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Assessing variability in snowmelt bypass among thermokarst lakes using water isotope tracers, Northwest Territories, Canada.

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Abstract. Snow represents the largest potential source of water for thermokarst lakes, but the runoff generated by snowmelt (freshet) can flow beneath lake ice and out of lakes without mixing with and replacing pre-snowmelt lake water. Although this phenomenon, called "snowmelt bypass", is common in ice-covered lakes, it is unknown what lake and watershed properties cause variation in snowmelt bypass among lakes. Understanding the variability of snowmelt bypass is important because the amount of freshet that is mixed into a lake affects the biogeochemical properties of the lake. To explore lake and watershed attributes that influence snowmelt bypass, we sampled 17 open-drainage thermokarst lakes for isotope analysis before and after snowmelt. Isotope data were used to estimate the amount of lake water replaced by freshet and to observe how the water source of lakes changed in response to the freshet. A median of 25.2% of lake water was replaced by freshet, with values ranging widely from 5.2 to 52.8%. For every metre lake depth increased, the portion of lake water replaced by freshet decreased by an average of 13%, regardless of the size of the lake's watershed. Vertical mixing is more restricted in deeper lakes, which reduces the relative thickness of the layer where freshet can mix with lake water, leading to more snowmelt bypass at deeper lakes. We expect a similar relationship between increasing lake depth and greater snowmelt bypass could be present at all ice-covered open-drainage lakes, since the limited vertical mixing conditions that lead to this relationship are present at all ice-covered lakes. The water source of freshet that was mixed into lakes was not exclusively snowmelt, but a combination of snowmelt mixed with rain-sourced water that was released as the soil thawed after snowmelt. As climate warming increases rainfall and shrubification causes earlier snowmelt timing relative to lake ice melt, snowmelt bypass may become more prevalent with the water remaining in thermokarst lakes post-freshet becoming increasingly rainfall sourced. However, if climate change causes lake levels to fall below the outlet level (i.e., lakes become closed drainage) more freshet may be retained by thermokarst lakes as snowmelt bypass will not be able to occur until lakes reach their outlet level.

1 Introduction

In the continuous permafrost zone of the Arctic, regions with thermokarst lakes have formed where ice-rich permafrost has thawed and the ground surface has subsided. Thermokarst lakes typically range from 1 – 5 m in depth, 0.01 – 1000 ha in area, can cover over 25% of the land area (Grosse et al., 2008; Burn and Kokelj, 2009; Turner et al., 2014; Farquharson et al., 2016) and mostly formed during a brief warm period following the last deglaciation of the northern hemisphere (Brosius et al., 2021). Comparison of aerial photography from the mid-1900s with more recent satellite imagery has revealed both increases

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and decreases in thermokarst lake area and number (Smith et al., 2005; Plug et al., 2008; Marsh et al., 2009; Jones et al., 2011; Finger Higgens et al., 2019). These changes are partially attributed to shifting thermokarst lake water balances: increased air temperatures (Woo et al., 2008), longer ice-free seasons (Surdu et al., 2014; Arp et al., 2015), permafrost thaw (Walvoord and Kurylyk, 2016), and shrub expansion leading to increased transpiration (Myers-Smith et al., 2011) and interception (Zwieback et al., 2019), all cause less inflow and more water to evaporate from thermokarst lakes. Contrarily, increasing precipitation can lead to more inflow to lakes, offsetting any rise in evaporation, interception and transpiration (Walsh et al., 2011; Stuefer et al., 2017; Box et al., 2019; MacDonald et al., 2021).

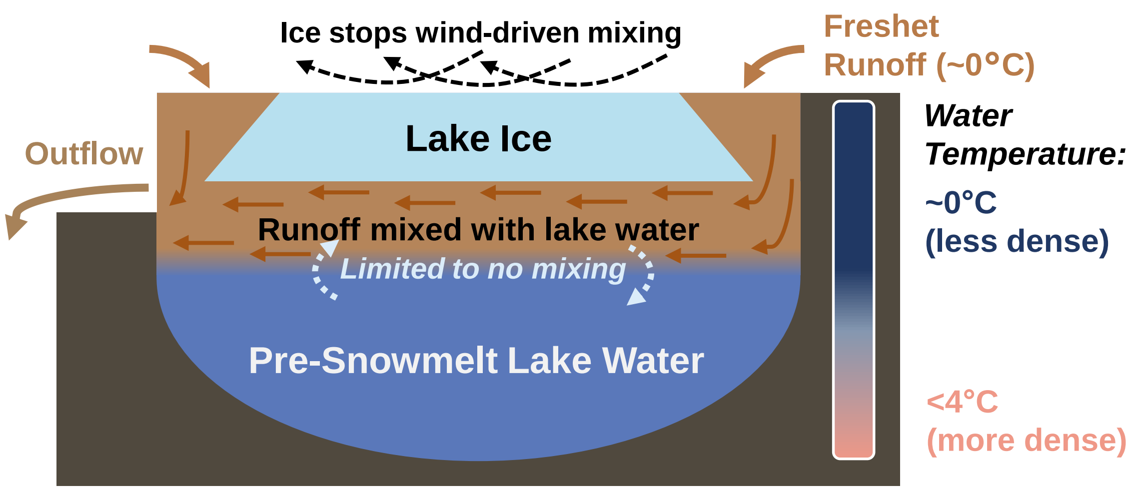

Figure 1. A conceptual cross-section of a open-drainage lake when freshet has begun. Freshet initially flows into the lake at the edge where lake ice has melted. A layer of snowmelt runoff mixed with lake water then remains buoyant on top of the warmer lake water, before flowing through the outlet (i.e., ‘snowmelt bypass’). Limited mixing occurs due to density differences between runoff and deeper lake water, and the lack of wind-driven mixing due to the presence of lake ice.

Runoff generated by snowmelt in lake watersheds represents a large potential water source for lakes, as snowfall comprises 40 to 80% of total precipitation in the Arctic (Bintanja and Andry, 2017). When snow melts in spring, the flow of runoff into lakes (freshet) generally results in the highest lake levels of the year (Woo, 1980; Roulet and Woo, 1988; Hardy, 1996; Pohl et al., 2009). When freshet is low, thermokarst lakes are prone to desiccation (Bouchard et al., 2013; Marsh and Bigras, 1988; Marsh and Lesack, 1996). It is a reasonable expectation that lakes which receive more freshet will also contain more freshet by the end of the snowmelt if they remain below their outlet level (i.e., closed-drainage lakes). However, for ice-covered lakes at or near their outlet level (i.e., open-drainage lakes), freshet may flow into and out of a lake without mixing with and replacing the pre-freshet lake water, resulting in “snowmelt bypass” (Bergmann and Welch, 1985) (Figure 1). While lake ice inhibits wind-driven mixing of lake water, the cooler, less dense freshet (~0 °C) cannot mix with the deeper, warmer, and denser lake waters (<4 °C). As a result, freshet water will flow into and out of a open-drainage lake without replacing the deeper, pre-snowmelt lake water until vertical mixing within the lake begins, which is initiated by the warming of lake

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waters from solar radiation penetrating through snow-free ice, and wind-driven mixing after the lake becomes ice-free (Cortés and MacIntyre, 2020). Snowmelt bypass is a common occurrence that has been observed in a wide variety of ice-covered open-drainage lakes around the world (Henriksen and Wright, 1977; Jeffries et al., 1979; Hendrey et al., 1980; Bergmann and Welch, 1985; Schiff and English, 1988; Edwards and McAndrews, 1989; Cortés et al., 2017). Although previous studies have established the mechanisms and conditions that cause snowmelt bypass, no studies have examined how lake and watershed characteristics affect snowmelt bypass. Understanding the factors that influence the amount of freshet retained by thermokarst lakes is important because of subsequent influence on lake water pH, nutrient composition, and suspended sediment (Henriksen and Wright, 1977; Marsh and Pomeroy, 1999; Finlay et al., 2006; Balasubramaniam et al., 2015).

In this study, we determine factors influencing the magnitude of snowmelt bypass for 17 open-drainage thermokarst lakes in the lake-rich tundra uplands east of the Mackenzie Delta in the Northwest Territories, Canada, during the freshet of 2018. This area contains thousands of thermokarst lakes that constitute up to 25% of the landscape and have changed in area and number during the past several decades in response to changing precipitation and permafrost thaw (Plug et al., 2008; Marsh et al., 2009). We applied isotope methods to estimate the proportion of lake water replaced by freshet during spring 2018 and evaluated relations between lake and watershed characteristics and the proportion of lake water replaced by freshet. Isotope tracers were also used to assess whether the freshet is sourced solely from snowmelt, or if other water sources contributed to freshet.

2 Study Area

The 17 studied lakes are situated in the taiga-tundra uplands east of the Mackenzie Delta, in the northwest region of the Northwest Territories, Canada (Figure 2). The landscape is comprised of rolling hills and is strongly influenced by permafrost thaw, as evidenced by the thousands of thermokarst lakes which formed between 13 000 and 8000 years ago (Rampton, 1988; Burn and Kokelj, 2009) that are typically 2 – 4 m in depth with a surface area from 10 – 1000 ha (Pienitz et al., 1997). The lakes sampled are situated along a ~70 km stretch of the Inuvik-Tuktoyaktuk Highway north of the town of Inuvik (Figure 2). The average area of the selected lakes is 14.2 ha (0.9 – 90.5 ha) and the average maximum depth is 2.2 m (1.0 – 4.1 m) (Table 1). All lakes have a defined outlet channel observed to be active during the spring melt, thus classifying them hydrologically as open-drainage, and many lakes have defined channelized inflows from their watersheds in the form of small streams or ice-wedge polygon troughs.

Soils in the region have evolved from fine-grained morainal tills, ice-contact sediment, and lacustrine deposits (Rampton and Wecke, 1987). Subsurface flow is efficiently conveyed by a network of interconnected peat channels 0.3 – 1.0 m across that exist between mineral earth hummocks (Quinton and Marsh, 1998). Lake watersheds contain tall shrub (>1 m), low-shrub (~0.5 m), and shrub-free landcover types comprising lichen, moss, and tussocks (Lantz et al., 2010; Grünberg et al., 2020). Mean annual air temperature in Inuvik is -8.2°C and mean annual precipitation is 241 mm of which 66% is snow, based on 1981-2010 climate normals. Snowmelt usually begins in mid-May, and lakes typically become ice-free in June and freeze-up in mid-October (Burn and Kokelj, 2009).

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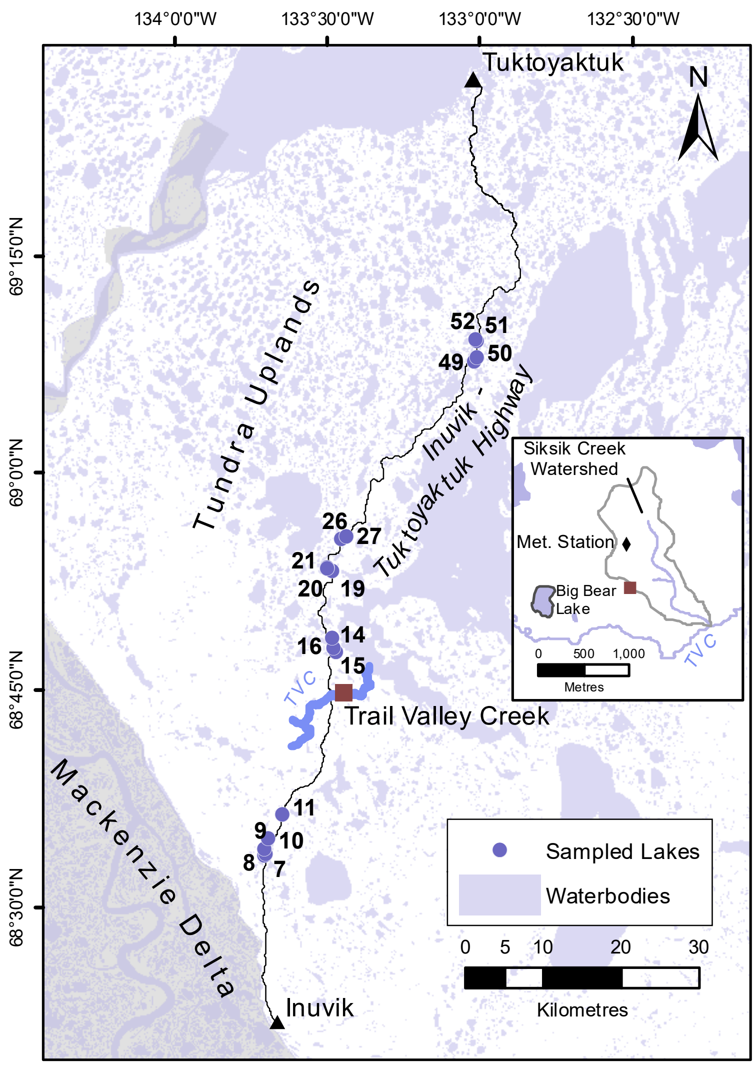

Figure 2. Blue circles indicate lakes that were sampled before and after snowmelt in 2018. Tundra uplands are in white while the Mackenzie Delta is in grey. In the inset, key locations near the Trail Valley Creek field station are shown.

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Table 1. Lake and watershed properties for sampled lakes. Lake locations are shown on Figure 2

Lake	Longitude	Latitude	Lake Elevation (m asl)	Lake Depth (m)	Ice Thickness (m)	Snow Depth (cm)	Lake Area (ha)	Watershed Area (ha)	Watershed Area/Lake Area
7	-133.76149	68.55745	89	2.24	0.81	22	2.81	6.45	2.62
8	-133.75566	68.55879	89	2.30	0.79	15	1.88	15.67	7.55
9	-133.76025	68.56446	86	1.02	0.84	30	59.56	203.56	3.42
10	-133.74651	68.57601	88	1.65	0.85	54	90.48	168.58	1.86
11	-133.70334	68.60390	83	1.91	0.97	11	0.92	21.76	17.32
14	-133.52093	68.78877	52	1.42	0.84	4	10.68	60.64	5.7
15	-133.52885	68.79452	57	1.57	1.14	19	5.66	29.83	4.99
16	-133.53196	68.80550	52	3.18	1.32	7	1.15	19.75	16.02
19	-133.52616	68.88175	39	2.46	1.24	11	5.68	38.98	7.09
20	-133.54301	68.88474	37	2.69	1.27	10	2.30	19.93	9.18
21	-133.54002	68.88721	36	1.78	1.19	11	2.61	10.91	3.96
26	-133.49557	68.91814	38	1.47	1.19	6	4.84	17.89	3.83
27	-133.47711	68.92095	45	3.10	1.22	10	1.13	8.57	6.7
49	-133.05281	69.11883	9	2.18	0.91	23	17.50	46.23	2.54
50	-133.04203	69.12333	8	1.65	0.86	19	8.16	31.92	3.67
51	-133.04142	69.14222	4	2.31	0.84	24	2.24	12.01	5.26
52	-133.04730	69.14389	6	4.14	0.86	18	23.52	49.92	2.05
Min	-133.76149	68.55745	4	1.02	0.79	4	0.92	6.45	1.86
Mean	-133.47497	68.83944	48	2.18	1.01	17	14.18	44.86	6.10
Max	-133.04142	69.14389	89	4.14	1.32	54	90.48	203.56	17.32

The 2018 snowmelt season was typical in comparison to recent decades. End of winter snow surveys conducted in the 58 km^2 watershed of Trail Valley Creek in 2018 (Figure 2) recorded an average snow water equivalent (SWE) of 141 mm, close to the average SWE recorded by these surveys of 147±35 mm from 1991 – 2019 (Marsh et al., 2019). At Trail Valley Creek, snowmelt began about May 1, with snow-free areas beginning to appear by May 8, while only the remnants of large snow drifts remained by June 3. Lake ice near Trail Valley Creek became snow-free by May 10, and lakes became ice-free on June 14. The mean air temperature at Trail Valley Creek during the sampling period from April 26 – June 15, 2018, was 0.4 °C, which was cooler than the average of 1.7 °C during 1999-2019 (Figure 3). Air temperatures roughly followed the average minimum and maximum daily air temperatures, with some temporal variability which can be expected for any given year. Maximum daily air temperatures were mostly above 0 °C after May 8, which was similar to the average timing of the first above 0 °C day during 1999-2019 (Figure 3).

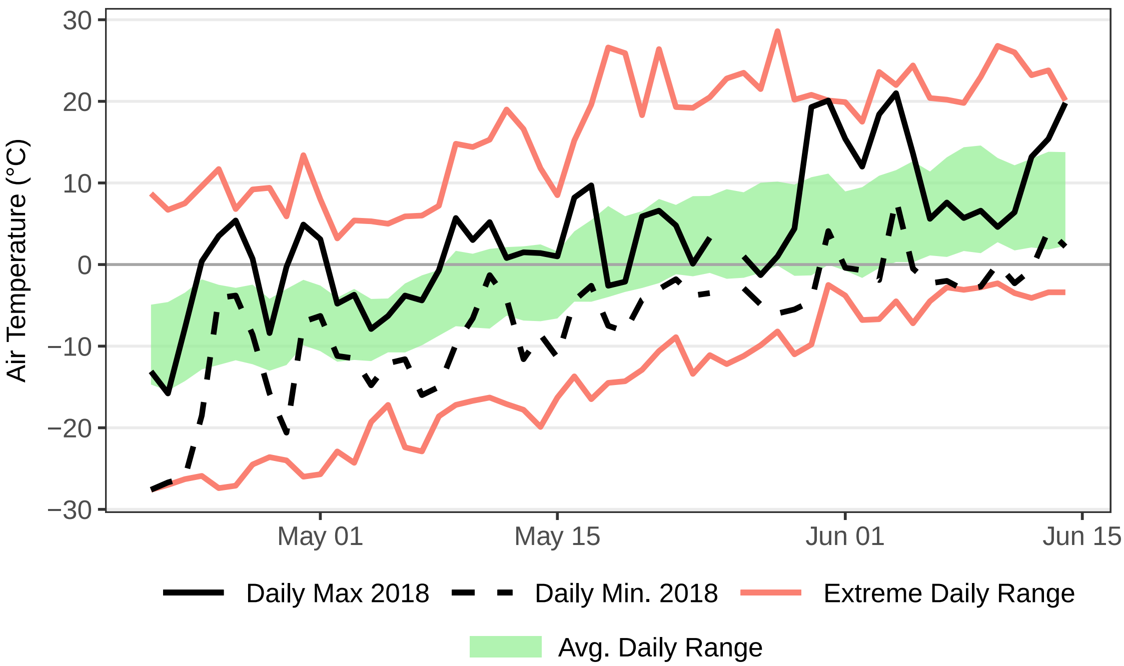

Figure 3. Maximum and minimum daily air temperature in comparison to the average and extreme values at the Trail Valley Creek field station for the period of 1999-2019.

3 Methods

3.1 Lake water and precipitation sampling for isotope analysis

Lake water samples isotope analysis were first collected from the 17 study lakes while they were ice covered (April 26 – May 1, 2018) and again soon after lakes became ice-free (June 15, 2018). Pre-snowmelt samples were obtained from a hole augured through the ice near the centre of each lake. These water samples were taken 10 cm below the water surface in the augured hole. Lake depth, snow depth on the ice, and ice thickness were recorded at the same time water samples were collected. Water samples were then collected post-snowmelt at the shore of each lake shortly after the lakes became fully ice-free. Isotope data were then used to estimate the portion of lake water that was replaced by freshet between the two sampling dates.

To estimate the Local Meteoric Water Line (LMWL) and the average isotope composition of precipitation (δ_P) in the study region, which are useful references for the interpretation of lake water isotope compositions, samples of end-of-winter snow on the ground in April 2018 and rainfall for the period May to September 2018 were obtained. Snow samples (n = 11) were collected from the study area by taking a vertical core of snow using a tube, completely melting the snow in a sealed plastic bag, and then filling a sample bottle with the meltwater. Rainfall (n = 13) was collected between May and September in Inuvik using a clean high-density polyethylene container, which was then transferred to a sample bottle shortly after the rain had stopped. The midpoint between the average isotope composition of snow samples and rain samples was calculated to define

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Table 2. Isotope, δ_I, and lake water replacement values for all lakes.

	Pre-Snowmelt Lake Water (01-05-2018)						Post-Snowmelt Lake Water (15-06-2018)				
Lake	$\delta^{18}O$	δ^2H	$\delta^{18}O$ Ice-Corrected	δ^2H Ice-Corrected	$\delta^{18}O_I$ Ice-Corrected	δ^2H_I Ice-Corrected	$\delta^{18}O$	δ^2H	$\delta^{18}O_I$	δ^2H_I	Average % lake water replaced
7	-15.41	-131.05	-14.07	-123.16	-22.84	-171.44	-15.48	-129.67	-21.11	-159.22	19.1
8	-15.94	-134.71	-14.7	-127.42	-24.68	-184.42	-16.71	-137.71	-22.9	-171.81	23.9
9	-17.17	-140.98	-12.02	-110.51	-17.94	-136.81	-15.45	-128.25	-19.78	-149.79	44.7
10	-15.68	-132.02	-13.54	-119.35	-20.84	-157.31	-14.86	-125.77	-20.29	-153.37	19.3
11	-20.56	-156.48	-18.48	-144.48	-20.47	-154.65	-19	-147.33	-20.59	-155.52	25.2
14	-19.85	-154.24	-17.23	-139.05	-21.3	-160.52	-18.66	-146.31	-21.09	-159.04	36.7
15	-18.17	-144.83	-14.35	-122.41	-19.35	-146.76	-16.03	-132.27	-20.82	-157.11	27.2
16	-20.57	-155.32	-18.99	-146.18	-19.96	-151.05	-19.05	-146.66	-20.05	-151.69	7.2
19	-18.92	-149.63	-16.85	-137.58	-21.86	-164.46	-18.26	-144.16	-21.05	-158.75	32.4
20	-19.32	-149.44	-17.44	-138.52	-20.02	-151.49	-17.86	-140.53	-19.96	-151.07	16.3
21	-19.61	-154.6	-16.33	-135.57	-22.83	-171.31	-17.72	-141.63	-21.29	-160.47	26.2
26	-17.5	-141.72	-12.59	-112.7	-17.47	-133.48	-16.15	-131.84	-19.87	-150.41	49.9
27	-17.95	-144.85	-16.48	-136.26	-22.68	-170.25	-17.79	-142.41	-21.65	-162.98	24.2
49	-14.72	-124.46	-13.12	-114.91	-17.44	-133.22	-14.75	-123.4	-18.51	-140.81	31.6
50	-14.99	-127.6	-12.8	-114.59	-18.86	-143.31	-15.43	-125.22	-17.7	-135.07	52.8
51	-16.28	-133.15	-14.95	-125.31	-19.33	-146.63	-15.8	-129.92	-19.72	-149.36	18.5
52	-13.98	-120.75	-13.29	-116.63	-18.51	-140.79	-13.59	-117.58	-18.03	-137.45	5.5
Min	-20.57	-156.48	-18.99	-146.18	-24.68	-184.42	-19.05	-147.33	-22.9	-171.81	5.2
Mean	-17.45	-140.93	-15.13	-127.33	-20.38	-154.00	-16.62	-134.74	-20.26	-153.17	27.1
Max	-13.98	-120.75	-12.02	-110.51	-17.44	-133.22	-13.59	-117.58	-17.7	-135.07	52.8

δ_P. All samples were collected in 30 mL high-density polyethylene bottles and were measured using Off-Axis Integrated Cavity Output Spectroscopy at the Environmental Isotope Laboratory at the University of Waterloo to determine the fraction of $^{18}O/^{16}O$ and $^2H/^1H$ in each sample. Isotope compositions are expressed in standard δ-notation, such that:

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

where R represents the ratio of $^{18}O/^{16}O$ or $^2H/^1H$, and VSMOW represents Vienna Mean Standard Ocean Water. Isotope values are normalized to Standard Light Antarctic Precipitation to $\delta^{18}O$ = -55.5‰ and δ^2H = -428‰; (Coplen, 1996). The analytical uncertainties were ±0.2‰ for $\delta^{18}O$ and ±0.8‰ for δ^2H. All isotope data from lakes is presented in Table 2.

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3.2 Estimating the replacement of lake water by freshet and lake source waters

The percentage of a lake's volume that has been replaced by a given water source can be estimated as follows:

$$\%\,lake\,water\,replaced = \frac{\delta_{L-Post} - \delta_{L-Pre}}{\delta_{I-Post} - \delta_{L-Pre}} * 100 \quad (2)$$

where $\delta_{\text{L-Pre}}$ is the lake isotope composition before snowmelt begins, $\delta_{\text{L-Post}}$ is the isotope composition of the lake after snowmelt is complete, and $\delta_{\text{I-Post}}$ is the isotope composition of the source water post-snowmelt. Application of this equation assumes minimal to no change in volume, which is reasonable given the lakes we samples are all open-drainage.

We calculated δ_{I} following the coupled isotope tracer approach outlined by Yi et al. (2008), using an isotope framework based on 2017 air temperature and humidity data for the typical ice-free period (June 15 – October 15) collected at the Trail Valley Creek meteorological station located 45 km NNE of Inuvik (Figure 2). The coupled isotope tracer approach assumes all lakes under the same meteorological conditions will evolve towards the same isotope composition (δ^*, the isotope composition of a lake at the moment of desiccation) as lakes evaporate along lake-specific evaporation lines. These lake-specific evaporation lines are defined by extrapolating from δ^* through δ_{L} until intersection with the Local Meteoric Water Line, which is used to estimate δ_{I} (Figure 4). We calculated δ_{I} for pre-snowmelt and post-snowmelt lake isotope compositions to identify whether the isotope composition of the source water changed after freshet. The percentage of lake water replaced was calculated using both δ^{18}O and δ^{2}H using Equation 2 and average values are reported. The average difference obtained using the two isotopes in the estimate of the percentage of lake volume replaced by runoff was minimal (1.8%). Details of the equations and variables used in the isotope framework are given in Appendix A.

As ice forms and preferentially incorporates water containing the heavy isotopes ^{18}O or ^{2}H, the lake water beneath the ice becomes increasingly depleted in ^{18}O and ^{2}H. Consequently, the water samples we collected pre-snowmelt were systematically isotopically depleted relative to pre-freeze-up lake water, and the magnitude of depletion depends on the fraction of lake water that had frozen into ice. We corrected $\delta_{\text{L-Pre}}$ for the fractionation of freezing water into ice using an equation developed by Gibson and Prowse (2002) that describes the fractionation of isotopes between water and freezing ice in a closed system:

$$\delta_{L-Pre} = -f^{\alpha_{eff}}(1000 * f^{\alpha_{eff}} - f * \delta_{L-BelowIce} - 1000 * f) \quad (3)$$

where $\delta_{\text{L-BelowIce}}$ is the isotope composition of the water beneath the lake ice, α_{eff} is the effective fractionation factor between ice and water, defined as $\alpha_{\text{eff}} = R_{\text{Ice}}/R_{\text{L}}$, and f is the fraction of unfrozen water remaining in the lake. α_{eff} is dependent on the thickness of the boundary layer between the forming ice and freezing water and the downwards freezing velocity of the ice. Since we did not have measurements of either of these variables, we relied on previously estimated values of α_{eff} (Souchez et al., 1987; Bowser and Gat, 1995; Ferrick et al., 2002) and boundary layer thickness (Ferrick et al., 1998, 2002; Gibson and Prowse, 2002). Using this information, we estimated values of α_{eff} that produced $\delta_{\text{L-Pre}}$ values that closely match lake water isotope compositions measured at the same lakes in August and September of 2018 (Figure B1). Additional information about the determination of α_{eff} values is provided in Appendix B.

To estimate fraction of unfrozen water remaining in lakes (f, Equation 3), bathymetry was collected at Big Bear Lake (Figure 2, a typical bowl-shaped thermokarst lake near the Trail Valley Creek meteorological station in June 2017 using a

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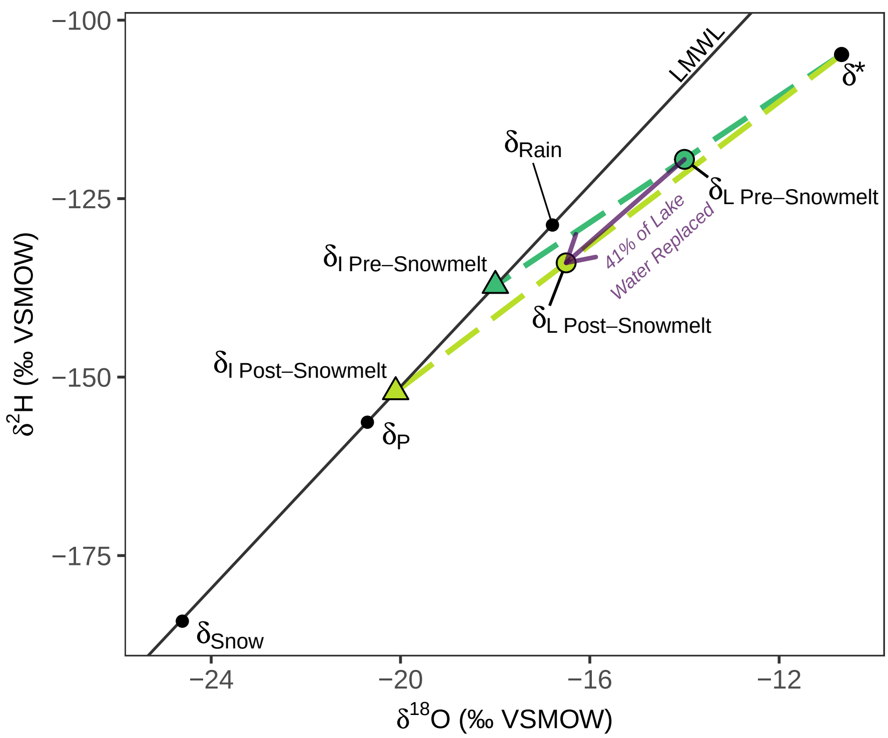

Figure 4. A hypothetical change in lake isotope composition from pre-snowmelt to post-snowmelt is shown. A visualization of how δ_{I} is calculated for an individual lake using a lake-specific evaporation line for both pre-snowmelt and post-snowmelt is also shown, where each lake's evaporation line (dashed line) extrapolates from δ^{*} through δ_{L} until intersecting the Local Meteoric Water Line to give δ_{I}. The Local Evaporation Line (LEL) is defined by the line between δ_{P} and δ^{*} (not shown).

Garmin echoMAP CHIRP 42dv fish finder. Bathymetry data was used to determine a relationship between lake volume and lake depth. We fit a quadratic equation to the bathymetric data to estimate the fraction of lake volume relative to the fraction of lake depth. The best fit quadratic equation (r^2 = 0.9997) was:

145 $$V_{Lake} = -0.0115D_{Lake}^2 + 2.1508D_{Lake} - 0.4857 \quad (4)$$

where V_{lake} is the fraction of total lake volume and D_{lake} is the fraction of total lake depth. However, this fitted equation does not reach 100% V_{lake} at 100% D_{lake}, or 0% V_{lake} at 0% D_{lake}, which is required to realistically represent the relationship between lake depth and lake volume. The equation was slightly adjusted to:

$$V_{Lake} = -0.001D_{Lake}^2 + 2D_{Lake} \quad (5)$$

150 in order to satisfy these requirements, resulting in a mean offset of 1.7% between the measured bathymetric data and the adjusted equation. Most lakes in this region have a bowl-shaped bathymetry because they were formed through thermokarst processes (Rampton, 1988; Burn and Kokelj, 2009), where subsidence caused by the thaw of ice-rich permafrost results in a

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waterbody which then expands outward radially in all directions. Bathymetric data for Big Bear Lake and a comparison of the equation between lake volume and depth are provided in Appendix C.

3.3 Quantifying lake and watershed properties

We quantified multiple lake and watershed properties to explore relations with the amount of lake water replaced by freshet. These properties included lake depth, lake volume, snow depth on the lake, ice thickness, lake area and watershed area. Lake depth, snow depth on the lake and ice thickness were measured at the same time as pre-snowmelt lake samples were collected. Lake volume was approximated by multiplying the product of lake depth and lake area by 0.7, which matched with the measured lake volume of Big Bear Lake. Watershed area was estimated by applying the D8 water routing algorithm (O'Callaghan and Mark, 1984) to the 2 metre resolution ArcticDEM (PGC, 2018) using ArcGIS 10.7.1 (ESRI, 2019).

4 Results

Correcting for ice fractionation using equation 3 resulted in an increase in estimated $\delta_{\text{L-Pre}}$ values as expected, with the median shifting from -17.50‰ (-19.32‰ to -15.68‰ IQR, inter-quartile range) to -14.70‰ (-16.85‰ to -13.29‰ IQR) for δ^{18}O (Figure 5a, Table 2). The corrected pre-snowmelt lake isotope compositions were distributed across a large range of the predicted LEL, spanning from near the LMWL to near δ^{*} (Figure 5a), reflecting that the lake waters were variably influenced by evaporation. Corrected pre-snowmelt lake isotope compositions also tightly cluster along the LEL, indicating that the predicted LEL is well characterized.

The change in lake isotope composition from pre-snowmelt to post-snowmelt was characterized by a small (~1.5‰ in δ^{18}O) shift towards δ_{P}, with median pre-snowmelt $\delta_{\text{L-Pre}}$ values of -14.70‰ (-16.85‰ to -13.29‰ IQR) and median $\delta_{\text{Lake-Post}}$ values of -16.15‰ (-17.86‰ to -15.45‰ IQR) for δ^{18}O (Figure 5b). The small change in lake isotope composition meant that most lakes retained an evaporated isotope signature post-snowmelt, overlapping with a substantial portion of $\delta_{\text{L-Pre}}$ and continuing to plot along the LEL (Figure 5b). Post-snowmelt, about half of the lakes (9 of 17) also plotted above the LEL, indicating that the δ_{I} of these lakes was more similar to rainfall than snowfall (Figure 5b). The shift in δ_{I} for lakes from pre-snowmelt to post-snowmelt shows a convergence of most δ_{I} values towards a value near δ_{P} and away from the isotope composition of the end-of-winter snow (δ_{Snow}) or rainfall (δ_{Rain}) (Figure 5c). The convergence of δ_{I} values towards δ_{P} and away from end-of-winter snow signify that a non-snow source of water, with a higher isotope composition than δ_{Snow}, was present in freshet.

Replacement of lake water by freshet ranged widely from 5.2 – 52.8%, with a median of 25.2% (19.1% to 32.4% IQR, Figure 6). A substantial proportion of this variation was explained by lake depth: deeper lakes had significantly less of their water replaced by freshet, with a reduction in lake water replacement of 13% for each additional metre of lake depth (R^2 = 0.53, $p < 0.001$, Figure 6, Table 3). Lake water replacement was not independently correlated with any other lake or watershed attribute including watershed area, lake volume, snow depth on the lake ice, lake ice thickness, lake area, and the ratio of lake area to watershed area (Table 3).

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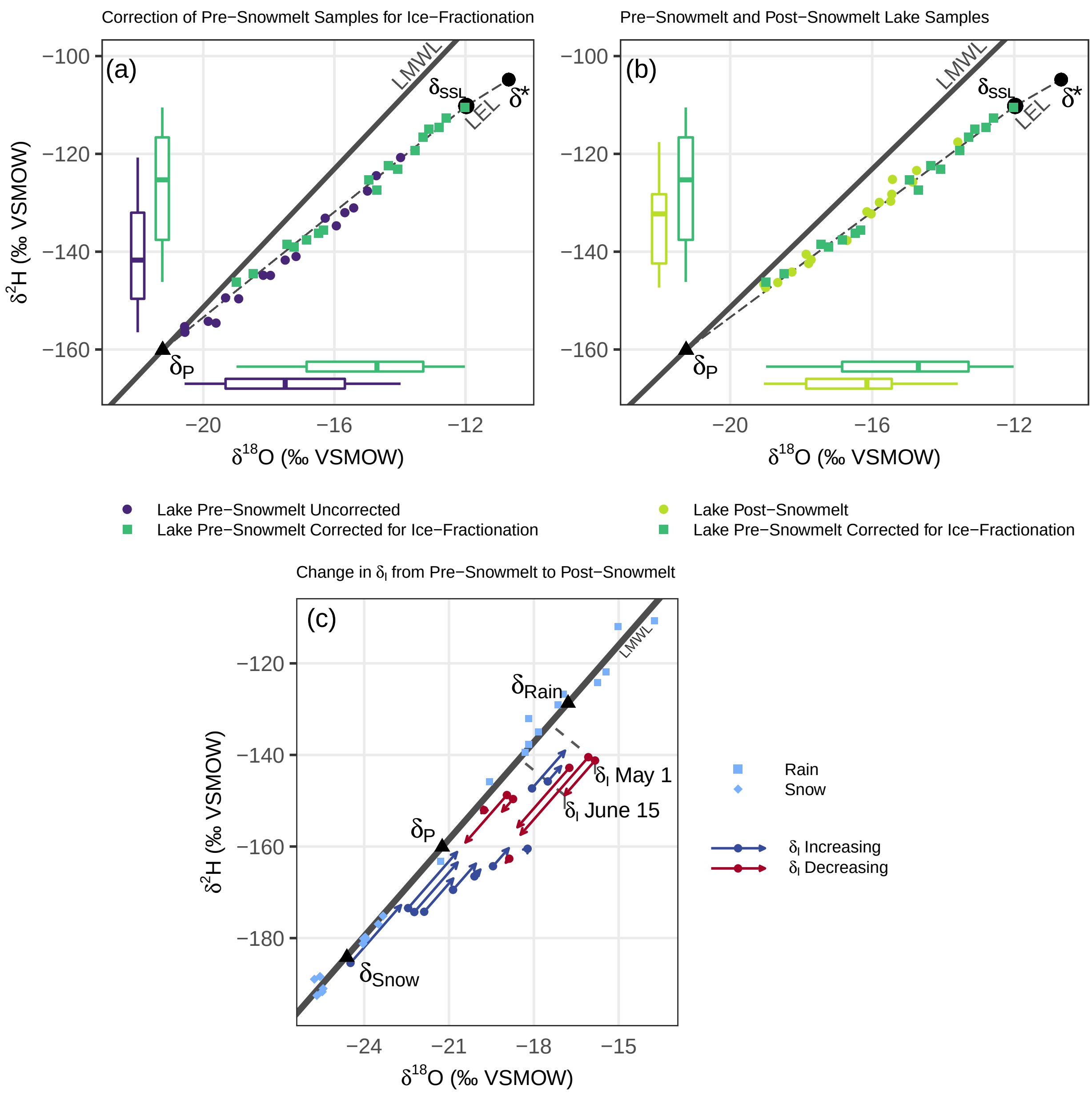

Figure 5. Lake water and precipitation isotope data are displayed on δ^{18}O-δ^{2}H graphs. The Local Meteoric Water Line (LMWL: δ^{2}H = 7.1*δ^{18}O – 10.0) is indicated by the solid line, while the Local Evaporation Line (LEL: δ^{2}H = 5.2*δ^{18}O – 48.9) is indicated by the dashed line. δ_P represents the average value of precipitation in the region, based on 2018 sampling of end-of-winter snow and rainfall from April to September. (a) Uncorrected and corrected for ice fractionation pre-snowmelt lake isotope data. (b) Corrected pre-snowmelt and post-snowmelt data. (c) The shift in δ_I from pre-snowmelt to post-snowmelt, as indicated by a circle for pre-snowmelt δ_I values and the end of the arrow for post-snowmelt δ_I values. All δ_I values are offset from the LMWL for visibility, as all δ_I values are constrained to the LMWL.

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Table 3. Results for a linear regression between total lake water replacement with multiple lake and watershed properties. The adjusted R^2 and p-value are shown for each isotope. Linear regressions were performed using the 'lm' function using R 4.0.2 (R Core Team, 2021)

	% Lake Volume Replaced by Freshet)	
Lake Attribute (unit)	Adjusted R^2	p-value
lake depth (m)	0.53	<0.001
watershed area (m^2)	0.02	0.274
lake volume (m^3)	-0.03	0.486
snow depth (cm)	-0.06	0.771
ice thickness (m)	-0.05	0.654
lake area (m^2)	-0.06	0.849
watershed area/lake area	-0.01	0.361

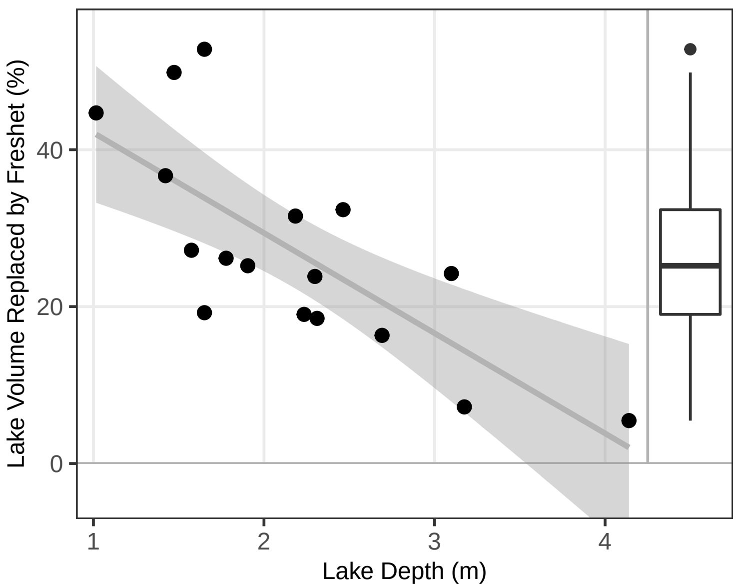

Figure 6. The relationship between the amount of lake water replaced by freshet and lake depth. The distribution of lake water replacement by freshet is shown by the boxplot on the right side of the plot. A linear regression is also displayed on the plot ($R^2 = 0.53$, $p < 0.001$). The shaded grey area represents the 95% confidence interval of the linear regression.

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5 Discussion

5.1 Influence of snowmelt bypass on the replacement of lake water by freshet

Characterization of the influence of snowmelt bypass required accurate determination of lake isotope compositions prior to freeze-up. Given that lake isotope samples are unavailable from Autumn 2017, and $\delta_{\text{L-Pre}}$ values were instead obtained from drilling through the lake ice before the lakes became ice-free, their isotope compositions required correction for the isotope fractionation caused by ice formation. Our novel approach to correcting $\delta_{\text{L-Pre}}$ values for the fractionation caused by lake ice formation provides a reasonable estimate of δ_{L} prior to lake ice formation. While our correction of $\delta_{\text{L-Pre}}$ involves some uncertainty, such as having to estimate the relationship between lake depth and lake volume, corrected $\delta_{\text{L-Pre}}$ values closely align with the general distribution of water isotope measurements from August and September 2018 of the same lakes (Appendix B1). Corrected $\delta_{\text{L-Pre}}$ values are also situated near or above the LEL, reasonably indicating a more rainfall-sourced δ_{I} that would be present in lakes at the time of freeze-up during the previous autumn. Prior to correction, most $\delta_{\text{L-Pre}}$ values plotted below the LEL (Figure 5a), indicating lakes had the majority of their inflow sourced from snow, which would be unlikely at the time when lake ice began forming during the previous autumn. We considered using δ_{L} values from September 2018 instead of correcting for ice fractionation, but 2018 was a cooler and wetter year than 2017, meaning the lake-specific δ_{L} values in September 2018 likely differed somewhat from September 2017.

The presence of a uniformly thick layer of freshet beneath lake likely explains the relationship between lake depth and the amount of lake water replaced by runoff (Figure 6). Previous studies have measured the thickness of the snowmelt bypass layer at the onset of freshet inflow to be ~100 – 200 cm (Henriksen and Wright, 1977; Bergmann and Welch, 1985). Since the mixed layer of pre-snowmelt lake water and freshet comprised a relatively larger volume in shallower lakes compared to deeper lakes (Figure 7a), a larger portion of lake water was able to be replaced with freshet in shallower lakes than in deeper lakes. Shallower lakes likely had colder lakebed temperatures, which allowed more mixing between pre-snowmelt lake water and freshet inflow due to the reduction in water density gradient between the bottom of the lake and the top of the lake. To our knowledge, the relationship between lake depth and freshet retention has not been described in previous literature, although estimates of freshet recharge in more than one lake are scarce (Falcone, 2007; Brock et al., 2008). We expect that a similar relationship between lake depth and snowmelt bypass could be present in other open-drainage lakes that experience snowmelt bypass, since the relationship between increasing lake depth and greater snowmelt bypass is caused by the typical water temperature gradient which is present in ice-covered lakes at the onset of freshet.

5.2 Sources of freshet

Following the freshet, the δ_{I} of lakes did not shift towards the isotope composition of snow (δ_{Snow}) as one may expect, but instead shifted towards the average isotope composition of precipitation (δ_{P}, Figure 5c). Other than the 21.3 mm rainfall that fell during the six-week period between the two isotope sampling dates, the only other potential source of water during this time period is water stored in the active layer, which mixed with snowmelt runoff as the soil thawed throughout the spring. The high infiltration capacity of the peat channels that convey runoff causes nearly all snowmelt to flow through subsurface routes,

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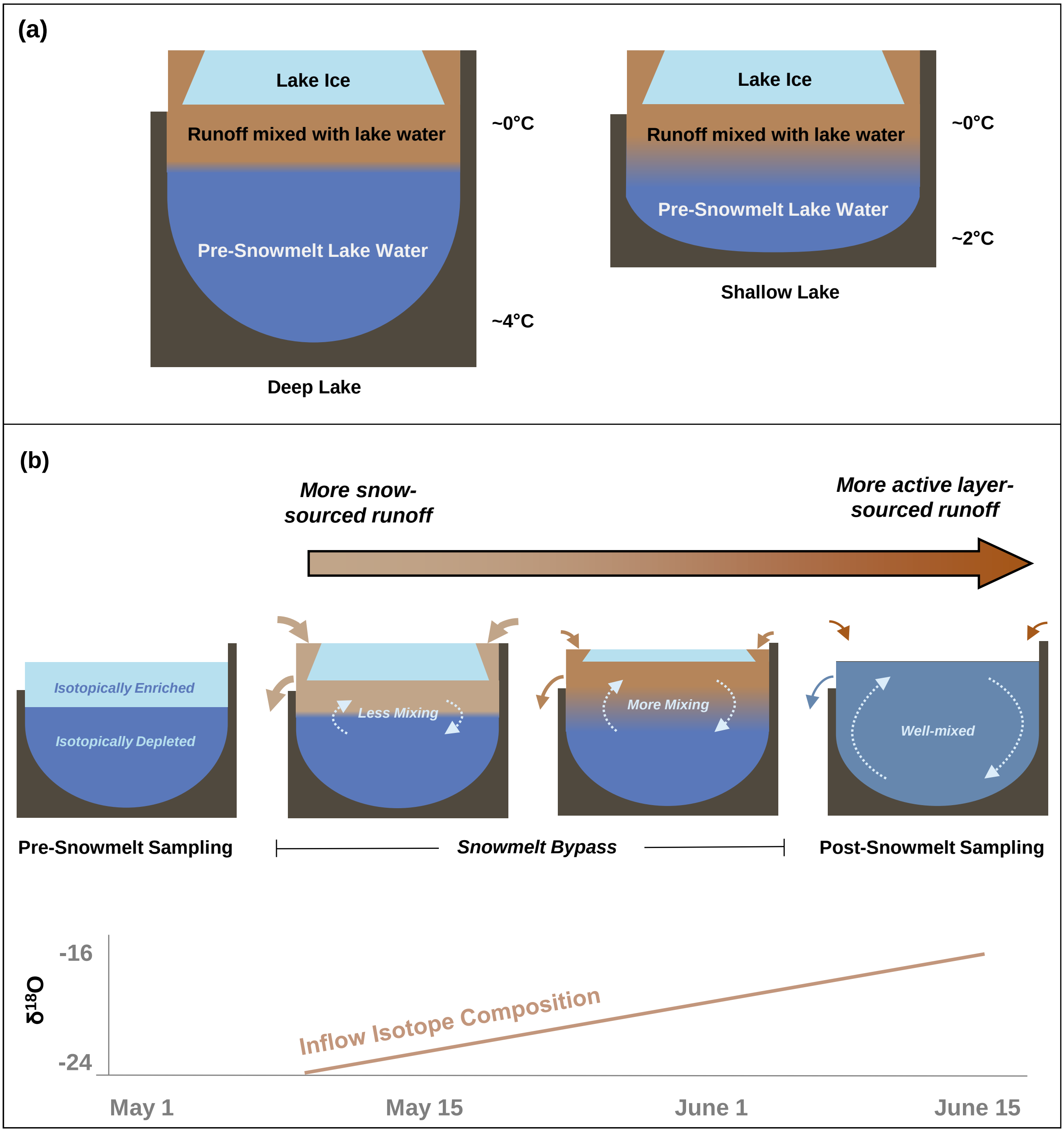

Figure 7. (a) A conceptual model showing the relative differences in snowmelt bypass between a shallow lake and a deep lake. Shallower lakes have a larger portion of their volume replaced by the runoff layer that flows beneath the ice, while a larger portion of water is isolated from mixing with runoff in deeper lakes. (b) A conceptual model of how snowmelt bypass occurred over the course of the snowmelt period. Pre-snowmelt samples were taken from water beneath lake ice which was isotopically depleted in comparison to the lake ice. As time goes on, the source of freshet shifts from snow-sourced to active layer-sourced, while mixing increases beneath the lake ice simultaneously.

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as was observed in the field and reported by Quinton and Marsh (1998). As water near the surface of the active layer is most likely to be comprised of rainfall from the previous year, we expect that much of the water stored in the top of the active layer would have been largely sourced from rainfall from the previous summer and autumn. In support of this inference, Tetzlaff et al. (2018) reported δ^2H values between -140‰ and -160‰ from August to September of 2014 in water samples taken at 10 cm soil depth at Siksik Creek, a watershed directly adjacent to the Trail Valley Creek camp (Figure 2). This range of soil water δ^2H values is higher relative to δ_{P} (δ^2H = -160.1 ‰, Figure 5c), indicating that a mixture of snowmelt runoff with this active layer water could result in a water source similar in isotope composition to δ_{P}.

Freshet flowing into lakes later during the snowmelt likely had a more rainfall-sourced isotope composition, replacing the more snow-sourced runoff that had entered the lake earlier during the freshet (Figure 7b). A shift from snow-sourced water towards rainfall-sourced water during the course of the snowmelt period has been observed using water isotope measurements at Siksik Creek by Tetzlaff et al. (2018). Additionally, the mixing of freshet beneath lake ice increases with time as the temperature and density gradient lessens between the top and bottom of the lake water column (Cortés et al., 2017). Based on our results and these previous studies, we conclude that the ability of the active layer to contribute runoff to lakes appears to be maximized at the same time that vertical mixing in the lake is stronger, while snowmelt runoff flows into lakes at a time when little vertical mixing is occurring and is also likely replaced by later runoff (Figure 7b). Such interplay between timing of snowmelt runoff, lake ice melt and hydrological behaviour of the active layer explains why the source of water to lakes is not solely snow-sourced, and that incorporation of active layer runoff into lakes is more important than the volume of freshet delivered to lakes, for the open-drainage lakes of this study.

5.3 Uncertainties and improving the estimation of pre-ice formation lake isotope compositions

In our estimation of lake water replacement by freshet we had to make some assumptions (Appendix B) when estimating $\delta_{\mathrm{L\text{-}Pre}}$ using Equation 3. Future studies could sample lakes in the previous autumn before ice formation begins to avoid these assumptions, as minimal hydrological activity occurs over the winter months at arctic lakes due to frozen soils and ice cover on lakes (Woo, 1980). If lakes cannot be sampled in the previous autumn, another option would be to take a lake ice core and sample the isotope composition at different points along the lake ice core. These isotope measurements could then be used to estimate the α_{eff} value used in Equation 3, as has been done by Souchez et al. (1987) and Bowser and Gat (1995). This approach would avoid the assumptions we made in estimating α_{eff} outlined in Appendix B. However with this approach, one still needs to know the volume of the lake ice relative to the volume of the remaining unfrozen water, and must rely on lake bathymetry data or a depth–to–volume relationship, such as the one we derived using bathymetry data from Big Bear Lake (Equation 4). Since we only have one survey of lake bathymetry, we do not know how well our depth–to–volume relationship fits to other lakes in the region, and it could be that this relationship varies as lakes increase in surface area or in areas of different surficial geology where thermokarst processes were stronger or weaker during lake formation.

Since we do not have measurements of lake temperature, we also assume our lakes have the typical thermal structure of ice-covered lakes that leads to snowmelt bypass. Even though snowmelt bypass is a common phenomenon in many types of ice–covered lakes around the world (Henriksen and Wright, 1977; Jeffries et al., 1979; Hendrey et al., 1980; Bergmann

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and Welch, 1985; Schiff and English, 1988; Edwards and McAndrews, 1989; Cortés et al., 2017), knowledge about how the thermal structure of our study lakes evolved over time, and varied between lakes of different depth, would have helped us better understand our results. Such data could have helped better explain why shallow lakes retain more freshet runoff than deeper lakes, and also could have helped confirm our hypothesis that water flowing into lakes later during the freshet mixes more readily with lake water.

5.4 Climate change and snowmelt bypass

Whether climate change allows lakes to continue being open drainage or causes lakes to become closed drainage will be an important distinction, given that snowmelt bypass can only occur when lakes are at or above their outlet level. There are multiple consequences of Arctic warming that will influence lake water balance: increases in rainfall (Bintanja and Andry, 2017) and snowfall (Brown and Mote, 2009; Ernakovich et al., 2014), increases in active layer thickness (Walvoord and Kurylyk, 2016; Tananaev and Lotsari, 2022; Koch et al., 2022), the proliferation of deciduous shrubs (Loranty et al., 2018), and longer lake ice-free periods (Woolway et al., 2020). Whether the combination of these changes will result in an increase or decrease in runoff to lakes is currently unknown (Blöschl et al., 2019), making it difficult to predict whether lakes will remain open drainage or if some lakes may shift to being closed drainage under future climate. Due to this uncertainty, we discuss potential future changes in snowmelt bypass under two scenarios: a) where lakes remain open drainage in the future, and b) where some lakes become closed drainage in the future.

If lakes are to remain open drainage in the future, we suspect the freshet that is incorporated into lakes may shift towards being more rainfall-sourced. Rainfall increases will likely leave the active layer in a more saturated state when the active layer freezes in autumn, potentially providing more water to lakes during the freshet. Other studies have already established a strong positive relationship between increased rainfall in the previous summer and a more efficient conversion of the snowpack into freshet (Bowling et al., 2003; Stuefer et al., 2017), indicating the importance of interactions between snowmelt runoff and water stored in the thawing active layer. Increasing shrub heights will advance snowmelt timing relative to lake ice breakup (Marsh et al., 2010; Wilcox et al., 2019), causing more snowmelt to flow into lakes at a time when there is limited below-ice vertical mixing. Combining earlier snowmelt timing with a more rain-saturated active layer could result in more freshet bypassing open-drainage lakes early during the freshet, with the active layer thawing deeper and shifting freshet more towards rainfall-sourced water by the time below-ice mixing begins.

We expect any lakes that become closed drainage will retain more freshet runoff than comparable open drainage lakes, because closed drainage lakes will retain any freshet that is required to recharge the lake to its outlet level. Since snowmelt bypass cannot occur until a closed drainage lake is recharged to its outlet level, we expect that freshet retention by closed drainage lakes will not be as influenced by lake depth. Lakes with smaller ratios of watershed area to lake area (WA/LA) are more prone to a more negative water balance (Marsh and Pomeroy, 1996; Gibson and Edwards, 2002; Turner et al., 2014; Arp et al., 2015; Wilcox et al., in prep.). Therefore, we expect lakes with relatively small WA/LA will be more prone to becoming closed drainage, relying on freshet to recharge them to their outlet level and retain more freshet as a result. A corollary of this prediction is that other ice-covered lakes which currently lie below their outlet level at the onset of the freshet (i.e., closed-

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drainage lakes) likely retain more freshet than open-drainage lakes of a similar lake depth. A more saturated active layer at the onset of snowmelt, combined with a greater amount of snowfall should increase the ability of freshet to recharge any closed-drainage lakes.

6 Conclusions

The large volume of freshet that flows into lakes every year is likely to bypass ice-covered, open-drainage lakes due to limited mixing between lake water beneath the lake ice and freshet. By estimating the percentage of lake water replaced by freshet at 17 lakes, we have been able to explore which lake and watershed attributes affect snowmelt bypass. Our data show that as lake depth increases the amount of lake water replaced by freshet decreases, because freshet is unable to mix with deeper lake water when lakes are ice-covered and the water column is stratified. Additionally, the volume of freshet flowing into the lakes seems to have no observable impact on the amount of lake water replaced by freshet. Estimation of the isotope composition of source water showed that the freshet remaining in lakes was not solely snow sourced – rainwater left in the active layer from the previous autumn had mixed with snowmelt before it entered lakes. Active layer-sourced water likely flows into lakes later on in spring and at a time when freshet can more easily mix with pre-snowmelt lake water, replacing the earlier more snow-sourced freshet.

Relationships observed between lake depth and lake biogeochemistry in ice-covered lakes may be explained by the differences in snowmelt bypass at lakes of different depth. Previous studies have observed that snowmelt runoff has a unique geochemical composition (Marsh and Pomeroy, 1999) that impacts lake biogeochemistry (Balasubramaniam et al., 2015); future research could investigate how the varying degrees of snowmelt bypass affects lake biogeochemistry. Research in Old Crow Flats, Yukon, identified that subarctic lakes with predominantly snow-sourced water have lower pH, lower specific conductivity, and higher concentrations of dissolved organic carbon (Balasubramaniam et al., 2015).

Models specialized for northern environments are rapidly improving their ability to represent the complicated processes present in permafrost regions, such as the effect of shrubs on snow accumulation, snowmelt and active layer thickness (Krogh and Pomeroy, 2019; Bui et al., 2020), lake ice formation and decay (MacKay et al., 2017) and the mixing processes that lead to snowmelt bypass (MacKay et al., 2017). Such models could be used to examine how freshet water sources may change in the future, which could have significant impacts on lake biogeochemistry (Finlay et al., 2006; Balasubramaniam et al., 2015). Additionally, current physically-based lake models can represent vertical mixing beneath lake ice (MacKay et al., 2017), and could be used to further evaluate the influence of lake depth or climate change on snowmelt bypass and resulting impacts on lake biogeochemistry.

Data availability. The data used in the paper are presented in tables in the manuscript and Appendix A and C. Isotope data and lake and watershed attribute data can be downloaded from the Trail Valley Creek Research Station Dataverse at https://doi.org/10.5683/SP3/AZE4ER.

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Appendix A: Isotope framework

Table A1. Variables used in isotope framework and sources of their calculation.

Parameter	Value	Reference
δ^* (‰)	δ^{18}O = -10.77, δ^2H = -104.97	(Gonfiantini, 1986)
h (%)	80.5	(Environment and Climate Change Canada, 2019)
T (K)	282.32	(Environment and Climate Change Canada, 2019)
α^*_{L-V}	^{18}O =1.0108, ^{2}H = 1.0981	(Horita and Wesolowski, 1994)
ε^*	^{18}O = 0.0108, ^{2}H = 0.0981	(Horita and Wesolowski, 1994)
ε_k	^{18}O = 0.0028, ^{2}H = 0.0024	(Gonfiantini, 1986)
δ_{Rain} (‰)	δ^{18}O = -16.79, δ^2H = -128.7	This study.
δ_{Snow} (‰)	δ^{18}O = -24.61, δ^2H = -184.2	This study.
LMWL Slope, Intercept (‰)	7.066, δ^2H = -10.0	This study.
LEL Slope , Intercept	5.114, 48.9	This study.

The isotope framework (i.e., establishment of the predicted Local Evaporation Line (LEL)) used for this study was based on the coupled isotope tracer method developed by Yi et al. 2008, following other studies that have investigated lake water balances using water isotope tracers (Turner et al., 2014; Remmer et al., 2020; MacDonald et al., 2021). Below are the variables and equations required to calculate δ^*, the terminal point on the LEL. The equation for δ^*, which represents the isotope composition of a lake at the point of desiccation, is as follows (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 (Horita and Wesolowski, 1994), calculated for δ^{18}O as:

$$\alpha^*_{L-V} = 2.718^{(-7.685+6.7123*\frac{10^3}{T}-1.6664*\frac{10^6}{T^2}+0.35041*\frac{10^9}{T^3})/1000} \tag{A2}$$

and calculated for δ^{18}O as:

$$\alpha^*_{L-V} = 2.718^{(1158.8*\frac{T^3}{10^9}-1620.1*\frac{T^2}{10^6}+794.84*\frac{(}{T}10^3-161.04+2.9992*\frac{10^9}{T^3})/1000} \tag{A3}$$

The term ε^* is a separation term where:

$$\varepsilon^* = \alpha^* - 1 \tag{A4}$$

The term h represents the relative humidity of the air above the water and δ_{As} is the isotope composition of atmospheric moisture during the open water season defined as:

$$\delta_{As} = \frac{\delta_{Ps} - \varepsilon^*}{\alpha^*} \tag{A5}$$

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where δ_{Ps} is the average isotope composition of precipitation (i.e., rainfall) during the open water season. The term ε_k is the kinetic fractionation separation term, defined as

$$\varepsilon_k = x * (1-h) \tag{A6}$$

335 where x = 0.0142 for δ^{18}O and x = 0.0125 for δ^2H (Gonfiantini, 1986).

Appendix B: Determination of α_{eff} values

In order to determine α_{eff}, two variables must be taken into account: the thickness of the ^{18}O or ^{2}H boundary layer across which heavy isotopes are diffusing from water into ice, and the downward velocity of the freezing ice (Ferrick et al., 2002). If these two variables are known, the fractionation factor can be estimated using a linear resistance model developed by Ferrick et al. 1998, which is similar in structure to the Craig and Gordon (1965) linear resistance model for evaporation. Ferrick et al. 1998 define the effective fractionation factor between ice and water as:

$$\alpha_{eff} = \frac{\alpha^*_{L-S}}{\alpha^*_{L-S} + (1-\alpha^*_{L-S})exp[\frac{-zv}{D_i}]} \tag{B1}$$

where α^*_{L-S} is the equilibrium fractionation factor between ice and water (1.002909 for ^{18}O/^{16}O, 1.02093 for ^{2}H/^{1}H (Wang and Meijer, 2018)), z is the ^{18}O or ^{2}H boundary layer thickness between the ice and water (mm), v is the velocity of ice growth (cm^2 day^{-1}), and D_i is the self-diffusion coefficient of $^1H_2{}^{18}O$ or $^1H^2H^{16}O$ at 0°C (cm^2 day^{-1}). As the boundary layer and the velocity of ice growth increase, α_{eff} moves from the value of α^*_{L-S} towards a value of 1 (no fractionation).

As we do not know the boundary layer thickness at the ice-water interface, or the exact ice growth velocity for the lakes studied here, we relied on multiple other sources of information to estimate a probable upper and lower bound of α_{eff}. We took into account previous estimates of α_{eff} for ice-water fractionation (Souchez et al., 1987; Bowser and Gat, 1995; Ferrick et al., 2002) and boundary layer thickness from other studies of lakes (Ferrick et al., 1998, 2002; Gibson and Prowse, 2002). The boundary layer thickness between water and freezing ice in a lake was estimated to be between 1 mm and 6 mm by Ferrick et al. 1998, however, they revised this estimate with a more rigorous diffusion model to 1 ±0.3 mm for ^{18}O and 0.4 ± 0.2 mm for ^{2}H (Ferrick et al., 2002). They also found that the boundary layer thickness remained mostly stable across different ice growth velocities, although the lowest ice growth velocity of ~0.9 cm day^{-1} had a boundary layer of ~1.8 mm (Ferrick et al., 2002). The mean ^{18}O α_{eff} values for two ice cores taken from the lake studied by Ferrick et al. 2002 were 1.0021 and 1.0020, with respective ice growth velocities of 3.7 and 4.1 cm day^{-1}. A 1 mm boundary layer was also estimated by Gibson and Prowse (2002) beneath river ice in northern Canada, however they also suggest that the boundary layer thickness can reach up to 4 mm in quiescent lake water. Therefore, we assume a minimum boundary layer thickness of 1 mm, and a maximum boundary layer thickness of 4 mm.

We estimated the minimum possible freezing velocity of our study lakes using the initial date of ice formation and the ice thickness we measured in spring. Based on Sentinel imagery (Sentinel Playground) all studied lakes became ice-covered by October 16th, 2017. Ice thickness was measured at Big Bear and Little Bear Lake (near Trail Valley Creek camp, Figure 1)

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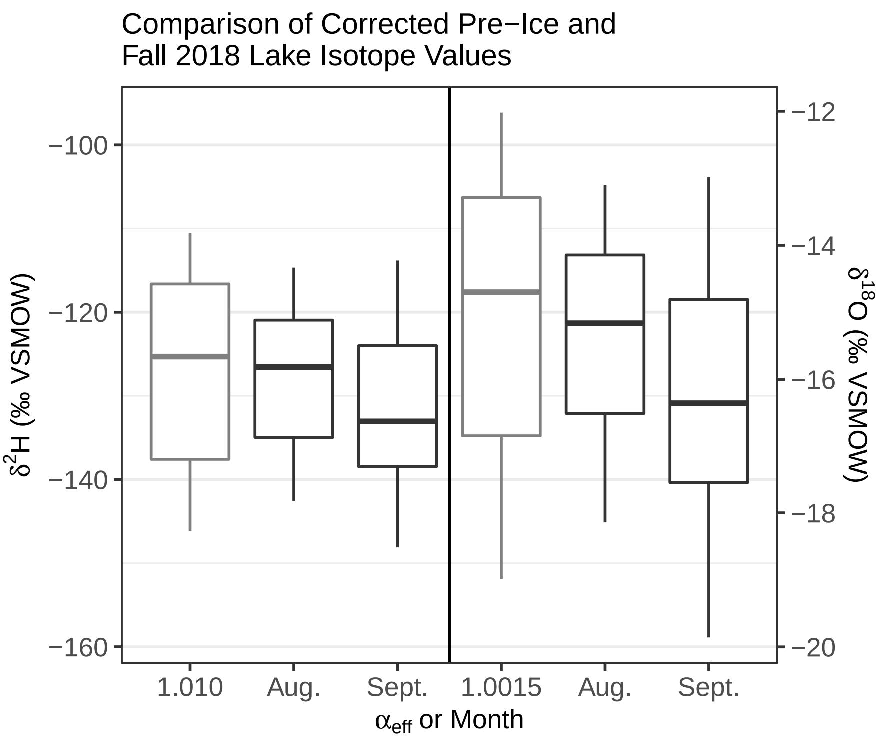

Figure B1. Comparison of Pre-Lake Ice Formation and August / September 2018 lake isotope compositions. The α_{eff} values (δ^2H α_{eff} = 1.010, δ^{18}O α_{eff} = 1.0015) match closely with August and September 2018 lake isotope compositions. This suggests that these α_{eff} estimates are appropriate to use for estimating the pre-ice formation lake water isotope compositions

on March 21st, 2018, and when ice thickness was remeasured again in late April, it had not become any thicker. Therefore, assuming ice growth began on October 16, 2017 and ceased on March 21, 2018, the ice growth velocities for our study lakes range from an average of 0.50 – 0.84 cm day^{-1} (0.78 – 1.32 m ice thickness). This only provides a lower bound estimate for ice growth velocity, as ice growth likely stopped earlier than March 21, 2018, and was more rapid during initial ice formation.

We further constrained our estimate of α_{eff} by assuming that α_{eff} values that result in lake water replacement estimates of >100% or <0% were not correct. Using all these sources of information, we calculated an upper bound of α_{eff} values based on the minimum possible ice freezing velocity (^{2}H α_{eff} = 1.0199, ^{18}O α_{eff} = 1.00286) and lower bound of α_{eff} values which still generate lake water replacement estimates that are > 0% (^{2}H α_{eff} = 1.010, ^{18}O α_{eff} = 1.0015).

Assuming a 2 mm boundary layer, which is within the range of our boundary layer thickness estimates, α_{eff} values of 1.0015 for ^{18}O and 1.010 for ^{2}H correspond to ice growth rates of 3.62 cm day-1 for ^{18}O and 3.34 cm day^{-1} for ^{2}H. Similar ice growth have been observed in Arctic lakes (Woo, 1980); greater ice growth rates were also estimated for a lake in a warmer climate by Ferrick et al. (2002). These α_{eff} values also compare well with other measured α_{eff} values in lakes: a range of α_{eff} = 1.013 – 1.015 for ^{2}H was found by Souchez et al. (1987) for a 4.4 cm thick lake ice cover; α_{eff} has been found to range from 1.0005

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to 1.0027 for ^{18}O in a single 50 cm ice core (Bowser and Gat, 1995). The $\delta_{\text{L-Pre}}$ values calculated using α_{eff} = 1.0015 for ^{18}O and 1.010 for ^{2}H also closely match the distribution of δ_{L} values from August and September of 2018, giving an indication that these α_{eff} values are realistic isotopic concentrations found in our study lakes (Figure B1). Therefore, we chose α_{eff} = 1.0015 for ^{18}O and α_{eff} = 1.010 for ^{2}H as our α_{eff} values, as they correspond well with other estimates α_{eff}, are within a range of probable ice-growth rates and lake water replacement by freshet and generate pre-ice formation isotope compositions that closely match the following summer's lake isotope composition.

Appendix C: Bathymetric data and volume – depth relationship

Table C1. Big Bear Lake bathymetric data and fit between modelled relationship between lake volume and lake depth as a percentage of total lake volume and depth.

Depth (m)	Cumulative Depth (% total)	Cumulative Volume (m^3)	Cumulative Volume (% total)	Modelled Cumulative Volume (% total)	Offset between data and modelled cumulative volume (%)
2.5	100	89326.83	100	100	0.00
2.25	90	88327.68	98.88	99	-0.12
2	80	85881.37	96.14	96	0.14
1.75	70	82441.90	92.29	91	1.29
1.5	60	77491.01	86.75	84	2.75
1.25	50	69384.35	77.67	75	2.67
1	40	58991.30	66.04	64	2.04
0.75	30	46548.63	52.11	51	1.11
0.5	20	32463.34	36.34	36	0.34
0.25	10	16944.40	18.97	19	-0.03
0	0	0	0	0	0.00

The modelled relationship between lake depth and volume is:

$$V_{lake} = (-0.01D_{lake})^2 + 2D_{lake} \tag{C1}$$

where V_{lake} is the cumulative lake volume as a percent of total and D_{lake} is the cumulative lake depth as a percent of total (Table C1).

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

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