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Received: 16 May 2024 | Accepted: 17 December 2024

DOI: 10.1002/vzj2.20396

Vadose Zone Journal

ORIGINAL ARTICLE

Insights into freeze–thaw and infiltration in seasonally frozen soils from field observations

Ines Sanchez-Rodriguez[1,2] | **Andrew Ireson**[1,2] | **Rosa Brannen**[1,3] | **Haley Brauner**[1,4]

[1]Global Institute for Water Security, University of Saskatchewan, Saskatoon, Saskatchewan, Canada

[2]School for Environment and Sustainability, University of Saskatchewan, Saskatoon, Saskatchewan, Canada

[3]Environment and Climate Change Canada, Ottawa, Ontario, Canada

[4]Water Security Agency, Saskatchewan, Canada

Correspondence
Andrew Ireson, Global Institute for Water Security, University of Saskatchewan, Saskatoon, SK, Canada.
Email: andrew.ireson@usask.ca

Assigned to Associate Editor Hailong He.

Funding information
University of Saskatchewan; Global Water Futures

Abstract

Snowmelt infiltration into frozen soils in seasonally frozen landscapes is a critically important hydrological process, with consequences for agriculture, water resources, and flooding. The partitioning of snowmelt between infiltration and runoff in any given location and in any given year is highly uncertain. While it is intuitive to expect lower infiltration capacities in frozen soils, extensive past field research has shown that infiltration is often the dominant flux over runoff during this process, and this is attributed to infiltration into air-filled macropores. Despite this understanding, we still lack models that can predict frozen soil infiltration reliably. In this study, we examine detailed field observations from the seasonally frozen Canadian Prairies to determine the controls on soil freeze/thaw, snowmelt partitioning, and groundwater recharge. We show how soil moisture, water table depth, snow water equivalent, and air temperature are all significant and confounding factors that determine soil freezing depth and snowmelt partitioning.

Plain Language Summary

Infiltration of snowmelt into seasonally frozen soil plays a key role in agriculture, water supply, and flood risk. However, predicting how much snowmelt will infiltrate into the soil versus runoff is uncertain. While it seems logical that frozen soils would absorb less water, research shows infiltration often dominates due to water entering air-filled channels called macropores. Despite this, reliable models to predict frozen soil infiltration are still lacking. This study uses data from the Canadian Prairies to explore what affects soil freezing, snowmelt behavior, and groundwater recharge. Our findings highlight that soil moisture, water table depth, snow amount, and air temperature all influence soil freezing and how snowmelt is split between infiltration and runoff.

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1 | INTRODUCTION

Seasonally frozen soils (SFS) are defined as soils where ice is present in the pore space for more than 15 days/year, but the soil profile is ice-free during the summer months (Zhang et al., 2003). SFS occur roughly where the mean annual temperature is just above 0° C, such that permafrost does not develop. They occur in regions cold enough to have significant snowfall during the winter, and an important characteristic of these soils is that snowmelt usually occurs while the soils are still frozen. During spring, snowmelt (M [mm/day]) and rainfall (P [mm/day]) combine to make the potential infiltration flux (I_p [mm/day]) as shown in Figure 1. I_P is partitioned into infiltration (I [mm/day]) and runoff (R [mm/day]). In bare soils, crops, and grassland, evaporation and transpiration can be assumed to be negligible during the melt period, though this is not the case in forested areas (Nehemy et al., 2022). Water that infiltrates into the soil profile may be retained in the soil as liquid water or ice or it may pass through the soil profile to form drainage (D[mm/day]) that goes on to recharge the groundwater. Drainage may occur immediately, in response to infiltration from rainfall or snowmelt events, or it may occur when the ice within the soil profile thaws, or some combination of the two. The soil water balance equation during the melt period is given by

$$\Delta S_L + \Delta S_I = I - D, \tag{1}$$

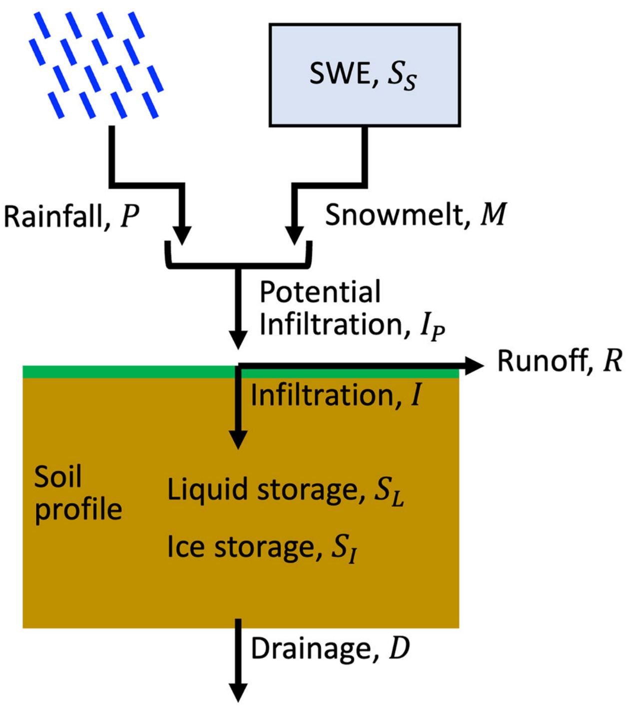

FIGURE 1 Soil water balance during the spring melt period. SWE, snow water equivalent.

> **Core Ideas**
>
> - Partitioning of snowmelt between infiltration and runoff is critically important in seasonally frozen soils, with important implications for flooding, soil moisture, and groundwater recharge.
> - Soil temperature, freezing depth, and infiltration capacity at the time of snowmelt depend on multiple confounding factors, namely, snow water equivalent, air temperature, soil water content, and water table depth.
> - Infiltration, drainage, and groundwater recharge occur rapidly in response to snowmelt, while the soils remain frozen, via preferential flow through air-filled macropores.

where ΔS_L (mm) and ΔS_I (mm of liquid water equivalent) represent the change in liquid and ice storage, respectively, and

$$I = I_P - R \tag{2}$$

and

$$I_P = P + M. \tag{3}$$

Partitioning of I_P into I and R depends on the state of the soil, which is in SFS will remain frozen during snowmelt, with varying liquid water and ice content from year to year, making prediction of I and R extremely challenging.

The Canadian Prairies are an area where SFS are present. The prairies contain about 80% of the agricultural land in Canada, where 90% of the nation's wheat is produced (Janzen et al., 1998; Jing-Qi et al., 2021). This area is semiarid and hence sensitive to climatic variability, with frequent floods and droughts (Basu et al., 2020; Bonsal et al., 2017; Lemmen & Canada et al., 1998; Sauchyn et al., 2011; St. Jacques et al., 2014). Historically, the worst floods in this area have been driven by snowmelt, for example, the 2011 floods (Ahmari et al., 2016; Brimelow et al., 2015; Buttle et al., 2016; Dumanski et al., 2015). Water availability for crops is a primary concern in the prairies (Sauchyn et al., 2011; St. Jacques et al., 2014). Much of the area is not irrigated, so soil moisture is recharged by snowmelt in the early spring (Lemmen & Canada et al., 1998; Shook et al., 2015). The partitioning of I_P on the prairies plays a critical role in determining the moisture availability for crops in the growing season (Appels et al., 2018), the amount of groundwater recharge (Mohammed et al., 2018), and the risk of flooding (Appels et al., 2018). This partitioning is highly uncertain in any given

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year and location (Mohammed et al., 2019), and we lack adequate methods to both predict this partitioning and subsequently apply these predictions to water management in the prairies. Direct observations of infiltration and runoff fluxes are not easy to obtain. Hence, we can make only indirect inferences about snowmelt water partitioning via observations of soil properties, soil moisture, streamflow, wetlands, and groundwater responses, which may be complemented by modeling and laboratory experiments.

One approach to quantifying snowmelt water partitioning is by direct or indirect measurements of surface runoff. In simple, small watersheds, the surface runoff can be estimated from observations of streamflow, snowpack, and precipitation (Shanley & Chalmers, 1999). In areas with closed topographic depressions, surface runoff may contribute directly to pond water level changes. In such a context, runoff can be approximated by determining the pond catchment area, the volume of snow on the ground, and the change in volume in the pond during melt (Mohammed et al., 2019). Hayashi and van der Kamp (2000) established area–depth (A–h) and volume–depth (V–h) relations for a number of ponds in St. Denis National Wildlife Area (SDNWA). At smaller scales, runoff plots can be constructed to capture runoff over some known contributing area of the field (Bayard et al., 2005). Mohammed et al. (2019) and Shanley and Chalmers (1999) used this approach to quantify infiltration partitioning on uplands in instrumented field sites in the Canadian Prairies in Alberta. They showed that infiltration was dominant over runoff. Their sites were subject to periodic midwinter melt events (unlike our site, where there is normally a single annual melt event), and they found that the runoff ratio increased later in season, due to refreezing of infiltration water in the soil macropores. They concluded that infiltration in frozen soils occurs as a preferential flow mechanism in macropores.

Another approach to quantifying snowmelt partitioning is to use information about the infiltration capacity of the frozen soil. Kane and Stein (1983) used infiltrometers to measure the infiltration capacity of frozen soils during the spring melt. In late fall, they removed the soil cover (plants and organic layer) and installed double-ring infiltrometers in open fields before the freezing process. During late winter, they removed the snow cover and measured the infiltration capacity using a constant head of water at 0°C. They found larger frozen soil infiltration rates were observed in dry versus wet soils and sandy versus silt loam soils, suggesting both texture and water content play a role in determining the infiltration capacity (Kane & Stein, 1983). van der Kamp et al. (2003) used infiltration rings to estimate the infiltration capacity at St. Denis, under cropped and grass landcover. They found much higher infiltration capacities in unfrozen conditions, but in frozen soils they found grass landcover had higher infiltration capacities because, they concluded, the macropores were undisturbed. Kane and Stein (1983) highlighted that a potential measuring error of the infiltration capacity might occur if the infiltrometer cylinders are not shaded to prevent a heat interchange with the frozen soil and consequent soil ice thawing/refreezing near the cylinder walls. In addition, soil disturbance to perform the gravimetric test may disrupt the soil conditions, leading to an inaccurate measure of the infiltration capacity. Therefore, researchers observed that this technique might not give accurate infiltration capacity measurements and may only be valid for qualitative analysis (Stähli et al., 1999).

Another approach to quantifying infiltration over the melt period is to use a water balance approach, which looks at the changes in observed water content over a vertical profile along with other water balance measurements. The volumetric liquid water content (θ_L [m^3/m^{-3}]) and volumetric ice content (θ_I [m^3/m^{-3}]) are related to the liquid and ice storage, S_L (mm) and S_I (mm of liquid water equivalent), respectively, by

$$\begin{aligned} S_L &= \int_{z=0}^{z_P} \theta_L \, dz, \\ S_I &= \int_{z=0}^{z_P} \frac{\rho_I}{\rho_L} \theta_I \, dz. \end{aligned} \tag{4}$$

And the total water content (θ_W [m^3/m^{-3}]) and total water storage (S_W [mm]) are given by

$$\begin{aligned} \theta_W &= \theta_L + \theta_I, \\ S_W &= S_L + S_I. \end{aligned} \tag{5}$$

Vereecken et al. (2008) and Lekshimi et al. (2014) describe different techniques to measure soil moisture. A very popular technique for measuring water content is the dielectric technique; however, this can only measure θ_L (Amankwah et al., 2022; Hayhoe et al., 1983; Smith et al., 1988; Spaans & Baker, 1995). Amankwah et al. (2022) and Patterson and Smith (1981) showed using a mixing model that ice is essentially invisible to dielectric probes, since the dielectric constant of ice is close to that of the soil solids and very different to that of liquid water. The gamma and neutron scattering techniques are able to measure the total water content (Lekshimi et al., 2014; Spaans & Baker, 1995; Yoshikawa & Overduin, 2005), yet due to the nature of the measurement (photon attenuation and neutron scattering), they are difficult to automate or use in cold winter conditions (Watanabe & Wake, 2009).

An important study in the Canadian Prairies that used the water balance method was reported by Granger et al. (1984). They collected data from 78 sites throughout the melt period. They assumed that I_P was equal to the change in observed snow water equivalent (SWE) over some time increment. They measured the total water content using (a) twin gamma probes to measure the change in total soil density and (b) soil core samples in-depth (every 10 cm) to calculate the dry soil

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density in the laboratory. The differences in dry and wet soil densities allowed them to estimate profiles of total water content, which is integrated over the profile to give S_W. Then, they assumed that the changes in S_W between sequential measurements is equal to the infiltration that occurred over that time increment, allowing them to compare I_P with I and thus also infer runoff. On the basis of this analysis, the authors described three types of infiltration: (a) *Unlimited infiltration*: All snowmelt infiltrates and there is no runoff, which they said occurs in cracked soils; (b) *Limited infiltration*: Some proportion of snowmelt infiltrates and some runs off, and they proposed an empirical equation to predict the partitioning (Equation 6); and (c) *Restricted infiltration*: All of the snowmelt runs off, due to a saturated frozen impervious layer at the soil surface, which may have been caused by refreezing of infiltrated water during midwinter snowmelt events.

$$I = 0.98\left(\frac{S_S}{\theta_{L,A}}\right)^{0.659}. \tag{6}$$

One limitation of the study of Granger et al. (1984) is that they ignore the fact that there may have also been soil drainage occurring between measurements, and therefore they potentially underestimate I, as is clear from Equation (1). Quantification of D is not simple, but the occurrence of drainage during the melt period can be inferred from responses in the shallow water table elevation during the melt period, as we report below. Despite this limitation, the results are insightful, and one important conclusion was that infiltration was very weakly associated with soil texture—the only soil physical property that determined infiltration was the presence or absence of macropores.

Field studies that have been undertaken to quantify infiltration rates in frozen soils using a range of different approaches are summarized in Table 1.

Abbreviations: SDNWA, St. Denis National Wildlife Area; TDR, time-domain reflectometry.

In this study, our research objective was to use field observations to develop qualitative insights into (a) the controls on soil freeze–thaw processes and (b) the mechanisms of runoff, infiltration, drainage, and groundwater recharge associated with snowmelt in SFS.

2 | MATERIALS AND METHODS

2.1 | Study area

The SDNWA is a conservation and research area established in 1968 (see Figure2). It is located at 52.2037 N, 106.1067 W in the Canadian province of Saskatchewan. SDNWA has an area of 361.5 ha (see Figure 2), and its altitude varies between 554 and 560 masl. The soil is brown and black clay loam. Land cover is managed by the Canadian Wildlife Service, and has varied between annual cropland, tame forage, and native grassland (Environment Canada–Canadian Wildlife Service, 2013; van der Kamp et al., 2003). In the period studied, landcover at both upland and lowland locations was native grassland.

The hydrography in SDNWA is dominated by wetland ponds (Bam et al., 2019) and the streamflow efficiencies are low (Shook et al., 2015). During spring, filled depressions can connect among them and eventually connect to a stream (fill and spill process; Spence & Woo, 2003).

The climate in SDNWA responds to the temperate seasons in the northern hemisphere (see Figure 2). According to the Climate Normals (1981–2010) reported by the Government of Canada (Government of Canada, 2021), about 80% of annual rainfall is concentrated in the period from May to September (cropping and harvesting months), with the average annual rainfall estimated to be around 263 mm. The annual average snowfall depth is 77 cm (Government of Canada, 2021), distributed mainly from November to March. The annual average temperature is 3.3° C; however, extreme cold and hot temperatures are expected during winter and summer. In wintertime (November–April), the average temperature varies between −13.9 ± 5°C, with extreme values reaching −43.9°C. During summer (June–August), the average air temperature varies between 18.2°C ± 1.8°C, with extreme values of 41°C (Government of Canada, 2021).

Two soil measurement transects have been established in the SDNWA, representative of lowlands and uplands. The lowland transect consists of three sets of soil profile instruments between ponds 107 and 108a (see Figure 2) and eight piezometers. The soil profiles are located at 555 and 557 masl and include instrumentation for soil water content, soil temperature, and matric potential at depths of 0, 5, 20, 50, 100, 200, and 300 cm (see Table 2). In this study, we only use soil profile 1 (Figure 2) to represent the lowland conditions, since the data quality in this profile was the highest (i.e., there were minimal gaps or outliers in the data, which can be caused by poor connectivity between the probe and the soil). The upland transect is near pond 125 at ∼550 and ∼557 masl; it contains four piezometers and a soil profile with the same instrumentation as the lowland site.

2.2 | Data

Observed data used in this paper come from two meteorological stations in the SDNWA and lowland/upland transects data (see Section 2.1). The set of instruments is listed in Table 1. We established soil transects in early 2014, and we have been collecting soil and groundwater data since then. These data have been reported before by Bam et al. (2019) and Brannen et al. (2015). Meteorological data are collected from a tower

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TABLE 1 Some strategies to measure the snowmelt water partitioning in seasonally frozen soils.

Author	Context	Snowpack	Changes in soil moisture	Drainage from soil column	Ground water monitoring	Comments
Granger et al. (1984)	Five snowmelt periods (from 1978 to 1983) in 90 sites on the Canadian Prairies	Snow surveys (during the snowpack accumulation until the snowmelt period)	Gamma density twin probes (measurements every 2 cm to a depth of 160 cm). Gravimetric test from core samples at the top of the soil column	Not measured	Not measured	No water table interactions were reported.
Stadler et al. (1996)	Two winter seasons (1992–1993 and 1993–1994) at Davos (Swiss Alps)	Snow depth (weekly during the wintertime)	TDR probes installed at 5-, 10-, 15-, 20-, 30-, and 40-cm depths	Not measured	Not measured	They measured the surface runoff and interflow using runoff plot.
Stähli et al. (1999)	Two winter seasons (1995–1996 and 1996–1997) at Ultuna (central Sweden)	Snow depth (weekly during the snowmelt period)	TDR probes installed at 10-, 17-, 25-, 35-, 50-, and 70-cm depths The total water content (ice plus liquid) was measured using a neutron probe	Outflow from four lysimeters.	Pressure transducers	No runoff was observed, indicating that snowmelt water infiltration was dominant.
Hayashi et al. (2003)	Three winter seasons (1998–1999, 1999–2000, and 2000–2001) at pond 109 of SDNWA (Canadian Prairies)	Snow surveys	TDR probe installed at 20-cm depth near to the depression. Gravimetric test from soil core samples at 20 cm over the representative points of the snow survey transects	Pond water levels measured manually and continuously by automated ultrasonic depth sensor and vibrating-wire pressure transducers	Piezometers installed near the pond 109S	The authors selected pond 109 because it does not receive runoff from other depressions and is representative of many small depressions in SDNWA. Complementary information from several depressions near was also collected.
Bayard et al. (2005)	Two winter seasons (2000–2001 and 2001–2002) at two sites in the southern Swiss Alps	Snow depth (daily during the snowmelt period)	TDR probes installed at 5-, 10-, 15-, 20-, 30- and 40-cm depths	Outflow from a lysimeter installed at 90-cm depth.	A piezometer was installed at 50-cm depth	The authors also measured the lateral surface and subsurface runoff. During the spring melt period, they did not find interactions between the vadose zone and the water table.
Iwata et al. (2008b)	One winter season (2002–2003) at two sites in Tokachi District, Hokkaido, Japan	Snow surveys (during the snowpack accumulation until the snowmelt period)	Water content reflectometer (WCR)—TDR. Depths of 0, 2, 5, 10, 20, 30, 40, 50, 60, 70, 80, 90, and 100 cm	Not measured	Not measured	Water table is located at 8-m depth. No water table interactions were reported.

(Continues)

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TABLE 1 (Continued)

Author	Context	Snowpack	Changes in soil moisture	Drainage from soil column	Ground water monitoring	Comments
Iwata et al. (2010)	One winter season (2005–2006) in one site located at Tokachi District of northern Japan	Snow surveys (twice a week during the winter period)	TDR probes installed at 5-, 10-, 25-, 35-, 45-, 55-, 65-, 75-, 85-, 95-, and 105-cm depth	It was calculated using soil moisture storage data at frozen and unfrozen depths		The water table is located at 8-m depth. No water table interactions were reported.
Mohammed et al. (2019)	One winter season (2016–2017) in three sites on the Canadian Prairies	Snow surveys (different date during the winter period)	TDR probes installed every 10 or 20 cm until 105 cm of depth	Not measured	Piezometers at different depths	Water table responses during the spring melt.

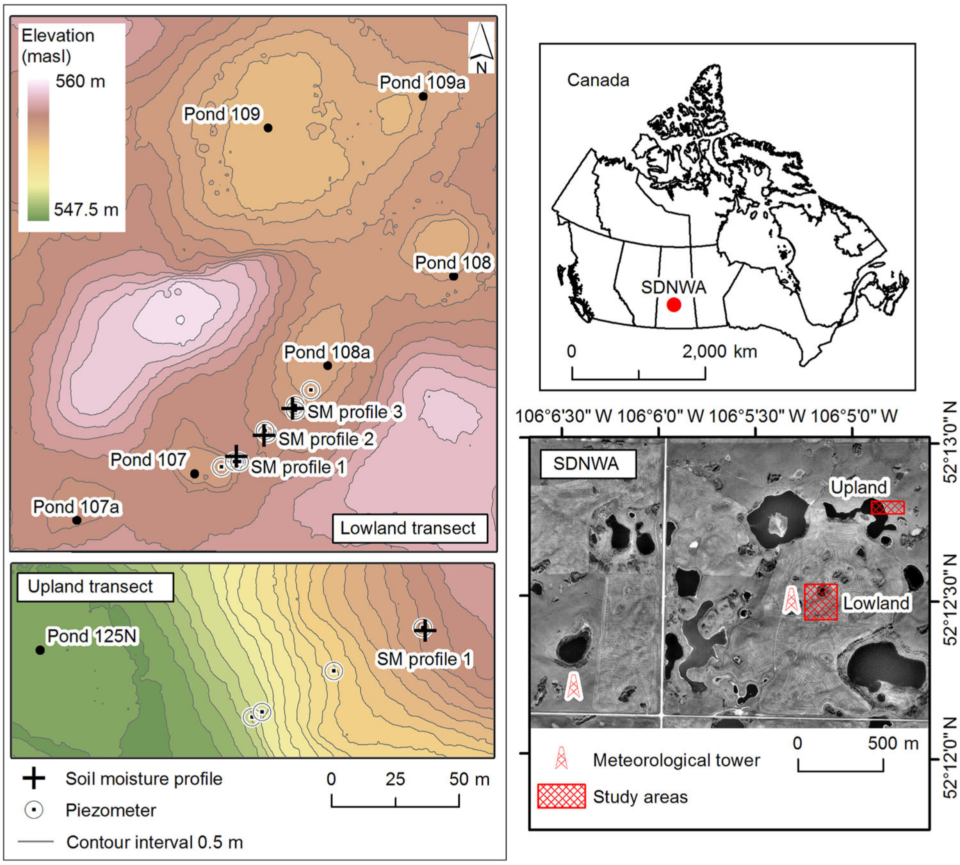

FIGURE 2 Research site map. From left to right: (a) elevation contour map where SM represents the soil moisture profiles of (upland and) lowland transects (see Figure 3), (b) location of St. Denis National Wildlife Area (SDNWA) in Canada, and (c) location of the meteorological stations.

on the site that was installed in 2013. High wind speeds impact the accuracy of the precipitation (rainfall + snowfall) measurements due to lateral mass losses or gains into the gauge. It has been estimated that corrections for solid precipitation under-catching may increase the registered amount from 13% to 19% (Pan et al., 2017). Furthermore, blowing snow affects the amount of snow on the ground at any given location (Pan et al., 2017). To estimate the SWE (mm) timeseries, we used the MESH model (Modélisation Environnementale communautaire—Surface Hydrology) (Davison et al., 2019; Dornes et al., 2008; Haghnegahdar & Razavi, 2017; Haghnegahdar et al., 2015; MacDonald et al., 2009; Mekonnen et al., 2014; Soulis et al., 2011; Yassin et al., 2019). The model was driven by local observations of precipitation measured by the Geonor T200-B weighing gauge, with a correction factor applied to account for blowing snow. The correction factor

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TABLE 2 List of data and instruments available in SDNWA to be used in the research.

Variable	Instrument	Time resolution
Volumetric water content (m^3/m^3) and soil temperature (˚C)	Stevens HydraProbe sensors at depths of 0, 5, 20, 50, 100, 200, and 300 cm bgl	hourly
Wind speed (2 and10m) and direction (10 m)	Met One 14A (2 m scaffolding tower), Campbell Scientific CSAT3 (10 m scaffolding tower) NRG 200 (10 m scaffolding tower)	30 min
Air temperature (˚C) and relative humidity (%)	Vaisala HMP45C (1.5-m scaffolding tower)	30 min
Net radiation (W/m^2)	Kipp and Zonen CNR4 (10-m scaffolding tower)	30 min
Precipitation	Geonor T200-B weighing gauge (mast tower) and Texas Electronics TE525m TBRG (0.5 m mast tower)	hourly
Water table elevation/hydraulic head	Solinist Leveloggers and Solinist Barologger (for barometric correction) installed in piezometers	hourly
Snow depth	Two traditional snow survey transects	annually

Abbreviation: SDNWA, St. Denis National Wildlife Area.

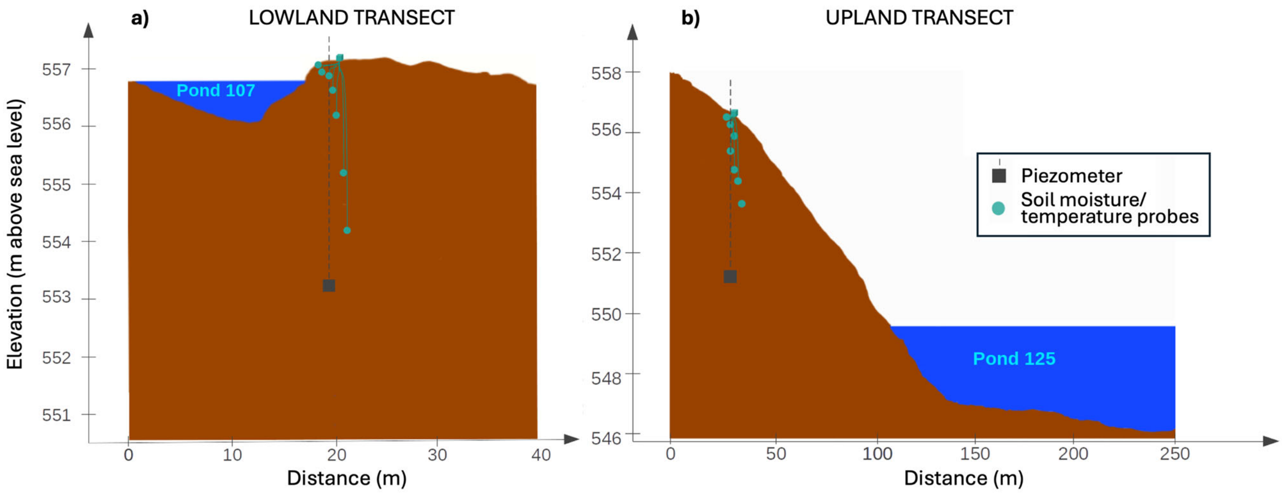

FIGURE 3 Cross-sections of the two instrumented transects in the lowland (a) and upland (b).

was optimized such that the simulated and observed SWE at peak snowpack matched.

3 | RESULTS

In Figure 4, observations of accumulated air temperature, SWE, soil freezing depth, liquid soil moisture storage, and water table depth for upland and lowland transects are plotted. Note that for air temperature and SWE we use site average data, whereas the soil and groundwater data are specific to each transect.

Accumulated temperature, with units of degree-days, is a useful measure of seasonal temperatures that is used for tracking plant growth as well as freeze–thaw processes. This shows large interannual differences in temperature at our site that are harder to recognize using daily air temperature alone. We found that 2015–2016 was the warmest year and 2018–2019 was the coldest year in our record.

There are large interannual differences in SWE, caused by differences in temperature and precipitation from year to year. For example, in 2016–2017, there was an early snowfall event that melted out completely in October. In 2019–2020, we had the greatest overall SWE and the latest snowmelt. 2015–2016 had a notably smaller snowpack. Since snow plays an important role in insulating the ground from changes in air temperature, early snowfall is expected to reduce soil freezing, while a late-melting snowpack is expected to result in delayed soil thaw.

The average freezing depth is ~80 cm in the lowland and ~100 cm in the upland and is always deeper at the upland site. 2015–2016 had very shallow freezing depths, meaning

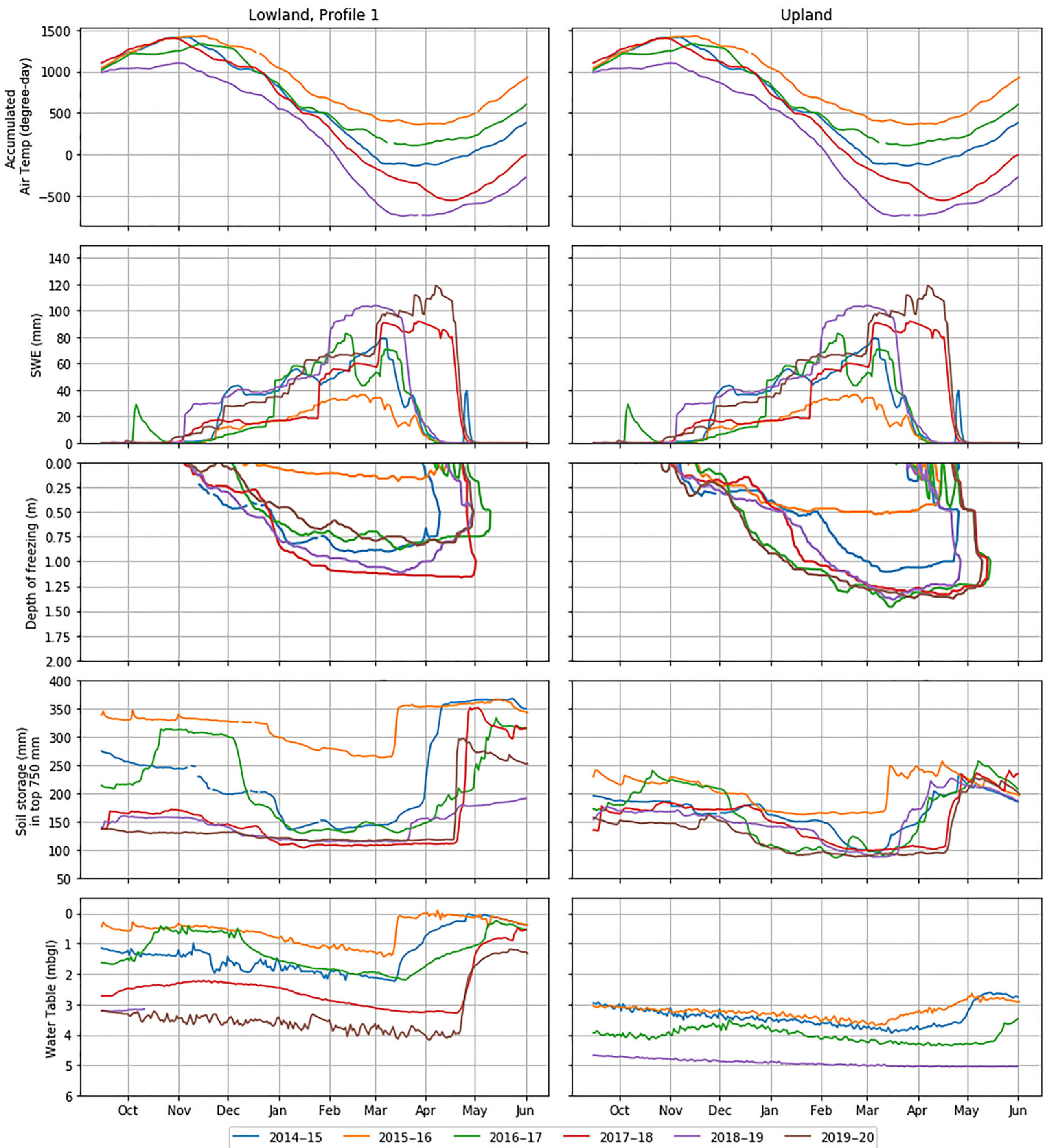

FIGURE 4 Observations of accumulated air temperature, snow water equivalent, soil freezing depth, soil liquid water storage, and water table depth, for lowland and upland sites, comparing the winter/melt periods for different years. SWE, snow water equivalent.

the soil temperatures in that winter were warmer than in other years. 2017–2018 had the deepest soil freezing depth at the lowland, while in the upland the maximum freezing depth was relatively consistent from 2016 to 2020. We discuss the reasons for these differences below.

In terms of soil liquid water storage, there is a seasonal pattern that is consistent with soils freezing in the winter and thawing in the spring. The increases in storage in the spring can be attributed to either infiltration of new liquid water or thawing of existing ice within the soil, and there is no direct way to determine the source of this water. In general, the lowland site has wetter soils than the upland, as would be expected, and this is especially so immediately after the soil has thawed. There is more variability in storage in the lowland site, with markedly wetter soils in 2015–2016, and with large variations in moisture storage at the end of the growing

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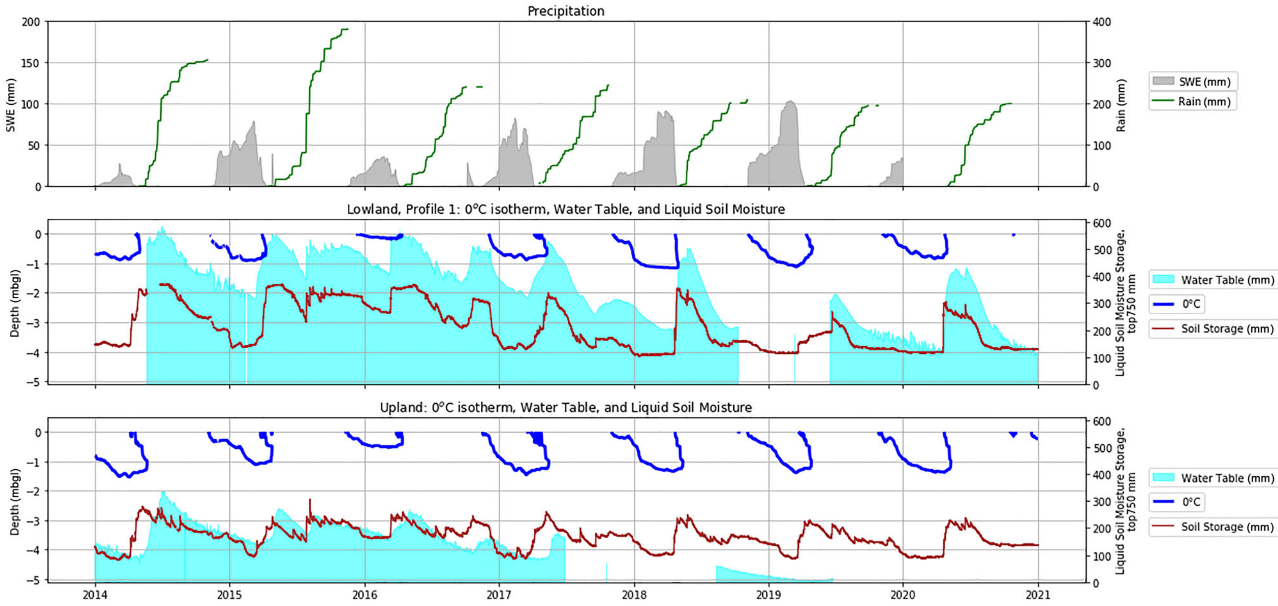

FIGURE 5 Snowpack depth, accumulated rainfall from spring to fall, the frozen depth, the liquid water soil storage, and the water table depths for lowland and upland site. SWE, snow water equivalent.

season (October) and the beginning of the growing season (June). The upland site has less variation and drier soils overall.

The water table elevations have a consistent seasonal pattern of gradual decline over the winter months, followed by rapid increases in the spring, coincident with snowmelt. The water table is deeper and less variable, with less rise in the spring, beneath the upland site. In the lowland site, the water table came up to the ground surface in 2014–2015 and 2015–2016, but in subsequent years, it was lower. 2015–2016 had the highest water table at both sites. 2019–2020 had the lowest water table at both sites, falling below the measuring depth of the piezometer in the upland.

In Figure 5, the snowpack depth, accumulated rainfall from spring to fall, the frozen depth, the liquid water soil storage, and the water table depths are plotted for the upland and lowland site as continuous time series. From 2015 to 2019, there is a decline in annual summer rainfall, while SWE amounts increase over the same period. There is a drop in the water table in the summers at both the upland and lowland site, indicating that we are moving from a wet period into a dry period. Patterns of soil liquid water storage in the lowland are consistent with the patterns of the water table and also show drying. In the upland, there is no clear declining trend in the soil liquid water storage, while the groundwater levels do decline. Figure 5 shows the timing of events in the spring period, when we see seasonal increases in soil moisture and groundwater are coincident with the timing of snowmelt, which happens while the soils remain partially frozen. Soil thaw is completed after snowmelt is complete and after the soils and groundwater levels have already responded.

4 | DISCUSSION

4.1 | Controls of soil freeze–thaw

Soil freezing is controlled by temperature (colder temperatures result in more freezing), snowpack insulation (more SWE, especially early in the winter, results in warmer soils and less freezing; Hardy et al., 2001; Hirota et al., 2006; Isard & Schaetzl, 1998; Iwata et al., 2008a, 2008b; Wilson et al., 2020; Zhang Tingjun, 2005), and soil moisture content (wetter soils have more stored latent heat which reduces freezing). In the field, these factors may vary in an independent manner, and the state of the soil will depend on the combination of these and not by any single control. As a result of this, there are a number of seemingly counterintuitive observations that can only be explained by considering all three factors listed above, including the following:

1. *The shallowest snowpack coincides with the shallowest freezing depth*: In 2015–2016, we have the shallowest snowpack, but also the warmest soil, contrary to what would be expected, all else being equal. This is because 2015–2016 was the warmest year, with the highest antecedent soil moisture coming into the winter, especially evident in the lowland site in Figure 4.
2. *The coldest year did not produce the deepest freezing depth*: The overall coldest year (2018–2019) had relatively early snowmelt and because thaw started earlier, it did not have the deepest depth of freezing, especially in the lowland, as seen in Figure 4.

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On the other hand, one control that did appear to behave in a manner consistent with expectations was soil moisture storage—it appears that wetter soils freeze the least, while drier soils freeze the most. This also explains differences in freezing depth between the upland site (consistently drier soils, consistently deeper freezing depths) and lowland site.

We see (coincidentally or not) that the year with the wettest antecedent soil moisture was also the year with the warmest winter (2015–2016). If this were not the case, it is unclear which control would dominate the outcome in terms of freezing—is temperature more important or is latent heat (higher in wetter soils) more important?

At the lowland site the water table was very close (within around 1 m) to the soil freezing front during freezing, whereas at the upland site this separation was greater than 2 m (Figure 5). It is possible that cryosuction, that is, the phenomenon of liquid water being drawn toward the freezing front, may be significant in the lowland site, since the shallow water table keeps the soils wet and relatively mobile (wetter soils have a higher hydraulic conductivity). This process has been described and modeled in previous studies (Hansson et al., 2004), but it is still unclear how significant this might be from a water and energy balance perspective. Since direct observations of subsurface fluxes of water are not possible, insights will come from future research that combines physically based numerical models with field observations.

4.2 | Snowmelt infiltration mechanisms

At the lowland site, we observe recharge to the soil and groundwater occur during the spring melt period. In most years (2018 being the one clear exception), the water table rise in the lowland begins at the onset of snowmelt, and takes place before the soils have completely thawed (Figure 5, middle panel). In 2018, snowmelt and soil thaw both occur relatively late and happen quickly, so the water table response, which is also rapid, coincides with both of these events. In the upland site, the water table is deeper (average depth around 3.5 m as compared with 1 m in the lowland). The water table responses are consistently delayed as compared to the lowland, but typically still occur while the soils remain frozen (2017 being an exception). The soil moisture responses in the lowland occur simultaneously with the water table responses, whereas in the upland the soil moisture recharge occurs more rapidly. Therefore, in the upland storage in the unsaturated zone, below the frozen layers, seems to act to delay groundwater recharge.

The infiltration and recharge that is occurring in the upland and lowland sites while the soils remain frozen is widely interpreted as happening via air-filled macropores (Granger et al., 1984; Huang et al., 2018; Mohammed et al., 2018, 2019, 2021). The soils at our field site are known to contain macropores (van der Kamp & Hayashi, 2009). In unfrozen soils, including soils with macropores, air entry occurs into the largest pores first, and progressively smaller pores are emptied as the matric potential reduces (i.e., becomes more negative). In a given unsaturated, unfrozen state, the largest water-filled pore size corresponds to the matric potential, and this is captured in the soil moisture characteristic relationship. Therefore, in soils with macropores, the macropores will generally be empty except when passing infiltration or close to/below the water table. When soils freeze, they are subject to freezing point depression, whereby the pore water in the largest pore freezes first (Amankwah et al., 2021; Kurylyk & Watanabe, 2013; Mohammed et al., 2018). As the temperature drops, progressively smaller pores freeze. As a result, in a given unsaturated and frozen state, the partitioning of liquid water, ice, and air in the pore space is determined by the total water content (which could be related to the pre-freezing matric potential, if no movement of water has occurred into/out of the soil control volume) and the soil temperature. This is depicted in Figure 6, where we see that ice exists in the medium-sized pores, with liquid water in the smallest pores and air in the largest pores. For this reason, it is expected that unless the soils were saturated at the time of freeze-up, the largest pores (whether macropores or not) will remain air-filled, and this controls the soil's infiltration capacity. In a modeling study, Huang et al. (2018) showed that models failed to reproduce the observed increases in soil moisture and groundwater without the inclusion of some form of preferential infiltration, which in that study was included in a very simplified way. A more physically robust modeling approach was developed by Mohammed et al. (2018) and applied in Mohammed et al. (2021). The dual permeability model assumed that flow can occur through air-filled macropores when the soils are frozen, and was able to simulate the observed behavior of frozen soils in laboratory and numerical experiments. There is therefore strong evidence from the field (this study as well as Mohammed et al. [2019]) and modeling studies (Huang et al., 2018; Mohammed et al., 2021) that preferential flow through macropores is a critically important process in frozen soils, and that snowmelt is able to recharge deeper soils and groundwater by bypassing the frozen zone of the soil.

Notable midwinter melt events occurred in January 2015 and 2017, where the snowpack was reduced (Figure 5, top panel). These events result in small water table responses in the lowland, suggesting again that water was able to percolate through the frozen soil by preferential flow through macropores. This was a major feature in the study by Mohammed et al. (2019) in Alberta, Canada, where midwinter melt events are more common than they are in Saskatchewan, Canada. They describe how refreezing of the infiltrated water associated with midwinter melt events can reduce the infiltration capacity of the soils so that the runoff ratios increase after midwinter melts have occurred.

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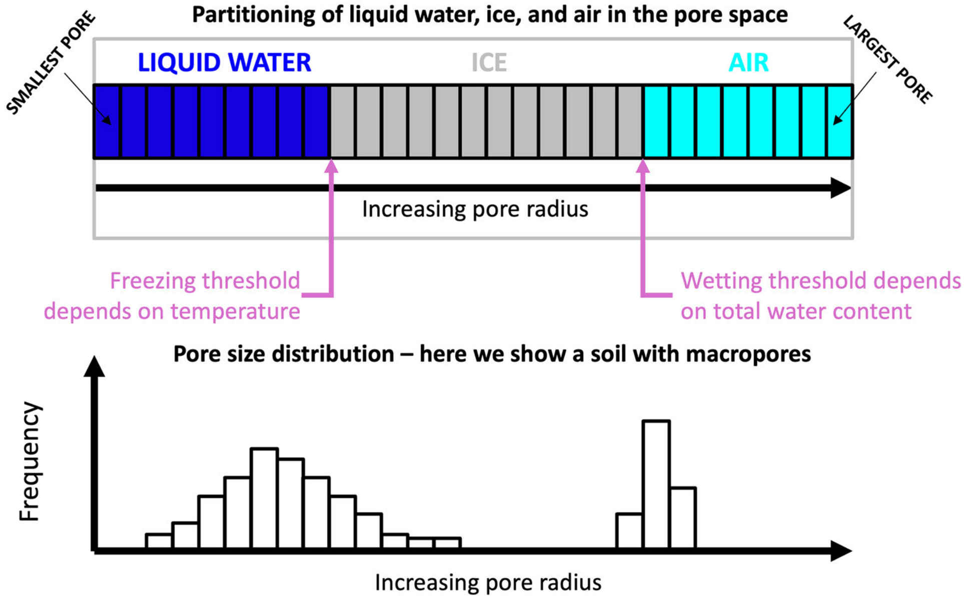

FIGURE 6 Conceptual model of the partitioning of liquid water, ice, and air in the soil pore space. The model should generalize for soils with or without macropores—the hypothetical distribution shown is for a soil with macropores.

In 2017, there is a secondary water table response that appears to coincide with the complete thaw of the soil (Figure 4, third and fifth panels on the left), but may also be driven by spring rainfall (Figure 5, top panel). It is not clear whether the thawing of the soil profile results in the release of water from the soils that recharge the groundwater, but this is a possible mechanism that warrants further study.

4.3 | A conceptual model for the hydrological processes during soil freeze–thaw in seasonally frozen regions

Consider a seasonally frozen soil column where liquid water, water vapor, ice, and air coexist along with the soil particles (see Figure 7). The soil column is considered frozen if the pore space is partially ice-filled (Appels et al., 2018). Seasonally frozen soils change from an unfrozen state to a frozen state in early winter (freezing process) and from a frozen to an unfrozen state (thawing process) in spring. We use the term "melt" to refer to the melting of the snowpack and "thaw" to refer to the thawing of the ice in the soil pore space. From early winter to spring, the system goes through two stages described below.

Stage 1—Early and midwinter: (from ~mid-November until ~mid-March; see Figure 7). The soil column loses heat to the atmosphere, which reduces soil temperature (Iwata et al., 2008b; JunPing et al., 2017). At temperatures below 0°C, the soil's freezing starts progressively as a freezing front. A portion of soil liquid water changes its phase to ice, but some unfrozen portions of water will remain as films around soil particles due to absorptive and capillary forces (Niu & Yang, Z., 2006; Spaans & Baker, 1996) and due to solutes in the pore water (Amankwah et al., 2021), both of which result in freezing point depression. If the soil is dry, the frozen front will reach deeper depths. As the frozen front progresses into the ground, the liquid water flows toward the frozen front (Hansson et al., 2004; JunPing et al., 2017; Kinosita, 1979). If the water table is shallow enough, it will interact with the frozen zone because frozen conditions create a vertical hydraulic head gradient that induces a mass flux from the deep soil column layers and water table to the upper soil layers.

Parallel to the soil freezing, the snowpack starts its formation, and it gets deeper through the midwinter due to continuing snowfall events. Eventually, the snowpack reaches its maximum thickness just before melting. At the same time, the frozen depth reaches its maximum depth just prior to the onset of thaw. During midwinter, the liquid water content within the frozen soil zone remains low, and the water table continues to drop due to lateral drainage and possible migration driven by cryosuction until the snowmelt starts.

Occasional changes in the air temperature to values above 0°C deplete the snowpack. Such meltwater infiltrates the soil and percolates through the soil column via macropores until it reaches the water table. In the transit of this water through the soil column, some portions may refreeze. Such phase changes the frozen depth by an interchange of latent heat into the soil matrix.

Stage 2—Late winter: (from ~mid-March until ~mid-April; see Figure 7). Once the snowpack starts to melt, some portions of the meltwater become surface runoff, and other portions infiltrate into the still-frozen soil column, and

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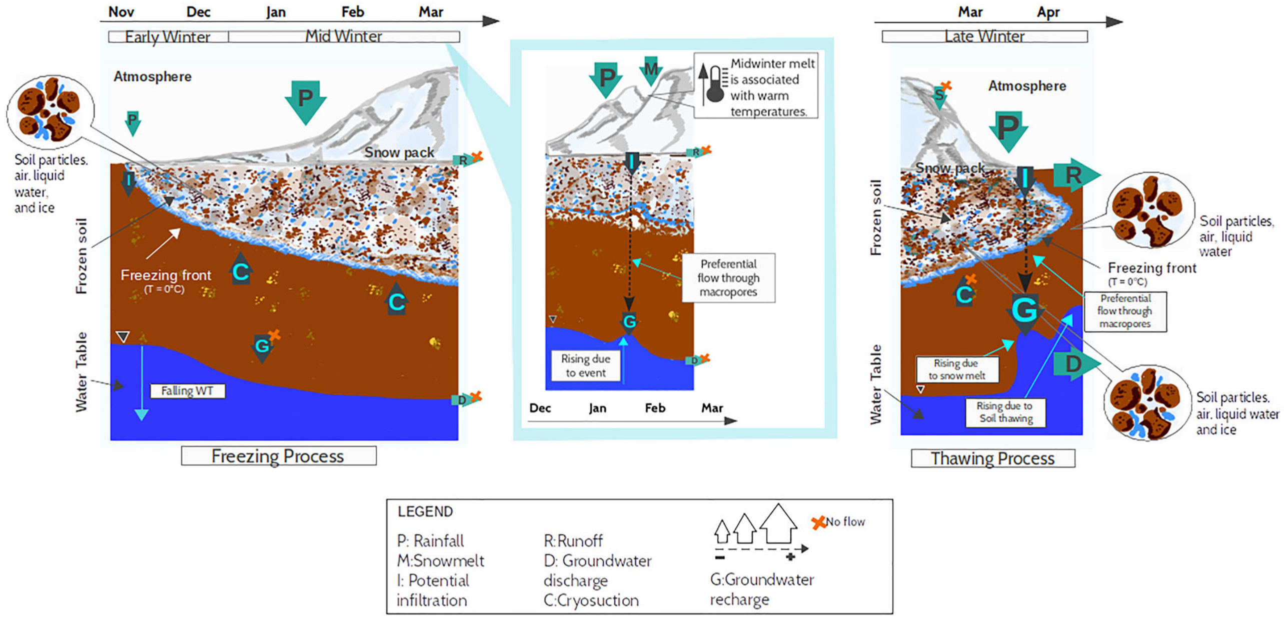

FIGURE 7 Diagram with a conceptual model for hydrological processes during soil freeze–thaw in seasonally frozen regions.

may be transmitted preferentially through the profile via air-filled macropores. Some portion of this infiltrated water will refreeze, and the remainder will percolate through the soil column into the deeper unsaturated zone or reach the water table. Percolated liquid water that reaches the water table generates the first pulse of recharge of groundwater in the early spring.

A second recharge pulse to the water table may occur in response to the complete thawing of the soil column. This happens after the snowpack has melted. Incoming energy to the system is used to change the ice to liquid water, firstly in the top layers of the soil column. This liquid soil moisture is now mobile, and gravity causes it to percolate down until it reaches the water table. The liquid water that did not percolate can refreeze during the night due to a diurnal freeze–thaw cycle. The total soil thawing is given in spring when the deeper frozen layers finally thaw, and the thawed liquid moisture percolates down by gravity until reaching the water table or evaporates.

5 | CONCLUSIONS

We have described the data collection and analysis methods used to study the physical controls of snowmelt water partitioning in seasonally frozen environments. Using information from meteorological stations and lowland/upland transects, we describe qualitative insights into the controls on soil freeze–thaw, snowmelt infiltration, and groundwater recharge. We propose a conceptual model for the hydrological processes that occur during soil freeze–thaw in seasonally frozen regions. Infiltration and percolation through air-filled macropores is a critical process, that explains how the water table and deep soil moisture can be recharged very rapidly during snowmelt, while the soils remain frozen. The partitioning of snowmelt into runoff versus infiltration, as described in our conceptual model, depends on the saturation of the soil and the temperature of the soil. The soil saturation depends on antecedent conditions in the fall, as well as the movement of moisture through the winter months due to infiltration of midwinter melt events, and the possible upward movement of water to the frozen soil from the shallow water table by cryosuction. The soil temperature depends on the seasonal air temperature, the soil saturation (wetter soils will be warmer as they store more latent heat), and the SWE (more SWE, especially early in the winter, will insulate the soils). Due to fact that these multiple controls are all important and are essentially independent, the runoff generation from SFS remains highly variable and hard to predict. Future insights are to be gained by the use of models to test hypotheses and corroborate field observations. Models must account for preferential macropore flow and for the multiple controls on soil saturation and temperature.

AUTHOR CONTRIBUTIONS

Ines Sanchez-Rodriguez: Conceptualization; data curation; formal analysis; investigation; methodology; software; validation; visualization; writing—original draft; writing—review and editing. **Andrew Ireson**: Conceptualization; formal analysis; funding acquisition; investigation; methodology; project administration; resources; supervision; validation; visualization; writing—original draft; writing—review and editing. **Rosa Brannen**: Data curation; formal analysis; visualization; writing—review and editing. **Haley Brauner**: Data curation; formal analysis; validation; writing—review and editing.

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ACKNOWLEDGMENTS

This project was supported by Ireson's NSERC Discovery Grant (RGPIN-2018-04274). Field instrumentation and technician support was provided by the Global Water Futures Program (GWF) and the Global Institute for Water Security (GIWS) at the University of Saskatchewan. We are grateful to Seth Amankwah and Bruce Johnson from GIWS for their support with field data collection. We are grateful to Chris Spence from Environment and Climate Change Canada (ECCC) for ongoing collaborations at St Denis.

CONFLICT OF INTEREST STATEMENT

The authors declare no conflicts of interest.

ORCID

Ines Sanchez-Rodriguez https://orcid.org/0000-0001-8065-313X
Andrew Ireson https://orcid.org/0000-0003-1957-7355

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How to cite this article: Sanchez-Rodriguez, I., Ireson, A., Brannen, R., & Brauner, H. (2025). Insights into freeze–thaw and infiltration in seasonally frozen soils from field observations. *Vadose Zone Journal*, *24*, e20396. https://doi.org/10.1002/vzj2.20396

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