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A Simple Temperature-Based Method to Estimate Heterogeneous Frozen Ground within a Distributed Watershed Model

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Abstract

Frozen ground can be important to flood production and is often heterogeneous within a watershed due to spatial variations in the available energy, insulation by snowpack and ground cover, and the thermal and moisture properties of the soil. The widely-used Continuous Frozen Ground Index (CFGI) model is a degree-day approach and identifies frozen ground using a simple frost index, which varies mainly with elevation through a temperature-elevation relationship. Similarly, snow depth and its insulating effect are also estimated based on elevation. The objective of this work is to more accurately represent the spatial heterogeneity of frozen ground in a distributed hydrologic model, Gridded Surface Subsurface Hydrologic Analysis (GSSHA), by modifying the CFGI method. Among the modifications, the snowpack and frost indices are simulated by replacing air temperature (a surrogate for the available energy) with a radiation-derived temperature that aims to better represent spatial variations in available energy. Ground cover is also included as an additional insulator of the soil. Furthermore, the modified Berggren Equation, which accounts for soil thermal conductivity and soil moisture, is used to convert the frost index into frost depth. The modified CFGI model is tested by application at six test sites within the Sleepers River Experimental Watershed in Vermont. Compared to the CFGI model, the modified CFGI model more accurately captures the variations in frozen ground between the sites, inter-annual variations in frozen ground depths at a given site, and the occurrence of frozen ground.

1 Introduction

Frozen ground (also known as frozen soil or soil frost) can play a significant role in the hydrologic response of watersheds because it impedes the infiltration of snowmelt and rainfall and enhances runoff. For example, Dunne and Black (1971) examined the Sleepers River Experimental Watershed (SREW) in northeastern Vermont and observed that almost 50% of meltwater left research plots as runoff due to frozen ground. Frozen ground has also been shown to contribute to the periodic flooding of the Red River in North Dakota (Stoner et al., 1993), the 1936 floods in New England (Diebold, 1938), and severe flooding and erosion events in the Pacific Northwest (Johnson and McArthur, 1973). Although the timing and

spatial pattern of frozen ground can greatly affect the amount and type of runoff within a watershed (Wilcox et al., 1997), they are often difficult to simulate in many hydrologic models.

The presence, spatial pattern, and depth of frozen ground are driven by mass (water) and energy balances. The energy available from the atmosphere to freeze or thaw the soil is subject to the insulation of the snowpack (Pearson, 1920; Willis et al., 1961) and ground cover including any vegetation, woody debris, and leaf litter (Brown, 1966; Fahey and Lang, 1975; Sartz, 1973). MacKinney (1929) found that ground cover reduced the depth of frost penetration by 40% at a test site in Connecticut. Additionally, the presence and depth of frozen ground is affected by soil moisture (Fox, 1992; Willis et al., 1961) and the thermal conductivity of the soil (Farouki, 1981; Johansen, 1977).

Frozen ground has proven difficult to simulate within distributed hydrologic models due to complex interactions of energy and water between the atmosphere, snowpack, and soil (Dun et al., 2010; Kennedy and Sharratt, 1998; Lin and McCool, 2006). Physically-based models of frozen ground, such as the Simultaneous Heat and Water (SHAW) model (Flerchinger and Saxton, 1989), have large data requirements, which restricts their applicability. To reduce the data requirements, simple temperature-index or degree-day methods have been developed (Molnau and Bissell, 1983; Rekolainen and Posch, 1993) and used within watershed models, including LISFLOOD (De Roo et al., 2001; Van Der Knijff et al., 2010), CREAMS (Rekolainen and Posch, 1993) and the Gridded Surface Subsurface Hydrologic Analysis (GSSHA) model (Downer and Ogden, 2004). Degree-day approaches typically accumulate the daily average temperature as a frost index (°C-days). When the frost index exceeds a threshold, the soil is considered frozen and impermeable to infiltration. Some degree-day approaches, such as the Continuous Frozen Ground Index (CFGI) (Molnau and Bissell, 1983), account for the insulating characteristics of the snowpack, but these approaches generally do not consider ground cover. Most degree-day models determine the spatial pattern of frozen ground using a temperature gage, the watershed elevations, and a lapse rate (which is also used to simulate spatial variations in snowpack that in-turn impacts frost depth). Degree-day methods have been successful in capturing frozen ground events that lead to increased runoff (Molnau and Bissell, 1983), and higher frost index values have been shown to correlate to deeper frost depths (Vermette and Christopher, 2008; Vermette and Kanack, 2012). However, testing of degree-day approaches has been limited because the frost index is not a physical property that can be compared to measurements.

The objective of this paper is to better estimate the spatial pattern of frozen ground in GSSHA by modifying the CFGI method in four ways. First, the CFGI method is coupled to an improved snowpack model. In past applications of GSSHA, the CFGI method has been coupled with a temperature-index (TI) snowpack model based on SNOW-17 (Anderson, 1973; Anderson, 2006). However, Follum et al. (2015) proposed a Radiation-derived Temperature Index (RTI) snow model that uses a proxy temperature instead of air temperature to represent the energy available to the snowpack. Compared to the TI model, the RTI model more directly includes the effects of shortwave radiation and canopy cover and was shown to better represent the spatial variations of snow cover and snow water equivalent (SWE) in the Senator Beck Basin in Colorado. The RTI model is adopted to simulate the snowpack in the present study. Second, the effects of shortwave radiation and canopy cover are included in the CFGI model when calculating the energy available at the snow or ground surface. These effects are

included by using a similar radiation-derived proxy temperature when calculating the frost index. Third, the insulation effects of ground cover are included by modifying the frost index equation. Fourth, frost depth is computed from the frost index value and is a model output. The modified Berggren Equation and similar Stefan Equation have been previously used to correlate degree-days to frost depth (Carey and Woo, 2005; DeWalle and Rango, 2008; Fox, 1992; Woo et al., 2004); a similar approach is used here to convert the frost index to frost depth.

The following sections first describe the existing TI and CFGI models within GSSHA. The combination of these two models serves as the baseline or control case for the experiments. Then, the RTI snow model and the modified CFGI frozen ground model (referred to as modCFGI) are described. Finally, the results of the TI/CFGI model and RTI/modCFGI models are compared to each other and to observations of snow depth, SWE, and frost depth at the SREW.

2 Methodology

2.1 TI Snowpack Model

The TI snow model was implemented into GSSHA by Follum et al. (2014), who provides additional information about the model. Although GSSHA allows a variable time step for multiple processes, it always uses an hourly time step (Δt) for snow calculations. GSSHA utilizes a structured grid in which each cell can have a different air temperature T_a ($°C$) and precipitation P (m h^{-1}). Air temperature is the primary driver of snowpack dynamics in the TI model and is estimated as:

$$T_a = T_g + \emptyset(E_g - E_c), \tag{1}$$

where T_g (°C) is the air temperature at a gage, $\emptyset$ is a linear lapse rate (°C km^{-1}), and E_g and E_c (m) are the elevations of the temperature gage and the grid cell where T_a is being calculated, respectively. Precipitation accumulates as SWE (m) when $T_a \leq T_{px}$, where T_{px} is the freezing point (0°C by default). Interception of snow by vegetation and the subsequent sublimation, melt, and fall-off are neglected in the TI model.

Before the snowpack begins to melt, its heat deficit (or cold content) must be overcome. The change in heat deficit ΔD_t (mm of SWE), due to a temperature difference between the snow surface and air, is calculated as:

$$\Delta D_t = N_{mf,\max}(dt/6)(M_f/M_{f,\max})(A_{TI} - T_{sur}), \tag{2}$$

where A_{TI} is the antecedent temperature index (°C), which is calculated using T_a and the antecedent snow temperature index parameter A_{TIPM}, and T_{sur} is the snow surface temperature (see Anderson (2006) for details regarding A_{TI} and T_{sur}). $N_{mf,\max}$ is the maximum negative melt factor (mm °C^{-1} (6 h)$^{-1}$), which is a parameter. M_f is the melt factor (mm °C^{-1} dt^{-1}), which is calculated as:

$$M_f = (dt/6)[S_v A_v(M_{f,\max} - M_{f,\min}) + M_{f,\min}], \tag{3}$$

where S_v and A_v are seasonal melt adjustments that change by Julian day, and $M_{f,\max}$ and $M_{f,\min}$ are the maximum and minimum melt factors (mm °C^{-1} (6 h)$^{-1}$), which are parameters.

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Once the heat deficit is overcome, SWE decreases as melt occurs. During normal conditions, the melt M (mm of SWE) is:

$$M = \left[M_f(T_a - T_{mbase}) + 0.0125 P f_r T_r\right] \Delta t \ , \tag{4}$$

where T_{mbase} is the temperature at which melt begins (0°C by default), f_r is the fraction of any precipitation that is rain (see Anderson (2006) for details), and T_r is the precipitation temperature (assumed equal to T_a or 0°C, whichever is greater). During rain-on-snow events (more than 1.5 mm of rainfall in the previous 6 h), M is calculated from a simple energy balance:

$$M = \sigma\,[(T_a + 273)^4 - 273^4]\Delta t + 0.0125 P f_r T_r + \ 8.5 f_u(\Delta t/6)[(r_h e_{sat} - 6.11) + 0.00057 P_a T_a], \tag{5}$$

where σ is the Stefan-Boltzmann constant, f_u is the average wind function (mm mb^{-1} (6 h)$^{-1}$) (see Anderson (2006) for details), r_h is the relative humidity (assumed to be 0.9 during rain-on-snow events) (Anderson, 1973, 2006), P_a is atmospheric pressure (mb) (either measured or calculated from elevation) (Anderson, 2006), and e_{sat} is the saturation vapor pressure (mb) (calculated based on Smith (1993)). The ripeness of the snowpack affects the amount of melt that is released and is controlled by the liquid holding capacity L_{hc}, which is a specified percentage of the ice in the snowpack (Anderson, 2006).

For frozen ground calculations, the snow depth is needed from the snow model. The snow depth D_s (cm) is found from the SWE and the snowpack density. GSSHA uses the single-layer snow density functions from SNOW-17 (Anderson, 1976; Anderson, 2006). The density of newly fallen snow ρ_n (gm cm^{-3}) varies between 0.05 ($T_a \leq -15°C$) and 0.15 ($T_a = 0°C$) according to:

$$\rho_n = 0.05 + 0.0017\ (T_a + 15)^{1.5}. \tag{6}$$

Increases in snowpack density ρ_x from compaction, destructive metamorphism, and melt metamorphism due to the presence of liquid water are calculated as (Koren et al., 1999):

$$\rho_{x,t} = \rho_{x,t-1}\left(\frac{e^{B_2}}{B_2}\right) e^{B_1}, \tag{7}$$

where:

$$B_1 = c_3\ c_5\ dt\ e^{c_4\,T_s - c_x\,\beta\,(\rho_{x,t-1} - \rho_d)}, \text{ and} \tag{8}$$

$$B_2 = W_{t-1}\ c_1\ dt\ e^{0.08\,T_s - c_2 \rho_{x,t-1}}. \tag{9}$$

The variable t is an index for time, W is the ice portion of the snow pack (cm, $W = 100\ S_{swe,t-1}$) where S_{swe} is the snow water equivalent on the ground in m, T_s is the average snow pack temperature (°C, calculated based on Anderson (2006)), and ρ_d is the threshold density above which destructive metamorphism decreases (ρ_d is set to 0.15 gm cm^{-1} based on Anderson (2006)). Finally, $\beta = 0$ if $\rho_{x,t-1} \leq \rho_d$, and $\beta = 1$ if $\rho_{x,t-1} > \rho_d$, and c_1 through c_5 are constants (see Anderson (2006) for details).

2.2 CFGI Frozen Ground Model

The CFGI model was originally developed as a lumped model for flood forecasting in the Pacific Northwest, but it has been used in distributed models as well (De Roo et al., 2001; Van Der Knijff et al., 2010). The rationale of the CFGI

method is that air temperature ultimately controls the ground temperature, but its impact is moderated by the insulating effects of any snowpack. The presence of frozen ground is determined by the frozen ground index F (°C-days), which is calculated as:

$$F_t = F_{t-1}A - T_a\, e^{-0.4K_sD_s}, \tag{10}$$

where A is a daily decay coefficient and K_s is the snow reduction coefficient (cm^{-1}). A controls the persistence of the F values, and K_s controls the insulation from the snowpack. Molnau and Bissell (1983) recommended changing K_s depending on whether T_a is above or below freezing (denoted as $K_{s,T_a>0°C}$ and $K_{s,T_a<0°C}$, respectively).

Higher values of F indicate a higher likelihood that the ground is frozen. Once F exceeds a specified threshold ($F_{threshold}$), the ground is considered frozen and infiltration is restricted. Molnau and Bissell (1983) found the ground to be frozen when $F > 83$ °C-days and thawed when $F < 56$ °C-days. When F is between these values, the ground could be either frozen or thawed. It is worth noting that the F does not depend on soil moisture, which is known to affect the initialization and depth of frozen ground (Kurganova et al., 2007; Willis et al., 1961).

2.3 RTI Snowpack Model

The RTI model makes two modifications to the TI model: (1) it accounts for interception of snow by the vegetation canopy and (2) it uses a radiation-derived temperature T_{rad} ($°C$) to better describe the available energy. Effective precipitation P_{eff} (m h^{-1}) is the amount of snow that reaches the ground and is calculated:

$$P_{eff} = P - I + D, \tag{11}$$

where I is the snow intercepted by the canopy (m h^{-1}), and D is the amount of previously intercepted snow that is unloaded to the ground surface (m h^{-1}). I is calculated using a physically-based method developed and tested by Pomeroy et al. (1998) and Hedstrom and Pomeroy (1998). As implemented by Liston and Elder (2006):

$$I_t = I_{t-1} + 0.0007(I_{\max} - I_{t-1})[1 - \exp(-1000\, P/I_{\max})], \tag{12}$$

where $I_{\max}$ is the maximum interception storage (kg m^{-2}), which is estimated based on leaf area index (L_{AI}) as $I_{\max} = 4.4\, L_{AI}$ (Hedstrom and Pomeroy, 1998). When $L_{AI} = 0$, P_{eff} is set equal to P. Unloading of intercepted snow D only occurs when $T_a > 0$°C and is calculated using a simple temperature index method (Liston and Elder, 2006):

$$D = 1.61 \times 10^{-8}(T_a - 273.16)\Delta t. \tag{13}$$

The amount of snow water equivalent residing in the canopy S_{swec} (m) is then calculated as:

$$S_{swec,t} = S_{swec,t-1} + I - S_{sub} - D, \tag{14}$$

where S_{subc} (m) is the amount of snow sublimated from the canopy. Following Liston and Elder (2006), S_{subc} is calculated as:

$$S_{subc} = C_e\, I\, \varphi\, \Delta t, \tag{15}$$

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where C_e is the non-dimensional canopy exposure coefficient and φ is the sublimation loss coefficient for an ice sphere. C_e is a function of both I and $I_{\max}$ (Pomeroy and Schmidt, 1993), and φ is calculated as a function of T_a and r_h (%) following Liston and Elder (2006).

The RTI model replaces T_a in Eq. (4) and (5) with a radiation-derived proxy temperature T_{rad} (°C). In those equations, T_a is used to conceptually represent the energy available to the snowpack. T_{rad} has a similar purpose but is intended to improve the estimation of available energy. T_{rad} is calculated by assuming that the radiation terms dominate the energy balance at the snow surface. Thus:

$$R_{LW\uparrow} = R_{SW,net} + R_{LW\downarrow}, \tag{16}$$

where $R_{LW\uparrow}$ is outgoing longwave radiation, $R_{SW,net}$ is the net incoming shortwave radiation, and $R_{LW\downarrow}$ is the downwelling longwave radiation. The right side of Eq. (16) represents the energy that is supplied to the snowpack via the atmosphere. $R_{LW\uparrow}$ (W m^{-2}) is the radiative response of the snowpack to that energy. Using the Stefan-Boltzmann Law, $R_{LW\uparrow}$ can be written in terms of a temperature T_{rad}:

$$T_{rad} = \left(\frac{R_{SW,net}+R_{LW\downarrow}}{\varepsilon_{snow}\,\sigma}\right)^{1/4} - 273.15, \tag{17}$$

where ε_{snow} is the emissivity of snow (assumed to be 1.0) and σ is the Stefan-Boltzmann constant.

$R_{SW,net}$ is calculated:

$$R_{\mathrm{SW,net}} = (1-\alpha_s)(R_{\mathrm{SW},0}\,\varphi_r\,\varphi_{atm}\,\varphi_c\,\varphi_v\,\varphi_s\,\varphi_t), \tag{18}$$

where $R_{SW,0}$ is the solar constant (Liou, 2002) and α_s is the albedo of the snowpack, which is calculated based on the time elapsed since the most recent snowfall and whether melt is occurring (Henneman and Stefan, 1999). φ_r accounts for distance from the Earth to the sun (based on Julian day (TVA, 1972)), φ_{atm} accounts for atmospheric scattering (based on elevation (Allen et al., 2005)), φ_c accounts for absorption by clouds (based on fractional cloud cover (TVA, 1972)), φ_v accounts for vegetation (set equal to the vegetation transmission coefficient (Bras, 1990)), φ_s accounts for the slope/aspect of the terrain (based on latitude, slope, and azimuth angle (Duffie and Beckman, 1980)), and φ_t accounts for topographic shading (based on elevation, azimuth angle, and solar elevation angle).

$R_{LW\downarrow}$ is calculated from the contributions of the atmosphere (including clouds) and the canopy:

$$R_{LW\downarrow} = \sigma\varepsilon_a(T_a+273.15)^4(1.0+0.17\,N^2)(1-F_c) + F_c\sigma\varepsilon_c\left(T_{canopy}+273.15\right)^4, \tag{19}$$

where ε_a is the air emissivity, N is the fractional cloud cover, F_c is the fractional canopy cover (estimated from L_{AI} following (Liston and Elder, 2006; Pomeroy et al., 2002)), ε_c is the canopy emissivity (assumed equal to 1 following Sicart et al. (2004)), and T_{canopy} is the canopy temperature (°C) which is assumed equal to T_a following DeWalle and Rango (2008).

Because the TI model uses T_a to drive snowpack dynamics, those dynamics are only directly associated with the downwelling longwave radiation from the air, which is a component of $R_{LW\downarrow}$. Furthermore, the spatial variations in the available energy depend only on the variations of T_a, which are inferred from elevation. T_{rad} in the RTI model considers both $R_{SW,net}$ and $R_{LW\downarrow}$ and thus accounts for heterogeneity in topographic orientation and shading as well as canopy cover.

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The TI model partially accounts for seasonal variation in solar radiation and snow albedo by empirically adjusting M_f as shown in Eq. (3). In the RTI model, seasonal variations in solar radiation and snow albedo are included in T_{rad}, so a constant melt factor M_f is used (Follum et al., 2015).

2.4 modCFGI Frozen Ground Model

The CFGI model is modified in three ways to create the modCFGI model. First, the proxy temperature T_{rad} is used in place of T_a to represent available energy. Second, ground cover (leaf litter, woody debris, etc.) is included as an insulator in the frozen ground index, and third, frost depth is calculated based on the frozen ground index.

The CFGI uses T_a in Eq. (10) to represent the energy that is available to heat the ground surface. In the modCFGI model, T_a is replaced with T_{rad}. T_{rad} is calculated using α_s (see Eq. (17) and (18)) when snow is present, and the albedo of the land cover when snow is not present. By using T_{rad}, the modCFGI model is expected to better represent the spatial heterogeneity of energy supply due to variations in the topography and canopy cover within a watershed.

The insulation by the ground cover is included by modifying Eq. (10) to become:

$$F_t = F_{t-1}A - T_{rad}\, e^{-0.4\left(K_s D_s + K_{gc} D_{gc}\right)}, \tag{20}$$

where K_{gc} is the ground cover reduction coefficient (cm^{-1}) and D_{gc} is the depth of ground cover (cm). This formulation retains the original form of the CFGI model but includes insulation from both snowpack and ground cover.

The frost depth is calculated based on the modified Berggren Equation. As originally proposed (and described by DeWalle and Rango (2008)), this equation relates the number of degree days in the freezing/thawing period U (ºC-days) to the maximum frost depth $Z_{\max}$ (m) as follows:

$$Z_{\max} = \lambda(48\, U\, \delta^{-1}\Omega_m)^{1/2}, \tag{21}$$

where λ is a dimensionless coefficient that accounts for changes in sensible heat of the soil, δ (J m^{-3}) is the latent heat of fusion of the soil, and Ω_m (J m^{-1} h^{-1} °C^{-1}) is the mean thermal conductivity of the frozen and unfrozen soil layers. The derivation and corresponding assumptions (i.e. linear soil temperature gradients (Aldrich, 1956)) do not reveal any major impediments to adapting this equation for a shorter time step. In addition, Fox (1992), Woo et al. (2004), and Carey and Woo (2005) have used a layered version of the Stefan Equation, which is similar to Eq. (21) to simulate daily frost depths with daily input data. Thus, the modified Berggren Equation is applied at a daily time scale and revised to become:

$$Z_d = \lambda[48\,(F - F_{threshold})\,\delta^{-1}\Omega_m]^{1/2}, \tag{22}$$

where Z_d is the depth of frozen ground (m). By using the difference between F and $F_{threshold}$, the degree-days from the current freezing/thawing period is utilized, which is similar to U in the original equation.

For the original modified Berggren Equation, λ can be estimated annually from Aldrich (1956) using U, the mean annual air temperature, and the soil water content ω (% of dry weight). Here, λ is calculated using daily differences between F and $F_{threshold}$, the mean annual air temperature, and daily ω values. Thus, soil moisture is included in the calculation of Z_d even though it is not included in the calculation of F. Furthermore, δ is estimated daily as:

$$\delta = \delta_f \rho\, \omega/100, \tag{23}$$

where δ_f is the latent heat of fusion of water (0.334 MJ kg^{-1} at 0°C) and ρ is the dry soil density. Ω_m is estimated as (Farouki, 1981; Johansen, 1977):

$$\Omega_m = (\Omega_{sat} - \Omega_{dry})\omega + \Omega_{dry}, \tag{24}$$

where Ω_{dry} and Ω_{sat} are the thermal conductivity of dry and saturated soil, respectively. Ω_{sat} is calculated as the geometric mean of the conductivities of the materials within the soil profile (Farouki, 1981; Johansen, 1977):

$$\Omega_{sat} = {\Omega_s}^{(1-n_{total})} {\Omega_{ice}}^{(n_{ice})} {\Omega_{water}}^{(n_{total}-n_{ice})}, \tag{25}$$

where Ω_s, Ω_{ice}, and Ω_{water} are the thermal conductivity of solids, ice, and water, respectively (Farouki, 1981). n_{total} is the porosity, and n_{ice} is:

$$n_{ice} = n_{total}\, Z_d/H, \tag{26}$$

where H (m) is the soil thickness.

3 Model Application

3.1 Study Area

The TI/CFGI and RTI/modCFGI models are tested at the W-3 sub-basin (Fig. 1) of the SREW. The study period is 1 Oct 2005 through 30 Sept 2010, which is water year (WY) 2006 through 2010. The SREW was founded in 1958 primarily for studies of snow accumulation, melt, and runoff (Anderson, 1973; Anderson, 1976; Dunne and Black, 1970a; Dunne and Black, 1970b; Dunne and Black, 1971; Shanley, 2000; Shanley and Chalmers, 1999). The W-3 sub-basin is located at 44° 29' N and 72° 09' W. Elevations range between 348 m and 697 m, and the area is approximately 8.5 km^2 (based on the National Elevation Dataset (Gesch et al., 2002)). The basin is primarily forested with deciduous (57.7%), evergreen (7.8%), and mixed (15.3%) trees (based on the 2006 National Land Cover Database (NLCD) (Fry et al., 2011)). Approximately 14.6% of the land cover is pasture/hay and cultivated crops. These open areas are typically below an elevation of 525 m, which is the approximate limit for viable agriculture (Shanley and Chalmers, 1999). The W-3 sub-basin is extensively gaged for both hydrometeorology and hydrology by the U.S. Geological Survey (USGS) and collaborators from federal agencies and universities. Additional basin information and data are provided by Anderson et al. (1979), Shanley et al. (1995), Shanley and Chalmers (1999) and the USGS website (https://nh.water.usgs.gov/project/sleepers/index.htm, accessed 7 November 2016).

Two snow sites and 35 frost sites within W-3 were monitored by the Vermont Field Office of the USGS. At the snow sites, SWE and snow depth were measured approximately weekly, and both sites are used in the present study. At the frost sites, snow depth and frost depth were measured periodically (between 0 and 14 measurements in a given winter). Frost depth was measured using gages described by Ricard et al. (1976) and Shanley and Chalmers (1999). The frost sites (labelled FS in Fig. 1) are clustered in six parts of the watershed. For this paper, one site from each cluster (FS4, FS11,

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FS21, FS24, FS30, and FS40) was selected for analysis. The selected sites are far enough apart to be relatively independent but still capture the variations in elevation and land cover classification within the watershed.

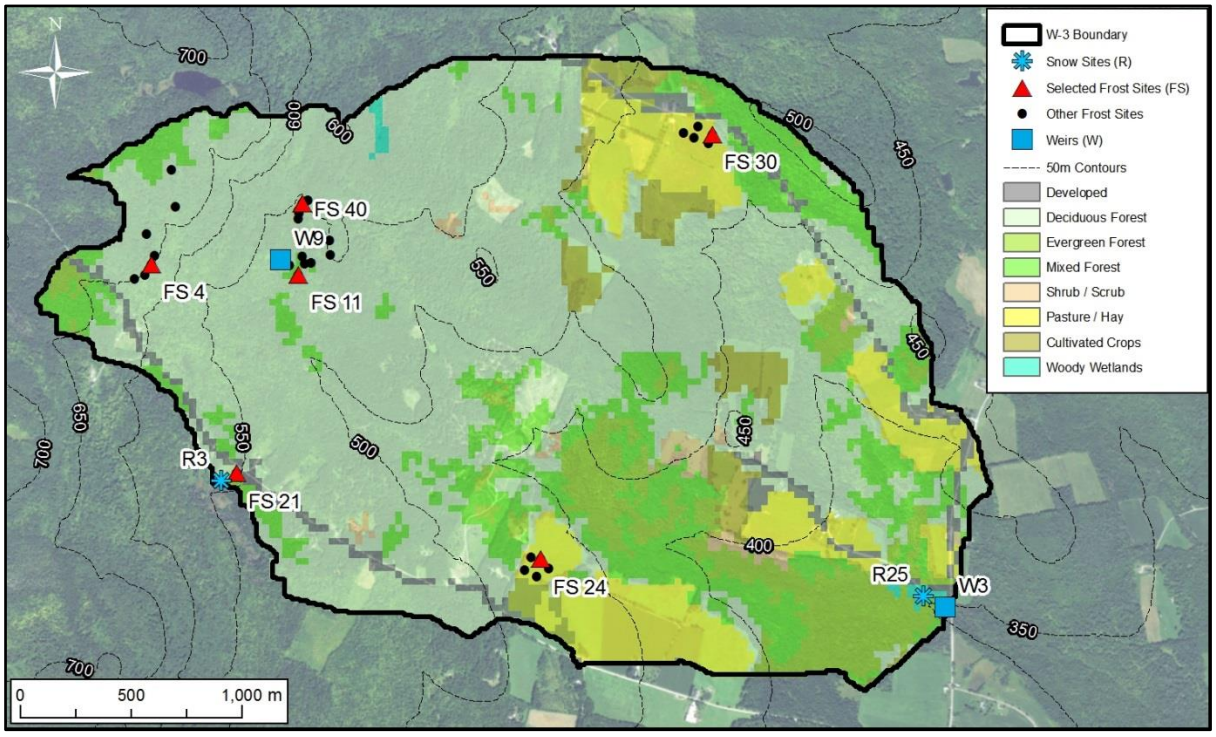

Figure 1. W-3 sub-basin in the Sleepers River Experimental Watershed. Sites used in this study are identified with red triangles and blue snowflakes. Basin delineation and elevation contours (m) are based on the 1/3-arc-second National Elevation Dataset, land cover classification based on the 2006 National Land Cover Database, and sources of the background imagery include ESRI, DigitalGlobe, Earthstar Geographics, CNES/Airbus DS, GeoEye, USDA FSA, USGS, Getmapping, Aerogrid, IGN, IGP, and the GIS User Community.

3.2 Model Forcing Data

Hourly precipitation and temperature data were obtained from the USGS. Precipitation was measured at the W9 weir and R3 snow site (Fig. 1). The USGS then creates a single spatially-averaged precipitation timeseries by weighting the measurements using the distribution of elevation (based on personal communication with Dr. James Shanley of the Vermont Field Office of the USGS on 14 November 2016). The W9 gage receives more weight because the watershed includes elevations both above and below this site. Hourly temperature was measured at the W9 site, which has an elevation of 520 m. All other hydrometeorological data were obtained from the National Centers for Environmental Information (NCEI, https://www.ncdc.noaa.gov/, accessed 7 November 2016). Specifically, hourly relative humidity, atmospheric pressure, and wind speed were obtained from the Fairbanks Museum in Saint Johnsbury, VT (11 km southeast of the basin). Hourly

cloud-cover classification data were obtained from both the Edward F. Knapp State Airport (44 km southwest of the basin) and the Morrisville-Stowe State Airport (36 km west of the basin). The cloud cover classifications (clear, few clouds, broken sky, etc.) were then converted to cloud cover percentages using the method from Follum et al. (2015). Relative humidity, atmospheric pressure, and wind speed data from the two airports were used to replace missing values within the Fairbanks Museum dataset. All of the hydrometeorological data described are required in long-term GSSHA simulations.

W-3 was delineated using the 1/3-arc-second (~9 m) National Elevation Dataset (Gesch et al., 2002). Land cover classifications were obtained from the 2006 National Land Cover Database (Fry et al., 2011), which has a 30-m resolution. The classifications of some grid cells were changed to match the land covers observed in the field. In particular, the grid cell containing R3 was changed from deciduous forest to pasture/hay, FS11 was changed from mixed forest to evergreen forest, and FS21 was changed from developed to mixed forest. Both FS24 and FS30 are classified as pasture/hay, where FS24 is a managed pasture and FS30 is an unmanaged pasture (personal communication with Ann Chalmers of the Vermont Field Office of the USGS on 15 November 2016). For example, during field observations in November 2016, FS24 had manure spread throughout the field, while FS30 was not fertilized. Soil classifications were obtained from the Digital General Soil Map of the Unites States (Soil Survey Staff, Natural Resources Conservation Service, United States Department of Agriculture, Web Soil Survey, available online at http://websoilsurvey.nrcs.usda.gov/. Accessed 10 August 2016). Almost the entire W-3 basin is classified as fine sandy loam. The Watershed Modeling System (Aquaveo, 2013) was used to develop the GSSHA model with a 30-m structured grid. This resolution is adequate to capture the spatial heterogeneity of the basin while remaining computationally efficient.

3.3 Parameter Estimation and Calibration

The Model-Independent Parameter Estimation and Uncertainly Analysis (PEST) method (Doherty et al., 1994) was used to calibrate the models. PEST is a nonlinear local search parameter estimator that calibrates numerous parameters simultaneously to produce the best fit between simulated results and observations. WY 2006 and 2007 were used as the calibration period. The TI and RTI snow models were calibrated first to minimize the sum of the squared residuals between simulated and observed snow depths at the 8 sites (6 frost sites and 2 snow sites).

Table 1 displays the allowable range, calibrated value, and sensitivity ranking for the calibrated snow parameters. Goodness of fit statistics as well as description of affects each parameter has on the snow simulations are described in the Results and Discussion section. The allowable ranges for A_{TIPM}, f_u, L_{hc}, $N_{mf,\max}$, M_f , $M_{f,\max}$, and $M_{f,\min}$ are based on physical limitations and typical ranges in the literature (Follum et al., 2014). L_{AI} can be estimated from seasonal and annual relationships to remotely-sensed normalized difference vegetation index (NDVI) values (Wang et al., 2005). However, snowpack affects the measurement of greenness in high latitude regions (Beck et al., 2006). Thus, L_{AI} was calibrated for each land cover classification using feasible ranges based on Liston and Elder (2006). Because the seasonal snowpack typically occurs after leaves have fallen, the minimum values of L_{AI} for each land cover type from Liston and Elder (2006) are considered. Separate L_{AI} and K_v values were calibrated for deciduous forest (including deciduous forest, woody

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wetlands, and mixed forest), evergreen forest, and pasture (including pasture/hay, cultivated crops, and shrub/scrub). T_{px} and T_{mbase} are not calibrated (both are 0°C) because the temperature data were post-processed by the Vermont USGS and are expected to be accurate. By comparing the temperature measurements at W9 and the Fairbanks Museum (elevation of ~212.4 m), $\emptyset$ was estimated at 6.6 °C km^{-1}. All snow density parameters are set constant based on Anderson (1973) and Anderson (2006).

The PEST results indicate that the TI model's snow depths are most sensitive to the melt factors ($M_{f,\min}$ and $M_{f,\max}$). For the RTI model, snow depths are most sensitive to vegetation related parameters including K_v for the open pasture sites and L_{AI} for the deciduous forest sites (Table 1). The calibrated pasture L_{AI} is at the top of the allowable range (0.250). Higher L_{AI} values produce decreased snowpack due to greater interception and $R_{LW\downarrow}$ values. Thus, the calibration is clearly attempting to match low snowpack levels at the pasture site. One potential cause for the low snowpack at this site is snow transport by wind in open areas, which is not included in either of the snow models considered here.

Table 1. Allowable ranges and calibrated values for the TI and RTI model parameters using PEST. Dashes indicate parameters that are not required in the associated model. The sensitivity ranking for each parameter is shown in parentheses.

Parameter	Units	Allowable Range	Calibrated Values TI	Calibrated Values RTI
$M_{f,\max}$	mm °C^{-1} (6 h)$^{-1}$	0.001-2.400	1.910 (2)	--
$M_{f,\min}$	mm °C^{-1} (6 h)$^{-1}$	0.001-1.600	1.600 (1)	--
M_f	mm °C^{-1} (6 h)$^{-1}$	0.001-2.400	--	1.080 (3)
$N_{mf,\max}$	mm °C^{-1} (6 h)$^{-1}$	0.001-2.400	0.001 (4)	0.130 (8)
f_u	mm mb^{-1} (6 h)$^{-1}$	0.001-1.000	0.500 (6)	0.500 (11)
A_{TIPM}	fraction	0.001-1.000	1.000 (3)	0.985 (7)
L_{hc}	fraction	0.001-0.100	0.001 (5)	0.001 (10)
$K_{v,\,deciduous}$	fraction	0.200-1.000	--	0.515 (4)
$K_{v,pasture}$	fraction	0.800-1.000	--	0.800 (1)
$K_{v,evergreen}$	fraction	0.200-0.800	--	0.305 (9)
$L_{AI,\,deciduous}$	m^2 m^{-2}	0.100-1.000	--	1.000 (2)
$L_{AI,\,pasture}$	m^2 m^{-2}	0.001-0.250	--	0.250 (5)
$L_{AI,\,evergreen}$	m^2 m^{-2}	1.000-4.000	--	1.000 (6)

The CFGI and modCFGI frozen ground models were calibrated to minimize the sum of squared residuals between the simulated and observed frost depths at the 6 frost sites. For purposes of comparison the modified Berggren equation was also added to the CFGI model to calculate frost depth. Table 2 displays the allowable range, calibrated value, and sensitivity

ranking of each calibrated frozen ground parameter. Goodness of fit statistics as well as description of affects each parameter has on the frost depth simulations are described in the Results and Discussion section. $F_{threshold}$ was calibrated for both the CFGI and modCFGI models with the upper range based on Molnau and Bissell (1983). Three K_{gc} values were calibrated for the modCFGI frozen ground model: one for the managed pasture site FS24 ($K_{gc,FS24}$), one for the unmanaged pasture site FS30 ($K_{gc,FS30}$), and one for the forested frozen ground sites ($K_{gc,forest}$).

Following Molnau and Bissell (1983), multiple combinations of A (0.8 and 0.97), and $K_{s,T_a<0°C}$ and $K_{s,T_a>0°C}$ (0.08, 0.2, and 0.5) values were tested with A = 0.97, $K_{s,T_a<0°C}$ = 0.08, and $K_{s,T_a>0°C}$ = 0.5 producing frost indices that best replicate the rise and fall of the frost depth as well as the timing of the peak frost depth. Depth of ground cover for each land cover type was obtained from field observations in November 2016. Specifically, $D_{gc} = 6$ cm for deciduous forest (fallen leaves), $D_{gc} = 2$ cm for evergreen forest (fallen leaves), $D_{gc} = 4$ cm for pasture (grass), and $D_{gc} = 0$ cm for all other land cover types.

The modified Berggren Equation requires soil moisture, which can be simulated using several methods in GSSHA (Downer and Ogden, 2006). To facilitate extension of these results to other hydrologic models, the commonly-used single-layer Green and Ampt infiltration model (Green and Ampt, 1911) with soil moisture redistribution between rainfall events (Ogden and Saghafian, 1997) is utilized to calculate infiltration. Soil moisture is tracked using a simple bucket approach, accounting for infiltration, evapotranspiration, and groundwater recharge as described in Downer (2007). The soil layer thickness (H) is set to 0.5 m for both the soil moisture calculations and frost depth equations. Soil infiltration parameters are set based on published values for the W-3 soil type (Downer and Ogden, 2006; Rawls et al., 1982; Rawls and Brakensiek, 1985; Rawls et al., 1983) and shown in Table 3. Evapotranspiration, which can reduce the soil moisture, is simulated using a Penman Monteith approach (Monteith, 1965; Monteith, 1981) with parameters estimated based on land cover (Downer and Ogden, 2006). The dry soil density (ρ = 1137 kg m^{-3}) and dry soil thermal conductivity (Ω_{dry} = 792 J m^{-1} h^{-1} °C^{-1}) are set based on measurements of fine sandy loam by Nikolaev et al. (2013).

For the CFGI model, the calibrated $F_{threshold}$ value (Table 2) is relatively close to the lower bound value of 56°C-days found in Molnau and Bissell (1983). For the modCFGI model, the calibrated $F_{threshold}$ value is at the lower bound. The $F_{threshold}$ value is expected to be lower for the modCFGI model than the CFGI model. The modCFGI model incorporates the insulation by ground cover directly using K_{gc} and D_{gc}, whereas the CFGI model can only account for those effects by adjusting the $F_{threshold}$ value. It is also worth noting that $K_{gc,FS30}$ has a very low value (minimum of allowable range), which suggests that insulation from grass in an unmanaged pasture is very small. This could be the result of snow falling within the grass of the unmanaged pasture, thus making any insulating contribution from the grass very small.

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Table 2. Allowable ranges and calibrated values for the CFGI and modCFGI model parameters using PEST. Dashes indicate parameters that are not required in the associated model. The sensitivity ranking for the modCFGI parameters are shown in parentheses.

Parameter	Units	Allowable Range	Calibrated Values CFGI	modCFGI
$F_{threshold}$	°C - days	10.00-83.00	40.52	10.00 (2)
K_{gc}	cm	0.001-1.000	--	0.719 (1)
$K_{gc,FS24}$	cm	0.001-1.000	--	1.000 (3)
$K_{gc,FS30}$	cm	0.001-1.000	--	0.001 (4)

Table 3. Values of soil parameters used to calculate soil moisture in the single-layer Green and Ampt infiltration model.

Parameter	Units	Value
saturated hydraulic conductivity	cm h^{-1}	2.040
effective porosity	cm^3 cm^{-3}	0.407
residual saturation	cm^3 cm^{-3}	0.038
field capacity	cm^3 cm^{-3}	0.166
wilting point	cm^3 cm^{-3}	0.075
capillary head	cm	8.570
pore distribution arithmetic mean	cm cm^{-1}	0.466

4 Results and Discussion

4.1 Snow Depth and SWE (TI vs RTI)

Figure 2 shows maps of simulated snow depth on 23 February 2007 from the TI and RTI snow models. The spatial variability in the TI snowpack is entirely based on elevation (due to the inference of local air temperature from elevation). Higher elevations have deeper snowpack due to lower air temperatures. The RTI snowpack also varies with elevation but also includes variation due to land cover. In particular, pasture areas have deeper snowpack than surrounding areas due to lower interception/sublimation rates and lower $R_{LW\downarrow}$. North-facing slopes also have more snow than south-facing slopes due to lower $R_{SW,net}$. Although no maps of observed snow depth are available for comparison, large-scale distributions of snowpack are known to be controlled by elevation, land cover, and slope/aspect (Fassnacht et al., 2017; Jost et al., 2007), which is more consistent with the RTI model.

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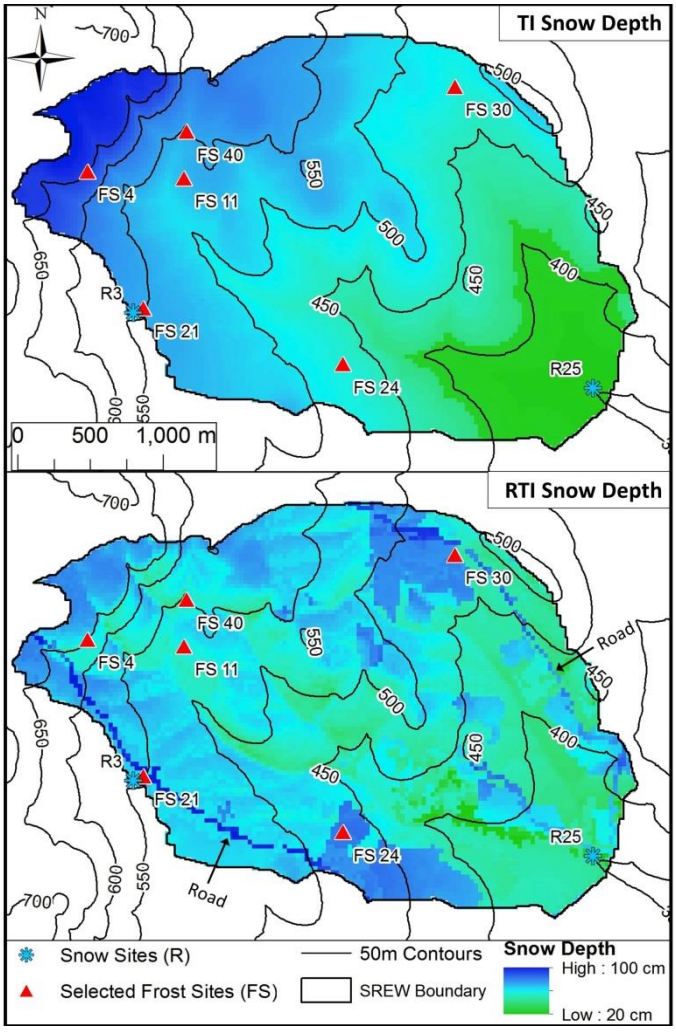

Figure 2. Simulated maps of snow depth (TI and RTI models) within the W-3 watershed for 23 February 2007. No observed maps of snow depth are available, but the map shows the differences between the temperature-based (TI) model and the modified (RTI) model.

Figure 3 shows the snow depths from the TI and RTI models at all 8 test locations and compares them to the observations. Root mean squared error (RMSE) and Nash-Sutcliffe Efficiency (NSE) are shown in Table 4 for the calibration period (WY 2006-2007), validation period (WY 2008-2010), and complete period (WY 2006-2010). The TI and RTI models track closely together at the 8 test locations despite differences in the snow depth shown in Fig. 2. The test sites are typically located on shallow slopes where terrain has limited influence on the energy available to melt the snowpack. Therefore, differences between the TI and RTI snowpack at the test sites are more related to the effects of vegetation. In Fig. 3, the two models tend to produce similar results at the pasture sites (see FS24, FS30, and R3), but they tend to differ at the deciduous forest sites (see FS4, FS21, and FS40). At the deciduous sites the RTI model often produces lower peak

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snowpack due to the interception and sublimation of snowpack by the canopy (which are not included in the TI model). Overall, the RTI model performs better than the TI model (lower RMSE values and higher NSE values in Table 4), suggesting that inclusion of sublimation/interception and use of T_{rad} improve the spatial representation of the snowpack. The observed snow depth is relatively low in WY2008 and 2009 at two of the pasture sites (FS24 and FS30) compared to the other sites. These two years of small snow depth are not captured within either model. The R3 site is also classified as pasture yet has a higher snowpack in WY2008 and 2009. The higher snowpack at this pasture site may be explained by the proximity of R3 to forested areas, which may reduce the wind and help preserve the snowpack. Neither model considers wind effects.

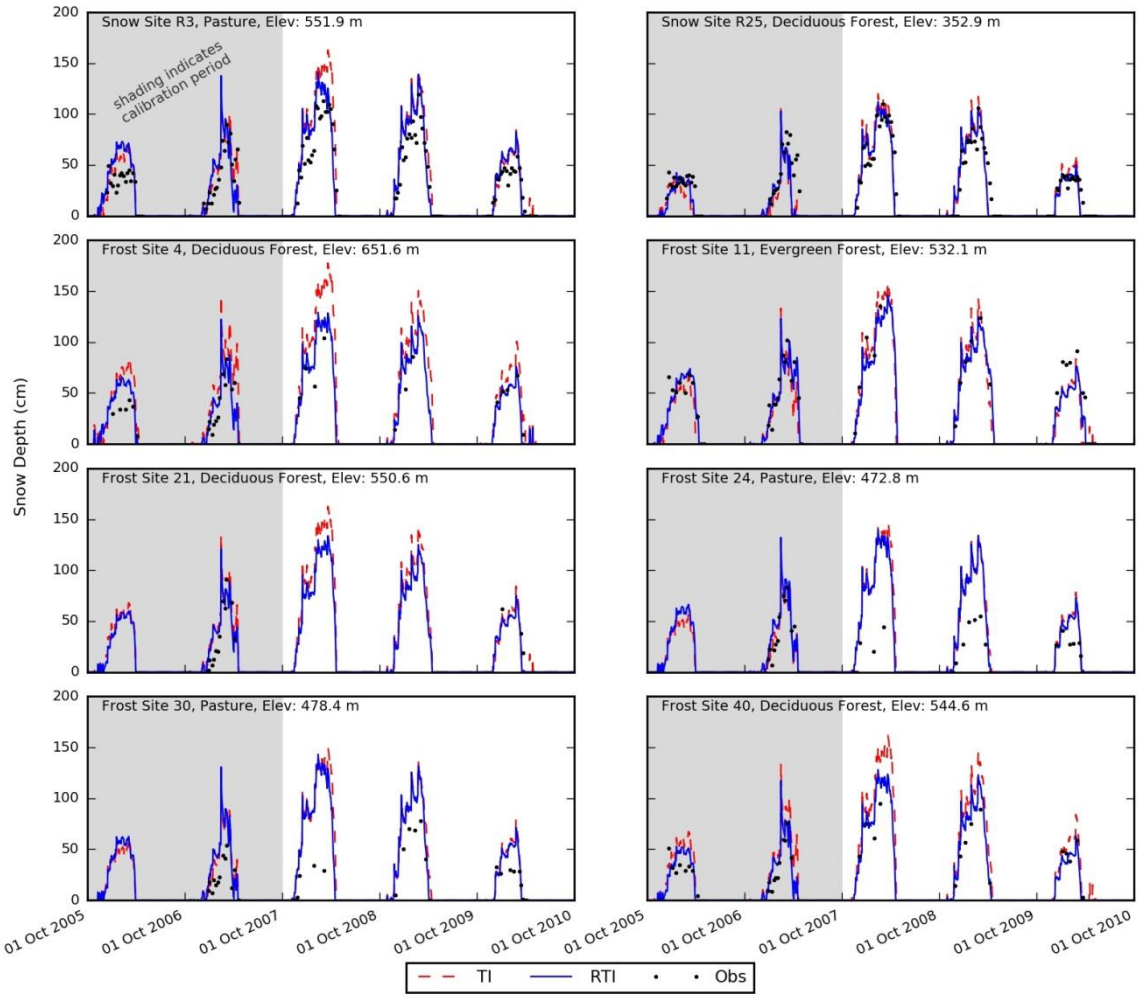

Figure 3. TI and RTI simulated snow depth at all eight test sites within the W-3 watershed.

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Table 4. Statistics for TI and RTI snow depth values at all 8 test sites, and statistics for TI and RTI SWE values at the R3 and R25 snow test sites. Values are shown for calibration period (WY 2006-2007), validation period (WY 2008-2010), and overall (WY 2006-2010). RMSE values closer to zero and NSE values closer to one indicate better fit.

	Site	Land Cover	Snow Model	Calibration		Validation		Overall	
				RMSE (cm)	NSE	RMSE (cm)	NSE	RMSE (cm)	NSE
Snow Depth	R3	Pasture	TI	14.0	0.60	23.6	0.58	20.8	0.59
			RTI	20.7	0.13	17.7	0.76	18.8	0.67
	R25	Deciduous Forest	TI	19.9	0.08	13.6	0.84	15.9	0.73
			RTI	19.5	0.12	12.7	0.86	15.2	0.76
	FS4	Deciduous Forest	TI	25.6	-0.40	28.7	-0.24	26.8	-0.14
			RTI	14.8	0.53	10.7	0.83	13.3	0.72
	FS11	Evergreen Forest	TI	16.2	0.65	18.8	0.79	17.5	0.75
			RTI	11.9	0.81	16.2	0.85	14.2	0.84
	FS21	Deciduous Forest	TI	17.5	0.62	13.5	0.41	16.8	0.60
			RTI	16.1	0.68	14.3	0.34	15.7	0.65
	FS24	Pasture	TI	8.1	0.90	38.6	-3.75	30.5	-0.88
			RTI	13.6	0.72	36.4	-3.22	29.6	-0.77
	FS30	Pasture	TI	19.8	-0.69	33.6	-1.01	29.1	-0.93
			RTI	24.5	-1.59	30.4	-0.65	28.3	-0.82
	FS40	Deciduous Forest	TI	18.8	-0.06	23.7	0.15	21.1	0.21
			RTI	12.7	0.52	10.8	0.82	11.9	0.75
SWE	R3	Pasture	TI	3.6	0.69	7.2	0.59	6.2	0.61
			RTI	7.1	-0.16	6.1	0.71	6.4	0.59
	R25	Deciduous Forest	TI	7.5	-0.47	4.7	0.77	5.8	0.58
			RTI	7.1	-0.32	4.6	0.79	5.5	0.62

Figure 4 shows the simulated (both TI and RTI models) and observed SWE values, and Table 4 shows the associated performance metrics at the R3 and R25 snow sites. The TI and RTI models are only calibrated to snow depth, but SWE is calculated first and then combined with snow density to determine snow depth. Both models use the same method to calculate snow density. Both models exhibit similar behaviour and performance at the two sites, which is consistent with their similar snow depths discussed earlier (Fig. 3 and Table 4). The TI model's overestimation of SWE at R3 during WY 2008 is associated with a similar overestimation of snow depth. Overall, these suggest that the snow density equations used within GSSHA are relatively accurate at the W-3 watershed. Thus, accurate estimates of snow depth typically correspond to accurate estimates of SWE as well.

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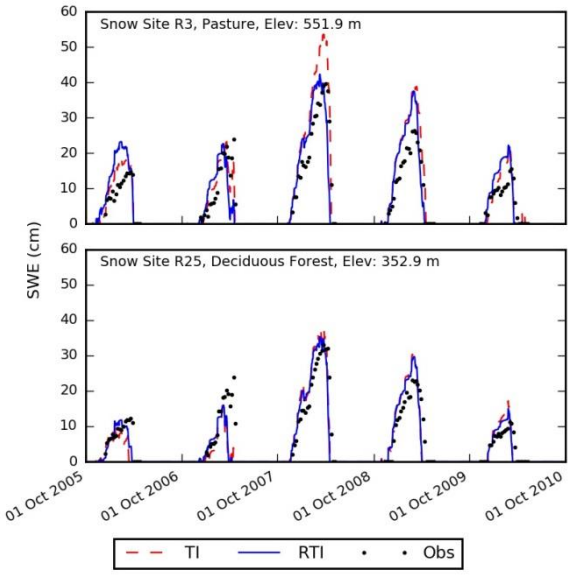

Figure 4. TI and RTI simulated SWE at R3 and R25 snow sites within the W-3 watershed.

4.2 Frost Depth (CFGI vs modCFGI)

Figure 5 shows simulated frost depth maps for 23 February 2007 using the CFGI and modCFGI models (no maps of observed frost depths are available for comparison). In the CFGI model, the frost depths mainly depend on elevation. Colder temperatures at higher elevations generally result in greater snowpack, which insulates the ground and produces smaller frost depths. However, at the beginning of the snow season when the snowpack is shallow, low temperatures at high elevations create deep frost in the higher elevations of the watershed. Later, deeper snowpack at high elevations insulate the ground, while the frost depth increases at lower elevations. This reversal in the elevation dependence can produce an inversion (localized minima in frost depth), as seen between the 400 and 450 m contour lines in Fig. 5. The modCFGI frost depth also has some elevation dependence, but it shows much more spatial variation. This variation is partly due to the use of T_{rad} and the increased heterogeneity in the snow depth. The effect of snowpack can be seen by comparing hillslopes with the same land cover but different orientations, such as along the 500 m contour south of FS11. Lower T_{rad} values on northeast-facing slopes result in deeper snowpack than the southwest-facing slopes (Fig. 2). This deeper snowpack produces shallower frost depths on the northeast-facing slopes due to insulation by the snow. However, the spatial pattern of frost depth is more heavily affected by the land cover. Land cover's impact largely occurs through the associated ground cover. This effect can be seen by comparing the deep frost at the unmanaged pasture (near FS30) with the shallower frost depth at the deciduous forest areas near FS4, FS21, and FS40. The low ground cover reduction coefficient at the unmanaged pasture ($K_{gc,FS30}$) reduces the insulation from the ground cover, creating deeper frost compared to the deciduous forest areas. The

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larger than expected role of ground cover in the modCFGI model may occur because ground cover is present during the initiation, deepening, and decrease of frost depth, while the snowpack is much more variable throughout the season.

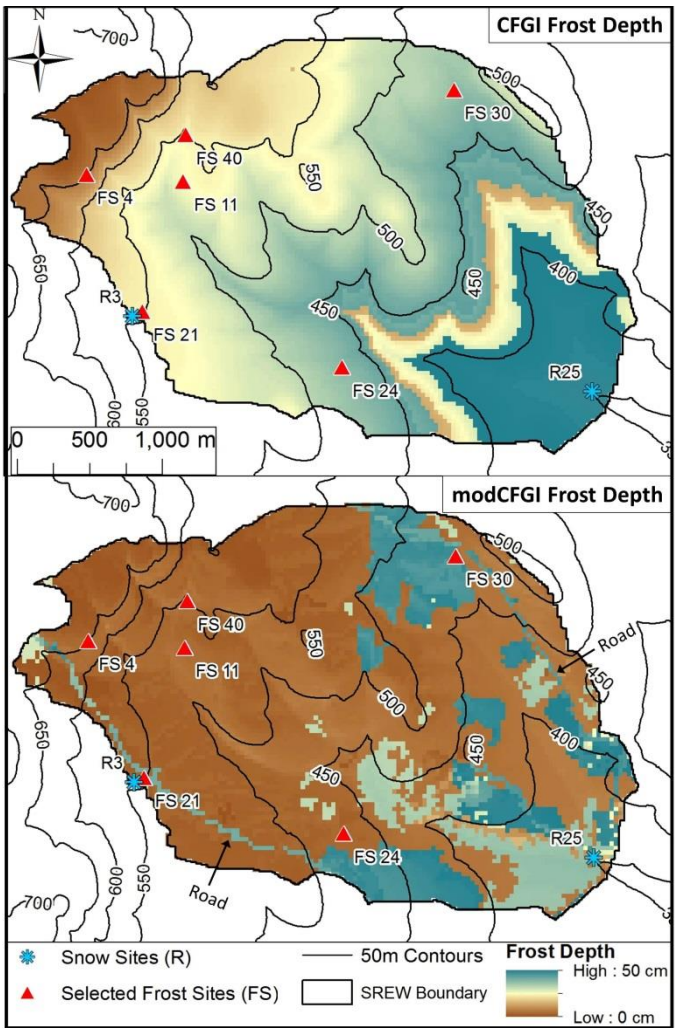

Figure 5. Simulated maps of frost depth (CFGI and modCFGI models) within the W-3 watershed for 23 February 2007. No observed maps of frost depth are available, but the map shows the differences between the temperature-based (CFGI) model and the modified (modCFGI) model.

Figure 6 shows the frost depths from the CFGI and modCFGI models along with the frost depth observations. The RMSE and NSE values during the calibration, validation, and overall periods are shown in Table 5. The simulated frost depth remains more constant amongst the sites when using the CFGI model, which produces similar maximum frost depths for a given year independent of the land cover. The modCFGI results deviate considerably from the CFGI results, producing greater frost depths at the unmanaged pasture (FS30) and evergreen (FS11) sites and smaller frost depths at the deciduous

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(FS4, FS21, and FS40) and managed pasture (FS24) sites. These simulated differences between the sites are consistent with the observations. The decreased frost depth in the deciduous forest and managed pasture result from their high measured litter depth (D_{gc} = 6 cm) and high reduction coefficient ($K_{gc,FS24}$ = 1.0 cm^{-1}), respectively. The two pasture sites (FS24 and FS30) differ considerably in the observed frost depth with FS30 consistently having deeper frost. This difference likely occurs because FS24 is managed and FS30 is not. With the exception of the validation period at FS30, the modCFGI model performs better (lower RMSE and higher NSE values) than the CFGI model. The difference in performance is most pronounced at the deciduous sites (FS4, FS21, and FS40) where the average overall NSE value is -7.26 for the CFGI model and -0.10 for the modCFGI model.

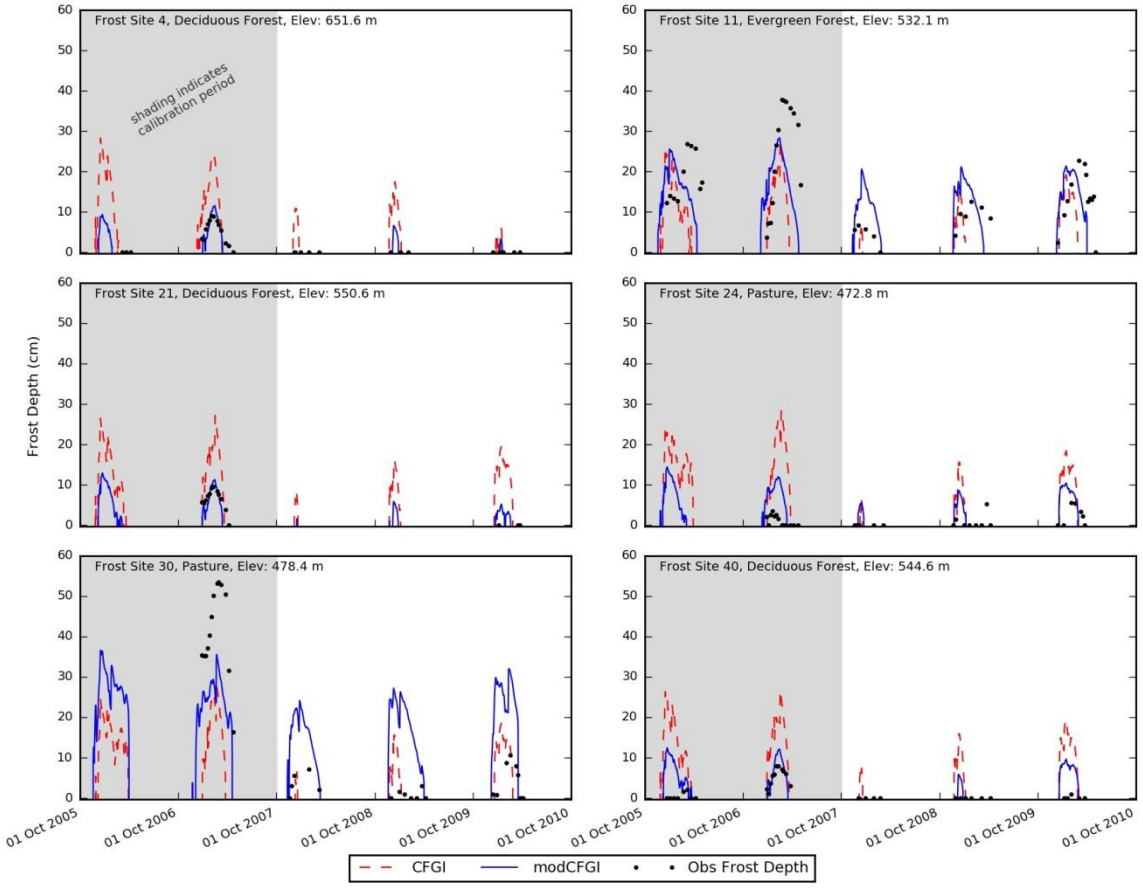

Figure 6. Observed frost depth compared against simulated (CFGI and modCFGI) frost depth at all 6 selected frozen ground test sites within the W-3 watershed.

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Table 5. Statistics for CFGI and modCFGI frost depth at all 6 frost sites. Values are shown for calibration period (WY 2006-2007), validation period (WY 2008-2010), and overall (WY 2006-2010). RMSE values closer to zero and NSE values closer to one indicate better fit. No frost was present at FS4 and FS21 during the validation period, resulting in an inability to calculate NSE. Statistics for a recalibrated modCFGI model without ground cover (labelled as "modCFGI no gc") are also shown.

Site	Land Cover	Frost Depth Model	Calibration		Validation		Overall	
			RMSE (cm)	NSE	RMSE (cm)	NSE	RMSE (cm)	NSE
FS4	Deciduous Forest	CFGI	7.5	-4.13	3.8	NA	6.2	-2.53
		modCFGI	1.5	0.79	0.0	NA	1.1	0.88
		modCFGI no gc	15.0	-19.5	13.9	NA	14.6	-18.5
FS11	Evergreen Forest	CFGI	17.2	-1.73	10.8	-2.07	14.4	-0.93
		modCFGI	13.9	-0.78	8.8	-1.06	11.7	-0.27
		modCFGI no gc	15.1	-1.10	10.6	-1.98	13.1	-0.6
FS21	Deciduous Forest	CFGI	8.6	-10.8	7.1	NA	8.3	-4.86
		modCFGI	2.4	0.10	2.0	NA	2.3	0.55
		modCFGI no gc	14.9	-34.6	11.4	NA	14.3	-16.3
FS24	Pasture	CFGI	16.0	-156.3	4.3	-3.69	10.6	-35.6
		modCFGI	6.6	-25.7	3.1	-1.41	4.8	-6.49
		modCFGI no gc	19.4	-229.9	13.0	-41.4	15.8	-80.8
FS30	Pasture	CFGI	28.7	-6.41	4.6	-0.95	18.1	0.18
		modCFGI	17.6	-1.78	14.8	-18.7	15.9	0.36
		modCFGI no gc	23.8	-4.12	10.5	-9.02	16.9	0.28
FS40	Deciduous Forest	CFGI	12.4	-16.9	7.6	-927.4	10.6	-14.4
		modCFGI	4.9	-1.78	3.8	-234.0	4.4	-1.73
		modCFGI no gc	21.0	-50.5	16.1	-4145	19.0	-48.8

In hydrologic models, capturing the presence of frozen ground may be more important than accurately simulating the frost depth because any frozen ground has the potential to impede infiltration and produce flooding. Therefore, the ability of the CFGI and modCFGI models to accurately capture the presence of frozen ground is evaluated. Whenever frost observations are available, the simulated frost depths are categorized as: True Positive (both simulated and observed data show frost), True Negative (both simulated and observed data show no frost), False Positive (simulated data shows frost but observed data shows no frost), or False Negative (simulated data shows no frost but observed data shows frost). Table 6 shows the number of observations in each category for each test site. The table also shows the model accuracy, which is calculated as the percent of the observations that are correctly classified (True Positive or True Negative). The CFGI and modCFGI models perform similarly in capturing True Positives in FS4, FS21, FS24, and FS40, while modCFGI has more True Positives at FS11 and FS30. The CFGI model also tends to have more False Negatives. Together the lower True Positives and higher False Negatives indicate that the CFGI model tends to underestimate the presence of frozen ground. Overall, both the CFGI and modCFGI models capture most of the frozen ground events, with the modCFGI model

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performing better than the CFGI model at 4 sites and worse at 2 sites. The average accuracy of the modCFGI model is 12.5% higher than the CFGI model, with the largest increase in accuracy at FS11 (29.8%).

Table 6. Number of True Positive (both simulated and observed data show frost depth), True Negative (both simulated and observed data show no frost depth), False Positive (simulated data shows frost depth but observed data does not), and False Negative (simulated data shows no frost depth but observed data shows frost depth) occurrences during the entire test period. The Accuracy is the sum of the True Positive and True Negative divided by the total number of observations.

Site	Land Cover	Elevation (m)	Model	True Positive	True Negative	False Positive	False Negative	Accuracy (%)
FS4	Deciduous Forest	651.6	CFGI	10	12	4	2	78.6%
			modCFGI	9	16	0	3	89.3%
FS11	Evergreen Forest	532.1	CFGI	22	2	0	23	51.1%
			modCFGI	37	1	1	8	80.9%
FS21	Deciduous Forest	550.6	CFGI	10	3	1	1	86.7%
			modCFGI	9	3	1	2	80.0%
FS24	Pasture	472.8	CFGI	8	14	7	4	66.7%
			modCFGI	9	14	7	3	69.7%
FS30	Pasture	478.4	CFGI	14	7	1	12	61.8%
			modCFGI	25	4	4	1	85.3%
FS40	Deciduous Forest	544.6	CFGI	13	11	9	1	70.6%
			modCFGI	12	11	9	2	67.6%
Total			CFGI	77	49	22	43	66.0%
			modCFGI	101	49	22	19	78.5%

A simple test is employed to explore the modification that contributes most to the increased accuracy of the modCFGI model. This test removes ground cover from the modCFGI model, recalibrates, and then compares the results to observations. When ground cover is removed, the calibrated $F_{threshold}$ value is 26.96 °C-days, which is closer to that of the CFGI model (40.52 °C-days) than the complete modCFGI model (10.00 °C-days). This change indicates that ground cover has a large impact on the appropriate value of this threshold. Figure 7 shows the simulated frost depths using the modCFGI model with and without ground cover for each test site. Performance metrics for the modCFGI model with and without ground cover are shown in Table 5. Variability in frost depth between the sites is diminished when ground cover is removed, leading to large errors between simulated and observed frost depth. When ground cover is removed, the frost depth results decrease in accuracy (higher RMSE values and lower NSE values) compared to the complete modCFGI model. The only exception is the validation period at FS30, which is also the only site and time period when the CFGI model outperforms the full modCFGI model. This result suggests that the inclusion of ground cover is an important reason why the modCFGI model outperforms the CFGI model.

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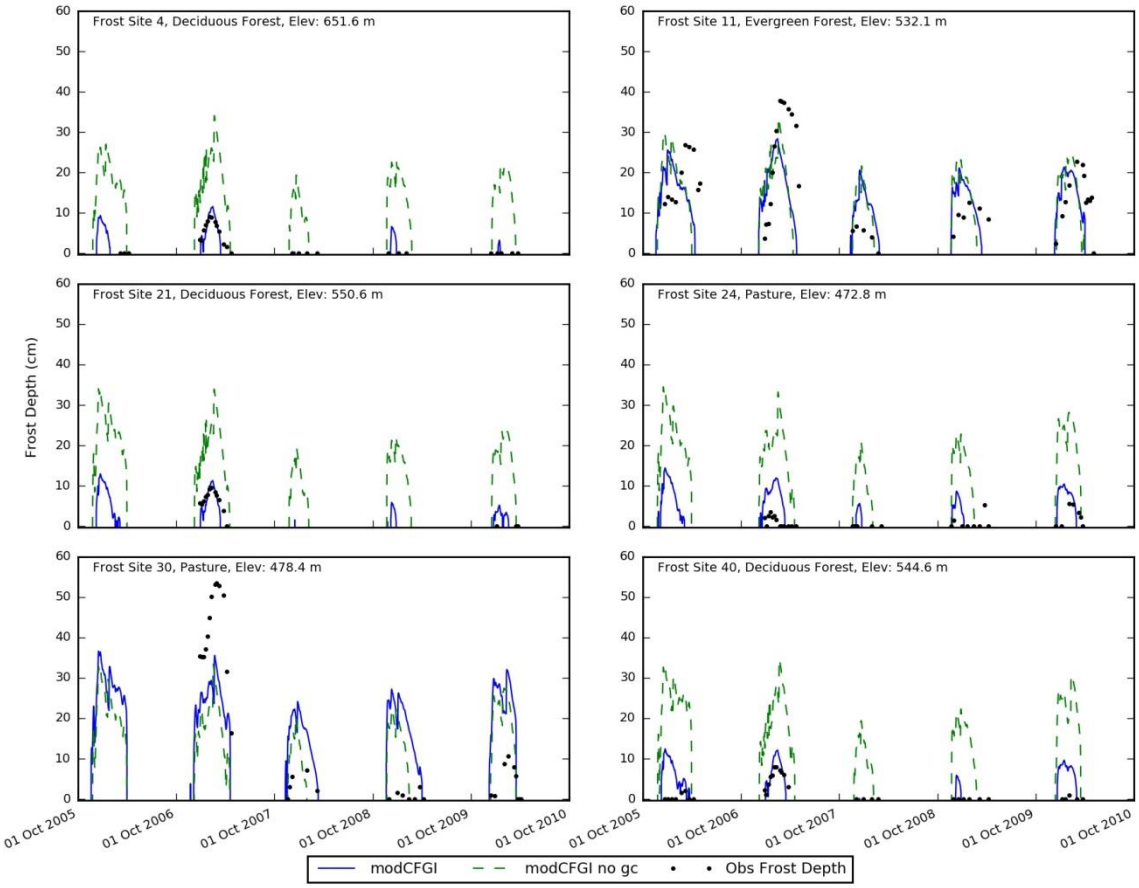

Figure 7. Observed frost depth compared against simulated (modCFGI with and without ground cover included) frost depth at all 6 selected frozen ground test sites within the W-3 watershed. The modCFGI model without ground cover is labelled as "modCFGI no gc".

The sensitivity of the modCFGI results to soil moisture is also examined. Soil moisture does not affect the calculation of F, but it is included within the modified Berggren Equation (Eq. (21) and (22)) in the calculation of δ (Eq. 23) and Ω_m (Eq. 24). Soil moisture was simulated using a single layer Green and Ampt approach. However, no soil moisture measurements are available at any of the test sites to evaluate the accuracy of the simulated values. Sensitivity of the modCFGI model to volumetric soil moisture is tested by artificially setting the soil moisture to either the residual soil moisture (θ_{low}) or the effective porosity (θ_{high}), which are the lower and upper bounds for soil moisture values within the model. Figure 8 shows the modelled frost depths from the modCFGI model using θ_{low}, θ_{high}, and the soil moisture from the Green and Ampt approach (θ_{sim}, which is identical to modCFGI in Fig. 6 and Fig. 7). Also shown are the observed frost depths for reference only. The frost depth from the θ_{sim} case is similar to the frost depth from the θ_{high} because the simulated soil moisture is usually close to the effective porosity. Frost depth increases when θ_{low} is used, which coincides with other studies (Fox, 1992; Willis et al., 1961). The timing of the frozen ground (when it begins and ends) is identical in

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all three of the simulations. The consistent timing occurs because soil moisture is not used to calculate F and the same $F_{threshold}$ (which controls when frozen ground begins) was used for all three simulations. This result highlights a deficiency in the modelling framework. Specifically, soil moisture should be considered for determining the initiation of frozen ground because wet soils have a higher specific heat capacity and require more energy to cool and freeze the soil (Kurganova et al., 2007).

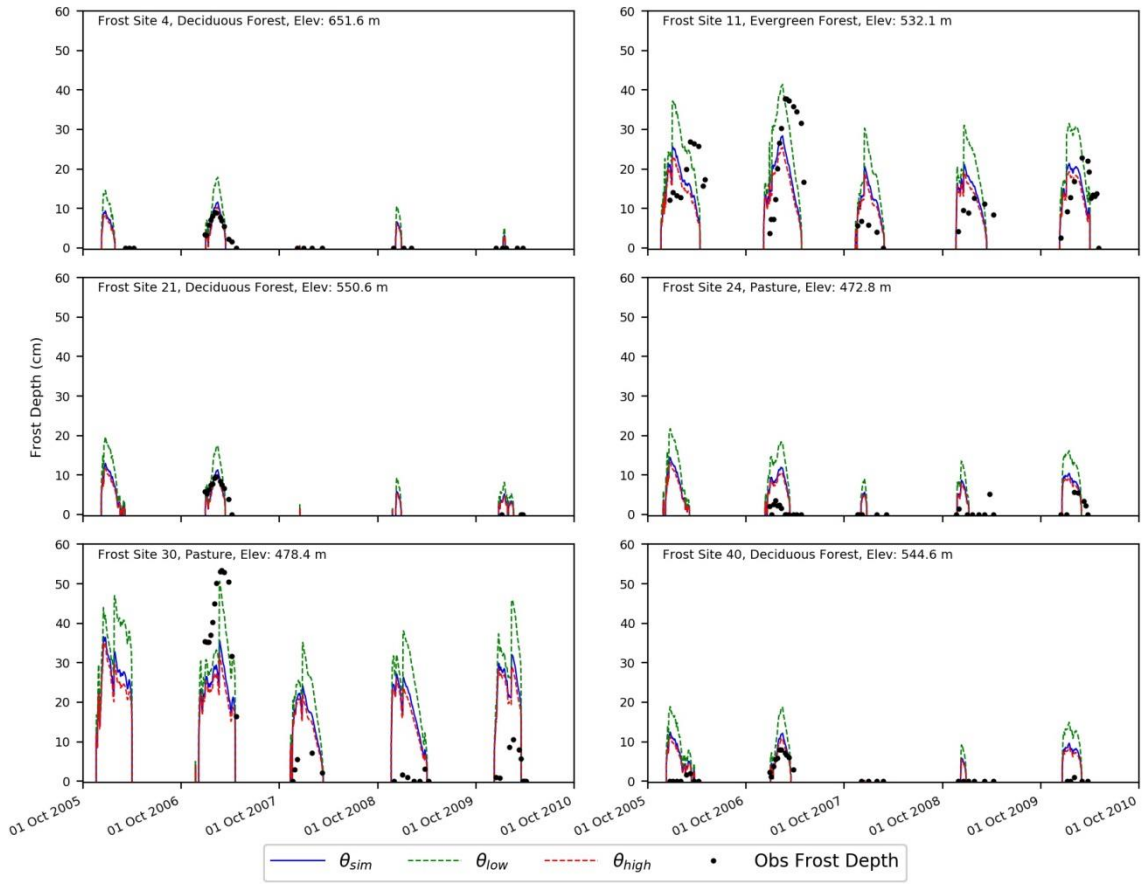

Figure 8. Simulated frost depths from the modCFGI model using simulated soil moisture (θ_{sim}), a constant high soil moisture (θ_{high}), and a constant low soil moisture (θ_{low}) at all 6 selected frozen ground test sites within the W-3 watershed.

5 Conclusions

The main purpose of this paper was to better estimate the spatial pattern of frozen ground for distributed watershed modelling by modifying an existing degree-day frozen ground model (CFGI), which uses a frost index value to determine whether the ground is frozen or not. The modifications to the CFGI model include: 1) use of a radiation-derived temperature index (RTI) snow model instead of a standard temperature-index (TI) snow model, 2) use of a radiation-derived proxy temperature (T_{rad}) instead of air temperature (T_a) in the calculation of the frost index, 3) inclusion of ground cover (litter,

debris, grass, etc.) as an insulator of the ground from air temperatures, and 4) implementation of a version of the modified Berggren Equation to calculate frost depths based on the frost index values. The CFGI and modCFGI models were tested using the GSSHA hydrologic model over a 5-year period within the W-3 watershed, which is part of Sleepers River Experimental Watershed in Vermont. The model results were compared against snow depth at eight sites, snow water equivalent at two sites, and frost depth at six sites. The primary conclusions of the paper are as follows:

1.) The RTI snow model produces much more complex spatial patterns of snow depth than the TI snow model for the W-3 watershed. The TI model, which is based on SNOW-17 (Anderson, 2006), only produces spatial variation using elevation. The RTI model accounts for elevation, hillslope orientation, canopy shading, and longwave radiation from the canopy through the use of the radiation-derived proxy temperature. It also includes snow interception/sublimation by the canopy. Thus, its snow depths exhibit spatial heterogeneity based on elevation, slope/aspect, and land cover, all of which are known to affect the largescale distribution of observed snow depths (Fassnacht et al., 2017; Jost et al., 2007).

2.) The RTI model produces more accurate estimates of snow depth than the TI model for the eight snow depth sites at W-3. Follum et al. (2015) previously compared the TI and RTI models for sites in Colorado that had the same land cover but diverse topographic attributes (e.g., aspect, slope, and topographic shading). Here, the test sites primarily differ in their land cover, which include pasture, deciduous forest, and evergreen forest. The RTI model more accurately simulates the snow depth at all eight sites, but both models perform poorly at two of the three pasture sites. This poor performance may occur because neither model accounts for wind effects.

3.) The modCFGI frost model produces more complex spatial patterns of frost depth than the CFGI frost model for the W-3 watershed. The CFGI model uses elevation to infer the spatial variation of air temperature. It also uses the TI model for snow depth, which also depends on elevation. Thus, the simulated frost depths at W-3 primarily reflect the watershed elevations. In contrast, the modCFGI model uses the radiation-derived proxy temperature to infer the energy available to heat the ground and the RTI model to simulate snow depth. Furthermore, it accounts for the insulating effects of ground cover (in addition to snowpack), which also depends on the land cover. Thus, the frost depths simulated by the modCFGI model at W-3 depend on the local elevation, hillslope orientation, and land cover, all of which are known to affect the distribution of frozen ground (Fox, 1992; MacKinney, 1929; Wilcox et al., 1997; Willis et al., 1961).

4.) The modCFGI model produces more accurate frost depths than the CFGI for all six test sites in the W-3 watershed. Overall, the modCFGI model more accurately captures the inter-annual variability in frost depth at a given site and variability of frost depth between sites. Although both the CFGI and modCFGI capture the majority of frozen ground events observed, the modCFGI model has 12.5% better accuracy in capturing the presence of frozen ground, which is expected to be important for capturing runoff that is produced by frozen ground.

5.) A key reason for the difference in performance between the two frost models is that the modCFGI model includes the insulation of the ground by ground cover while the CFGI model does not. When ground cover is removed from

the modCFGI model, its results for W-3 are less accurate and the variability in simulated frost depth between the sites is limited. Ground cover is likely important in this watershed because it is relatively thick and is also present at all stages of the winter while snowpack is not.

Four main avenues are available for future research. First, the modCFGI model should be generalized to include the effects of wind (as it relates to the snowpack) and more completely consider the role of soil moisture. Soil moisture is not considered when calculating the frost index, so it does not impact the initiation or duration of frozen ground. This limitation results from using a degree-day approach and may be important in some cases (Kurganova et al., 2007; Willis et al., 1961). Second, the modCFGI model should be tested further. Additional testing should consider other areas where snow and frozen ground are known to affect runoff, such as the Upper Midwest region of the United States. Additional testing should also better characterize the insulation properties of ground cover under different management scenarios. Third, future research should also determine the effects of spatial heterogeneity of snow and frost depth on runoff and streamflow at both the local and watershed scales. Finally, although this paper focuses on the simulation of frost depth in the context of watershed modelling, the methods described could also be used for agriculture, overland mobility modelling, and infrastructure where snow and frost depth are major concerns.

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