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Soil frost effects on streamflow recessions in a subarctic catchment

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Abstract

The Arctic is warming rapidly. Changing seasonal freezing and thawing cycles of the soil are expected to affect river run-off substantially, but how soil frost influences river run-off at catchment scales is still largely unknown. We hypothesize that soil frost alters flow paths and therefore affects storage–discharge relations in subarctic catchments. To test this hypothesis, we used an approach that combines meteorological records and recession analysis. We studied streamflow data (1986–2015) of Abiskojokka, a river that drains a mountainous catchment (560 km^2) in the north of Sweden (68° latitude). Recessions were separated into frost periods (spring) and no-frost periods (summer) and then compared. We observed a significant difference between recessions of the two periods: During spring, discharge was linearly related to storage, whereas storage–discharge relationships in summer were less linear. An analysis of explanatory factors showed that after winters with cold soil temperatures and low snowpack, storage–discharge relations approached linearity. On the other hand, relatively warm winter soil conditions resulted in storage–discharge relationships that were less linear. Even in summer, relatively cold antecedent winter soils and low snowpack levels had a propagating effect on streamflow. This could be an indication that soil frost controls recharge of deep groundwater flow paths, which affects storage–discharge relationships in summer. We interpret these findings as evidence for soil frost to have an important control over river run-off dynamics. To our knowledge, this is the first study showing significant catchment-integrated effects of soil frost on this spatiotemporal scale.

KEYWORDS

Arctic, hydrology, permafrost, recession analysis, snowmelt, soil frost, thawing, warming

1 | INTRODUCTION

Arctic regions have experienced stronger warming than lower latitudes during the last decades (Bekryaev, Polyakov, & Alexeev, 2010; Francis, Vavrus, & Cohen, 2017). This warming of large amounts of subsurface carbon storage may cause the Arctic to shift from sink to source in the global carbon cycle (Cole et al., 2007; Zimov, Schuur, & Chapin III, 2006). Groundwater flow in Arctic ecosystems fulfils an important, climate-sensitive, ecosystem function: It conveys thawed water and solutes from the defrosting subsurface to streams and rivers (Vonk et al., 2015). This directly affects stream processes and carbon export from terrestrial systems. The flow paths that groundwater follows along its way to a stream largely determine the variation in stream biogeochemistry (Seibert et al., 2009). Frozen ground can affect these flow paths. Permafrost thawing due to climate warming changes flow paths (and subsequent stream chemistry) on long timescales (Frey & McClelland, 2009; S. Lyon et al., 2009).

On shorter timescales, seasonal soil frost temporarily affects flow paths (Bayard, Stähli, le Parriaux, & Flühler, 2005; Cherkauer, Bowling, & Lettenmaier, 2003; Nyberg, Stähli, Mellander, & Bishop, 2001), but long-term trends and future projections of soil frost vary among different geolocations. In Germany, mean soil temperatures are expected to increase, and freeze thaw cycles decrease with warmer climate conditions (Kreyling & Henry, 2011). On the other hand in New Hampshire, USA, no clear relation between warmer climate and decreasing soil frost depth has been reported, and simulations show that little change in frost depth can be expected in the coming decades (Campbell et al., 2010). However, earlier snow manipulations in the same study area show potential increase of soil freezing when snow packs decrease (Hardy et al., 2001). In Japan, the insulating effect of increasingly thicker snow packs have been correlated with decreasing soil frost depths (Hirota et al., 2006). Scandinavian studies predict increasing spatial variability in snow and soil frost distribution (Mellander, Löfvenius, & Laudon, 2007; Venäläinen et al., 2001).

1.1 | Conceptualizing soil frost effects on the hydrology of (sub) Arctic catchments

Especially in the spring, during snowmelt and rain on snow events, frozen soils can be an important control on subsurface water flow (Henry, 2008; Laudon, Seibert, Köhler, & Bishop, 2004). However, directly measuring soil frost processes on river basin scales is unfeasible, and therefore, the effects of soil frost on river flows remain poorly understood (Carey & Woo, 2001). Soil frost is a complex phenomenon that is difficult to monitor due to its variation in spatial extent and temporal character (Cherkauer et al., 2003). The presence of soil frost depends on thermal, hydraulic, and mechanical processes, which again depend on local soil properties, water availability, and temperature (Liu, Sun, & Yu, 2012). Hillslope studies indicate that soil frost, on some occasions, affects run-off generation and that winter conditions and local soil properties control the extent of this influence (Nyberg et al., 2001; Stähli, Nyberg, Mellander, Jansson, & Bishop, 2001). Surface run-off and blockage of infiltration are the most pronounced effects. A confined flow between a solid frozen soil and a partially thawed snowpack drastically alters the flow paths and travel times (Edwards, Creasey, & Cresser, 1986). These studies all found local effects of soil frost on water flow paths. However, the integrated catchment scale effects of soil frost for larger river basins have not yet been quantified.

Soil frost complicates hydrological systems, because it is a phenomenon affected by external drivers (soil moisture, snow, and temperature), which temporarily changes catchment characteristics that are often thought to be relatively constant (hydrological conductivity, storativity, and water retention; Seibert & Meerveld, 2016). We conceptualize the effect of soil frost on the typically shallow, vertical soil profile as (partial) deactivation of flow paths (Figure 1, top left panel). For this concept, we assume a profile (Figure 1, conductivity columns) of high hydraulic conductivity in the top soils, which decreases with depth due to the presence of bedrock and/or loam deposits within 2 m below the ground surface (Bishop, Seibert, Köhler, & Laudon, 2004; Rupp & Selker, 2005, 2006). Frozen soil conditions promote flow over the surface and through highly conductive (organic) top soils (Figure 1, top right panel). When there is no soil frost, all flow paths over the conductivity profile are active (Figure 1, lower panels). With river discharge being an integrator of catchment processes, we expect that streamflow records contain information of the temporary catchment scale effect of soil frost on water flow paths contributing to the stream. In this paper, we analyse discharge of a subarctic catchment ($566\ km^2$) with seasonally frozen soils. Our aim is to present an approach that allows to identify the catchment scale effects of seasonal soil frost on stream flows.

1.2 | Recession analysis

The use of discharge-based methods, such as recession analysis, is particularly valuable for subarctic and Arctic hydrological systems, because river discharge is one of the few observation types with reasonable spatial and temporal coverage in this harsh environment. Streamflow recession analysis aims to link catchment subsurface and surface characteristics to observed streamflow, relatively independent of the external drivers (Troch et al., 2013). Under receding streamflow conditions (e.g., after peak flows due to snowmelt or rainfall events), discharge can be conceptualized by draining of subsurface storages (Brutsaert & Nieber, 1977). When assumed that there are no significant inputs, such as precipitation, and other outputs than the discharge itself, such as evapotranspiration (ET), the water balance of a catchment with only storage and discharge can be rewritten to yield:

$$-\frac{\partial Q}{\partial t} = Q\frac{dQ}{dS} \approx \alpha Q^{\beta}, \qquad (1)$$

where Q is discharge, usually expressed in mm/day, coefficient α [$day^{\beta-2}/mm^{\beta-1}$] relates to aquifer properties and coefficient β[−] to the (non-)linearity of reservoir drainage (Troch et al., 2013). $\frac{dQ}{dS}$ is the sensitivity of discharge, Q, to a change in storage S, conveniently parameterized by α and β. We hypothesize that soil frost affects this sensitivity (i.e., the relation between storage and streamflow) and therefore that the effects of soil frost can be analysed through the values of parameters α and β. Because both the value and unit of α depends strongly on the value of β, α cannot be evaluated independently of β. Therefore, in this study, we focus our analysis on the effects of soil frost on storage–discharge linearity β, meaning that β approaching 1 represents linear decrease in discharge with decreasing storage. In this study, we

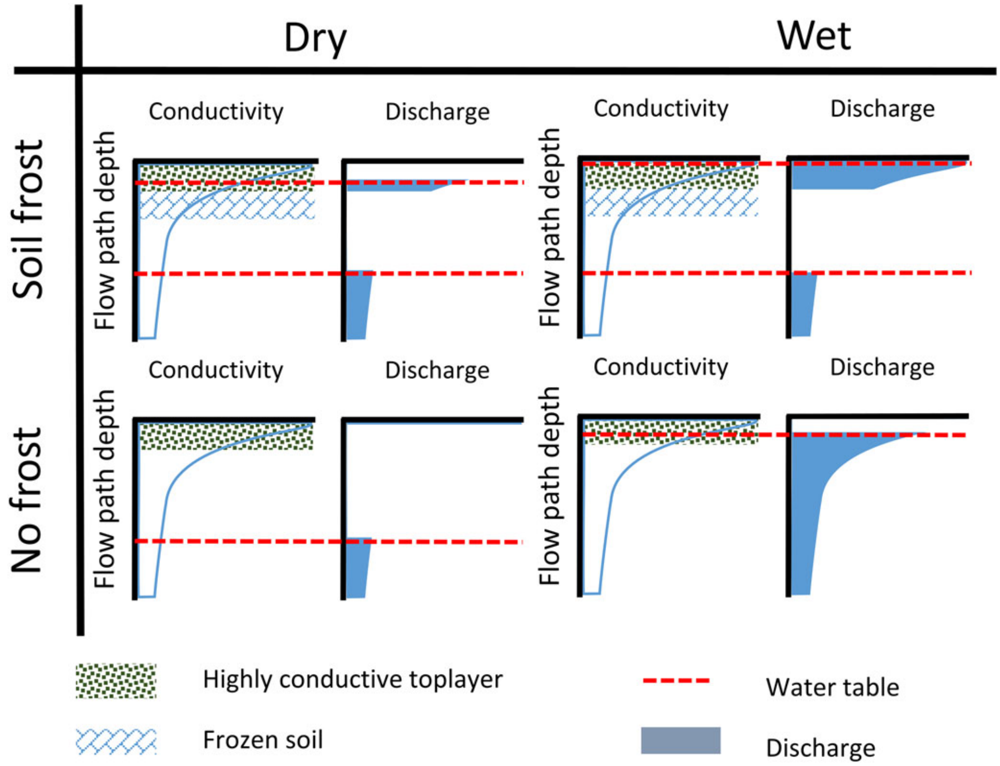

FIGURE 1 Conceptual interpretation of soil frost effects on the vertical soil profile. Contributions to streamflow are conceptualized as the cumulative horizontal flow component from active flow paths over the soil profile, assuming exponentially decreasing hydraulic conductivity with depth (lower right panel). Frozen soils cause a disconnection between shallow and deep groundwater (top panels). The shallow, high conducting top layer acts as a linear reservoir and dominates the storage–discharge relationship under wet conditions (top right panel). In thawed state, the soil profile provides a continuous contribution of shallow and deeper flow paths yielding less linear storage–discharge relationships (lower panels)

do not evaluate the curvature of recessions in the log–log space, although this can be interpreted as a (non-)linearity as well.

Typically, studies that use recession analysis focus on the spatial and/or temporal changes in coefficients α and/or β to understand and quantify changes in the catchment scale relationship between storage and discharge (Troch et al., 2013). Although recession analysis is a common method in hydrology, approaches differ widely (Stoelzle, Stahl, & Weiler, 2013), for example, using lumped data or individual recessions (Shaw & Riha, 2012). Also, methodological decisions (recession selection criteria) are variable and partially subjective, which should be accounted for in the interpretation and comparison of studies (D. N. Dralle, Karst, Charalampous, Veenstra, & Thompson, 2017). When recession analysis is used for estimation of model parameters, the temporal resolution, period, and duration of available data affect performance (Melsen, Teuling, Berkum, Torfs, & Uijlenhoet, 2014). Changes in recessions have been analysed by using a flow record from a single catchment with different conditions in time (Brauer, Teuling, Torfs, & Uijlenhoet, 2013) or by using flow records from multiple catchments with different characteristics and similar forcings (Bogaart, Van Der Velde, Lyon, & Dekker, 2016; Shaw & Riha, 2012). Specifically for northern latitudes, recession analysis has proven insightful in addressing spatial differences in natural- and human-induced landscape evolution (Bogaart et al., 2016), investigating the long-term change of aquifer properties (S. W. Lyon & Destouni, 2010; S. Lyon et al., 2009) and estimating thawing rates of permafrost in Canada and Sweden (Sjöberg, Frampton, & Lyon, 2013). Similarly, Brutsaert and Hiyama (2012) derived increasing active layer depths from streamflow records in Siberia.

1.3 | Snowmelt and ET effects on recessions

If soil frost is obstructing water flow in the subsurface, this is expected to be visible in streamflow recession behaviour. If soil frost partially will partition meltwater and rainwater into shallow and deeper flow paths and affect connectivity of water stores in the landscape, we expect particularly a change in the reservoir-linearity parameter β (Troch et al., 2013). The additional challenge we hereby specifically address is to separate soil frost effects from other seasonal effects on streamflow recessions, such as snowmelt and evapotranspiration (ET).

When precipitation and ET fluxes are included in the water balance relationship, Equation (1) can be written as (Teuling, Lehner, Kirchner, & Seneviratne, 2010)

$$-\frac{\partial Q}{\partial t} = \alpha Q^{\beta}\left(1 + \frac{E}{Q} - \frac{P}{Q}\right) \tag{2}$$

With E [mm/day] being ET and P [mm/day] an influx such as precipitation or snowmelt. During spring snowmelt conditions, we expect that additional inputs in the form of snowmelt affect recessions. Following Equation (2), snowmelt leads to less linear storage–discharge relationships, which corresponds to an increased slope, or in other words, this leads to steeper regression lines in a log–log space as in Figure 2. Similarly, Equation (2) suggests that ET leads to a more linear storage–discharge behaviour (flatter lines, Figure 2). It must be noted that we highlight here the slope changes by snowmelt and ET, not the curvature of the lines in log–log space. This effect has been demonstrated in data by Shaw and Riha (2012).

1.4 | Hypotheses

Combining our conceptual model with recession analysis theory, we formulated multiple hypotheses. For our study catchment, we consider that soils have a decrease in flow path conductivity with depth (Figure 1, lower left panel), which have been shown to yield non-linear storage–discharge relations (Rupp & Selker, 2005, 2006). Therefore, we expect more linear storage–discharge relations when flow dominantly occurs through shallow flow paths above the frozen soil layer. On the other hand, we consider the possibility that snowmelt and ET effects can lead to respectively less linear and more linear storage–discharge relations.

In order to test our central hypothesis that soil frost affects streamflow and to quantify this effect, we addressed the following specific hypotheses:

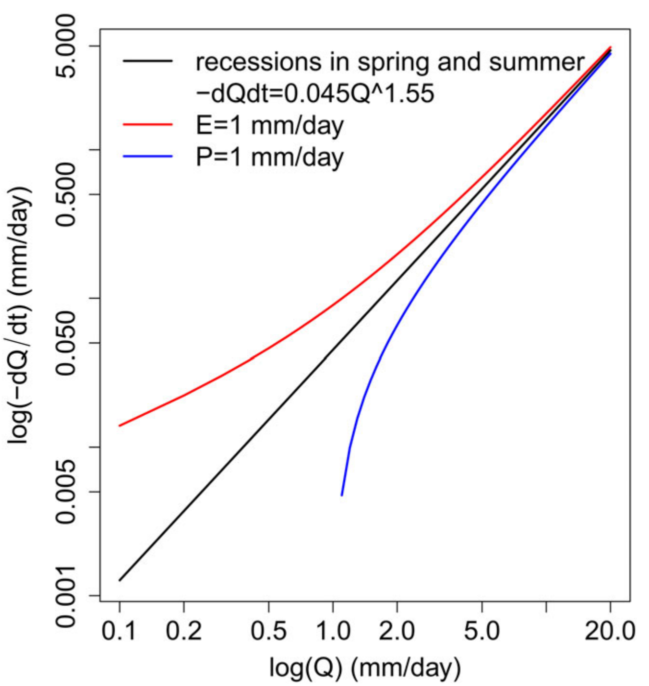

FIGURE 2 Conceptualization of evapotranspiration (ET) and water inputs (P) effects on recessions. On the *x*-axis, the average daily discharge (Q) is plotted over the negative change in discharge ($-dQ/dt$) on the *y*-axis. The black line represents Equation (1), with coefficients $\alpha = 0.045$ and $\beta = 1.55$. The effect of 1 mm/day of ET is presented as a red line, and 1 mm/day of precipitation presented in blue line (Equation 2)

- Hypothesis Test 1: If soil frost affects river recessions, we expect a significantly lower parameter β during spring recessions (with more frozen soils) compared with summer recessions (with less frozen soils).
- Hypothesis Test 2.ii: If snowmelt affects river recessions, we expect that winters with a lot of snow show a significantly higher parameter β during spring recessions compared with low snow winters. Moreover, when snowmelt is assumed, the dominant driver for the β-difference found under Hypothesis Test 1, we expect that low and high snow winters show a difference in β in the opposite direction as the difference found between spring (with snow) and summer (no snow).
- Hypothesis Test 2.iii: If summer ET affects river recessions, we expect that summers with high air temperatures and low cloud cover show a significantly lower parameter β than summers with low temperature and high cloud cover. Moreover, when ET is assumed the dominant driver for the β-difference found under Hypothesis Test 1, we expect that low and high ET summers show a difference in β in the same direction as the difference found between spring (low ET) and summer (high ET).

In the following sections, each of these steps are explained in detail.

2 | METHODS

2.1 | Study area

The Abisko research station in northern Sweden is one of the few high-latitude stations where this study can be performed, as long overlapping measurement time series of stream discharge, soil temperatures, snow depth, and meteorological variables allow testing of our hypothesis in the shrinking network of high-latitude research facilities (Laudon et al., 2017). Abiskojokka is a large subarctic headwater catchment (566 km^2) in northern Sweden (68°21′57.9″N 18°47′25.7″E) with an elevation gradient from mountain Adnetjårro at 1,755 to 330 m asl at the outlet into Lake Torneträsk (Figure 3). The postglacial landscape mainly consists of birch forest and small wetlands in the valleys, and subalpine and alpine vegetation zones on the slopes and high elevated areas. These (sub)alpine areas predominantly consist of heath and small shrub vegetation or exposed rock and thin soils. In the valleys, the soil development is estimated to be between 1.7 and 5.3 m, with regolith depths up to 30 m (Smedberg, Humborg, Jakobsson, & Mörth, 2009). In the Abisko region, sporadic and discontinuous permafrost is present of which the distribution is mainly determined by slope orientation and altitude (Niessen, Van Horssen, & Koster, 1992).

2.2 | Data

Daily streamflow of the Abiskojokka at the outlet is available for the period 1986–2015. Additional observations from Swedish Meteorological and Hydrological Institute, for both Abisko and the neighbouring

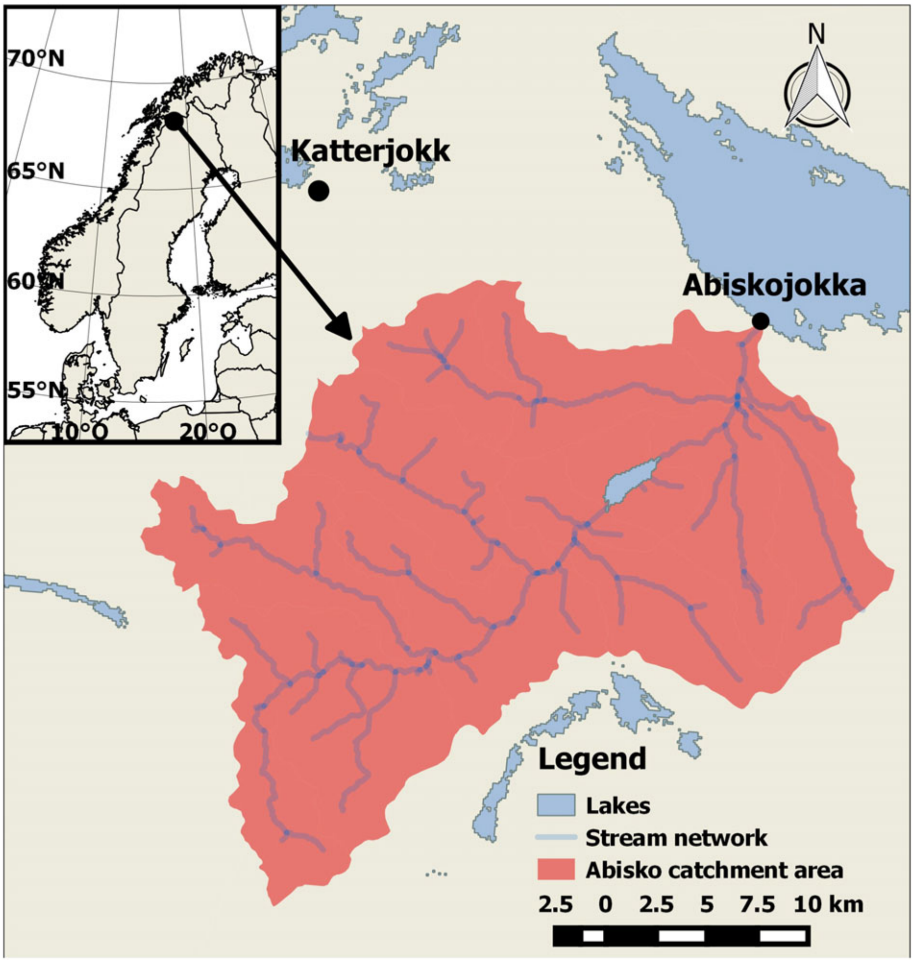

FIGURE 3 The study area Abisko in northern Sweden. The catchment is indicated in red (540 km^2) with the stream network in blue. The outlet of the river Abiskojokka is indicated in black. Katterjokk is a meteorological station nearby that represents high-altitude areas of the catchment

Katterjokk station, and the observatory of Abisko Research station were used over the same period. Katterjokk is a Swedish Meteorological and Hydrological Institute monitoring location near Abisko, which is representative for more elevated areas of the catchment. The observed variables used in this study are as follows: air temperature at midnight (in Abisko and Katterjokk), soil temperature at 0.5-m depth measured every fifth day (Abisko Research Station), daily average cloud cover (based on three hourly data from automated cloud altimeter at Katterjokk), and daily snow depth (Katterjokk). The air temperature records of Abisko and Katterjokk were linearly gap filled for respectively 4 and 13 (non-consecutive) values. The cloud cover data for June 1988 are missing but are available for the rest of the study period.

2.3 | Data processing

2.3.1 | Deconstructing the discharge record into spring and summer recessions

The onset of spring was defined using a degree day approach, which is the sum of daily air temperatures above 0°C, at midnight. A series of consecutive midnight temperatures above zero are in practise a good predictor of the snowmelt onset in spring. Every time the temperature fell below 0°C, the degree days are reset to zero. We defined the start of spring as the first day of the year when the degree day approach exceeded 20 degree days (this date varied between April 11th and May 25th). The blue sections in Figure 4 show the extent of the period of which spring recessions were considered.

After the spring, the start of summer is defined using snow depth records. Although snow disappeared rapidly in the lower latitude (Abisko) and midlatitude (Katterjokk) stations, the snowpack at higher latitudes can stay for an extended period and subsequently provide snowmelt water. For that reason, the summer threshold was defined as the first day that snow depth was 0 at Katterjokk (550 m asl) plus 21 days to account for delayed meltwater effects from the higher elevations (transitions were all between June 14 and July 24). This delay was based on fieldwork expertise in the area of Abisko. Moreover, varying the 21 days of delay (up to 2 days less, and 2 days more delay) showed no notable changes in the results. The end of the summer period was defined by three consecutive days with midnight temperatures below 0°C, which were between September 27 and November 21. The vertical bars in Figure 5a show the thresholds for that particular year. The orange section in Figure 4 shows the extent of the summer period.

With the spring and summer periods defined based on meteorological conditions, the next step was the selection of recessions. The

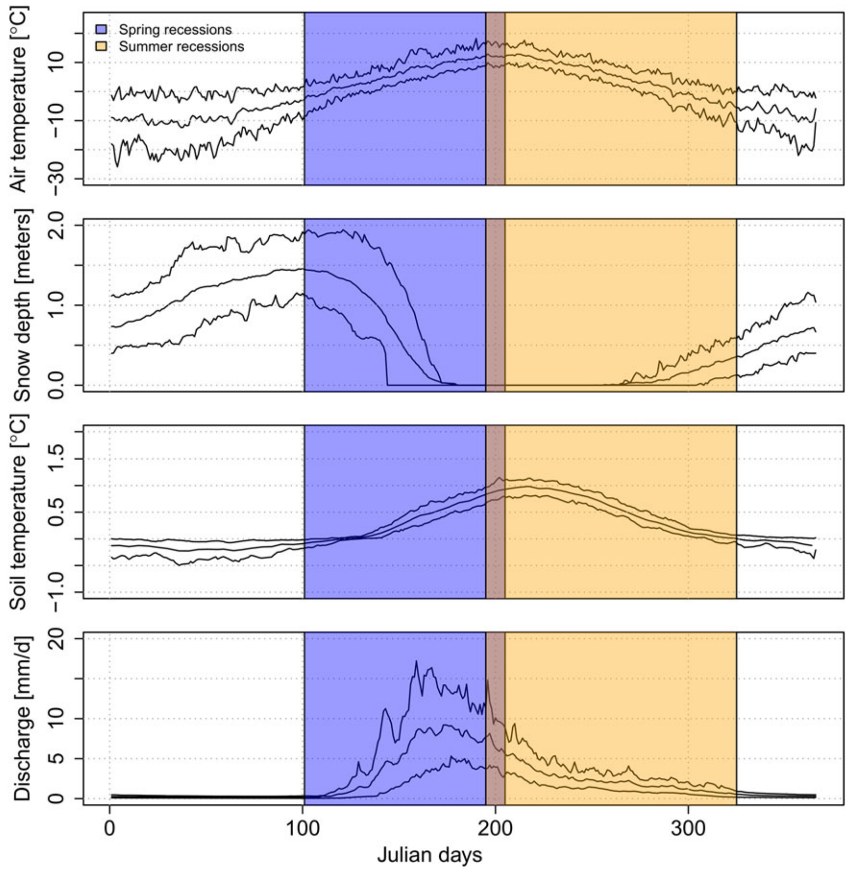

FIGURE 4 Daily regime in Abisko from 1986 until 2015. The black lines in the four panels represent daily mean, and 0.1 and 0.9 quantiles of respectively air temperature, snow depth, soil temperature, and discharge at the outlet of Abiskojokka. In blue and orange, the 0.1 and 0.9 quantiles of the spring and summer recession periods are indicated. These periods are for each year determined based on the governing snow depths and air temperatures

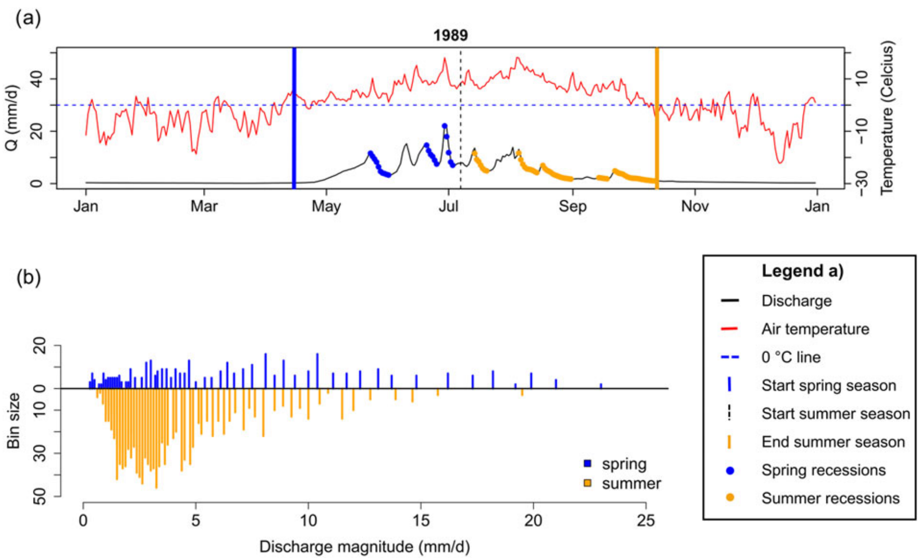

FIGURE 5 Processing of the Abiskojokka streamflow record into spring and summer data. In panel (a), the hydrograph for 1989 is shown (left *y*-axis) in black with selected recessions in blue dots for spring and orange dots for summer recessions. On the right *y*-axis, air temperature is plotted as a solid red line. The dashed blue line indicates the 0°C line. The vertical bars show the domain of both spring and summer seasons for the year 1989. In panel (b), the exemplar distribution of the bin size of the spring (blue, n = 57) and summer (orange, n = 68) recessions is presented over the range of discharge magnitudes that the recessions cover

methodological decisions in selecting recessions can affect the outcomes of recession analysis (D. N. Dralle et al., 2017). We defined recessions as a period of at least 5 days of consecutive decrease in discharge (dQ/dt < 0). To exclude effects of uncertainty in precipitation timing, the first day of each recession was removed. Because low discharges were usually overrepresented, we applied a binning procedure following Kirchner (2009; Figure 5b). Each bin covered at least 1% of the logarithmic discharge range, and within each bin, the standard error of the data points was smaller than half of the mean ($SD(-dQ/dt) \leq$ mean$(-dQ/dt)/2$). The recessions were plotted in a log–log space with the negative change in discharge ($-dQ/dt$) on the vertical axis and the mean discharge on the horizontal axis (as in Figures 2 and 6a). Linear regressions was applied to the bins, which resulted in a fit with intercept and slope, also referred to by coefficients α and β, respectively (Equation 1). We bootstrapped the recession data and repeated the procedure of binning and linear regression (n = 1,000), resulting in distributions of α and β for both spring and summer.

2.3.2 | Hypothesis Test 1: Spring versus summer recessions

To test the significant difference between the distributions of coefficients α and β for spring and summer, a non-parametric statistical test was applied with the use of bootstrapping. The motivation for this test compared with more conventional tests for significant differences between distributions, such as the student t test, was to avoid the assumption of Gaussian distribution. A schematic overview of this approach can be found in Figure S1. In our test, we determined if the squared mean difference between both distributions was significantly larger ($p < 0.01$) than the squared mean difference of a random selection from the total dataset (both distributions combined to a new distribution).

2.3.3 | Hypothesis Test 2.i: Soil frost

To further test the effect of soil frost on river recessions, we separated the spring and summer recessions from Hypothesis Test 1, based on the antecedent winter soil temperature. Winter soil temperature was considered as the mean soil temperature measured at 0.5 m depth in Abisko (from January until April). We considered cold winter soils (years when winter soil temperature was below or equal to the median) and warm winter soils (years when winter soil temperature was above the median). This resulted in four sets of recessions to which we applied the recession analysis and the non-parametric statistical test as described in Section 2.3.2. With this approach, we tested if there was a significant difference in recession coefficients α and β, based on cold and warm antecedent winter soil temperature, and if this difference corroborated the difference found under Hypothesis Test 1.

2.3.4 | Hypothesis Test 2.ii: Snowmelt

The continuous snowmelt input during the spring season is considered as a possible explanation for the difference between spring and summer recessions. To test this alternative hypothesis, we repeat the approach in Section 2.3.3, but instead of winter soil temperature, we used snow depth to split our recession datasets. The variable snow depth is defined as the annual maximum snow depth in metres, measured at the Katterjokk station (maximum snow depths occurred around March/April). We tested if there was a significant difference in recession parameters between both datasets and if this difference is stronger and in the same direction as the difference found for soil temperature under Hypothesis Test 2.i.

2.3.5 | Hypothesis Test 2.iii: ET

Several studies have found a strong effect of ET on discharge recessions (Shaw & Riha, 2012; Szilagyi, Gribovszki, & Kalicz, 2007; Teuling et al., 2010). Therefore, we tested if ET could also have caused the same significant difference as observed under Hypothesis Test 1. To this end, both the air temperature and cloud cover datasets were used to split the summer recession dataset into years with above and below median summer temperature and summer cloud cover (average from May until October). Both variables were expected to control summer ET. We tested if there was significant difference in recession parameters between both datasets and if the direction of this difference can

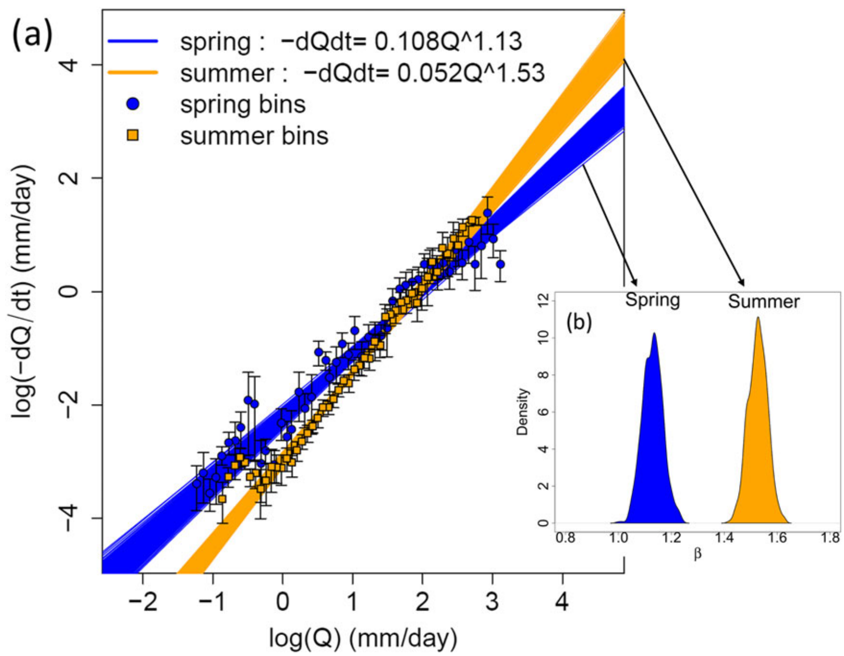

FIGURE 6 Recession plot in logarithmic space of the binned streamflow data. In panel (a), the means of all bins from the bootstrapping procedure are plotted for spring (blue points) and summer (orange points) with whiskers indicating the standard deviation in −dQ/dt for each mean bin. The x-axis shows the average daily discharge (Q), and the y-axis shows the negative change in discharge (−dQ/dt). The coloured lines (blue and orange) show the collection of fitted lines (n = 1,000) for both spring and summer as a result of the bootstrapping procedure. Panel (b) shows the distributions of coefficient β for spring and summer recessions, which correspond to the collection of fitted lines (n = 1,000) in panel (a)). The spring distribution is presented in blue and the summer distribution in orange

also explain the difference found under Hypothesis Test 1. We excluded the spring recessions in this analysis as it is not meaningful that future summer conditions affect prior spring flow.

3 | RESULTS

The climate regime in the Abisko region is dominated by an extended period of snow cover, in combination with freezing conditions (Figure 4). The mean annual air temperature in Katterjokk was −7.4°C and 0.1°C in Abisko. The Abisko air temperature record showed a weak significant trend (R^2 = 0.13, p = 0.05) in mean annual air temperature of 0.033°C/year. Mean summer temperatures increased 0.046°C/year (R^2 = 0.22, p = 0.008) in the studied period. These trends corresponded to the reported trends in the region (Irannezhad, Chen, & Kløve, 2015), with the exception that winter and spring temperatures alone did not show any trends in Abisko. Snowmelt amounts and timing did not show any trends. A visual inspection of hydrographs and temperature records (as in Figure 5a) showed that stream flow variation follows air temperature fluctuations especially in the spring periods. Soil temperature had an attenuated but similar regime as air temperature. The moment that soil temperatures at 0.5 m approximated 0°C was fairly constant: The third panel in Figure 4 shows that at the first of May, or Julian day 120, the 0.1 and 0.9 quantiles approximate the mean.

3.1 | Deconstructing the discharge record into spring and summer recessions

The number of days with discharge recession in the 30-year discharge record were 423 days during spring and 1,215 days during summer. These days were divided over respectively 60 and 183 individual recession events. These differences correspond with the higher flow regime in the spring period and with the decreasing flow in summer (Figure 4, lowest panel). The recessions were divided into respectively 58 and 70 bins, spanning the range of measured discharges (Figure 5 b). Discharge magnitudes of the bins in the spring period ranged from 0.9 to 22.9 mm/day, with a median of 5.3 mm/day. The median discharge of the spring is higher than the summer bins, and the bins are distributed over a wider discharge range compared with the summer recessions, which had a median of 2.4 mm/day and ranged from 0.4 mm/day to a maximum of 15.3 mm/day.

3.2 | Hypothesis Test 1: Spring versus summer recessions

Discharge recession analysis revealed a significant difference between spring (blue) and summer (orange) recessions (Figure 6a). The spring recessions (β = 1.13) have a significantly lower slope than the summer recessions (β = 1.54), indicating a close-to-linear storage–discharge relation in spring and less linear relation in summer. The regressions resulting from the bootstrapping procedures had a mean R^2 = 0.89 for spring and 0.95 for summer. For both coefficients α and β, the fitted parameter distributions do not overlap. Therefore, there is a significant difference in streamflow recessions between spring and summer periods.

3.3 | Hypothesis Test 2.i: Soil frost

The spring slope of β = 1.13 from Hypothesis Test 1 decreased further to 1.08 in spring when winter soil temperatures were low (below or equal to the median). This result showed that the effect of a lower (more linear) β in spring compared with summer (i.e., Hypothesis Test 1) is even stronger following winters with cold soils. The spring recessions following warm winter soils resulted in higher slopes (β = 1.20; Table S1), which suggests that a weaker effect was observed in winters with less extensive soil frost. The summer recession slope (β = 1.54; Figure 6a) was also changed as a result of low winter soil temperatures (β = 1.50), and the warm winter soil temperatures resulted in an increased slope of 1.64 (Figure 7).

Overall, the recession slopes in springs that followed winters with cold soils continued to decrease (from 1.13 to 1.08), which confirms our hypothesis that soil frost results in a stronger effect in the same direction as under Hypothesis Test 1. However, these findings do not exclude effects of snowmelt or ET, and therefore, we consider in the following sections the alternative hypotheses that snowmelt and ET explain the found recession difference in spring and summer.

3.4 | Hypothesis Test 2.ii: Snowmelt

Also for snow depth, the spring recessions were significantly different if we compared winters with low snow levels (β = 1.08) to winters with high snow levels (β = 1.20; Figure 7, Table S1). These are similar differences as found for soil frost under Hypothesis Test 2.i. Snow has a ground insulating effect; hence, there is a possible correlation between low snow levels and low soil temperatures. However, no significant correlations were found between our snow and soil temperature records. Also in summer, the results are very similar to the soil temperature effects: Winters with low snow levels resulted in summers with a slightly decreased slope (β = 1.53 to 1.51). High snow levels increased recession slope (β = 1.64). This finding is in line with the increase in β as a result of snowmelt input according to Equation (2) (Figure 2). All together, we found that winters with low snow levels resulted in a stronger linearization of spring recessions, which suggests that the effect found under Hypothesis Test 1 cannot be explained by snowmelt inputs. This leads us to reject Hypothesis Test 2.ii; however, it provided us with the insight that winters with high snow levels counteract the effect found under Hypothesis Test 1. This is also in accordance with the theory following Equation 2 and the blue line in Figure 2.

3.5 | Hypothesis Test 2.iii: ET

Cold summers resulted in the summer recessions to have a lower slope (β = 1.49) compared with the initial summer slope of 1.55 (Figure 7). Vice versa, warm summers produced an increasingly less linear flow recession (β = 1.64). Complementary to the air temperature, low cloud cover served as an indicator of summers with high ET rates. In summers with clear skies, the slope remains nearly the same (β = 1.54), but during cloudy summers, recession slope increases (β = 1.60). These changes do

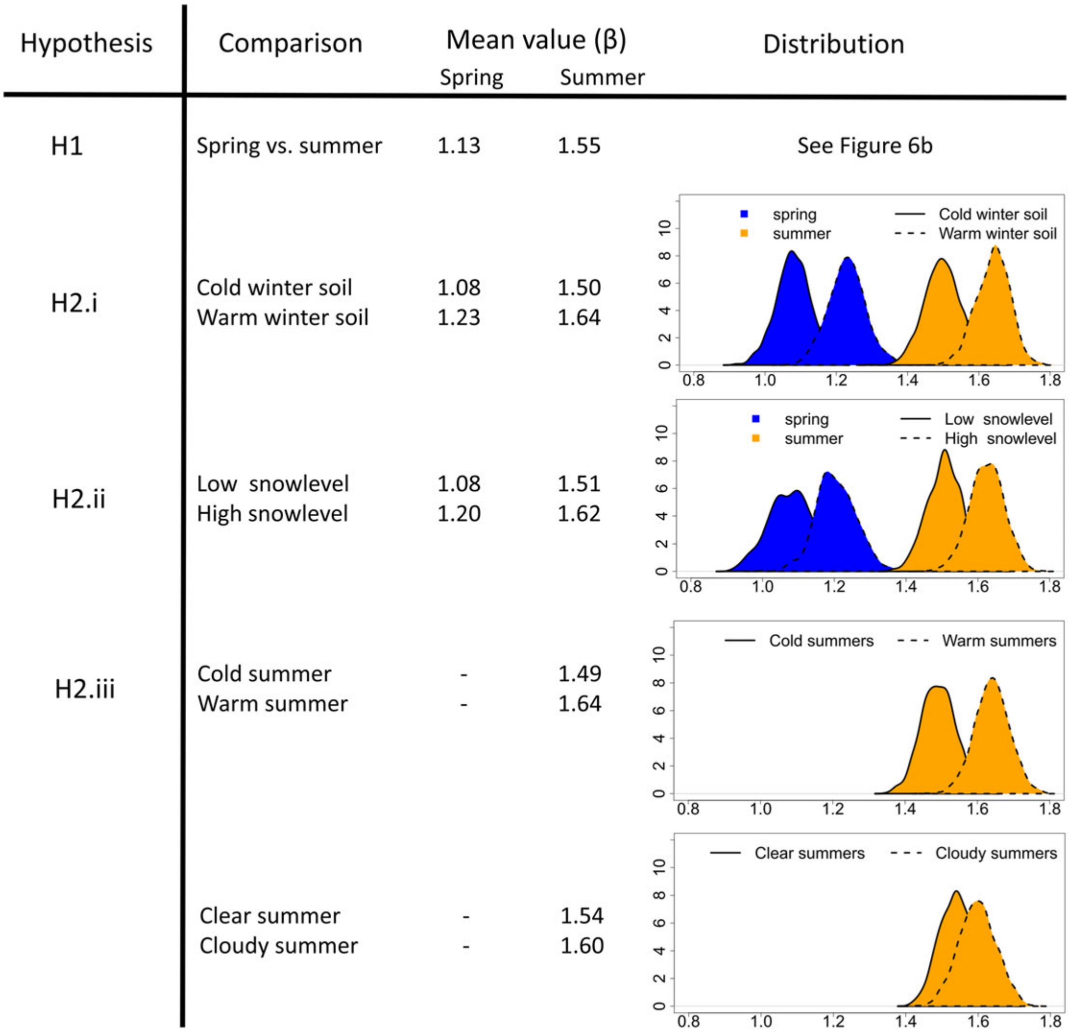

FIGURE 7 Matrix of changes in coefficient β under Hypothesis Tests 2.i, 2.ii, and 2.iii. In the first column, the comparison of each hypothesis is presented. The change in β of respectively spring and summer recessions is presented in the second and third columns. The last column shows how the distribution from Figure 6b changes according to the comparison. For example, the blue distribution in Figure 6b, with β = 1.13, shifts to β = 1.08 for spring recession that follow after winters with cold soils. Similarly, the β coefficient shifts to 1.23 for springs that follow winters with warm soils. The results for summer effects on previous spring recessions are left out, as these are not meaningful. However Tables S1 and S2 provide all values and corresponding statistical outputs, including coefficient α (also presented as 1/α)

not follow the expected lower slopes due to ET based on Equation (2) and the red line in Figure 2. Our results showed that the slope increased in the opposite direction due to ET of the spring versus summer difference found under Hypothesis Test 1. This suggested that the increased linearity in spring relative to summer is unlikely an effect of ET. However, the available records only served as proxies for ET differences between years. On a final note, the slope of the spring recessions were not altered by summer temperature and cloud cover.

4 | DISCUSSION

Based on our results, we consider our first hypothesis confirmed, namely, that during subarctic spring conditions, streamflow recession slopes significantly decrease (β = 1.13) compared with the summer period (β = 1.53). The more linear storage–discharge relation during spring compared with summer is interpreted as a propagating effect of preceding winter conditions. In our follow-up hypotheses, in-season differences of winter soil temperatures, snow depth, and ET during summer were examined in more detail. These were tested as likely candidates to explain this seasonal change in the recession slope. The in-season changes could have a variety of causes, but it demonstrated how robust our interseasonal findings are to variation and provides further insights in the change in slope found under Hypothesis Test 1.

Our second hypothesis that the difference in slope can be attributed to soil frost is considered plausible because spring recession slopes continue to decrease with cold winter soils (β = 1.08) and to increase with warm winter soils (β = 1.23). Previously, streamflow recessions have been found to become more non-linear with increasing discharge magnitudes and snowmelt inputs (Brutsaert & Nieber, 1977; Shaw, 2015). In our case, the highest discharges occurred during snowmelt in spring, which were found to approach linear storage–discharge behaviour. The lower discharge recessions during summer showed less linear storage–discharge behaviour. Also, Clark et al. (2009) showed that non-linear storage–discharge relationships are not necessarily observed with increasing flow magnitudes.

Hypothesis Test 2.ii showed that snow depth is also a likely control on the change in recession slope in spring, but not as suggested by Equation (2). Based on Equation (2) and Figure 2, increasing slopes would be expected by snowmelt inputs, but our results showed decreasing slopes in spring. However, spring recessions following winters with low snow depths show similar slope (1.08) as winters with cold soils. Winters with thick snow packs resulted in slopes in spring increasing again (from β = 1.13 to β = 1.20 in Figure 7), which possibly indicates that associated higher spring flow volumes (as the blue line in Figure 2) counteract the effect of soil frost. This finding could be the result of the relationship between snow depth and frost penetration (Henry, 2008), meaning that the insulating effects of snow and frost are correlated. As a result, Hypothesis Tests 2.i and 2.ii could show the same process but displayed as different variables. However, we consider this unlikely, given the lack of correlation between our snow and soil temperature records.

Poor insulation of the soil in winter could explain the control of winter soil temperature and snow depth on the linearity of summer discharge recessions. During summer, winter soil frost continues to significantly affect recessions. This could be explained by water being discharged through highly conducting top layers during spring, when soil frost disconnects deeper flow paths such as bedrock fractures, glacial deposits, or sedimentary deposits. A stronger disconnection between shallow and deep flow paths due to severe soil frost increases the deficit in deep groundwater. This could result to limited run-off generation from deep flow paths in summer. It may take far into the relatively dry Abisko summer before the continuum of shallow to deep flow paths is fully replenished and contributing to discharge. Another explanation for the significant effect of winter soil frost on summer discharge recessions is that frozen grounds persist far into summer and thus continue to affect river discharge during summer in high-altitude regions and north-facing areas. Propagation of winter processes to summer streamflow has readily been shown to cause droughts in cold regions (Van Loon et al., 2015).

The hypothesis (Hypothesis Test 2.iii) that ET could have caused the difference between spring and summer storage–discharge relationships is considered unlikely. The direction and magnitude of change of recessions (β = 1.53) in warm summers and summers with low cloud cover (1.54 and 1.60) did not suffice to consider Hypothesis Test 2.iii to be plausible as an explanation for the found differences under Hypothesis Test 1. Karlsen et al. (2016) found in Krycklan, a boreal catchment 600 km south of Abisko, lower slopes and less variability of β in spring compared with other seasons. In that study, ET showed a strong influence on β values. ET can explain seasonal variations in flow recession slopes as demonstrated by Teuling et al. (2010). However Bart and Hope (2014) showed that, for example, antecedent storage could lead to interseasonal variability recession rates as well. We showed in Figure 2 that an increase in ET is expected to yield more linear storage–discharge behaviour as found in the mentioned studies. However, this is in contrast to what we find in Abiskojokka, where spring with less ET compared with the summer yields more linear versus less linear storage–discharge relationships during high ET summers. Therefore, we reason that ET alone cannot explain the more linear storage–discharge behaviour during spring compared with summer observed in Abiskojokka. In addition, it is likely that the complexity and size of the Abiskojokka catchment is an additional factor that might influence the visibility of ET, compared with more simple or smaller catchments as used in studies that to find ET effects (Karlsen et al., 2016; Teuling et al., 2010). The similarly linearizing spring flows in boreal catchments (Karlsen et al., 2016) could be studied in more detail to compare soil frost and ET as a potential driver of seasonal changes in storage–discharge relationships.

Our lumped approach using bins, based on Kirchner (2009), differentiates from recession analysis based on individual recessions (D. N. Dralle et al., 2017). Alternatively, we could have used regression on each individual recession and lump these afterwards. However, the spring recessions are often short and have a high flow magnitude compared with summer recessions. Summer recessions can be long and can consist of long low flow periods. The binning approach from Kirchner (2009) allows to equally represent the entire flow range. Figure 6a shows that the slope of the regression is obtained over bins that are not entirely linear in log–log space: The bins at low log(Q) magnitudes (log(Q) < 0.5) for both spring and summer recessions seem to deviate from the regression. Low flow recession points can approach the limit of measurement error and can therefore be susceptible to interpreting an analytical artefact (Roques, Rupp, & Selker, 2017). An additional analysis excluding the low flow data resulted in a minor change in spring slope under cold winter soils: The β coefficient shifted from 1.08 back to 1.12, whereas for springs with low snow levels, the slope remained the same (Figure S2c and S2d). This shows that low flow magnitudes do seem to matter for soil frost effects on recessions (using a lumped approach). Conceptually, we expected that soil frost will mostly affect the high- and mid-range flow magnitudes, as soil frost likely affects the connectivity in the catchment and the partitioning between overland flow and infiltration. Nonetheless, the other hypotheses still are rejected, and we argue that soil frost remains as the most accountable cause for the more linear slopes in spring. A future study using an approach based on individual recessions could shed more light on the specific causes of temporarily linearized storage–discharge relationships in spring.

Although this study focused on changes in the β coefficient, also, the α coefficient can be studied to understand catchment scale processes. For example, S. Lyon et al. (2009) readily showed that the long-term trend of the α coefficient in Abiskojokka corresponded to increased thawing of frozen soil. With our methodological approach (analysing change in β within a catchment based on binned data), the interpretation of changes in the α coefficient is limited (D. Dralle, Karst, & Thompson, 2015). Considering α to represent aquifer properties for values of β close to 1 (Brutsaert & Nieber, 1977), future studies focused on changes in α could potentially provide insights on the relation between seasonal dynamics in stream flow generation and frost-induced, temporary changes in aquifer properties. Furthermore, our approach, using binned data to fit coefficients and comparing changes in β under certain meteorological conditions, allows us to compare outcomes within the catchment but has limited power to either extrapolate our findings to other Arctic catchments or infer parameters that could be used, for example, to model streamflow (D. N. Dralle et al., 2017). More appropriately, individual recessions could be studied to derive such information (Shaw & Riha, 2012). To further corroborate our conceptual understanding of catchment-integrated soil frost effects on streamflow generation, it would be valuable to compare Abiskojokka with other catchments in the region. Abiskojokka is a mountainous catchment, with shallow soils and relatively limited amount of lake and mire surface. Comparing catchments, Bogaart et al. (2016) showed that different landscape units (lakes, open areas, and forest) result in different coefficients. In a similar fashion, it would be interesting to study soil frost effects on streamflow of subarctic catchments that are more mire- or lake-dominated.

In summary, our results indicated that soil frost is an important control on the discharge recessions of the Abisko river. The Abiskojokka catchment is situated in an area where climate is warming rapidly, which will likely have large impacts on the hydrology. However, given the intricate interplay between groundwater, snow depth, soil frost, and air temperature, predictions of the spatial extent and depth of soil frost remains speculative. With more pronounced seasonal variations, our results and conceptual model indicate that yearly changing soil frost conditions will affect the discharge behaviour of subarctic rivers, resulting in increased variability of spring melt water distribution.

5 | CONCLUSIONS

In this paper, we tested whether the occurrence of seasonal soil frost affects streamflow of a subarctic river. The results indicate that streamflow in spring periods exhibits more linear storage–discharge behaviour (β = 1.13), whereas summer periods are much less linear (β = 1.54). Winter soil temperature anomalies best explain the significant changes in recession behaviour that propagate throughout the spring and summer months. This means that, compared with the majority of frost studies from boreal and alpine regions, this subarctic study observed three exceptional findings, namely, (a) soil frost is of major hydrological importance on a catchment integrated scale; (b) soil frost causes catchments to behave more like a linear reservoir (i.e., discharge varies linearly with storage); and (c) the seasonal dynamics in recession behaviour of this subarctic river cannot be attributed to snowmelt and ET, but instead is primarily controlled by the soil frost state. Finally, this study showed that there are still significant knowledge gaps in the hydrological functioning of high latitude catchments, which emphasizes the need to reverse the decreasing trend of Arctic research and long-term monitoring infrastructures.

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