AGU PUBLICATIONS

Geophysical Research Letters

RESEARCH LETTER
10.1002/2015GL065855

Key Points:
- Snow disappearance explained significant variability in peak soil moisture timing in all ecoregions
- Changes in peak soil moisture timing from warming were larger in maritime versus continental regions
- Soil hydrology modifies the impacts of earlier snowmelt to streamflow and ecological response

Supporting Information:
- Figures S1–S4 and Tables S1–S3

Correspondence to:
A. A. Harpold,
adrian.harpold@gmail.com

Citation:
Harpold, A. A., and N. P. Molotch (2015), Sensitivity of soil water availability to changing snowmelt timing in the western U.S., *Geophys. Res. Lett.*, *42*, 8011–8020, doi:10.1002/2015GL065855.

Received 20 AUG 2015
Accepted 19 SEP 2015
Accepted article online 25 SEP 2015
Published online 14 OCT 2015

Sensitivity of soil water availability to changing snowmelt timing in the western U.S.

Adrian A. Harpold[1] and Noah P. Molotch[2,3]

[1]Natural Resources and Environmental Science Department, University of Nevada, Reno, Nevada, USA, [2]Department of Geography, Institute of Arctic and Alpine Research, University of Colorado Boulder, Boulder, Colorado, USA, [3]Jet Propulsion Laboratory, California Institute of Technology, Pasadena, California, USA

Abstract The ecohydrological effects of changing snowmelt are strongly mediated by soil moisture. We utilize 259 Snow Telemetry stations across the western U.S. to address two questions: (1) how do relationships between peak soil moisture (PSM) timing and the day of snow disappearance (DSD) vary across ecoregions and (2) what is the regional sensitivity of PSM timing to earlier DSD associated with warming and drying scenarios? All western U.S. ecoregions showed significant relationships between the timing of PSM and DSD. Changes in the timing of PSM based on warming predicted for the middle and end of the 21st century ranged from 1 to 9 days and from 6 to 17 days among ecoregions, respectively. The maritime ecoregions PSM timing were 2–3 times more sensitive to warming and drying versus the interior mountain ecoregions. This work suggests that soil hydrology modifies the effects of earlier snowmelt on regional streamflow response and vegetation water stress.

1. Introduction

Snowmelt is a principal control on water availability and runoff generation globally [*Barnett et al.*, 2005]. Western U.S. snowpacks have generally been declining [*Mote et al.*, 2005; *Regonda et al.*, 2005; *Harpold et al.*, 2012], and snow disappearance has shifted earlier over the past 50+ years [*Dye and Tucker*, 2003], with regional sensitivity arising from differences in elevation, climate, and warming temperatures [*Bales et al.*, 2006]. These decreases in snow accumulation have reduced the length of the snowmelt season by 1–5 days/decade over the last 30–50 years [*Stewart et al.*, 2005; *Harpold et al.*, 2012]. The hydrologic implications of these changes is significant as snowmelt partially controls groundwater recharge [*Earman et al.*, 2006; *Jasechko et al.*, 2014], streamflow generation [*Moore et al.*, 2011; *Berghuijs et al.*, 2014], base flow [*Godsey et al.*, 2014], and transpiration [*Hu et al.*, 2010; *Trujillo et al.*, 2012]. Importantly, most snowmelt water infiltrates and moves through the soil profile [*Frisbee et al.*, 2011; *Jasechko et al.*, 2014], and therefore, the coupling of snowmelt and soil moisture partially dictates ecohydrologic sensitivity to climate change.

The importance of snowmelt and soil moisture is well established with regard to forest productivity [*Hu et al.*, 2010; *Peng et al.*, 2010; *Parida and Buermann*, 2014] and disturbance [*Westerling et al.*, 2006]. A consistent finding among these previous works is that earlier snowmelt extends the length of the growing season and leads to increased water stress late in the growing season, particularly in regions that receive little summer rainfall. As a result, earlier snowmelt has been linked to decreased gross primary productivity [*Hu et al.*, 2010; *Trujillo et al.*, 2012; *Peng et al.*, 2010], earlier flowering phenology [*Cayan et al.*, 2001; *Inouye et al.*, 2002; *Dunne et al.*, 2003], and increased occurrence of high-intensity fire [*Westerling et al.*, 2006]. Despite the consistent linkages between snowmelt timing and ecosystem function, little work has connected snowmelt with soil moisture. Consequently, soil moisture represents a missing link in our understanding of ecosystem response to changing snow water inputs.

In mountainous regions of the western U.S. and elsewhere, peak annual soil moisture often coincides with snowmelt [*Molotch et al.*, 2009; *Williams et al.*, 2009; *Bales et al.*, 2011; *Harpold et al.*, 2015]. In these snow-dominated areas, shifts to earlier snow disappearance could affect the timing of peak soil moisture. However, site-specific interactions between climate and soil properties, vegetation water use and rooting depth, and topography dictate the sensitivity of soil moisture dynamics to changing snowmelt magnitude and timing [*Bales et al.*, 2011]. In particular, the magnitude and timing of summer rainfall plays an important role in determining the duration of water stress associated with a shift toward earlier snowmelt [*Seyfried*, 1998; *Williams et al.*, 2009]. Earlier snowmelt has additional implications with regard to streamflow generation, as snowmelt strongly affects antecedent soil moisture conditions and modulates water table heights

[*Huntington and Niswonger*, 2012; *Godsey et al*., 2014]. At the plot scale, soil depth, soil water retention, and rooting depth provide a first-order control on the amount of soil water storage available for transpiration [*Smith et al*., 2011]. While the larger-scale movement of water is strongly influenced by topography and lateral redistribution [*Western et al*., 1999; *McNamara et al*., 2005; *Williams et al*., 2009; *Tague and Peng*, 2013], soil water retention characteristics control the water accessible to vegetation at low-matric potential, and the development of downward pressure gradients and drainage out of the soil profile at water contents above field capacity [*Seyfried et al*., 2009; *Williams et al*., 2009]. Understanding how the interactions between soil properties and variable snowmelt regimes control soil water availability across diverse western U.S. ecosystems remains a critical research gap.

The installation of soil moisture sensors at numerous Snow Telemetry (SNOTEL) stations creates an opportunity to evaluate regional-scale sensitivities of peak soil moisture timing to changes in the timing of snowmelt [*Harpold et al*., 2015; *Maurer and Bowling*, 2014]. Using soil moisture, snowpack, and air temperature observations from this network, we address two questions: (1) how do relationships between peak soil moisture (PSM) timing and the day of snow disappearance (DSD) vary across ecoregions and (2) what is the regional sensitivity of PSM timing to earlier DSD associated with warming and drying?

2. Data and Methods

Observations of snow water equivalent (SWE, made with a snow pillow), accumulated precipitation (weighing gauge), air temperature (naturally aspirated enclosure), and soil moisture were obtained from the SNOTEL network (U.S. National Resources Conservation Service). Soil moisture was measured based on soil dielectric permittivity (Stevens Hydraprobe I and II, Stevens Water Monitoring Systems, Inc.) using a standard calibration for all soil types with a measurement uncertainty of 3.4% [*Seyfried et al*., 2005]. A total of 259 SNOTEL stations were used that met the criteria of at least 5 years of daily soil moisture data through water year (WY) 2013. Of the 259 stations, 197 were forested [*Xian et al*., 2009], with the remaining in grassland, shrubland, and alpine areas. A quality assurance/quality control (QA/QC) procedure was applied to the SNOTEL data sets. First, unrealistic values were removed, including negative SWE values or soil moisture values below zero or above unity. Second, all daily soil moisture data outside of three standard deviations from the mean were removed. Third, a manual screening was performed on the soil moisture data to identify shifts, spurious trends, and other artifacts not captured by the automated procedures. Following QA/QC, the station record lengths ranged from 4 to 17 years with a total of 2159 station years.

To answer question 1, we regressed the timing of snow disappearance against the timing of peak soil moisture. The DSD was estimated as the first snow-free day at the snow pillow following annual maximum SWE. Ablation season air temperature was calculated as the average daily temperature from the day of maximum SWE to DSD. To minimize the effects of soil texture variability, we analyzed only the timing of PSM and not the magnitude of volumetric water content (VWC). The timing of PSM was determined at a resolution of 0.01 VWC and evaluated between the dates of initial snowpack accumulation and 1 September of every station year at 10 cm (PSM_{10}), 20 cm (PSM_{20}), and 50 cm (PSM_{50}) soil depths. Linear regressions were developed for the entire domain and for individual ecoregions; based on the Level III North American Terrestrial Ecoregions [*Wiken et al*., 2011] that had > 60 station years of data. The slopes of these regression equations were compared to evaluate the sensitivity of PSM timing to DSD across the different ecoregions. The R^2 values, relating DSD and PSM, were computed as one minus the ratio of the error sum of squares to the total sum of squares; p values were computed based on F statistics.

To answer question 2, we investigated the effects of changing DSD on PSM timing. Future predictions of DSD were made using historical relationships between air temperature and maximum SWE magnitude and timing. From first principles, the timing of snow disappearance is dictated by the timing of peak SWE and the length of the snowmelt period [*Trujillo and Molotch*, 2014]. The length of the snowmelt period is dictated by the magnitude of peak SWE and the snowmelt rate. Hence, we developed a multiple linear regression model (equation (1)) to predict the length of melt (days) as a function of maximum SWE (cm), date of maximum SWE (SWE_{date}), and the average daily number of positive degree days (DD) during the melt period:

$$Length\,of\quad melt = \left(a\,\frac{\mathrm{SWE}}{\mathrm{DD}} + b\,\mathrm{SWE}_{date} + c\right) \tag{1}$$

Table 1. Site, Topographic, Climate, and Soil Moisture Information From the Ten Ecoregions, All Other Stations, and All Stations[a]

	# Stations	# Station Years at 5–10 cm	Elevation (m)	Max. SWE (cm)	Postmelt *P* (cm)	Ablation Temperature (°C)	PSM_{10} (WY Day)	DSD (WY Day)	PSM_{10}-DSD (days)
Sierra Nevada	26	224	2423 (2116, 2546)	57 (33, 94)	4 (3, 9)	3.5 (1.8, 4.8)	222 (205, 247)	227 (211, 249)	−7 (−20, 5)
Blue Mountains	10	95	1666 (1580, 1776)	43 (23, 66)	13 (9, 17)	3.6 (2.1, 4.8)	220 (201, 239)	227 (207, 242)	−6 (−20, 5)
Eastern Cascades	8	69	1638 (1554, 1689)	30 (15, 46)	9 (6, 14)	2.6 (1.1, 3.7)	207 (181, 225)	212 (194, 228)	−5 (−21, 13)
Central Basin Range	19	176	2469 (2301, 2681)	31 (14, 45)	12 (6, 22)	2.1 (0.1, 4.3)	216 (200, 235)	221 (205, 235)	−5 (−13, 2)
Northern Basin Range	15	103	2108 (2011, 2167)	30 (16, 44)	12 (7, 19)	1.7 (0.1, 4)	225 (207, 240)	220 (206, 234)	−2 (−9, 7)
Idaho Batholith	9	78	2155 (1963, 2310)	36 (11, 58)	9 (6, 18)	1.9 (−1.8, 4.7)	231 (217, 245)	235 (226, 249)	−5 (−11, 1)
Southern Rockies	34	218	2976 (2851, 3322)	38 (25, 57)	17 (9, 23)	1.2 (−3, 4.5)	233 (214, 247)	234 (223, 248)	−5 (−10, 5)
Wasatch/Uintas	81	739	2679 (2438, 2901)	42 (30, 60)	14 (9, 20)	3 (0.2, 4.9)	225 (211, 241)	226 (213, 239)	−2 (−9, 6)
Middle Rockies	20	120	2437 (2183, 2722)	34 (21, 50)	14 (8, 20)	0.1 (−2.9, 4.5)	243 (227, 257)	240 (226, 255)	1 (−9, 10)
Northern Rockies	10	67	1410 (1306, 1433)	48 (16, 66)	16 (11, 23)	4.2 (1.4, 5.6)	227 (210, 244)	227 (213, 238)	−2 (−9, 8)
Other	21	264	2240 (2030, 2355)	30 (18, 63)	15 (8, 24)	3.5 (0.8, 5.1)	224 (198, 253)	214 (196, 243)	−2 (−10, 16)
All Stations	259	2153	2456 (2069, 2798)	39 (24, 60)	13 (7, 20)	2.9 (0.2, 4.8)	226 (208, 244)	227 (211, 243)	−3 (−11, 6)

[a]Elevation and climate are shown as the median and the 25% and 75% percentile in parentheses.

where *a*, *b*, and *c* are fitted parameters for each ecoregion. Equation (1) is supported by the literature in that melt length will become shorter via reduced SWE [*Trujillo and Molotch*, 2014], increased DD (i.e., melt rate), and later SWE_{date} [*Jepsen et al.*, 2012; *Trujillo and Molotch*, 2014]. It should be noted that the use of DD in equation (1) is analogous to a temperature index snowmelt model without inclusion of the melt parameter that links DD (°C) and melt rate (mm d^{-1}). It should also be noted that the melt parameter (*a*, in this case) can vary over small distances [*Kumar et al.*, 2013] and thus can be a source of regional uncertainty. The utility of temperature index models stems from the high correlation between temperature and other energy balance components [*Sato et al.*, 1984], including both shortwave and longwave radiation [*Rango and Martinec*, 1995; *Ohmura*, 2001]. By including SWE_{date} as an independent variable, the equation accounts for temporal variability in the relationship between DD and melt rate associated with differences in solar irradiance due to time of year. Thus, the negative *b* parameter found in all regions (see supporting information) suggests that the melt length is shortened with later SWE_{date}. Using equation (1), we then predicted length of melt in each ecoregion under plausible future scenarios of 1, 2, 3, and 4°C warming and 10%, 20%, 30%, and 40% reduced maximum SWE; warming and SWE reduction scenarios were evaluated independently and not concurrently. DSD was then determined for each scenario by adding the predicted length of melt to the historical mean date of peak SWE; hence, scenario estimates of DSD are conservative because shifts to earlier timing of peak SWE are not accounted for. The timing of PSM under these scenarios was then predicted using the PSM-DSD relationships described above.

3. Results

The timing and magnitude of seasonal precipitation and soil moisture was variable among and within 10 western U.S. ecoregions (Table 1). The maximum SWE was greatest in the Sierra Nevada ecoregion (median of 57 cm across all station years) and smallest (median of 30 cm) in the Northern Basin and Range. The median postmelt rainfall was 13 cm across all station years or 23% of maximum SWE on average. The median DSD was 16 May, which was earliest in the Eastern Cascades (26 April) and latest in the Middle Rockies (1 June). Consistent with DSD, the median date of PSM was 15 May, with earliest PSM in the Eastern Cascades (26 April) and latest in the Middle Rockies (1 June). A linear relationship based on DSD explained 40% of the variability in PSM_{10}, with 65% of station years having PSM_{10}-DSD deviations of <14 days (Figure 1a). The best fit slope was 0.79 days PSM_{10} per

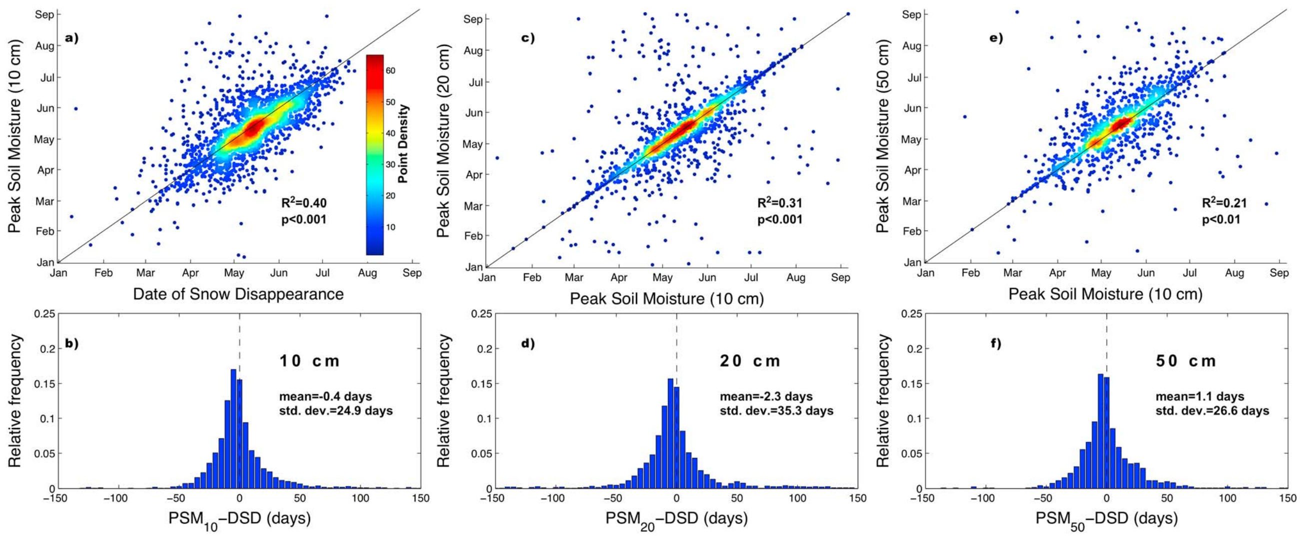

Figure 1. The relationship between peak soil moisture timing (PSM) and date of snow disappearance (DSD) for each station year at (a, c, and e) 10, 20, and 50 cm depths. The colors represent point density from 1 to >60 (black line is 1:1). (b, d, and f) The corresponding histogram of PSM-DSD deviations for the three depths are shown below (black dashed line is 1:1 relationship).

DSD, indicating that a 5 day shift in DSD resulted in a 4 day shift in PSM_{10} (Figure 1a). The timing of PSM_{10} was correlated to PSM_{20} and PSM_{50} ($R^2 = 0.31$, $p < 0.001$ and $R^2 = 0.21$, $p < 0.001$, respectively) (Figures 1b and 1c). Not surprisingly, however, the deviation of PSM and DSD increased at deeper depths, suggesting that PSM_{50} was lagged behind PSM_{10}. We focus our analysis on 10 cm soil depths (PSM_{10}) because of the larger number of station years compared to 20 and 50 cm depths.

All western U.S. ecoregions showed significant relationships ($p < 0.001$) between PSM_{10} and DSD (Figures 2a–2j), with R^2 values ranging from 0.33 (Blue Mountains, Figure 2b) to 0.66 (Central Basin and Range, Figure 2d). The corresponding slope of the best fit regression varied between 0.66 in the Blue Mountains (Figure 2b) to 0.99 in the Southern Rockies (Figure 2g), indicating significant regional variation in the sensitivity of PSM_{10} timing to DSD. The sensitivity of PSM_{10} to earlier DSD was a function of the best fit slope. For example, in the Sierra Nevada the slope was 0.95 (Figure 2a) indicating that PSM timing was quite sensitive to DSD. Conversely, the Middle Rockies had a slope of 0.75 indicating less sensitivity in PSM_{10} to DSD (Figure 2i).

Empirical relationships used to predict DSD under future conditions (equation (1)) were robust with R^2 values ranging from 0.91 to 0.97 and RMSE values of 6.1 to 11.8 days compared to historical DSD observations. While challenges arise in extrapolating these relationships to warming outside of historical conditions, the warming scenarios provided an unbiased comparison of ecoregion sensitivity. An increase of 1°C led to a 3 day earlier PSM_{10} on average, with 4 °C warming leading to between 6 and 17 days earlier PSM among ecoregions (Figure 3a). The DSD was on average 2 days earlier for each 10% reduction in maximum SWE, which resulted in 5 to 15 days earlier PSM for a 40% decline in SWE among ecoregions (Figure 3b).

Warmer ecoregions with higher accumulated degree days and lower postmelt rainfall (i.e., Sierra Nevada, Blue Mountains, and East Cascades) were most sensitive to warming and drying. These results are intuitive in that warmer regions have less ability to buffer the effects of regional warming, and wetter postmelt regions have an additional source of soil water after snowmelt. The regional sensitivity was modulated by the slope of the relationship between DSD and PSM_{10} (Figure 2); i.e., ecoregions with lower slope values had reduced sensitivity of PSM_{10} to earlier DSD. Hence, despite larger changes in DSD from 4°C warming in the Blue Mountains versus the Sierra Nevada (22 and 17 days, respectively), changes in the timing of PSM_{10} were smaller in the Blue Mountains versus the Sierra Nevada (14 and 17 days, respectively) because of the relatively lower sensitivity of PSM_{10} timing to earlier DSD in the Blue Mountains.

The most sensitive ecoregions to warming and drying (i.e., Sierra Nevada and Blue Mountains) had shifts in PSM_{10} timing that were 2–3 times larger than the least sensitive ecoregions (i.e., Northern Rockies and Central and Northern Great Basins) (Figures 3a and 3b). Ecoregions most sensitive to warming were also most

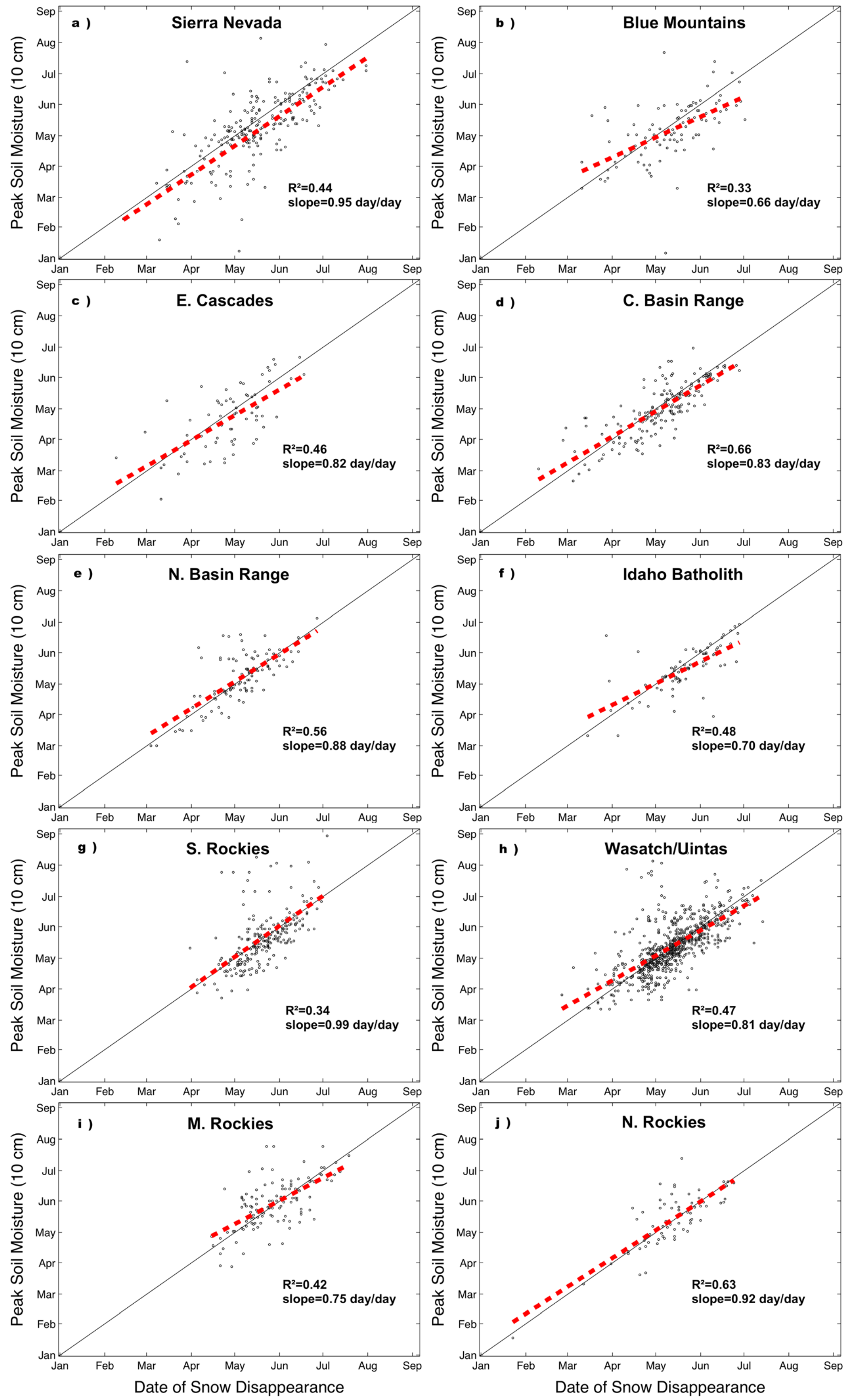

Figure 2. Best fit linear relationship (dotted red line) between PSM and DSD in the 10 ecoregions. All stations show significant relationships between PSM and DSD ($p < 0.001$). Solid line is a 1:1 relationship.

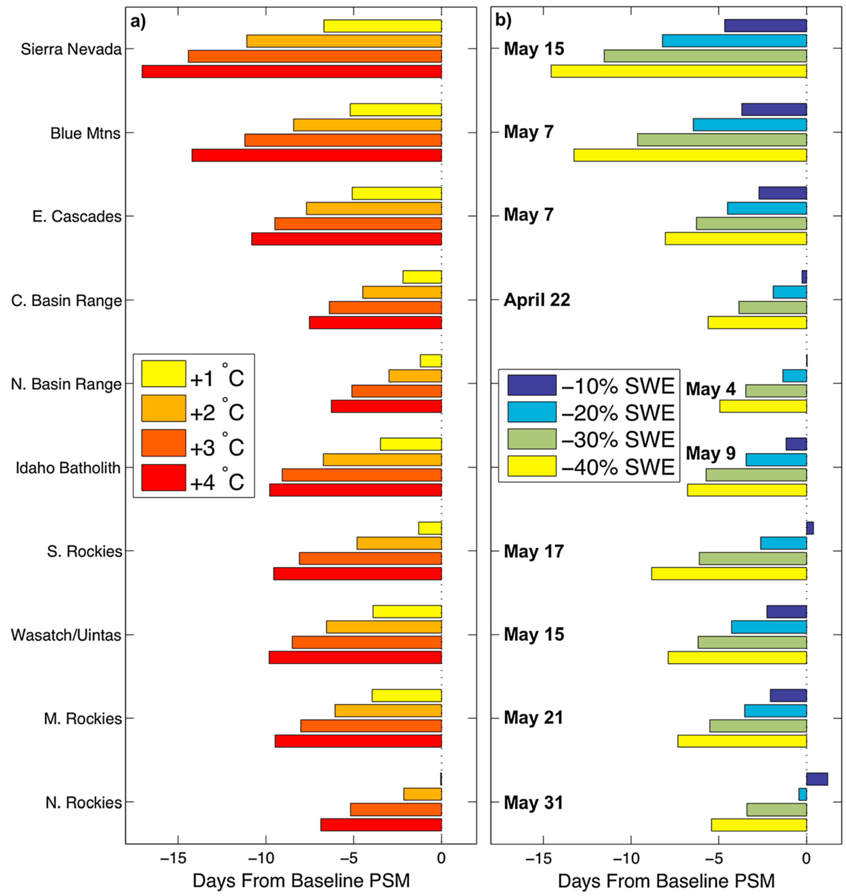

Figure 3. Effects of (a) warming and (b) drying on peak soil moisture (PSM) timing using a multiple linear regression relationship in each of 10 ecoregions. Both warming and drying resulted in earlier (negative) PSM relative to the baseline (i.e., historical mean) in nearly all cases, but the magnitude of change varied substantially across ecoregions. The historical mean PSM date is shown in Figure 3b.

sensitive to drying, which has the potential to accentuate regional differences. For example, warming of only 1°C in areas like the Sierra Nevada, Blue Mountains, and East Cascades would exacerbate changes in PSM_{10} associated with interannual variability or long-term declines in SWE accumulation (i.e., 10–40% less SWE). Conversely, regions like the Northern Rockies and Great Basin would have negligible changes in PSM_{10} timing from a 1°C warming. It is also likely that these estimates of earlier PSM_{10} may be conservative, as both reduced maximum SWE and warmer temperatures are expected to shift the beginning of melt (i.e., date of maximum SWE) earlier [*Harpold et al.*, 2012; *Kapnick and Hall*, 2012], which was not considered here.

4. Discussion

The sensitivity of PSM to earlier DSD varied across western U.S. ecoregions (Figure 2), with regional differences becoming more accentuated under warming and drying scenarios (Figure 3). The warming scenarios applied (i.e., 1–4°C) generally align with expected mid-21st century warming of 0.8–1.7°C [*Barnett et al.*, 2005] and end of century warming of 3–5°C in the western U.S. [*Leung et al.*, 2004; *Christensen et al.*, 2007]. The relatively high sensitivity of PSM_{10} timing to warming in the Sierra Nevada and Pacific Northwest, versus other parts of the western U.S., corresponds with observed spatial changes in twentieth century snowmelt [*Kapnick and Hall*, 2012] and streamflow timing [*Cayan et al.*, 2001; *Stewart et al.*, 2005]. The reduced maximum SWE scenarios (i.e., 10–40% decreased SWE) are consistent with expected decreases in snow accumulation across the western U.S. that have been estimated at 20–70% by midcentury [*Leung et al.*, 2004]. The combined effects of warming and reduced SWE on PSM timing were similar in magnitude and spatial distribution to midcentury estimates of changes in streamflow timing from hydrological simulations;

i.e., 5–30 days earlier with the largest impacts in the maritime mountains [*Hamlet et al.*, 2007; *Rauscher et al.*, 2008]. However, our results demonstrate that attributing earlier streamflow timing solely to shifts in snowmelt timing [e.g., *Cayan et al.*, 2001] neglects the role of regional-scale soil hydrology in modifying streamflow response.

Our results represent the first broad-scale, observation-based study to document the sensitivity of the timing of soil water availability to changes in climate. The observations indicate that warmer ecoregions with lower summer rainfall will have heightened sensitivity to earlier snow disappearance. Evaluation of the mechanisms causing uneven regional sensitivity will require detailed vegetation, climate, and soil properties that are not available at SNOTEL stations. In addition, the SNOTEL network is limited to relatively flat areas and does not sample gradients in slope, aspect, and elevation [*Molotch and Bales*, 2006; *Meromy et al.*, 2013]. The flat topography and lack of soil moisture transects impeded investigations into the role of lateral redistribution of water, which can have important impacts on streamflow and hydrological partitioning [*McNamara et al.*, 2005; *Tague and Peng*, 2013]. Moreover, variability in soil properties within the SNOTEL network makes an explicit comparison of soil moisture magnitudes unwise given their sensitivity to soil properties (i.e., porosity and soil water retention curves). Other locational biases in the SNOTEL network prevent analysis of areas with intermittent snow cover which may become more widespread as the climate warms; changes in PSM-DSD relationships should be expected as a consequence of multiple DSD during a given water year. The SNOTEL stations also lack observations of radiative and turbulent heat fluxes that are needed to force more complex snowmelt models (e.g., SNOWPACK; *Lehning et al.*, 2006). As a result, the work presented here did not include possible snowmelt sensitivities to light-absorbing impurities, such as dust, which strongly influence snowmelt rates in some environments [*Painter et al.*, 2007; *Skiles et al.*, 2015]; it is important to note, however, that snowmelt processes alone are not solely responsible for PSM-DSD relationships observed here. Instead, differences in infiltration and soil water storage must partially explain uneven PSM-DSD coupling across regions. This study provides an important phenomenological response to changing snowpack; however, fully understanding the associated ecohydrological implications will require hillslope-scale soil moisture and soil property observations and modeling across western U.S. ecoregions [*Williams et al.*, 2009; *Bales et al.*, 2011]. Development of new soil moisture observations from emerging in situ networks [*Larson et al.*, 2008; *Zreda et al.*, 2012] and remote sensing [*Dubois et al.*, 1995; *Entekhabi et al.*, 2010] hold promise for identifying the mechanisms driving regional differences in PSM-DSD relationships.

In addition to providing new insights into soil moisture sensitivity to climate change, our findings elucidate the processes delivering water to streams and forest ecosystems. We found stronger coupling ($p < 0.10$) between the timing of PSM_{10} and DSD at individual stations (124 of 259 stations) relative to comparisons of peak SWE and average summer soil moisture (16 of 254 stations) performed by *Maurer and Bowling* [2014]. This suggests that the timing of soil water availability may be more predictable than the magnitude of soil water availability with respect to future climate warming. Our results support the findings of *Maurer and Bowling* [2014] that deep soil moisture exhibits significant sensitivity to snowmelt timing and magnitude, which was contrary to *Blankinship et al.* [2014] who found little sensitivity of deep soil moisture to variable snowmelt. It is important to note that summer rainfall complicated these snowmelt-soil moisture relationships, as 19% of station years analyzed here had PSM_{10} occurring >2 weeks after DSD. We infer that earlier DSD has the potential to reduce antecedent wetness conditions and thus alter the role of summer rainfall in streamflow generation, e.g., reducing hydrological connectivity [*McNamara et al.*, 2005; *Tague and Peng*, 2013] and groundwater recharge [*Huntington and Niswonger*, 2012; *Godsey et al.*, 2014].

Our results have significant implications for how changes in climate will impact vegetation phenology, productivity, and transpiration. Earlier snowmelt and associated alterations to the timing of soil water availability will shift the onset of the growing season earlier in the year [*Hu et al.*, 2010; *Sacks et al.,* 2007; *Cayan et al.*, 2001]. The predictions that PSM_{10} timing will advance by 1.6–4.3 days per 1°C of warming were consistent with observed earlier bloom dates of 4.5 days for lilac and honeysuckle and 5.5 days for subalpine meadow species per 1°C of warming found during the later part of the twentieth century in the western U.S. [*Cayan et al.*, 2001; *Dunne et al.*, 2003]. Vegetation phenology response to changing snowmelt timing is complex, however, with species-level differences, elevation range, frost damage, magnitude of soil moisture, and soil temperature all exerting some influence on the timing of flowering [*Sparks and Carey*, 1995; *Inouye et al.*, 2002; *Dunne et al.*, 2003]. Both ground-based and remote sensing observations indicate that gross primary productivity and greenness of mountain forest ecosystems increases during larger snowpack years, despite shorter growing seasons [*Hu et al.*, 2010; *Trujillo et al.*, 2012]. As a result, late summer forest transpiration and productivity will likely be reduced unless water availability is maintained by soil water storage or summer rainfall [*Hu et al.*, 2010; *Scott-Denton et al.*, 2013; *Parida and Buermann*, 2014]. Therefore, water limitations on forest transpiration and productivity will be particularly acute in ecosystems with thin, well-drained soils, shallow-rooted plants, and low summer precipitation.

Our results add mechanistic explanation to the growing body of literature suggesting regional warming causes increased drought stress in forests [*Allen et al.*, 2010; *Williams et al.*, 2013] and vulnerability to disturbance [*Lenihan et al.*, 2003; *Westerling et al.*, 2006]. A shift toward earlier PSM has important implications for species defenses to common tree pathogens [*Kaiser et al.*, 2013; *Hart et al.*, 2014] and increasing fuel for severe forest fires [*Lenihan et al.*, 2003; *Westerling et al.*, 2006]. *Westerling et al.* [2006] showed that years with earlier snowmelt (e.g., 5 days earlier than average) had 3–10 times greater frequency of high-intensity forest fires relative to years with late snowmelt at the elevations examined here. Our empirical model indicated that a 1°C warming would shift PSM_{10} 5 days earlier in the most sensitive ecoregions (i.e., the Sierra Nevada and Blue Mountains), and a 3°C warming would lead to >5 days earlier PSM_{10} in all ecoregions (Figure 3a). Consequently, the degree of soil moisture-snowmelt coupling represents an important mechanism to identify high-elevation forests at risk for increased disturbance as a consequence of changes in regional climate.

5. Conclusions

Using 2159 station years of paired soil moisture and snow observations from across the western U.S., we showed that the DSD explains 40% of the variability in PSM_{10}. On average, a 5 day earlier DSD led to a 4 day earlier PSM_{10}. PSM_{10} shifted between 3.3 and 5.0 days earlier per 5 day change in DSD for the 10 western U.S. ecoregions considered. Empirical relationships used to estimate future DSD and PSM under scenarios of 1–4°C air temperature increases showed that PSM timing occurs 1.6–4.3 days earlier for each 1°C of warming. Scenarios evaluating PSM timing with 10%–40% reductions in maximum SWE indicated that PSM timing occurs 1.2–3.6 days earlier with each 10% reduction in maximum SWE. Estimated changes in the timing of PSM based on warming estimated for the middle and end of the 21st century ranged from 1 to 9 days and from 6 to 17 days among ecoregions, respectively. Thus, warming temperatures could substantially exacerbate peak soil moisture variations due to typical interannual snowpack variations, as well as long-term changes in snow accumulation. Due to differences in winter air temperature and summer rainfall, maritime ecoregions (e.g., the Sierra Nevada, Blue Mountains, and Eastern Cascades) were 2–3 times more sensitive to warming and drying scenarios than interior mountain ranges (e.g., the Middle and Southern Rockies). Our observations suggest that soil hydrological processes can both exacerbate and moderate changes in snowmelt timing. This new information requires that mechanistic explanations for uneven regional streamflow response to warming expand from only focusing on snowmelt variability [e.g., *Cayan et al.*, 2001] to more fully consider the role of soil hydrology [*Hamlet et al.*, 2007; *Koster et al.*, 2010]. Consequently, our predictions of future changes in PSM timing associated with warming and drying have new and broad implications for runoff generation mechanisms, forest productivity and phenology, and vulnerability to disturbance.

Acknowledgments
Randall Julander provided insights regarding the SNOTEL data. The SNOTEL data sets are publicly available at http://www.wcc.nrcs.usda.gov/snow/. NSF post-doc fellowship (EAR 1144894) supported A.A. Harpold. Awards from U.S. Department of Agriculture (2012-67003-19802) and NSF (EAR 1141764) supported N.P. Molotch. NSF award (EAR 1331828) Boulder Creek Critical Zone Observatory supported both authors.

The Editor thanks Jeff Dozier and an anonymous reviewer for their assistance in evaluating this paper.

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