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Snowpack affects soil microclimate throughout the year

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Abstract
Variations in snow depth have complex effects on soil microclimate. Snow insulates soil and thus regulates, along with air temperature, the nature, and extent of soil freezing. There is great uncertainty about the main drivers of soil freezing, which have important effects on ecosystem carbon and nitrogen cycling processes and might change as climate warms and snowfall decreases as part of climate change. Here, we utilitze sites from a variety of elevations and aspects within the northern hardwood forest at the Hubbard Brook Experimental Forest (New Hampshire, USA) to investigate relationships between seasonal snowpack, soil freezing, and soil microclimate across this gradient using 8 years of bi-weekly snowpack and soil frost-depth measurements, and continuous soil climate monitoring. We utilize a time-integrated snowpack descriptor and find that snowpacks with lower seasonal snow water equivalents result in more soil temperature variation and deeper soil frost but have no effect on variation in soil moisture. Seasonal snow water equivalent of the snowpack influences the date of rapid soil warming in the spring, which in turn influences both summer soil moisture and an index of annual cumulative soil heat. These results show that snowpack dynamics, which are highly sensitive to changes in climate, have wide-ranging effects on soil microclimate year-round and thus could have important implications for ecosystem carbon and nitrogen cycling processes.

Keywords Snowpack · Soil microclimate · Soil frost · Winter climate change

1 Introduction

Soil microclimate is one of the most important controls on terrestrial ecosystem processes. Climate models forecast continued increases in both air temperatures and precipitation in the northeastern USA (Campbell et al. 2010, Hayhoe et al. 2008, Ahmed et al. 2013), and such increases are likely to influence soil microclimate through changes in the soil physical

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Extended author information available on the last page of the article environment and plant-soil interactions (Kardol et al. 2010). However, the scope and magnitude of these changes are difficult to predict and might be associated with a wide suite of changes in ecosystem processes. For example, soil metabolic rates, as indicated by soil respiration, are tightly correlated to soil temperature (e.g., Davidson et al. 1998, Fahey et al. 2005a). Similarly, nitrogen (N) cycling is sensitive to soil temperature, moisture, and their interactions (e.g., Hong et al. 2005, Groffman et al. 2009).

Snowpack is a key mediator of how soil microclimate will respond to climate change. A developed snowpack insulates soil and can prevent soils from freezing even when air temperatures are below zero (Brooks et al. 1997, Isard and Schaetzl 1998, Groffman et al. 2001, Decker et al. 2003, Liptzin et al. 2009). In contrast, freezing temperatures on a thin snowpack or bare soil can create soil frost, which can have a host of ecological effects. Nitrogen cycling (Blankinship and Hart 2012, Duran et al. 2014, 2016, 2017, Morse et al. 2015, Campbell et al. 2014, Brooks and Williams 1999), root dynamics (Kreyling et al. 2012, Repo et al. 2014, Tierney et al. 2001, Cleavitt et al. 2008, Sorenson et al. 2016), tree physiology (Comerford et al. 2013), tree growth (Reinmann et al. 2018), nutrient export (Matzner and Borken 2008, Austnes et al. 2008, Fitzhugh et al. 2001, Judd et al. 2007), soil respiration (Contosta et al. 2016a, Liptzin et al. 2009), and the activity and diversity of soil invertebrates (Sulkava and Huhta 2003, Christenson et al. 2017, Templer et al. 2012) are all affected when soils freeze. Warming air temperatures will have the potentially contrasting effects of reducing the severity of cold temperatures above soils, while also reducing the amount of snow available to buffer the soils from those temperatures, raising the question of whether we will have "colder soils in a warmer world" (Groffman et al. 2001) due to the loss of insulating snowpack. Current models reflect this complexity, suggesting increases in soil freeze-thaw events (Henry 2008, Campbell et al. 2010), but decreases in total frost duration and little if any change in maximum frost depth (Campbell et al. 2010). A recent 100-year regional analysis detected minimal changes in soil freezing potential despite clear reductions in snowpack (Contosta et al. 2019).

Snowpack characteristics have already changed in response to higher winter average air temperatures over the past few decades. Recent analyses have quantified declines throughout eastern North America, affecting characteristics of both winter and spring (Contosta et al. 2019, Campbell et al. 2010). Snowpacks affect the energy balance of ecosystems through their high albedo, high latent heat, low thermal conductivity, and high emissivity (Zhang 2005), and as snowpack characteristics change, landscapes are experiencing increasing periods of time where their soils are more closely coupled to the surrounding air temperatures. This is likely to lead to significant changes in the soil energy balance, as well as changes in the time between snowmelt and the beginning of the growing season in spring (Groffman et al. 2012, Contosta et al. 2016b). Snowpacks are also hydrologically important, storing the precipitation of the winter months and then fully wetting soils prior to the growing season. Reduced snowpacks shift the timing of peak snowmelt to earlier in the year (Huntington et al. 2009, Hodgkins et al. 2003, Campbell et al. 2011, Harpold and Molotch 2015), which can increase water limitation of growing season plant productivity (Hu et al. 2010; Buermann et al. 2018; Cooper et al. 2020). Snowpack changes can thus have cascading effects through ecosystems due to their influence on soil temperatures and water availability.

During the winter to spring transition period, the snowpack and soil microclimate undergo a series of rapid changes (Kondo and Yamazaki 1990, Contosta et al. 2016a). Snowmelt raises soil moisture and streamflow (Contosta et al. 2016b), and importantly exposes the comparatively low albedo soils to rapid warming, sometimes rising as much as 8 °C in 48 h (Groffman et al. 2012). Microbial activity is highly correlated with soil temperatures (Classen et al. 2015, Campbell et al. 2005), so changes in the date of spring soil warming and biological "re-activation" will result in temporal changes in the availability of nutrients, potentially resulting in asynchronies with plant demand and subsequent exposure to hydrologic loss through subsurface drainage (Muller and Borman 1976, Zak et al. 1990). Additionally, shifts in snowmelt and soil warming to earlier in the year could increase the annual flux of carbon from the soils, given that soil respiration is tightly correlated with soil temperatures (Fahey et al. 2005a).

In this study, we aimed to understand how snow regulates winter and spring soil microclimate, as well as to test whether changes in soil warm-up dates might be a significant contributor to the soil microclimate in the growing season. To do so, we used a natural climate gradient resulting from a variety of sites with contrasting elevations and aspects within a northern hardwood forest and investigated relationships between seasonal snowpack and soil microclimate using 8 years of bi-weekly snowpack and soil frost measurements along with continuous data from soil climate sensors.

We hypothesized that:

1. Snowpack characteristics will mediate the relationship between air temperatures and the soil microenvironment. Snowpacks with low seasonal snow water equivalent (SWE) will result in higher soil temperature variation, deeper soil frost, and higher soil moisture variation when compared with snowpacks with high seasonal SWE.
2. Snowpack characteristics will determine the date of rapid soil warming in the spring, and
3. The date of rapid soil warming in spring will influence soil temperature and moisture during the growing season.

2 Methods

2.1 Site description

The study was conducted at the Hubbard Brook Experimental Forest (HBEF) in New Hampshire (USA; 43° 56′ N 71° 45′ W). The HBEF is administered by the USDA Forest Service and is described in detail in other publications (see Holmes and Likens 2016). In short, the climate is cool and continental with temperatures ranging from a July average of 18 °C to a January average of − 9 °C and a snowpack that typically forms in December and persists until mid-April (Campbell et al. 2010). Soils are generally Typic Haplorthods with relatively thick (3–15 cm) organic horizons (Likens and Bormann 1995) and evident lateral podzolization influenced by bedrock outcrops and topography (Bailey et al. 2014). The vegetation is dominated by approximately 100-year-old sugar maple (*Acer saccharum* Marsh.), yellow birch (*Betula alleghaniensis* Britt.), and American beech (*Fagus grandifolia* Ehrh.), with balsam fir (*Abies balsamea* (L.) Mill), red spruce (*Picea rubens* Sarg.), and paper birch (*Betula papyrifera* Marsh.) common at higher elevations.

2.2 Climate gradient plots

In 2010, a network of plots was established utilizing variation in elevation and aspect in order to represent the range of current climate conditions supporting northern hardwood forests. The initial objectives of the study were to understand the role of winter climate in influencing belowground ecosystem processes, including nitrogen cycling processes (Durán et al. 2014, Morse et al. 2015, Durán et al. 2016, Durán et al. 2017), microbial biomass and activity (Durán et al. 2017), trace gas fluxes (Morse et al. 2015), root-microbe interactions (Sorensen et al. 2016), soil water nutrient mobilization (Fuss et al. 2016), and soil arthropod abundance and activity (Christensen et al. 2017). Plots were circular and either 20-m diameter for a subset of six plots or 10-m diameter for the remaining eight plots used in this study. The plots encompass an elevation range of 375 to 770 m asl on both north- and south-facing aspects (Fig. 1).

2.3 Measurements and data processing

Snow depth, snow water equivalent (SWE), and soil frost were measured in the subset of six larger plots on a biweekly basis each year from the first measurable snowfall until each plot was snow and frost-free in the spring (Groffman 2019a). Each plot had three equidistant sampling sites centered on a soil frost tube (Ricard et al. 1976, Hardy et al. 2001), where the freezing depth was measured. Snow depth and snow water equivalent (SWE) measurements were made adjacent to each frost tube using a Federal snow sampling tube (Rickly Hydrological Company, Columbus, OH, United States). Each frost and snow measurement was recorded as the average for each plot/measurement period. In order to describe the complex nature of a seasonal snowpack using a single variable, we created a time-integrated snow variable by plotting the bi-weekly measurements over time and utilizing a trapezoidal approach to calculate the area under the curve (Duran et al. 2014). This produced a variable which reflected both the magnitude and longevity of the seasonal snowpack (SNOW). Snow water equivalent was chosen for analysis of the snow measurements, but we note a generally high correlation between snow depth and SWE (Hill et al. 2019). Maximum frost depth per season (MAXFROST) was used as an indicator of how much of the rooting zone was likely to be exposed to freezing conditions.

All 14 plots were outfitted with Decagon (now Meter™) EM50 data loggers connected to a pair of 5TM combination temperature and volumetric water content (VWC) probes inserted

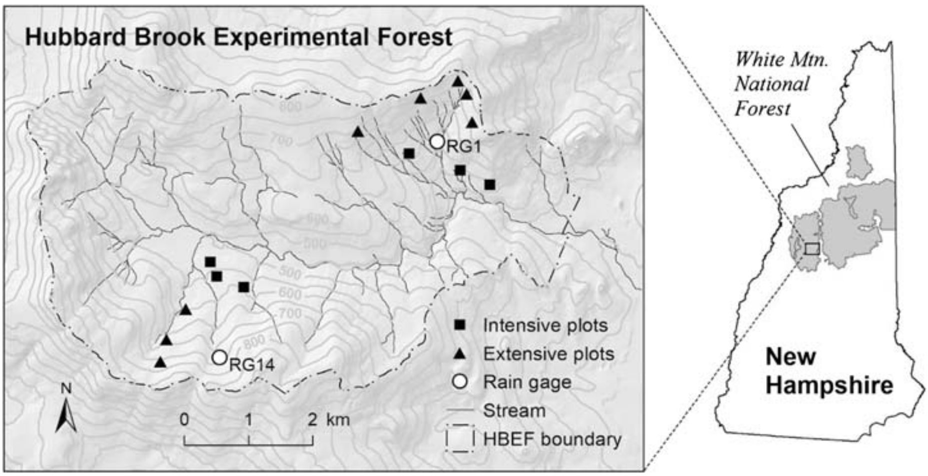

Fig. 1 Location of Hubbard Brook Experimental Forest, including the sites used in this study

into the forest floor vertically from 5 to 10 cm depth. Soil temperature and VWC were measured hourly from the beginning of the study (November 2010) through December 2018, providing eight full years of site soil conditions for analysis. Outliers were excluded with an algorithm that performed a range check and evaluated extreme deviations from the other sensors (see supplementary methods). An ensemble regression approach was then used to fill gaps in the data by regressing the probe in question against the 27 other probes and calculating the median prediction for the missing value (See et al. 2020). Hourly temperature and moisture averages for each site were calculated as the mean of the two probes for each hourly reading, after which a daily site temperature value was calculated by averaging the 24-hourly values (Groffman 2019b).

Soil microclimate characteristics were calculated using the daily values from each site. Specifically, we calculated winter soil temperature variability (WSTV) as the standard deviation of the log-transformed daily soil temperature observations (Durán et al. 2014) and used this as an indicator of the coupling of soil and air temperatures. Similarly, we calculated winter soil moisture variation (WSMV) using the same method as an indicator of coupling between soil moisture and atmospheric conditions, particularly those producing thawing temperatures. Soil warm-up date (WARMUP) was calculated as the day of the year when the daily soil temperature exceeded 5 °C (Groffman et al. 2012) and used as an indicator of the vernal onset of increased metabolic activity in soils (Fahey et al. 2005a, Groffman et al. 2012). Soil activity degree days (SADD) was calculated as the cumulative sum of daily mean soil degree days between when the soil temperatures first rise above the 5 °C threshold in the spring until they drop below that threshold during the approach to winter. Figures S1 and S2 show the soil temperature and moisture time series for each site. To account for variability in soil characteristics among sites, volumetric water content (VWC) was normalized by dividing by the field capacity VWC, defined as the average VWC during the 1 week after soil warm-up, a period with freshly wetted soils which have been gravity-drained and experienced little evapotranspiration (Weil and Brady 2017). As an indicator of summer soil moisture status (VWCsummer), we used the median daily VWC over the period from 1 June to 31 August for each site.

2.4 Data analysis

To test our hypotheses, we used a linear mixed model (LMM) approach, except in one case where a non-linear mixed model was used because a residual analysis indicated the linear approach was not suitable. Repeated observations were grouped by site. To account for the repeated measures, several covariance types were selected for consideration: four spatial types (power, exponential, Gaussian, and linear), variance components, unstructured, and Toeplitz. Each covariance structure was considered for the full model containing all independent variables and each of the dependent variables (WSTV, WSMV, MAXFROST, WARMUP, SADD, and VWCsummer). Each dependent variable therefore had a set of seven Aikake Information Criteria corrected (AICc) values (Littell et al. 1996, Akaike 1998), with the lowest AICc value of the set indicating the best covariance structure for further modeling. Once the best covariance structure was selected, independent variables were tested for significance and dropped if nonsignificant. For hypothesis 1, three separate models were explored for WSTV, WSMV, and MAXFROST. Each model initially included SNOW and YEAR as independent variables. We use YEAR to refer to the calendar year containing the end of each winter season and the subsequent growing season. For hypothesis 2, WARMUP was the dependent variable in the LMM and was modeled similarly to hypothesis 1. The same approach was initially applied to explore whether the date of rapid soil warming in spring influenced soil temperatures (SADD) or moisture (VWCsummer) during the growing season (hypothesis 3). We used all 14 sites with both SADD and VWCsummer as the dependent variable and WARMUP and YEAR as potential independent variables. The LMM approach was found suitable for the analysis of WARMUP and VWCsummer, but analysis of residuals indicated that this was not appropriate for analysis of the relationship between WARMUP AND SADD, so nonlinear mixed models were explored. The best modeling approach based on residual analysis was a breakpoint regression, with multiple potential breakpoints tested manually. Repeated measures were accounted for using the same covariance structures as the LMM's with AICc values determined using the Nlinmix macro in SAS version 9.4 (SAS Institute Inc. 2012). The nonlinear model was expressed as $\text{SADD} = \beta_0$ if $\text{WARMUP} < 100$ or $\text{SADD} = \beta_0 + \beta_1(\text{WARMUP-100})$ if $\text{WARMUP} > 100$.

A post hoc investigation was used to gain insights into the amount of snowpack SWE required to effectively decouple the soil temperatures from air temperatures. We paired the daily soil temperatures from each site containing snow measurements with the nearest available daily air temperature data from two weather stations located on the north- and south-facing slopes of the HBEF (see Fig. 1; USDA Forest Service 2019). The three north-facing plots were paired with the weather station on the north-facing side of the valley and the three south-facing plots were paired with the weather station on the south-facing side of the valley. For each period between field measurements, we calculated the mean SWE from each plot and the ordinary least-squares regression of the soil temperature to air temperature relationship. The slope from that relationship was plotted against the mean SWE during that period to gain insight into how the snowpack regulated the soil-to-air temperature relationship.

3 Results

3.1 Variation of snowpack and soil microclimate

The winters of 2011 and 2018 had the highest and 2012 and 2016 had the lowest amounts of snow (Fig. 2a). In general, the snowiest years had the greatest, and the least snowy years had the smallest intersite variation. The spatial variation shown in Fig. 2b reflects both elevation and aspect effects. Thus, the north-facing sites consistently had higher SNOW than the south-facing sites, despite being comparable in elevation with the highest of the south-facing sites. Within each aspect group, increases with elevation were evident. Average annual soil temperatures varied slightly from year to year (Fig. 2c) and reflected effects of both elevation and aspect. The north-facing sites all had, on average, colder soils than the south-facing sites (Fig. 2d).

A consistent seasonal pattern was evident, with soil frost developing early in the season along with the accumulation of snow but peaking before the maximum snowpack and beginning to thaw as the snowpack continued to accumulate (Fig. 3a). Soil moisture also showed a seasonal pattern, with peaks during snowmelt, declines until mid-October, and increases until the onset of freezing temperatures, as indicated by both the accumulation of snow and soil frost (Fig. 3 a and b) and the average air temperature crossing the 0 °C

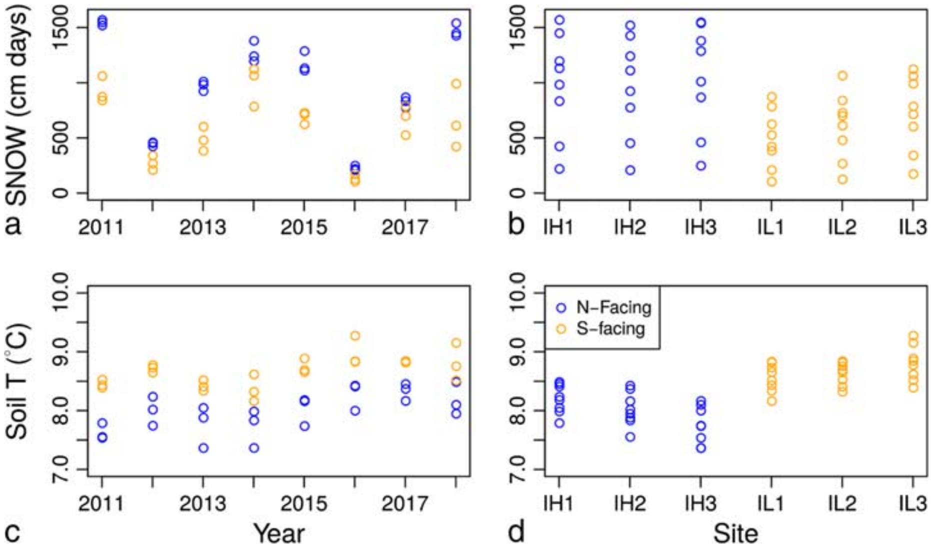

Fig. 2 Interannual and intersite variation in snowpack and mean annual temperature. Panels **a** and **b** show variation in snowpack (SNOW) across sites within each year of observation (**a**) and across years within each site (**b**). Panels **c** and **d** show the variation in mean annual soil temperature at 5 cm across sites within each year (**c**) and across years within each site (**d**). The sites (panel **b** and **d**) are organized by aspect (IH is north facing and IL is south facing) and then secondarily by elevation (the lowest elevation on the left and the highest elevation on the right). In each panel, the north-facing sites are colored in blue and the south-facing sites are colored in orange

threshold. Soil temperatures generally follow air temperatures during the growing season but become decoupled when the snowpack is present (Fig. 3 a and b).

3.2 Winter soil conditions (hypothesis 1)

We found consistent relationships between the snowpack and soil temperature metrics, but not between snowpack and soil moisture. Thus, SNOW had a significant negative effect on WSTV and MAXFROST, ($p < 0.0001$ and $p = 0.0002$, respectively; Table 1, Fig. 4a–c), but had no significant effect on WSMV, $p = 0.34$ (data not shown). There was more variation around the regression line between SNOW and MAXFROST for values below 600 cm days of SNOW than above this value (Fig. 4b). We did not find a significant effect of YEAR in the models for WSMV. Similarly, residual analysis of the model for WSTV indicated a poor fit using YEAR and it was removed from consideration. The relationship between SNOW and MAXFROST implied by the final model suggests maximum frost depth differences of approximately 7 cm between the lowest and highest observed snowpacks (Fig. 4b).

Post hoc investigation of the relationship between daily SWE and the coupling of soil and air temperatures suggested that SWE values of approximately 5–8 cm effectively decoupled the air temperatures from the shallow soil temperatures (Fig. 5 a and b).

3.3 Spring warm-up (hypothesis 2)

The soil warm-up date was positively related to both SNOW ($p < 0.0001$; Table 1, Fig. 6), and YEAR ($p = 0.0148$; Table 1, Fig. S3a). For every 100 cm-day increase in SNOW, WARMUP

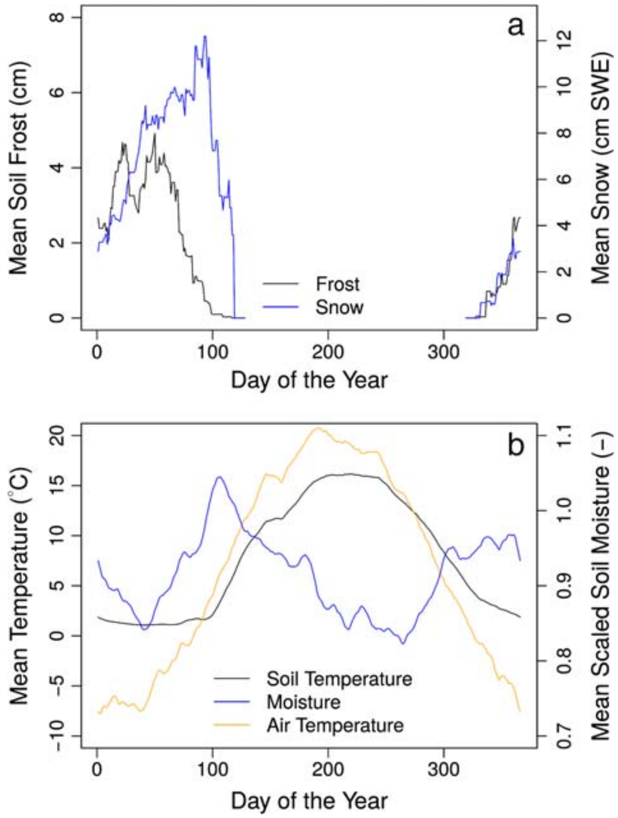

Fig. 3 Generalized seasonal variation of snowpack (SWE), soil frost, soil temperature, and soil moisture at the sites used in this study. Air temperature on the south-face of the Hubbard Brook valley is added in panel **b** for reference. Panel **a** shows the mean snowpack (SWE) and mean soil frost at the six intensively monitored sites, and panel **b** shows the mean soil temperature and moisture curves at all 14 soil sensing sites. The curves represent all 8 years of data, and use a 14-day moving window to smooth the seasonal curve

Table 1 Linear mixed model (LMM) results for hypotheses 1, 2, and the summer soil moisture portion of hypothesis 3. Winter soil moisture variation (WSMV) was not significantly correlated with the dependent variable and is not shown

Hypothesis	Dependent variable	Independent variables	Estimate	Lower 95% confidence	Upper 95% confidence	*p* value
1	Scaled WSTV	Intercept	1.5868	1.4293	1.7443	
		SNOW	− 0.00025	− 0.00031	− 0.00019	< 0.0001
1	MAXFROST	Intercept	10.1616	7.2882	13.0349	
		SNOW	− 0.00502	− 0.00743	− 0.00261	0.0002
2	WARMUP	Intercept	78.2167	70.3195	86.1139	
		SNOW	0.02595	0.02104	0.03086	< 0.0001
		YEAR	1.1722	0.2428	2.1015	0.0148
3	VWCsummer	Intercept	0.7824	0.6691	0.8958	
		WARMUP	0.001724	0.000873	0.002575	0.0001
		YEAR	− 0.01734	− 0.02545	− 0.00923	< 0.0001

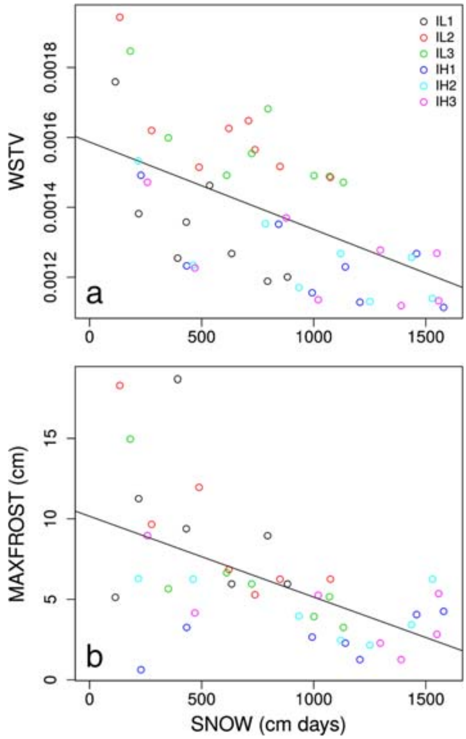

Fig. 4 The relationships between snowpack (SNOW) and both winter soil temperature variation (WSTV; panel **a**) and maximum soil frost depth (MAXFROST; panel **b**). Colors indicate each specific plot, repeated over the 8 years of the analysis. In each figure, the linear mixed model (LMM) line is shown in black. Panel **a** shows snowpack vs. winter soil temperature variation (WSTV) and panel **b** shows snowpack vs. maximum frost depth (MAXFROST)

increased by 2.1 to 3.0 days. The range in the warm-up date was approximately 39 days between the smallest and largest observed snowpacks (Fig. 6). The warm-up date tended to be earlier at the south-facing sites by almost 10 days (Fig. S3b).

3.4 Warm season effects (hypothesis 3)

The soil warm-up date (WARMUP) negatively affected the total soil activity degree days of the subsequent warm season (SADD), but only when the warm-up date was after the 100th day

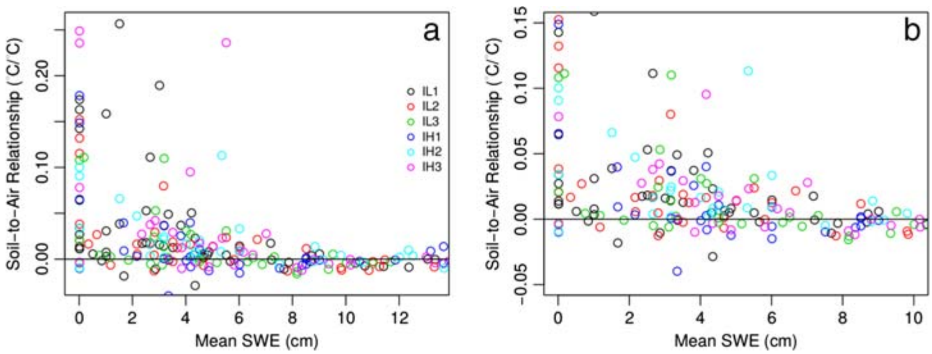

Fig. 5 The slope of the soil to air temperature relationship versus the mean snow water equivalent (SWE) between field measurements for the full range of SWE (**a**) and a smaller range of SWE (**b**) to help see the SWE threshold where soil and air temperatures decouple

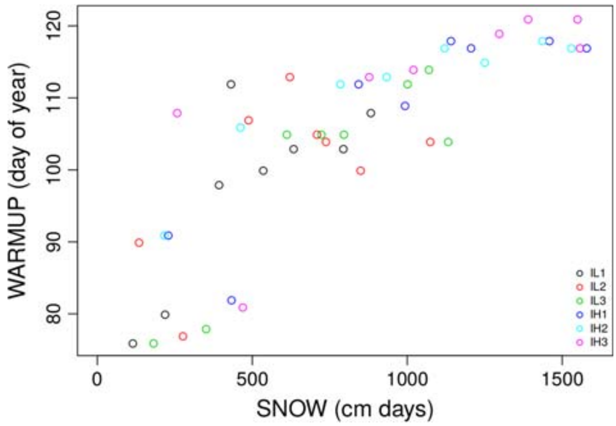

Fig. 6 The relationship between SNOW and soil WARMUP day. No model line is shown because the final model included year-to-year variability as a second significant variable

of the year (April 9–10, Table 2, Fig. 7). Soil activity degree days varied with elevation and aspect, with north facing and higher elevation sites generally having a lower total SADD than south facing or lower elevation sites (Fig. S4 a and b). Year-to-year variability (YEAR) was not significant (Fig. S4b). The latest wake-up date we observed was day 126 (May 5–6), which indicates potential snowpack influences on annual SADD of up to 18% in our dataset.

Summer soil moisture (VWCsummer) was correlated with both the soil warm-up date (WARMUP) and YEAR, with WARMUP being positively, and YEAR being negatively correlated with VWCsummer (Table 1). Earlier warm-up dates (WARMUP), which were associated with smaller seasonal snowpacks (Fig. 6), led to lower average soil moisture during the peak months of the growing season (June–August). The negative correlation with year indicated a decrease in soil moisture over the 2011–2018 period.

4 Discussion

The results from 8 years of monitoring demonstrate that snowpack is an important factor regulating soil temperature dynamics in northern hardwood forests. We show that snowpack decouples soil temperatures from air temperatures, and significantly influences winter soil temperature variability and maximum soil frost, especially under larger snowpacks (Fig. 4). It appears that snowpacks of approximately 5 to 8 cm of SWE fully decouple shallow soil temperatures from the air temperatures at our site, but this is a preliminary observation which warrants further study (Fig. 5). Snow measurements were made biweekly, which provided an opportunity to represent the seasonal snowpack by integrating over time but were not frequent

Table 2 Results of non-linear, breakpoint regression describing the relationship between soil activity degree days (SADD) and soil warm-up date (WARMUP). Best-fit tests were performed manually and the optimal breakpoint was determined to be day 100. The nonlinear model was expressed as SADD = $\beta0$ if WARMUP < 100 or SADD = $\beta0 + \beta1$(WARMUP-100) if WARMUP > 100

Effect	Estimate	Standard error	Alpha	Lower 95% confidence	Upper 95% confidence	*p* value
Beta0	1650.29	22.3124	0.05	1602.1	1698.5	< 0.0001
Beta1	− 11.3367	1.7615	0.05	− 14.833	− 7.841	< 0.0001

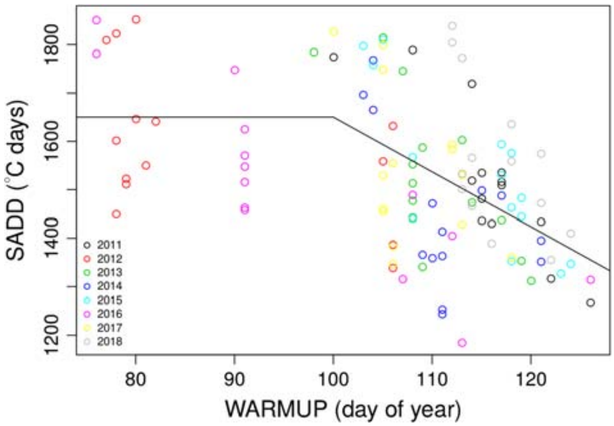

Fig. 7 The relationship between the soil warm-up date (WARMUP) and the warm-season soil activity degree days (SADD) with the non-linear, breakpoint regression (see Table 2) in black

enough to observe shorter timescale variation. We note that our median snow density is 0.23 cm SWE per cm of snow depth, which produces a threshold snow depth of 22 to 35 cm, similar to the snow depth of 30 cm discussed by Brooks et al. (1997) as sufficient to insulate soil temperatures in the US Rocky Mountains (Brooks et al. 1997, Crawford and Legget 1957).

Somewhat surprisingly, winter soil moisture did not appear to be sensitive to the size of the snowpack. We expected to see more variation in soil moisture under snowpacks with lower seasonal SWE because they would be associated with less storage of precipitation in frozen form and thus more liquid water, or with more periods of thawing temperatures, or both. However, the relative stability of winter soil moisture is consistent with the observations of Maurer and Bowling (2014), who observed little variation in normalized winter soil moisture within a variety of sites in the western USA once a snowpack was established. This stability is likely due to the ability of an existing snowpack to store liquid water or to the relatively small amounts of water released from thaws of a cold snowpack (Colbeck 1975). It is also possible that soil moisture increases were obfuscated by the frozen water more common under thinner snowpacks because of challenges sensing frozen soil water content (Spaans and Baker 1996).

After snowmelt, soil temperatures again become responsive to air temperature changes, and the date at which this happens is closely related to the size of the seasonal snowpack (Fig. 6). In turn, this soil temperature warm-up date was clearly and negatively related to the overall magnitude of the soil heat storage (as indicated by the soil activity degree days), but only when it occurred after approximately the 100th day of the year (Fig. 7). These results suggest a threshold from which the warm-up date arises as a strong negative driver of the overall seasonal temperature environment of the soil.

The snowpack decouples the soil temperature from air temperatures because of both its insulating properties and its albedo (Zhang 2005; Ge and Gong 2010). The implications of the decoupling between the winter air temperatures and soils are evident in the negative relationship we observed between snowpack and maximum soil frost depth. In northern hardwood forests, 83% of the root biomass occurs in the top 20 cm of the soil, with 40% of fine roots occurring in the topmost, organic horizons (Yanai et al. 2006). Every instance of frost depth deeper than 10 cm occurred at SNOW values of 600 cm days or lower, so the 7 cm range of frost depth described by the model occurs through a zone of very high root density. Root damage is understood to be the driver of increased nutrient loss during soil freezing (Tierney et al. 2001, Kreyling 2010), and has been linked to decreased growth in adult trees (Reinman et al. 2018). The relationship we observed between snowpack and the variation in winter soil temperatures is consistent with models, suggesting that a future with reduced snowpacks will expose soils to more freeze-thaw events (Henry 2008, Campbell et al. 2010). However, most models suggest no increases in maximum frost depth coincident with reduced snowpacks in the future, due to concurrent warming temperatures (Campbell et al. 2010). Our multi-year observational study suggests a more nuanced picture. Overall, we observed consistent negative relationships between seasonal snowpack and soil frost depth, with shallower snowpacks leading to deeper freezing. However, there appeared to be a threshold in snowpack SWE of about 600 cm days below which the variability in seasonal frost depth increases (Fig. 4b). This increased variability in soil frost under smaller snowpacks could be due to variations in microsite characteristics or in air temperatures which are masked at deeper snowpacks, and suggests changes in the snowpack are an important factor driving soil microclimate until air temperatures warm enough to reduce soil freezing conditions. Our observations are consistent with Durán et al.'s (2014) observations from these sites in their first 2 years of operation, Brown and DeGaetano (2011) in the broader northeastern USA, and the review by Zhang (2005) in arctic systems.

Our study documented the close relationship between seasonal snowpack and soil warm-up date, which will improve predictions of soil responses to changing seasonal snowpacks. The amount of snow that accumulates in winter has been documented to regulate the timing of this soil warm-up date due to the total cold content of the snowpack (snowpack mass and temperature; Zhang 2005; Contosta et al. 2016b). Thus, winter snow accumulation drives the warm-up date by delaying the rapid change in albedo due to the radiative, sensible, and latent heat required to melt a snowpack (Zhang 2005, Contosta et al. 2016b). Our model describes a significant increase in warm-up date over time (by 0.24 to 2.10 days/year). However, this increase is apparently driven by a combination of the very early warm-up dates in 2012 and the late warm-up dates of 2018 (Fig. S3), and is unlikely to hold true over longer timescales, especially given the current trends toward reduced snowpacks (Campbell et al. 2010).

The soil warm-up date clearly impacts metrics of growing season soil temperature and moisture. The annual soil temperature sum was responsive to changes in the warm-up date but showed an apparent threshold where only later warm-up dates were associated with changes in the soil temperature sum (Fig. 7). These results highlight the complexity of how the winter-to-spring transition impacts conditions during the growing season. Intuitively, one might expect that earlier warm-ups would lead to a higher total soil activity degree days since there would simply be more warm days added to the seasonal total. However, early warm-up dates do not necessarily lead to overall seasonally warmer soils, whereas warm-up dates after about the middle of April clearly depress the growing season heat balance. This apparent threshold in the relationship between snowpack disappearance and growing season microclimate could be because some drivers of the soil heat budget, such as incoming radiation and sensible heat flux, are not as intense in the early spring months due to lower solar radiation and cooler temperatures so they do not strongly affect the yearly balance. Identifying the importance of the various drivers of soil heating in the early season is important for predicting soil microclimate changes in a warming world with diminishing snowpacks since the drivers will respond differently to changing atmospheric conditions. The positive relationship between later warm-up dates and soil activity degree days is also complex due to the influence of snowmelt-driven soil moisture on the heat capacity of soils, with moist soils requiring more incoming energy to heat than drier soils (Jury and Horton 2004). Thus, our observation of a growing season soil moisture relationship with the warm-up date may have contributed to the relationship between soil warm-up and the growing season soil temperature sum. These observations together highlight the need for additional studies on how soil water and energy budgets are influenced by changing seasonality.

Our observation that the annual soil heat balance and summer soil moisture conditions can be affected by the snowpack size is important because it ties winter conditions to net ecosystem productivity. Previous work on warm-up dates in the northeastern USA has concentrated primarily on their role in controlling the length of the vernal window—defined as the period between soil warm-up and tree canopy closure—and asynchronies between temperature and moisture mediated soil metabolism and increased plant nutrient demand associated with tree leaf-out (Groffman et al. 2012). The fact that the effect of the warm-up date on soil characteristics can be measured across an entire growing season has additional long-term implications for analyses of carbon cycling and nutrient availability given the close relationships between microbial activity and soil temperature and moisture (Fahey et al. 2005a, Conant et al. 2011). For instance, changes in growing season soil temperature could cause the overall annual carbon flux from the soil pool to be lower in years with persistent snowpacks and higher in years with shorter-lived snowpack. A trend toward shallower snowpacks in the future could thus result in higher losses of CO_2 from forests and reduced carbon storage in soils through increased soil respiration. In contrast, although soil respiration is not generally moisture limited in our forests (Fahey et al. 2005a), the summer soil moisture was drier with shallower snowpacks, suggesting greater potential for periods of moisture limitation on soil respiration as snowpacks diminish. Forest primary productivity could also potentially be affected by changes in snowpack through changes in soil moisture (Ollinger et al. 1998, Buermann et al. 2018, Knowles et al. 2018, Cooper et al. 2020) or the availability of nutrients (Brooks et al. 2011). Intriguingly, other recent work in our region has indicated that the warm-up date is a major control on the seasonal maximum photosynthetic rate, with later warm-up dates correlated with higher maximum photosynthesis (Ouimette et al. 2018). Thus, spring soil warming (and by implication, snowpack) appears to affect the two largest carbon fluxes in our forests (Fahey et al. 2005b). The mechanism behind the observed relationship with photosynthesis is not clear, but it is consistent with previous work from these plots showing higher annual rates of nitrogen mineralization and larger pools of available nitrogen in plots with more snow and less soil frost (Sorensen et al. 2016, Durán et al. 2017). It is also consistent with the idea that shorter vernal windows move the timing of maximum soil resource availability and increased plant demand closer together (Groffman et al. 2012). Further research on the role the snowpack plays in regulating the cycling of carbon and nitrogen in soil will be essential for improving predictions of forest productivity and soil carbon storage. Current models predicting soil carbon storage at Hubbard Brook agree that soil carbon flux will increase with increasing air temperatures, but this may be offset by increases in forest productivity (Dib et al. 2014). Predicted future increases in temperature variability will have unknown, but potentially important effects on the dynamics studied here. The overall energy budget of the soil will change due to growing season changes in temperature, moisture, and biological influences, in addition to changes in winter, but the interaction of these effects is uncertain and warrants further attention.

Our analysis of the relationship between the warm-up date and soil activity degree days suggested a threshold that corresponds with a SNOW value of about 600 cm days. At this point, the relationship between SNOW and warm-up date also appears to change, with less variability and a possible change in slope (Fig. 6). Similarly, at SNOW values of about 600 cm days, the relationship between SNOW and MAXFROST appears to become less variable (Fig.

4b), perhaps indicating a point where individual site characteristics become less important. Altogether, these patterns indicate a possible threshold below which snow loses its major influence on the soil heat budget (in both winter and summer) as a latent heat buffer (Contosta et al. 2016b). Further work toward identifying a threshold in snow characteristics will greatly aid in efforts to predict future soil conditions and ecosystem function.

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Authors' contributions GW and MG led the project by first organizing the data, including gap filling and data hygiene, performing initial data analysis, and sharing the writing responsibilities. JB performed the final data analysis. JC and PG were involved with the project since its inception and repeatedly reviewed drafts of the manuscript. PG was the lead PI on the grant that initiated the project. JM and JD were post-docs who designed the original study and oversaw instrument installation, published initial papers from the study, and provided helpful comments and review of the manuscript toward its final stages.

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Data availability Soil temperature and moisture data: https://doi.org/10.6073/pasta/7409e6255a71e18f6d0c4b726f68b65f. Snow water equivalent and soil frost depth data: https://doi.org/10.6073/pasta/3958640a5f5ed3af7b5e40a5cc710b40. Air temperature data: https://doi.org/10.6073/pasta/5885076607dd57101dfd6129758d5adc.

Compliance with ethical standards

Conflict of interest The authors declare they have no conflict of interest.

Ethical approval The authors state that this manuscript is their original work and has not been submitted or published elsewhere.

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Code availability Data processing and analysis was performed in SAS version 9.4 (SAS institute 2012) and R version 3.3.3 (R Core Team 2017).

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