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Reviews and Analyses

Core Ideas

- Review highlights the hydrological importance of macropore flow in frozen soils.
- Governing flow mechanisms and infiltration and refreezing dynamics are discussed.
- Research is needed to integrate macropore flow and soil freeze–thaw theory.
- Dual-domain models of macropore flow should be adapted to frozen ground.
- A conceptual framework for modeling frozen macroporous soils is proposed.

Snowmelt Infiltration and Macropore Flow in Frozen Soils: Overview, Knowledge Gaps, and a Conceptual Framework

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Macropore flow in frozen soils plays a critical role in partitioning snowmelt at the land surface and modulating snowmelt-driven hydrological processes. Previous descriptions of macropore flow processes in frozen soil do not explicitly represent the physics of water and heat transfer between macropores and the soil matrix, and there is a need to adapt recent conceptual and numerical models of unfrozen macropore flow to account for frozen ground. Macropores remain air filled under partially saturated conditions, allowing preferential flow and meltwater infiltration prior to ground thaw. Nonequilibrium gravity-driven flow can rapidly transport snowmelt to depths below the frost zone or, alternatively, infiltrated water may refreeze in macropores and restrict preferential flow. As with unfrozen soils, models of water movement in frozen soil that rely solely on diffuse flow concepts cannot adequately represent unsaturated macropore hydraulics. Dual-domain descriptions of unsaturated flow that explicitly define macropore hydraulic characteristics have been successful under unfrozen conditions but need refinement for frozen soils. In particular, because pore connectivity and hydraulic conductivity are influenced by ice content, modeling schemes specifying macropore–matrix interactions and refreezing of infiltrating water are critical. This review discusses the need for research on the interacting effects of macropore flow and soil freeze–thaw and the integration of these concepts into a framework of coupled heat and water transfer. As a result, it proposes a conceptual model of unsaturated flow in frozen macroporous soils that assumes two interacting domains (macropore and matrix) with distinct water and heat transfer regimes.

Abbreviations: SFC, soil freezing characteristic; SMC, soil moisture characteristic.

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Frozen soil plays a critical role in hydrological processes by controlling the partitioning of snowmelt flux between runoff and infiltration (Gray et al., 2001; Lundberg et al., 2016). Preferential pathways also play an important role in partitioning and routing of snowmelt (Stähli et al., 1996). Preferential flow in frozen soil is largely enabled by macropores, such as root holes and fractures, which facilitate the rapid movement of water and solutes that can bypass portions of the bulk soil matrix (Beven and Germann, 1982, 2013). Here, the term *macropore flow* is used to describe this general nonequilibrium behavior where vertical flow in larger soil pores is rapid relative to the rate of lateral equilibration of water pressures (or temperatures) in the surrounding soil matrix, generating substantial lateral differences or discontinuities across short (typically centimeter to decimeter) distances (Jarvis, 2007). Macropore flow can affect the spatial and temporal characteristics of snowmelt infiltration and related processes such as runoff generation, soil moisture distribution, and shallow groundwater recharge (Espeby, 1992; Baker and Spaans, 1997; Daniel and Staricka, 2000; van der Kamp et al., 2003).

Early studies showed that soil frost generally impedes water movement (e.g., Burt and Williams, 1976). However, subsequent studies have also shown that frozen soil can remain permeable and rapidly infiltrate snowmelt water via macropores (e.g., Granger et al., 1984; Stähli et al., 1996; Stadler et al., 2000). In recent reviews of unfrozen macropore flow literature (Beven and Germann, 2013; Jarvis et al., 2016, 2017), researchers concluded that, despite increased attention, there are still critical limitations in process understanding and modeling approaches of macropore flow, specifically regarding macropore hydrodynamic behavior and its deviation from traditional Darcy–Richards capillary-based assumptions. Jarvis et al. (2016) also specifically cited uncertainty regarding the effects of soil freeze–thaw as one reason why current approaches to modeling macropore flow lag behind process understanding. Traditional water retention concepts, such as the capillary-bundle model and the soil moisture characteristic, have been widely applied to frozen soils (e.g., Watanabe and Flury, 2008). However, models based on these concepts cannot capture the timing and magnitude of snowmelt infiltration (Stähli et al., 1996; Weigert and Schmidt, 2005), partly due to the lack of proper representation of macropore flow. Dual-domain descriptions combining diffuse flow in the soil matrix and rapid flow through macropores have successfully described preferential flow under unfrozen conditions, as they inherently assume that flow in macropores is subject to different physical controls and processes than flow in the soil matrix (Beven and Germann, 1981; Jarvis et al., 1991; Gerke and van Genuchten, 1993a; Nimmo, 2010). These conceptual models emphasize gravity-driven flow, macropore–matrix interactions, and characteristics of water supply at the ground surface as the dominant controls on preferential flow. However, the underlying process descriptions need further refinement for adaptation to frozen soil environments. In particular, because the pore space available for flow is influenced by ice content, accurate conceptualization of the processes controlling the refreezing of infiltrating water is critically important.

The aims of this review are to highlight the hydrologic importance of macropore flow during snowmelt infiltration in seasonally frozen soils and discuss a framework necessary to adapt dual-domain conceptual models of macropore flow to the study of frozen-ground infiltration and soil freeze–thaw dynamics.

Hydrological Processes in Frozen Soil

Snowmelt rate, or more generally water-input rate at the ground surface, constrains the infiltration response because it determines the rate of water and associated energy input into the soil system. The timing and rate of snowmelt are controlled by the energy balance near the snow surface, determined by radiation fluxes, turbulent fluxes, ground heat flux, advected energy from possible rain-on-snow events, and energy storage within snowpacks (Male and Granger, 1981). Melt begins when snowpack temperatures reach 0°C. Snowmelt rates range from around 3 mm d^{-1} in sheltered environments (e.g., forests) to up to 25 mm d^{-1} in more open settings (e.g., grasslands) (Ohta et al., 1993; Lundberg et al., 2016). Snowmelt usually occurs in a diurnal cycle, resulting in meltwater infiltration during daytime and refreezing within the snow and soil at night, which may cause ice-sealing of the ground surface and blockage of soil pores and may influence infiltration dynamics (Nyberg et al., 2001). Snow cover has a large insulation capacity and plays an important role in modulating frost penetration. Thick snow cover generally results in reduced soil freezing compared with thin snowpacks (Zhang, 2005; Iwata et al., 2010).

When unsaturated soils begin to freeze, the smaller pores are filled with water and the larger pores are usually air filled (Fig. 1a). As the freezing progresses, thin films of unfrozen water remain on the surface of soil particles, as capillary and adsorptive forces depress the freezing point and keep this bound water unfrozen. Figure 1b illustrates the three key phases in frozen soil: liquid water in the smallest pores, liquid water and ice in the intermediate pores, and air in the largest pores (Koopmans and Miller, 1966; Miller, 1980). The energy transfer processes in soil are strongly coupled with water transfer because water flows under matric and thermal potential gradients, while moving water transports energy and affects soil thermal properties (Hoekstra, 1966; Dirksen and Miller, 1966). Consequently, snowmelt infiltration is influenced by coupled heat and water transfer between the ground surface and the underlying soil (Zhao et al., 1997; Stähli et al., 1999). The primary mechanism controlling ground heat flux is thermal conduction, although flowing water can transfer substantial heat by advection during snowmelt infiltration (Kane et al., 2001). Heat and water fluxes are generally coupled through the vertical heat transport equation (e.g., Jansson and Karlberg, 2001):

$$\frac{\partial(C_s T)}{\partial t} - \rho_i H_f \frac{\partial \theta_i}{\partial t} = \frac{\partial}{\partial z}\left(\lambda \frac{\partial T}{\partial z}\right) - C_w \frac{\partial(q_w T)}{\partial z} \quad [1]$$

where C_s (J K^{-1} m^{-3}) is the bulk volumetric heat capacity of the soil, T (K) is temperature, t (s) is time, H_f (J kg^{-1}) is the latent heat of fusion, ρ_i is the density of ice (kg m^{-3}), θ_i (dimensionless) is volumetric ice content, λ (W m^{-1} K^{-1}) is the bulk thermal conductivity of the soil system (soil, air, water), C_w (J K^{-1} m^{-3}) is the volumetric heat capacity of water, z (m) is elevation, and q_w (m s^{-1}) is the vertical flux of liquid water. The first term on the left represents the rate of change in sensible heat storage, while the second term on the left represents latent energy released (absorbed) during freezing (melting) of pore water. The terms on the right represent the divergence of conductive and advective heat fluxes, respectively. Equation [1] highlights the tight coupling of the soil heat and water fluxes because advective heat flux is dependent on water flow; λ and C_s are influenced by soil liquid water, ice, and air content; and the pore water

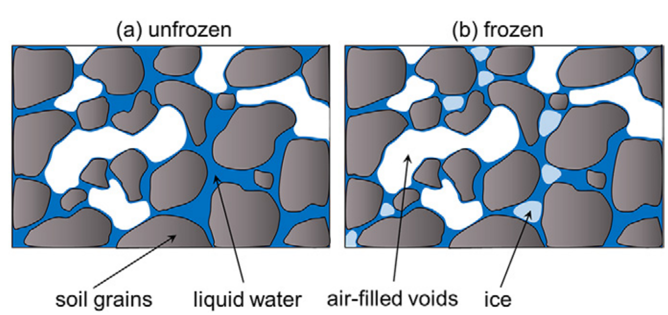

Fig. 1. Conceptual model of the constitutive fluid phases present in partially saturated (a) unfrozen and (b) frozen soils.

available for phase change is dependent on the water distribution. The thermodynamic equilibrium between pressure and temperature is commonly expressed by the Clausius–Clapeyron equation, assuming that ice pressure is constant (Williams and Smith, 1989):

$$\frac{dP_w}{dT}=\frac{H_f}{TV_w} \qquad [2]$$

where P_w (Pa) is the liquid water pressure and V_w (m^3 kg^{-1}) is the specific volume of liquid water. Equation [2] shows that freezing decreases the liquid water content and decreases soil matric potential (termed *cryo-suction*), resulting in steep potential gradients that redistribute water from unfrozen soil to the freezing front (e.g., Gray and Granger, 1986; Iwata et al., 2010). If the antecedent moisture content is high enough, this freezing-induced moisture redistribution may promote pore-ice formation, which reduces the soil infiltrability (Kane and Stein, 1983) as the freezing of water causes a volume expansion of about 9% (Haynes, 2014). Larger water-filled pores will freeze first, as the water is held under less capillary force. Once these pores are blocked, hydraulic conductivity decreases, causing a reduction in infiltrability (Granger et al., 1984; Lundberg et al., 2016). Unless the soil is near saturation when it freezes, macropores retain very little water and generally remain air filled and open for infiltration after soil freezing, thus providing preferential flow paths in the frozen soil profile (Espeby, 1990; Watanabe and Kugisaki, 2017).

Field Evidence of Macropore Flow in Frozen Soil

Field studies in cold regions clearly illustrate the hydrological importance of macropore flow in frozen soil (summarized in Table 1). Higher antecedent soil moisture on soil freezing often results in the development of an ice-rich zone near the ground surface that impedes snowmelt infiltration, but large open pores can still rapidly infiltrate meltwater (Stoeckeler and Weitzman, 1960; Kane, 1980; Kane and Stein, 1983; Granger et al., 1984). These rapid infiltration and subsurface responses to snowmelt events were explained by rapid drainage of "free water" (i.e., not bound by adsorption and capillarity) via an interconnected macropore network in the soil profile (e.g., Greminger, 1984; Espeby, 1992). It was also hypothesized that this flow was restricted to macropores, as freezing temperatures and pore ice would severely limit matrix flow (Komarov and Makarova, 1973; Steenhuis et al., 1977; Lundin, 1989). These studies stressed the importance of the three-phase composition (water, ice, air) within the soil and its influence on infiltrability. Granger et al. (1984) proposed a conceptual model that divided the infiltrability of frozen prairie soils into three categories: (i) unlimited—gravity flow dominates and most snowmelt infiltrates through macropores; (ii) limited—capillary flow dominates and infiltration is influenced primarily by soil texture and soil frost conditions; and (iii) restricted—soil infiltrability is restricted by soil surface conditions. This implied that macropores can remain air filled on freezing and thereby allow water to bypass large portions of the soil matrix, enabling the rapid infiltration of water while significant soil frost is still present. Subsequent studies also highlighted the effects of soil temperature and refreezing of infiltrated water on subsequent infiltration dynamics (Jansson and Gustafsson, 1987; Thunholm et al., 1989).

The concept of water flow in air-filled pores was formalized from snowmelt infiltration observations of Johnsson and Lundin (1991) and quantified by Stähli et al. (1996). Both studies observed infiltration through frozen soil that was too rapid to be explained by assuming solely capillary-driven liquid water movement through the matrix. Rather, the results were indicative of gravity-driven flow through larger pores that were air filled at the beginning of snowmelt. Stähli et al. (1996, 1999) made no clear distinction between macropores and air-filled pore space in the matrix but did make a distinction between "low-flow" and "high-flow" mechanisms and highlighted that the refreezing of infiltrating meltwater was dependent on the thermal conditions of the bulk frozen soil. These findings were transcribed into a numerical model described below.

Later studies in the Canadian Prairies have shown that perennial grasslands can have a much higher frozen soil infiltrability due to the presence and development of a macropore network compared with croplands, where annual cultivation breaks the macropore network (van der Kamp et al., 2003; Bodhinayake and Si, 2004). In northern prairie landscapes, meltwater collected in topographic depressions following snowmelt can supply large volumes of water for infiltration. Under these conditions, macropores can cause rapid infiltration of snowmelt and groundwater recharge through partially frozen ground and can facilitate preferential contaminant transport to aquifers (Baker and Spaans, 1997; Daniel and Staricka, 2000; Derby and Knighton, 2001).

Macropore flow is also important in permafrost regions for rapidly conveying snowmelt deeper within the active layer before complete thaw and for modulating runoff (Mackay, 1983; Boike et al., 1998; Scherler et al., 2010). When snowmelt infiltration is restricted, meltwater is rapidly conveyed to the watershed outlet and typically produces a flashy hydrologic response at the catchment scale (Roulet and Woo, 1986). However, unsaturated macropores allow infiltration of meltwater through the frozen active layer, which may reduce high stream flows associated with snowmelt events and increase soil moisture storage (Mackay, 1983; Boike et al., 1998; Scherler et al., 2010). Importantly for permafrost environments, which are undergoing rapid warming due to climate change, macropore flows provide conduits for advective heat flux, which can contribute to thawing of the active layer and shallow permafrost (Roth and Boike, 2001; Ishikawa et al., 2006; Koch et al., 2013).

Collectively, these studies identified that the critical subsurface factors influencing snowmelt infiltration dynamics are (i)

Table 1. Summary of frozen soil studies with key insights on macropore flow in frozen soil.

Study	Location and landscape†	Observations	Hydrological implications
Stoeckeler and Weitzman (1960)	Minnesota, USA; agricultural field (SF)	frozen ground infiltration	hypothesized that infiltration into frozen soil was restricted to macropores
Mackay (1983)	Alaska, Canada, Russia, China (PF)	meltwater infiltration into frozen active layer	snowmelt infiltration can migrate through frozen soil to permafrost
Granger et al. (1984)	Saskatchewan, Canada; agricultural field (SF)	all snow cover infiltrates into cracked frozen soils	development of frozen soil infiltration conceptual model and infiltration index empirical model
Greminger (1984)	Switzerland; coniferous forest (SF)	almost instantaneous infiltration of snowmelt	infiltration response caused by an interconnected macropore system
Gray et al. (1985)	Saskatchewan, Canada; agricultural field (SF)	frozen soil infiltration index model improves simulation of streamflow from snowmelt	application of frozen soil infiltration index empirical model into operational streamflow model
Jansson and Gustafsson (1987)	Sweden; agricultural field (SF)	infiltration below frost zone due to macropores	macropores significantly enhance frozen soil hydraulic conductivity
Thunholm et al. (1989)	Sweden; agricultural field (SF)	freezing of infiltrated meltwater in macropores linked to soil temperature	midwinter snowmelt infiltration refroze and reduced infiltrability in spring
Espeby (1992)	Sweden; coniferous forest, glacial till hillslope (SF)	rapid infiltration due to interconnected macropore network	macropores allow bypass flow through frozen soil
Johnsson and Lundin (1991)	Sweden; agricultural field (SF)	infiltration prior to ground thaw; infiltration below frost zone	dual-domain frozen soil hydraulic conductivity concept proposed
Stähli et al. (1996)	Sweden; agricultural field (SF)	dual-domain frozen soil model improved snowmelt infiltration simulations	dual-domain frozen soil hydraulic conductivity concept implemented into SOIL model
Baker and Spaans (1997)	Minnesota, USA; agricultural field (SF)	rapid frozen ground infiltration; infiltration below frost zone	infiltration through large pores allow meltwater to bypass frozen soil
Boike et al. (1998)	Siberia; tundra (PF)	infiltration into frozen ground	observed preferential penetration of dye-tracer halfway through frost zone
Daniel and Staricka (2000)	Minnesota, USA; agricultural field (SF)	infiltration below frost zone; groundwater recharge through frozen ground	macropores allow channelling of snowmelt to groundwater
Derby and Knighton (2001)	North Dakota, USA; agricultural field (SF)	infiltration below frost zone; groundwater recharge through frozen ground	contaminant transport during these events relevant to protection of aquifers
van der Kamp et al. (2003)	Saskatchewan, Canada; grassland/agricultural (SF)	rapid frozen ground infiltration	soils with well-developed macropore network have higher frozen soil infiltrability
Stähli et al. (2004)	Switzerland; alpine forest (SF)	infiltration below the frost zone	irregular dye-stained pattern indicated that water bypassed some of the frost zone
Ishikawa et al. (2006)	Mongolia; grassland (PF)	rapid frozen ground infiltration and soil moisture responses	macropores promote active layer warming via “pipe-like” snowmelt infiltration to deeper soil layers
Scherler et al. (2010)	Switzerland; alpine scree slope (PF)	rapid frozen ground infiltration; refreezing of infiltrated meltwater	preferential flow promotes infiltration into frozen active layer
Koch et al. (2013)	Alaska, USA; boreal forest (PF)	meltwater reaching permafrost table via macropores	meltwater reaching permafrost increases watershed’s carbon export
Stadler et al. (2000)	laboratory, soil column	water flow through initially air-filled pores	infiltration almost exclusively through macropores
Weigert and Schmidt (2005)	laboratory, soil column	high frozen soil hydraulic conductivity	macropore flow dominated the hydraulic regime of the sample
Watanabe and Kugisaki (2017)	laboratory, soil column	water flow through initially air-filled pores; refreezing of flowing water	refreezing of flowing water blocked macropores; ice formed from macropore surface

† SF, seasonally frozen ground; PF, permafrost.

antecedent soil moisture, (ii) freezing-induced moisture redistribution, (iii) increased infiltrability due to larger air-filled pores, and (iv) refreezing of meltwater reducing infiltrability. Just as importantly, the field evidence shows that when a macropore network is present, macropore flow dominates over matrix flow in frozen soils to an even greater extent than in unfrozen soils, since freezing temperatures and pore ice greatly reduce the hydraulic conductivity of the soil matrix.

Laboratory Studies

Compared with field studies affected by numerous uncontrolled variables, laboratory data can isolate specific processes and provide new insights. Although limited in number compared with field investigations, focused laboratory experiments on frozen macroporous soils have advanced the understanding of rapid frozen soil infiltration and the role of air-filled macropores, as shown in Table 1. Stadler et al. (2000) performed infiltration experiments on a frozen undisturbed soil column and observed a dye tracer moving through macropores that remained air filled during freezing. Weigert and Schmidt (2005) repacked soil columns of sandy and loamy soil and froze them to a temperature of −4°C before allowing water to infiltrate. The loamy soil developed macropores due to desiccation during freezing, and as a result antecedent moisture had little influence on infiltration. The measured hydraulic conductivity was nearly independent of initial soil moisture, implying that the soil matric potential (i.e., capillary forces) had little effect on the hydrodynamic behavior of frozen macropore flow.

Watanabe and Kugisaki (2017) performed a soil-column infiltration experiment on packed frozen cores with artificial cylindrical macropores, providing the first direct observational evidence of how water freezes in macropores. The major observation was that water infiltrating along macropores was cooled sufficiently by the surrounding frozen soil matrix to freeze within the macropores, thereby blocking further water migration. These results clearly demonstrated that macropores remain open during soil freezing but can be blocked by the freezing of infiltrated water, even when flow rates are relatively high. Watanabe and Kugisaki (2017) also highlighted an important problem regarding macropore flow in frozen soil: the question of how water freezes within a macropore. Their results support the hypothesis that flowing water in a macropore freezes first along macropore walls where dents and microcavities can trap water in a reduced energy state. This is in contrast to smaller saturated matrix pores, in which ice formation first occurs in the center of pores (Fig. 1), as adsorptive forces suppress the freezing temperature at the soil–water interface.

A key development in improving understanding of macropore flow has been the emergence of suitable methods for visualizing and quantifying aspects of flow dynamics. Previous laboratory experiments on macropore flow under unfrozen conditions have yielded key insights through the use of dye and chemical tracers (e.g., Wildenschild et al., 1994) and novel experimental techniques to simulate macropore flow under unsaturated conditions (e.g., Tokunaga and Wan, 1997). Adaption of these methodologies to frozen soils will require careful consideration of the influence of soil ice and temperature effects on measurement techniques and observations. The use of geophysics and imaging technologies also offers promising tools for advanced understanding of frozen soil processes. For example, Koestel et al. (2009) demonstrated that Brilliant Blue dye could be detected nondestructively using electrical resistivity tomography (ERT) of soil columns. Electrical resistivity tomography has already been applied to imaging the evolution of frozen ground during snowmelt infiltration (French and Binley, 2004), but the image resolution in the field is too coarse to discern pore-scale observations. Other nondestructive imaging techniques, such as X-ray tomography and neutron radiography, have been used to visualize macropore structures, pore-fluid configuration, and infiltration patterns in undisturbed soil columns (e.g., Badorreck et al., 2012; Sammartino et al., 2012, 2015). Compared with ERT, which has a spatial resolution of millimeters to centimeters, X-ray tomography and neutron radiography can resolve macropore structures down to the micrometer range (Koestel and Larsbo, 2014). Thus, the combination of geophysical imaging techniques and chemical tracers at the core scale in a controlled laboratory environment offers a promising avenue for potential experimental research. More research on how to obtain the necessary data at the scales of interest is required, and controlled laboratory studies are crucial for such advancement.

Modeling Approaches

Modeling Flow in Unfrozen Macropores

The consensus that preferential flow in macroporous soils cannot be modeled solely with a single-continuum capillary approach has led to the development of a number of models with varying capabilities and underlying concepts to simulate combined macropore–matrix flow. Most models include distinct flow systems for the macropore network and the soil matrix, either as discrete fractures or macropores embedded within the soil matrix or as separate flow domains (dual continuum) (Šimůnek et al., 2003). Most importantly, appropriate representations of fluid flow dynamics in macropores and macropore–matrix mass transfer are crucial to simulating flow in both unfrozen and frozen soil (Beven and Germann, 2013; Jarvis et al., 2017).

To capture the dual nature of flow, two flow equations are typically coupled: one for the highly permeable macropore domain and one for the less permeable matrix. Matrix flow is formulated with Darcy–Buckingham and Richards equations, using traditional hydraulic conductivity and soil moisture characteristic relations to represent hydraulic properties. There is no current consensus on the appropriate descriptions for flow in the macropore domain (Jarvis et al., 2016). Using a similar capillary-bundle approach to simulate macropore flow, the Richards equation has been adopted for the macropore domain with some success (Gerke and van Genuchten, 1993a; Alberti and Cey, 2011). However, others have argued that macropore flow is mainly gravity driven, with negligible influence of soil capillarity on flow rates (Tokunaga et al., 2000; Germann, 2001; Nimmo, 2010). Supporting experimental evidence reveals that macropores frequently transport water as thin, free-surface films or rivulets along the macropore surface under unsaturated conditions (Tokunaga and Wan, 1997; Su et al., 1999; Dragila and Wheatcraft, 2001; Nimmo, 2003; Cey and Rudolph, 2009). In this situation, fluid viscosity controls the velocity profile in the water film flowing between the solid–water and air–water interfaces as illustrated in Fig. 2 (Tokunaga and Wan, 1997; Nimmo, 2010). This flow behavior is conceptually well captured by a kinematic wave framework, which has long been used to model preferential flow in soils (e.g., Germann, 1985; Chen and Wagenet, 1992; Larsson and Jarvis, 1999; Larsbo et al., 2005) and has been shown to be a natural generalization of conceptual pore-scale models of film and rivulet flow (Dragila and Wheatcraft, 2001; Germann, 2001; Jarvis et al., 2017). It should be noted that macropores do not always conduct water as free-surface films or

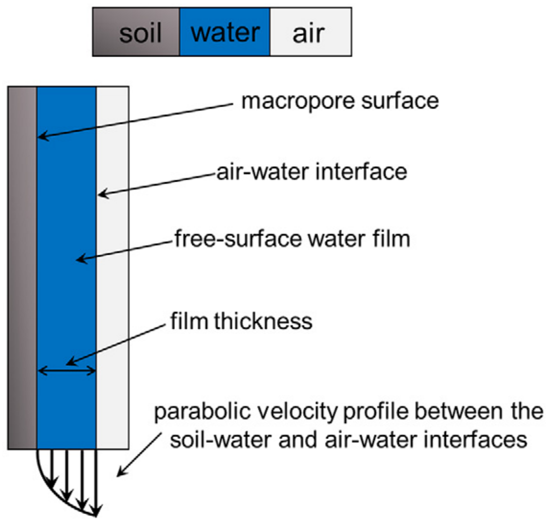

Fig. 2. Conceptual model of free-surface film flow in unfrozen macropores.

rivulets but can also support fully saturated flow under certain conditions (e.g., ponded infiltration) (Jarvis, 2007; Sammartino et al., 2012, 2015). Jarvis et al. (2017) showed how the various modes of flow can be captured with the kinematic wave equation.

Conceptualizing macropore flow as gravity-driven film flow or kinematic waves has proven to be a more physically realistic description than capillary-bundle models because these hydrodynamic models capture the key characteristics realized from field and laboratory studies, namely rapid vertical flow through connected macropores under partially saturated conditions in response to an infiltration source (Cey and Rudolph, 2009; Nimmo, 2010; Jarvis et al., 2016). Additionally, because flow in macropores tends to respond quickly to changes in the water input rate, any macropore flow model needs to be able to respond sensitively to changes in water input at the soil surface (Germann and Di Pietro, 1999; Dragila and Wheatcraft, 2001; Nimmo, 2010). This "source-responsive" characterization, i.e., flow is only active when a water source is present, is implicitly captured by film flow and kinematic wave models, which combine high flow capacity with limited fluid storage. Models based on these principles have significantly improved the simulation of preferential flow events observed in unsaturated flow and transport studies (Germann and Di Pietro, 1999; Larsson and Jarvis, 1999; Dragila and Wheatcraft, 2001; Nimmo, 2010; Cuthbert et al., 2013). Observations of rapid, source-responsive flow in frozen soils with air-filled macropores suggest that a similar modeling approach holds promise for capturing preferential flow in frozen ground.

Equally critical in all dual-domain (dual-porosity or -permeability) models are the coupling terms describing mass and energy transfer between the macropore and matrix domains (Šimůnek et al., 2003). Water transfer is often described based on water content or pressure head differences between domains (Šimůnek et al., 2003). Geometrically, transfer occurs along the specific contact area between the matrix and macropore domains (Gerke and van Genuchten, 1993b; Jarvis 1994; Hincapié and Germann, 2009). The definition is crucial because the specific contact area is the interface across which water and solutes are exchanged between the macropore and matrix domains (Köhne et al., 2009). Importantly for frozen soil, the contact area will dictate heat transfer between domains, which will have implications for soil freeze–thaw and related flow dynamics in both the macropore and matrix. During infiltration, macropore hydraulic properties and the specific contact area combine to control infiltration, mass exchange with the matrix, and ultimately the flow and transport (solute and thermal) processes within the soil profile.

Frozen Soil Infiltration and Soil Water Modeling

A variety of models have been designed to simulate frozen ground effects on hydrological processes. The simplest approaches are water balance models, where model structure is based on different conceptualizations and empirical equations governing water flow and storage between different soil layers (e.g., Fox, 1992; Luo et al., 2000). Soil freezing is usually estimated empirically using a temperature index or a heat conduction algorithm, with soil frost assumed to form below a specified temperature. Soil infiltrability is then decreased accordingly due to soil freezing (e.g., Schroeder et al., 1994; Mohammed et al., 2013), although some use additional empirical bypass flow routines to account for high frozen soil infiltrability due to macropores (e.g., Chung et al., 1992). Distributed hydrological models use similar empirical frozen soil routines to modulate infiltration in snowmelt–runoff relations, to better represent hydrograph responses to snowmelt (Gray et al., 1985; Prévost et al., 1990; Pomeroy et al., 2007). The focus here, however, is on physically based models specifically conceptualized to simulate water flow and freeze–thaw in the subsurface.

Table 2 lists several commonly used process-based models of varying levels of complexity for simulating variably saturated flow in frozen soil. Most physically based numerical models of water flow in frozen soil are formulated by coupling a modified Richards equation for water flow to a heat transfer equation, similar to Eq. [1], that includes latent energy exchange associated with soil freeze–thaw. The one-dimensional form of the Richards equation is expressed as (Li et al., 2010)

$$\frac{\partial \theta_l}{\partial t}+\frac{\rho_i}{\rho_l}\frac{\partial \theta_i}{\partial t}=\frac{\partial}{\partial z}\left[K_F\left(\frac{\partial \psi}{\partial z}+1\right)\right] \qquad [3]$$

where θ_l is the liquid water content (dimensionless), K_F is the frozen soil hydraulic conductivity (m s^{-1}), ψ is matric potential head (m), and ρ_l is density of liquid water (kg m^{-3}). Koopmans and Miller (1966) demonstrated the similarity of the soil freezing characteristic (SFC) to the soil moisture characteristic (SMC) and proposed that this could be used in soil moisture retention models for predicting the relative hydraulic conductivity of saturated frozen soils. Further work (Jame and Norum, 1980; Miller, 1980) showed that the unfrozen water content is largely independent of the total water (ice + liquid) content under unsaturated

Table 2. Physically based hydrological numerical models capable of simulating soil freeze/thaw and variably saturated flow.

Model	Model type	Water flux	Thermal processes	Soil freezing†
Guymon and Luthin (1974)	1D continuum	Richards	conduction, advection, latent heat	SFC-SMC linked via Clapeyron equation
Jame and Norum (1980)	1D continuum	Richards	conduction, advection, latent heat	SFC to define liquid WC and HC
SHAW, Flerchinger and Saxton (1989)	1D continuum	Richards	conduction, advection, latent heat	SFC-SMC linked via Clapeyron equation
HydrogeoSphere, Therrien et al. (2010)	3D continuum	Richards	conduction	Temperature index/heat conduction algorithm
Zhao et al. (1997)	1D continuum	Richards	conduction, advection, latent heat	SFC-SMC linked via Clapeyron equation
VIC, Cherkauer and Lettenmaier (1999)	distributed hydrologic response unit	Richards	conduction, latent heat	SFC-SMC linked via Clapeyron equation
DRAINMOD, Luo et al. (2000)	1D water balance	Richards	conduction, latent heat	SFC-SMC linked via Clapeyron equation
COUP (SOIL), Jansson and Karlberg (2001)	1D continuum	Richards/dual-domain frozen soil hydraulic conductivity	conduction, advection, latent heat	SFC-SMC linked via Clapeyron equation
Ippisch (2001)	3D continuum	Richards	conduction, advection, latent heat	SFC to define liquid WC and HC
HYDRUS-1D, Hansson et al. (2004)	1D continuum	Richards	conduction, advection, latent heat	SFC-SMC linked via Clapeyron equation
White and Oostrom (2006)	3D continuum	Richards	conduction, advection, latent heat	SFC to define liquid WC and HC
SUTRA-ICE, McKenzie et al. (2007)	3D continuum	Richards	conduction, advection, latent heat	SFC to define liquid WC and HC
Dall'Amico et al. (2011)	3D continuum	Richards	conduction, advection, latent heat	SFC to define liquid WC and HC
Tan et al. (2011)	3D continuum	Richards	conduction, advection, latent heat	SFC to define liquid WC and HC
Liu and Yu (2011)	3D continuum	Richards	conduction, advection, latent heat	SFC to define liquid WC and HC
ATS, Painter et al. (2016)	3D continuum	Richards	conduction, advection, latent heat	SFC-SMC linked via Clapeyron equation

† SFC, soil freezing characteristic; SMC, soil moisture characteristic; WC, water content; HC, hydraulic conductivity.

freezing conditions. This implied that the unfrozen water content is controlled by temperature even under unsaturated conditions and that SMC–SFC relationships could be extended to determine the hydraulic conductivity of unsaturated frozen soils. Figure 3 depicts this relation, which assumes that soil freezing occurs in an analogous fashion to soil drying. The total water content is obtained based on the pre-freezing pressure head from the SMC curve. At temperatures below freezing, the unfrozen water content can then be determined from the SFC curve, which can then be used to calculate frozen soil hydraulic conductivity using existing SMC–conductivity relations for unfrozen soils (e.g., Mualem, 1976). This approach allows the application of hydraulic conductivity models that have already been tested and parameterized for a number of soil types.

Following this approach, one-dimensional models have been developed that allow for capillary-driven water fluxes in frozen soils using a hydraulic conductivity function related to the SFC and SMC (e.g., van Genuchten 1980) and various frozen ground infiltration algorithms (e.g., Flerchinger and Saxton, 1989; Zhao et al., 1997; Zhang et al., 2010). Several researchers (e.g., Newman and Wilson, 1997; Painter, 2011; Azmatch et al., 2012) have demonstrated that SMC–conductivity equations could be applied to frozen soils without modifications to account for the flow resistance due to the presence of ice in pore spaces. In this case, either the matric potential during freezing could be calculated using the Clausius–Clapeyron equation (Eq. [2]) or the liquid water content during freezing could be obtained from the SFC (Fig. 3). However, the volumetric expansion of ice and the fact that the pore space available for flow has a different geometry in frozen (soil and ice) vs. unfrozen soil (soil) creates problems with applying SMC–SFC-derived hydraulic conductivities. As a result, others (e.g., Jame and Norum, 1980; Flerchinger and Saxton, 1989; Lundin, 1990; Hansson et al., 2004) have used an additional empirical impedance factor to account for the additional hydraulic resistance of ice in partially frozen soil compared with the analogous presence of air in unfrozen, drying soil. The impedance concept is expressed as (Lundin, 1990)

$$K_F = 10^{-EQ} K_U \qquad [4]$$

where K_U is the hydraulic conductivity of the unfrozen soil at the equivalent liquid water content and matric potential (m s^{-1}), and 10^{-EQ} is the empirical impedance factor (dimensionless), where Q is the mass ratio of ice to total water (dimensionless) and E is an empirical constant that accounts for the reduction in hydraulic conductivity due to the presence of ice. More recent studies have favored bimodal or multimodal porosity–hydraulic conductivity relationships, as opposed to an impedance factor, to address the complex character of hydraulic conductivity resulting from multimodal pore-size distributions (Watanabe et al., 2010; Kurylyk and Watanabe, 2013). However, these equilibrium approaches for linking the SFC and SMC are not well suited for macroporous frozen soils, as one might expect since Beven and Germann (1982) explicitly identified the shortcomings of using SMC relations (e.g., Brooks and Corey, 1964; van Genuchten, 1980) to predict macropore flow under unfrozen conditions. As a result, these models have difficulty reproducing the subsurface response to snowmelt, with simulated infiltration and drainage lagging field measurements (e.g., Johnsson and Lundin, 1991).

Conceptual models for preferential flow through frozen soils have proposed that water flow occurs though macropores and bypasses a portion of the frozen soil profile (Komarov and Makarova, 1973; Johnsson and Lundin, 1991; Espeby, 1992). Utilizing this framework, a few mathematical models of snowmelt infiltration into frozen soil have been developed to account for preferential flow. Espeby (1992) modified the SOIL model (Jansson and Halldin, 1980) to incorporate an empirical bypass function accounting for macropore flow and was able to better simulate the rapid response in groundwater levels and runoff from snowmelt events. Stähli et al. (1996) expanded on this and developed a physically based description of infiltration into frozen soil that accounted for high infiltrability due to the presence of air-filled pores. Stähli et al. (1996) integrated a dual-domain flow concept (Fig. 4) into the SOIL model as a composite water content–hydraulic conductivity (θ–K) function. They divided the θ–K relation into two separate functions, representing the high- and low-flow domains, and specified mass transfer between domains. In the high-flow domain, water was assumed to flow in the previously air-filled pores, a unit gravitational gradient was assumed, and the hydraulic conductivity was defined as (Stähli et al., 1996)

$$K_{\mathrm{HF}} = K_{\mathrm{U}}\left(\theta_{\mathrm{HF}} + \theta_{\mathrm{i}} + \theta_{\mathrm{LF}}\right) - K_{\mathrm{U}}\left(\theta_{\mathrm{i}} + \theta_{\mathrm{LF}}\right) \quad [5]$$

where θ_{HF} is the liquid water content in the high-flow domain, θ_{i} is the ice content, θ_{LF} is the liquid water content in the low-flow domain, $K_{\mathrm{U}}(\theta_{\mathrm{HF}} + \theta_{\mathrm{i}} + \theta_{\mathrm{LF}})$ is the hydraulic conductivity of the total pore volume occupied by ice or water, and $K_{\mathrm{U}}(\theta_{\mathrm{i}} + \theta_{\mathrm{LF}})$ is the hydraulic conductivity of the pore volume occupied by ice and water in the low-flow domain. The high-flow domain relied on the SMC to define the unsaturated hydraulic conductivity (Mualem, 1976):

$$K_{\mathrm{U}}\left(\theta_{\mathrm{HF}} + \theta_{\mathrm{i}} + \theta_{\mathrm{LF}}\right) = K_{\mathrm{s}} S_e^{n+2+2/b} \quad [6]$$

where K_{s} is the saturated hydraulic conductivity, S_{e} is the effective water saturation (liquid + ice), n is a tortuosity factor, and b is the pore-size distribution index (Brooks and Corey, 1964). The freezing of water infiltrated in the high-flow domain released

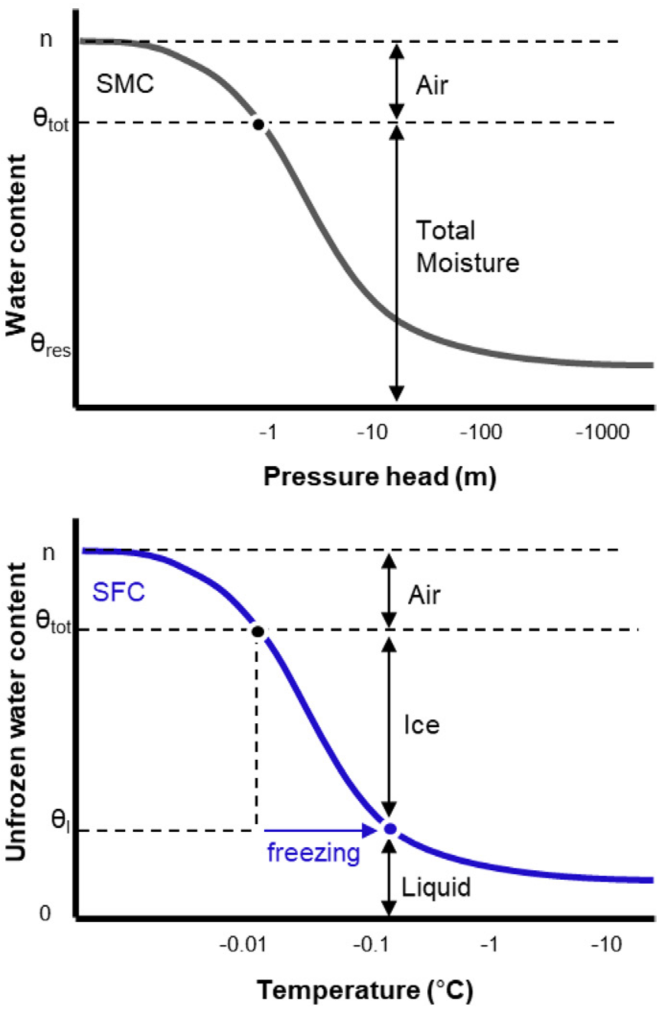

Fig. 3. Conceptual illustration of the soil moisture characteristic (SMC) (above) and soil freezing characteristic (SFC) (below). The SMC partitions between air and total moisture, and the SFC then partitions total moisture between ice and liquid at freezing temperatures; n = porosity; θ_{tot} = total water content; θ_{l} = liquid water content; θ_{res} = residual water content (adapted from Kurylyk and Watanabe, 2013).

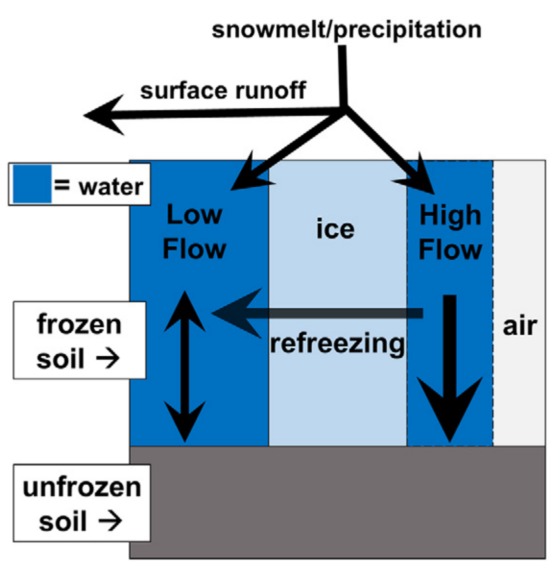

Fig. 4. Dual-domain hydraulic conductivity concept (adapted from Stähli et al., 1996).

energy to the low-flow domain and caused melting in the smallest pores and freezing of water in the larger pores (Fig. 4). Stähli et al. (1996) treated this as a water transfer from the high-flow to the low-flow domain, effectively shifting the relative size of the two model domains with time and temperature by redistributing liquid water from the high-flow domain to the low-flow domain without changing the ice content during the lateral redistribution. Thus, this description does not represent two physical domains in the sense of matrix and macropores but rather two flow regimes where water tightly bound to the soil surface moves under matric potential gradients in the low-flow domain and "free water" not subject to adsorption and capillary forces flows in the high-flow domain where gravity-driven flow occurs. Simulations using this modeling framework were able to better capture soil moisture dynamics at the onset of snowmelt infiltration and the refreezing of infiltrating water. Application of the model indicated that relatively rapid water flow occurs through the largest pores, which are air filled at the beginning of snowmelt, along with a much slower flow regime in the unfrozen smaller pores of the bulk matrix (Stadler et al., 1997; Stähli et al., 1999). The SOIL model (Jansson, 1998), which has been since incorporated into the larger COUP model (Jansson and Karlberg, 2001), was the first to simulate water fluxes in frozen soils by assuming two water-conducting domains (Stähli et al., 1996). The model has performed well at simulating the onset of infiltration in frozen soils from both field and experimental data (Stähli et al., 1996; Stähli and Stadler, 1997;) but still underestimated the magnitude and depth of infiltration (Stadler et al., 1997; Stähli et al., 1999). Stähli et al. (1999) concluded that a better physical description of the "high-flow mechanism" would improve model capabilities. More recently, Weigert and Schmidt (2005) modified the concept of Stähli et al. (1996) by using the saturated hydraulic conductivity of the high-flow domain to calculate the hydraulic conductivity of the high-flow domain. Applying this model to their previously described experiments, they needed to increase their theoretical frozen hydraulic conductivity of the high-flow domain by a factor of 15 to match measured values of 9×10^{-4} m s^{-1} and concluded that macropore flow dominated the hydraulic regime of the frozen sample. These results clearly demonstrate that fundamental assumptions of the water retention and transmission characteristics are incomplete for frozen macroporous soils, and an improved hydrodynamic description is still required.

Synthesis and Proposed Modeling Framework

Processes Unique to Water Flow in Unsaturated Frozen Soil

Based on previous cold regions studies, the main subsurface factors governing snowmelt infiltration are: (i) infiltration of snowmelt water into unfrozen and air-filled matrix pores, where capillary flow (driven by matric potential gradients) dominates; (ii) infiltration into air-filled macropores that remain open during soil freezing, where gravity-driven flow dominates; (iii) freezing-induced moisture migration and blockage of some initially air-filled matrix pores; and (iv) freezing of infiltrated water and blockage of pores (matrix and macropores) until ground thaw.

In most frozen soil models, liquid water content is calculated from temperature using Eq. [2] by assuming similarity between the SMC and SFC (e.g., Flerchinger and Saxton, 1989; Hansson et al., 2004; Watanabe and Flury, 2008). This assumes that freezing of unsaturated soil occurs in a fashion similar to saturated soil, where ice first begins to form in the largest pores, and implies that the liquid–ice interface is geometrically equal to the liquid–air interface. Thus when frozen, pore ice is assumed to be located in the center of the conductive large pores regardless of how it was formed at freezing, i.e., soil freezing–thawing occurs similar to soil drying–wetting, where the largest pores freeze (drain) first and thaw (saturate) last. Under this assumption, hydraulic conductivity decreases as ice begins to form in the largest pores, and the capacity for flow in most frozen soils is considered very low or negligible. However, if the soil is unsaturated when freezing occurs, the largest pores are air filled and not occupied by ice (Fig. 1). Capillarity plays a role in some of the air-filled pore space, where freezing-induced moisture redistribution can cause blockage of some of the initially air-filled space, but macropores are generally unaffected and remain open during soil freezing, only closing due to freezing of infiltrating water (Watanabe and Kugisaki, 2017). Thus, there is an important difference in the mechanisms governing water movement and freezing in macropores and matrix pores. A distinction should be made based on a pore-size threshold that marks the transition between these two flow and freezing regimes.

The concept of two-domain water flow in frozen soil developed by Stähli et al. (1996) implicitly incorporates increased infiltrability due to air-filled pores and also considers the refreezing of infiltrating water. However, the description does not explicitly represent the physics of macropore flow and, as such, does not incorporate the influence of macropore flow under all (frozen and unfrozen) conditions. It seems reasonable, when investigating macropore flow in frozen soils, to introduce newly developed understandings of partially saturated macropore flow and integrate them with frozen soil processes. This means developing model descriptions that explicitly define macropore hydraulic characteristics. Alternative hydrodynamic descriptions like kinematic wave and film flow models (Jarvis, 1994; Germann, 2001; Hincapié and Germann, 2009; Nimmo, 2010) circumvent the need for SMC-derived descriptions of macropore hydraulic conductivity and, additionally, capture the influence of variations in water input on the initiation and cessation of macropore flow (Jarvis et al., 2016, 2017). However, complexity arises in frozen ground because water fluxes are strongly coupled to soil heat transfer. Research is required to integrate these concepts into a framework adapted carefully to consider the influence of heat transfer and associated soil-moisture phase change on water flow within macropores. More importantly for frozen soils, adaptation of dual-domain methodologies will require research on how macropore–matrix exchanges between domains are influenced by freeze–thaw processes and vice versa. A necessary modification must include the effect of freezing of infiltrating water and the reduction in pore space available for flow in the macropore domain. In frozen soils, the specific contact area at the macropore–matrix interface will determine heat exchange between domains and, subsequently, the freeze–thaw dynamics within macropores. In terms of heat transfer, this will be the surface over which infiltrated water may be cooled by the surrounding frozen matrix and may subsequently freeze, causing ice formation and blockage of macropore flow paths. Depending on the temperature of both the infiltrating water and the matrix, vertical macropore flow can be effectively reduced to zero due to the freezing of infiltrated water, or macropores can remain open and flowing.

Conceptualizing these phenomena as a dual-domain process within a numerical modeling framework allows for the influence of both diffuse and preferential flow in unsaturated frozen soils. This framework, linked to surface energy balance and snowmelt dynamics at the ground surface, would improve the ability of physically based models to simulate frozen-soil infiltration and its consequential effects on cold regions hydrological processes, such as soil freeze–thaw, snowmelt redistribution, runoff generation, soil moisture distribution, and groundwater recharge.

A Matrix–Macropore Conceptual Framework for Water and Heat Transfer in Unsaturated Frozen Soil

To better represent the process understanding described above, a conceptual framework of macropore flow needs to (i) emphasize the dynamic and interacting nature of soil freeze–thaw and macropore fluxes, and (ii) transition between frozen and unfrozen conditions while still representing the underlying physics.

Based on this and other findings reviewed, we propose a modification to the conceptual framework developed by Johnsson and Lundin (1991) and Stähli et al. (1996) (Fig. 5 and 6). Because the hydraulic regimes differ considerably, the subsequent heat transfer in both domains and their interactions are also taken into account. In the matrix domain, the model uses traditional concepts of diffuse flow through frozen soil in a three-phase water–ice–air system, with the sequential freezing of pore water with decreasing pore size described by the SFC. These pores are subject to dominant capillary forces and can be blocked by ice during the redistribution of soil moisture from unfrozen soil below. The revised conceptual model includes a distinct macropore domain that is not subject to strong capillary forces and is thus unaffected by freezing-induced moisture redistribution. The lack of capillarity in the macropore domain allows gravity-driven flow, enabling large volumes of water to infiltrate and redistribute under frozen conditions. The newly proposed model domains would be physically defined by intrinsic

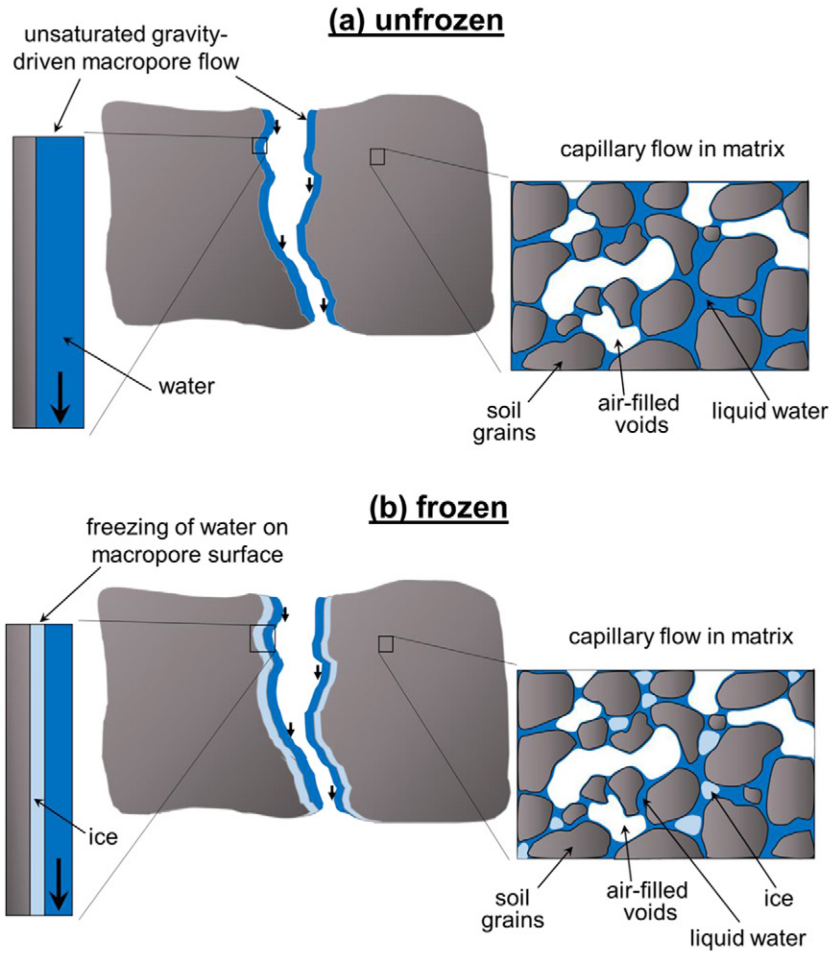

Fig. 5. Conceptual model of frozen combined matrix–macropore flow with capillary flow in the matrix and gravity flow in macropores (adapted from Nimmo, 2010).

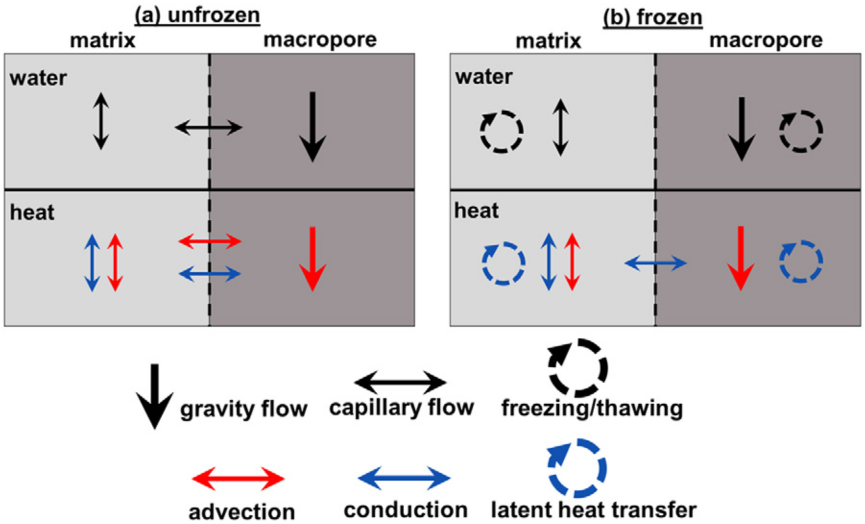

Fig. 6. Conceptual model of dual-domain coupled water and heat transfer for (a) unfrozen soil and (b) frozen soil.

parameters (e.g., macroporosity) of the medium, such that the domain boundary does not shift in time with freezing–thawing. For practical use, this conceptualization assumes a non-deformable medium and does not include changes in domain size due to the expansion of water on freezing. This allows the new model to explicitly consider heat and fluid transfer without changing the domain boundary, as the underlying flow processes are sufficiently distinct. Macropores can then only be blocked by the freezing of infiltrated water and subsequently cannot be thawed until enough energy is conducted from above or the surrounding matrix to thaw the ice. Ice nucleation occurs at the macropore surface (Watanabe and Kugisaki, 2017), creating a solid barrier along the macropore–matrix interface (Fig. 5) and effectively blocking flow between domains. As water flows through the macropore network, it may refreeze depending on the surrounding soil matrix temperature. Simultaneously, the latent heat released from freezing in the macropore domain is transferred to warm the matrix.

The conceptual model of heat transfer illustrated in Fig. 6 is summarized here. Within the matrix domain, three types of heat transfer are considered: (i) heat conduction due to vertical temperature gradients, (ii) advective heat transfer due to vertically moving water, and (iii) latent heat exchange due to soil moisture water–ice phase change. In contrast, only advective and latent heat transfers are considered in the macropore domain. Two types of thermal interactions between domains are considered: (i) heat conduction and (ii) advective heat transfer due to water transfer between domains only when the matrix is unfrozen. Under frozen conditions, heat conduction across domains determines the latent heat transfer for freezing water in the macropore domain. As outlined, these modes of heat transfer can significantly affect water fluxes in both flow domains and their interaction.

It should be noted that this framework enables increased infiltrability at low antecedent moisture and ice contents. This is an important point to make because the air or ice content of the largest pores plays an important role for infiltration in the matrix (i.e., soils without macropores). The revised framework proposed here is an effort to incorporate the consensus that flow in macropores (frozen or unfrozen) differs considerably in terms of flow processes (gravity- vs. capillary-driven flow) and that water in macropores freezes as it is cooled by the surrounding matrix (Watanabe and Kugisaki, 2017). Both of these crucial flow and freezing processes can be implemented in a dual-domain model framework explicitly incorporating macropores and soil heat transfer.

The significance of these processes can be highlighted with some illustrative scenarios. The temperature of melting snow is probably close to 0°C, and thus latent heat is the major source of energy available for melting the frozen matrix. However, the large thermal mass of the matrix relative to macropores may provide a large heat sink. Thus, the rate of heat conduction across the macropore–matrix interface, or the "cooling potential" of the matrix, determines whether water may freeze during flow. The specific contact area of the moving water with the matrix determines the degree of heat transfer between the two domains. Depending on the competing thermal conditions of the matrix and infiltrating water, water in macropores may begin to freeze, reducing the macropore flow capacity and restricting further infiltration. Alternatively, if snowmelt provides a large enough source of water to macropores, thermal advection may be an important thaw mechanism (Roth and Boike, 2001; Ishikawa et al., 2006). In this case, when macropores are actively transmitting water, the thermal regime may be dominated by downward thermal advection. Taking these processes under consideration, both heat and water exchange between domains are treated as source–sink terms, adding or taking energy away from the downward movement of mass and energy in the macropore domain (Fig. 6).

This conceptual model provides a physically based framework that specifically allows the flow regime, hydraulic characteristics, and partitioning of ice in the macropore domain to be linked to the thermal conditions in the matrix via macropore–matrix interaction. This thermal interaction will dictate when macropores can transport water or be blocked with ice. As such, it provides a platform to address several key questions, including:

1. When does infiltration into frozen soil begin with respect to snowmelt?

2. How fast are infiltration rates in frozen soils?
3. Under what set of dynamic conditions does preferential flow bypass the frost zone?
4. Alternatively, when does infiltrating water freeze and restrict subsurface flow?

Such a model could be used to improve the evaluation of factors controlling frozen soil infiltration and redistribution (e.g., snowmelt dynamics, soil thermal and hydraulic properties, antecedent soil moisture, degree of macroporosity) and to explore related hydrologic processes at the hillslope and watershed scales (e.g., runoff, streamflow, groundwater recharge).

Conclusions and Future Research Directions

Hydrological studies spanning seasonally frozen and permafrost environments have shown that snowmelt infiltration in frozen soil can be strongly affected, and even dominated, by macropore flow. Despite these findings, a detailed understanding of the mechanisms of macropore flow in frozen soil and how it varies in response to different soil thermal regimes remains uncertain. A critical limitation has been the lack of clear conceptualization of the dominant flow mechanisms, controls on flow initiation, and infiltration–refreezing dynamics. Current modeling approaches have mainly focused on capillary-based flow concepts, i.e., Richards' equation and the SMC–SFC relationship, which do not adequately represent macropore hydrodynamics. It is hoped that the conceptual model presented here will provide an effective framework for understanding these processes by integrating knowledge of macropore flow with soil freeze–thaw behavior. Improved understanding of the coupling of these heat and water transfer processes will be critical to simulating and evaluating the implications of frozen-soil macropore flow processes in the context of larger watershed-scale snowmelt partitioning (e.g., runoff generation) and other hydrological processes (e.g., contaminant transport). Addressing these issues will require further development of existing macropore flow descriptions and modeling methodologies across a range of scales.

Dual-domain flow models have been successful in simulating preferential flow dynamics in the vadose zone, but the physical descriptions of macropore flow and macropore–matrix interactions require refinement for frozen soils. Specifically, research is required to integrate these concepts into a framework that includes soil heat transfer, freeze–thaw, and pore-water phase change. Because the pore space available for flow is influenced by the spatial configuration and volume of unfrozen water and ice, model parameters and schemes linking macropore–matrix heat transfer to the freezing of infiltrating water will be of critical importance. New modeling methodologies to test these concepts and quantify these dynamics will ultimately enable us to address how fast water flows within a frozen macroporous soil and investigate the conditions that enable water to bypass the frozen zone or, in opposing fashion, cause water to freeze within macropores. Integrating these questions with the significant advances made thus far will enable a better understanding of macropore flow in frozen soils and its hydrological consequences in a changing cryosphere.

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