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# Permafrost and lakes control river isotope composition across a boreal Arctic transect in the Western Siberian lowlands

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Supplementary material for this article is available online

## Abstract

The Western Siberian Lowlands (WSL) store large quantities of organic carbon that will be exposed and mobilized by the thawing of permafrost. The fate of mobilized carbon, however, is not well understood, partly because of inadequate knowledge of hydrological controls in the region which has a vast low-relief surface area, extensive lake and wetland coverage and gradually increasing permafrost influence. We used stable water isotopes to improve our understanding of dominant landscape controls on the hydrology of the WSL. We sampled rivers along a 1700 km South–North transect from permafrost-free to continuous permafrost repeatedly over three years, and derived isotope proxies for catchment hydrological responsiveness and connectivity. We found correlations between the isotope proxies and catchment characteristics, suggesting that lakes and wetlands are intimately connected to rivers, and that permafrost increases the responsiveness of the catchment to rainfall and snowmelt events, reducing catchment mean transit times. Our work provides rare isotope-based field evidence that permafrost and lakes/wetlands influence hydrological pathways across a wide range of spatial scales (10–$10^5$ $km^2$) and permafrost coverage (0%–70%). This has important implications, because both permafrost extent and lake/wetland coverage are affected by permafrost thaw in the changing climate. Changes in these hydrological landscape controls are likely to alter carbon export and emission via inland waters, which may be of global significance.

## 1. Introduction

The Western Siberian Lowlands (WSL) store a substantial amount of the global terrestrial carbon pool in widespread organic peat soils (Sheng *et al* 2004, Smith *et al* 2012). Soil carbon is largely immobilized by permafrost, but rising global air temperature is causing substantial thawing of permafrost, which is expected to have implications for carbon mineralization and release from peatlands (Frey and Smith 2005, White *et al* 2007, Walvoord and Kurylyk 2016). Carbon release is of special importance as the greenhouse gases ($CO_2$ and $CH_4$) emitted from the soil carbon stocks will have a positive feedback effect, further enhancing global warming (Frey and McClelland 2009, Schaefer *et al* 2014, Grosse *et al* 2016). Such biogeochemical processes are intimately linked with cryogenic and hydrological processes, which need to be better understood to predict future environmental change in the WSL (Yi *et al* 2012, Shiklomanov *et al* 2013, Zakharova *et al* 2014). Although environmental change such as increasing river flows has been reported in Arctic rivers (Peterson *et al* 2002, White *et al* 2007, St Jacques and Sauchyn 2009), persistent trends in precipitation or discharge have not been found for the major watersheds of the WSL (Berezovskaya *et al* 2004, Karlsson *et al* 2012). However, in parts of the WSL, long-term trends in decreased streamflow variability (Karlsson *et al* 2012) and increased surface storage (Zakharova *et al* 2011) and mid-summer and winter flows (Yang *et al* 2004), all indicative of permafrost thaw, are reported. Thermokarst lake development is also associated with permafrost thaw, and Smith *et al* (2005) reported increasing lake numbers in the WSL for the continuous permafrost zone, and decreasing numbers for the discontinuous zone. However, work by Karlsson *et al* (2012, 2014) did not find any long-term trends in WSL lake abundance.

High-latitude hydrological research in North America has improved our understanding of the process by which permafrost influences runoff generation (Quinton and Marsh 1999, Carey and Quinton 2004, Walvoord *et al* 2012), how lakes and wetlands interact with the hydrological cycle (Gibson 2001, Hayashi *et al* 2004, Woo and Mielko 2007, Turner *et al* 2010) and how this is reflected in river water sources (Gibson *et al* 2016, Lachniet *et al* 2016). Although the same physical processes operate in the WSL, in the context of permafrost-influenced circumpolar areas it is unique in the spatial extent of its low-relief terrain (Kirpotin *et al* 2009, Zakharova *et al* 2009, Pokrovsky *et al* 2015, Manasypov *et al* 2015). Hydrological studies in the WSL have shown that these specific characteristics result in a large water storage capacity in both organic soils and open waterbodies (Smith *et al* 2012, Zakharova *et al* 2014, Ala-aho *et al* 2018), leading to higher than average evaporation loss (in a high-latitude context) (Serreze *et al* 2002, Slater *et al* 2007, Zakharova *et al* 2011). All these processes are likely to have important consequences for both water transit time and hydrological connectivity in the WSL, which, in turn, are likely of importance for biogeochemical processes Pokrovsky *et al* (2015). However, data availability and suitable methodologies for studying remote, largely ungauged areas is a major persistent challenge in WSL hydrology (Sivapalan *et al* 2003, Shiklomanov *et al* 2013).

Stable water isotopes can provide integral signals of the functioning of a hydrological catchment. They are proven tools for assessing and quantifying catchment functionality and responsiveness to rainfall and snowmelt events by estimating mean transit times and water travel time distributions (Soulsby and Tetzlaff 2008, Soulsby *et al* 2015, Benettin *et al* 2017), percentage of event water contributions (Rodhe 1981, Laudon *et al* 2007, Tetzlaff *et al* 2014) and, more recently, young water fractions (Jasechko *et al* 2016, Kirchner 2016). The average time water takes to travel through a catchment, commonly referred to as mean transit time (MTT), is an important metric that can also be used to understand biogeochemical processes, such as leaching or the kinetics of biogeochemical processes (McGuire and McDonnell 2006).

Previous work has shown that topography (McGuire *et al* 2005), soil type (Soulsby *et al* 2006, Laudon *et al* 2007, Hrachowitz *et al* 2009, Tetzlaff *et al* 2009a) and geology (Asano and Uchida 2012, Hale and McDonnell 2016) exert primary controls on catchment MTTs. So far, the majority of work on transit times has been done in temperate regions with no frozen soil influence, and in small catchments ($<100\,km^2$) with steep slopes, variable geological or pedological characteristics and few lakes and wetlands. However, the resulting insights transfer poorly to high latitudes, where cryogenic processes (snow, permafrost) have a strong influence on seasonality of the hydrological regime and the ability of the catchment to partition water to surface flow and deeper flow paths (Woo *et al* 2008, Tetzlaff *et al* 2015, Walvoord and Kurylyk 2016, Ala-aho *et al* 2017). Studies in the WSL, and other high-latitude regions, also show that the influence of lakes/bogs and permafrost can be important for runoff attenuation or enhancing near-surface flow, including the shallow subsurface (suprapermafrost) water (Cooper *et al* 1993, Hayashi *et al* 2004, Cooper *et al* 2008, Yi *et al* 2012, Connon *et al* 2015, Streletskiy *et al* 2015, Lachniet *et al* 2016).

A major challenge in using isotopes to estimate catchment water transmission metrics is that long and frequent temporal time series (e.g. fortnightly sampling over a year) are typically required to reliably parameterize the used distributions or models (McGuire and McDonnell 2006). However, previous work has shown that simpler methods relating tracer input (precipitation) and output (streamflow) variability can be used to derive proxies for catchment MTTs (Soulsby and Tetzlaff 2008, Tetzlaff *et al* 2009b, Buttle 2016). Such simple metrics are promising for remote high-latitude regions, where access and data collection are significant challenges and data sets are sparse.

The main contribution of this paper is the use of physical catchment characteristics to explain metrics of catchment hydrological functioning, derived from water isotopes, for the WSL.

Our specific objectives are to:

- assess the usefulness of river isotope monitoring, targeting primarily the main seasonal hydrological periods (winter baseflow, spring flood, summer low flows), to serve as a proxy for catchment responsiveness and hydrological connectivity in a large remote region;
- use these isotope data, collected from variable sized catchments (10–$10^5$ $km^2$) along a 1700 km transect from permafrost-free in the south to continuous permafrost in the north, to assess the influence of catchment characteristics on hydrological functioning of the WSL

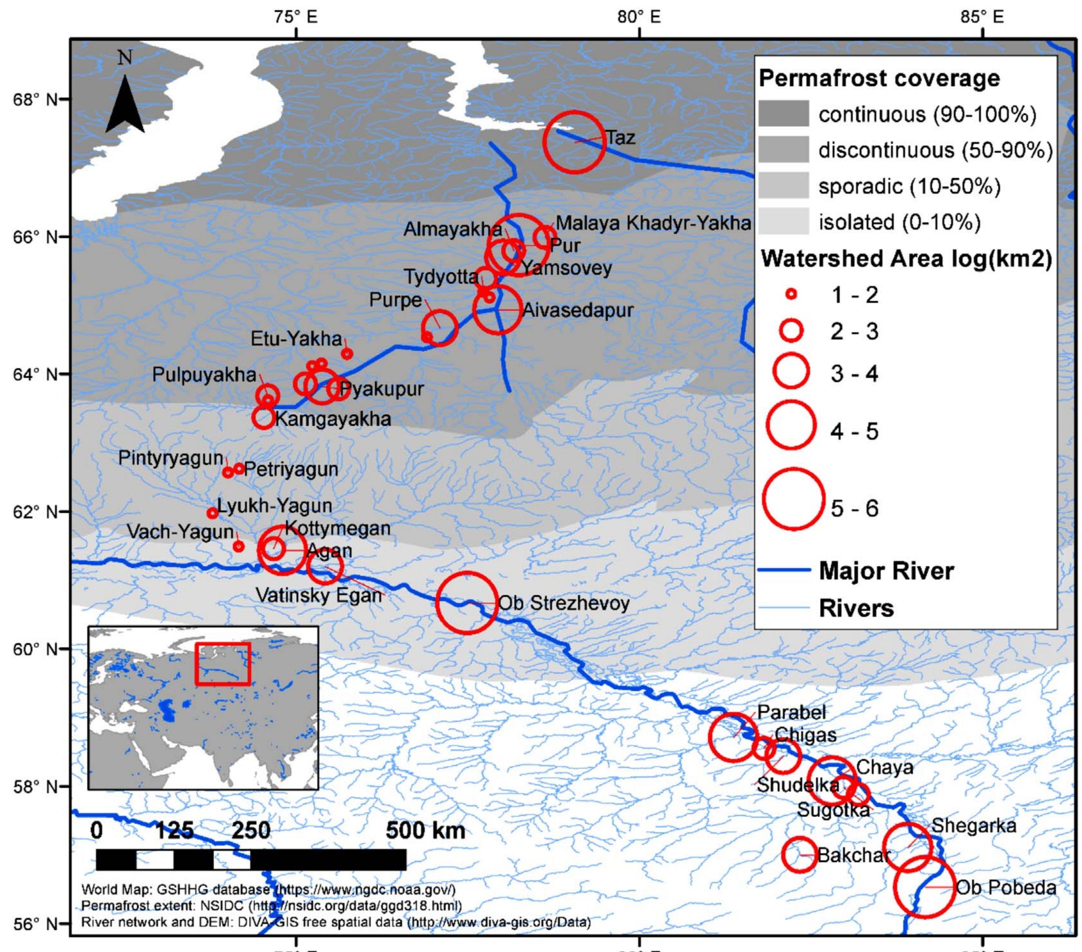


**Figure 1.** The sampled rivers along the study transect in the WSL. The centre of the circle indicates the sampling location, and the circle size shows the relative size of the catchment area (logarithm transformed). Different shades of grey describe the permafrost influence, according to Brown *et al* (2002).

With this approach, we are able to increase our understanding of the main hydrological controls in this large and hydrologically unique data-sparse region, whose hydrology has global-scale relevance in underpinning changes in water and carbon fluxes in a changing climate.

## 2. Materials and methods

### 2.1. Study area

The WSL (figure 1) is a vast low-lying area of over $3 \times 10^6$ km$^2$. Most forms part of the River Ob watershed, but it also encompasses the Nadym, Taz and Pur watersheds in the northern, permafrost-affected part. Geologically, the WSL comprises sedimentary rocks, mantled by Quaternary aeolian fluvial sand, silt and clay deposits from a few meters to 200–250 m in thickness (Ulmishek 2003, Pokrovsky *et al* 2015). The soils present as a thick peat layer (Dystric Hemic Histosols (Gelic)) overlying mineral deposits.

The area's hydrology is dominated by the cold climate and the extensive low-relief areas. As a result of low gradients and poor drainage, the landscape is characterized by waterlogged peatlands with a mosaic of lakes and wetlands (Frey *et al* 2007, Kirpotin *et al* 2009). The lake coverage is 5.7% on average but achieves 30% in some places (Polishchuk *et al* 2017). An abundance of open water bodies and shallow water tables makes summer evaporation a larger source of water loss in the WSL compared with other major circumpolar watersheds (Serreze *et al* 2002, Slater *et al* 2007). The area has modest differences in precipitation (550 mm in the south, 650–700 mm in the middle ranges, and 400–500 mm in the north). Mean annual air temperature decreases from south (−0.5 °C) to north (–9.5 °C). Cold mean annual air temperatures have a strong, but spatially variable, control on the soil thermal regime. The southern areas experience seasonally frozen soils during winter, but towards the north, permafrost, i.e. perennially frozen soil, becomes more common (figure 1). All of the above factors make the WSL a unique large-scale natural laboratory for the study of hydrological processes.

### 2.2. River isotope sampling

We collected samples for stable isotope analysis from tributaries of the Ob, Taz and Pur rivers, and the

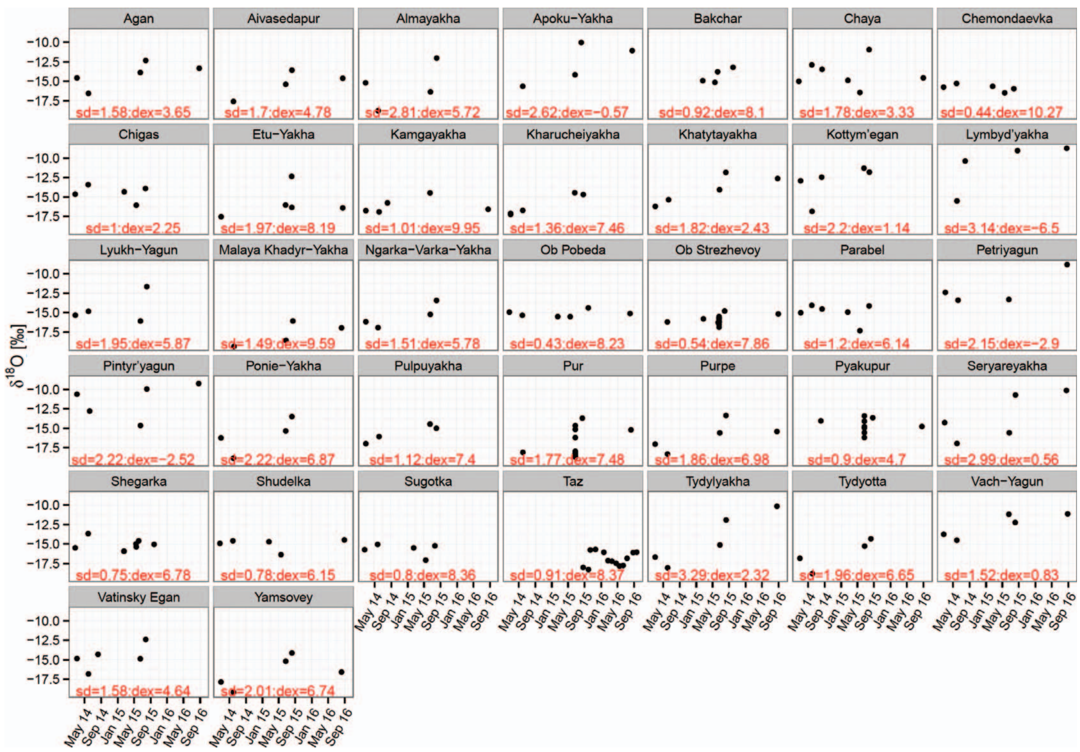


**Figure 2.** Time series of $\delta^{18}$O variability in individual catchments with four or more samples during the monitoring period. sd is the $\delta^{18}$O standard deviation, dex is the mean d-excess for each catchment.

main stems of the Ob and Taz rivers ($n = 37$) along a south–north transect in the WSL (figure 1). Catchment areas of the sampling locations varied between 12 km$^2$ and $7.7 \times 10^5$ km$^2$. Catchment characteristics (table 1) were determined by digitalizing available 1:1 000 000 and 1:500 000 maps of the permafrost distribution, physico-geographical parameters, geocryology, soil and lithology. Sampling was conducted between January 2014 and October 2016 (figure 2). The remote, inaccessible locations restricted the sampling to individual campaigns (2–3 weeks each), and due to resource restrictions we could not sample all rivers during each campaign. Therefore, the campaigns targeted the dominant periods of the hydrological year in circumpolar regions: winter baseflow when the area is frozen/snow covered, spring flood induced by snowmelt, and summer low flows. During the campaigns, more rivers were sampled (for the full dataset see Ala-aho *et al* 2018), but here we included only rivers with four or more samples, which we considered the minimum number of samples to produce a proxy for the isotopic variability in rivers. Sampling was done by grab samples collected from the middle of the river channel wherever possible, or from the river bank at 0.5 m depth from actively flowing water. Samples were collected in 3.5 ml glass vials and stored in the dark at 4 °C–6 °C. Analysis used a Los Gatos DLT-100 laser isotope analyzer (precision ±0.4‰ for $\delta^2$H; ±0.1‰ for $\delta^{18}$O) at the University of Aberdeen. Results use $\delta$-notation with respect to Vienna Standard Mean Ocean Water standards.

### 2.3. Isotope standard deviation and deuterium excess as proxies for catchment MTTs and water sources

We use the standard deviation (SD) of measured $\delta^{18}$O in stream water as an index of isotope damping to provide insight into catchment hydrological responsiveness. Although $\delta^2$H was also measured, in our variability analysis we used $\delta^{18}$O due its lower sensitivity to evaporation fractionation. The ratio between precipitation SD and stream SD has been shown to correlate with catchment MTT (McGuire *et al* 2005, Soulsby and Tetzlaff 2008), and therefore it can provide a simple, easily determined proxy for catchment MTT. This is because greater isotopic damping in stream water is indicative of greater mixing and longer MTTs (Hrachowitz *et al* 2009).

Due to insurmountable logistical difficulties we could not collect representative precipitation samples across the region, and therefore could not directly assess the precipitation variability (SD) over the study transect. Continentality and temperature effects may introduce systematic changes in the precipitation isotope composition (Gat 1996), and thereby affect the isotope variability of streamflow. We explored the likely isotope variability in precipitation in the WSL using historical data (collected between 1973 and 1990) in the IAEA GNIP database (IAEA/WMO 2017) from seven observation stations (figure S1 available at stacks.iop.org/ERL/13/034028/mmedia) enveloping the study transect. Based on the Bartlett test on homogeneity of variances there were no statistically

**Table 1.** Catchment characteristics.

| Catchment name | Lat. | Long. | Area (km$^2$) | Mean annual temp (°C) | Annual precip. (mm) | Average slope [‰] | Mean elevation (m) | Annual discharge (m$^3$ s$^{-1}$) | Bog (%) | Forest (%) | Lake (%) | Soil, sand (%) | Soil loam (%) | Perma-frost (%) |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Agan | 61.43 | 74.80 | $2.9 \times 10^4$ | −3.6 | 670 | 0.07 | 95 | 288 | 46.9 | 42.5 | 10.6 | 35 | 10 | 5 |
| Aivasedapur | 64.93 | 77.94 | $2.6 \times 10^4$ | −5.7 | 670 | 0.13 | 88 | 256 | 40.1 | 45.5 | 14.4 | 50 | 4 | 20 |
| Almayakha | 65.79 | 78.17 | $2.5 \times 10^2$ | −7.7 | 620 | 0.27 | 26 | 2.4 | 76.3 | 4.2 | 19.5 | 24 | 0 | 76 |
| Apoku-Yakha | 64.15 | 75.37 | $2.8 \times 10^1$ | −5.7 | 670 | 0.67 | 83 | 0.2 | 75.5 | 12.8 | 11.7 | 24 | 0 | 38 |
| Bakchar | 57.00 | 82.34 | $2.7 \times 10^3$ | −0.7 | 570 | 0.23 | 122 | 11.8 | 39.3 | 60.7 | 0.2 | 0 | 61 | 0 |
| Chaya | 58.08 | 82.81 | $2.7 \times 10^4$ | −0.9 | 575 | 0.06 | 115 | 75.6 | 59.3 | 39.5 | 1.2 | 0 | 41 | 0 |
| Chemondaevka | 57.87 | 83.19 | $2.1 \times 10^2$ | −1.1 | 580 | 0.67 | 101 | 0.5 | 10.4 | 49.6 | 0.04 | 0 | 90 | 0 |
| Chigas | 58.55 | 81.81 | $6.3 \times 10^2$ | −1.6 | 580 | 0.76 | 91 | 3.2 | 39.4 | 46.2 | 1.58 | 0 | 61 | 0 |
| Etu-Yakha | 64.29 | 75.74 | $8.3 \times 10^1$ | −5.8 | 670 | 0.48 | 76 | 0.6 | 23.4 | 71.5 | 1.96 | 77 | 0 | 23 |
| Kamgayakha | 63.37 | 74.53 | $2.0 \times 10^2$ | −5.2 | 670 | 0.56 | 121 | 1.3 | 23.7 | 76.2 | 0.1 | 76 | 0 | 12 |
| Kharucheiyakha | 63.86 | 75.14 | $7.8 \times 10^2$ | −5.6 | 670 | 0.73 | 111 | 7.3 | 44.6 | 54 | 1.4 | 55 | 0 | 44 |
| Khatytayakha | 63.61 | 74.59 | $2.8 \times 10^1$ | −5.4 | 670 | 0.67 | 89 | 0.1 | 75.3 | 13.2 | 10.8 | 25 | 0 | 38 |
| Kottymegan | 61.45 | 74.67 | $2.4 \times 10^2$ | −2.7 | 670 | 0.19 | 43 | 1.4 | 77.6 | 12 | 10.4 | 22 | 0 | 0 |
| Lymbydyakha | 63.78 | 75.62 | $1.0 \times 10^2$ | −5.6 | 670 | 0.26 | 68 | 0.5 | 59.3 | 6.1 | 34.6 | 41 | 0 | 30 |
| Lyukh-Yagun | 61.97 | 73.78 | $9.2 \times 10^1$ | −3.5 | 670 | 0.58 | 68 | 1 | 62.2 | 17.9 | 19.5 | 38 | 0 | 0 |
| Malaya Khadyr-Yakha | 65.99 | 78.62 | $6.5 \times 10^2$ | −7.9 | 580 | 0.44 | 55 | 6.9 | 14.8 | 84.9 | 0.3 | 31 | 53 | 15 |
| Ngarka-Varka-Yakha | 64.11 | 75.24 | $4.8 \times 10^1$ | −5.6 | 670 | 0.5 | 86 | 0.2 | 52.1 | 32.6 | 15.3 | 48 | 0 | 26 |
| Ob Pobeda | 56.53 | 84.16 | $2.64 \times 10^5$ | −0.2 | 560 | | | 4030 | 0.7 | 53.4 | 0.23 | | | 0 |
| Ob Strezhevoy | 60.67 | 77.49 | $7.7 \times 10^5$ | −2.5 | 510 | 0.04 | 200 | 5080 | 10 | 71.4 | 0.66 | | | 0 |
| Parabel | 58.71 | 81.37 | $2.4 \times 10^4$ | −1.4 | 502 | | 110 | 151 | 28.8 | 69.4 | 0.8 | 0 | 31 | 0 |
| Petriyagun | 62.62 | 74.17 | $4.1 \times 10^1$ | −4 | 670 | 0.33 | 92 | 0.4 | 57.2 | 6.7 | 36.1 | 43 | 0 | 5 |
| Pintyryagun | 62.56 | 74.01 | $1.7 \times 10^1$ | −4 | 670 | 0 | 82 | | 61 | 0 | 39 | 39 | 0 | 8 |
| Ponie-Yakha | 65.39 | 77.76 | $1.5 \times 10^2$ | −6.9 | 640 | 0.2 | 30 | 1.3 | 66 | 17.7 | 16.3 | 34 | 0 | 70 |
| Pulpuyakha | 63.68 | 74.59 | $1.8 \times 10^2$ | −5.4 | 670 | 0.57 | 111 | 0.5 | 27.8 | 61.8 | 9.28 | 72 | 0 | 15 |
| Pur | 65.88 | 78.25 | $1.1 \times 10^5$ | −5.8 | 650 | 0.11 | 81 | 761 | 56.9 | 34.4 | 8.7 | 43 | 0 | 34 |
| Purpe | 64.67 | 77.09 | $5.0 \times 10^3$ | −5.8 | 670 | 0.2 | 80 | 50.2 | 48 | 34 | 15 | 52 | 0 | 48 |
| Pyakupur | 63.82 | 75.38 | $9.9 \times 10^3$ | −5.2 | 670 | 0.15 | 94 | 102 | 45 | 40 | 12 | 55 | 0 | 34 |
| Seryareyakha | 64.54 | 76.91 | $2.9 \times 10^1$ | −5.9 | 670 | 0.31 | 49 | 0.3 | 61.2 | 19.4 | 19.4 | 39 | 0 | 60 |
| Shegarka | 57.11 | 83.91 | $1.2 \times 10^4$ | −0.6 | 575 | 0.17 | 126 | 22.3 | 19.7 | 41.4 | 1.1 | 0 | 80 | 0 |
| Shudelka | 58.43 | 82.10 | $3.6 \times 10^3$ | | | | | | 68.2 | 31.8 | 0 | 0 | 32 | 0 |
| Sugotka | 57.98 | 82.98 | $2.8 \times 10^2$ | −1.1 | 580 | 0.93 | 103 | 0.6 | 6.99 | 62.6 | 0 | 0 | 93 | 0 |
| Taz | 67.37 | 79.06 | $1.4 \times 10^5$ | −6.5 | 600 | 0.06 | 114 | 1569 | 38 | 59 | 3 | 62 | 0 | 40 |
| Tydylyakha | 65.11 | 77.82 | $1.7 \times 10^1$ | −6.5 | 650 | 0.63 | 31 | 0.2 | 49.4 | 37.4 | 12.7 | 51 | 0 | 49 |
| Tydyotta | 65.20 | 77.73 | $1.2 \times 10^1$ | −6.5 | 650 | 0.16 | 60 | 40.8 | 53 | 43 | 4 | 47 | 0 | 25 |
| Vach-Yagun | 61.49 | 74.16 | $9.9 \times 10^1$ | −2.8 | 670 | 0.33 | 47 | 0.6 | 77.9 | 17.2 | 1.7 | 22 | 0 | 0 |
| Vatinsky Egan | 61.20 | 75.42 | $3.1 \times 10^3$ | −2.6 | 670 | 0.11 | 80 | 27.9 | 67.3 | 27.5 | 5.2 | 31 | 18 | 0 |
| Yamsovey | 65.70 | 78.02 | $4.1 \times 10^3$ | −7.5 | 600 | 0.26 | 67 | 40.5 | 53.7 | 38.7 | 7.6 | 46 | 0 | 54 |
| **Average** | **62.40** | **77.75** | $\mathbf{3.8 \times 10^4}$ | **−4.32** | **632** | **0.35** | **85.6** | **358.2** | **46.4** | **38.8** | **9.4** | **33.8** | **16.4** | **20.8** |

significant differences (Bartlett's $K^2 = 2.26$, $p$-value $= 0.89$) in the variability in $\delta^{18}$O between stations. This supports the use of only the SD in stream samples as a proxy for MTT, implicitly assuming that the range of $\delta^{18}$O variation in precipitation is constant across the area and does not introduce systematic differences in stream $\delta^{18}$O variability.

Deuterium excess (d-excess) was calculated for samples according to Dansgaard (1964) with d-excess $= \delta^2\mathrm{H} - 8 \times \delta^{18}\mathrm{O}$. Values lower than 10‰ are associated with kinetic fractionation, which is indicative of a sample being subjected to evaporation. In addition to inland evaporation processes, d-excess can also reflect environmental conditions at the moisture sources of precipitation (Gat 1996). We considered the d-excess signal as a proxy for the river water being sourced by water bodies that have undergone evaporation, i.e. water from lakes, wetlands and surficial soils (suprapermafrost flow or shallow subsurface waters).

It should be noted that isotopic enrichment occurring in lakes and bogs can bias the isotope SD in streamflow if it is to be used as a proxy for catchment responsivity to rainfall/snowmelt events. In other words, higher stream SD is not only a result of a more rapid catchment response but is also a product of evaporative enrichment. In order to isolate the variability in the river isotope ratios reflecting only the catchment responsiveness and not evaporation effects, we constructed a linear model to explain the isotope variability in SD caused specifically by lake and bog coverage:

$$\mathrm{SD}_E = a + b \cdot L_{\%} + c \cdot B_{\%} \quad (1)$$

where $\mathrm{SD}_E$ is the catchment isotope variability (response variable), $L_{\%}$ and $B_{\%}$ are the lake and bog percentages, respectively, and $a$, $b$ and $c$ are constants of the linear model obtained through model fitting. To exclude the effect on lake and bog evaporated water, the river SD in each catchment was 'normalized' by:

$$\mathrm{SD}_{\mathrm{norm}} - \mathrm{SD}_E \quad (2)$$

where $\mathrm{SD}_{\mathrm{norm}}$ is effectively the residual term of the linear model $\mathrm{SD}_E$. From a physical perspective we considered $\mathrm{SD}_{\mathrm{norm}}$ to be a similar proxy for isotope variability as the original SD. High (positive) values of $\mathrm{SD}_{\mathrm{norm}}$ correspond to catchments with high isotope variability (SD) after removing the systematic variance caused by the influence of lakes and bogs with equation (2). In catchments where the SD is low, $\mathrm{SD}_{\mathrm{norm}}$ may become negative, indicating low isotope variability also after normalization.

## 3. Results

### 3.1. Measured isotopes and their temporal variability

Figure 2 presents the $\delta^{18}$O time series for each catchment showing the data from which the hydrological responsivity and connectivity proxies (SD and d-excess, respectively) are derived. Typical (average) isotope values vary between catchments, with consistently enriched values in some (e.g. Lymbydyakha, Vach-Yagun) and greater depletion in others (e.g. Malaya Khadyr-Yakha, Taz). The average $\delta^{18}$O, however, has no bearing on the analysis. The variability in river isotopes, expressed as the SD, forms the basis of analysis, with a low SD value being interpreted as a subdued catchment response indicative of a relatively long MTT, and a high SD value as a more responsive catchment with a shorter MTT. The SD values vary between 0.43‰ in Pobeda, Ob and 3.29‰ in Tydylyakha.

Catchment average d-excess (dex) is also given for each site in figure 2, but is not intuitively shown in the figure because only $\delta^{18}$O is plotted as the time series. The average d-excess values vary from most negative (−6.5‰) in Lymbydyakha indicating a strong evaporation signal, to values near average global precipitation (10.3‰) in Chemondaevka, suggesting negligible influence of evaporation.

The dual-isotope plot shows seasonal differences in river isotope composition between spring, summer and winter (figure S2) Importantly, all seasons have some overlap—with some enriched samples in spring and depleted in the summer—suggesting a complex and prevalent hydrological connectivity in the region.

### 3.2. Correlation between isotope proxies and catchment characteristics

The variability in stream isotopes, used as a proxy for catchment responsivity and expressed as SD, is positively correlated with latitude, and permafrost, lake and bog coverage (figure 3, top row). The positive correlations imply that these landscape characteristics decrease the catchment MTTs. Negative correlations with SD are found for forest coverage and logarithm of watershed area, inferring an increasing effect on catchment MTT. All correlations are statistically significant at the 0.05 confidence level.

Catchment average d-excess, here used as a proxy for river connectivity with evaporated water sources residing in the landscape, is positively correlated with forest coverage and watershed area, and negatively with lake and bog percentage. No correlation is evident for latitude and permafrost coverage (figure 3, middle row). In the context of d-excess, for which more negative values imply a stronger evaporation signal, the results indicate that the streams with a high lake and bog percentage in the watershed carry more evaporated water.

The significant negative correlation with lake and bog percentage and d-excess implies that the observed isotope variability (expressed as SD) is partially caused by evaporation enrichment taking place in lakes and bogs. We found a linear model $\mathrm{SD}_E = 0.63 + 0.033L_{\%} + 0.015B_{\%}$ (equation 1) to explain 0.52% of the variability in the river SD.

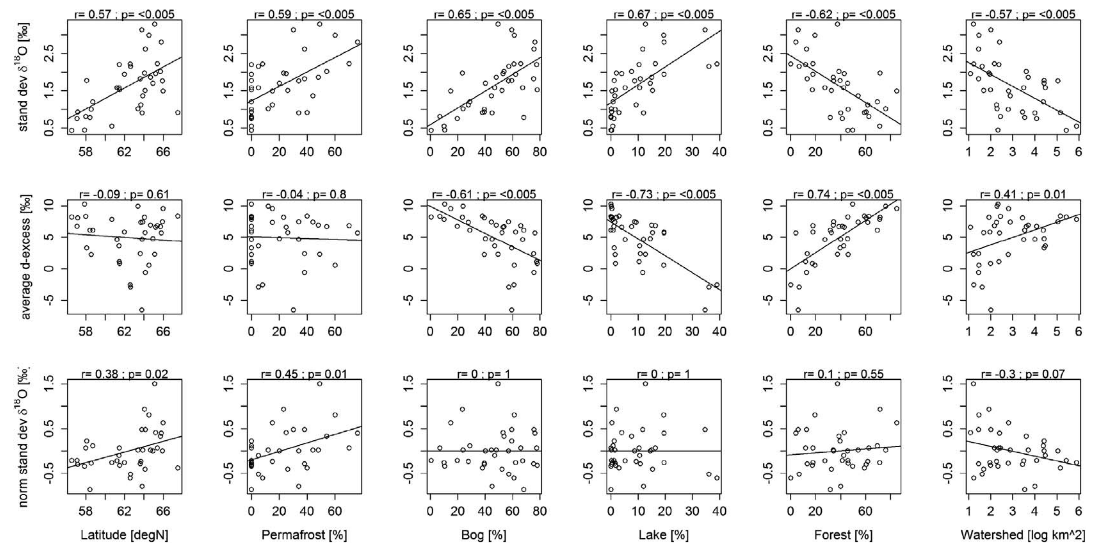


**Figure 3.** Scatterplots for landscape characteristics explaining isotope standard deviation for $\delta^{18}$O (top row) average d-excess (middle row) and standard deviation for $\delta^{18}$O after normalizing to lake and bog percentage (bottom row). The plot title gives the Pearson correlation coefficient, and the *p*-value indicating its statistical significance.

After normalizing the SD for the influence of lake and bog evaporation using equation (2), i.e. analyzing the residual term of $SD_E$, the correlation between lakes, bog and forest coverage disappears, as expected (figure 3, bottom row). The only statistically significant explanatory variables remaining to explain the isotope variability are latitude and permafrost coverage. The negative correlation between isotope variability and watershed area persists, but it is no longer statistically significant at the 0.05 confidence level.

## 4. Discussion

### 4.1. Utility and limitations of isotope MTT proxies from sparse datasets in remote high-latitude regions

We could relate the isotope variability in rivers draining the WSL across a 1700 km transect to hydrologically relevant catchment characteristics using our unique large-scale dataset. It should be stressed, however, that the relatively small (and variable) number of samples per catchment, from which variability (SD) is approximated (figure 2), causes uncertainty in the results. The relatively few samples per river were a direct consequence of the vast areal extent and challenging accessibility in our study transect. As a result, not all rivers could be sampled in all of the campaigns, which brings an additional constraint on the comparability between catchments. The shortcomings due to sparse data were mitigated by the sampling design, where specific flow conditions are targeted to capture the major seasonal isotope variability. Isotope signals were expected to be most depleted during snowmelt-induced spring floods, and most enriched during mid-summer due to evaporation and enriched summer precipitation, which is confirmed by figure S2. Still, it is unlikely that the absolute minimum and maximum isotope ratios were captured in all rivers. Finally, systematic variation in the precipitation isotope composition across the study region can potentially introduce a bias in the analysis. We did not find systematic changes in the precipitation isotope variability in the historical GNIP monitoring data (figure S1), but the data predated the study period, leaving uncertainty in the contemporary precipitation signal. However, despite these limitations, the result that we see, i.e. strong covariance with SD and average d-excess isotope proxies and landscape characteristics, indicates that the data capture some important interactions controlling runoff generation in this vast, sparsely monitored lowland area.

Previous work has successfully related catchment characteristics to similar MTT proxies, such as the SD ratio between stream and precipitation isotopes, albeit determined from a higher number of samples. However, their analysis has been complicated by either a smaller number (<10) of catchments (Rodgers *et al* 2005, Buttle 2016) or catchments from very different geomorphic provinces (Tetzlaff *et al* 2009b).

A key finding in prior isotope studies has been the strong control of topography and subsurface characteristics (soil type, geology) on catchments MTTs (McGuire *et al* 2005, Soulsby and Tetzlaff 2008, Hrachowitz *et al* 2010, Hale and McDonnell 2016). However, the WSL as an area is extremely flat, and given the size, is relative homogeneous in its geology and relief (Pokrovsky *et al* 2015), implying that the typical controls on MTT may not be applicable (Devito *et al* 2005). In fact, we did not explore the covariance of topographically derived indices and isotope proxies in the WSL, which is typically done in similar studies (Laudon *et al* 2007, Tetzlaff *et al* 2009b). In contrast, the unique geomorphic province of WSL allowed us to use the isotope data to focus on assessing the influence of surface water (lakes and bogs) and permafrost presence, which previous work has found to be a major hydrological control in high-latitude environments (McNamara *et al* 1997, Hayashi *et al* 2004, Woo *et al* 2008, Lyon *et al* 2010, Tetzlaff *et al* 2015, Walvoord and Kurylyk 2016).

### 4.2. Influence of catchment characteristics on the hydrological functioning of the WSL

The d-excess signal in our isotope data show that rivers in the WSL carry water that has been subjected to evaporation, and that the d-excess signal is negatively correlated with lake and bog percentage (figure 3). It is known that streams connected to lakes tend to have a more enriched isotopic composition due to evaporation (Gibson *et al* 2005, Laudon *et al* 2007), and isotope studies have shown that water can be displaced from peatlands adjacent to stream networks (Carey and Quinton 2004, Rodgers *et al* 2005, Sprenger *et al* 2016). Gibson *et al* (2015) found bog cover and permafrost to be the dominant hydrological controls in northeast Alberta, Canada, where they calculated water yields and runoff ratios using nine years of isotope data collected from 50 lakes. We also observed positive correlation with forest cover and d-excess and negative correlation between forest cover and SD. However, forest percentage is negatively correlated with lake and bog percentage (Pearson's $r = 0.73$ and 0.81, respectively). Based on the d-excess signal in catchments with lake and bog influence, we hypothesize, from a physical perspective, that causality between lakes and bogs, not forests, is the reason for the response in both SD and d-excess.

Latitude and permafrost coverage do not appear to co-vary with d-excess. Some work suggests that permafrost reduces hydrological connectivity in the landscape (Wright *et al* 2009, Zakharova *et al* 2009, Connon *et al* 2015, Manasypov *et al* 2015), and that runoff in complex permafrost terrain is generated through 'fill and spill' where lakes/bogs connect to streams mainly during snowmelt for a very short period (Quinton and Roulet 1998, Woo and Mielko 2007, Spence 2010). Following this logic, a stronger permafrost influence would lead to shorter time windows when lakes and streams connect in the spring. During this period, river water would be dominated by snowmelt with typically high d-excess values—leading presumably to a positive correlation between d-excess and permafrost cover. The lack of such a correlation in our data suggests that lakes and bogs in the WSL experience connectivity with rivers, and old, evaporated, water residing in the landscape is displaced to rivers after mixing during rainfall and snowmelt events even in permafrost environments. We hypothesize that this connectivity is achieved via water movement along the permafrost table in the thawed active layer, in the form of so-called suprapermafrost flow between peat bogs and the lakes, and further to the rivers.

Latitude, and the associated increase in permafrost coverage, was positively correlated with the isotope variability (SD) in rivers (figure 3). According to previous results (e.g. Soulsby and Tetzlaff 2008, Tetzlaff *et al* 2009b, Buttle 2016), isotope variability is directly related to catchment MTT. With this reasoning, our data suggest that permafrost decreases the catchment-scale water MTT. However, as our d-excess analysis shows, isotope fractionation from bogs and lakes probably causes variability in the stream isotope signal, which is not necessarily related to the MTT in our catchments. Therefore using the sole SD as a proxy for MTT would be biased in the WSL, which exhibits a strong influence of lakes and bogs on the river isotope composition. Another process that may distort the observed isotope variability, in addition to lake/wetland evaporation, is water released from thawing permafrost (Walvoord and Striegl 2007, Gibson *et al* 2015), but it is difficult to disentangle because of the mixed and variable isotope composition of the frozen soil water (Streletskiy *et al* 2015, Throckmorton *et al* 2016).

After eliminating the influence of lakes and bogs on the isotope variability, a statistically significant correlation between isotope variability and latitude and permafrost coverage prevails (figure 3, bottom row). This suggests that partial disappearance of permafrost in the north can increase water travel times in affected catchments. This is in general agreement with the accumulated understanding of permafrost hydrology, suggesting that permafrost leads to more responsive watersheds due to limited subsurface storage and rapid water transmission on frozen ground (Woo *et al* 2008, Walvoord and Kurylyk 2016). In the presence of widely documented recent permafrost thaw, which is expected to accelerate because of climate change, changes in high-latitude hydrological regimes are likely to result in increased riverine emission of $CO_2$ (Serikova *et al* in review). Longer MTTs due to thawing permafrost may either increase the riverine export of dissolved organic carbon (Frey and Smith 2005, Prokushkin *et al* 2011) or decrease it because of a lower degree of photo- and bio-degradation in inland waters (Mann *et al* 2014, Spencer *et al* 2015).

It should be noted that normalizing the SD for the influence of lakes and bogs is not conceptually straightforward despite the obvious evaporative influence. Wetlands in boreal regions have been shown to transmit water more quickly with respect to the rest of the catchment (Hayashi *et al* 2004, Connon *et al* 2015), especially when the wetland surface is frozen during early snowmelt (Roulet and Woo 1986, Laudon *et al* 2007). Therefore, perhaps some of the isotope variability (SD) caused by rapid response over the lakes and wetlands is removed when the SD is normalized for evaporation. The true relationship between permafrost and isotope variability would therefore lie somewhere between the original and normalized data (top and bottom rows in figure 3) and is clearly an issue that requires further research.

Finally, there appears to be a negative correlation between catchment area and isotope variability (figure 3). Previous work has found no evidence for a relationship between isotopically derived MTTs (or their SD proxies) and catchment area (McGuire *et al* 2005, Laudon *et al* 2007, Tetzlaff *et al* 2009b, Hrachowitz *et al* 2010, Jasechko *et al* 2017), suggesting that topography and/or soil characteristics control catchment-scale solute transport. However, very few studies have looked at scales beyond mesoscale catchments (i.e. >100 km$^2$) (though see Tetzlaff *et al* 2011). After normalizing for the influence of lakes and bogs, the relationship between watershed area and isotope variability is not statistically significant by a small margin, although it cannot be discounted. In the absence of strong controls for topography or soil type it may be that the larger mixing volumes associated with bigger watersheds may have a damping influence on solute transport.

## 5. Conclusions

Our work shows the utility of stable water isotopes sampled with relatively sparse, but targeted, temporal resolution, and extensive spatial coverage, in helping to identify the dominant landscape controls on hydrology of the WSL. We were able to show how lake/bog percentage and permafrost are linked to hydrological connectivity and catchment responsiveness, respectively, in the WSL. Thereby, we provide large-scale isotope-based evidence for the relative influence of permafrost and lake/bog coverage influencing hydrological processes in high-latitude catchments with variable size (10–10$^5$ km$^2$). Although tentative, the established new relationships with landscape characteristics and isotope proxies show promise for relatively cost-effective monitoring of the WSL—a large, remote and data-poor region with difficult access. The fact that our results highlight lake, wetland and permafrost coverage as first-order controls on the region's hydrology has important implications. Climate change, and the resulting permafrost thaw, is expected to alter not only the permafrost regime but also the abundance of lakes and wetlands in the region. Our work further supports the hypothesis that high-latitude regions, and the WSL in particular, are likely to experience hydrological changes because of thawing permafrost. Our work provides large-scale evidence that permafrost leads to more responsive watersheds, thus agreeing with previous findings that thawing of permafrost is likely to result in altered flow path dynamics and increased travel times.

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