https://doi.org/10.5194/hess-2022-279
Preprint. Discussion started: 5 September 2022

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Hydrological, meteorological and watershed controls on the water balance of thermokarst lakes between Inuvik and Tuktoyaktuk, Northwest Territories, Canada

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Abstract. Thermokarst lake water balances are becoming increasingly vulnerable to change in the Arctic as air temperature increases and precipitation patterns shift. In the tundra uplands east of the Mackenzie Delta in the Northwest Territories, Canada, previous research has found that lakes responded non-uniformly to changes in precipitation, suggesting that lake and watershed properties moderate the response of lakes to climate change. To investigate how lake and watershed properties and meteorological conditions influence the water balance of thermokarst lakes in this region, we sampled 25 lakes for isotope analysis five times in 2018, beginning before snowmelt on May 1 and ending on September 3. Water isotope data were used to calculate the ratio of evaporation-to-inflow (E/I) and the average isotope composition of lake source water (δ_I). We identified four distinct water balance phases as lakes responded to seasonal shifts in meteorological conditions and hydrological processes. During the freshet phase from May 1 to June 15, the median E/I ratio of lakes decreased from 0.20 to 0.13 in response to freshet runoff and limited evaporation due to lake ice presence that persisted for the duration of this phase. During the following warm, dry, and ice-free period from June 15 to July 26, designated the evaporation phase, the median E/I ratio increased to 0.19. During the brief soil wetting phase, E/I ratios did not respond to rainfall between July 26 and August 2, likely because watershed soils absorbed most of the precipitation which resulted in minimal runoff to lakes. The median E/I ratio decreased to 0.11 after an unseasonably cool and rainy August, identified as the recharge phase. Throughout the sampling period, δ_I remained relatively stable and most lakes contained a greater amount of rainfall-sourced water than snow-sourced water, even after the freshet phase due to snowmelt bypass. The range of average E/I ratios we observed at lakes (0.00 – 0.43) was relatively narrow and low compared to thermokarst lakes in other regions, likely owing to the large watershed area to lake area (WA/LA), efficient preferential flow pathways for runoff, and a shorter ice-free season. WA/LA strongly predicted average lake E/I ratio (R^2 = 0.74), as lakes with smaller WA/LA tended to have higher E/I ratios because they received relatively less inflow. We used this relationship to predict the average E/I ratio of 7340 lakes in the region, finding that lakes are not vulnerable to desiccation in this region, given that all predicted average E/I values were <0.33. If future permafrost thaw and warming cause less runoff to flow into lakes, we expect that lakes with smaller WA/LA will be more influenced by increasing evaporation, while lakes with larger WA/LA will be more resistant to lake-level drawdown. However under wetter conditions, lakes with larger WA/LA will likely experience greater increases in lake level and could be more susceptible to rapid drainage as a result.

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1 Introduction

Thermokarst lakes are common features in ice-rich permafrost terrain, occupying up to 25% of the land area (Woo, 2012). The water balance of thermokarst lakes are changing as arctic warming causes precipitation to shift from snowfall to rainfall (Bintanja and Andry, 2017; MacDonald et al., 2021), permafrost thaw alters the hydrological connectivity of lake watersheds (Wolfe et al., 2011; Walvoord and Kurylyk, 2016; Tananaev and Lotsari, 2022; Koch et al., 2022), longer ice-free periods increase evaporation (Prowse et al., 2011; Arp et al., 2015), and increased vegetation growth changes snow redistribution and snowmelt timing (Sturm et al., 2001; Essery and Pomeroy, 2004; Pomeroy et al., 2006). In many regions, expansion and contraction of thermokarst lakes has been observed demonstrating that lake water balances are not responding uniformly to climate change (Smith et al., 2005; Plug et al., 2008; Marsh et al., 2009; Arp et al., 2011; Jones et al., 2011; Andresen and Lougheed, 2015; Travers-Smith et al., 2021).

Previous studies have demonstrated that knowledge of meteorological conditions and lake and watershed attributes, and their influence on lake water balance, can explain why lakes react non-uniformly to climate change (Brock et al., 2009; Wolfe et al., 2011; Turner et al., 2014; Nitze et al., 2017; Wan et al., 2020). Key drivers of lake water balances include the relative size of a lake within its watershed (Watershed Area/Lake Area, WA/LA), rainfall and snowfall patterns, permafrost dynamics, wildfire, vegetation cover, and ice-free season length (Turner et al., 2014; Arp et al., 2015; MacDonald et al., 2017; Wan et al., 2020). For example, Turner et al. (2014) found that thermokarst lakes in Old Crow Flats, Yukon, with increased evaporation-inflow ratios (E/I) tended to have smaller WA/LA ratios, reflecting the control of watershed area on the amount of inflow a lake receives. Arp et al. (2015) found that lakes in the Alaskan Coastal Plain with bedfast ice became ice-free sooner than lakes with floating ice, causing bedfast lakes to lose more water to evaporation as a result of a longer ice-free season.

In several studies, water isotope (^{18}O and ^{2}H) analysis has been the primary method used to efficiently characterize the water balance of a large number of thermokarst lakes because the isotope composition provides an integrated measure of influential hydrological processes (Edwards et al., 2004). Two key metrics of lake water balance modelled using water isotope data include the average isotope composition of lake source waters (δ_{I}), which can then be related to the measured isotope composition of different water sources (Yi et al., 2008) and the ratio of evaporation to inflow (E/I) (Gibson et al., 1993). Sampling lake water for isotope analysis multiple times throughout the year provides data to assess the response of lake water balances to different hydrological processes, which can be related to meteorological conditions and compared to lake and watershed attributes. Recently, MacDonald et al. (2017) compared E/I and δ_{I} of thermokarst lakes from six regions in northern North America and found a wide range of water balances that were influenced by local meteorological conditions, permafrost extent and vegetation characteristics. In some locations, such as the Alaskan Coastal Plain, E/I ratios were mostly below 0.2, whereas lakes in the Hudson Bay Lowlands of northeastern Manitoba typically possessed E/I ratios >0.5, with some lakes trending towards desiccation.

This study aims to evaluate how meteorological conditions, hydrological processes, and lake and watershed attributes influence the δ_{I} and E/I of lakes located in the tundra uplands east of the Mackenzie Delta in the Northwest Territories, Canada (Figure 1). A previous study in this region showed that only 25% of pre-snowmelt lake water is replaced by freshet on aver-

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age, because snowmelt bypass leads to minimal mixing between freshet and pre-snowmelt lake water (Bergmann and Welch, 1985; Wilcox et al., under rev.). With this study we explore whether snowmelt bypass influences summertime lake water balances, given that in other regions thermokarst lakes are prone to desiccation when they retain little snowmelt (Bouchard et al., 2013). We also investigate how watershed characteristics that have been identified to influence lake water balance in other regions, such as WA/LA, influence lake water balances. We then use the relationships found between average E/I and watershed characteristics to predict average E/I for all lakes in the region and assess their vulnerability to climate change.

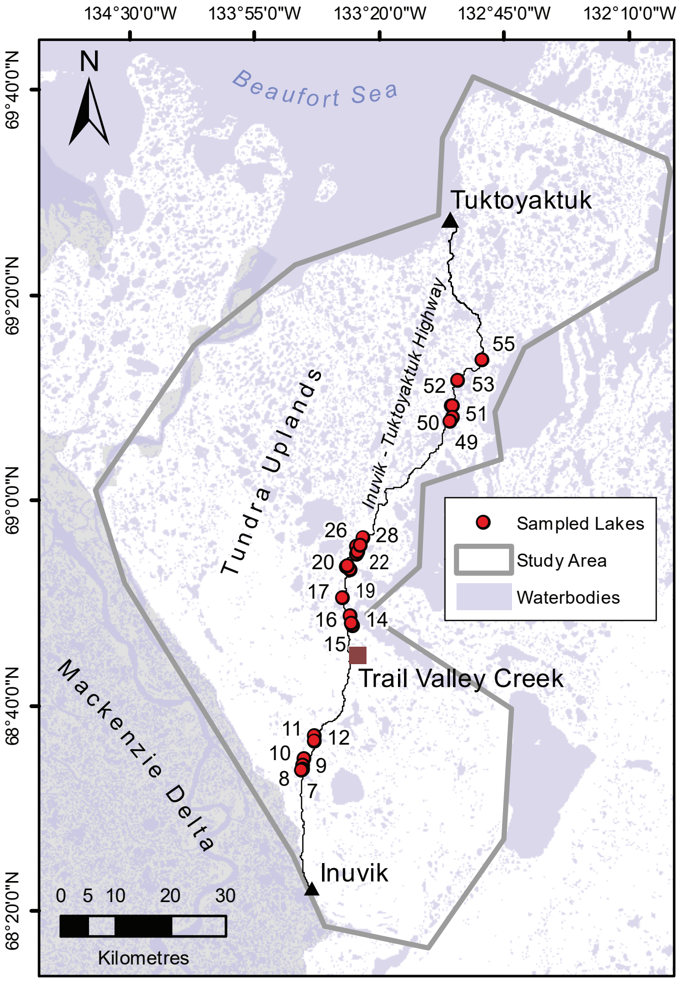

Figure 1. Study area showing locations of thermokarst lakes where water samples for isotope analysis were taken. The locations of the meteorological station in Inuvik and the Trail Valley Creek research station are also shown.

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2 Study Area

The thermokarst lakes in this study are located in the tundra uplands east of the Mackenzie Delta in the northwest region of the Northwest Territories, Canada (Figure 1). The landscape is comprised of rolling hills and has been shaped by the thaw of ice-rich permafrost, evidenced by the thousands of thermokarst lakes in the region (Rampton, 1988; Burn and Kokelj, 2009). Hillslopes are well-drained by the network of peat channels that facilitate subsurface flow between mineral earth hummocks (Quinton and Marsh, 1998), while flatter areas are typically drained by high-centred ice-wedge polygons (Burn and Kokelj, 2009). Vegetation consists of tall shrub (>1 m), low-shrub (~0.5 m), and shrub-free landcover types containing lichen, moss, and tussocks (Lantz et al., 2010; Grünberg et al., 2020).

We selected 25 lakes to cover a range of watershed sizes, surface areas, among other characteristics along a ~70 km stretch of the Inuvik - Tuktoyaktuk Highway (Figure 1). All lakes we selected were either headwater lakes or downstream of lakes that we sampled. The average area of the studied lakes is 10.43 ha (0.37 to 90.48 ha) and the average maximum depth is 2.28 m (1.02 to 4.14 m) (Table 1). Twenty of the lakes are headwater lakes and the other five lakes are downstream of other sampled lakes (Table 1). All lakes have a defined outlet channel and many lakes have defined channelized inflows from their watersheds in the form of small ephemeral streams or ice-wedge polygon troughs.

3 Methods

3.1 Lake water and precipitation sampling and analysis of isotope data

Water samples from the 25 thermokarst lakes were collected on five days (May 1, June 15, July 26, August 2, September 3) during 2018 for water isotope analysis. Samples were collected on May 1 through a hole augured through the ice in the centre of each lake to capture the pre-snowmelt lake water balance conditions. The May 1 samples also represent the hydrological status of lakes during the freeze-up period of 2017, since virtually no hydrological activity occurs during the winter months due to the complete freezing of the soil and lake surface. Because May 1 water samples were influenced by fractionation that occurred during lake ice formation, the isotope compositions were corrected to represent the isotope composition of the lake before lake ice formation took place (Wilcox et al., under rev.). June 15 was chosen as it marked the first day of the ice-free season for most lakes and was intended to capture the influence of snowmelt bypass, a phenomenon where snowmelt runoff to flows underneath lake ice and out of lakes without mixing with pre-snowmelt lake waters (Bergmann and Welch, 1985). More information about ice-correction and the response of lakes to snowmelt bypass is provided by Wilcox et al. (under rev.). Next, samples were obtained on July 26 to capture the effects of evaporation and rainfall in the early open-water season, and before a large, forecasted rainfall event occurred. Samples were then collected on August 2 to capture the influence of the week-long rainy period following the sampling date on July 26. Samples were lastly taken on September 3 to assess lake water balance response to the relatively rainy and cool August weather prior to freeze-up. The four time periods between the five sampling dates are identified as “P1” (May 1 to Jun. 15), “P2” (Jun. 15 to Jul. 26), “P3” (Jul. 26 to Aug. 2), and “P4” (Aug. 2 to Sept. 3).

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Table 1. Lake and watershed attributes and isotope derived hydrological indicators for all lakes.

Lake	Latitude (dec deg.)	Lake Depth (m)	Lake Elevation (m asl)	Polygon Extent (% Watershed)	Watershed Slope (°)	Lake Area (ha)	Watershed Area (ha)	WA/LA	Headwater Lake?	Upstream Lake(s)	Drainage Density (m ha^{-1})	E/I Average	δ_I Average (‰ $\delta^{18}O_I$)
7	68.557	2.24	89.0	0.00	1.30	2.94	6.45	2.62	yes	none	0.00	0.28	-19.82
8	68.559	2.30	88.9	0.00	2.63	2.07	15.67	4.41	no	7	0.00	0.21	-20.58
9	68.564	1.02	86.4	29.84	1.20	59.56	203.56	3.77	no	7,8,10	22.28	0.40	-18.19
10	68.576	1.65	87.9	5.17	1.42	90.48	168.58	1.86	yes	none	7.40	0.35	-18.85
11	68.604	1.91	82.7	2.53	2.10	1.26	21.76	17.32	yes	none	12.68	0.04	-19.84
12	68.613	2.79	90.8	0.00	2.45	8.19	23.05	2.81	yes	none	0.51	0.43	-21.84
14	68.789	1.42	51.9	8.07	3.01	10.65	60.64	5.70	yes	none	1.24	0.09	-20.35
15	68.795	1.57	57.2	1.62	2.73	5.98	29.83	4.99	yes	none	0.00	0.17	-18.81
16	68.805	3.18	51.5	12.57	4.04	1.23	19.75	16.02	yes	none	4.86	0.04	-19.65
17	68.836	1.09	39.1	8.23	4.38	3.27	39.88	12.20	yes	none	11.53	0.08	-20.21
19	68.882	2.46	39.2	4.91	4.27	5.50	38.98	7.09	yes	none	5.79	0.10	-19.69
20	68.885	2.69	36.7	2.76	4.69	2.17	19.93	9.18	yes	none	3.04	0.10	-19.33
21	68.887	1.78	35.5	1.87	5.33	2.75	10.91	10.31	no	20	6.86	0.16	-20.07
22	68.907	3.66	33.6	4.11	5.33	3.41	23.67	18.25	no	23,24	6.12	0.12	-19.75
23	68.910	3.02	34.9	1.22	5.35	1.83	22.73	20.40	no	24	4.88	0.04	-19.22
24	68.911	3.02	38.3	4.49	4.69	0.37	11.10	30.10	yes	none	9.69	-0.01	-19.15
26	68.918	1.47	38.4	4.71	4.69	4.68	17.89	3.83	yes	none	0.00	0.25	-18.45
27	68.921	3.10	45.4	0.00	6.21	1.28	8.57	6.70	yes	none	0.00	0.17	-20.52
28	68.933	2.24	24.1	9.58	4.66	2.32	92.08	39.69	yes	none	22.73	0.01	-19.28
49	69.119	2.18	8.6	4.06	3.18	18.20	46.23	2.54	yes	none	0.79	0.31	-17.45
50	69.123	1.65	8.3	2.92	3.85	8.70	31.92	3.67	yes	none	0.00	0.23	-17.53
51	69.142	2.31	3.8	0.00	4.90	2.29	12.01	5.26	yes	none	0.00	0.19	-18.46
52	69.144	4.14	6.3	1.26	3.93	24.32	49.92	2.05	yes	none	0.00	0.32	-17.01
53	69.184	2.29	4.8	8.71	6.54	0.46	11.21	24.21	yes	none	10.05	0.13	-19.38
55	69.215	1.83	2.3	0.00	2.12	1.61	8.11	5.03	yes	none	0.00	0.18	-17.71
Min	68.557	1.02	2.3	0.00	1.2	0.37	6.45	1.86	n/a	n/a	0.00	-0.01	-21.84
Avg.	68.871	2.28	43.4	4.75	3.8	10.43	39.78	10.40	n/a	n/a	5.218	0.16	-19.25
Max	69.215	4.14	90.8	29.84	6.54	90.48	203.56	39.69	n/a	n/a	22.73	0.43	-17.01

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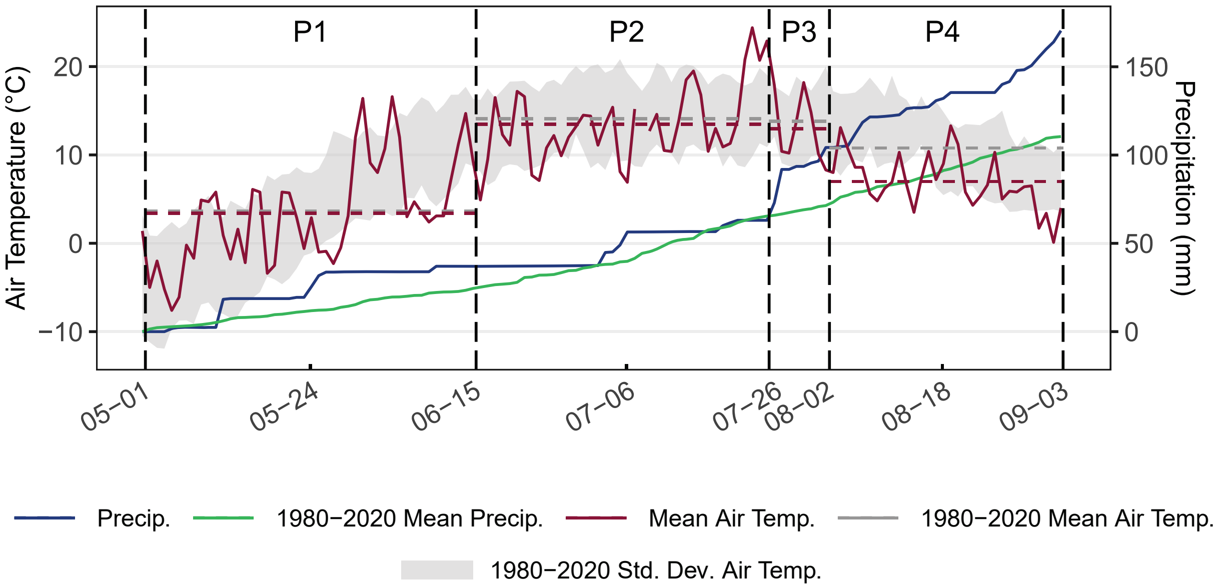

Figure 2. Air temperature and precipitation during the sampling period and the 1980-2020 mean, as recorded at Inuvik (WMO ID: 71364) (Figure 1). The five days when water samples were taken from lakes are shown by the vertical dotted lines. The periods between the sampling dates are labeled P1, P2, P3, and P4. Mean daily air temperature for 2018 and 1980-2020 for each period is indicated by the dashed lines.

From June 15 onward, water samples were collected at the shoreline of each lake where water was freely circulating. Samples of end-of-winter snow and open-water season rainfall were obtained to determine the isotope composition of rain (δ_R) and snow (δ_S). Snow samples (n = 11) were collected from the region by taking a vertical core of snow using a tube, completely melting the snow in a sealed plastic bag, and then filling a sampling bottle with the melted snow. Rainfall (n = 13) was collected between May and September in Inuvik using a clean high density polyethylene container, which was then transferred into a sample bottle shortly after precipitation ceased. All samples were stored in 30 mL high-density polyethylene bottles. A Los Gatos Research (LGR) Liquid Water Isotope Analyser, model T-LWIA-45-EP was used to measure the ratio of $^{18}O/^{16}O$ and $^{2}H/^{1}H$ in each sample at the Environmental Isotope Laboratory at the University of Waterloo. Vienna Mean Standard Ocean Water (VSMOW) and Vienna Standard Light Antarctic Precipitation (VSLAP) standards provided by LGR were used to calibrate the instrument. VSMOW and VSLAP standards were used during the analysis to check the calibration of the instrument. Isotope compositions are expressed in standard δ-notation, such that:

$$\delta_{sample} = \frac{R_{sample}}{R_{VSMOW}} - 1 * 10^3 \quad (1)$$

where R represents the ratio of $^{18}O/^{16}O$ or $^{2}H/^{1}H$, and VSMOW represents Vienna Mean Standard Ocean Water. Isotope values are normalized for Standard Light Antarctic Precipitation to $\delta^{18}O$ = -55.5 ‰ and $\delta^{2}H$ = -428 ‰ (Coplen, 1996). Every fifth sample was run through the analyzer a twice to determine the analytical uncertainties, which were ±0.1‰ for $\delta^{18}O$ and ±0.6‰ for $\delta^{2}H$, calculated as two standard deviations away from the difference between the duplicate samples.

3.2 Isotope framework, δ_{I} and E/I

The isotope data were initially described and interpreted with respect to an 'isotope framework' that consists of two fundamental linear relationships that form when isotope data is plotted with δ^{18}O on the x-axis and δ^{2}H on the y-axis ('δ-δ space'). The Local Meteoric Water Line (LMWL) represents the expected isotope composition of precipitation in the region and was estimated using a linear regression through δ_{S} and δ_{R} values, with δ_{P} representing the average of δ_{S} and δ_{R}. Evaporated waters tend to plot along a Local Evaporation Line (LEL), which can be defined independent of measured lake water isotope compositions. Using this approach and for the case of a lake fed by waters of mean annual isotope composition of precipitation, the LEL was anchored at δ_{P} and the theoretical isotope composition of a lake at the point of desiccation (δ^{*}), which is dependent on air temperature, relative humidity and the isotope composition of atmospheric moisture in the region (Gonfiantini, 1986). Along the LEL exists a useful reference point where the amount of evaporation a water body is experiencing is equal to the amount of water input, defined as δ_{SSL}. When lake water isotope compositions are plotted in δ-δ space, they typically plot near the LEL, with more evaporated lakes plotting closer to δ^{*} and less evaporated lakes plotting closer to δ_{P}. Lakes preferentially sourced by rainfall will tend to plot above the LEL, whereas lakes preferentially influenced by snowmelt will tend to plot below the LEL.

Isotope compositions of the lakes were used to estimate the average isotope composition of source water (δ_{I}) and the ratio of evaporation-to-inflow (E/I), following the coupled isotope tracer method introduced by Yi et al. (2008) which has been applied in a variety of northern environments (Turner et al., 2014; MacDonald et al., 2017; Remmer et al., 2020). E/I was calculated as:

$$E/I = \frac{\delta_I - \delta_L}{\delta_E - \delta_L} \tag{2}$$

where δ_{E} is the isotope composition of evaporated vapour from the lake (Gonfiantini, 1986). More detailed information about the calculation of the isotope framework components, δ_{I} and E/I is provided in Appendix A.

3.3 Meteorological conditions and lake and watershed attributes

The end of winter snowpack at Trail Valley Creek contained 141 mm of snow water equivalent, nearly equivalent to the 1991-2019 mean of 147 mm (Marsh et al., 2019). Air temperature at Inuvik was within a degree of the long-term mean during P1, P2 and P3, but P4 was 3.8°C cooler than the long-term mean (Figure 2). P1 and P2 experienced a combined 72.9 mm of rainfall, close to the long-term mean of 66.0 mm for this period, while P3 and P4 experienced 107.3 mm of rainfall, more than double the long-term mean of 45.0 mm for this period (Figure 2).

Multiple lake and watershed attributes for each lake were quantified for comparison with δ_{I} and E/I. Lake-specific properties included surface area, depth, latitude, and elevation, while watershed-specific properties included surface area, mean hillslope angle, drainage density, vegetation cover and ice-wedge polygon coverage (Table 1). Within each sampled lake watershed, the areas of ice-wedge polygons were identified visually from satellite imagery and digitized manually. Drainage density was calculated as the length of all flowpaths with a contributing area greater than 5000 m^{2}, and then divided by the total area of

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the watershed. A threshold of 5000 m^2 was chosen as this was roughly the threshold when water tracks became visible in optical satellite images. Vegetation height in each catchment was quantified using the remote sensing vegetation height product produced by Bartsch et al. (2020), which provides vegetation height at 20 metre spatial resolution with a RMSE of 45 cm in vegetation height. All spatial analysis was carried out using ArcMap 10.7.1 (ESRI, 2019).

To evaluate whether lake and watershed properties were correlated with lake water balance components derived from water isotope data, linear regressions between lake and watershed properties and E/I and δ_I were tested. Only correlations where p <0.01 were considered significant. The distribution of each variable was plotted and data were transformed if the distribution was non-uniform. Statistical analysis was performed using R 4.1.0 (R Core Team, 2021).

We identified a strong relation was identified between WA/LA and average E/I and used this relationship to predict average E/I for all lakes in the region. Watershed area was estimated for all lakes >0.25 ha (n = 7454) in the study area outlined in Figure 1, including the 25 lakes sampled for isotope composition, by applying the D8 water routing algorithm (O'Callaghan and Mark, 1984) to the 2 m resolution the digital elevation model ArcticDEM (PGC, 2018). The ratio of WA/LA was calculated for each lake by dividing the total watershed area by the total area of the lake(s) in the watershed. In rare cases watersheds were not delineated accurately and part of the watershed was clipped to the lake boundary, due to slight offsets between the lake layer and ArcticDEM. We rejected all watersheds where WA/LA <1.5, as all watersheds we inspected that met this criteria were improperly delineated due to the offset error just described. After rejecting all watersheds where WA/LA <1.5, a total of 7340 watersheds remained.

4 Results

4.1 Seasonal evolution of lake water balances

Lake water isotope compositions span from near the LMWL to approaching δ_{SSL}, reflecting a range of lake water balance conditions (Figure 3). Lake water isotope compositions plot close to the LEL, indicating that the isotope framework has been well estimated using the Trail Valley Creek meteorological data. No lakes plot between δ_{SSL} and δ^*, indicating that E/I was <1 for all lake samples and lakes were not trending towards desiccation during any sampling episode. As expected, rain and snow samples plot in two distinct clusters along the isotopically-enriched and depleted segments of the LMWL, respectively, permitting rain-influenced and snow-influenced δ_I to be distinguished. Most lakes plot slightly above the LEL, signifying that the δ_I of most lakes is more reflective of rain than snow.

We observed distinct shifts in lake water isotope composition along the LEL, $\delta^{18}O_I$ and E/I between sampling dates. Shifts in E/I between sampling dates were often more pronounced than shifts in $\delta^{18}O_I$. During P1 (May 1 to June 15), $\delta^{18}O$ decreased from a median of -15.04‰ (-17.53‰ to -13.81‰ inter-quartile range [IQR]) to -16.15‰ (-17.89‰ to -15.43‰ IQR) and δ^2H values decreased from -130.35‰ (-139.08‰ to -120.88‰ IQR) to -132.27‰ (-142.41‰ to -125.77‰ IQR) (Figure 3). At the same time, the range of $\delta^{18}O_I$ values decreased and the median decreased very slightly from -19.36‰ (-20.33‰ to -18.34‰ IQR) to -19.59‰ (-20.34‰ to -18.81‰ IQR) (Figure 4a). The median (and range) of E/I values also decreased during P1

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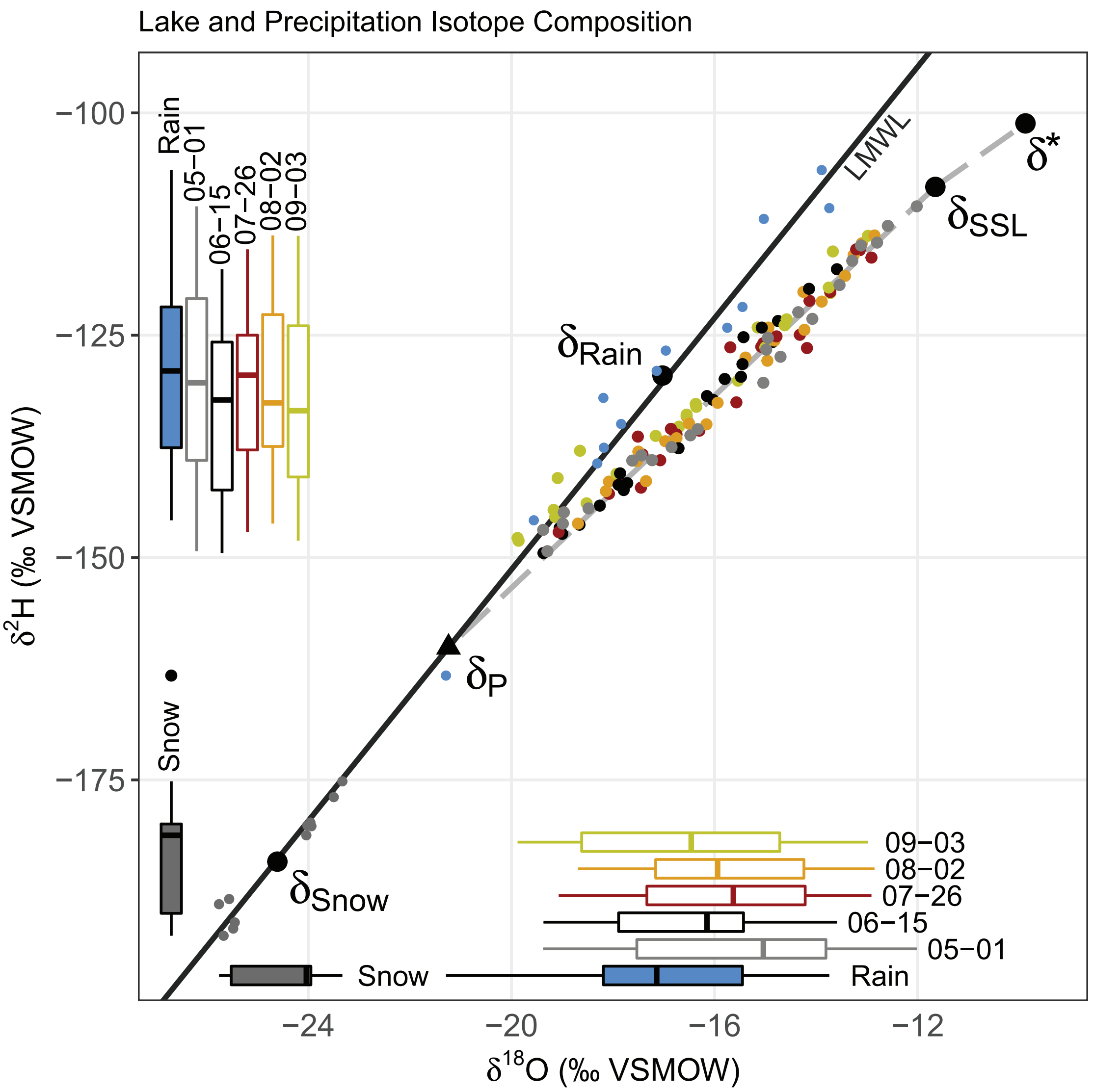

Figure 3. Isotope values and their respective ranges for lake and precipitation samples are plotted in δ^{18}O-δ^{2}H space, as deviations per mil from Vienna Standard Mean Ocean Water (VSMOW). The Local Meteoric Water Line (LMWL, black solid line, δ^{2}H = 7.1* δ^{18}O – 10.0) is calculated as a regression through snow samples collected throughout the study region and rain samples collected from Inuvik in 2018 (n = 24). The Local Evaporation Line (LEL, grey dashed line, δ^{2}H = 5.2*δ^{18}O – 48.9), the isotope composition of a waterbody at desiccation (δ^{*}), and the point at which evaporation is equal to inflow (E/I = 1, δ_{SSL}) are added for reference.

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from 0.20 (0.07 to 0.38 IQR) to 0.13 (0.07 to 0.18 IQR) (Figure 4b). The median reduction in E/I was -0.08 (Figure 4c). The reduction in E/I values that occurred during P1 was larger than in any other period (Figure 4c).

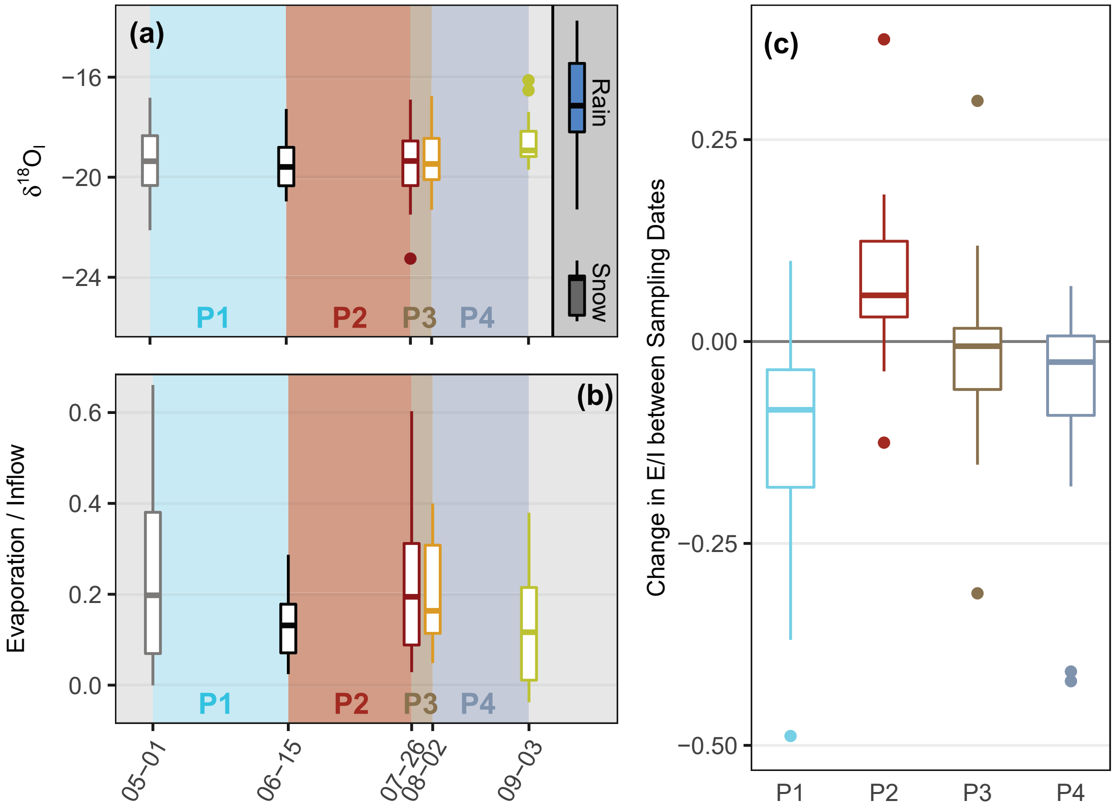

Figure 4. (a) The distribution of $\delta^{18}O_I$ for each sampling date with rain and snow samples for reference. (b) The distribution of E/I values at all lakes across the four sampling dates. (c) The distribution of lake-specific change in E/I between the sampling dates.

During P2 (June 15 to July 26), lake water isotope compositions increased slightly. The median δ^{18}O increased from -16.15‰ (-17.89‰ to -15.43‰) to -15.63‰ (-17.33‰ to -14.22‰ IQR) and the median δ^2H increased from -132.27‰ (-142.41‰ to -125.77‰ IQR) to -129.50‰ (-137.90‰ to -124.99‰ IQR) (Figure 3). During P2, $\delta^{18}O_I$ increased very slightly from -19.59‰ (-20.34‰ to -18.81‰ IQR) to -19.35‰ (-20.34‰ to -18.55‰ IQR) (Figure 4a). The median E/I ratio increased from 0.13 (0.07 to 0.18 IQR) to 0.19 (0.09 to 0.31 IQR) (Figure 4b). The median increase in E/I was 0.06 (0.03 to 0.12 IQR) although three lakes experienced a decrease in E/I (Figure 4c).

Lake water isotope compositions decreased marginally during P3 (July 26 to Aug 2). Median δ^{18}O decreased from -15.63‰ (-17.33‰ to -14.22‰ IQR) to -15.94‰ (-17.16‰ to -14.24‰ IQR) and median δ^2H decreased from -129.50‰ (-137.90‰ to -124.99‰ IQR) to -132.60‰ (-137.51‰ to -122.68‰ IQR) (Figure 3). E/I and $\delta^{18}O_I$ values also shifted minimally during P3. Median $\delta^{18}O_I$ decreased from -19.35‰ (-20.33‰ to -18.55‰ IQR) to -19.47‰ (-20.10‰ to -18.45‰ IQR) (Figure 4b). The

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median change in E/I was -0.01 (-0.06 to 0.02 IQR) with a near equal number of lakes experiencing an increase or decrease in E/I (Figure 4c).

During P4 (August 2 to September 3), median δ^{18}O values decreased from -15.94‰ (-17.16‰ to -14.24‰ IQR) to -16.55‰ (-18.65‰ to -15.00‰ IQR) and δ^{2}H values decreased from -132.60‰ (-137.51‰ to -122.68‰ IQR) to -133.49‰ (-140.94‰ to -123.94‰ IQR) (Figure 3). On September 3, some lakes plotted close to the LMWL, indicating that their waters had experienced negligible amounts of evaporation (Figure 3). The median $\delta^{18}O_I$ values increased from -19.47‰ (-20.10‰ to -18.45‰ IQR) to -18.94‰ (-19.18‰ to -18.14‰ IQR) and the range of values narrowed and became more similar to rain (Figure 4a). Median E/I values also returned towards June 15 values by the end of P4, decreasing to a median of 0.11 (0.00 to 0.21 IQR) (Figure 4b). Two thirds of lakes experienced a decrease in E/I and the median change in E/I was -0.03 (-0.09 to 0.01 IQR) (Figure 4c).

4.2 Correlation analysis between lake and watershed attributes and lake water balance metrics

While a distinct seasonal trend was observed in lake water isotope composition, E/I, and $\delta^{18}O_I$, variability among lakes in average E/I and $\delta^{18}O_I$ was partially explained by statistically significant relationships with lake and watershed properties. Average E/I was negatively correlated with WA/LA ($p < 0.0001$, Table 2, Figure 5a) and average watershed slope ($p = 0.0085$, Table 2) and was positively correlated with the log of lake surface area ($p < 0.0001$, Table 2). Thus, lakes that had higher E/I values tended to have relatively smaller watershed size, have a flatter watershed, and be larger in surface area. When these three variables are combined into a linear model, only WA/LA remained a significant predictor of average E/I ($p = 0.0002$) while lake area ($p = 0.1949$) and average watershed slope ($p = 0.8472$) became insignificant effects (Table 2). Both lake surface area and average watershed slope were correlated with WA/LA (Table 2). For the relationship between average E/I and WA/LA, the majority of downstream lakes were outliers, having higher than typical E/I ratios for a given WA/LA (Figure 5a).

Average $\delta^{18}O_I$ was only correlated with two attributes: lake elevation ($p = 0.0029$) and latitude ($p = 0.0027$) (Table 2). Lakes at lower elevations and higher latitudes tended to have higher $\delta^{18}O_I$ values (Figure 5b). Lake elevation and latitude are also nearly perfectly correlated ($R^2 = 0.97$, Table 2), with lake elevation decreasing as latitude increases.

5 Discussion

5.1 Influence of meteorological conditions on lake water balances

Water isotope measurements from five discrete time points provide context for characterizing the seasonal evolution of thermokarst lake water balances as a series of phases in relation to meteorological conditions and influential hydrological processes (Figure 6). A previous study by Wilcox et al. (under rev.) observed the presence of snowmelt bypass during the Freshet Phase (P1; Figure 6). Initial snowmelt runoff flowing into lakes did not mix with water underneath lake ice because the ~0°C snowmelt runoff was less dense than the deeper, warmer waters beneath the lake ice, causing it to flow into and out of lakes without mixing with lake water (Henriksen and Wright, 1977; Hendrey et al., 1980; Bergmann and Welch, 1985; Wilcox et al., under

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Table 2. Significant correlations between δ_I, E/I, and explanatory variables. In cases where there are multiple explanatory variables, the p-values for each explanatory variable are listed respectively. p-values >0.05 are *italicized*.

Response Variable	Explanatory Variable(s)	p-value(s)	Adjusted R^2
Average E/I	log(WA/LA)	<0.0001	0.7361
Average E/I	log(Lake Area)	<0.0001	0.5092
Average E/I	Average Watershed Slope	0.0085	0.2332
Average E/I	log(WA/LA) + log(Lake Area) + Average Watershed Slope	0.0002, *0.1949*, *0.8472*	0.7339
Average Watershed Slope	log(Lake Area)	0.0040	0.2771
log(WA/LA)	log(Lake Area)	<0.0001	0.5150
log(WA/LA)	Average Watershed Slope	0.0022	0.3128
Average $\delta^{18}O_I$ (‰)	Lake Latitude	0.0027	0.3010
Average $\delta^{18}O_I$ (‰)	Lake Elevation	0.0029	0.2954
Lake Elevation	Lake Latitude	<0.0001	0.9695

rev.). This resulted in a much smaller reduction in E/I than would have occurred if freshet runoff was able to mix with the entire lake water column (Wilcox et al., under rev.). While snowmelt bypass does limit the recharge of lake waters during the Freshet Phase, the sheer volume of snowmelt runoff and lack of evaporation due to lake ice cover results appears to compensate for the impact of snowmelt bypass and results in a reduction in E/I greater than any other period (Figure 4c).

During the Freshet Phase, $\delta^{18}O_I$ shifted towards the value of δ_P and not towards δ_S even though minimal rainfall fell during the period (Figure 2) and snowmelt runoff was flowing into lakes. Wilcox et al. (under rev.) determined that by the time snowmelt runoff could mix with lake waters, the soil had begun to thaw and allowed snowmelt runoff to mix with soil water from the previous year before flowing into lakes. This resulted in a post-snowmelt $\delta^{18}O_I$ that was a mixture of snow-sourced and rain-sourced water (Figure 4a). Since only a small amount of snow-sourced water was incorporated into lakes during the Freshet Phase, the $\delta^{18}O_I$ of lakes remained primarily rain-sourced throughout the entire study period (Figure 4a).

The Evaporation Phase (P2; Figure 6) was characterized by minimal change in $\delta^{18}O_I$ (Figure 4a) and rising E/I ratios at nearly all lakes (Figure 4c), caused by the relatively dry and warm conditions that are typical for this region. Previous water balance studies conducted in this region also found that lake evaporation rates were higher as a result of substantial incoming solar radiation in June and July compared to August and September, and when inflow is minimal due to low precipitation (Marsh and Bigras, 1988; Pohl et al., 2009).

During the brief Soil Wetting Phase (P3; Figure 6), there was minimal change in $\delta^{18}O_I$ (Figure 4a) and E/I (Figure 4b) despite the 41.2 mm of rainfall, which represented 25% of the rainfall during the study period. This evidence suggests little runoff was generated by this event and that rainfall that fell directly on lakes had only a minor effect on lake water balances. We hypothesize that dry conditions preceding the rainfall event allowed soils to absorb the rainfall without becoming saturated sufficiently to generate much lateral flow. The lack of runoff in response to this rainfall event is expected given that antecedent

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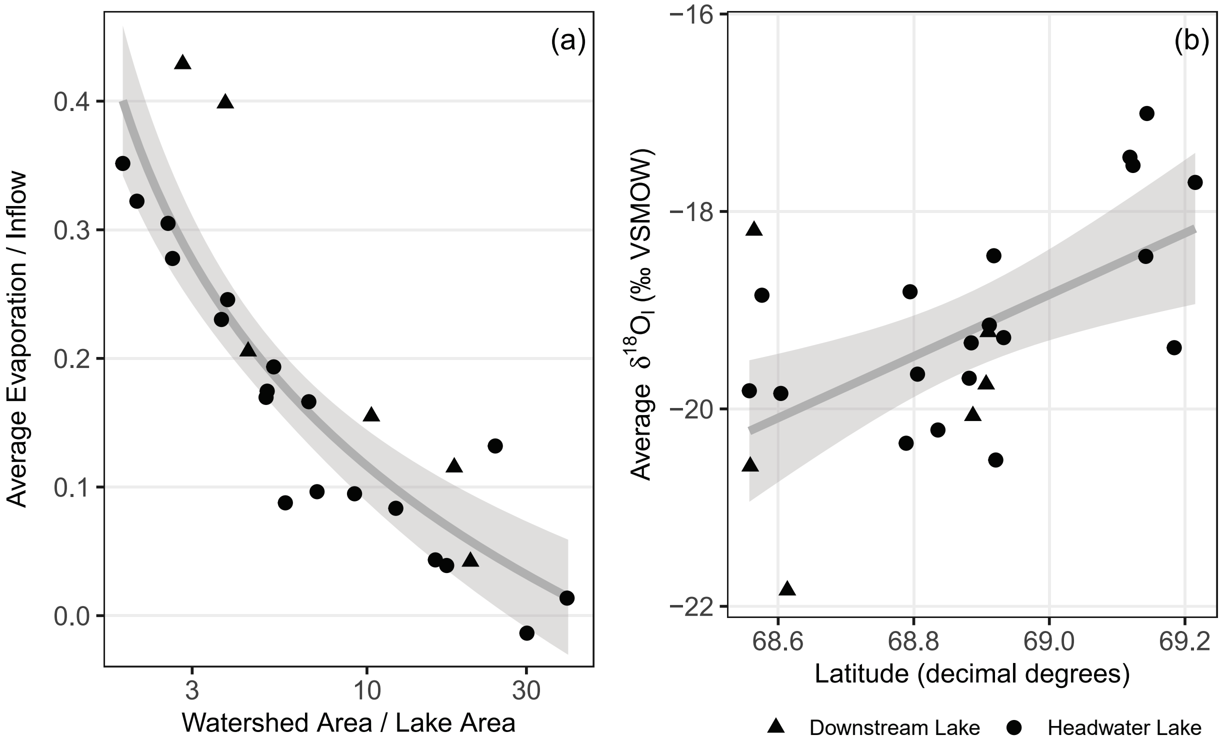

Figure 5. (a) Relationship between average evaporation/inflow ratio and watershed area/lake area for sampled lakes. The grey line represents a line of best fit between watershed area/lake area and the average evaporation/inflow ratio, such that y = log(x), with the 95% confidence interval shown in grey shading (R^2 = 0.74, p <0.0001). Note the logarithmic x-axis. (b) The relationship between latitude and δ_I. The grey line represents a linear line of best fit between δ_I and latitude, with the 95% confidence interval shown in grey shading (R^2 = 0.30, p = 0.0027).

soil moisture conditions have been observed to greatly influence the efficiency of runoff generation from rainfall events in areas underlain by continuous permafrost (Roulet and Woo, 1988; Kane et al., 1998; Favaro and Lamoureux, 2014; Stuefer et al., 2017). Since the first half of the summer is typically drier than the second half (Figure 6), soils generally experience drying throughout the first half of the summer before becoming rewetted.

The Recharge Phase (P4; Figure 6) was defined by the $\delta^{18}O_I$ of lakes becoming more similar to rainfall (Figure 4a) and E/I ratios decreasing to values similar to the beginning of the Freshet Phase (Figure 4b). The Recharge Phase of 2018 was 3.8°C cooler than average and experienced 66.1 mm of precipitation, exceeding the long-term mean of 37.2 mm (Figure 2). Given the cooler air temperatures, it is likely that evaporation rates also decreased and contributed to the decrease in E/I. Since the meteorological conditions were wetter and cooler than the long-term means, we expect the water balance conditions of the Recharge Phase captured in 2018 to be more representative of a later sampling date during a more normal climate year

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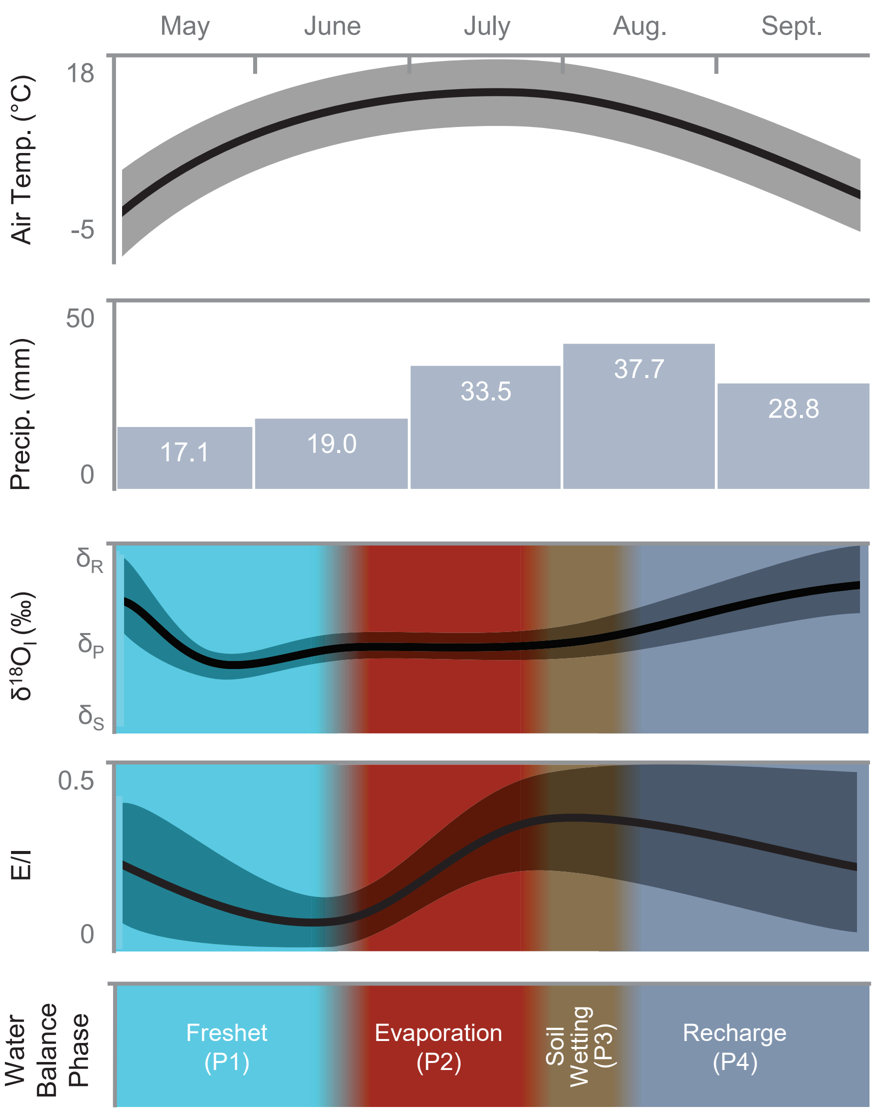

Figure 6. Relations between meteorological conditions and the response of $\delta^{18}O_I$ and E/I used to designate water balance phases. Shaded areas around the $\delta^{18}O_I$ and E/I lines represent the potential variability caused by lake and watershed attributes or meteorological conditions.

(e.g., late September). E/I ratios on May 1 were greater than September 3, suggesting that lakes received less inflow during the Recharge Phase during 2017 than during 2018 (Figure 4b).

Findings from previous studies of thermokarst lakes in this region imply that the four phases of lake water balance we observed are typical for this region. The four phases of seasonal water balance evolution we identified (Freshet, Evaporation, Soil Wetting, Recharge) roughly follow the pattern of seasonal changes in lake surface area observed by Cooley et al. (2019) who analyzed near-weekly satellite imagery of a large region which encompassed the lakes we studied. Cooley et al. (2019) observed initial decreases in total lake surface area during the Evaporation Phase, followed by stabilizing or increasing trends in lake surface area by the end of August. Additionally, Pohl et al. (2009) modeled the water balance for a single lake in our study region across a 30-year period, and identified that maximum lake level was most likely to occur in early June or late August/early September, corresponding with the start of the Evaporation Phase and end of the Recharge Phase.

In comparison to the five thermokarst lake regions examined in a synthesis of isotope data by MacDonald et al. (2017), the relatively low E/I ratios and rainfall-like $\delta^{18}O_I$ values found at our study lakes compare most closely to lakes in the Alaskan Coastal Plain (ACP), where the majority of lakes also have an E/I < 0.25 and $\delta^{18}O_I$ similar to rain. This region is cooler during the summer and has a shorter ice-free season (Arp et al., 2015) than our study region, but receives less precipitation and is more lake rich (MacDonald et al., 2017), suggesting lakes likely have smaller watersheds than in our study region. The cooler and shorter summers in ACP, which decrease evaporation, may be offset by smaller watersheds and less precipitation, which decrease inflow, leading to similar E/I values as our study region. Our lakes differed from the more nearby OCF, where most lakes have an E/I between 0.25 and 0.75 but have a smaller WA/LA of ~3, whereas the average WA/LA of lakes we sampled was 9.5 (Table 1). OCF differs from our study region in that it is situated in a post-glacial lake bed underlain by fine-grained glaciolacustrine sediments (Hughes, 1972), resulting in a relatively flat landscape with poorer ability to convey runoff in comparison to our study region, where rolling hills are well drained by networks of peat channels with high hydraulic conductivity (Quinton and Marsh, 1998). These differences between OCF and our study region may explain the greater E/I values at the former, however our study year was cooler and wetter than average, which may also have contributed to the comparatively low E/I ratios.

5.2 Effects of lake and watershed attributes on E/I

While shifts in E/I over time can be attributed to changing meteorological conditions and hydrological processes, differences in E/I among lakes can be further explained by lake and watershed attributes. Most of the variability in average E/I among lakes can be explained by WA/LA (R^2 = 0.74, Figure 5a), as lakes with smaller WA/LA likely receive less inflow and have greater E/I ratios as a result. A similar inverse relationship between E/I ratios and WA/LA has also been observed in OCF (Turner et al., 2014), and the taiga-shield of the Northwest Territories (Gibson and Edwards, 2002). Four of the six downstream lakes we sampled had anomalously high E/I ratios compared to their WA/LA (Figure 5a). Downstream lakes receive evaporated inflow from their upstream lakes, which is then evaporated further in the downstream lake, producing an enhanced E/I ratio. When downstream lakes are removed from the regression between E/I and WA/LA, the R^2 improves from 0.74 to 0.82.

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We estimated E/I for all lakes in the study area using the strong relationship between average E/I and WA/LA. This was done by delineating the watersheds of all lakes larger than 0.25 ha in the study area (Figure 1) and applying the fitted regression between log(WA/LA) and average E/I for the headwater lakes we sampled:

285 $$AverageE/I = -0.10867 * \log(WA/LA) + 0.37007 \quad (3)$$

The resulting histogram of average E/I (Figure 7b) is primarily skewed towards larger values, because WA/LA is skewed

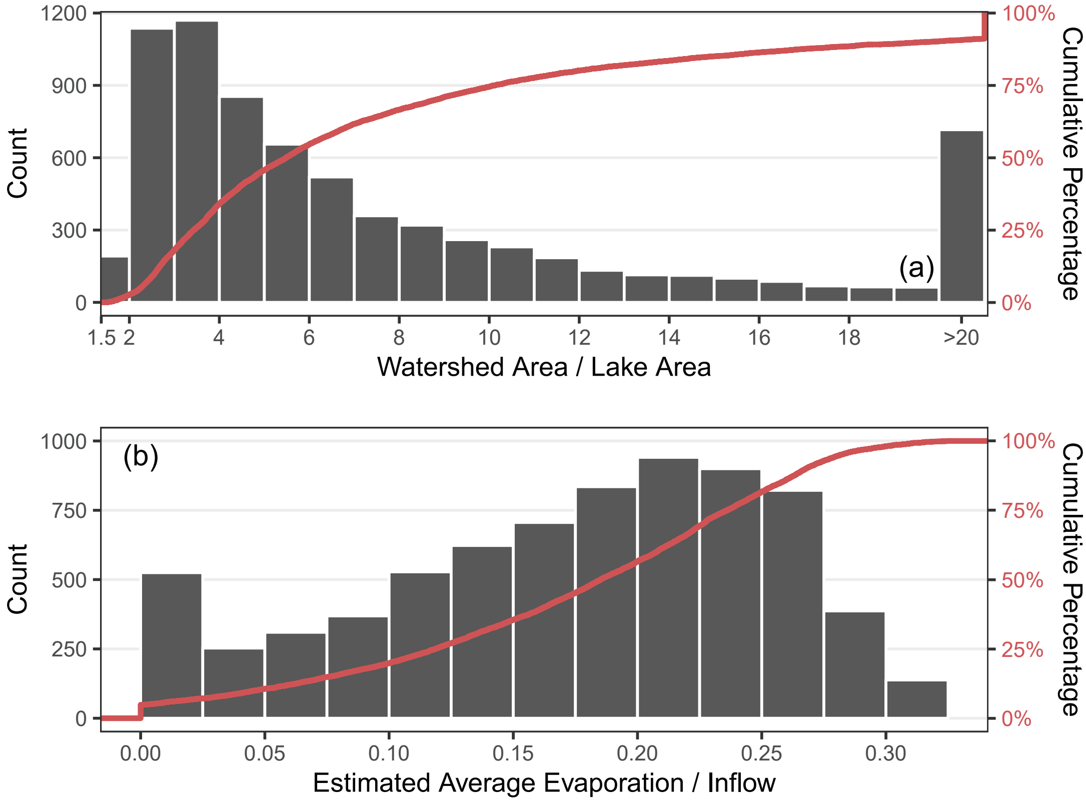

Figure 7. Distribution of (a) watershed area/lake area and (b) evaporation/inflow for 7340 lakes and their watersheds in the study area. Average evaporation/inflow is estimated using the relationship with watershed area/lake area of headwater lakes. Note that the leftmost bin in panel (a) is only half the width of other bins, reflecting the rejection of all WA/LA <1.5 that we applied when filtering the data.

towards smaller values (Figure 7a). An average E/I between 0.2 and 0.225 is most common for lakes indicating the lakes are dominated by inflow. At WA/LA >30, the estimated E/I becomes 0, and as a result 7% of lakes are predicted to have an average E/I of 0 (Figure 7b). None of the lakes appear to be approaching desiccation.

290 We hypothesize that the response of lakes in this region to climate change will be strongly influenced by their WA/LA. Recent predictions of future Arctic precipitation indicate greater rainfall, snowfall and more yearly precipitation overall (Brown and Mote, 2009; Bintanja and Andry, 2017; Bintanja et al., 2020) and in OCF increased rainfall is already reducing lake E/I

ratios (MacDonald et al., 2021). We would expect that under wetter conditions E/I ratios would decrease for all lakes, and the threshold where average E/I = 0 would also decrease. The logarithmic nature of the relationship between WA/LA and E/I indicates that average E/I is more sensitive to changes in inflow as WA/LA decreases. Therefore, we expect that lakes with smaller WA/LA would experience greater reductions in average E/I under a wetter future climate. We also hypothesize that lakes with larger WA/LA could be more vulnerable to rapid lake drainage because they receive more inflow and likely experience greater fluctuations in lake level as a result; rapid drainage is typically triggered when extremely high water levels lead to the rapid thermo-mechanical erosion of a new lake outlet (Mackay, 1988; Brewer et al., 1993; Turner et al., 2010).

Alternatively, climate change may lead to a drier future for lakes, as ice-free periods lengthen, warmer air temperatures increase evaporation and permafrost thaw leads to landscape drying (Walvoord and Kurylyk, 2016; Koch et al., 2022). A combination of such conditions have already caused lake contraction in western Greenland (Finger Higgens et al., 2019). If drier conditions prevail, given the logarithmic nature of the relationship between WA/LA and average E/I, we expect lakes with smaller WA/LA will experience greater increases in average E/I. If future climate change causes dry enough conditions to cause lake desiccation, a large number of lakes in the study area could potentially be affected, given that the distribution of WA/LA in the region is skewed towards smaller values of WA/LA, with many lakes possessing a WA/LA <4 (Figure 7a). The 'drier future' scenario may seem less likely to result in lake desiccation given that lakes are currently dominated by inflow, however the portion of precipitation converted into runoff can be halved during dry periods when compared to wetter periods (Stuefer et al., 2017), further reducing runoff to lakes.

Future studies could build on our hypothesis that WA/LA will mediate the response of lakes to climate change by comparing past changes in lake surface area, via remote sensing and paleohydrological analyses, with WA/LA. We would expect that during drier periods, lakes with smaller WA/LA experienced greater reductions in lake surface area than lakes with larger WA/LA. Previous studies from this region have already found that lakes change in surface area in response to seasonal (Cooley et al., 2019) and multi-year (Plug et al., 2008) shifts in precipitation, indicating that changes to lake water balances can be observed by tracking changes in water surface area.

6 Conclusions

Water isotope-derived metrics were used to derive distinct seasonal phases of lake water balances for 25 thermokarst lakes in the tundra uplands east of the Mackenzie Delta (Northwest Territories, Canada). The freshet phase saw lakes experience a reduction in E/I and a shift of δ_I values towards δ_P, as a mixture of soil water and snowmelt-sourced freshet entered lakes and evaporation was minimal due to lake ice cover. Following this period was the evaporaiton phase, where minimal precipitation and warm and sunny conditions lead to increasing E/I ratios and no change in δ_I. Then, a brief and intense period of rainfall lead to minimal response in E/I and δ_I at lakes, as dry soils absorbed most of the precipitation during the soil wetting phase. In the final stages of summer during the recharge phase, air temperatures dropped and precipitation was unseasonably high, causing reductions in E/I and a shift of δ_I towards δ_R as lakes received increased runoff from their watersheds and evaporation rates reduced.

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Comparison of water isotope-derived lake water balance components with lake and watershed attributes shows that WA/LA explains the majority of variability in E/I among lakes. The strong relationship between average E/I and WA/LA allowed us to predict the average E/I for 7340 lakes in the study region. Predicted average E/I values were low compared to other regions of thermokarst lakes, only reaching as high as 0.33, indicating that lakes are not currently near risk of desiccation. We hypothesize that lakes with larger WA/LA will be more prone to rapid drainage if future conditions are wetter.

Few studies have directly investigated the influence of WA/LA on E/I (Gibson and Edwards, 2002; Turner et al., 2014), but given the strong relationship we found, WA/LA could serve as a useful metric in other permafrost environments for characterizing thermokarst lake water balances and predicting which lakes may be vulnerable to climate change. The non-uniform response of lake surface area to past climate change (Smith et al., 2005; Plug et al., 2008; Arp et al., 2011; Jones et al., 2011; Andresen and Lougheed, 2015) may also be explained by WA/LA; future studies could investigate whether lakes with smaller WA/LA are more likely to decrease in surface area than lakes with larger WA/LA. Further links between WA/LA and biogeochemical properties of lakes may also exist, as E/I has been linked to the biogeochemistry of a wide range of lakes, from tropical to tundra environments (Kosten et al., 2009).

Data availability. Data used in the manuscript are presented in Tables 1, 2 and the Appendix. Lake isotope and attribute data can be downloaded from the Trail Valley Creek Dataverse at https://doi.org/10.5683/SP3/AZE4ER. Meteorological data were retrieved from Environment and Climate Change Canada at https://climate.weather.gc.ca/historical_data/search_historic_data_e.html.

Appendix A: Isotope framework

The calculation of the point of maximum evaporative isotopic enrichment, or the isotope composition of lake at the point of desiccation (δ^*), Gonfiantini (1986):

$$\delta^* = \frac{h * \delta_{As} + \varepsilon_k + (\varepsilon^*/\alpha^*)}{h - \varepsilon_k - (\varepsilon^*/\alpha^*)} \tag{A1}$$

where α^* is the fractionation factor between the liquid and vapour phase of water, ε^* and ε_k are the equilibrium and kinetic separation terms, with ε^* serving as a convenient expression of α^* , where $\varepsilon^* = \alpha^* - 1$. The term h represents the relative humidity for the open water season (see below). Equilibrium fractionation (α^*) was calculated following equations from Horita and Wesolowski (1994):

$$1000 ln * \alpha^* = -7.685 + 6.7123 * \frac{10^3}{T} - 1.6664 * \frac{10^6}{T^2} + 0.35041 * \frac{10^9}{T^3} \tag{A2}$$

for δ^{18}O and

$$1000 ln * \alpha^* = 1158.8 * \frac{T^3}{10^9} - 1620.1 * \frac{T^2}{10^6} + 794.84 * \frac{T}{10^3} - 161.04 + 2.9992 * \frac{10^9}{T^3} \tag{A3}$$

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for δ^2H, where T represents the temperature of the interface (K) (see below). The kinetic separation term ε_k was calculated as:

355 $$\varepsilon_k = x * (1-h) \tag{A4}$$

where $x = 0.0142$ for δ^{18}O and $x = 0.0125$ for δ^2H (Gonfiantini, 1986). δ_{As} represents the isotope composition of atmospheric vapour, which we assume is in equilibrium with the isotope composition of summertime precipitation (δ_{Ps}, Gibson and Edwards, 2002). We can therefore estimate δ_{As} as:

$$\delta_{As} = (\delta_{Ps} - \varepsilon^*/\alpha^*) \tag{A5}$$

360 The reference point of when E/I = 1 (δ_{SSL}) was calculated using:

$$\delta_{SSL} = (\alpha^* * \delta_{Ps} * (1 - h + \varepsilon_k)) + \alpha^* * h * \delta_{As} + \alpha^* * \varepsilon_k + \varepsilon^* \tag{A6}$$

where δ_{Ps} represents summertime precipitation during the open-water season (Gonfiantini, 1986).

Data for air temperature and relative humidity were collected at Trail Valley Creek near the centre of the study area. The time period used for the average air temperature and relative humidity spans from when lakes became ice-free (June 15, 2018) until
365 the last day of sampling (September 3, 2018) to match the time span between the first and last sampling dates. For a few dates, air temperature was not recorded by the TVC meteorological station maintained by the Meteorological Service of Canada, and air temperature from another meteorological station at TVC was used instead.

The isotope composition of the lake-specific input water (δ_{I}) was calculated following Yi et al. (2008), where δ_{I} is estimated as the intersection of the LMWL and the lake-specific LEL, defined as the line between the measured isotope composition
370 of the lake (δ_{L}) and δ_{E}, which is the isotope composition of vapour evaporating from the lake. δ_{E} was calculated following Gonfiantini (1986):

$$\delta_E = \frac{(\delta_L - \varepsilon^*)/(\alpha^* - h * \delta_{As} - \varepsilon_k)}{1 - h - \varepsilon_k} \tag{A7}$$

δ_{I}, δ_{E} and δ_{L} were then used to calculate the ratio of evaporation to inflow (E/I) as described by Yi et al. (2008) and others as:

$$\frac{E}{I} = \frac{\delta_I - \delta_L}{\delta_E - \delta_L} \tag{A8}$$

375 assuming that lakes are well mixed and in hydrological and isotopic steady state.

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Table A1. Parameters used in isotope framework.

Parameter (unit)	Source	Value(s)
T (K)	Trail Valley Creek (WMO ID: 71683)	281.95
h (%)	Trail Valley Creek (WMO ID: 71683)	78.3
δ_P(^{18}O, ^{2}H) (‰)	Samples from study region.	-21.24,-160.1
δ_{SSL}(^{18}O, ^{2}H) (‰)	Equation A6	-11.65, -108.3
δ* (18O, 2H) (‰)	Equation A1	-9.88, -101.1
δ_{As} (^{18}O, ^{2}H) (‰)	Equation A5	-27.58, -207.7
α^* (^{18}O, ^{2}H) (‰)	Equation A2, A3	1.0109, 1.0986
ε_k (^{18}O, ^{2}H)	Equation A4	3.08, 2.71

Author contributions. The study design and sampling plan was developed by E.J.W. with input from B.B.W. and P.M. The field data collection and sample preparation for lab analysis was done by E.J.W. Analysis of the data was completed by E.J.W. with input from B.B.W. The manuscript was primarily written by E.J.W. with input from B.B.W. and P.M.

Competing interests. The authors declare that they do not have any conflict of interest.

Acknowledgements. E.J.W was funded by a W. Garfield Weston Award for Northern Research and an Ontario Graduate Scholarship. We acknowledge funding from ArcticNet, Northern Scientific Training Program, Polar Continental Shelf Program, the Canada Research Chairs program, and the Natural Sciences and Engineering Research Council of Canada. The research license (No. 16237) was administered by the Aurora Research Institute in Inuvik, Northwest Territories, Canada, and can be found at http://data.nwtresearch.com/Scientific/16237. The authors are grateful for the assistance of the Arctic Hydrology Research Group members for their help in completing the field work. The authors also thank the staff of the UW-EIL for completing the isotope analyses.

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