Influences of Macropores on Infiltration into Seasonally Frozen Soil

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Water frozen in soil can reduce the soil infiltrability, depending on the water content. We hypothesize that air-filled macropores control the infiltration of a seasonally frozen soil under high saturation degrees. Sprinkling experiments with different intensities on a seasonally frozen soil were conducted in two winters at high initial water contents. Brilliant Blue FCF (BB) was sprinkled on four plots equipped with soil moisture and temperature probes to mark flow paths. Frozen layer thickness was measured with infrared thermography of soil sections and overlaid with BB images. The frost depth of the experiments was 8 to 15 cm. Infiltration rates showed reduced infiltration compared with unfrozen conditions. By impeding refreezing of the infiltrating water with added NaCl, infiltration rates of 23 to 29 mm h⁻¹ were measured. Without the addition of salt, the infiltration rates decreased to 5 to 10 mm h⁻¹, attributed to pore blockage by refreezing water. Temperature measurements revealed that the frozen layer only thawed close to the soil surface during the experiments. Blue-stained areas indicated that water was channeled through the frozen layer into the unfrozen soil. In addition, the soil moisture probes below the frozen layer measured an increase in unfrozen water content, whereas total water content in the frozen layer was constant. These observations were explained by a connected air-filled porosity, such as biopores, which allowed water flow even under high initial water contents. These results illustrate the importance of macroporosity in relation to frost depth in controlling the infiltrability of seasonally frozen soils.

Abbreviations: BB, Brilliant Blue FCF; EC, electrical conductivity; IRT, infrared thermography.

More than 50% of the uppermost soils in the northern hemisphere are perennially or seasonally frozen (Zhang et al., 2003). Frozen soil conditions can strongly decrease infiltration rate and hydraulic conductivity depending on the soil water content and soil temperature due to pore blockage with ice (Kane, 1980; Kane and Stein, 1983; Watanabe and Osada, 2016). Because snowmelt is a governing factor for groundwater recharge in many northern regions, reduced infiltration by frozen soil can have an important impact on the water balance (Bayard et al., 2005; van der Kamp et al., 2003). Furthermore, reduced infiltration and thereby higher surface runoff can lead to erosion, higher pollutant or nutrient transport, or increased flood risk (Lundberg et al., 2016; Nyberg et al., 2001; Stadler et al., 2000). Most studies focusing on the effect of frozen soils on infiltration have been performed at high altitudes or latitudes in Canada, Fennoscandia, the European Alps, or northern Japan (Bayard et al., 2005; Gray et al., 2001; Iwata et al., 2008; Suttinen et al., 2008). However, seasonally frozen soils are also present in many uplands of Central Europe (Weigert and Schmidt, 2005). These regions have a different climatic regime than most of the studies described above, with shorter freezing periods and higher minimum temperatures.

Measuring infiltration and water flow in frozen soils is challenging because both energy and mass exchange must include phase change contributions, which are in turn tightly coupled to soil temperature. Therefore, many studies have used numerical models of coupled mass and energy transport to understand infiltration into frozen soil (Fléchinger and Saxton, 1989; Harlan, 1973; Janson and Karlberg, 2001; Watanabe and Flury, 2008). Stähli et al. (1996) and Stadler et al. (1997) developed two-domain models for frozen soils, additionally accounting for preferential transport in the initially air-filled macroporosity.
Besides modeling, measurements have mostly been directed to the determination of hydraulic conductivity of frozen soils, and these have been performed in the laboratory on a cubic centimeter scale with dynamic controls of the boundary conditions (Watanabe et al., 2013; Weigert and Schmidt, 2005; Zhao et al., 2013b). Few studies have measured infiltration or observed snowmelt infiltration and water flow during winter conditions (e.g., Bayard et al., 2005; Iwata et al., 2008; Stähli et al., 1999). Measured infiltration rates have often yielded surprisingly high values, attributed to conducting macropore networks (Johnsson and Lundin, 1991; van der Kamp et al., 2003; Zuzel and Pikul, 1987), although few studies have focused on these macropore effects. Higher snowmelt infiltration was observed in a cracked prairie soil than in uncracked areas (Granger et al., 1984). Espeby (1990) analyzed chemical parameter ratio and that these soils can retain higher infiltration rates than those where the flow relies on the soil matrix. The thickness of the frozen layer, its extent in space, and its thermal state are important variables to predict the influence of macropores on infiltration. Especially because seasonally frozen soils in uplands often have temperatures close to 0°C, small temperature changes (e.g., due to different rainfall temperatures) can influence the spatial extent of the frozen layer and affect water flow path.

The aim of this study was to detect the presence and investigate the influence of macropore flow in shallow frozen soils close to 0°C with respect to the depth of the frozen layer and irrigation rates. We further quantify the effect of energy exchange and phase change in a macroporous frozen soil and the resulting effect on the infiltration capacity. We equipped several experimental plots with soil water content and temperature sensors and conducted irrigation experiments on seasonally frozen soils near 0°C during the winter season combining IRT images with BB infiltration patterns. We show that combining these methodologies (BB and IRT) improves understanding of preferential flow in seasonally frozen soils, information that is sparse in the literature.

Materials and Methods

Study Site

The study site was a typical montane pasture at the Schauinsland Mountain in the Black Forest in southwestern Germany (Fig. 1). The site (47°9′42″15′′ N, 7°90′8877″ E) was at an altitude of 1196 m asl with a slope of 5°. Climate data were obtained from a meteorological station 50 m away from the study site, operated by the German Environmental Protection Agency. The mean annual temperature is 6.0°C; February is the coldest month, with a mean temperature of −1.4°C. The mean annual precipitation is 1936 mm, much of which falls as snow in the winter months. The nearby station of the German Weather Agency at the Feldberg mountain (1490 m asl) reported three winters with frozen soil conditions at depths of 5 or 10 cm from 2009 to 2015. The area is frequently subject to rain-on-snow events (Garvelmann et al., 2015).

The soil at the site is a Cambisol developed over gneiss with a 17-cm Ah horizon followed by a 23- to 33-cm B horizon, underlain by a C horizon with a high rock content (>60%). The soil texture is a sandy loam (58% sand, 29% silt, 13% clay) measured by wet sieving and a PARIO particle size analyzer (METER Environment), which uses the integral suspension pressure method (Durner et al., 2017). The bulk density was 0.93 g cm−3 for the Ah horizon and 1.09 g cm−3 for the B horizon. The porosity of the Ah horizon was estimated as 64% (SD ±4%) determined from 14 soil cores using air pycnometry (Danielson and Sutherland, 1986). The constant infiltration capacity measured on unfrozen soil using a double ring infiltrometer with falling head conditions (water head 12.9–9.2 cm) was 360 mm h−1. Approximately 30 macropores (diameters ≥ 2 mm) per square meter were found at the 10-cm depth, as measured by digging a horizontal profile. The highest root density was found in the upper 15 cm. Ott and Uhlenbrook
(2004) noted that no infiltration excess overland flow should occur in that region of the Black Forest.

**Sprinkling Experiments**

**Flow Path Experiments without Refreezing**

Four plots were prepared in September and October 2015, each with an area of 0.6 m² (1.2 by 0.5 m), and were installed with 5TE capacitance probes (Decagon Devices, METER Environment) measuring volumetric soil moisture (θ), temperature, and electrical conductivity (EC) at depths of 5, 10, 15, 25, and 35 cm (Fig. 1). Sensor performance was confirmed (permittivity ±1) by checking reported values when the sensors were placed in pure ethanol and dry sand. Soil pits were dug upslope of the plots to facilitate excavation during the frozen soil experiments in winter. A 10-cm soil buffer zone between the pit and the plot was then insulated by a coating of tarpaulin, expanding foam, and wooden boards. At two of the four plots, two additional 5TE probes were installed at depths of 10 and 25 cm opposite the five main probes to ensure no horizontal temperature gradient developed as a result of the soil pit. Due to warm winter conditions in 2015 and 2016, the insulating snow was removed regularly to increase soil freezing.

Sprinkling experiments with added NaCl were performed in February and March 2016 after 1 to 2 wk of temperatures below 0°C (Table 1). Two plots were irrigated for 1 h with a lower (30 mm h⁻¹) and a higher (47 mm h⁻¹) sprinkling rate using water at a temperature of 3°C. Natural wintertime precipitation temperature and phase differ among events (Dai, 2008), and infiltration rates of frozen soils close to 0°C might be sensitive to these differences in rainfall temperatures. Therefore, the two additional plots were irrigated for 1 h with sprinkling rates of 37 and 41 mm h⁻¹ but with higher water temperatures (8–10°C) to detect the effect of water temperature on infiltration and the soil thermal regime. Surface runoff was collected every 5 min downslope of the plot using a gutter with clay as sealing material between the gutter and soil. Water infiltration at the surface was calculated from the value of the imposed rainfall rate and the measure of runoff with consideration of the water budget. The dye tracer Brilliant Blue FCF (4 g L⁻¹) was added to the irrigation water to visualize water flow path in the soil (Flury et al., 1994; Weiler and Flühler, 2004). To observe the potential flow path and maximum infiltration into the frozen soil, NaCl was dissolved in the water to lower the freezing point to be equal to the temperature of the frozen layer and thereby prevent it from refreezing. Salt (5 g L⁻¹ NaCl) was added to decrease the freezing point together with the BB to a temperature of −0.35°C. The higher EC values of the sprinkling water could also be detected as increased bulk EC by the 5TE probes. One soil core per plot was collected at each of two depths (0–5 and 5–10 cm) before the experiments to obtain the initial water content of the topsoil. Additionally, one core per plot at the 0- to 5-cm depth

<table>
<thead>
<tr>
<th>Treatment</th>
<th>Date</th>
<th>Topsoil initial water content</th>
<th>Sprinkling intensity</th>
<th>Sprinkling water temperature</th>
</tr>
</thead>
<tbody>
<tr>
<td>Unfrozen control</td>
<td>0.31 (dry)</td>
<td>47 mm h⁻¹</td>
<td>3</td>
<td></td>
</tr>
<tr>
<td>Nov. 2015</td>
<td>0.41 (wet)</td>
<td>51 mm h⁻¹</td>
<td>10</td>
<td></td>
</tr>
<tr>
<td>Frozen Salt</td>
<td>Feb. 2016</td>
<td>0.48</td>
<td>47</td>
<td>3</td>
</tr>
<tr>
<td>Mar. 2016</td>
<td>0.53</td>
<td>37 mm h⁻¹</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>No salt</td>
<td>Feb. 2017</td>
<td>0.56</td>
<td>41</td>
<td>8</td>
</tr>
</tbody>
</table>

**Table 1. Overview of the different sprinkling experiments and the related conditions.**
was sampled 1 h after the sprinkling experiment to determine changes in water content in the frozen layer. Spatial representation of the frozen soil layer was obtained using ground-based IRT. Thermograms of the vertical soil profile were taken before the sprinkling. Additionally, during two experiments (February 2016), we took time-lapse thermograms (one image per minute) to track changes in thermal regime. To do so, the buffer zone to the trench was excavated shortly before the irrigation, and pictures were made from a distance of ~1 m using an InfraTec VARIOCam high-definition infrared camera (InfraTec GmbH) with a resolution of 640 by 480 pixels (~1.1 mm pixel−1). This camera measured thermal infrared emission between 7.5 and 14 μm with a thermal resolution of 0.08°C (at 30°C). All plots were protected against direct sunlight to reduce radiative heating.

Comparative unfrozen soil experiments were performed in November 2015 under wet (θ0–15 cm = 0.41) and dry (θ0–15 cm = 0.31) initial conditions for 1 h with irrigation rates of 51 and 47 mm h−1, respectively (without thermal imagery). To achieve dry initial conditions, the plot was covered for several weeks to shield it from precipitation.

For the comparison of measured infiltration to pure soil matrix infiltration, HYDRUS 1D (Simunek et al., 2013) simulations were performed (without accounting for energy exchange and phase change). The frozen soil hydraulic parameters (van Genuchten–Mualem) for a sandy loam were taken from Zhao et al. (2013b, Treatment 3) but using measured air-filled porosity and unfrozen water content of our soil as the water content at saturation (θs). The simulated frozen layer thickness was 10 cm, divided into two layers (θs0–5 cm = 0.21, θs5–10 cm = 0.36), above 20 cm of unfrozen soil. Unfrozen soil parametrization (sandy loam) was taken from Carsel and Parrish (1988). Initial unfrozen volumetric soil water content for the frozen soil was set to 0.15 for the 0- to 5-cm layer, 0.25 for 5- to 10-cm layer, and 0.30 for the unfrozen soil. We used free drainage as the lower boundary condition and 57 mm h−1 constant rainfall for 1 h as the upper boundary.

Refreezing Infiltration Experiments
Refreezing experiments were conducted in February 2017. Snow removal was not necessary to ensure soil freezing. Soil temperature and frost depth were determined (soil temperature <0°C) at several positions using a laboratory hand-held device Greisinger GMH 3750 (GHM Messtechnik GmbH) with a thermal resolution of 0.01°C and an error ≤0.03°C. One plot was irrigated with 54 mm h−1 for 1 h, and 5 g L−1 NaCl was added for comparison to the conditions of 2016. Furthermore, two sprinkling experiments (52 and 57 mm h−1) were performed for 1 h without the addition of salt to observe the effect of refreezing. The area of the plots was 0.4 m² (0.8 by 0.5 m). No soil moisture sensors were installed during the refreezing experiments. Surface runoff was collected every 5 min, and soil cores at 0 to 5 cm depth were taken before the experiment to determine the initial water content of the frozen layer. The sprinkling water temperature was 1°C.

Data Analysis
Brilliant Blue
Two vertical soil profiles at each plot were dug 18 to 24 h after the end of the sprinkling experiments, and pictures were taken for the analysis of the BB flow pattern with a Nikon D5100 under diffuse light conditions. The pictures had a spatial resolution of ~0.17 mm pixel−1. A frame and gray scale were included in the picture, which was analyzed for percentage of blue-stained pixels using the method of Weiler and Flühler (2004). The method includes geometric correction, color adjustment, and classification. For two profiles, we took thermograms to combine flow and thermal information 1 d after the irrigation. The transformed BB images had a spatial resolution of 1 mm pixel−1, similar to the thermograms.

Infrared Thermography
The temperature obtained by IRT is based on an electromagnetic radiation measurement. Planck’s law integrated over the spectral range of 7.5 to 14 μm gives the relation of surface temperature to emitted radiation of a black body (Vollmer and Möllmann, 2010):

\[ M_\lambda(T) = \frac{1}{\lambda^5} \frac{2\pi h c^2}{\exp(hc/\lambda kT) - 1} \]

where \( M_\lambda \) is the spectral emittance across the range of 7.5 to 14 μm (W m−2 μm−1), \( h \) is the Planck constant (6.626 × 10−34 J s), \( c \) is the speed of light (2.998 × 108 m s−1), \( \lambda \) is the wavelength (here 7.5–14 μm), \( k \) is the Boltzmann constant (1.381 J K−1), and \( T \) is the temperature of the surface (K).

Because the soil is not an ideal black body, the temperature obtained by the camera must be corrected by the emissivity (\( \varepsilon \)) of the measured surface (here unfrozen and frozen soil). Furthermore, measured emittance must be corrected for reflection of infrared radiation emitted by the surrounding area. Atmospheric transmissivity (\( \tau \)) was found by the IRBIS 3.0 plus software (InfraTec GmbH) using the distance between the camera and the plot (~1 m), air temperature, and relative humidity. The transmissivity effect was too small (\( \tau \approx 1 \)) to influence the results. Therefore, the infrared radiation balance measured by the IRT can be written as (Aubry-Wake et al., 2015)

\[ M_\lambda(T_{IRT}) = M_\lambda(T_{soil}) + M_\lambda(T_{ref})(1 - \varepsilon) \]

with the temperature (K) measured by the IRT (\( T_{IRT} \)), the temperature of the soil (\( T_{soil} \)), and of the surrounding area (\( T_{ref} \)); \( T_{ref} \) was measured for every profile using a crumbled and reflattened piece of aluminum foil in front of the object of interest, which served as Lambert reflector with a low emissivity and thereby a high reflectivity (Fokaiades and Kalogirou, 2011).

The Greisinger GMH 3750 hand-held device was used for reference measurements of near-surface temperature of the frozen layer by vertical temperature profiles of 5-cm spatial resolution. The 5TE temperature data were used for the
unfrozen part of the soil. By taking thermograms of the same soil section, without interfering factors (e.g., direct sunlight), all parts of Eq. [2] except for the emissivity are known. Values for the apparent emissivity for the frozen \( \varepsilon_{\text{fr}} \) and unfrozen soil \( \varepsilon_{\text{un}} \) in the spectral range of 7.5 to 14 \( \mu \text{m} \) were calculated and yielded values of \( \varepsilon_{\text{fr}} = 0.920 \) and \( \varepsilon_{\text{un}} = 0.924 \). These values were used for calculating the soil temperature of all soil profiles. Additionally, thermograms were compared with reference measurements using the Greisinger GMH 3750 thermometer and 5TE data. The images showed a mean offset in the range of \( \sim 0.5 \) K for the initial and \( \sim 2 \) K for the BB thermograms compared with the reference measurement and were corrected accordingly. The offset can be explained by slight heating of the soil surface due to diffusive light from the snow surrounding the profiles. Thermal pictures were then used to estimate the mean depth of the frozen layer by classifying the images for areas \(<0^\circ \text{C}\). This 0°C isotherm can be interpreted as the depth of the freezing front, and the frozen water content within this layer is heterogeneous, depending on soil temperature. Greisinger GMH 3750 hand-held device measurements were further used for validation of the estimated frost depth.

Sensor Measurements

In contrast to free water, soil water lower than 0°C is only partly frozen, depending on the relationship between temperature and matric potential, which can be described by the Clausius–Clapeyron equation (Kurylyk and Watanabe, 2013; Spaans and Baker, 1996). Therefore, liquid and frozen water are simultaneously present in frozen soils at temperatures <0°C. Because the dielectric constant of ice differs from that of water (water \( \approx 80\) at \( 20^\circ \text{C} \), ice \( \approx 3 \); Watanabe and Wake [2009]) and the most relative permittivity–water content relationships were developed for unfrozen conditions, soil moisture measurements based on permittivity perform rather poorly for frozen conditions (e.g., Topp equation; Seyfried and Murdock, 1996). Birchak et al. (1974) and Dobson et al. (1985) introduced a fourth phase, which was used in several studies as the ice phase (e.g., Bittelli et al., 2003; Roth and Boike, 2001; Stähli and Stadler, 1997):

\[
\theta_u = \frac{\kappa_c - (1 - \phi) \kappa_l - (\phi - \theta_{\text{tot}}) \kappa_f - \theta_{\text{tot}} \kappa_i}{\kappa_c (T_{\text{soil}}) - \kappa_l}
\]  

[3]

where \( \theta_u \) is the unfrozen water content (dimensionless); \( \kappa \) is the relative permittivity (dimensionless) of the whole soil \( (\kappa_c) \) and the solid \( (\kappa_s = 3.9) \), liquid water \( (\kappa_l = 1) \), and ice \( (\kappa_i = 3.27) \) phase (Roth et al., 1990; Watanabe and Wake, 2009); \( \phi \) is the porosity (dimensionless); \( \theta_{\text{tot}} \) is the total water content. Temperature dependence of \( \kappa_w \) was taken into account using the relationship of Roth et al. (1990) because soil temperature varied substantially over the measurement period:

\[
\kappa_w (T_{\text{soil}}) = 78.54 \left[ 1 - 4.579 \times 10^{-3} (T_{\text{soil}} - 25) + 1.19 \times 10^{-5} (T_{\text{soil}} - 25)^2 - 2.8 \times 10^{-8} (T_{\text{soil}} - 25)^3 \right]
\]  

[4]

The soil temperature \( T_{\text{soil}} \) measured continuously by the 5TE sensors was used for the permittivity correction (accuracy, \( \pm 1^\circ \text{C}; \) resolution, 0.1°C). These probes can export raw permittivity values \( (\kappa_c) \). As shown by Rosenbaum et al. (2011), soil permittivity values from the 5TE have to be corrected for temperature and EC effects due to the low frequency of the probes (70 MHz) and the dielectric losses resulting from ionic conduction (Visconti et al., 2014). The \( \kappa_c \) correction functions of Rosenbaum et al. (2011) for EC and temperature were used to correct the 5TE permittivity measurements. Electrical conductivity was calculated as a function of temperature (removing internal sensor correction to 25°C). For 5TE measurements and soil temperature above 0.3°C, the dielectric mixing model was used excluding the ice phase; 0.3°C was chosen as a threshold temperature because a water content decrease was observed to start at this temperature. Unfrozen water content before and after our sprinkling experiments could be calculated because porosity and total soil water content of the frozen layer were known from the soil cores taken before and after the experiment. The porosity measured in autumn 2015 was used for the unfrozen part of the soil. For an overview of the unfrozen water content throughout the winter of 2015–2016, we assumed the total soil water content during each individual freezing period to be constant at the magnitude measured before the onset of freezing (Stähli and Stadler, 1997). These initial soil water contents were taken from the soil core measurements or from the 5TE readings for the freezing periods without soil core measurements.

The added salt allowed the calculation of the relative EC at a specific depth for each observation time (Öhrström et al., 2004):

\[
\sigma_{\text{rel}} = \frac{\sigma_w - \sigma_{\text{in}}}{\sigma_p - \sigma_{\text{in}}}
\]  

[5]

where \( \sigma_w \) is the EC of the pulse \( (<10 \text{ dS m}^{-1}) \), and \( \sigma_p \) and \( \sigma_{\text{in}} \) are the EC of the pore water during the experiment and the initial pore water EC before the experiment, respectively. Pore water EC was calculated from bulk EC \( (\sigma_b) \) using the model of Hilhorst (2000):

\[
\sigma_{\text{w/in}} = \frac{\kappa_w \sigma_b}{\kappa_w - \kappa_q}
\]  

[6]

where \( \kappa_q \) is the intercept \( (\sigma_b = 0) \) of the \( \sigma_b - \kappa_a \) relationship, which was set to 5.87 derived from laboratory freezing experiments of the soil.

Energy Budget Calculations

Because water flow in frozen soil is always a coupled energy and mass transport, we also estimated the effect of the infiltrating water on the thermal regime. This supports the temperature measurements and can help to interpret observations. This was done by estimating the required energy input of infiltrating water
to change the temperature of the frozen layer. We assumed that energy transfer only takes place between the frozen layer and the infiltrating water without other energy fluxes in or out of the frozen layer (closed energy system) and that temperature is in equilibrium between all soil phases. Thawing of the frozen layer was considered, but refreezing of infiltrating water was only included in the calculation when no salt was added to the irrigation water. The total amount of water at a given temperature that is required to increase soil temperature was calculated with $\theta_{\text{tot}} = 0.53$ (March 2016), $\theta = 0.64$, an initial soil temperature of $-0.2^\circ\text{C}$, and a 10-cm frozen layer thickness.

According to Iwata et al. (2008), the volumetric heat capacity ($J \text{ kg}^{-1} \text{K}^{-1}$) of a soil is

$$ C = C_w \rho_w + C_i \rho_i + C_b \rho_b $$

with the specific heat of solids ($C_s = 1110 \text{ J kg}^{-1} \text{K}^{-1}$, gneiss-granite), water ($C_w = 4217 \text{ J kg}^{-1} \text{K}^{-1}$ at $0^\circ\text{C}$), ice ($C_i = 2117 \text{ J kg}^{-1} \text{K}^{-1}$ at $0^\circ\text{C}$), and $\rho$ the density of water ($\rho_w = 1000 \text{ kg m}^{-3}$ at $0^\circ\text{C}$), ice ($\rho_i = 917 \text{ kg m}^{-3}$ at $0^\circ\text{C}$), and the soil bulk density ($\rho_b = 930 \text{ kg m}^{-2}$) (Eppelbaum et al., 2014; Iwata et al., 2008); $\theta_w$ and $\theta_i$ are the volumetric fractions of unfrozen water and ice, respectively, and were taken from the freezing point depression curve. The freezing point depression curve was calculated following Cherkauer and Lettenmaier (1999), with the pore size distribution index and air entry value of a sandy loam (geometric mean) taken from Rawls et al. (1982).

The energy input ($J$) for the experiments can be calculated with

$$ E_{\text{in}} = m_w C_w \Delta T_m + L m_w $$

where $m_w$ is the mass of water that was sprinkled (kg), $L$ the latent heat of fusion (333,700 J kg$^{-1}$; Bittelli et al., 2003), and $\Delta T_m$ is the temperature difference (K) of the sprinkled water to the soil. The second term ($L m_w$) describes the refreezing of water in the frozen layer. A complete freeze-up of all sprinkled water was assumed to test for the maximum temperature change due to latent heat release. For simulating the experiments with salt that prevent refreezing, the second term was not considered. Different from Iwata et al. (2008), we assumed that the energy input into the soil is not only consumed by the soil heat content (Eq. [7]) but also by thawing of ice, depending on the freezing point depression characteristics of the soil. This energy consumption ($E_{\text{con}}$) can be found as

$$ E_{\text{con}} = V_{\text{soil}} (\Delta \theta_i L + \Delta T_{\text{soil}} C) $$

where $V_{\text{soil}}$ (m$^3$) is the volume of simulated soil, and $\Delta T_{\text{soil}}$ describes the soil temperature increase (K) that goes along with the melting of the ice fraction $\Delta \theta_i$.

**Results**

Figure 2 shows the pre-experimental conditions in winter 2016 with three main frost periods. The experiments were performed in mid-February and March. Air temperature reached a minimum of $-13.2^\circ\text{C}$ on 18 January. Figure 2 shows a time lag between air and soil temperature, and as soil temperature reaches the freezing point a simultaneous decrease of the mean unfrozen water content in the corresponding depth can be observed. During the freezing period in January, the mean unfrozen water content decreased in the topsoil to 15.1% at the 5-cm depth and to 16.9% at the 10-cm depth. The soil pit did not influence the soil temperature conditions of the plots. The control measurements showed that no horizontal temperature gradient developed in the same soil depth between the sensors next to the pit and opposite without a soil pit. Only one sensor pair had a horizontal temperature offset at the 10-cm depth after a 3-wk freezing period in January. The freezing period in February and March occurred after a snowmelt period and showed a relatively high initial total soil water content (February $\theta_{0–10\text{cm}} = 48%$ [v/v], SD = 0.42%; and March $\theta_{0–10\text{cm}} = 53%$ [v/v], SD = 0.02%). We measured a mean frozen layer temperature of $-0.15^\circ\text{C}$ for the experiments in February and $-0.29^\circ\text{C}$ in March using the Greisinger GMH 3750 handheld device. The mean frost depth derived by the image analysis of the corrected thermograms was 7.9 cm for the experiments in February and 10.1 cm for the experiments in March. None of the soil temperature depth profiles measured by the hand-held device indicated a frozen layer extent deeper than the mean frost depth measured by IRT.

Infiltration rates were calculated as the difference between sprinkling rate and measured surface runoff. The first two measurements (5 and 10 min) were combined to compensate for the water that starts ponding at the soil surface before the initialization of surface runoff. Because we had problems with the clay lining for one of the experiments in February 2016 irrigated with 37 mm h$^{-1}$, we were not able to measure the surface runoff of this
plot correctly. The infiltration rates of the other three frozen soil plots in 2016 and the unfrozen comparisons are depicted in Fig. 3. All applied water was infiltrating into the wet unfrozen soil without any occurrence of surface runoff. The dry soil plot (autumn 2015) showed lower infiltration at the beginning of the experiment and a continuous increase over time. In all frozen soil experiments, little surface runoff started directly in the first minute after the beginning of irrigation. All of the plots showed the same overarching behavior: a decrease in infiltration rate in the first minutes before it increases to reach a steady value, independent of the experimental date, frost depth, sprinkling intensity, or water temperature. Only the time required for the infiltration rate to reach a constant value was affected by the sprinkling intensity (i.e., it was longer for low intensities). After 35 min, all plots showed a constant infiltration rate, with mean values between 23 and 29 mm h⁻¹. Similarly, this constant rate was independent of other factors. Comparing these results against the unfrozen wet conditions, the actual infiltration was reduced by more than 50%, or, compared with infiltrability measured by a double ring infiltrometer, infiltration rate was reduced by more than 90%. Cumulative infiltration varied between 26 mm for the high sprinkling rate and 17.5 mm for the lowest. The modeled matrix infiltration using HYDRUS 1D suggested that infiltration rates of ~2.5 mm h⁻¹ should be expected in frozen soil without macropores.

The image analysis of the frozen soil BB profiles depicts many small, stained patterns indicating macropore flow (Fig. 4). Compared with the unfrozen wet conditions, the frozen soil showed fewer stained areas and a layer of lower dye coverage near the surface (Fig. 5). All eight frozen soil BB images showed an increased dye coverage below the layer of lower dye coverage, with a mean peak depth of 9.0 cm for the February and 10.6 cm for the March experiments. This was only slightly deeper than the mean frost depth measured for these plots. The unfrozen dry experiment shows a similar dye distribution as the frozen soil. Furthermore, thermograms provided the possibility to combine BB patterns with the spatial extent of the frozen layer (Fig. 6). The zone of lower dye coverage matches the frozen soil layer extent, and the highest staining could be observed below that layer. Furthermore, Fig. 6 shows the typical behavior wherein most flow occurred in areas where the frozen layer was thin or discontinuous. No changes in soil thermal regime due to infiltrating water could be observed in the time-lapse thermograms during the experiments. Only a thin soil layer at the surface was thawed (0.5–1 cm) through the irrigation, and no further thawing ($T_{soil} > 0°C$) could be determined using all temperature measures (5TE, hand-held device, IRT), independent of sprinkling water temperature.

The mean total topsoil water content (0–5 cm) of the soil cores showed no significant difference before and after the irrigation experiments (paired t-test, $p = 0.916$). During excavation of

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**Fig. 3.** Measured frozen soil infiltration rates of 2016 (refreezing prevented by salt) since the beginning of sprinkling compared with unfrozen conditions in autumn 2015.

**Fig. 4.** Brilliant Blue (BB) profiles of the two plots excavated in February 2016. Blue colors indicate the BB flow patterns, green areas show the soil surface, and black symbolizes dark spots on the pictures (not classifiable).

**Fig. 5.** Percentage of dye coverage for the Brilliant Blue profiles in 2016 with depth. The blue-shaded area shows the SD of the eight images. The black horizontal line indicates the mean frost depth analyzed by infrared thermography. The orange and red line show the dye coverage for the unfrozen conditions in fall 2015.
the soil cores next to the BB profiles, air-filled macropores were clearly visible (example in Fig. 7). The unfrozen soil moisture response of the 5TE was heterogeneous because the data only represent point measurements. The readings of one soil moisture sensor at a depth of 25 cm were removed due to a sensor problem. The soil moisture probes showed an increase of the mean unfrozen water content after the sprinkling at 10 and 15 cm depth (Fig. 8). The increase at a depth of 10 cm was greatest directly after the irrigation and decreased with time, whereas the increase at the 15-cm depth took several hours. One hour after the irrigation, the standard deviation of $\Delta \theta_u$ at 10 cm depth was $\pm 0.01$. Due to the heterogeneous EC response between the four profiles, the mean $\sigma_{rel}$ value is not representative but indicates the fast response in several depths. The fastest mean $\sigma_{rel}$ change was observed at the 5-cm depth but was also a little delayed at depths of 25 and 10 cm (Fig. 9). No responses could be detected at depths of 15 and 35 cm.

In 2017, during the refreezing experiments, the soil froze more deeply than in 2016 (~15 cm), with soil temperatures between $-0.16$ and $-0.24^\circ C$ at the 5-cm depth and a total soil water content of $\theta_{0-10cm} = 0.56$. As observed in 2016, infiltration rate increased with time to a steady value of 18 mm h$^{-1}$ (Fig. 10). A decreased infiltration rate in the first minutes of the experiments was not observed. In the experiments without NaCl added to the sprinkling water (refreezing of water in soil pores possible), the infiltration rate decreased rapidly in the first minutes and reached constant values between 5 and 10 mm h$^{-1}$, which is lower than the refreezing-impeded infiltration. The amount of cumulative infiltration was 7 and 13 mm for the two plots. Again, no thawing could be measured in the frozen layer by the hand-held device except for the upper 0.5 to 1 cm.

Fig. 6. Combined information of the frozen layer from infrared thermography (orange) and Brilliant Blue images for two profiles in March 2016.

Fig. 7. Air-filled macropore (diameter >5 mm) that was found during excavation of the soil cores next to the Brilliant Blue profiles.

Fig. 8. Mean unfrozen volumetric water content change of the four plots with depth since end of irrigation.

Fig. 9. Mean relative electrical conductivity of the pore water (four plots) for different depth since the start of the experiments irrigated with saline water.
To verify the observed behavior of the frozen layer thermal regime during the experiment (temperature stays <0°C except for the upper centimeter), energy balance computations were performed (Eq. [7], [8], and [9]). We calculated the energy input of infiltrating water that is necessary to detect an increase in soil temperature by 0.1°C. For the 2016 experiments without refreezing of infiltrating water but with latent heat consumption due to thawing of ice, an energy input of ~72 mm of 3°C sprinkling water or 28 mm of 8°C water would be necessary to increase the frozen layer temperature from −0.2 to −0.1°C. The 2017 experiments allowed for refreezing of water in soil pores. According to the freezing point depression curve of our soil, an increase of soil temperature from −0.2 to −0.1°C would correspond to a volumetric ice content of 3% that has to be melted. If we include this energy consumption in our calculations, 2.8 mm of the 1°C sprinkling water would be needed in 2017 to increase soil temperature by 0.1°C.

**Discussion**

**Infiltration Rates of Seasonally Frozen Soil**

Many frozen soil studies have measured snowmelt infiltration with melt rates lower than the infiltration capacity, thus not reaching infiltration capacity and accounting for macropore flow. In contrast, we used sprinkling experiments with high intensities to explore the limits of infiltration into a seasonally frozen soil. Measured infiltration capacities were high enough under all applied conditions to infiltrate even high snowmelt rates in our study region. Garvelmann et al. (2014, 2015) found maximum snowmelt rates of 3 mm h⁻¹ for rain-on-snow events in the southern part of the Black Forest. This is in accordance with studies in other regions where most of the snowmelt water could infiltrate into frozen soil without producing large amounts of overland flow (Bayard et al., 2005; Iwata et al., 2011; Sutinen et al., 2008). In our study, soil water contents were relatively high, and therefore the air-filled porosity (8–16%) in the upper 10 cm was dominated by macropores, allowing infiltration even under frozen soil conditions (Granger et al., 1984). The importance of air-filled pores was also highlighted in the study of van der Kamp et al. (2003). However, they measured much higher infiltration rates compared with our study, which can be explained by high air-filled porosities in the upper 15 cm of their soils. van der Kamp et al. (2003) observed higher infiltration in a grassland compared with an agricultural field and related this to a developed macropore network in the...
grassland. Furthermore, Zhao et al. (2013a) attributed the higher magnitude of snowmelt infiltration into frozen ungrazed soils to the higher macroporosity observed at these sites compared with grazed sites.

The infiltration rates strongly decreased compared with the unfrozen experiments, although unfrozen water content was ~15% (v/v) prior to the experiments, and air-filled macropores were available. This reduction is attributed to the decrease in soil matrix hydraulic conductivity due to the freezing of previously water-filled pores (Zhao et al., 2013b). The fact that infiltration rates in the first 10 min appear higher is partially caused by the method used to calculate infiltration: it was estimated as the difference between the surface runoff and the water application rate. Water ponding on the surface before runoff started was assigned to infiltration. Observed surface runoff of the pond water at later time steps results in an increased “apparent” infiltration rate with time. The system reached a steady state wherein the surface runoff and infiltration became constant, which can be attributed to the soil properties. As expected, due to more rapid filling of surface detention, steady state can be seen to have been reached earlier with higher sprinkling intensities. Additionally, increasing infiltration rates during the experiment could be indicative of the initiation of macropore flow from a saturated soil layer (Weiler and Naef, 2003) because thawing and saturation of the upper 0.5 to 1 cm was observed in all frozen soil experiments. Higher sprinkling water temperature caused neither higher infiltration rates nor a systematically quicker increase in constant rates. Additionally, an increase in infiltration rate with time can be caused by a thin basal ice sheet that blocks pores and that must thaw before infiltration can start. Bayard et al. (2005), for example, observed such a basal ice sheet under a snow layer. The mean potential infiltration rate (salt added) of 24 mm h⁻¹ observed within this study compared with the HYDRUS-1D calculations or other modeling studies of pure matrix flow (e.g., Zhao and Gray, 1999) suggests significant macropore flow. Also, the modeling experiments of Stadler et al. (1997) and Stähli et al. (1996) underestimated measured snowmelt infiltration without adding a second fast flow domain.

Without the addition of salt, the infiltration rate decreased more rapidly and achieved a lower rate than when refreezing was impeded. This behavior and the range of infiltration rates matches in general with other frozen soil studies, which found rates between 0 and 8 mm h⁻¹ (Hayashi et al., 2003; Kane and Stein, 1983; Zhao and Gray, 1999; Zuzel and Pikul, 1987). Initial conditions in 2016 and 2017 were in the same range. Only a slightly higher initial water content resulted in a lower final infiltration rate. Hence, the observed reduction in infiltration between the salt and no-salt experiments, combined with observation of primary flow in the macropores, indicates that refreezing in macropores should be considered in the establishment of infiltration capacity.

Flow Path and Frozen Layer Extent

The observed BB images support the theory of macropore flow-dominated infiltration at high water contents. Initially air-filled macropores channel the water through the frozen layer, and hence water bypasses this layer of low matrix permeability. Based on the dye-stained patterns, little interaction with the frozen soil matrix was observed. This is due to high saturation and restricted mixing by pore blockage of ice. The interaction of water with the matrix, however, increases in the lower unfrozen layer. The main peak in blue-stained areas can be seen near the frozen–unfrozen boundary because water has the potential to flow into the soil matrix. Furthermore, the stained areas in the unfrozen layer show not merely the presence of air-filled macropores but also the connectivity of those macropores. The quickly increasing mean water content at the 10-cm depth and the rapid increase in EC further indicates fast preferential flow. The total water content in the frozen layer was constant, however, indicating again that water is flowing though large pores that cannot hold the water against gravity.

We showed that IRT is an efficient method for measuring the spatial variability of the frozen layer, particularly when merged with the dye patterns from the BB analysis. This approach is especially important to evaluate the infiltration rate in frozen soil if the soil frost is patchy rather than continuous (Jones and Pomeroy, 2001). However, to have a fully quantitative method will require further effort dedicated specifically to spatial validation of the frozen layer extent measured with IRT. Many roots were found where the frozen layer was thin. This possibly supports the finding of Watanabe and Kugisaki (2016) that macropores can reduce the freezing depth by condensation of water in themselves. Furthermore, IRT was shown to provide information about the thermal status of the soil. Except in the upper centimeters, soil did not thaw during the BB experiments, which is due to the large amounts of energy needed to thaw a frozen layer. Based on the energy balance calculation, we showed that nearly the complete sprinkled water amount of the 2016 experiments (23 mm, 8°C) had to be stored in the frozen layer to increase the temperature by a measurable difference (0.1°C). However, there should be almost no energy exchange between water bypassing the frozen layer and the frozen layer itself. This explains why no large temperature increase and thawing was measured.

The dry unfrozen comparison shows a BB peak at a similar depth as for the frozen soil. The temporal dynamics of overland flow for the dry unfrozen soil were similar to observations on hydrophobic soils (e.g., Butzen et al., 2015). A hydrophobic layer could have evolved during the drying process in autumn, which in turn could have triggered preferential flow (Nimmo, 2012) by macropore initiation at the soil surface (Weiler and Flühler, 2004). Because both macropore-controlled infiltration processes (dry, frozen) showed a stained area peak at similar soil depths, a majority of biopores appeared to have ended at this depth. Rooting depths could be relevant insofar as it creates connected macroporosity. Brown et al. (2010) measured rooting depth of common grass species which were present at our study site. They found the highest amount of root mass in the upper 15 cm, which is deeper than the measured frost depth in our experiments. This finding supports
the hypothesis that frost depth is a minor factor in infiltration rate if the air-filled porosity (e.g., the rooting depth) is connected to zones deeper than frost depth. The supposition and evidence that water in frozen soil mainly flowed in root-created macropores concurs with the findings of Espeby (1990), who found snow water chemical signatures (pH, alkalinity, and $\delta^{18}$O) in the water of frozen soil root channels after snowmelt.

Compared with other studies (e.g., Iwata et al., 2011; Stähli et al., 2004), the frozen layer in this study was rather thin and did not have a strong impact on infiltration rate between the experiments (February 2016, March 2016, February 2017). Other studies have found that frozen layer thickness has a strong impact on infiltration. Iwata et al. (2010, 2011) compared snowmelt infiltration on agricultural soils with high porosity and different frost depths in northern Japan and found that thin frozen layers (11–21 cm) did not impede snowmelt infiltration, whereas deeper frost (37–42 cm) did. In contrast to our study, Stähli et al. (2004) observed more blue-stained areas in the frozen layer on a grassland–Ericaceae shrubland. This was probably due to the higher air-filled porosity of ~30% in the frozen layer of their study but a missing connection of air-filled porosity to the unfrozen zones below 40 to 50 cm. This also supports the process assumption of reduced water channeling through the frozen layer if the air-filled porosity ends in “dead end” pores in the frozen soil layer. Therefore, connected biopores, in relation to frost depth, seem to be an important parameter for predicting infiltration into seasonally frozen soil.

**Refreezing and Thermal Regime**

Energy balance calculation reveals that large amounts of water are necessary to increase soil temperature and to thaw the frozen layer if phase change is impeded by salt. Nevertheless, refreezing of water in soil pores strongly decreases the required amount of infiltrating water for increasing soil temperature due to the latent heat of freezing. Because no large temperature change and thawing was observed in the frozen layer, the question was whether we could exclude refreezing of water in macropores as an explanation for the measured decrease in infiltrability. By using energy balance calculations, Iwata et al. (2010) concluded that there was no refreezing in the soil with frozen layer temperatures of $\sim$0.1°C. In our soil, results indicate that 2.8 mm (35% of the macroporosity in the frozen layer) had to freeze to increase the frozen layer temperature by 0.1°C. Therefore, a freeze-up of a large fraction of the air-filled porosity seems plausible without a detectable increase in soil temperature even with the assumption of a homogenous soil and full energy transfer of the irrigated water into the frozen layer. Because frozen soil temperatures were relatively high (approximately $\sim$0.2°C), refreezing of water in macropores is supposed to be a refreezing and blocking of some pores close to the surface (“plugs”) rather than a complete freeze-up of the macropore network (Watanabe and Kugisaki, 2016). This process could explain the thawing in the upper centimeter in our study. In addition to our findings, other studies measured a reduction of infiltration rate of frozen soils and supposed a connection to refreezing meltwater (Bayard et al., 2005; Granger et al., 1984). The refreezing of water in macropores as observed in our tests suggest that this process could be important even at temperatures not far below 0°C. Although the soil surface thawed 0.5 to 1 cm in all our experiments, an enhanced thawing of thin basal ice patches due to the addition of salt and higher water temperatures in the 2016 experiments cannot be excluded and may have occurred in addition to the refreezing of water in macropores.

**Implications for Catchment and Cold Region Hydrology**

It is also important to discuss the effect of macropores on the infiltration into frozen soils, as found in our plot experiments, in the context of catchment-scale reaction. The numerous connected and persistent air-filled macropores created by flora and soil fauna combined with low frost depth (Watanabe and Kugisaki, 2016) seem to explain why many studies did not find an effect of frozen soils on catchment runoff responses in forests (Fuss et al., 2016; Lindström et al., 2002; Nyberg et al., 2001; Stähli, 2017). Therefore, surface runoff due to frozen soils seems to be more common in catchments with a deep frozen layer (>30 cm) or high water content in autumn and a low connected macroporosity by, for example, biopores. Hence, agricultural catchments might have a higher potential of overland flow and erosion under frozen soil conditions, as indicated by Shanley and Chalmers (1999).

The relevance of macropores on infiltration in cold regions is highlighted in Fig. 11 as a conceptual diagram of freezing processes and their implication on infiltration on a plot to field scale. A deep frozen layer (greater than ~30 cm) is more impermeable and conducts water with the low hydraulic conductivity of the frozen soil matrix (Zhao et al., 2013b). In this case, depending on the ecosystem, macropores may rarely be connected to these deeper zones. In contrast, for a thinner frozen layer that develops under a more moderate climate, flora and soil fauna seem to be important modifiers of soil properties through creation of connected macropores. These macropores are air filled even under the high water contents that are common during winter conditions in many climates. In addition to climates with short freezing periods or higher temperatures, infiltration processes near 0°C are important broadly for regions with frozen soil beneath a thick snowpack. In the case of a thick snowpack, soil is thermally decoupled from the atmosphere and often reaches temperatures near 0°C before the start of snowmelt (Bayard et al., 2005; Nyberg et al., 2001; Zhang, 2005). As indicated in Fig. 11, climate change can result in a shift from permafrost and deep frozen soils toward thin seasonally frozen layers (Saito et al., 2007). Hence, the effect of connected air-filled macroporosity can be even more important for infiltration in cold regions in the future. For a more reliable prediction of the water balance in cold regions, it is therefore important to include more information on the depth distribution of biopores for different soils and land uses and to incorporate phase change processes into hydrological models.
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Conclusions

Our measurements showed that even a thin seasonally frozen soil layer (8–15 cm thick) at high saturation (>50% v/v) and close to 0°C strongly reduces the infiltration rate compared with unfrozen conditions. However, connected macropores retain an infiltration capacity of around an order of magnitude higher than the calculated matrix infiltration rate, when refreezing of infiltrating water is impeded. For our experiment, an irrigation rate of 30 mm h⁻¹ activated the macropore network, and higher sprinkling rates did not yield higher infiltration capacities. However, the required sprinkling rate to activate the macropore network is also dependent on the matrix infiltration rate of the frozen soil. Partly refreezing of infiltrating water in macropores (“plugs”) can potentially further reduce the infiltration capacity, even for those soils close to 0°C. Due to the bypassing of most infiltrating water through the frozen layer, energy exchange between the water and frozen layer and therefore thawing was minor, independent of sprinkling water temperature. Infrared thermography could be used to detect the depth and heterogeneity of the frozen layer and, in combination with the Brilliant Blue images, could reveal zones of higher connected porosity where the frozen layer was thin. Further research is necessary to identify the effect of connected macroporosity to frost depth for different cold climates, ecosystems, land uses, and initial water contents. Macropore flow initiation and refreezing processes during infiltration should be studied in more detail to determine its relation to soil temperature and the consequences on infiltration rate (Mohammed et al., 2018). Especially in the light of climate change, detailed knowledge of seasonally frozen soil could be important for the future assessment of the water balance, runoff generation, and biogeochemistry in cold climates.

References


