. 1}$$

$$\frac{Vd\delta_L + \delta_L dV}{dt} = I\delta_I - Q\delta_Q - E\delta_E \qquad \text{Eq. 2}$$

where V is the volume of the lake, t is time, dV is the change of volume over time dt, I is instantaneous inflow where $I = I_F + I_G + P$ (I_F being surface inflow, I_G groundwater inflow and P precipitation on the lake surface); Q is instantaneous outflow where $Q = Q_R + Q_G$ (Q_R is surface outflow and Q_G is groundwater outflow), E is evaporation; and δ_L, δ_I, δ_Q and δ_E are the isotopic compositions of the lake, inflow, outflow and evaporative flux respectively. Assuming (1) that the lake maintains a near-constant volume on the long-term (*i.e.*, $dV = 0$ and $dt \rightarrow \infty$) (Darling et *al*., 2005) and (2) that physical outflow does not cause isotopic fractionation (*i.e.*, $\delta_Q = \delta_L$) (Gibson and Edwards, 2002; Yi et *al*., 2008), Eq. 1 and 2 can be simplified and rewritten as follows:

$$I = Q + E \qquad \text{Eq. 3}$$

$$I\delta_I = Q\delta_Q + E\delta_E \qquad \text{Eq. 4}$$

E can be related to I assuming that the lakes are in isotopic steady state. This assumption seems well justified as the lakes have had sufficient time in the past to reach their isotopic steady-state which is reflective of the local climate and their mean hydrological status, and can be defined by its water balance, which corresponds to the ratio of the total inflow to the evaporation rate (Isokangas et *al*., 2015). The evaporation-to-inflow ratio of the lake E_L/I_L can be calculated by combining Eq. 3 and 4 (Gibson and Edwards, 2002; Yi et *al*., 2008):

$$\frac{E_L}{I_L} = \frac{\delta_I - \delta_L}{\delta_E - \delta_L} \qquad \text{Eq. 5}$$

where δ_I was computed as the intersection of the Local Meteoric Water Line (LMWL) with the Local Evaporation Line (LEL) (Gibson et *al*., 1993; Yi et *al*., 2008); δ_L is the isotopic composition of the lake water sample and δ_E was estimated using the Craig-Gordon model (Craig and Gordon, 1965) formulated by Gonfiantini (1986) as follows:

$$\delta_E = \frac{(\delta_L - \varepsilon^*)/\alpha^* - h\delta_A - \varepsilon_k}{1 - h + \varepsilon_k} \qquad \text{Eq. 6}$$

where δ_L is the isotopic composition of lake water, ε^* is the equilibrium isotopic separation term, α^* is the liquid–vapour equilibrium fractionation factor, h is the relative humidity, δ_A is the isotopic composition of the local atmospheric moisture, and ε_k is the kinetic separation term between the liquid and vapour phases. The ε^* and α^* parameters which are temperature dependent can be calculated using empirical equations for δ^{18}O as follows (Horita and Wesolowski, 1994):

$$\varepsilon^* = -7.685 + 6.7123\left(\frac{10^3}{T}\right) - 1.6664\left(\frac{10^6}{T^2}\right) + 0.35041\left(\frac{10^9}{T^3}\right) \qquad \text{Eq. 7}$$

$$\alpha^* = exp\left(-\frac{7.685}{10^3} + \frac{6.7123}{T} - \frac{1666.4}{T^2} + \frac{350410}{T^3}\right) \qquad \text{Eq. 8}$$

where T is the air temperature in Kelvins. ε_k (Eq. 9) is expressed for δ^{18}O by (Gonfiantini, 1986):

$$\varepsilon_k = (0.0142\,(1 - h))1000 \qquad \text{Eq. 9}$$

The equation for δ_E was modified according to Gibson and Edwards (2002) to directly utilize isotopic data in per mil rather than as a decimal fraction and expressed as follows:

$$\delta_E = \frac{\alpha^*\delta_L - h\delta_A - \varepsilon}{1 - h + 10^{-3}\varepsilon_k} \qquad \text{Eq. 10}$$

where ε is the total isotopic separation factor that includes both ε^* and ε_k expressed as:

$$\varepsilon = \varepsilon^* + \varepsilon_k \qquad \text{Eq. 11}$$

δ_A was originally estimated with the original model that assumes isotopic equilibrium between atmospheric moisture and precipitation as follows (Gibson, 2002):

$$\delta_A = \frac{\delta_P - \varepsilon^*}{1 + 10^{-3}\varepsilon^*} \qquad \text{Eq. 12}$$

where δp was computed as the average isotopic composition of annual precipitation from February 1997 to November 2010 (data collected by CNIP). The same procedure was used to calculate E/I ratios of streams and groundwater springs.

4 Results

4.1 Temperature and water stable isotopes

During the August 2014 field campaign, all types of bodies of water differed from one another by their temperature: springs are characterized by low temperatures (6 -12°C) whereas streams have slightly higher temperature (12 -16°C) and lakes ranged from (16 -18°C). It should be noted however that spring temperature is not an accurate indication of groundwater temperature due to alterations induced by the velocity of the discharging spring and the resulting warming of the discharging water as it reaches the surface.

Monthly precipitation isotopic data from CNIP from February, 1997 to November, 2010 from Bonner Lake, about 100 km NNW of Timmins, show progressive enrichment in values between winter, spring, fall and summer on the global and local meteoric water lines, which are similar (**Fig. 2a**). The water samples (*i.e.* groundwater springs, streams and lakes) displayed a wide range of water stable isotope values (-14.7‰ to -6.8‰ for δ^{18}O and -105.5‰ to -68.6‰ for δ^{2}H), which fell on a Local Evaporation Line ($r = 0.99$, $p < 0.001$) (**Fig. 2b**). Groundwater springs have isotopic values very similar to mean annual precipitation (-14.7‰ to -13.1‰ for δ^{18}O and 105.5‰ to -96.0‰ for δ^{2}H). Streams have comparable isotopic composition to groundwater springs (-13.8‰ to -11.3‰ for δ^{18}O, and -100.8‰ to -91.9‰ for δ^{2}H). By contrast, lakes are characterized by large variations in water isotopic composition, ranging from values comparable to groundwater springs (*ca*. -14‰ for δ^{18}O and -100‰ for δ^{2}H) to more enriched values (*ca*. -6‰ for δ^{18}O and -70‰ for δ^{2}H).

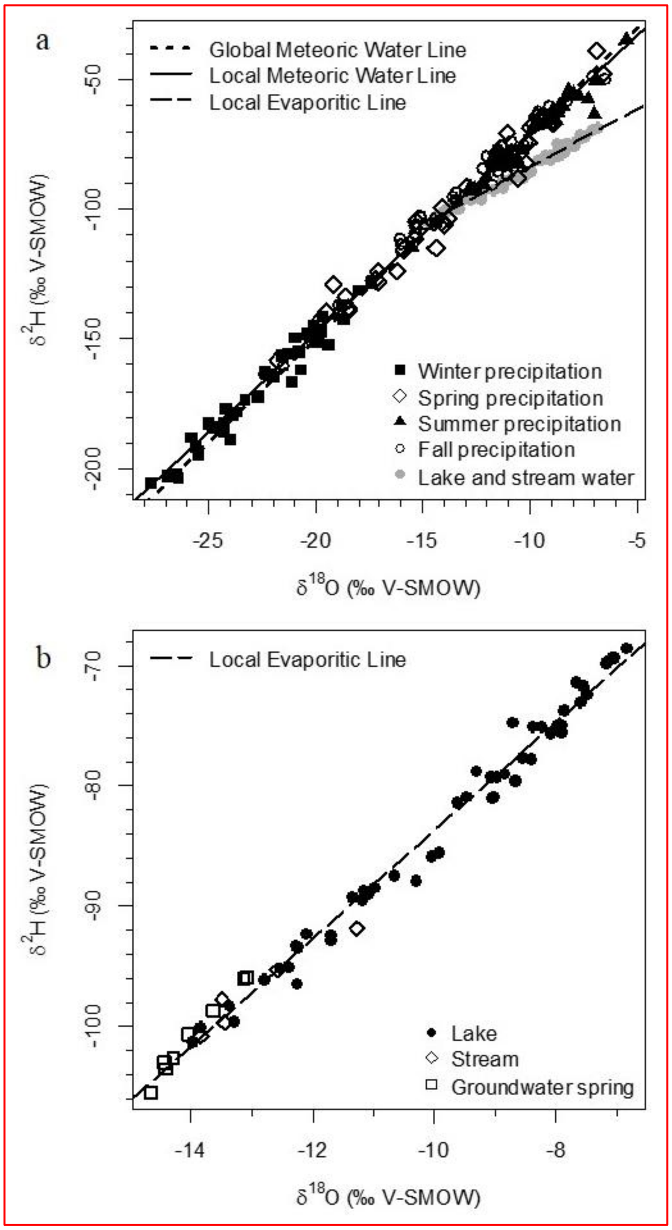

Fig. 2: Isotopic composition of precipitation in relation to the Global Meteoric Water and Local Meteoric Water lines ($r = 0.99$, $n = 166$, $p < 0.001$) (a), and isotopic composition of collected water samples in relation to the Local Evaporitic Line ($r = 0.99$, $n = 68$, $p < 0.001$) (b).

A significant correlation exists between δ^{18}O and elevation ($r = 0.53$, $n = 50$, $p < 0.001$) (**Fig. 3a**), suggesting that elevation is a variable explaining lake-water isotopic composition in lakes. Lakes sampled at lower elevations are more depleted while lakes sampled at higher elevations are enriched in ^{18}O and ^{2}H.

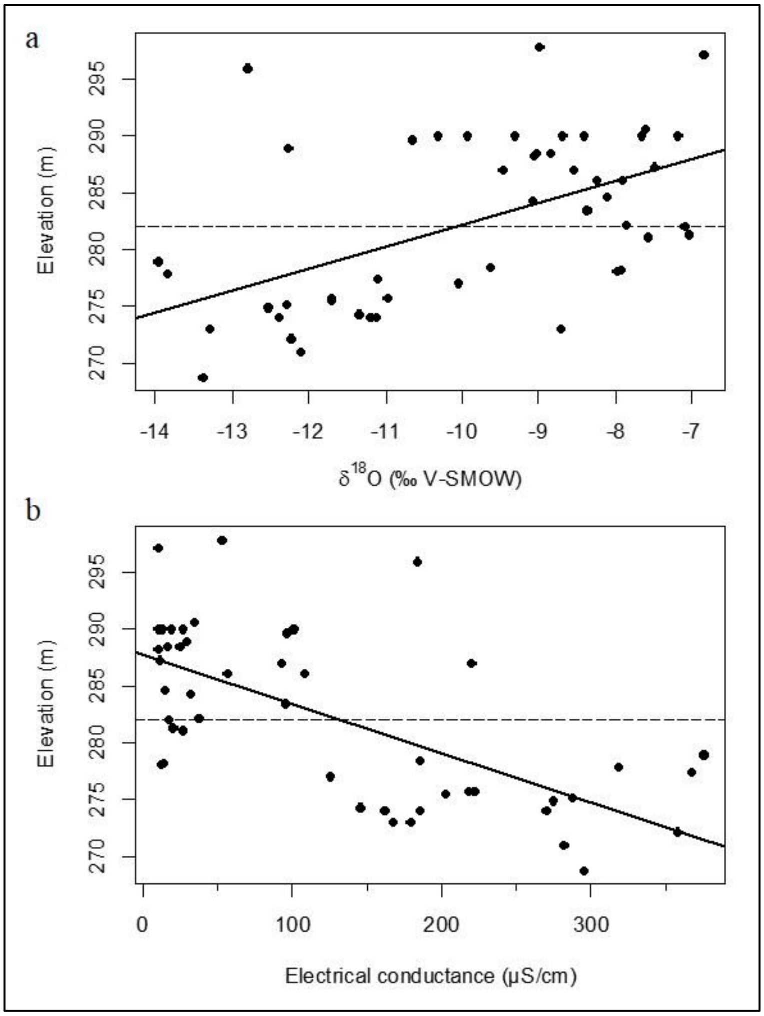

Fig. 3: Relation between δ^{18}O in water samples and elevation (r = 0.53, n = 50, $p < 0.001$) (a), and specific conductivity and elevation (r = -0.67, n = 50, $p < 0.001$) (b). The horizontal dashed line indicates the position of a breakpoint analysis which occurs at an elevation of ~282 m asl, in both cases (**Tab. 1**)

Evaporation-to-inflow ratio calculations show that groundwater springs and streams have E/I ratios close to 0 due to their short residence times while lakes have E/I values ranging from values similar to groundwater springs and streams to above E/I ~ 1 owing to their longer residence times that expose them to evaporation (**Fig. 4a**).

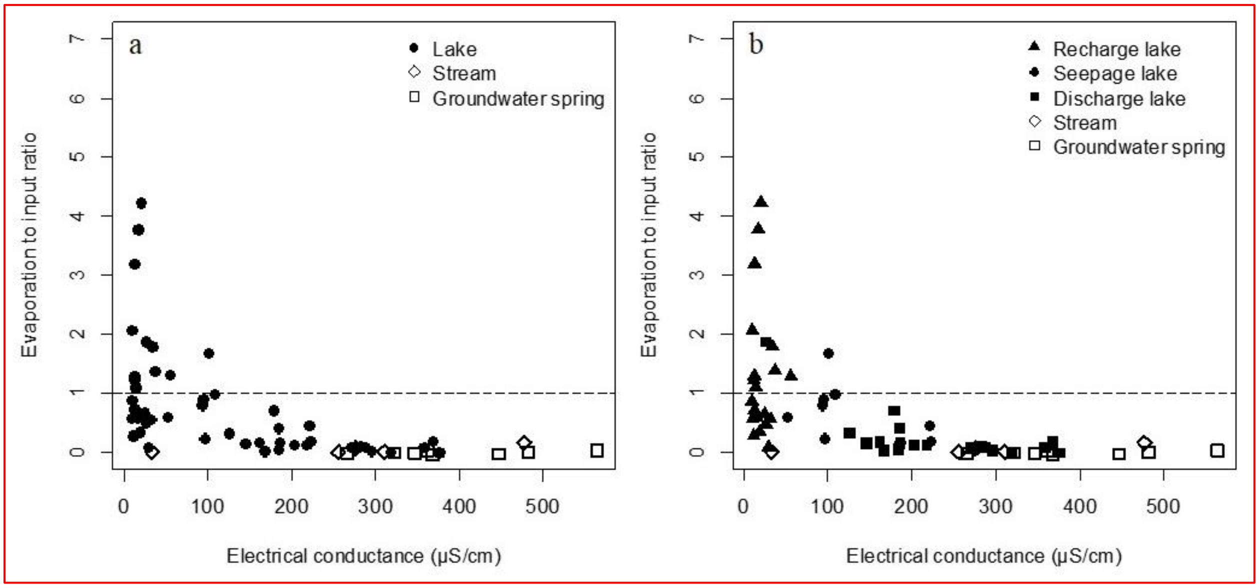

Fig. 4: Relation between electrical conductance and calculated evaporation-to-inflow ratios. Lakes in panel b are coded according to a classification scheme developed in this paper

4.2 Solutes and dissolved organic matter

As is the case with stable isotope values, water bodies reveal a wide range for non-conservative ions, and the overall sum of ions indicated by lake-water specific conductance, all of which are significantly correlated (see correlation matrix **Tab. A1**). Groundwater springs have the highest specific conductance (300-550 µS/cm) while streams have values *ca.* 300 µS/cm. Lakes show a wide range of solute content and values of specific conductance, from values similar to groundwater springs (*ca.* 400 µS/cm) to very low values similar to precipitation (as low as 10 µS/cm) (**Fig. 3b**). A significant correlation exists between the specific conductance and elevation ($r = -0.67$, $n = 50$, $p < 0.001$), suggesting that elevation is a variable explaining specific conductance in lakes.

Only 36 lakes were sampled for dissolved organic carbon (DOC) and nitrogen (TN). Unlike non-conservative ions and conservative isotopic tracers, no significant relation was found between dissolved organic elements and elevation ($r = -0.04$, $n = 36$, $p = 0.84$ for DOC; $r = 0.05$, $n = 36$, $p = 0.77$ for TN). However, significant, or marginally significant correlations were observed between DOC and mean lake depth, with deeper lakes having lower concentrations of DOC ($r = -0.58$, $n = 36$, $p = 0.11$). A similar pattern was observed between TN and mean lake depth ($r = -0.71$, $n = 36$, $p = 0.02$).

4.3 Correlations between water tracers

There is a strong and significant correlation between lake water isotopic values and specific conductance ($r = 0.80$, $n = 50$, $p < 0.001$). However, the slopes of the linear regressions for the water $\delta^{18}O$ (**Fig. 3a**) and specific conductance (**Fig. 3b**) do not match the data points perfectly as there seem to be a distinct transition between similar values found in higher- and lower-elevation lakes, which was further examined using a breakpoint analysis of the lake water properties. Nine available environmental variables had a statistically significant breakpoint when regressed over elevation and significant breakpoints were within a narrow range of elevation with a mean of 282.4 m asl (**Tab. 1**).

Tab. 1: Results showing significant breakpoints in nine water chemistry variables and lake elevation (lower and upper elevation ranges calculated from the standard deviation)

Environmental Variable	Mean Elevation	Lower Elevation	Upper Elevation
DIC	281.5	279.6	283.4
K^+	281.8	278.2	285.4
TC	281.8	280.2	283.4
Mg^{2+}	282.0	280.3	283.7
d^2H	282.0	279.9	284.1
*d**	282.2	280.3	284.1
$d^{18}O$	282.2	280.2	284.2
EC	284.0	281.4	286.6
Ca^{2+}	284.2	281.9	286.5
Breakpoint line	282.4		

**d* corresponds to deuterium excess

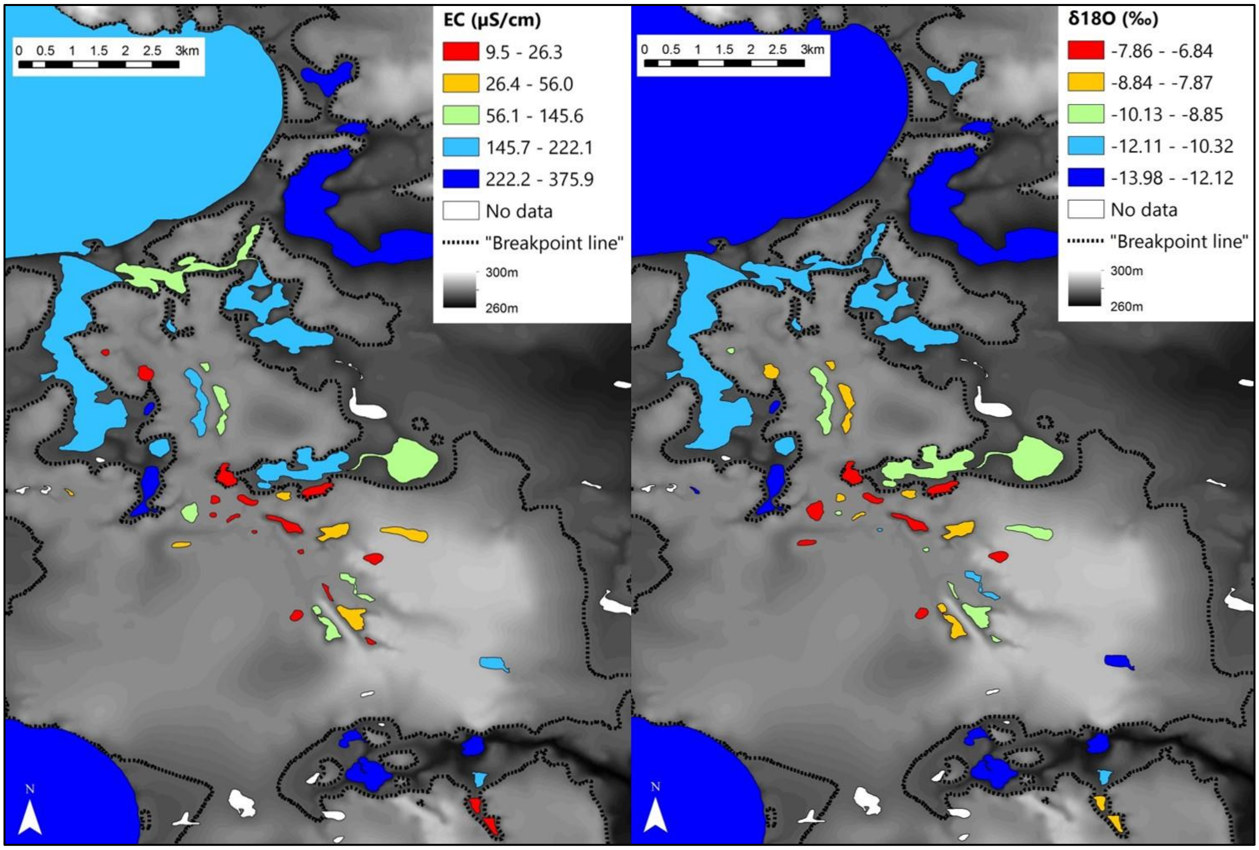

Fig. 5: Spatial depiction between elevation and lake-water specific conductance (µS/cm) (a), and lake-water $\delta^{18}O$ (‰) (b). The elevation of the breakpoint line (284 m) is shown as a dashed line

To further assess differentiation of the lakes in terms of conservative and non-conservative water tracers, a non-metric multidimensional scaling (NMDS) based on these factors was undertaken, with lakes above an elevation of 282 m coded as groundwater recharge lakes, and lakes below 282 m coded as groundwater discharge lakes, indicating good separation of the groups based on water geochemistry (**Fig. 6a**).

The existence of two distinctive types of lakes was used to develop a lake typology to explain changes in water biogeochemistry across the studied lakes. In order to better understand how water tracers vary in those two zones, individual ANOVAs were undertaken for all lake-water variables that were above detection limits (**Tab. A2**). All anion and cations (and correlated variables including specific conductance), with the exception of nitrogen species, individually were significantly different above and below an elevation of 282 m asl. This was also the case of the isotopic variables. Variables that were not significantly different included: temperature, forms of nitrogen, the C/N, and lake depth.

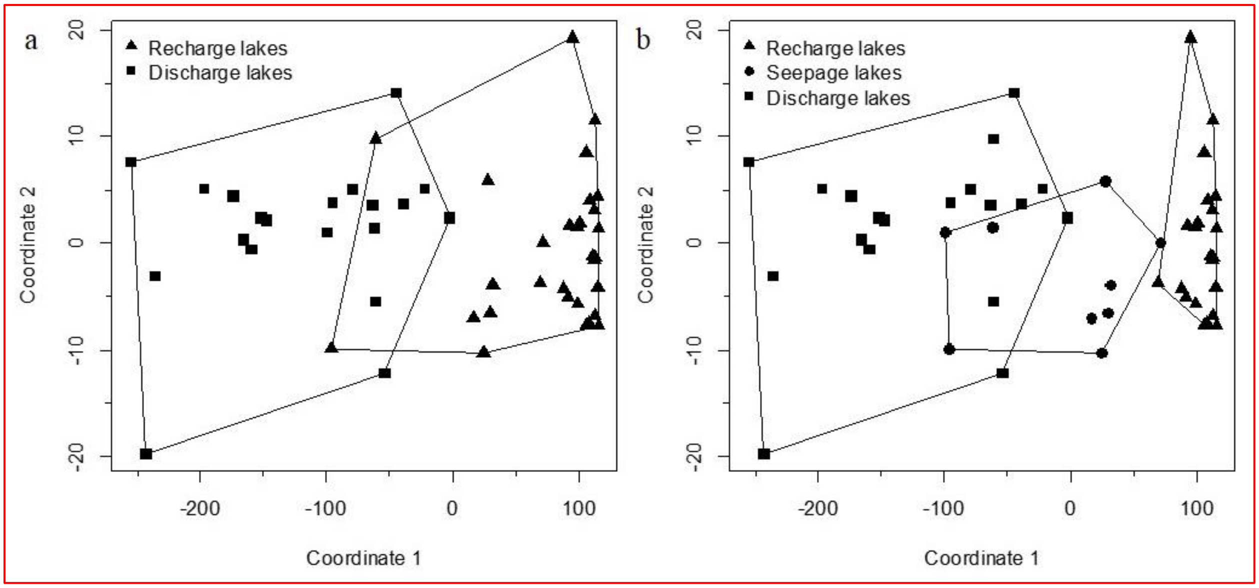

Fig. 6: Non-metric multidimensional scaling (NMDS) applied to conservative and non-conservative water tracers. Lakes that occurred at an elevation of greater than 282 m asl are labelled as 'recharge' lakes, whereas lakes as located at an elevation of less than 282 m are labelled as 'discharge' lakes (a). A third group of lakes (labelled 'seepage') is further discerned based on chemical and isotopic characteristics (b). Stress value for the NMSD is low (stress = 0.023)

4.4 Short-term evolution of water tracers

Short-term water balance variability was observed towards the end of the growing season (**Fig. A1b,d**) and between years (**Fig. A2a,c**). Lakes located at an elevation greater than 282 m asl underwent marked changes in water balance during the three sampling campaigns whereas lakes located below 282 m asl underwent little or no change, especially the lower lakes (**Fig. A2a,b**). Similarly, the water chemistry gradient between upland solute-poor lakes and lowland solute-rich lakes changed seasonally and between years. Lakes above 282 m asl did not respond chemically to short-term hydroclimatic change while lakes below 282 m asl displayed significant solute changes (**Fig. A2c,d**).

5 Discussion

5.1 Interpretation of water tracers

The wide range of lake isotopic values and E/I ratios suggest that lakes on the esker are heterogeneous in terms of water balance and hydrological characteristics. Groundwater springs have the most depleted isotopic values because their water residence times are short and they only undergo limited evaporation

(Gazis and Feng, 2004). Streams display isotopic values similar to groundwater springs since they originate from groundwater and experience relatively low evaporation rates due to more continuous water flow. The wide range of isotopic values for lakes can be explained by their position in the landscape, particularly relative to their location in the esker aquifer system for which elevation appears to be a good proxy. This suggests that lowland lakes are primarily fed by groundwater inflow while upland lakes receive a much lower contribution from the local groundwater in their respective water budgets. Nonetheless, it is challenging to distinguish the influence of groundwater from precipitation in lake water balance as they both have similar isotopic signatures (**Fig. 2b**) (Gibson and Edwards, 2002; Gibson et *al*., 2008; Yi et *al*., 2008). Yet, given that i) our study was carried out on a small spatial scale (*i.e.* a rectangular zone of ~12 km by ~6 km); ii) the close proximity of the lakes; iii) the terrain homogeneity (with boreal forest as the dominant land cover); iv) the limited topography; and v) the good correlation between water isotopes and specific conductance ($r = 0.63$, $n = 50$, $p < 0.001$), it is unlikely that there are significant differences in terms of precipitation patterns within our study area. Therefore, it is a reasonable assumption that groundwater connectivity is the main control on lake water balance. Lake isotopic values can also be influenced by the isotopic composition of surface inflowing waters. But those are for the most groundwater springs (which are made up of groundwater), their volume is supposed to be small in comparison to the lake volume and, some of those streams are intermittent in the sense that they were not flowing during each of the three field campaigns. Thus, it can reasonably be assumed that this influence is limited.

Similar patterns are observed with the major ions in water. Groundwater springs have the highest solute load likely due to chemical processes associated with mineral surface exchanges and weathering (Ala-aho et *al*., 2013). Lakes, however, displayed a wide range of water chemistries, and range from high solutes characteristic of groundwater to values close to zero, typical of precipitation (**Fig. 2**), suggesting that the heterogeneity of lakes on the esker are a result to the degree to which they interact with groundwater. As it is the case for isotopic values, chemical composition of a given lake depends on elevation. Lakes sampled at lower elevation are higher in solutes; this indicates that lowland lakes reflect interaction with intermediate or regional groundwater flows subject to more mineral weathering and dissolution (Tóth, 1963). Upland lakes on the other hand reflect interaction with local groundwater flow paths with correspondingly reduced mineral weathering and dissolution (Tóth, 1963). There are only minor differences in terms of the relative solute composition among the samples suggesting that the esker subsurface material is geochemically relatively homogeneous and reflects the carbonate-rich nature of the glaciofuvial outwash that makes up the esker (Cummings et *al*., 2011). The carbonate-rich sediment originates from Paleozoic carbonates of the Hudson sedimentary Platform *ca*. 150 km to the north and is localized to glacial surficial sediments (Roy et *al*., 2011).

Other characteristics of the lake-water (temperature and dissolved organic matter) do not reflect the changes seen in the stable isotopes and ionic concentration. Water temperatures of groundwater springs are coolest because the temperature of groundwater is typically close to the mean annual temperature of the region, while lake water temperature varies strongly with season. There is little differences of temperature between lakes and the latter are mainly the result of lake morphology than their connection with cool groundwater as lakes of smaller volumes have a lower thermal inertia than larger ones. Differences in dissolved organic content between the lakes is also a result of lake morphology as smaller lakes tend to have higher concentrations as they mix more easily and receive greater inputs due to their high catchment size-lake volume ratios (Knoll et *al*., 2015). However, upland lakes tend to have slightly higher amounts likely due to their higher water residence times.

5.2 Lake hydrological classification

The correlation between lake water δ^{18}O and specific conductance ($r = 0.80$, $n = 50$, $p < 0.001$) shows there is a clear relation between conservative water tracers (which are indicative of water source and evaporation) and non-conservative water tracers (which are indicative of water flow paths), signifying that lowland lakes will receive a significant portion of their water as groundwater flows and will geochemically reflect this origin, while upland lakes will receive most of their water through precipitation and will be geochemically dilute.

Consequently, the observed breakpoint elevation appears to correspond to an effective hydraulic separation between groundwater recharge and discharge areas (Winter et *al*., 1998). This contrast between the contribution of groundwater flow is evident in the distribution of lake-water composition (**Fig. 5**, **Fig. A1**). Lakes characterized by groundwater discharge are spatially distinct from higher elevation lakes in the groundwater recharge zone. Lake position has been used as a classification criterion in several studies (e.g. Winter, 1997; Born et *al*., 1979). Thus, upland lakes in the recharge zone are known as groundwater recharge lakes or recharge lakes and, conversely, lowland lakes in the discharge zone, also called underflow zone, will be referred to as groundwater discharge lakes or discharge lakes (**Fig. 7**) (Winter et *al*., 1998). Because discharge lakes receive a substantial amount of water from groundwater, they are considered to be groundwater-fed or minerotrophic whereas recharge lakes which receive the majority of their water from precipitation and feed the aquifer are said to be precipitation-fed or ombrotrophic (Webster et *al*., 1996). NMDS (**Fig. 6a**) and ANOVA (**Tab. A2**) analysis showed that all conservative and non-conservative tracers are statistically different between the discharge zone and the recharge zone, except for DOC, TN and temperature. An analysis of similarity (ANOSIM) between of the recharge and discharge lakes shows that within group similarity is significantly greater than between group similarity, as illustrated by a large and significant r value (r = 0.77, n = 50, significance = 0.001 on 1000 permutations).

Nevertheless, a few solute-rich alkaline lakes are located in the groundwater recharge zone which is supposed to be depleted in solutes (**Fig. 5a**). Furthermore, the two primary groups in ordinal space display a small overlap (**Fig. 6a**), suggesting the existence a third category of hybrid lakes referred to as seepage or flow-through lakes (Winter, 1976; Webster et *al*., 1996; Winter et *al*., 2003). In those lakes, water comes in as groundwater in-seepage and is returned to the groundwater system as out-seepage (Anderson and Munter, 1981). Seepage lakes can be found both in the recharge zone and the discharge zone and they can be contribute a recharge or a discharge function (Anderson and Munter, 1981). In the groundwater recharge zone, seepage lakes differ from recharge lakes by their pH, which is more alkaline while in the groundwater discharge zone, seepage lakes differ from discharge lakes by the absence of an outlet, meaning they receive less input than discharge lakes (**Fig. 8**). Lake water chemistry and the presence of outlets have been used in several studies as a classification criterion (e.g. Newton and Driscoll, 1990; Winter, 1977). NMDS (**Fig. 6b**) and ANOVA (**Tab. A3**) analysis showed that all conservative and non-conservative tracers are statistically different between the three types of lakes, except for DOC, TN and temperature.

As noted earlier, lakes can first be classified according to their location within the groundwater system, particularly above and below the hydraulic midline, in this case, at or near the 282 m elevation in this study that indicated a breakpoint in the many isotopic and limnological variable (282.4 m) associated with differences in hydrological inputs. In this study, elevation appears to be a proxy of the boundary between the groundwater recharge and discharge zones on the esker. Seepage lakes, however, can be found on each side of the hydraulic midline, thus generating additional classes of lakes (see conceptual diagram, **Fig. 7**) distinguished from on the presence or absence of a lake outlet (**Fig. A1**) (Stauffer and Wittchen, 1992). Lakes with outlets were classified as “groundwater discharge lakes” and lakes without outlets as “discharge seepage lakes”.

Lakes with outlets are typically found at the contact between the impermeable glaciolacustrine clay plain and the permeable glaciofluvial esker sand contact where groundwater springs emerge (**Fig. 1b**). Furthermore, in the study area, there is a significant relation between elevation and lake watershed maximum slope (r = 0.75, n = 50, $p < 0.001$). Lithology is also a variable used as a classification criterion in several studies (Winter, 1997; Martin et *al*., 2011). Lakes located at the edge of the esker tend to have steeper watershed slopes and have low elevations, which increases the likelihood of lakes to be in contact with deeper groundwater flows (Winter, 1976). These lakes receive a substantial inflow of groundwater due to their geological setting and their water isotopic and chemical composition is similar to the one of groundwater springs and outlets (Ala-aho et *al*., 2013).

Only two lakes in the groundwater discharge zone are without outlets and are classified as discharge seepage lakes. The following hypothesis could potentially explain the absence of outlet: the ability of the lakes to lose water to groundwater, or a reduced input from groundwater or surface inflow. Yet, some lakes with outlets, like Pit Lake, can be found at elevations corresponding to the groundwater recharge zone (**Fig. 5**). The presence of Pit Lake at a distinctly high elevation (296 m) is nonetheless explained by the location of this lake at the clay-sand interface (**Fig. 1b**). This suggests that elevation is just a putative variable or, in other words, a proxy of the actual hydraulic midline delineated by the clay-sand interface. This is confirmed numerically by applying a logistic regression to the presence or absence of a lake outlet and lake elevation, which provided a poor relation (Mc-Fadden $r = 0.40$, $n = 50$, $p < 0.01$). On the other hand, the logistic regression of the presence or absence of a lake outlet to the closest distance to the sand-clay contact provided a better relation (Mc-Fadden $r = 0.69$, $n = 50$, $p < 0.001$), although it is challenging to find the real clay-sand interface as the lateral sands can mask the real exact location (**Fig. 1b**) and surficial geological maps have errors in tens of meters in comparison to field observations.

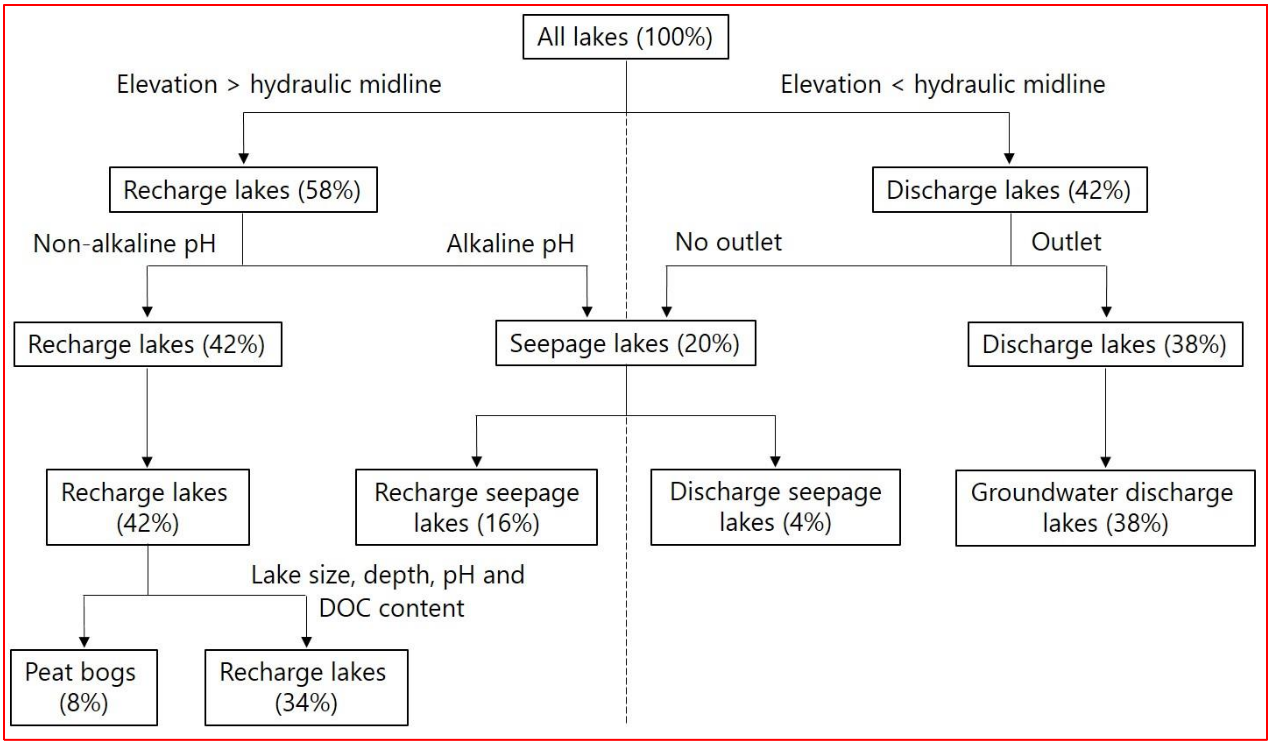

Fig. 7: Illustration of a lake typology for the study lakes. Lake elevation was used as the primary factor in separating lakes, with elevation being an effective surrogate for dividing lakes based on contribution of

groundwater flow to a lake. The lakes were then separated based on presence of an outlet, as well as lake water pH and other morphological characteristics of the lakes

Some lakes in the recharge zone, called “recharge seepage lakes”, have an alkaline or a circumneutral pH and higher solute content, signifying that they interact to some extent with groundwater. This could occur for a number of reasons related to the existence of significant groundwater input including: i) slightly steeper lake watershed slopes, ii) geographical proximity to the clay-sand interface, and iii) lake relative deepness allowing interactions with deeper groundwater flows, or a combination of those.

Recharge lakes can be further subdivided into two types of recharge lakes: “classical” recharge lakes and peat bogs (**Fig. 7**), adding a fifth type of lake in the typology (Newton and Driscoll, 1990). Sampled peat bogs are very small and shallow lakes rimmed by floating mats of vegetation (typically less than 1 ha and 1 to 2 m deep), acidic, characterized by a very low amount of solutes (maximum 30 µS/cm), a relatively high amount of dissolved organic carbon (above 10 ppm) and by water isotopic composition controlled by short-term hydroclimatic conditions (enriched in heavy water isotopes during drier periods and similar to the isotopic composition of precipitation during wet periods). This could result from relative hydrological isolation from the groundwater system due to the thick layer of peat at their bottom formed by the successive accumulation of sedge and sphagnum characterized by a low hydraulic conductivity (Newton and Driscoll, 1990). As a consequence, direct precipitation would be the predominant source of water, making these lakes sensitive to hydroclimatic variability.

A sixth type of lake can be added to the proposed lake typology: ephemeral kettle ponds, commonly called dry kettles, even though they are not lakes *per se*. They can be identified on geological maps, aerial photography and satellite imagery. Dry kettles consist of small kettle holes adjacent to the esker crest that are usually but not always too high in elevation and shallow to be connected to the water table and will be dry during the summer months and most of the year, creating patches of bare land or wetlands in the forested landscape.

Based on this typology, 42% of the sampled lakes were recharge lakes, 20% seepage lakes and 38% discharge lakes. Other studies report slightly different proportions (e.g. Anderson and Munter, 1981) as might be expected based on the geographical location, the size of the study area, and other factors relative to the groundwater system. The subdivision of lakes into recharge and discharge lakes is consistent with Ala-aho et *al*. (2013) who worked in similar settings and established their classification based on water solutes. The three-fold typology compares well with Turner et *al*. (2010) who also established a three category lake typology although the terminology of the three lake types differ and E/I values were used as the primary classification criteria. Hence, the threefold typology of **Fig. 7** is particularly relevant for water resources management in esker complexes as it uses readily available variables (*i.e.* elevation, specific conductance/pH and presence/absence of an outlet).

Discharge (ground-water discharge) lakes have the highest pH, and amounts of solutes, while the opposite is observed for recharge lakes. Seepage lakes have intermediate values compared to the other two lake types. The same patterns are observed for isotopic values: recharge lakes tend to be enriched in heavier water isotopes, discharge lakes tend to be depleted in heavier isotopes, and seepage lakes characterized by intermediate values. This reinforces the interpretation that seepage lakes are a hybrid type between recharge and discharge lakes. Seepage lakes thus tend to be closer to recharge lakes on an isotopic basis, but more similar to discharge lakes on a chemical basis. This is explained by the fact that most sampled seepage lakes are located in the recharge zone and those receive some intermediate groundwater flows.

Recharge lakes contain higher amounts of dissolved organic matter likely due to substantial water residence times, whereas seepage lakes and discharge lakes can have sustained inflow of groundwater that decreases water residence times. The concentration of DOC in discharge lakes is relatively higher than expected as groundwater usually contain limited organic matter. This is likely an artefact of lake morphology and anthropogenic activities. Indeed, values of DOC for discharge lakes display a high standard deviation due to a subgroup of smaller lakes that are relatively deep and without cottages, combined with another group of larger lakes that are relatively shallow with the presence of cottages. Relatively shallow lakes are known to have significantly higher amounts of DOC than deeper ones as a result of smaller volumes relative to inputs, and potential reworking of bottom organic matter and nutrients due to the shallow water columns. DOC values for discharge lakes also contain two sizeable outliers that skew the average and median: Nighthawk Lake and Frederick House Lake, both of which are two regional high-order and large lakes (10,701 ha and 3,888 ha respectively) that are relatively shallow (maximum depth of 4.6 m and 12.0 m respectively) and are heavily used for recreational purposes.

5.3 Lake morphometry and water geochemistry

There are additional morphometric factors that influence lake water isotopic composition in these settings. Evaporation over a water body is enhanced when there is a strong gradient of temperature and water vapour pressure between the lake water and the adjacent air. Assuming that there is no large microclimate differences within the esker complex, evaporation rates are influenced by: i) the lake fetch

(Granger and Hedstrom, 2011); ii) relative depth; and iii) the steepness of the slopes surrounding the lake. In the case of the latter, steep slopes tend to reduce evaporation rates by blocking air flows over the lake, thus reducing wind speed, water-air temperature and water vapour pressure contrasts.

Lake-water chemistry can also be influenced by lake morphometry. Lake depth has been known to play a role (Winter, 1976). It seems that *a priori* there is no relation between lake maximum depth and specific conductance ($r = 0.00$, $n = 50$, $p = 0.64$). However, to determine whether the relation between specific conductance and lake maximum depth varied with lake landscape position, two separate regressions analyses were carried out for recharge ($r = 0.01$, $n = 50$, $p = 0.66$) and discharge lakes ($r = 0.30$, $n = 50$, $p = 0.05$). The two regressions show that maximum depth can acts as a control on specific conductance only in lakes in the discharge zone: the deeper the lake, the deeper solutes-rich groundwater flows seep into the lake. It is also expected that small lakes with a large ratio of perimeter to surface area ($r = 0.24$, $n = 48$, $p < 0.001$) will have a higher amounts of solutes as the majority of groundwater seepage into the lake is typically localised near the shoreline (Rosenberry et *al*., 2015).

Given this, it can be argued that lake morphometry can play a significant role on lake-water geochemistry and introduce some complexity in the lake typology. As a fact, some lake typologies incorporate morphometric features as defining characteristics (e.g. Knoll et *al*., 2015; Martin et *al*., 2011; Winter, 1977)). Nevertheless, water tracers show a stronger correlation with elevation than with morphometric features, implying that elevation accounts for much more variance in isotopic composition and specific conductance and trumps the influence of variations in evaporation between lake basins.

5.4 Implications for water balance, hydrochemistry and response to external stressors

Results from the study show that the water balance of individual lakes is highly dependent on the nature of their interactions with groundwater that is in turn determined by landscape position at highly localized scale. Discharge lakes undergo minimal changes in water balance (usually E/I ~ 0) as the continuous flow of isotopically light groundwater masks evaporative enrichment whereas recharge lakes are highly sensitive to evaporation (in most cases $E/I > 1$) as they rely on precipitation as their primary source of water. Seepage lakes constitute a hybrid between recharge and discharge and their water balance is between the other two lake types (typically $0 < E/I < 1$) (**Fig. 4b**). Consequently, upland recharge lakes will be more prone to evaporative drawdown and therefore more sensitive to short-term climate change and droughts, while discharge lakes will be buffered by groundwater inflow and affected by hydroclimatological changes of greater duration and persistence that alter water table position (**Fig. 8**).

Seepage lakes will presumably be sensitive to drought but not to the same extent as recharge lakes as they are have stronger interactions with groundwater (**Fig. 8**). Under extreme hydroclimatic conditions, seepage lakes can become recharge lakes or discharge lakes (Anderson and Munter, 1981), suggesting that the esker hydrological system can respond to a large-scale hydroclimatic forcing (e.g. prolonged drought) in a manner that affects individual lakes differently. The degree of interaction with groundwater by an individual lake will also dictate the response to strong hydroclimatic forcings and introduce time lags (Webster et *al*., 2000).

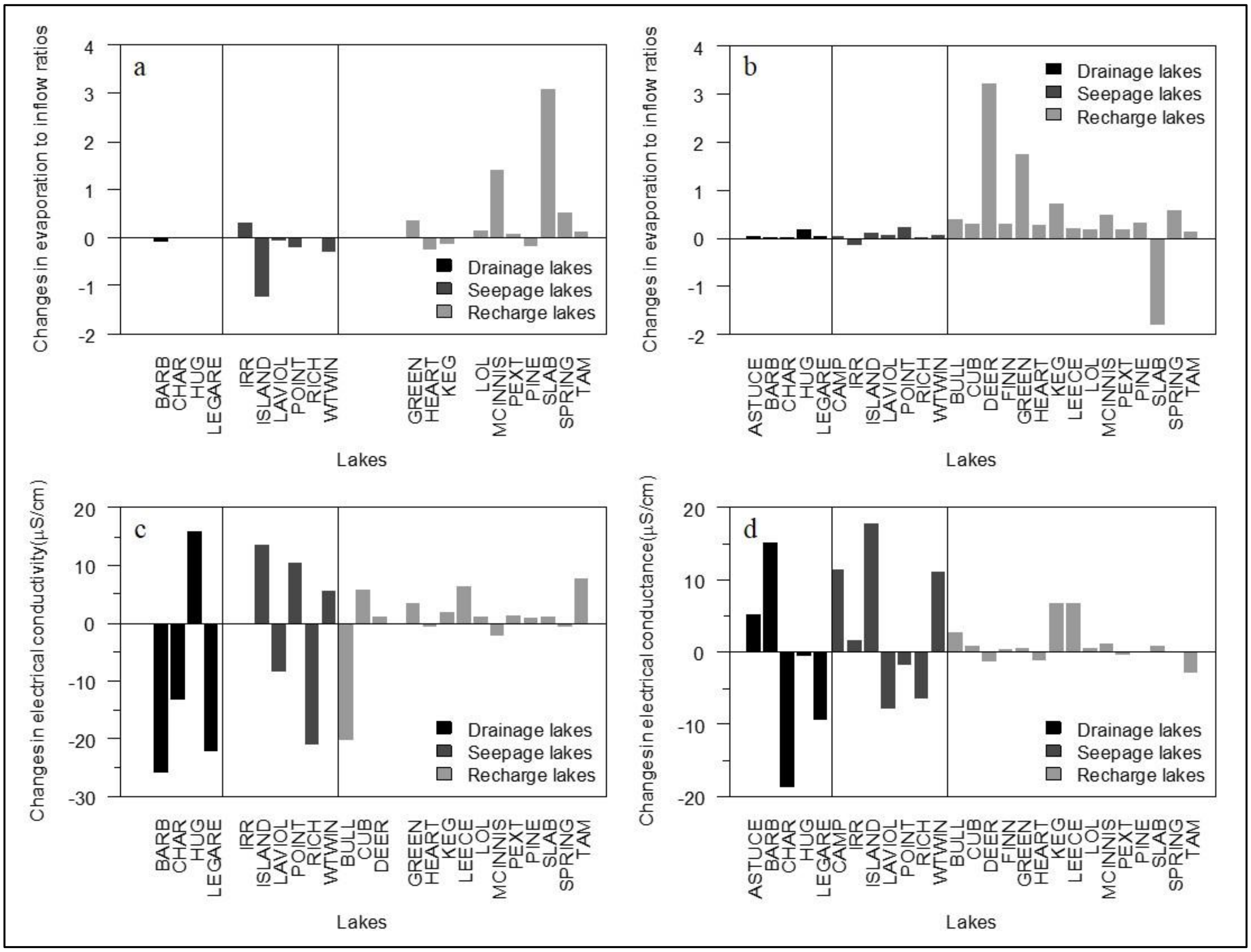

Fig. 8: Changes in E/I between June 2013 and June 2014 (E/I 2014 – E/I 2013) (a) and changes in E/I between August 2014 and June 2014 (E/I Aug – E/I Jun) (b), and changes in specific conductance between June 2013 and June 2014 (EC 2014 – EC 2013) (c) and changes in electrical conductance between August 2014 and June 2014 (EC Aug – EC Jun) (d) by lake type

Similarly, the water-chemistry gradient between upland solute-poor lakes and lowland solute-rich lakes could make recharge lakes more vulnerable to acidification than seepage and discharge lakes. The effects of changing hydroclimatic conditions on lake-water chemistry across the landscape observed during successive sampling in June 2013, June 2014 and August 2014 resulted in heterogeneous pattern depending on the lake type similar to the findings of Webster et *al*. (1996). Recharge lakes did not respond chemically to short-term hydroclimatic change as they feed the groundwater system by lake out-seepage thus producing no net change in solutes, while seepage and discharge lakes displayed significant solute changes. Based on the limited sampling frequency, it is difficult to draw conclusions on which of the seepage or discharge lake types undergo the most chemical variation. However, Webster et *al*. (1996) suggest that discharge lakes and seepage lakes respond chemically to evaporative drawdown in opposite ways, with seepage lakes showing a decline in solutes during droughts as inputs from groundwater diminish due to the lowering of the water table. By contrast, discharge lakes have been noted as susceptible to evaporative enrichment of solutes and increased relative contribution of solute-rich groundwater during drought periods (Kratz et *al*., 1997). For these reasons, Webster et *al*. (1996) suggested that climate change could amplify anthropogenic impacts and make lakes more vulnerable to other stressors, such as lake acidification.

Due to their varied hydrological characteristics, the lake types identified in this study will have a different susceptibility to direct anthropogenic impacts. Because recharge lakes have lower groundwater inflow, they are characterized by relatively long water residence times, making them highly vulnerable to inputs in comparison to seepage lakes and discharge lakes that tend to have a greater watershed, which results in increased flushing. However, it can be argued that differences in material permeability between the esker and the clay plain can produce the opposite effect. Sand extraction and mining activities in the groundwater discharge zone could potentially influence the water levels of upland lakes in the recharge zone (Klove et *al*., 2011). This potential was acknowledged by an impact study undertaken for an aggregate pit project in the study area, and one recommendation was the construction of an engineered frozen earth barrier to prevent ground water flow into the proposed pit in order to minimize the effects on the water table and surrounding lake levels (Cochrane, 2006).

Finally, when considering groundwater-fed lakes for the purposes of paleohydrological reconstruction (e.g. Laird et *al*., 2012), it is critical to explore their modern hydrology to be able to correctly interpret the probable causes of isotopic, chemical, and biological change and variability recorded in the sediment through time along with their potential hydroclimatic drivers. Because the degree to which lakes interact with groundwater produces differences in hydrologic response to a same hydroclimatic forcing, the interpretation of paleolimnological records can be complex (Fritz, 2000) and multiple site selection seem therefore necessary as a discharge lake may show long-term stability while a recharge lake may display significant short-term and long-term variability (Bennett et *al*., 2007).

6 Conclusion

Lakes located in an esker complex in northeast Ontario showed strong systematic and localized differences in terms of water balance and hydrochemistry, similar to other esker complexes in other settings. Results from this study indicated that elevation is a critical factor explaining water chemistry and water balance across the landscape. As eskers are structurally complex and often characterized by high hydraulic conductivity, groundwater interactions are an important component of lake water hydrology. Low elevation lakes are likely interconnected with solute-rich intermediate and regional groundwater flowpaths while upland lakes are only interacting with local solute-poor groundwater flowpaths. This threshold in water chemistry is also accompanied by strong contrasts in lake-water balance. Upland lakes tend to be isotopically-enriched and more sensitive to evaporation, while lowland lakes are more depleted and subject to groundwater inflows. Limnological variables including lake depth, DOC and forms of nitrogen were a bit related to elevation, and likely influenced by lake morphometry and watershed activities. Thus, these results confirm other studies that indicate that at the local scale, landscape position is the main control on lake water chemistry and balance.

The physical and chemical characteristics of lake water allowed the development of a lake typology that is made up of three main types of lakes: i) higher elevation groundwater recharge lakes essentially fed by precipitation and characterized by higher evaporations rates and lower amounts of solutes; ii) seepage lakes that both gain and lose water to the groundwater characterized by intermediate rates of evaporation and amounts of solutes; and iii) lower elevation groundwater discharge lakes, that are continuously fed by groundwater inflow and characterized by almost no evaporation rates and higher amounts of solutes. The obtained typology provides insights about lake vulnerability to environmental stressors, particularly short- and long-term hydroclimatic change. Recharge lakes will be more prone to evaporative drawdown and therefore more sensitive to short-term droughts, while discharge lakes will be buffered by groundwater inflow and affected by hydroclimatological changes of greater persistence. Similarly, recharge lakes will likely be more subject to other anthropogenic impacts, in comparison to discharge lakes.

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Author contributions

Maxime Boreux collected data in the field and processed samples in the lab, conducted data analysis and interpretation (with input from all authors), generated figures and tables. Scott Lamoureux provided guidance, funding, reviewed and edited the manuscript in his function as supervisor, as did Brian Cumming. The manuscript was written by Maxime Boreux with input from all authors.

Data availability

The data can be made available by contacting the corresponding author.

Competing interests

The authors declare that they have no conflict of interest.

Appendix

Tab. A1a: Correlation matrix between hydrological tracers and morphometric variables

	EC	pH	T	Ca^{2+}	Mg^{2+}	K^+	Na^+	Cl^-	SO_4^{2-}	NO_3^-	NH_4^+	$\delta^{18}O$	δ^2H
EC	**1.00**	0.73	-0.22	0.97	0.93	0.73	0.60	0.62	0.36	-0.13	-0.18	-0.80	-0.75
pH	0.73	**1.00**	-0.05	0.76	0.73	0.41	0.29	0.29	0.20	-0.15	-0.18	-0.47	-0.41
T	-0.22	-0.05	**1.00**	-0.24	-0.25	-0.32	0.06	0.05	-0.12	0.19	-0.16	0.34	0.33
Ca^{2+}	0.97	0.76	-0.24	**1.00**	0.94	0.66	0.43	0.48	0.40	-0.12	-0.18	-0.83	-0.78
Mg^{2+}	0.93	0.73	-0.25	0.94	**1.00**	0.72	0.36	0.37	0.26	-0.12	-0.18	-0.77	-0.73
K^+	0.73	0.41	-0.32	0.66	0.72	**1.00**	0.56	0.49	0.37	-0.01	0.05	-0.55	-0.51
Na^+	0.60	0.29	0.06	0.43	0.36	0.56	**1.00**	0.95	0.12	-0.06	-0.07	-0.31	-0.29
Cl^-	0.62	0.29	0.05	0.48	0.37	0.49	0.95	**1.00**	0.06	-0.06	-0.04	-0.35	-0.33
SO_4^{2-}	0.36	0.20	-0.12	0.40	0.26	0.37	0.12	0.06	**1.00**	0.03	-0.05	-0.33	-0.30
NO_3^-	-0.13	-0.15	0.19	-0.12	-0.12	-0.01	-0.06	-0.06	0.03	**1.00**	0.04	0.16	0.16
NH_4^+	-0.18	-0.18	-0.16	-0.18	-0.18	0.05	-0.07	-0.04	-0.05	0.04	**1.00**	0.14	0.16
$\delta^{18}O$	-0.80	-0.47	0.34	-0.83	-0.77	-0.55	-0.31	-0.35	-0.33	0.16	0.14	**1.00**	0.99
δ^2H	-0.75	-0.41	0.33	-0.78	-0.73	-0.51	-0.29	-0.33	-0.30	0.16	0.16	0.99	**1.00**
d	0.84	0.53	-0.34	0.87	0.82	0.59	0.33	0.37	0.35	-0.16	-0.11	-0.99	-0.96
E/I	-0.49	-0.40	0.35	-0.50	-0.46	-0.32	-0.21	-0.21	-0.15	0.12	-0.04	0.68	0.69
TC	0.96	0.75	-0.43	0.96	0.95	0.74	0.59	0.42	0.36	-0.17	-0.03	-0.88	-0.85
DOC	-0.12	-0.25	-0.05	-0.09	-0.19	0.26	0.00	-0.14	0.59	-0.03	0.50	-0.06	-0.08
DIC	0.96	0.81	-0.38	0.95	0.97	0.62	0.56	0.44	0.16	-0.16	-0.21	-0.82	-0.79
TN	-0.22	-0.30	-0.09	-0.21	-0.29	0.16	0.00	-0.09	0.44	-0.06	0.53	0.08	0.06
C/N	-0.13	-0.21	0.28	-0.10	-0.10	0.12	-0.10	-0.20	0.35	0.04	0.17	-0.07	-0.08
H	-0.67	-0.57	0.04	-0.69	-0.72	-0.55	-0.31	-0.28	-0.25	0.08	0.16	0.53	0.49
Z	0.09	0.20	0.17	0.06	0.18	0.06	-0.01	0.00	-0.18	0.08	0.01	0.01	0.05
S	0.51	0.50	-0.04	0.52	0.53	0.45	0.22	0.16	0.39	-0.10	-0.24	-0.33	-0.27
A	0.35	0.40	0.23	0.37	0.38	0.32	0.20	0.17	0.18	-0.06	-0.08	-0.24	-0.21
P	0.42	0.53	0.25	0.45	0.45	0.35	0.21	0.17	0.24	-0.07	-0.13	-0.29	-0.27
P/A	-0.48	-0.73	-0.28	-0.48	-0.49	-0.29	-0.29	-0.23	-0.40	0.00	0.21	0.10	0.02

Note: electrical conductance (EC) is expressed in µS/cm, temperature (T) in °C, dissolved ions (Ca^{2+}, Mg^{2+}, K^+, Na^+, Cl^-, SO_4^{2-}, NO_2^--NO_3^-, NH_4^+) in ppm, water stable isotopes ($\delta^{18}O$, δ^2H) and deuterium excess (*d*) in V-SMOW, total carbon (TC), dissolved organic carbon (DOC), dissolved inorganic carbon (DIC) and total nitrogen (TN) in ppm, elevation (H), lake maximum depth (Z) and perimeter (P) in m, maximum lake watershed slope (S) in % and area (A) in ha.

Tab. A1b: Correlation matrix between hydrological tracers and morphometric variables (con’t)

	d	E/I	TC	DOC	DIC	TN	C/N	H	Z	S	A	P	P/A
EC	0.84	-0.49	0.96	-0.12	0.96	-0.22	-0.13	-0.67	0.09	0.51	0.35	0.42	-0.48
pH	0.53	-0.40	0.75	-0.25	0.81	-0.30	-0.21	-0.57	0.20	0.50	0.40	0.53	-0.73
T	-0.34	0.35	-0.43	-0.05	-0.38	-0.09	0.28	0.04	0.17	-0.04	0.23	0.25	-0.28
Ca^{2+}	0.87	-0.50	0.96	-0.09	0.95	-0.21	-0.10	-0.69	0.06	0.52	0.37	0.45	-0.48
Mg^{2+}	0.82	-0.46	0.95	-0.19	0.97	-0.29	-0.10	-0.72	0.18	0.53	0.38	0.45	-0.49
K^+	0.59	-0.32	0.74	0.26	0.62	0.16	0.12	-0.55	0.06	0.45	0.32	0.35	-0.29
Na^+	0.33	-0.21	0.59	0.00	0.56	0.00	-0.10	-0.31	-0.01	0.22	0.20	0.21	-0.29
Cl^-	0.37	-0.21	0.42	-0.14	0.44	-0.09	-0.20	-0.28	0.00	0.16	0.17	0.17	-0.23
SO_4^{2-}	0.35	-0.15	0.36	0.59	0.16	0.44	0.35	-0.25	-0.18	0.39	0.18	0.24	-0.40
NO_3^-	-0.16	0.12	-0.17	-0.03	-0.16	-0.06	0.04	0.08	0.08	-0.10	-0.06	-0.07	0.00
NH_4^+	-0.11	-0.04	-0.03	0.50	-0.21	0.53	0.17	0.16	0.01	-0.24	-0.08	-0.13	0.21
$\delta^{18}O$	-0.99	0.68	-0.88	-0.06	-0.82	0.08	-0.07	0.53	0.01	-0.33	-0.24	-0.29	0.10
$\delta^{2}H$	-0.96	0.69	-0.85	-0.08	-0.79	0.06	-0.08	0.49	0.05	-0.27	-0.21	-0.27	0.02
d	**1.00**	-0.64	0.90	0.05	0.85	-0.11	0.05	-0.57	0.04	0.40	0.28	0.32	-0.20
E/I	-0.64	**1.00**	-0.60	-0.15	-0.53	-0.14	-0.03	0.39	-0.05	-0.13	-0.19	-0.23	-0.05
TC	0.90	-0.60	**1.00**	0.05	0.94	-0.05	0.01	-0.64	0.03	0.50	0.31	0.38	-0.27
DOC	0.05	-0.15	0.05	**1.00**	-0.29	0.93	0.66	-0.04	-0.29	0.08	0.00	-0.12	0.54
DIC	0.85	-0.53	0.94	-0.29	**1.00**	-0.37	-0.21	-0.60	0.12	0.45	0.29	0.39	-0.40
TN	-0.11	-0.14	-0.05	0.93	-0.37	**1.00**	0.40	0.05	-0.43	-0.03	-0.01	-0.10	0.60
C/N	0.05	-0.03	0.01	0.66	-0.21	0.40	**1.00**	-0.09	0.08	0.18	0.07	-0.03	0.03
H	-0.57	0.39	-0.64	-0.04	-0.60	0.05	-0.09	**1.00**	-0.26	-0.77	-0.46	-0.50	0.53
Z	0.04	-0.05	0.03	-0.29	0.12	-0.43	0.08	-0.26	**1.00**	0.30	-0.02	0.04	-0.40
S	0.40	-0.13	0.50	0.08	0.45	-0.03	0.18	-0.77	0.30	**1.00**	0.33	0.35	-0.52
A	0.28	-0.19	0.31	0.00	0.29	-0.01	0.07	-0.46	-0.02	0.33	**1.00**	0.91	-0.49
P	0.32	-0.23	0.38	-0.12	0.39	-0.10	-0.03	-0.50	0.04	0.35	0.91	**1.00**	-0.55
P/A	-0.20	-0.05	-0.27	0.54	-0.40	0.60	0.03	0.53	-0.40	-0.52	-0.49	-0.55	**1.00**

Note: electrical conductance (EC) is expressed in µS/cm, temperature (T) in °C, dissolved ions (Ca^{2+}, Mg^{2+}, K^+, Na^+, Cl^-, SO_4^{2-}, NO_2^--NO_3^-, NH_4^+) in ppm, water stable isotopes ($\delta^{18}O$, $\delta^{2}H$) and deuterium excess (*d*) in V-SMOW, total carbon (TC), dissolved organic carbon (DOC), dissolved inorganic carbon (DIC) and total nitrogen (TN) in ppm, elevation (H), lake maximum depth (Z) and perimeter (P) in m, maximum lake watershed slope (S) in % and area (A) in ha.

Tab. A2. Results of analysis of variance (ANOVA) of individual conservative and non-conservative hydrological tracers using the 2-class lake typology as the categorical variable. Non-significant *p*-values are indicated in bold. Lakes in the 'recharge' zone are defined as lakes above an elevation of 282 m asl, whereas lakes in the 'discharge' zone are defined as lakes located at an elevation of less than 282 m asl.

	Lakes in the recharge zone	Lakes in the discharge zone	ANOVA analysis	
Variable	Mean (SD)	Mean (SD)	F value	p-value
EC (µS/cm)	47.6 (53.5)	230.4 (88.4)	82.6	$p < 0.001$
pH	6.7 (0.9)	7.9 (0.4)	34.9	$p < 0.001$
T (°C)	17.0 (1.9)	16.5 (2.6)	**0.7**	**0.409**
Ca^{2+} (ppm)	4.8 (6.6)	29.4 (10.7)	101.0	$p < 0.001$
Mg^{2+} (ppm)	1.1 (1.4)	7.1 (3.1)	83.4	$p < 0.001$
K^{+} (ppm)	0.4 (0.2)	1.0 (0.4)	28.0	$p < 0.001$
Na^{+} (ppm)	0.8 (1.9)	5.3 (7.9)	8.9	0.005
Cl^{-} (ppm)	1.1 (3.9)	9.5 (15.0)	8.3	0.006
SO_4^{2-} (ppm)	0.9 (0.8)	6.6 (13.9)	4.8	0.034
NO_2^{-}-NO_3^{-} (ppm)	0.0 (0.3)	0.0 (0.0)	**0.9**	**0.339**
NH_4^{+} (ppm)	0.0 (0.0)	0.0 (0.0)	**2.1**	**0.159**
$\delta^{18}O$ (‰)	- 8.7 (1.4)	- 11.5 (1.6)	43.0	$p < 0.001$
$\delta^{2}H$ (‰)	- 78.2 (7.1)	- 90.6 (7.6)	35.0	$p < 0.001$
d (‰)	- 8.5 (4.5)	1.7 (5.5)	52.3	$p < 0.001$
E/I	1.4 (1.6)	0.2 (0.4)	10.4	0.002
TC (ppm)	11.4 (5.9)	30.4 (8.7)	60.2	$p < 0.001$
DOC (ppm)	5.9 (3.2)	6.2 (4.9)	**0.1**	**0.813**
DIC (ppm)	5.5 (6.4)	24.2 (10.6)	44.2	$p < 0.001$
TN (ppm)	0.5 (0.2)	0.5 (0.3)	**0.2**	**0.652**
Atomic C/N	13.2 (3.3)	13.6 (4.0)	**0.1**	**0.739**
Elevation (m)	287.7 (4.8)	275.1 (2.8)	116.5	$p < 0.001$
Maximum depth (m)	12.4 (6.2)	13.2 (6.7)	**0.2**	**0.694**
Maximum lake watershed slope (%)	9.8 (6.9)	20.5 (5.7)	33.7	$p < 0.001$

Tab. A3: Results of analysis of variance (ANOVA) of conservative and non-conservative hydrological tracers using the 3 class lake typology as the categorical variable. Non-significant *p*-values are indicated in bold. Lakes in the 'recharge' zone are defined as lakes above an elevation of 282 m asl and non-alkaline, lakes in the 'seepage' zone are defined as lakes above an elevation of 282 m asl and alkaline whereas lakes in the 'discharge' zone are defined as lakes located at an elevation of less than 282 m asl.

	Recharge lakes	Seepage lakes	Discharge lakes	ANOVA analysis	
Variable	Mean (SD)	Mean (SD)	Mean (SD)	F value	p-value
EC (µS/cm)	20.5 (11.9)	135.9 (60.8)	233.2 (92.5)	114.4	$p < 0.001$
pH	6.2 (0.7)	7.8 (0.3)	7.9 (0.4)	88.4	$p < 0.001$
T (°C)	17.1 (2.0)	16.4 (1.7)	16.5 (2.8)	**0.7**	**0.402**
Ca^{2+} (ppm)	1.4 (1.4)	16.8 (8.6)	29.5 (11.2)	130.3	$p < 0.001$
Mg^{2+} (ppm)	0.4 (0.3)	3.5 (1.7)	7.2 (3.3)	95.2	$p < 0.001$
K^{+} (ppm)	0.5 (0.3)	0.4 (0.1)	1.0 (0.4)	19.3	$p < 0.001$
Na^{+} (ppm)	0.4 (0.7)	1.9 (2.9)	5.6 (8.3)	9.2	0.004
Cl^{-} (ppm)	0.4 (1.3)	3.3 (6.5)	9.9 (15.7)	8.1	0.006
SO_4^{2-} (ppm)	0.7 (0.6)	1.7 (1.1)	7.0 (14.6)	4.7	0.035
NO_2^{-}-NO_3^{-} (ppm)	0.1 (0.4)	0.0 (0.0)	0.0 (0.0)	**0.9**	**0.339**
NH_4^{+} (ppm)	0.0 (0.0)	0.0 (0.0)	0.0 (0.00)	**2.6**	**0.114**
$\delta^{18}O$ (‰)	-8.5 (1.3)	- 9.7 (1.6)	- 11.6 (1.7)	48.0	$p < 0.001$
$\delta^{2}H$ (‰)	- 77.4 (6.8)	- 82.1 (7.8)	- 90.8 (8.0)	36.5	$p < 0.001$
d (‰)	- 9.6 (3.7)	- 4.5 (5.4)	1.9 (5.7)	64.6	$p < 0.001$
E/I	1.6 (1.8)	0.6 (0.5)	0.2 (0.4)	14.0	$p < 0.001$
TC (ppm)	8.7 (4.0)	19.9 (6.6)	30.8 (9.4)	74.7	$p < 0.001$
DOC (ppm)	6.7 (3.5)	3.9 (1.2)	6.7 (5.2)	**0.1**	**0.733**
DIC (ppm)	2.0 (1.3)	15.9 (7.2)	24.1 (11.5)	65.7	$p < 0.001$
TN (ppm)	0.6 (0.2)	0.4 (0.1)	0.5 (0.3)	**0.9**	**0.341**
Atomic C/N	14.1 (3.5)	11.1 (1.4)	14.0 (4.3)	**0.1**	**0.784**
Elevation (m)	287.0 (4.6)	286.7 (7.6)	275.1 (3.0)	37.0	$p < 0.001$
Maximum depth (m)	12.0 (6.6)	12.4 (5.7)	13.5 (6.8)	**0.4**	**0.555**
Maximum lake watershed slope (%)	10.2 (7.4)	11.3 (6.7)	19.9 (6.7)	19.1	$p < 0.001$

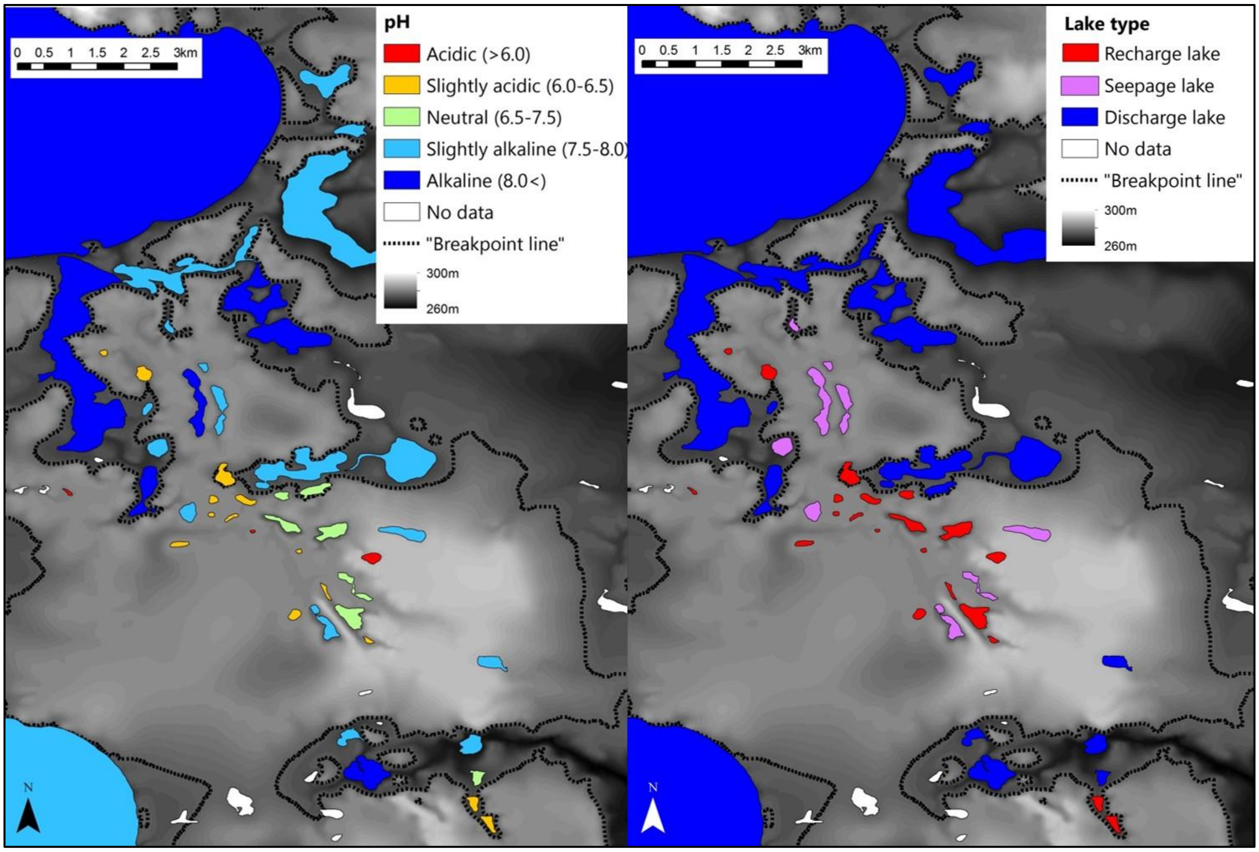

Fig. A1: Spatial depiction between elevation and lake-water pH (a) and lake type (b). The elevation of the breakpoint line (284 m) is shown as a dashed line

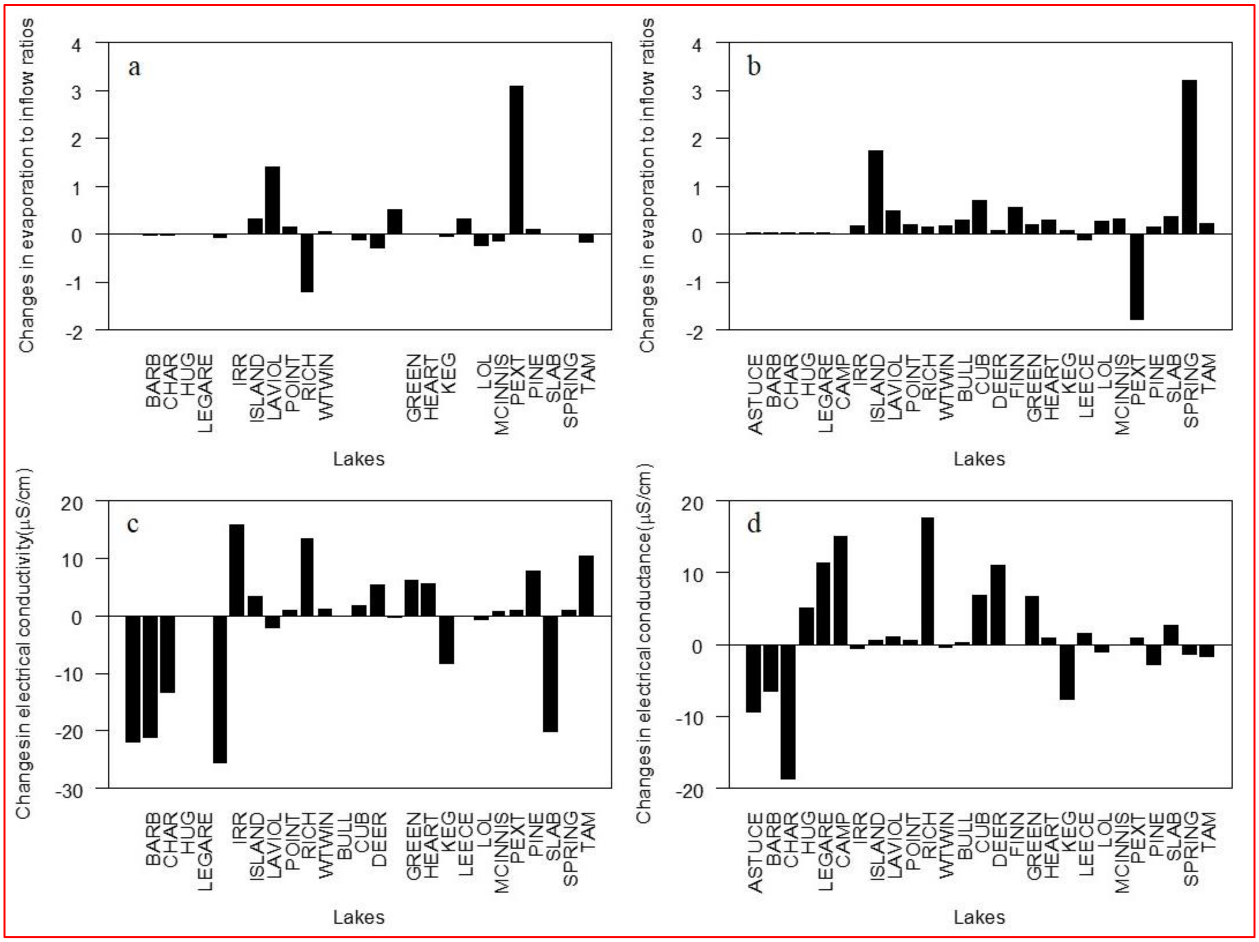

Fig. A2: Changes in E/I between June 2013 and June 2014 (E/I 2014 – E/I 2013) (a) and changes in E/I between August 2014 and June 2014 (E/I Aug – E/I Jun) (b), and changes in specific conductance between June 2013 and June 2014 (EC 2014 – EC 2013) (c) and changes in electrical conductance between August 2014 and June 2014 (EC Aug – EC Jun) (d). Lakes are organized along an elevation gradient (lower to higher elevation).