AGU PUBLICATIONS

Water Resources Research

RESEARCH ARTICLE

10.1002/2013WR014452

Key Points:
- Soil temperature and moisture data were examined for western U.S. mountains
- Seasonal snowpack characteristics influence the soil environment
- This has potential impacts for ecosystems and biogeochemical processes

Supporting Information:
- Readme
- Appendix A
- Appendix B

Correspondence to:
G. E. Maurer,
greg.maurer@utah.edu

Citation:
Maurer, G. E., and D. R. Bowling (2014), Seasonal snowpack characteristics influence soil temperature and water content at multiple scales in interior western U.S. mountain ecosystems, *Water Resour. Res.*, *50*, 5216–5234, doi:10.1002/2013WR014452.

Received 22 JUL 2013
Accepted 8 JUN 2014
Accepted article online 11 JUN 2014
Published online 27 JUN 2014

Seasonal snowpack characteristics influence soil temperature and water content at multiple scales in interior western U.S. mountain ecosystems

Gregory E. Maurer[1] and David R. Bowling[1]

[1]Department of Biology, University of Utah, Salt Lake City, Utah, USA

Abstract Mountain snowpacks directly and indirectly influence soil temperature (T_{soil}) and soil water content (θ). Vegetation, soil organisms, and associated biogeochemical processes certainly respond to snowpack-related variability in the soil biophysical environment, but there is currently a poor understanding of how snow-soil interactions vary in time and across the mountain landscape. Using data from a network of automated snowpack monitoring stations in the interior western U.S., we quantified seasonal and landscape patterns in T_{soil} and θ, and their dependence on snowpack characteristics over an eleven year period. Elevation gradients in T_{soil} were absent beneath winter snowpacks, despite large gradients in air temperature (T_{air}). Winter T_{soil} was warmer and less variable than $T_{air,}$ but interannual and across-site variations in T_{soil} were likely large enough to impact biogeochemical processes. Winter θ varied between years and across sites, but during a given winter at a site it changed little between the start of snowpack accumulation and the initiation of spring snowmelt. Winter T_{soil} and θ were both higher when early winter snow accumulation was greater. Summer θ was lower when summer T_{air} was high. Depending on the site and the year examined, summer θ was higher when there was greater summer precipitation, a larger snowpack, later snowpack melt, or a combination of these factors. We found that snowpack-related variability in the soil environment was of sufficient magnitude to influence biogeochemical processes in snow-dominated ecosystems.

1. Introduction

Snowfall is the dominant hydrologic input to the mountain watersheds of the western U.S., making up 40–70% of annual precipitation [*Serreze et al.*, 1999]. Winter snowpacks persist for a large portion of each year and are primary controllers of the energy and water balance of soils in the region. Snowpack effects on soil temperature and water content directly and indirectly influence vegetation, soil microbial communities, and associated biogeochemical processes during the cold season and the warm season [*Lipson et al.*, 2002; *Monson et al.*, 2006b; *Litaor et al.*, 2008]. The western U.S. experiences high interannual and spatial variability in snowpack size, duration, and melt timing, but at present, there is no comprehensive understanding of how this variability influences the soil environment.

The rates of many biogeochemical processes vary with temperature and moisture. Studies of soil carbon cycling across elevation gradients, for example, have found that changes in soil respiration, rates of organic matter decomposition, and the storage of soil carbon are linked to soil temperature and moisture [*Amundson et al.*, 1989; *Trumbore et al.*, 1996; *Conant et al.*, 2000; *Kueppers and Harte*, 2005]. Despite colder temperature, these and other ecologically important processes occur beneath winter snowpacks. Below-snowpack soil respiration accounts for anywhere from ~12% to 50% of the annual carbon dioxide loss in ecosystems with persistent winter snowpacks [*Liptzin et al.*, 2009]. In addition, decomposition [*Hobbie and Chapin*, 1996; *Williams et al.*, 1998; *Kueppers and Harte*, 2005; *Baptist et al.*, 2009], nitrogen mineralization and immobilization by microbial communities [*Brooks and Williams*, 1999; *Grogan et al.*, 2004; *Schimel et al.*, 2004; *Kielland et al.*, 2006], and the production and consumption of greenhouse gasses such as methane and nitrous oxide [*Sommerfeld et al.*, 1993; *Mast et al.*, 1998; *Schurmann et al.*, 2002; *Groffman et al.*, 2006; *Filippa et al.*, 2009] all occur beneath seasonal snowpacks. Winter snowpack characteristics can influence soil temperature in ways that alter soil carbon cycling during the warm season [*Nowinski et al.*, 2010]. It is unknown how much these biogeochemical processes vary in time and space due to a poor understanding of how snowpacks influence the temperature and moisture environment of soils.

The energy and water balance of the soil surface changes dramatically beneath a snowpack. Because snow has high shortwave albedo and low thermal conductivity, snowpacks decouple soil energy exchange from the radiative and thermal environment at the snowpack surface [*Sturm et al*., 1997; *Grundstein et al*., 2005]. During winter, this slows cooling of soil through radiative, sensible, and latent heat exchange, and when energy availability increases in the spring, it slows warming of the soil by the same processes [*Sokratov and Barry*, 2002; *Bartlett et al*., 2004; *Zhang*, 2005]. Snowpacks temporarily store water, thereby isolating soil from winter precipitation until sufficient energy is available to melt snow and deliver water to soils, streams, or the subsurface [*McNamara et al*., 2005; *Hamlet et al*., 2007; *Williams et al*., 2009; *Bales et al*., 2011]. Winter precipitation can be lost through sublimation or redistributed by wind, vegetation interception, topographic effects, and lateral water movement through the snowpack [*Daly et al*., 1994; *Clark et al*., 2011; *Ohara et al*., 2011; *Eiriksson et al*., 2013]. The impact of these processes on soil temperature and moisture varies depending on snowpack size, distribution, duration, and other snowpack and climate characteristics. Because the interannual and spatial variability in snowpack characteristics and climate are high in the western U.S., it is likely that soil temperature, soil moisture, and associated biogeochemical processes will be highly variable in response.

Numerous studies have identified declining trends in snow cover extent, duration, and snowpack size in the western U.S. [*Hamlet et al*., 2005; *Mote et al*., 2005; *Mote*, 2006; *Dyer and Mote*, 2007]. Model projections tend to agree that these trends will continue and intensify in the coming century [*Brown and Mote*, 2009; *Seager and Vecchi*, 2010]. Although observed changes have been most pronounced for maritime climates, snowpack changes have also been reported in the interior western U.S. [*Clow*, 2010; *Nayak et al*., 2010; *Harpold et al*., 2012]. Researchers have found trends toward earlier spring runoff timing [*Dettinger and Cayan*, 1995; *McCabe and Clark*, 2005; *Stewart et al*., 2005; *Hamlet et al*., 2007] and a larger proportion of precipitation falling as rain instead of snow [*Hamlet et al*., 2005; *Regonda et al*., 2005; *Knowles et al*., 2006; *Gillies et al*., 2012]. Climatic phenomena that influence snowpack size, distribution, and duration are linked to perturbations of ecosystems and human communities in this area, such as widespread increases in wildfire [*Westerling et al*., 2006], drought [*Cayan et al*., 2010], tree mortality [*Anderegg et al*., 2011], and insect outbreaks [*Logan et al*., 2010]. Understanding the relationships between climate, snowpack variability, and the soil environment is critical to predict how ecosystems and biogeochemical processes will respond to future changes in climate.

Here we examine the extant variability in soil temperature and water content in the mountains of the interior western United States and how it is influenced by seasonal snowpack size, environmental conditions during snowpack accumulation, and melt timing. Our study area has a continental climate with cold winters, a seasonal precipitation pattern, and variable winter snowpacks. Sites with maritime climates, which are warmer and have more frequent late winter/early spring snowpack melt and rain-on-snow events [*Knowles et al*., 2006; *Mote*, 2006; *Kapnick and Hall*, 2012], were deliberately excluded from our analysis because we expect them to have different snowpack, soil temperature, and soil moisture dynamics. This study takes advantage of a long-term data set collected by the USDA Natural Resources Conservation Service (NRCS) Snowpack Telemetry (SNOTEL) network. We examine the following hypotheses:

1. There are no elevation gradients in soil temperature when seasonal snowpacks are present.

2. Soil temperature is dependent on snowpack characteristics such as snowpack size and the timing of accumulation.

3. Winter soil moisture (a) changes minimally between the start of snowpack accumulation and the initiation of snowpack melt and (b) is dependent on fall and early winter conditions.

4. Warm season soil moisture is dependent on snowpack size and the timing of snowpack melt.

We show that snowpack-related variability in soil temperature and moisture is of sufficient magnitude to influence soil biological activity, and we discuss the relevance of this complex biophysical environment for ecosystems and biogeochemical processes.

2. Methods

2.1. Study Area and Sites Description

The SNOTEL network is composed of automated stations located in middle to upper elevation basins throughout the western U.S. Data and maps of SNOTEL site locations are available on the NRCS SNOTEL website (http://www.wcc.nrcs.usda.gov/snow/). This network's purpose is to forecast water supply in regions

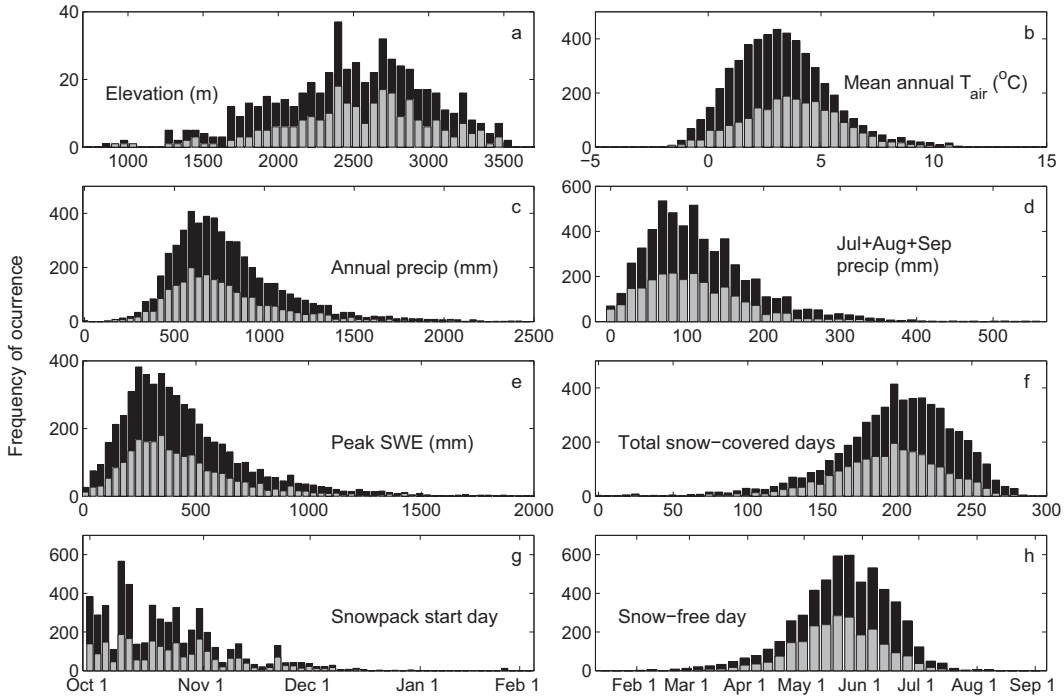

Figure 1. Frequency distributions for selected climate and snowpack characteristics during water years 2001–2011, inclusive. Distributions are shown for the full set of SNOTEL stations in the interior western U.S. (black bars, 574 sites in AZ, CO, ID, MT, NM, NV, UT, and WY) and for the subset of those sites that have soil sensor profiles installed (gray bars, 252 sites).

where snowfall makes up a significant portion of annual precipitation. Our study area includes all sites in Arizona, Colorado, Idaho, Montana, Nevada, New Mexico, Utah, and Wyoming (574 stations—which we refer to as all sites). We excluded all SNOTEL stations in coastal states (CA, OR, WA) because they include mountain ranges with a maritime climatic influence that is distinct from the climate of the interior western U.S. Typically, SNOTEL stations are located in natural or artificial clearings within forested areas and do not span the entire topographic range of the watersheds in which they are operated. Our results, therefore, do not fully represent watershed-scale hydrological processes.

The standard set of SNOTEL measurements includes snow water equivalent (SWE, snow pillow), accumulated precipitation (storage gauge), snow depth (ultrasonic depth sensor), and air temperature ($T_{air,}$ naturally ventilated extended range thermistor). Instrument specifications for these measurements are documented in the NRCS Snow Survey and Water Supply Forecasting National Engineering Handbook [*Natural Resources Conservation Service*, 2010]. In our eight-state study area, a subset of 252 stations (which we refer to as soil sites) were equipped with sensors (Stevens Hydraprobe I and II, Stevens Water Monitoring Systems, Inc., Portland, OR, USA) that monitor vertical profiles of soil temperature (T_{soil}) using integrated thermistors, and soil volumetric water content (θ) using a calibrated measurement of soil dielectric permittivity. The calibration equations used to determine T_{soil} and θ are the same for all sensors and soil types [*Seyfried et al.*, 2005] and are not updated after sensors are installed (T. Tolsdorf, NRCS, personal communication, 2014). The instrument uncertainties for temperature and water content measurements are specified at $\pm 0.26°$C and 3.4%, respectively [*Seyfried et al.*, 2005; K. Bellingham and M. Fleming, Evaluation of the Stevens Hydra Probe's temperature measurements from 230 to 40 degrees celsius, http://www.stevenswater.com/catalog/stevensProduct.aspx?-SKU=%2793640%27]. Because the dielectric properties of ice and liquid water are different, measurements of θ decline sharply as soil water enters the solid phase [*Spaans and Baker*, 1996]. We did not correct for this effect. The number and placement of soil sensors varied among the soil sites, so we used only data from sensors at 5, 20, and 50 cm below the top of the mineral soil horizon for consistency. Soil sensor profiles were typically located within 20 m of the location of the standard SNOTEL instrumentation.

Our study sites spanned a range in elevation from 875 to 3542 m (Figure 1a), in mean annual temperature from -2.8 to 11.3°C (Figure 1b), and in latitude from 32.9°N to 49.0°N (data not shown). For the period

Table 1. Mean and Standard Deviation of Elevation, Snowpack Metrics, and Selected Climate Variables for the Years 2001–2011 (Inclusive)[a]

	All Sites		Soil Sites	
Variable	Mean	SD	Mean	SD
Elevation (m)	2511.4	513.5	2549.8	483.0
Mean annual T_{air} (°C)	3.4	2.1	3.9	2.2
Annual precipitation (mm)	821.1	322.1	791.6	301.1
Summer quarter precipitation (mm)	124.0	73.3	114.1	68.3
Peak SWE (mm)	463.6	285.9	456.9	268.4
Total snow covered days (days)	204.1	39.6	197.8	37.6
Snowpack start day	24 Oct.	17.8	26 Oct.	17.5
Snow-free day	23 May	25.2	20 May	23.1

[a]Data for all sites (n = 574) and the soil sites (n = 252) are shown.

from 2001 to 2011 (inclusive), these sites had a broad range in snowpack size, snowpack start day, snow-free day, and other climatic variables (defined below, see Figure 1). Statistics for snowpack characteristics and selected climatic variables for our study sites during the 2001–2011 period are shown in Table 1.

2.2. Data Processing

We examined hourly T_{soil} and θ data for all available years through 2011 from the soil sites. On average, there were 6.3 years of soil sensor data at these sites. We also examined daily measurements of SWE, precipitation, and air temperature at all sites for the years 2001 to 2011, or for longer periods in cases where the soil sensor record extended to before 2001 (mean = 10.1 years).

The USDA/NRCS provides limited maintenance and quality assurance of the data from SNOTEL soil sensors. For this reason, we created our own quality assurance procedures that excluded a large amount of problematic data. Measurements flagged as errors by the data logger were removed and files with irregular measurement times (other than hourly) were excluded from analysis. Each individual sensor time series was then plotted and visually screened to identify and remove problematic data. When T_{soil}, θ, SWE, or T_{air} data were more than three standard deviations from the moving-window mean (24 h window for hourly data, 10 day for daily data) of a time series, they were classified as outliers and removed. Because soils have a broad range of textural and hydraulic properties, soil θ measurements were not directly comparable between individual sensors. To facilitate comparison across all sensors and sites, θ data for each sensor were normalized linearly according to its full observed range of values (lowest = 0, highest = 1). These procedures are documented, with examples and summary data, in supporting information Appendix A.

Following the quality assurance steps above, we calculated a number of statistics from each time series. The mean and standard deviation of T_{air}, SWE, T_{soil}, and θ were calculated for months and quarters (3 month means of OND, JFM, AMJ, and JAS) at all sites. We calculated accumulated precipitation for each warm season month (M, J, J, A, or S), and for the summer quarter (JAS). Time series of SWE were used to calculate several snowpack metrics. Peak SWE was calculated as the maximum SWE during a water year. Snowpack start day was the first day of persistent snow cover (>5mm of SWE lasting 2 or more days) after 1 October. Snow-free day was the first snow-free day following the day that peak SWE occurred. Total snow covered days was the number of days with >5mm of SWE. For the below-snow period between the snowpack start and snow-free days, we calculated the mean and standard deviation of T_{soil}, θ, and T_{air}. Finally, we calculated presnowpack T_{soil}, θ, and T_{air} for each water year, defined as the mean of each quantity during the 2 week period immediately prior to snowpack start day. When calculating any of the values above from these time series, time periods missing more than 5% of data (~28% of all calculations) were excluded.

2.3. Hypothesis Testing

We examined both interannual and intersite variability in the quantities described above, and used both types of variability to test our hypotheses. Interannual variability refers to variation in a measured quantity over multiple years at one site. To test a hypothesis using interannual variability, we performed least squares linear regression using all years of data from a site. We then repeated the same test for every site and summed the number of sites with significant relationships ($p < 0.05$). To test whether the slopes of

Table 2. Summarized Results for Linear Regression of Mean Below-Snow T_{soil} and Mean Winter Quarter θ on a Number of Explanatory Variables[a]

	Below-Snow T_{soil}			Winter Quarter θ		
Explanatory Variables	5 cm	20 cm	50 cm	5 cm	20 cm	50 cm
Peak SWE	14***	11***	12***	16***	14***	6***
Snowpack start day	13	14	15	13	12	16*
Presnowpack T_{air}	8***	8***	10***	7	10***	4***
Below-snow period T_{air}	12***	12***	13***	14***	13***	6***
Snow-free day	5***	4*	4	8***	9***	8**
Mean Nov. SWE	23***	25***	25***	31***	25***	19***
Mean Dec. SWE	40***	42***	29***	53***	46***	27***

[a]Results from 5 cm, 20 cm, and 50 cm soil depths are shown (n = 252 sites). All regression coefficients (not shown) indicated positive relationships to the explanatory variable. For each variable, numbers represent the total number of sites in which simple linear regression was significant ($p < 0.05$). Asterisks denote the level of significance of the explanatory variable in a multilevel linear model using site as the random variable (*** for $p < 0.001$; ** for $p < 0.01$; * for $p < 0.05$).

these relationships were significant in the aggregate, we fit a multilevel linear model to data from all sites using site as a random variable.

Intersite variability refers to variation in a measured quantity across sites during one or multiple years. When a hypothesis involved clear two-variable relationships across sites, we used simple linear regression (e.g., temperature-elevation gradients or across-site relationships between soil θ at two time periods). Hypotheses involving intersite relationships between more than one explanatory variable were tested using a combination of principal component analysis (PCA) and multiple regression.

As is common with environmental data, many of our explanatory variables were correlated, which makes interpretation of multiple regression results unreliable. To overcome this limitation, we performed two PCAs, one for the below-snow period and one for the warm season. These used our calculated snowpack, soil, and climate statistics (see section 2.2 for a description) as explanatory variables to produce a number of new, uncorrelated principal component axes. All observations in our data set then received a score for each axis. We used these scores as explanatory variables in multiple regression analysis of observations from all years together and subsets of individual year observations (2007, 2009, and 2011). These tests added statistical support for some hypotheses beyond that found using linear regression. A brief summary of the PCA results and our interpretation of the axes will be given in section 3.6. A detailed description of PCA and multiple regression methods and results is presented in supporting information Appendix B.

2.3.1. Hypothesis 1

We examined elevation gradients in T_{soil} and T_{air} using simple linear regression with data from all soil sites. To minimize the influence of latitude or continental location, we also performed the analysis with a geographically constrained subset of sites (Utah). The elevation gradients (slopes of the regressions) were examined for January and July.

2.3.2. Hypothesis 2

Interannual relationships between mean below-snow T_{soil} and several explanatory variables, including snowpack characteristics (Table 2), were examined using simple linear regression at each individual site, and a multilevel linear model to test slope significance for all sites together. We tested the significance of intersite relationships between these variables using multiple regression, with mean below-snow T_{soil} (in individual years, and all years together) as the dependent variable and below-snow principal component axes as explanatory variables.

2.3.3. Hypothesis 3a

We examined within-year variation in below-snow soil θ using two metrics. First, we quantified the month-to-month changes in mean soil θ from October to May at every soil site, in every available year. Second, we calculated the cumulative change between presnowpack soil θ and mean monthly θ in October through May.

2.3.4. Hypothesis 3b

To test this hypothesis, we used simple linear regression between mean winter quarter (JFM) θ and the same explanatory variables used for Hypothesis 2 (Table 2) at each site. We used a multilevel linear model

Table 3. Multiple Regression Results for Three Dependent Variables[a]

Dependent Variables	Explanatory Variables	All Years	2007	2009	2011
Below-snow T_{soil}	Spring snowmelt (PC 1)	−0.02 **	0.01	−0.01	0.03 **
	Winter temperature (PC 2)	0.14 ***	−0.08 ***	−0.14 ***	−0.12 ***
	Snowpack start temperature (PC 3)	−0.04 *	−0.07 *	0.12 **	0.08 **
	Fall snow/soil (PC 4)	0.13 ***	−0.25 ***	−0.16 **	0.02
	Model Adjusted R^2	*0.23*	*0.27*	*0.26*	*0.33*
Winter quarter θ	Spring snowmelt (PC 1)	0.00 *	−0.01	0.01 *	0.02 ***
	Winter temperature (PC 2)	0.05 ***	0.02 **	−0.05 ***	−0.04 ***
	Snowpack start temperature (PC 3)	0.01 **	0.05 ***	−0.01	−0.01
	Fall snow/soil (PC 4)	0.04 ***	−0.10 ***	−0.09 ***	−0.08 ***
	Model Adjusted R^2	*0.24*	*0.36*	*0.42*	*0.38*
Summer quarter θ	Summer T_{air} (PC 1)	−0.02 ***	0.03 ***	−0.01 **	−0.03 ***
	Spring snowmelt/summer precip (PC 2)	0.01 ***	0.02 ***	0.00	−0.01
	Winter T_{soil} (PC 3)	0.03 ***	0.01	−0.02 *	−0.03 **
	Model Adjusted R^2	*0.19*	*0.31*	*0.08*	*0.18*

[a]Mean below-snow T_{soil} and winter quarter θ were regressed against principal component scores from the below-snow PCA, and mean summer quarter θ was regressed against scores from the warm season PCA (see supporting information Appendix B for PC axis details). Each multiple regression model was tested using data from all years together and data from each of three individual years. Regression coefficients for each PC axis and asterisks denoting their significance as explanatory variables in the model (*** for $p < 0.001$; ** for $p < 0.01$; * for $p < 0.05$) are shown.

to test slope significance for all sites together. We also used multiple regression with below-snow principal component axes (Table 3) as explanatory variables.

2.3.5. Hypothesis 4

We tested this hypothesis using simple linear regression of summer quarter (JAS) θ versus a number of warm season variables and snowpack characteristics (see Table 4) at each site. We used a multilevel linear model to test slope significance for all sites together. We also used multiple regression with warm season principal component axes (Table 3) as explanatory variables. As an additional test for intersite differences in summer quarter θ, we compared groups of sites with high and low elevation (a proxy for air temperature), SWE, and summer rainfall. Sites in high summer rainfall groups received greater than 20% of total annual precipitation during the summer quarter (JAS). High and low thresholds for SWE and elevation were selected above and below the mean for all sites, at a value that allowed greater than seven sites in each group.

3. Results

3.1. Snowpack and the Soil Environment at One Site

To illustrate the relationships between snowpack characteristics, $T_{soil,}$ and θ, we highlight multiple years of observations at Currant Creek, Utah. In Figure 2a, 10 consecutive 1 year time series of SWE are plotted on a common time axis. Despite similarities in the shape of the SWE hydrographs, there were large interannual differences. Total snow covered days ranged between 133 and 185 days. Snowpack start day ranged between 22 October and 1 December, and snow-free day ranged between 1 April and 11 May (both varied by ~40 days). Peak SWE ranged between 96 and 400 mm. The data in Figure 2b illustrate the interannual variability and within-year stability of below-snow $T_{soil.}$ Mean below-snow T_{soil} across years ranged between −0.5 and 2.3°C. Below-snow T_{soil} varied little within any given year even though T_{air} consistently dropped far below 0°C in December through February

Table 4. Summarized Results for Linear Regression of Mean Summer Quarter θ on a Number of Explanatory Variables[a]

Explanatory Variables	5 cm	20 cm	50 cm
Peak SWE	11***	18***	21***
Snow-free day	11***	12***	16***
Summer quarter T_{air}	10(-)***	8(-)***	9(-)***
Summer quarter precipitation	26***	16***	7***
Winter quarter 5cm T_{soil}	9	5	3

[a]Results from 5, 20, and 50 cm soil depths are shown (n = 252 sites). Negative regression coefficients are indicated in parentheses, all others were positive. For each variable, numbers represent the total number of sites in which simple linear regression was significant ($p < 0.05$). Asterisks denote the level of significance of the explanatory variable in a multilevel linear model using site as the random variable (*** for $p < 0.001$; ** for $p < 0.01$; * for $p < 0.05$).

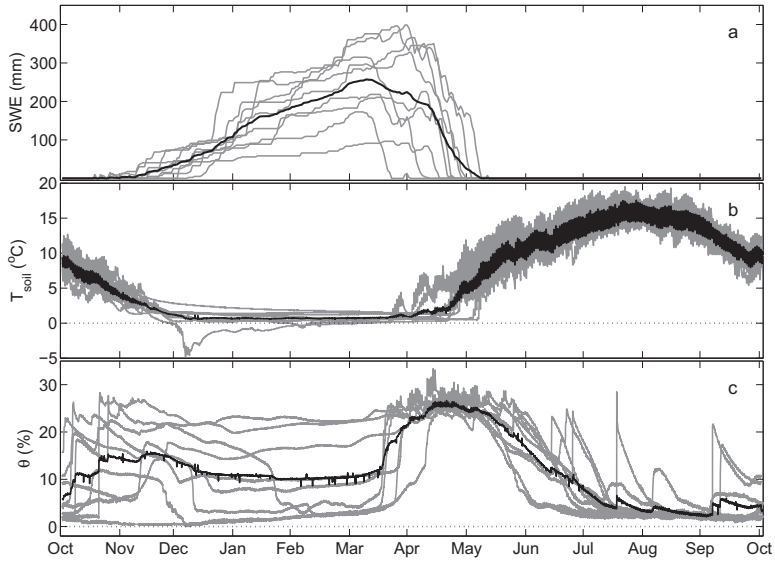

Figure 2. Time series of (a) SWE, (b) 20 cm T_{soil}, and (c) 20 cm θ from 2002 to 2011 at the Currant Creek site (UT). One time series for each individual year since installation of the soil sensors is plotted in gray, and the mean of all these years is plotted in black.

(data not shown). During the coldest year in the record (2010), T_{soil} dropped to almost $-5°$C during December and remained well below 0°C for most of the remainder of winter. The transition to springtime warming of the soil began at the snow-free date, and in some years this occurred after mean T_{air} had climbed above 0°C. The beginning of spring soil warming varied between years by ~40 days (Figure 2b). Below-snow θ changed little until the spring melt began, even as large amounts of precipitation accumulated in the snowpack (Figure 2c). There are exceptions to this, however. In 2010, below-snowpack θ dropped to near zero during the cold soil event described above. This and similar events may indicate the freezing of soil water. Winter quarter θ at the site had high interannual variability, ranging between 3 and 23% (θ not normalized here). In a given year, peak θ coincided roughly with the snow-free date and then declined over the next 2 months. The timing of peak θ varied between years by ~40 days.

3.2. Change in Temperature With Elevation

In the warm season (July), both T_{soil} and T_{air} declined with elevation across all sites, but in January the T_{soil} elevation gradient was absent (Hypothesis 1; Figures 3a and 3b). Results were similar when sites were geographically restricted (Utah, Figures 3c and 3d). The Utah sites had a July T_{soil} (20 cm depth) elevation gradient of $-4.2°$C/km (Figure 3c, $p < 0.001$), which was slightly smaller than the July T_{air} gradient (Figure 3d, $-5.0°$C/km, $p < 0.001$). In January, the T_{soil} elevation gradient for the Utah sites was minimal, but statistically distinguishable from no relationship ($-0.7°$C/km, $p < 0.001$), while a gradient in T_{air} remained ($-2.9°$C/km, $p < 0.001$). The difference between T_{soil} and T_{air} ($T_{soil} - T_{air}$) during January increased with elevation (2.0°C/km, $p < 0.01$) in both groups of sites (data not shown).

3.3. Stability of Winter Soil Moisture

Once a snowpack accumulated, there were only small month-to-month changes in normalized soil θ (averaged across all sites) until the snowpack began to melt (Hypothesis 3a; Figure 4). Between October and November, monthly mean θ increased by ~0.1 (normalized units, dimensionless). There was a slight decline in θ of surface soils (5 and 20 cm depths) possibly due to soil freezing between November and December, followed by little month-to-month change from December to February. There was an increase in θ again in March (Figure 4a). Cumulative changes in mean winter month θ were small (Figure 4b), increasing, on average across all sites, by less than 0.25 (normalized units) between the presnowpack period and March.

3.4. Interannual Variability in Below-Snow Soil Temperature

Interannual variability in below-snow T_{soil} was related to snowpack characteristics (Hypothesis 2). During water year 2005 at the Mosby Mountain site (Utah, Figure 5), for example, a large snowpack accumulated

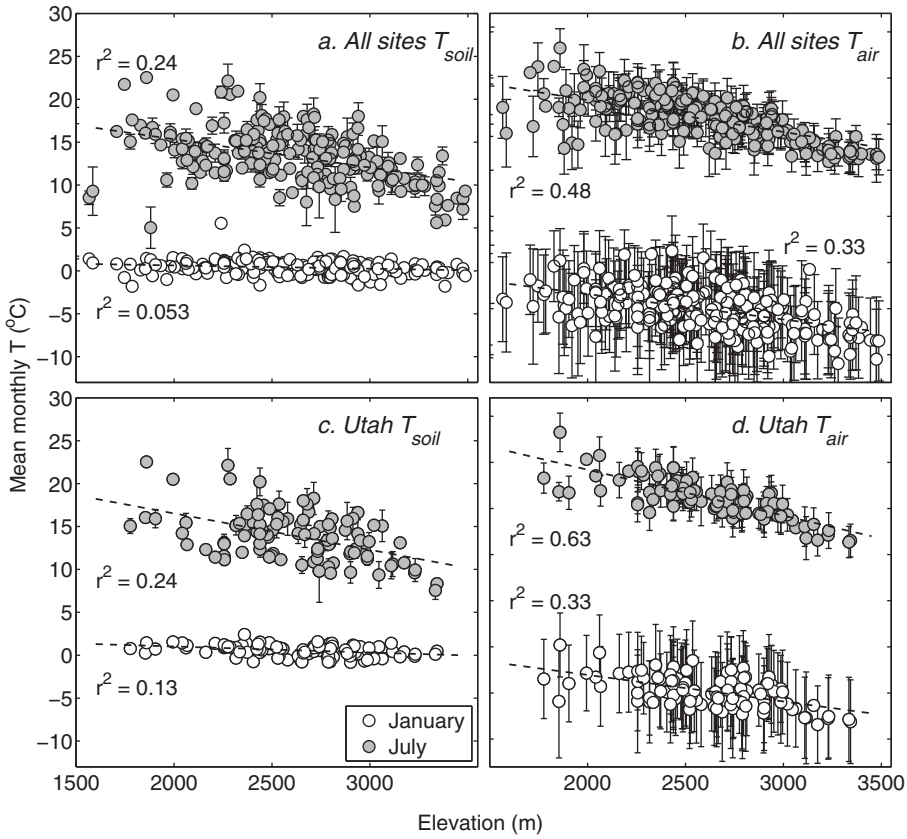

Figure 3. Elevation gradients in mean monthly (a and c) T_{soil} and (b and d) T_{air}. January and July data at all soil sites (n = 252) are shown in plots a and b, and at Utah soil sites (n = 102) in plots c and d. All points are multiyear means of January or July measurements from all available water years, and error bars are 1 standard deviation (some are smaller than the symbols). Dashed lines are least squares linear regressions. T_{soil} measurements are from 20 cm depth. Regression equations for plots a and b: July mean $T_{soil} = -3.3x + 21.94$; January mean $T_{soil} = -0.4x + 1.40$; July mean $T_{air} = -3.4x + 24.49$; January mean $T_{air} = -2.8x + 2.08$. All slopes are significantly different than zero ($p < 0.001$). Utah regression coefficients are given in the text.

early and T_{soil} never dropped below 0°C. In contrast, during water year 2010, the snowpack accumulated slowly and was thin during the early winter. This allowed the soil to cool, and T_{soil} remained well below 0°C for most of the winter. Similar occurrences of low below-snow T_{soil} (<0°C) during years with small early winter snowpacks were widespread in our study area (Figure 6).

Mean below-snow T_{soil} was warmer in years when mean November, December, and January SWE were higher (Figure 7a, one site for December; Table 2, all significant results, January data not shown), and when mean T_{air} during the below-snow period was higher (Table 2). These relationships, however, were only significant at 23–42 sites, depending on soil depth (Table 2). At some sites, T_{soil} was positively correlated with snowpack start day and below-snow period T_{air} (12–15 sites, Table 2), meaning later snowpack accumulation or warmer winter weather was associated with warmer T_{soil} at those sites. The multilevel linear model (Table 2) and multiple regression (section 3.6) provided additional statistical support for some of these relationships.

3.5. Interannual Variability in Soil Moisture

Interannual variability in winter quarter soil θ was dependent on fall and early winter snowpack conditions (Hypothesis 3b). At 19–53 sites (depending on soil depth), mean winter quarter θ was higher in years when mean November, December, or January SWE were higher (Figure 7b, one site for December; Table 2, all significant results, January data not shown). Some sites had higher winter quarter θ in years with a later snowpack start day (12–16 sites, Table 2). Winter quarter θ was also positively related to winter T_{air} at around 6–14 sites and to peak SWE at around 6–16 sites (depending on depth of θ measurements, Table 2).

Interannual variability in summer quarter θ was dependent on summer precipitation, snowpack characteristics, and summer air temperature (Hypothesis 4). At 7–26 sites (depending on soil depth), mean summer

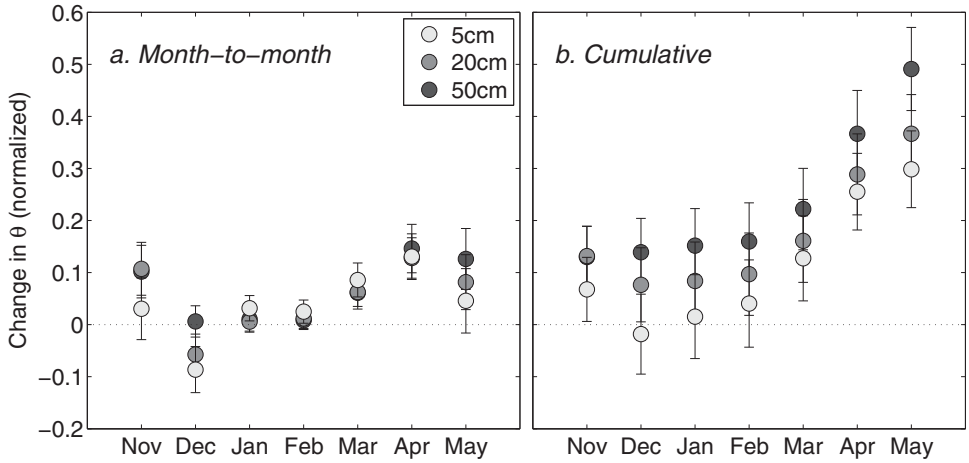

Figure 4. Monthly change in normalized soil θ (dimensionless) at three soil depths. (a) One month changes in mean θ (mean month θ – mean prior month θ). (b) The cumulative change in soil θ since the presnowpack period as described in the text (mean month θ – presnowpack θ). Points represent the mean change for the soil sites (n = 252) at the indicated depth. Error bars are 1 standard error. A dotted line indicating no change in θ is plotted for reference.

quarter θ was higher in years with greater summer quarter precipitation (one site shown in Figure 7c; Table 4, all significant results). This relationship was significant most often at the 5 cm measurement depth (26 sites). Summer quarter θ was also higher in years with greater peak SWE at 11–21 sites (depending on soil depth), but this relationship was significant more often at the 50 cm measurement depth (21 sites, Table 4). At some sites (9–16 sites, soil depth dependent), summer quarter θ was higher in years with a later snow-free date, and lower in years with warmer summer T_{air} (Table 4). Again, multilevel linear models and multiple regression added statistical support to some of these relationships (Tables 2 and 4, section 3.6).

3.6. Intersite Variability in Soil Temperature and Water Content

There was high intersite variability in below-snow T_{soil}, winter quarter soil θ, and summer quarter soil θ in our study area. Mean January T_{soil}, for example, had a range of 11°C across the soil sites, about half the range in mean January T_{air} (Figure 8). To test whether intersite differences in these variables were related to snowpack and other climatic variables across our study sites, we used multiple regression analysis with PCA

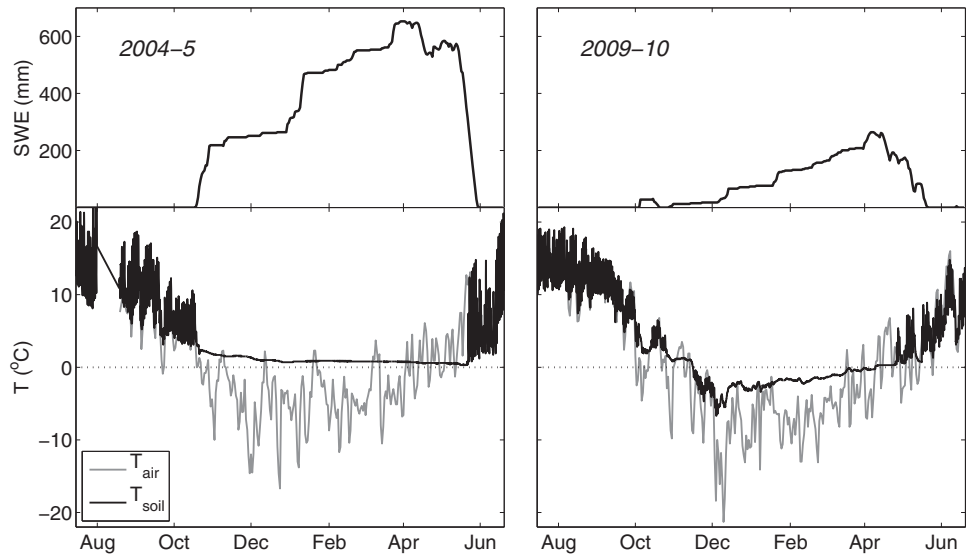

Figure 5. Daily T_{soil} (5 cm depth), T_{air}, and SWE at Mosby Mountain site (UT) during 2 contrasting years. In water year 2005, a large snowpack (SWE) accumulated early, leading to stable, above-zero T_{soil} during the entire below-snow period. In water year 2010, a small early season snowpack led to subzero T_{soil} for much of the below-snow period.

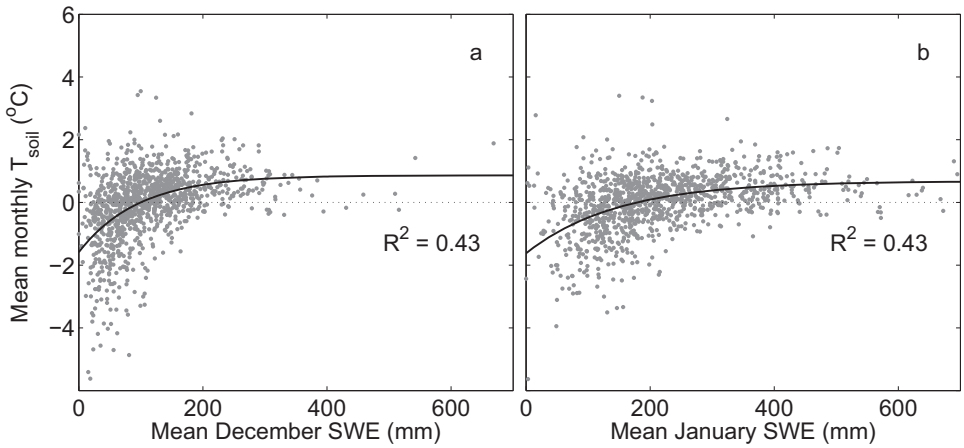

Figure 6. Mean monthly T_{soil} as a function of mean monthly SWE in early winter for all sites (n = 252). Each point represents the mean T_{soil} at 5 cm depth for 1 month at an individual site. The solid lines are the least squares fit to a bounded exponential function ($y=a(1-be^{-cx})$). The fitted values of the upper temperature bounds in December and January are 0.89 and 0.67°C, respectively. The fitted values of SWE at 90% of these upper bounds are 308.6 and 480.3 mm, respectively. Data for December and January of all available water years are shown here, but similar patterns were present during February (R^2 = 0.35) and at other depths (not shown, 20 cm R^2 values = 0.34–37, 50 cm R^2 values = 0.27–0.31). Low early season T_{soil} occurred more frequently with a small snowpack.

scores as the explanatory variables. Detailed PCA and multiple regression results are presented in supporting information Appendix B, but we summarize these results here and in Table 3.

The first four principal component axes from our below-snow PCA were significant as explanatory variables for mean below-snow T_{soil} and winter quarter θ (20 cm depths) in multiple regression analyses (Hypotheses 2 and 3b; Table 3). Based on their explanatory variable loadings (supporting information Table B2), we interpreted these axes as the spring snowmelt axis (PC1), the winter temperature axis (PC2), the snowpack start temperature axis (PC3), and the fall snow/soil axis (PC4). Mean below-snow T_{soil} was significantly higher at sites with warmer winter T_{air} (PC2) and warmer presnowpack T_{soil} and T_{air} (PC3). Sites with warmer presnowpack temperatures tended to be those with an early snowpack start day (supporting information Table B2). Below-snow T_{soil} was also significantly warmer at sites with higher early winter SWE accumulation (PC1 and 4). Mean winter quarter θ was significantly higher at sites with warmer winter T_{air} (PC2), but unlike $T_{soil,}$ it was lower at sites with warm presnowpack T_{soil} and $T_{air.}$ Winter quarter θ was significantly higher at sites with greater October and November SWE and sites with higher presnowpack θ (PC4). Some of these axes were not significant when individual years of data were tested with these multiple regression models.

The first three principal component axes from our warm season PCA (testing Hypothesis 4) were significant explanatory variables for mean summer quarter θ (20 cm; Table 3). We interpreted these axes (supporting information Table B6) as the summer T_{air} axis (PC1), the spring snowmelt/summer precip axis (PC2), and the winter T_{soil} axis (PC3). Mean summer quarter θ was significantly lower at sites with warmer summer T_{air} (PC1). Summer quarter θ was significantly higher at sites with greater warm season precipitation, higher peak SWE, and later snow-free date (PC2 and 3). Again, the significance of some of these axes changed when individual years of data were used in the model. Some explanatory variable loadings for the warm season PCA changed between individual years (supporting information Table B6).

Examination of summer quarter soil θ distributions (Hypothesis 4) revealed differences between groups of sites with high and low elevation, SWE, and summer rainfall (Figure 9). We found that the high summer rainfall sites had, on average, higher summer quarter θ than low summer rainfall sites. Groups with high peak SWE and high elevation had higher summer quarter θ when compared to groups with lower peak SWE or elevation.

4. Discussion

4.1. Soil Temperature Variation Below Seasonal Snowpacks

Temperature in the bulk atmosphere and near-surface air declines with elevation (Figure 3). Hence, one might expect T_{soil} to also decline with elevation. Soil temperature showed little dependence on elevation

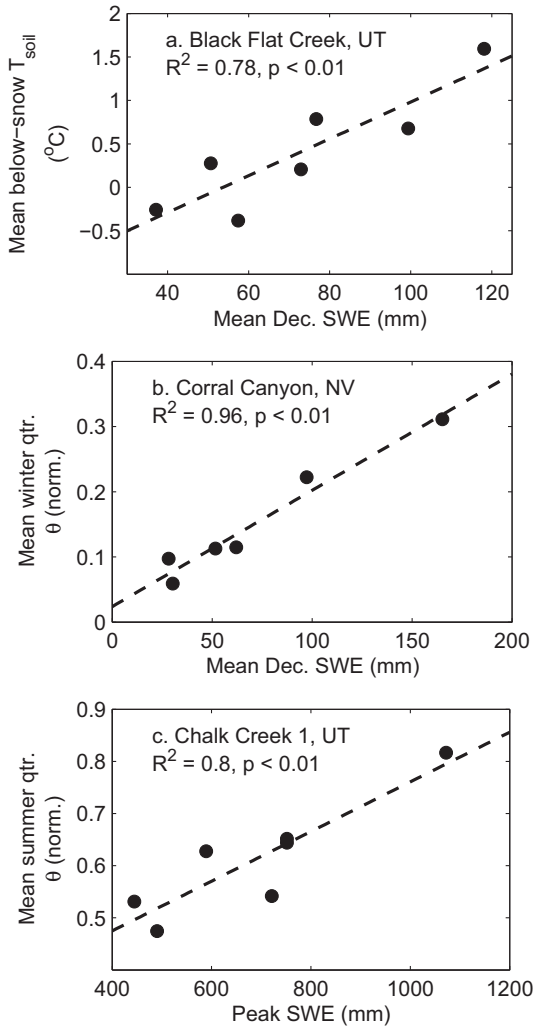

Figure 7. Simple linear regressions of (a) mean below-snow T_{soil} versus mean December SWE, (b) mean winter quarter (JFM) 20 cm θ versus December mean SWE, and (c) mean summer quarter (JAS) 50 cm θ versus peak SWE during different years (interannual variability) at individual SNOTEL sites. These are shown as examples of the regression results presented in Tables 2 and 4.

when a snowpack was present, despite large gradients in T_{air} in our study area (Figure 3). The moist adiabatic lapse rate is generally between 3 and 7 °C/km [*Whiteman*, 2000] and we observed July T_{air} and T_{soil} elevation gradients similar to this across our sites. Elevation gradients in T_{soil} were much smaller than T_{air} gradients when a snowpack was present (Figure 3). These data support our first hypothesis that seasonal snowpacks remove elevation gradients in T_{soil} and are evidence that insulation by snow dramatically reduces energy exchange at the soil surface.

Insulation by snowpacks kept soils warmer than air during the winter. Across all sites, we found mean below-snow T_{soil} values of 0.3, 0.7, and 1.3°C at 5, 20, and 50 cm depths, respectively, all of which were warmer than mean T_{air} during the same period (−1.8°C, Figures 3 and 8). Other studies have shown similar T_{soil} patterns, with below-snowpack T_{soil} exceeding T_{air} when a snowpack is present [*Brooks et al.*, 1995; *Van Miegroet et al.*, 2000; *Hardy et al.*, 2001; *Seyfried et al.*, 2001; *Körner and Paulsen*, 2004; *Monson et al.*, 2006a; *Lundquist and Lott*, 2008; *Sutinen et al.*, 2009; *Masbruch et al.*, 2012; *Schmid et al.*, 2012; *Raleigh et al.*, 2013], but to our knowledge, these landscape-scale changes in T_{soil} gradients have not been demonstrated.

Despite insulation by snowpacks, there was considerable variability in T_{soil} during winter. We found interannual and intersite ranges in below-snow T_{soil} as large as 7 (mean = 1°C) and 11°C (mean = 6°C), respectively, in our study area (Figure 8). To our knowledge, interannual variability in winter T_{soil} has only been quantified in a few isolated studies in western U.S. mountains. At Niwot Ridge, Colorado, for example, there was a 1.5°C range in below-snowpack T_{soil} over a 6 year period [*Monson et al.*, 2006b]. Spatial variability in below-snowpack T_{soil} has been shown to be linked to snowpack depth and T_{air} in arctic environments [*Taras et al.*, 2002]. Studies in snow-dominated mountains are few, but have demonstrated that below-snowpack T_{soil} is often related to snowpack, as well as slope position and aspect [*Körner and Paulsen*, 2004; *Tyler et al.*, 2008; *Scherrer and Körner*, 2010].

Much of the observed variability in below-snow T_{soil} was related to fall and early winter conditions, including snowpack size, presnowpack T_{air} and $T_{soil,}$ and snowpack start day. Snowpack thermal resistance increases with depth, and at greater snow depths soil temperature stops responding to seasonal surface temperature fluctuations [*Sturm et al.*, 1997; *Sokratov and Barry*, 2002; *Bartlett et al.*, 2004; *Grundstein et al.*, 2005; *Zhang*, 2005]. We found that soils were frequently warmer when there was greater early winter SWE accumulation (Tables 2 and 3, PC1 and PC4). Cold soils (mean monthly $T_{soil} < 0$°C) during early winter months were more common at sites with small snowpacks, while sites with large

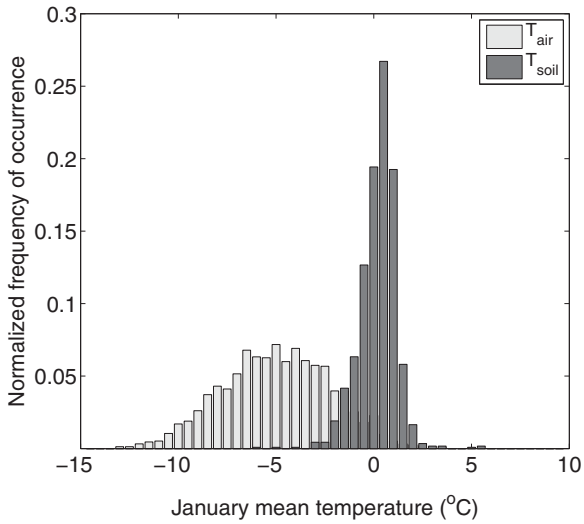

Figure 8. Frequency distributions of mean January soil (20 cm) and air temperature for all sites and all years of data (2001–2011). The histograms are standardized to show the fraction of data in each temperature bin.

snowpacks were generally above 0°C (Figure 6, only December and January shown). We estimated the SWE at which fitted T_{soil} was within 90% of its upper temperature bound to be 308–480 mm. At 30% snow density, this is equivalent to a 1–1.6 m snowpack. This is higher than the estimate of 0.4 m in *Brooks and Williams* [1999]. The model of *Bartlett et al.* [2004] predicts that a snow depth of 1 m insulates the ground from most seasonal T_{air} fluctuations and halts the early winter decline in soil temperature. These results support our second hypothesis that winter soil temperature is dependent on snowpack characteristics. Below-snow T_{soil} was also warmer in years with later snowpack start days (Table 2) at some sites, which is inconsistent with our expectations. A number of sites had higher θ in years with late snowpack start days, so it is possible that warmer T_{soil} in late accumulation years can be accounted for by the high heat capacity of water in the soil or by latent heat release during soil freezing [*Brooks et al.*, 2011].

4.2. Soil Moisture Variation Below Seasonal Snowpacks

Soil moisture below the snowpack was generally stable for several months within a given winter, providing support for our hypothesis (3a) that soil moisture changes minimally between the start of snowpack accumulation and the initiation of snowpack melt. After November, there was little month-to-month or cumulative change in mean monthly θ, and below-snow θ remained similar to presnowpack θ until February (Figure

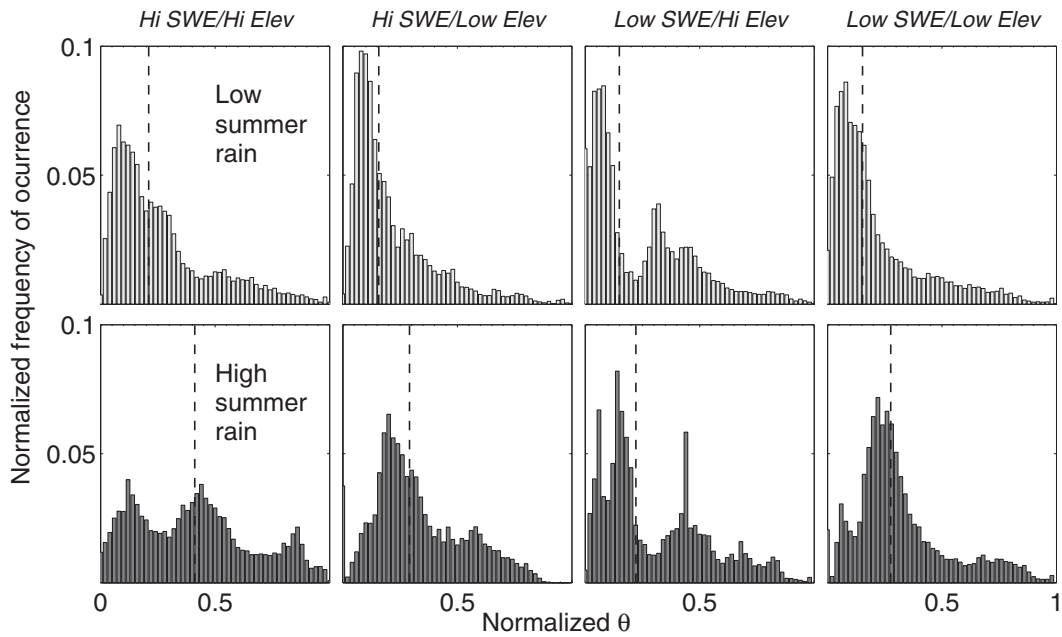

Figure 9. Frequency distributions of mean summer quarter θ (normalized 20 cm data for July, August, and September) for subsets of soil sites with contrasting profiles of elevation, mean snowpack size, and summer precipitation. Sites in the top row received less than 20% of total annual precipitation in the 3 summer months. Sites in the lower row received greater than 20%. High and low elevation and SWE groups are defined in the text. The same 6 years of data, 2006–2011, are used in each group of sites. Median summer quarter soil θ for each group is plotted with a dashed vertical line.

4). Both are evidence that evapotranspiration was low, and little precipitation or snowmelt water infiltrated into soils for 3 winter months or more. In March and April, month-to-month and cumulative increases in θ were observed, suggesting that snowmelt began to reach the soil at this time (Figure 4).

Winter quarter soil moisture was dependent on fall and early winter snowpack and soil conditions. On average, mean winter quarter θ was around 0.4 (normalized) suggesting that, in general, soil moisture was not fully recharged in fall and early winter months. Winter quarter θ was higher when there was greater early winter SWE accumulation or greater presnowpack θ (Tables 2 and 3, PC4). In some years, winter quarter θ was lower at sites where presnowpack T_{soil} and T_{air} were high (Table 3, PC3), indicating that higher evapotranspiration during this period may have dried soils. These observations, coupled with the stability of soil θ during the cold season (Figure 4), provide support for our hypothesis (3b) that midwinter θ was determined by conditions in fall and early winter. We also found, however, a positive relationship between winter quarter θ and winter T_{air} (Table 2; Table 3—PC2), suggesting that winter melt events at warmer sites or in warm years may lead to some recharge of soil moisture.

The fall and early winter period can be viewed as a transitional state between the relative stability of the warm and cold seasons. During this transition, the soil environment is highly sensitive to variability in temperature and precipitation [*Grayson et al*., 1997; *McNamara et al*., 2005]. This is understandable because the phase (rain or snow) of precipitation, and the likelihood that snowfall will melt and recharge soil θ, are both highly sensitive to temperature fluctuations during this time. We did not use fall and early winter precipitation or snowmelt as explanatory variables in multiple regression analysis, and it is possible that these would have provided some additional information. Whatever the dominant drivers of θ are during this fall and early winter transition period, it appears that winter θ is sometimes determined at this time.

4.3. Warm Season Soil Moisture and Snowpack Variability

We found some evidence that summer quarter air temperature, rainfall, and prior spring snowpack characteristics influenced summer soil moisture. Summer quarter θ was lower during warmer years (Table 4), but only at 8–13 sites (depending on soil depth). Sites with warmer T_{air} (Table 3—PC1) also had lower summer quarter θ. Low summer quarter θ may have been the result of high evapotranspiration rates in warm years that removed water from soil. Evapotranspiration is enhanced by warmer air temperature and associated higher evaporative demand. Soil water is primarily recharged by water pulses from snowmelt or summer rain events. Accordingly, we found higher summer quarter θ when there was greater summer precipitation, larger prior spring snowpacks, and later snow-free dates (Tables 4 and 3, PC2 and 3). These relationships were not significant at all sites or in all individual years tested, indicating that the importance of precipitation and snowpack varied in time and space. This provides limited support for our hypothesis (4) that warm season soil moisture is influenced by snowpack characteristics. Warm season air temperature, however, was a more consistent explanatory variable. In our comparison of sites grouped by summer rainfall, elevation, and snowpack size, the group with the highest mean summer quarter θ was the one with sites at high elevations (cooler), with large snowpacks, and large amounts of summer rainfall (Figure 9). High summer rainfall sites were generally wetter than sites with less summer rainfall, and median summer θ was lower at low elevation and low SWE groups. We also found evidence that warm season rainfall events primarily wet the upper layers of the soil profile, while snowmelt recharged θ at greater depth (Table 4).

These results, though complex, agree with other studies of soil water recharge at catchment [*Seyfried*, 1998; *McNamara et al*., 2005; *Williams et al*., 2009] and regional scales in the western U.S. [*Loik et al*., 2004; *Hamlet et al*., 2007]. Both *Seyfried* [1998] and *Williams et al*. [2009] found that spatial variability in snowpack size and melt timing explained spatial variability in θ early in the warm season. As θ declined after the snowpack melted, however, those spatial patterns were replaced by soil moisture patterns determined by summer rain. Mountain soils are often shallow and have a small water storage capacity that limits soil moisture recharge by snowmelt water [*Smith et al*., 2011]. A possible explanation for the weak relationships we observed between summer quarter θ and snowpack is that snowmelt-derived soil water was depleted prior to the summer quarter at many sites. This is consistent with recent observations in the region [*Molotch et al*., 2009]. Local controls, such as soil texture, vegetation, and topography can also greatly influence soil water storage and the rate of θ drawdown during the warm season [*Litaor et al*., 2008; *Williams et al*., 2009; *Bales et al*., 2011]. These and other site-specific variables are undoubtedly important and highly variable in our study area.

4.4. Implications for Ecosystems and Biogeochemical Processes

Soil microbial activity occurring near the freezing point of water is highly sensitive to temperature. This has been observed in laboratory [*Fang and Moncrieff*, 2001; *Mikan et al*., 2002; *Öquist et al*., 2009] and field studies of soil biogeochemical processes [*Brooks et al*., 1996; *Elberling and Brandt*, 2003; *Monson et al*., 2006b]. Other than the effect of temperature on biochemical reaction kinetics, several explanations for this phenomenon have been made, including changes in the availability of liquid water [*Mikan et al*., 2002; *Öquist et al*., 2009] and organic carbon substrates [*Brooks et al*., 2005; *Schimel and Mikan*, 2005; *Davidson and Janssens*, 2006], and the exponential growth of soil microbial communities at low temperatures [*Schmidt et al*., 2009]. Because of this temperature sensitivity, seemingly minor changes in winter soil temperature can have major effects on biogeochemical processes, even at the ecosystem level. In the study by *Monson et al*. [2006b], for example, an interannual range in below-snow T_{soil} from -1.5 to 0°C was responsible for a 21% variation in cumulative annual net ecosystem CO_2 exchange at Niwot Ridge, Colorado. We found that below-snow T_{soil} averaged around 0°C across our western U.S. study sites, but interannual and intersite ranges in below-snow T_{soil} were large enough to significantly impact rates of biological activity in soils (Figure 8).

Soil frost events become less likely in temperate mountain ecosystems as the sizes of seasonal snowpacks increase. Frost formation damages root and microbial biomass and because some soil organisms are more cold sensitive than others, soil community composition can change [*DeLuca et al*., 1992; *Sutinen et al*., 1999; *Tierney et al*., 2001; *Feng et al*., 2007; *Comerford et al*., 2013]. Frost damage is thought to release labile carbon and nutrient rich cell contents into the soil [*Matzner and Borken*, 2008], and a variety of effects on soil biogeochemical processes have been observed following freeze-thaw events. These include increases in soil respiration [*Schimel and Clein*, 1996; *Brooks et al*., 1997; *Feng et al*., 2007], higher soil inorganic nitrogen concentration and N_2O emission [*DeLuca et al*., 1992; *Brooks et al*., 1996; *Groffman et al*., 2001, 2006], and greater export of carbon, nitrogen, and other nutrients from soils in solution [*Boutin and Robitaille*, 1995; *Brooks et al*., 1998; *Fitzhugh et al*., 2001; *Haei et al*., 2010]. Some studies, however, have found that soil frost events have little net effect on, or reduce the rates of these same biogeochemical processes [*Lipson et al*., 2000; *Grogan et al*., 2004; *Hentschel et al*., 2009; *Muhr et al*., 2009; *Groffman et al*., 2011]. We found indirect evidence of soil frost at one site (Figures 2b and 2c), and extensive evidence that fall and early winter conditions influenced whether soil temperature dropped below 0°C during the winter (Figure 6).

Soil moisture also has a well-recognized influence on soil biological activity and associated biogeochemical processes [*Orchard and Cook*, 1983; *Borken and Matzner*, 2009]. Below-snow soil microbial processes, such as those that emit carbon dioxide, methane, and nitrogen oxides during winter, respond to variations in soil moisture [*Mast et al*., 1998; *Filippa et al*., 2009; *Liptzin et al*., 2009; *Aanderud et al*., 2013]. There is some evidence that the availability of soil water beneath melting spring snowpacks stimulates the upregulation of photosynthesis and transpiration in conifer forests in our study area [*Monson et al*., 2005; *Zarter et al*., 2006]. Within a given winter, we generally found stability in below-snow soil θ (Figure 4), but considerable interannual and intersite variability was driven by fall and early winter snow and temperature conditions.

Winter biological and biogeochemical activity can be substantial given the below-snow T_{soil} and moisture conditions found in our study area. Below-snow soil respiration, for example, has been shown to account for anywhere from ~12 to 50% of the annual respiration flux in seasonally snow covered ecosystems [reviewed in *Liptzin et al*., 2009]. Aside from some studies of soil processes along elevation transects in our region [*Amundson et al*., 1989; *Trumbore et al*., 1996; *Kueppers and Harte*, 2005], there is little data on how biogeochemical processes vary spatially and temporally in seasonally snow covered mountain ecosystems. There has been some effort to synthesize aspects of the interactions between snow, soil, and winter biogeochemical cycling into a conceptual model [*Brooks and Williams*, 1999; *Liptzin et al*., 2009; *Brooks et al*., 2011]. In this framework, snowpacks limit soil biological activity when they are shallow or transient enough to allow frozen soil for long periods or permanent enough to restrict warm season primary production and thereby reduce the supply of carbon for soil heterotrophs. The majority of our study sites fall between these extremes. Short duration frost events occur, often in response to fall and early winter snow and weather conditions. These may enhance nutrient availability via organic matter fragmentation [*Hobbie and Chapin*, 1996] and turnover of microbial biomass [*Schimel and Clein*, 1996; *Brooks and Williams*, 1999]. Typically,

however, soils are thawed during winter, permitting the activity and growth of a large below-snowpack soil microbial community [*Lipson et al*., 1999; *Schmidt et al*., 2009]. The decomposition of autumn plant litter inputs provides a carbon source for the growth of this community and fuels the winter biogeochemical activity discussed above [*Taylor and Jones*, 1990; *Hobbie and Chapin*, 1996; *Schmidt and Lipson*, 2004].

The influence of winter snowpacks on the soil biophysical environment also extends to the warm season. Following the winter growth of large below-snow microbial communities, the spring melt is accompanied by a change in microbial community and a rapid decline in microbial biomass [*Brooks et al*., 1996; *Lipson et al*., 1999]. The subsequent flush of nutrients can be lost in spring runoff [*Hood et al*., 2003] or exploited by plants during the warm season [*Brooks et al*., 1998; *Jaeger et al*., 1999; *Lipson et al*., 1999]. The spring snowmelt also marks the beginning of the growing season for most plant communities, and changes in the timing of melt can alter the timing of plant phenological events, such as greening and flowering, in alpine plant communities [*Steltzer et al*., 2009]. Warm season activity by plant and soil communities in snow-dominated ecosystems depends heavily on snowmelt water [*Brown-Mitic et al*., 2007; *Litaor et al*., 2008; *Riveros-Iregui and McGlynn*, 2009], and differences in snowpack size and melt timing can have significant effects on forest productivity [*Molotch et al*., 2009; *Tague et al*., 2009; *Hu et al*., 2010]. Our results support the idea that snowmelt enhances warm season soil moisture availability, but this effect is variable and dependent on snowpack size, melt timing, and summer air temperature for a particular site or year.

4.5. Limitations and Future Research

There are a number of limitations to this study, many of which provide opportunity for future investigation. We focused our study on elucidating the climatic drivers of T_{soil} and θ, and consequently ignored many site-specific variables that influence the soil biophysical environment. Soils vary widely in composition and texture, for example, which have significant effects on water retention and thermal or hydraulic conductivity [*Campbell et al*., 1994; *Abu-Hamdeh and Reeder*, 2000; *Haverkamp et al*., 2005]. Our study sites also vary in topographic position and vegetation cover, which may strongly influence precipitation accumulation, evapotranspiration rate, soil and groundwater flow, and soil surface energy balance. None of these site-specific variables, or other potential sources of uncertainty, are accounted for in our study The statistical models we fit in this study explained only a small amount of the variance in T_{soil} and θ across our study sites (R^2 of 0.07–0.42, Table 3), and it is likely that inclusion of additional site-specific variables and uncertainties would have improved this analysis.

Another limitation stems from our use of artificial, rather than hydrologically defined, seasonal periods. Averaging data into quarterly or monthly values, which are arbitrary with respect to the annual hydrologic cycle, risks losing important information about hydrologic events and processes. In studies examining inter-site or interannual variability, such as ours, it may be advantageous to compare hydrologically based events and seasons rather than artificially imposed ones. Such an approach has been successfully used to study interannual variability in forest ecohydrological processes [*Thomas et al*., 2009].

Finally, though provided by a trusted government agency, the data we used are somewhat provisional and limited in quality. Our own quality assurance procedure for T_{soil} and θ data (see supporting information Appendix A) removed the majority of problematic data, but additional sources of uncertainty remain in the data set. We corrected for obvious instances of sensor change at each site, but there may be cases where sensors changed during the time series used to examine interannual variability. Additionally, soil sensor profiles are not precisely colocated with other SNOTEL measurements (SWE, $T_{air,}$ precipitation) and this may have introduced a mismatch between these measurements and T_{soil} or θ data. These limitations illustrate that publically available data sets are not always what they appear and researchers should approach them with appropriate caution. Nevertheless, we consider the USDA/NRCS SNOTEL data set a valuable one with significant potential to inform ecosystem studies in the western U.S.

5. Conclusions

We found that seasonal snowpack characteristics had significant effects on the soil biophysical environment. First, snowpacks decoupled T_{soil} from $T_{air,}$ reducing elevation gradients in T_{soil} across the landscape during the cold season. Second, below-snow T_{soil} was greatly influenced by the timing and magnitude of snow accumulation, and low early winter snowpacks led to cooler soil and higher likelihood of freeze-thaw

events. Third, θ changed little between the start of snowpack accumulation and the initiation of snowpack melt. Fourth, winter quarter θ was influenced by fall and early winter precipitation and temperature. Finally, snowmelt-derived soil moisture was a limited resource, but availability of this resource was more likely with large snowpacks and later melt timing.

These findings suggest that seasonal snowpack change in the western U.S. will be accompanied by shifts in spatial and temporal patterns of soil temperature and soil moisture. Of particular importance are changes in fall and early winter snowpack development, as seasonal snowpacks isolate the soil environment until spring snowpack ablation begins. Temperature, precipitation, and snowpack variations during this transition from fall to winter give rise to below-snowpack T_{soil} and θ differences large enough to impact soil biological activity and associated biogeochemical processes. This appears to be important at many locations across the western U.S. Snowpack size (peak SWE) and melt timing, while critical to the hydrological processes of the western U.S., only significantly impacted warm season soil water availability at a few of our study sites.

There is growing appreciation for the importance of seasonal snowpacks for ecosystem and biogeochemical processes. This research highlights the important role that spring and fall transitions between snow covered and snow-free states have in setting the stage for these processes in the montane ecosystems of the western U.S. Studies of current hydroclimate, and projected trends in this region indicate that snowpack and temperature changes during these seasons are underway and likely to intensify [*Brown and Mote*, 2009; *Seager and Vecchi*, 2010; *Barichivich et al*., 2012; *Kapnick and Hall*, 2012]. We therefore anticipate changes to the soil temperature and soil moisture environment of the region and a significant response from ecosystems and biogeochemical processes.

Acknowledgments
The authors wish to acknowledge Anne Nolin and two anonymous reviewers for their attentive comments, which greatly improved this manuscript. We also thank a number of staff and scientists of the NRCS Snow Survey program. Tim Bardsley, Jim Marron, Tony Tolsdorf, Karen Vaughan, and Randall Julander provided valuable information and/or helpful discussions on using and interpreting SNOTEL data. Undergraduate Richard Malyn and high-school student Davis Unruh donated time collecting SNOTEL data files. Greg Maurer was partially supported during this study by Office of Science (BER), U.S. Department of Energy, grant DE-SC0005236. The data used in this study are available at http://www.wcc.nrcs.usda.gov/snow/.

References

Aanderud, Z. T., S. E. Jones, D. R. Schoolmaster Jr., N. Fierer, and J. T. Lennon (2013), Sensitivity of soil respiration and microbial communities to altered snowfall, *Soil Biol. Biochem*., *57*, 217–227, doi:10.1016/j.soilbio.2012.07.022.

Abu-Hamdeh, N. H., and R. C. Reeder (2000), Soil thermal conductivity, *Soil Sci. Soc. Am. J*., *64*(4), 1285–1290, doi:10.2136/sssaj2000.6441285x.

Amundson, R. G., O. A. Chadwick, and J. M. Sowers (1989), A comparison of soil climate and biological activity along an elevation gradient in the eastern Mojave Desert, *Oecologia*, *80*(3), 395–400, doi:10.2307/4219063.

Anderegg, W. R. L., J. A. Berry, D. D. Smith, J. S. Sperry, L. D. L. Anderegg, and C. B. Field (2011), The roles of hydraulic and carbon stress in a widespread climate-induced forest die-off, *Proc. Natl. Acad. Sci. U. S. A*., *109*(1), 233–237, doi:10.1073/pnas.1107891109.

Bales, R. C., J. W. Hopmans, A. T. O'Geen, M. Meadows, P. C. Hartsough, P. Kirchner, C. T. Hunsaker, and D. Beaudette (2011), Soil moisture response to snowmelt and rainfall in a Sierra Nevada mixed-conifer forest, *Vadose Zone J*., *10*(3), 786–799, doi:10.2136/vzj2011.0001.

Baptist, F., N. G. Yoccoz, and P. Choler (2009), Direct and indirect control by snow cover over decomposition in alpine tundra along a snowmelt gradient, *Plant Soil*, *328*(1-2), 397–410, doi:10.1007/s11104-009-0119-6.

Barichivich, J., K. R. Briffa, T. J. Osborn, T. M. Melvin, and J. Caesar (2012), Thermal growing season and timing of biospheric carbon uptake across the northern hemisphere, *Global Biogeochem. Cycles*, *26*, GB4015, doi:10.1029/2012GB004312.

Bartlett, M. G., D. S. Chapman, and R. N. Harris (2004), Snow and the ground temperature record of climate change, *J. Geophys. Res*., *109*, F04008, doi:10.1029/2004JF000224.

Borken, W., and E. Matzner (2009), Reappraisal of drying and wetting effects on C and N mineralization and fluxes in soils, *Global Change Biol*., *15*(4), 808–824, doi:10.1111/j.1365-2486.2008.01681.x.

Boutin, R., and G. Robitaille (1995), Increased soil nitrate losses under mature sugar maple trees affected by experimentally induced deep frost, *Can. J. For. Res*., *25*(4), 588–602.

Brooks, P., M. Williams, D. Walker, and S. Schmidt (1995), The Niwot Ridge snow fence experiment: Biogeochemical responses to changes in the seasonal snowpack, in edited by K. A. Tonneseen, Biogeochemistry of Seasonally Snow-Covered Catchments, pp. 293–302, IAHS, Boulder, Colo.

Brooks, P., M. Williams, and S. Schmidt (1998), Inorganic nitrogen and microbial biomass dynamics before and during spring snowmelt, *Biogeochemistry*, *43*(1), 1–15, doi:10.1023/A:1005947511910.

Brooks, P. D., and M. W. Williams (1999), Snowpack controls on nitrogen cycling and export in seasonally snow-covered catchments, *Hydrol. Processes*, *13*(14-15), 2177–2190, doi:10.1002/(SICI)1099-1085(199910)13:14/15<2177::AID-HYP850>3.0.CO;2-V.

Brooks, P. D., M. W. Williams, and S. K. Schmidt (1996), Microbial activity under alpine snowpacks, Niwot Ridge, Colorado, *Biogeochemistry*, *32*(2), 93–113, doi:10.2307/1469255.

Brooks, P. D., S. K. Schmidt, and M. W. Williams (1997), Winter production of CO_2 and N_2O from alpine tundra: Environmental controls and relationship to inter-system C and N fluxes, *Oecologia*, *110*(3), 403–413, doi:10.1007/PL00008814.

Brooks, P. D., D. McKnight, and K. Elder (2005), Carbon limitation of soil respiration under winter snowpacks: Potential feedbacks between growing season and winter carbon fluxes, *Global Change Biol*., *11*(2), 231–238, doi:10.1111/j.1365-2486.2004.00877.x.

Brooks, P. D., P. Grogan, P. H. Templer, P. Groffman, M. G. Öquist, and J. Schimel (2011), Carbon and nitrogen cycling in snow-covered environments, *Geogr. Compass*, *5*(9), 682–699.

Brown, R. D., and P. W. Mote (2009), The response of northern hemisphere snow cover to a changing climate*, *J. Clim*., *22*(8), 2124–2145, doi:10.1175/2008JCLI2665.1.

Brown-Mitic, C., W. Shuttleworth, R. Chawn Harlow, J. Petti, E. Burke, and R. Bales (2007), Seasonal water dynamics of a sky island subalpine forest in semi-arid southwestern United States, *J. Arid Environ*., *69*(2), 237–258, doi:10.1016/j.jaridenv.2006.09.005.

Campbell, G. S., J. D. Jungbauer Jr., W. R. Bidlake, and R. D. Hungerford (1994), Predicting the effect of temperature on soil thermal conductivity, *Soil Sci.*, *158*(5), 307–313.

Cayan, D. R., T. Das, D. W. Pierce, T. P. Barnett, M. Tyree, and A. Gershunov (2010), Future dryness in the southwest US and the hydrology of the early 21st century drought, *Proc. Natl. Acad. Sci. U. S. A.*, *107*(50), 21271–21276, doi:10.1073/pnas.0912391107.

Clark, M. P., J. Hendrikx, A. G. Slater, D. Kavetski, B. Anderson, N. J. Cullen, T. Kerr, E. Örn Hreinsson, and R. A. Woods (2011), Representing spatial variability of snow water equivalent in hydrologic and land-surface models: A review, *Water Resour. Res.*, *47*, W07539, doi:10.1029/2011WR010745.

Clow, D. W. (2010), Changes in the timing of snowmelt and streamflow in Colorado: A response to recent warming, *J. Clim.*, *23*(9), 2293–2306, doi:10.1175/2009JCLI2951.1.

Comerford, D. P., P. G. Schaberg, P. H. Templer, A. M. Socci, J. L. Campbell, and K. F. Wallin (2013), Influence of experimental snow removal on root and canopy physiology of sugar maple trees in a northern hardwood forest, *Oecologia*, *171*(1), 261–269, doi:10.1007/s00442-012-2393-x.

Conant, R. T., J. M. Klopatek, and C. C. Klopatek (2000), Environmental factors controlling soil respiration in three semiarid ecosystems, *Soil Sci. Soc. Am. J.*, *64*(1), 383–390, doi:10.2136/sssaj2000.641383x.

Daly, C., R. P. Neilson, and D. L. Phillips (1994), A statistical-topographic model for mapping climatological precipitation over mountainous terrain, *J. Appl. Meteorol.*, *33*(2), 140–158, doi:10.1175/1520-0450.

Davidson, E., and I. Janssens (2006), Temperature sensitivity of soil carbon decomposition and feedbacks to climate change, *Nature*, *440*(7081), 165–173.

DeLuca, T. H., D. R. Keeney, and G. W. McCarty (1992), Effect of freeze-thaw events on mineralization of soil nitrogen, *Biol. Fert. Soils*, *14*(2), 116–120, doi:10.1007/BF00336260.

Dettinger, M. D., and D. R. Cayan (1995), Large-scale atmospheric forcing of recent trends toward early snowmelt runoff in California, *J. Clim.*, *8*(3), 606–623.

Dyer, J. L., and T. L. Mote (2007), Trends in snow ablation over North America, *Int. J. Climatol.*, *27*(6), 739–748.

Eiriksson, D., M. Whitson, C. H. Luce, H. P. Marshall, J. Bradford, S. G. Benner, T. Black, H. Hetrick, and J. P. McNamara (2013), An evaluation of the hydrologic relevance of lateral flow in snow at hillslope and catchment scales, *Hydrol. Processes*, *27*, 640–654, doi:10.1002/hyp.9666.

Elberling, B., and K. K. Brandt (2003), Uncoupling of microbial CO_2 production and release in frozen soil and its implications for field studies of arctic C cycling, *Soil Biol. Biochem.*, *35*(2), 263–272, doi:10.1016/S0038-0717(02)00258-4.

Fang, C., and J. B. Moncrieff (2001), The dependence of soil CO_2 efflux on temperature, *Soil Biol. Biochem.*, *33*(2), 155–165.

Feng, X., L. L. Nielsen, and M. J. Simpson (2007), Responses of soil organic matter and microorganisms to freeze–thaw cycles, *Soil Biol. Biochem.*, *39*(8), 2027–2037, doi:10.1016/j.soilbio.2007.03.003.

Filippa, G., M. Freppaz, M. W. Williams, D. Helmig, D. Liptzin, B. Seok, B. Hall, and K. Chowanski (2009), Winter and summer nitrous oxide and nitrogen oxides fluxes from a seasonally snow-covered subalpine meadow at Niwot Ridge, Colorado, *Biogeochemistry*, *95*(1), 131–149, doi:10.1007/s10533-009-9304-1.

Fitzhugh, R. D., C. T. Driscoll, P. M. Groffman, G. L. Tierney, T. J. Fahey, and J. P. Hardy (2001), Effects of soil freezing disturbance on soil solution nitrogen, phosphorus, and carbon chemistry in a northern hardwood ecosystem, *Biogeochemistry*, *56*(2), 215–238.

Gillies, R. R., S.-Y. Wang, and M. R. Booth (2012), Observational and synoptic analyses of the winter precipitation regime change over Utah, *J. Clim.*, *25*, 4679–4698, doi:10.1175/JCLI-D-11-00084.1.

Grayson, R. B., A. W. Western, F. H. S. Chiew, and G. Blöschl (1997), Preferred states in spatial soil moisture patterns: Local and nonlocal controls, *Water Resour. Res.*, *33*(12), 2897–2908, doi:10.1029/97WR02174.

Groffman, P., J. Hardy, S. Fashu-Kanu, C. Driscoll, N. Cleavitt, T. Fahey, and M. Fisk (2011), Snow depth, soil freezing and nitrogen cycling in a northern hardwood forest landscape, *Biogeochemistry*, *102*(1), 223–238.

Groffman, P. M., C. T. Driscoll, T. J. Fahey, J. P. Hardy, R. D. Fitzhugh, and G. L. Tierney (2001), Colder soils in a warmer world: A snow manipulation study in a northern hardwood forest ecosystem, *Biogeochemistry*, *56*(2), 135–150, doi:10.1023/A:1013039830323.

Groffman, P. M., J. P. Hardy, C. T. Driscoll, and T. J. Fahey (2006), Snow depth, soil freezing, and fluxes of carbon dioxide, nitrous oxide and methane in a northern hardwood forest, *Global Change Biol.*, *12*(9), 1748–1760, doi:10.1111/j.1365-2486.2006.01194.x.

Grogan, P., A. Michelsen, P. Ambus, and S. Jonasson (2004), Freeze–thaw regime effects on carbon and nitrogen dynamics in sub-arctic heath tundra mesocosms, *Soil Biol. Biochem.*, *36*(4), 641–654, doi:10.1016/j.soilbio.2003.12.007.

Grundstein, A., P. Todhunter, and T. Mote (2005), Snowpack control over the thermal offset of air and soil temperatures in eastern north dakota, *Geophys. Res. Lett.*, *32*, L08503, doi:10.1029/2005GL022532.

Haei, M., M. G. Öquist, I. Buffam, A. Ågren, P. Blomkvist, K. Bishop, M. Ottosson Löfvenius, and H. Laudon (2010), Cold winter soils enhance dissolved organic carbon concentrations in soil and stream water, *Geophys. Res. Lett.*, *37*, L08501, doi:10.1029/2010GL042821.

Hamlet, A. F., P. W. Mote, M. P. Clark, and D. P. Lettenmaier (2005), Effects of temperature and precipitation variability on snowpack trends in the western United States, *J. Clim.*, *18*(21), 4545–4561.

Hamlet, A. F., P. W. Mote, M. P. Clark, and D. P. Lettenmaier (2007), Twentieth-century trends in runoff, evapotranspiration, and soil moisture in the western United States, *J. Clim.*, *20*(8), 1468–1486.

Hardy, J. P., P. M. Groffman, R. D. Fitzhugh, K. S. Henry, A. T. Welman, J. D. Demers, T. J. Fahey, C. T. Driscoll, G. L. Tierney, and S. Nolan (2001), Snow depth manipulation and its influence on soil frost and water dynamics in a northern hardwood forest, *Biogeochemistry*, *56*(2), 151–174.

Harpold, A., P. Brooks, S. Rajagopal, I. Heidbuchel, A. Jardine, and C. Stielstra (2012), Changes in snowpack accumulation and ablation in the intermountain west, *Water Resour. Res.*, *48*, W11501, doi:10.1029/2012WR011949.

Haverkamp, R., F. J. Leij, C. Fuentes, A. Sciortino, and P. J. Ross (2005), Soil water retention, *Soil Sci. Soc. Am. J.*, *69*(6), 1881–1890, doi:10.2136/sssaj2004.0225.

Hentschel, K., W. Borken, T. Zuber, C. Bogner, B. Huwe, and E. Matzner (2009), Effects of soil frost on nitrogen net mineralization, soil solution chemistry and seepage losses in a temperate forest soil, *Global Change Biol.*, *15*(4), 825–836.

Hobbie, S. E., and F. S. Chapin (1996), Winter regulation of tundra litter carbon and nitrogen dynamics, *Biogeochemistry*, *35*(2), 327–338, doi:10.1007/BF02179958.

Hood, E. W., M. W. Williams, and N. Caine (2003), Landscape controls on organic and inorganic nitrogen leaching across an Alpine/Subalpine ecotone, Green Lakes Valley, Colorado Front Range, *Ecosystems*, *6*(1), 0031–0045, doi:10.1007/s10021-002-0175-8.

Hu, J., D. J. P. Moore, S. P. Burns, and R. K. Monson (2010), Longer growing seasons lead to less carbon sequestration by a subalpine forest, *Global Change Biol.*, *16*(2), 771–783, doi:10.1111/j.1365–2486.2009.01967.x.

Jaeger, C. H., III, R. K. Monson, M. C. Fisk, and S. K. Schmidt (1999), Seasonal partitioning of nitrogen by plants and soil microorganisms in an alpine ecosystem, *Ecology*, *80*(6), 1883–1891.

Kapnick, S., and A. Hall (2012), Causes of recent changes in western North American snowpack, *Clim. Dyn.*, *38*(9-10), 1885–1899, doi:10.1007/s00382-011-1089-y.
Kielland, K., K. Olson, R. W. Ruess, and R. D. Boone (2006), Contribution of winter processes to soil nitrogen flux in taiga forest ecosystems, *Biogeochemistry*, *81*(3), 349–360, doi:10.1007/s10533-006-9045-3.
Knowles, N., M. D. Dettinger, and D. R. Cayan (2006), Trends in snowfall versus rainfall in the western United States, *J. Clim.*, *19*(18), 4545–4559, doi:10.1175/JCLI3850.1.
Körner, C., and J. Paulsen (2004), A world-wide study of high altitude treeline temperatures, *J. Biogeogr.*, *31*(5), 713–732, doi:10.1111/j.1365–2699.2003.01043.x.
Kueppers, L. M., and J. Harte (2005), Subalpine forest carbon cycling: Short- and long-term influence of climate and species, *Ecol. Appl.*, *15*(6), 1984–1999, doi:10.1890/04–1769.
Lipson, D. A., S. K. Schmidt, and R. K. Monson (1999), Links between microbial population dynamics and nitrogen availability in an alpine ecosystem, *Ecology*, *80*(5), 1623–1631, doi:10.1890/0012-9658(1999)080[1623:LBMPDA]2.0.CO;2.
Lipson, D. A., S. K. Schmidt, and R. K. Monson (2000), Carbon availability and temperature control the post-snowmelt decline in alpine soil microbial biomass, *Soil Biol. Biochem.*, *32*(4), 441–448.
Lipson, D. A., C. W. Schadt, and S. K. Schmidt (2002), Changes in soil microbial community structure and function in an alpine dry meadow following spring snow melt, *Microbial Ecol.*, *43*(3), 307–314.
Liptzin, D., M. W. Williams, D. Helmig, B. Seok, G. Filippa, K. Chowanski, and J. Hueber (2009), Process-level controls on CO_2 fluxes from a seasonally snow-covered subalpine meadow soil, Niwot Ridge, Colorado, *Biogeochemistry*, *95*(1), 151–166, doi:10.1007/s10533-009-9303-2.
Litaor, M. I., M. Williams, and T. R. Seastedt (2008), Topographic controls on snow distribution, soil moisture, and species diversity of herbaceous alpine vegetation, Niwot Ridge, Colorado, *J. Geophys. Res.*, *113*, G02008, doi:10.1029/2007JG000419.
Logan, J. A., W. W. Macfarlane, and L. Willcox (2010), Whitebark pine vulnerability to climate-driven mountain pine beetle disturbance in the Greater Yellowstone ecosystem, *Ecol. Appl.*, *20*(4), 895–902, doi:10.1890/09-0655.1.
Loik, M. E., D. D. Breshears, W. K. Lauenroth, and J. Belnap (2004), A multi-scale perspective of water pulses in dryland ecosystems: Climatology and ecohydrology of the western USA, *Oecologia*, *141*(2), 269–281.
Lundquist, J. D., and F. Lott (2008), Using inexpensive temperature sensors to monitor the duration and heterogeneity of snow-covered areas, *Water Resour. Res.*, *44*, W00D16, doi:10.1029/2008WR007035.
Masbruch, M. D., D. S. Chapman, and D. K. Solomon (2012), Air, ground, and groundwater recharge temperatures in an alpine setting, brighton basin, Utah, *Water Resour. Res.*, *48*, W10530, doi:10.1029/2012WR012100.
Mast, M. A., K. Wickland, R. Striegl, and D. Clow (1998), Winter fluxes of CO_2 and CH4 from subalpine soils in Rocky Mountain National Park, Colorado, *Global Biogeochem. Cycles*, *12*(4), 607–620.
Matzner, E., and W. Borken (2008), Do freeze-thaw events enhance C and N losses from soils of different ecosystems? a review, *Eur. J. Soil Sci.*, *59*(2), 274–284, doi:10.1111/j.1365-2389.2007.00992.x.
McCabe, G. J., and M. P. Clark (2005), Trends and variability in snowmelt runoff in the western United States, *J. Hydrometeorol.*, *6*(4), 476–482.
McNamara, J. P., D. Chandler, M. Seyfried, and S. Achet (2005), Soil moisture states, lateral flow, and streamflow generation in a semi-arid, snowmelt-driven catchment, *Hydrol. Processes*, *19*(20), 4023–4038, doi:10.1002/hyp.5869.
Mikan, C. J., J. P. Schimel, and A. P. Doyle (2002), Temperature controls of microbial respiration in arctic tundra soils above and below freezing, *Soil Biol. Biochem.*, *34*(11), 1785–1795, doi:10.1016/S0038-0717(02)00168-2.
Molotch, N., P. Brooks, S. Burns, M. Litvak, R. Monson, J. McConnell, and K. Musselman (2009), Ecohydrological controls on snowmelt partitioning in mixed-conifer sub-alpine forests, *Ecohydrology*, *2*(2), 129–142.
Monson, R. K., J. P. Sparks, T. N. Rosenstiel, L. E. Scott-Denton, T. E. Huxman, P. C. Harley, A. A. Turnipseed, S. P. Burns, B. Backlund, and J. Hu (2005), Climatic influences on net ecosystem CO_2 exchange during the transition from wintertime carbon source to springtime carbon sink in a high-elevation, subalpine forest, *Oecologia*, *146*(1), 130–147.
Monson, R. K., S. P. Burns, M. W. Williams, A. C. Delany, M. Weintraub, and D. A. Lipson (2006a), The contribution of beneath-snow soil respiration to total ecosystem respiration in a high-elevation, subalpine forest, *Global Biogeochem. Cycles*, *20*, GB3030, doi:10.1029/2005GB002684.
Monson, R. K., D. L. Lipson, S. P. Burns, A. A. Turnipseed, A. C. Delany, M. W. Williams, and S. K. Schmidt (2006b), Winter forest soil respiration controlled by climate and microbial community composition, *Nature*, *439*(7077), 711–714.
Mote, P. W. (2006), Climate-driven variability and trends in mountain snowpack in western North America, *J. Clim.*, *19*(23), 6209–6220.
Mote, P. W., A. F. Hamlet, M. P. Clark, and D. P. Lettenmaier (2005), Declining mountain snowpack in western North America, *Bull. Am. Meteorol. Soc.*, *86*(1), 39–49.
Muhr, J., W. Borken, and E. Matzner (2009), Effects of soil frost on soil respiration and its radiocarbon signature in a norway spruce forest soil, *Global Change Biol.*, *15*(4), 782–793.
Natural Resources Conservation Service (2010), *National Engineering Handbook*, part 622, U.S. Dep. of Agric., Washington, D. C.
Nayak, A., D. Marks, D. G. Chandler, and M. Seyfried (2010), Long-term snow, climate, and streamflow trends at the Reynolds Creek Experimental Watershed, Owyhee Mountains, Idaho, United States, *Water Resour. Res.*, *46*, W06519, doi:10.1029/2008WR007525.
Nowinski, N. S., L. Taneva, S. E. Trumbore, and J. M. Welker (2010), Decomposition of old organic matter as a result of deeper active layers in a snow depth manipulation experiment, *Oecologia*, *163*(3), 785–792, doi:10.1007/s00442-009-1556-x.
Ohara, N., M. Kavvas, D. Easton, E. Dogrul, J. Yoon, and Z. Chen (2011), Role of snow in runoff processes in a subalpine hillslope: Field study in the Ward Creek watershed, Lake Tahoe, California, during 2000 and 2001 water years, *J. Hydrol. Eng.*, *16*(6), 521–533, doi:10.1061/(ASCE)HE.1943-5584.0000348.
Öquist, M. G., T. Sparrman, L. Klemedtsson, S. H. Drotz, H. Grip, J. Schleucher, and M. Nilsson (2009), Water availability controls microbial temperature responses in frozen soil CO2 production, *Global Change Biol.*, *15*(11), 2715–2722, doi:10.1111/j.1365-2486.2009.01898.x.
Orchard, V., and F. Cook (1983), Relationship between soil respiration and soil moisture, *Soil Biol. Biochem.*, *15*(4), 447–453.
Raleigh, M. S., K. Rittger, C. E. Moore, B. Henn, J. A. Lutz, and J. D. Lundquist (2013), Ground-based testing of MODIS fractional snow cover in subalpine meadows and forests of the Sierra Nevada, *Remote Sens. Environ.*, *128*, 44–57, doi:10.1016/j.rse.2012.09.016.
Regonda, S. K., B. Rajagopalan, M. Clark, and J. Pitlick (2005), Seasonal cycle shifts in hydroclimatology over the western United States, *J. Clim.*, *18*(2), 372–384.
Riveros-Iregui, D. A., and B. L. McGlynn (2009), Landscape structure control on soil CO_2 efflux variability in complex terrain: Scaling from point observations to watershed scale fluxes, *J. Geophys. Res.*, *114*, G02010, doi:10.1029/2008JG000885.
Scherrer, D., and C. Körner (2010), Infra-red thermometry of alpine landscapes challenges climatic warming projections, *Global Change Biol.*, *16*, 2602–2613, doi:10.1111/j.1365-2486.2009.02122.x.

Schimel, J. P., and J. S. Clein (1996), Microbial response to freeze-thaw cycles in tundra and taiga soils, *Soil Biol. Biochem.*, *28*(8), 1061–1066, doi:10.1016/0038-0717(96)00083-1.

Schimel, J. P., and C. Mikan (2005), Changing microbial substrate use in arctic tundra soils through a freeze-thaw cycle, *Soil Biol. Biochem.*, *37*(8), 1411–1418, doi:10.1016/j.soilbio.2004.12.011.

Schimel, J. P., C. Bilbrough, and J. M. Welker (2004), Increased snow depth affects microbial activity and nitrogen mineralization in two arctic tundra communities, *Soil Biol. Biochem.*, *36*(2), 217–227, doi:10.1016/j.soilbio.2003.09.008.

Schmid, M.-O., S. Gubler, J. Fiddes, and S. Gruber (2012), Inferring snowpack ripening and melt-out from distributed measurements of near-surface ground temperatures, *Cryosphere*, *6*(5), 1127–1139, doi:10.5194/tc-6-1127-2012.

Schmidt, S., K. Wilson, R. Monson, and D. Lipson (2009), Exponential growth of "snow molds" at sub-zero temperatures: An explanation for high beneath-snow respiration rates and q 10 values, *Biogeochemistry*, *95*(1), 13–21.

Schmidt, S. K., and D. A. Lipson (2004), Microbial growth under the snow: Implications for nutrient and allelochemical availability in temperate soils, *Plant Soil*, *259*(1-2), 1–7, doi:10.1023/B:PLSO.0000020933.32473.7e.

Schurmann, A., J. Mohn, and R. Bachofen (2002), N_2O emissions from snow-covered soils in the Swiss Alps, *Tellus, Ser. B*, *54*(2), 134–142, doi:10.1034/j.1600-0889.2002.00295.x.

Seager, R., and G. A. Vecchi (2010), Greenhouse warming and the 21st century hydroclimate of southwestern North America, *Proc. Natl. Acad. Sci. U. S. A.*, *107*(50), 21277–21282, doi:10.1073/pnas.0910856107.

Serreze, M. C., M. P. Clark, R. L. Armstrong, D. A. McGinnis, and R. S. Pulwarty (1999), Characteristics of the western United States snowpack from snowpack telemetry (SNOTEL) data, *Water Resour. Res.*, *35*(7), 2145–2160.

Seyfried, M. (1998), Spatial variability constraints to modeling soil water at different scales, *Geoderma*, *85*(2–3), 231–254, doi:10.1016/S0016-7061(98)00022-6.

Seyfried, M. S., G. N. Flerchinger, M. D. Murdock, C. L. Hanson, and S. Van Vactor (2001), Long-term soil temperature database, Reynolds Creek Experimental Watershed, Idaho, United States, *Water Resour. Res.*, *37*(11), 2843–2846, doi:10.1029/2001WR000418.

Seyfried, M. S., L. E. Grant, E. Du, and K. Humes (2005), Dielectric loss and calibration of the Hydra Probe soil water sensor, *Vadose Zone J.*, *4*(4), 1070–1079, doi:10.2136/vzj2004.0148.

Smith, T. J., J. P. McNamara, A. N. Flores, M. M. Gribb, P. S. Aishlin, and S. G. Benner (2011), Small soil storage capacity limits benefit of winter snowpack to upland vegetation, *Hydrol. Processes*, *25*(25), 3858–3865, doi:10.1002/hyp.8340.

Sokratov, S. A., and R. G. Barry (2002), Intraseasonal variation in the thermoinsulation effect of snow cover on soil temperatures and energy balance, *J. Geophys. Res.*, *107*(D10), doi:10.1029/2001JD000489.

Sommerfeld, R. A., A. R. Mosier, and R. C. Musselman (1993), CO_2, CH_4 and N_2O flux through a Wyoming snowpack and implications for global budgets, *Nature*, *361*(6408), 140–142, doi:10.1038/361140a0.

Spaans, E. J. A., and J. M. Baker (1996), The soil freezing characteristic: Its measurement and similarity to the soil moisture characteristic, *Soil Sci. Soc. Am. J.*, *60*(1), 13–19, doi:10.2136/sssaj1996.03615995006000010005x.

Steltzer, H., C. Landry, T. Painter, J. Anderson, and E. Ayres (2009), Biological consequences of earlier snowmelt from desert dust deposition in alpine landscapes, *Proc. Natl. Acad. Sci. U. S. A.*, *106*(28), 11629–11634, doi:10.1073/pnas.0900758106.

Stewart, I. T., D. R. Cayan, and M. D. Dettinger (2005), Changes toward earlier streamflow timing across western North America, *J. Clim.*, *18*(8), 1136–1155.

Sturm, M., J. Holmgren, M. König, and K. Morris (1997), The thermal conductivity of seasonal snow, *J. Glaciol.*, *43*(143), 26–41.

Sutinen, M.-L., T. Holappa, and K. Kujala (1999), Seasonal changes in soil temperature and in the frost hardiness of Scots pine roots under subarctic conditions: Comparison with soil temperature and snow-cover under different simulated winter conditions, *Phyton*, *39*, 213–218.

Sutinen, R., A. Vajda, P. Hänninen, and M.-L. Sutinen (2009), Significance of snowpack for root-zone water and temperature cycles in subarctic lapland, *Arct. Antarct. Alp. Res.*, *41*(3), 373–380, doi:10.1657/1938-4246-41.3.373.

Tague, C., K. Heyn, and L. Christensen (2009), Topographic controls on spatial patterns of conifer transpiration and net primary productivity under climate warming in mountain ecosystems, *Ecohydrology*, *2*(4), 541–554, doi:10.1002/eco.88.

Taras, B., M. Sturm, and G. E. Liston (2002), Snow–Ground interface temperatures in the Kuparuk River Basin, arctic Alaska: Measurements and model, *J. Hydrometeorol.*, *3*(4), 377–394, doi:10.1175/1525-7541(2002)003<0377:SGITIT>2.0.CO;2.

Taylor, B. R., and H. G. Jones (1990), Litter decomposition under snow cover in a balsam fir forest, *Can. J. Bot.*, *68*(1), 112–120, doi:10.1139/b90-016.

Thomas, C. K., B. E. Law, J. Irvine, J. G. Martin, J. C. Pettijohn, and K. J. Davis (2009), Seasonal hydrology explains interannual and seasonal variation in carbon and water exchange in a semiarid mature ponderosa pine forest in central oregon, *J. Geophys. Res.*, *114*, G04006, doi:10.1029/2009JG001010.

Tierney, G. L., T. J. Fahey, P. M. Groffman, J. P. Hardy, R. D. Fitzhugh, and C. T. Driscoll (2001), Soil freezing alters fine root dynamics in a northern hardwood forest, *Biogeochemistry*, *56*(2), 175–190.

Trumbore, S., O. Chadwick, and R. Amundson (1996), Rapid exchange between soil carbon and atmospheric carbon dioxide driven by temperature change, *Science*, *272*(5260), 393–396.

Tyler, S. W., S. A. Burak, J. P. McNamara, A. Lamontagne, J. S. Selker, and J. Dozier (2008), Spatially distributed temperatures at the base of two mountain snowpacks measured with fiber-optic sensors, *J. Glaciol.*, *54*(187), 673–679.

Van Miegroet, H., M. T. Hysell, and A. D. Johnson (2000), Soil microclimate and chemistry of spruce-fir tree islands in northern Utah, *Soil Sci. Soc. Am. J.*, *64*(4), 1515–1525.

Westerling, A. L., H. G. Hidalgo, D. R. Cayan, and T. W. Swetnam (2006), Warming and earlier spring increase western US forest wildfire activity, *Science*, *313*(5789), 940–943.

Whiteman, C. D. (2000), *Mountain Meteorology: Fundamentals and Applications*, Oxford Univ. Press, N. Y.

Williams, C. J., J. P. McNamara, and D. G. Chandler (2009), Controls on the temporal and spatial variability of soil moisture in a mountainous landscape: The signature of snow and complex terrain, *Hydrol. Earth Syst. Sci.*, *13*(7), 1325–1336, doi:10.5194/hess-13-1325-2009.

Williams, M. W., P. D. Brooks, and T. Seastedt (1998), Nitrogen and carbon soil dynamics in response to climate change in a high-elevation ecosystem in the Rocky Mountains, U.S.A., *Arct. Alp. Res.*, *30*(1), 26–30, doi:10.2307/1551742.

Zarter, C. R., B. Demmig-Adams, V. Ebbert, I. Adamska, and W. W. Adams (2006), Photosynthetic capacity and light harvesting efficiency during the winter-to-spring transition in subalpine conifers, *New Phytolog.*, *172*(2), 283–292, doi:10.1111/j.1469-8137.2006.01816.x.

Zhang, T. (2005), Influence of the seasonal snow cover on the ground thermal regime: An overview, *Rev. Geophys.*, *43*, RG4002, doi:10.1029/2004RG000157.

