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Discussions

Partitioning snowmelt and rainfall in the critical zone: effects of climate type and soil properties

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Abstract

Streamflow generation and deep groundwater recharge in high elevation and high latitude locations may be vulnerable to loss of snow, making it important to quantify how snowmelt is partitioned between soil storage, deep drainage, evapotranspiration, and runoff. Based on previous findings, we hypothesize that snowmelt produces greater streamflow and deep drainage than rainfall and that this effect is greatest in dry climates. To test this hypothesis we examine how snowmelt and rainfall partitioning vary with climate and soil properties using a physically based variably saturated subsurface flow model, HYDRUS-1D. To represent climate variability we use historical inputs from five SNOTEL sites in each of three mountain regions with humid to semiarid climates: Northern Cascades, Sierra Nevada, and Uinta. Each input scenario is run with three soil profiles of varying hydraulic conductivity, soil texture, and bulk density. We also create artificial input scenarios to test how the concentration of input in time, conversion of snow to rain input, and soil profile depth affect partitioning of input into deep drainage and runoff. Results indicate that event-scale runoff is higher for snowmelt than for rainfall due to higher antecedent moisture and input rates in both wet and dry climates. At the annual scale, surface runoff also increases with snowmelt fraction, whereas deep drainage is not correlated with snowmelt fraction. Deep drainage is less affected by changes from snowmelt to rainfall because it is controlled by deep soil moisture changes over longer time scales. However, extreme scenarios with input highly concentrated in time, such as during melt of a deep snowpack, yield greater deep drainage below the root zone than intermittent input. Soil texture modifies daily wetting and drying patterns but has limited effect on annual scale partitioning of rain and snowmelt, whereas increases in soil depth decrease runoff and increase deep drainage. Overall these results indicate that runoff may be substantially reduced with seasonal snowpack decline in all climates. These mechanisms help explain recent observations of streamflow sensitivity to changing snowpack and emphasize the need to develop strategies to mitigate impacts of reduced streamflow generation in places most at risk for shifts from snow to rain.

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1 Introduction

Snowmelt is the dominant source of streamflow generation and groundwater recharge in many high elevation and high latitude locations (Regonda et al. 2005; Stewart et al. 2005; Earman et al., 2006; Clow, 2010; Jefferson, 2011; Furey et al., 2012). Soils modulate the partitioning of snowmelt into subsurface storage, deep drainage, evaporative losses and surface runoff. Snow persistence shows declines around the globe (Hammond et al., 2018b), and these snow losses may lead to changes in water input magnitude and timing (Harpold et al., 2015; Harpold et al., 2017). As areas of "at risk snow" become more apparent (Nolin and Daly, 2006), there is an urgent need for mechanistic studies that quantify the partitioning of snowmelt in the critical zone among vapor losses, surface flow, and subsurface flow and storage (Brooks et al., 2015; Meixner et al., 2016).

Changes in precipitation phase from snow to rain can modify hydrological partitioning by altering the timing and rate of inputs. Snowmelt rates may not reach the extreme intensities of rainfall (Yan et al., 2018), but unlike rainfall, which is typically episodic, snow can accumulate over time, then melting as a concentrated aggregate of soil water input (Loik et al., 2004). Such concentrated snowmelt events can lead to a large proportion of runoff and deep drainage (Earman et al., 2006; Berghuijs et al., 2014; Li et al., 2017). With climate warming, future snowmelt rates may be reduced in many areas because earlier melt occurs when solar radiation is lower (Musselman et al., 2017). Along with warmer temperatures, increasing atmospheric humidity is leading to more frequent and greater magnitude mid-winter melt events in humid regions, and increased snowpack sublimation and/or evaporation in dry regions (Harpold and Brooks, 2018). Some areas (i.e. the Cascades) are predicted to receive more intense water inputs with rainier futures, whereas others (i.e. Southern Rockies) will likely experience declines in input intensity with snow loss (Harpold and Kohler, 2017). Given the considerable heterogeneity in the factors that affect hydrological partitioning, such as climate, soils, topography, and vegetation, different locations may not respond in the same way to loss of snow.

Water inputs from rain or snowmelt during periods of low potential evapotranspiration and higher antecedent moisture conditions are more likely to generate runoff and deep drainage (Molotch et al., 2009). Prior research has shown that near-surface soil moisture response is closely related to snow disappearance (Harpold and Molotch, 2015; Webb et al., 2015; Harpold et al., 2015) with strong links between snowmelt and soil moisture dynamics at multiple spatial and temporal scales (Loik et al. 2004; Williams et al. 2009; Blankinship et al. 2014; Kormos et al., 2014; Harpold and Molotch, 2015; Webb et al. 2015; Kampf et al. 2015). Earlier snow disappearance can lead to lower average soil moisture conditions not as conducive to streamflow generation as later snowmelt (Kampf et al. 2015). The effects of earlier snowmelt on soil moisture dynamics may also vary with precipitation after snowmelt. Late-spring precipitation can overwrite the signal of earlier snowmelt timing on spring and summer soil moisture (Liator et al., 2008, Conner et al., 2016), whereas a lack of spring and summer precipitation can cause effects of earlier snowmelt on soil moisture to persist longer (Blankenship et al, 2014; Harpold, 2016). Earlier snow disappearance can lead to diverging patterns in growing season length; a longer growing season if energy hinders

vegetation growth, or a shorter growing season when soil water stress limits productivity (Harpold et al., 2015; Harpold, 2016; Hu et al., 2010).

Both runoff and deep drainage are affected by soil texture, soil depth, rooting depth (Cho and Olivera, 2009; Seyfried et al., 2005) and topography. These properties have limited variability over timespans of hydrologic analysis and can produce temporally stable spatial patterns of soil moisture, where some parts of the landscape are consistently wetter than others (Williams et al., 2009; Kaiser and McGlynn, 2018). Aspect modifies the snowpack energy balance, leading to higher sustained moisture content on north-facing slopes compared to south-facing slopes with the same input (in the northern hemisphere); landscape evolution due to wetter conditions on north-facing slopes may lead to deeper profiles and more deeply weathered rock conducive to deep drainage in some locations (Hinckley et al., 2014; Langston et al., 2015). Where soils are shallow, winter precipitation may exceed the soil storage capacity, leading to both runoff generation and deep drainage (Smith et al., 2011). Deeper soil profiles have greater storage capacity, which can sustain streamflow, even with snow loss; however, given consecutive years of low input these profiles will be depleted of storage leading to lower flows (Markovich et al., 2016). Deeper soils can also help sustain transpiration during the late spring and summer, when shallow soils have already dried (Foster et al. 2016; Jepsen et al., 2016). Streamflow can be insensitive to inputs under dry conditions, but respond rapidly once a threshold soil moisture storage value is exceeded (McNamara et al., 2005; Liu et al., 2008; Seyfried et al., 2009). McNamara et al. (2005) hypothesized that when dry-soil barriers are breached, there is sudden connection to upslope soils, leading to delivery of water to areas that were previously disconnected. In their semi-arid study area, such breaching of dry-soil barriers was only observed for periods of concentrated and sustained input from high-magnitude spring snowmelt. Together the complex interactions of soil properties, antecedent conditions, water inputs, and evaporative demand make it challenging to determine how changes from snow to rain affect hydrologic response even in idealized settings.

The goal of this study is to improve our understanding of how changes in precipitation phase from snow to rain affect hydrological partitioning in a one-dimensional (1-D) representation of the critical zone. Partitioning of precipitation input, P, can be into runoff, Q, defined as lateral export of water from the domain; evaporation, E; transpiration, T; deep drainage below the root zone, D; and storage within the soil root zone, ΔS. Over a given time increment, partitioning can be tracked using the water balance (equation 1).

$$P = Q + E + T + D + \Delta S \quad (1)$$

We address the questions: (1) Are snowmelt and rain partitioned differently between Q, ET, and D? and (2) How is snowmelt and rain partitioning affected by climate, soil type, and soil depth? We use a physically-based 1-D modeling approach to address these questions and systematically test hypotheses about hydrologic response to snow loss.

We hypothesize that reducing the fraction of precipitation falling as snow leads to lower runoff and deep drainage because it reduces the concentration of input in time (Figure 1). Concentrated input during melt of a seasonal snowpack often saturates soils, causing saturation excess runoff and deep drainage below the root zone (Hunsaker et al., 2012; Kampf et al., 2015; Webb et al., 2015; Barnhart et al., 2016). Diffuse input over time reduces the likelihood of saturation because it allows time for water redistribution and evapotranspiration between inputs. We also hypothesize that snowmelt is critical for runoff generation and deep drainage in dry climates and deep soils, where snowmelt is the dominant cause of soil saturation (McNamara et al., 2005; Tague and Peng, 2013), whereas the partitioning of rain and snowmelt may be more similar in wet climates and shallow soils, which are more frequently saturated by either rain or snowmelt inputs (Loik et al., 2004) (Figure 1).

2 Methods

To evaluate soil moisture response to rainfall and snowmelt over a wide range of climate and soil conditions we used HYDRUS-1D (Šimůnek et al. 1998), a physically-based finite element numerical model for simulating one-dimensional water movement in variably saturated, multi-layer, porous media.

2.1 Study design, site selection, and data sources

We utilized daily input data from five United States Department of Agriculture Natural Resources Conservation Service (NRCS) snow telemetry (SNOTEL) sites in each of three regions that span a wide range of climate and snow conditions: the Cascades, Sierra Nevada, and Uinta mountains for a total of 15 sites. Daily rather than hourly data were chosen to limit the effects of missing and incorrect values on the analyses. The three regions chosen to represent dominant climate types in the western U.S., and within each region, sites were selected to span a snow persistence (SP) gradient, which is the mean annual fraction of time that an area is snow covered between Jan 1 and Jul 3 (Moore et al., 2015) over the ~35 years of record (Figure 2a, Table 1).

With each climate scenario we simulated vertical profiles of volumetric water content (VWC), which were depth-integrated to compute soil moisture storage (S). For these simulations deep drainage (D) is any flux of water downward below the root zone. Runoff (Q) is any water that does not infiltrate into the soil, either because of saturated conditions or because input rates exceed infiltration capacity. Evaporation (E) is direct evaporation from the soil, and transpiration (T) is mediated by plant roots; for this study, these values are combined into evapotranspiration (ET) to represent the bulk loss of water to the atmosphere.

Daily precipitation (P), snow water equivalent (SWE), and volumetric water content (VWC) at 5, 20, and 50 cm were obtained for each SNOTEL site using the NRCS National Weather and Climate Center (NWCC, 2016) Report Generator (Table 1). The records were quality controlled to ensure reasonable precipitation, SWE and VWC values as in Harpold and Molotch (2015). Unrealistic values were removed (i.e. negative SWE, VWC below zero or above unity); all daily VWC outside of three standard deviations from the mean were removed, and a manual screening

was performed on VWC data to identify shifts and other artifacts not captured by the first two automated procedures. Daily potential evapotranspiration (PET) was extracted from daily gridMET (Abatzoglou, 2013) for the 4 km pixel containing each SNOTEL site. This product uses the ASCE Penman-Monteith method to compute PET.

We chose three SNOTEL sites (432 Currant Creek, 698 Pole Creek R.S., 979 Van Wyck) to represent soil profile characteristics. While 365 of the 747 SNOTEL sites in the western U.S. have soil moisture sensors, only a fraction of these sites have detailed soil profile data. The sites with soil profile data have information obtained from soil samples taken in the soil pits and processed in the NRCS Soil Survey Laboratory in Lincoln, NE for texture, water retention properties, and hydraulic conductivity. We obtained detailed soil profile data, in the form of pedon primary characterization files from the NRCS, and selected three profiles (Figure 2b, Table 2) that represent the range of soil textures and hydraulic conductivity values present at SNOTEL locations. Each had detailed NRCS pedon primary characterizations to depths greater than 100 cm and >15 years of daily soil moisture records at 5, 20 and 50 cm depths.

2.2 Simulations

In HYDRUS-1D, we simulated water flow and root water uptake for a vertical domain 10 m deep. The domain was discretized into 500 nodes with higher node density near the surface (~0.15 cm for top 5 cm to ~5 cm for the bottom of the profile). For the surface boundary, we used a time variable atmospheric boundary condition, which allows specifying input (snowmelt and rain) and potential evapotranspiration (PET) time series. Runoff can also be generated at the surface boundary. For the lower boundary, we allowed free drainage from the bottom of the soil profile at 10 m. Surface soil water input was calculated by totaling snowmelt and rainfall input at the daily time step from SNOTEL precipitation and SWE values. Melt was computed for any day when SWE decreased; if SWE decreased, and the precipitation was greater than 0, total soil water input was assumed to be melt plus precipitation. The atmospheric boundary condition requires PET, leaf area index (LAI), and a radiation extinction coefficient used in the estimation and separation of potential evaporation and transpiration. We assigned a constant LAI of three, as this value generally fits the mixed conifer forests (Jensen et al., 2011) where SNOTEL sites are installed and assumed a radiative extinction coefficient of 0.39, which is the default value. Root water uptake in the model was estimated using Feddes parameters for a conifer forest (Lv, 2014: h1 0 cm, h2 0 cm, h3h -5,100 cm, h3l -12,800 cm, h4 -21,500 cm, T_{Plow} 0.5 cm/d, T_{Phigh} 0.1 cm/d)), with roots uniformly distributed from the soil surface to the interface with a lower hydraulic conductivity layer, as we lacked any more detailed information on root distribution or soil depth at these sites.

We created soil layers with depths and textures taken from the NRCS soil pedon measurements. From this information we applied the neural network capability of HYDRUS-1D, which draws from the USDA ROSETTA pedotransfer function model (Schaap et al., 2001), to determine soil hydraulic parameters. Using the NRCS pedon primary characterizations we input percent sand, silt and clay, bulk density, wilting point, and field capacity; the neural network model estimates soil hydraulic parameters based on these inputs. Below the depth of the soil pedon

measurements, we configured the simulations to have a deep "bedrock" or regolith layer with lower saturated hydraulic conductivity (Ks) but the same water retention parameters as the layer above. Any water entering this lower layer is considered deep drainage. The hydraulic conductivity of this lower layer was set at one tenth that of the layer above; this value was determined through iterative testing of Ks values (see Supplementary). We extended the "bedrock" or regolith layer to 10 m depth to allow for deep drainage to occur without boundary effects that could be caused by a shallower regolith. The initial VWC for all layers in each simulation was 0.2, and simulations were run with a year of surface boundary condition inputs to establish initial conditions. We tested the simulation configuration by comparing to observed VWC at 5, 20 and 50 cm depths for the three selected soil profile sites (Figure S1, Table S1). Rather than force-fitting, our goal was to produce simulations with similar timing of wetting and drying to observations. This approach is consistent with other studies using HYDRUS – 1D, which also started with basic soils data and application of the ROSETTA pedotransfer function (Scott et al., 2000) and then calibrated to observed water content measurements by adjusting permeability of the "bedrock" layer (Flint et al., 2008).

We applied climate scenarios from each of the 15 SNOTEL sites selected (Table 1) to each of the soil profiles to examine how climate and soil type affect partitioning. We then conducted additional experiments to modify inputs using just the loam profile. First to examine whether snowmelt and rainfall are partitioned differently, we changed all precipitation to rain and compared the simulation output to those with the original climate data. Second, to examine the effects of input concentration, we artificially produced intermittent input (four five-day periods of low magnitude) and concentrated input (one twenty-day period of high magnitude) of the same annual total for one wet (559) and one dry (375) site using the loam profile (1056) for all years of data. Third, to examine how soil depth affects partitioning we altered the depth of rooting zones to 0.5, 1.5 and 2 times their original depth. For 0.5 depth scenarios, we replaced soil layers deeper than 0.5 times the original depth with the bedrock/regolith layer. For 1.5x and 2x scenarios, the layer above bedrock/regolith was extended downward, and the rooting zone extended to the new soil depth.

2.3 Analysis

Using the simulation results, we examined how rain and snowmelt were partitioned into soil storage (S), deep drainage (D), evapotranspiration (ET), and runoff (Q). Daily soil storage is reported as the total soil water within the rooting zone only, and D is any water passing below the rooting zone (106-127 cm depending on the soil profile). We assessed partition components both in units of length (cm) and as ratios to total input (unitless, e.g. Q/P) at both event and annual time scales.

To analyze hydrologic partitioning at event time scale we defined rainfall events as days with precipitation while SWE equaled zero and snowmelt events for days with declining SWE and no simultaneous precipitation. To focus on differences between rainfall and snowmelt, only events with entirely rainfall or entirely snowmelt input were considered in this analysis; mixed events were excluded. Events could last as long as the conditions were

continuously satisfied, and only those followed by at least five days of no input were used in analysis. Total depths of each variable were computed for each defined event time period; input rain and snowmelt were summed over the event time period, and response variables (Q, ET, D) also included the day after the event ended to account for lag in event response. Antecedent S for each event was determined by taking the root zone storage from the day prior to the first event input.

At the annual scale, soil water input and partitioning components (rain, snowmelt, Q, ET, D) were totaled for each year, and the change in water year storage (ΔS) determined by subtracting the values of S at the end of the year from the value at the beginning of the year. In addition to ΔS, mean saturation (Sat) at each observed depth was calculated as the average annual VWC divided by soil porosity. We use mean saturation (Sat) as an alternative to change in water year storage (ΔS) because mean saturation is much easier to quantify at a field site than root zone storage, and this extends the application of our study to other areas with daily VWC data. Sat also provides a measure of soil water conditions throughout the year as opposed to ΔS which represents only changes between the start and end of the water year.

To characterize climate conditions at the mean annual scale, each site was classified as dry (precipitation deficit, PET>P) and wet (precipitation surplus, PET<P). This separation by aridity index is based on our hypothesis that the influence of concentrated snowmelt is greater in dry climates than in wet climates (Hammond et al, 2018a). We also report the maximum SWE and snowmelt fraction as the annual total snowmelt divided by annual total input. Following the methods for computing the precipitation concentration index (PCI), which represents the continuity or discrete nature of input through time (Martin-Vide, 2004; Raziei et al., 2008; Li et al., 2011), we computed the input concentration index (ICI) using snowmelt and rain input. Pearson correlation tests were conducted between explanatory variables (P, PET, P/PET, peak SWE, average melt rate, and ICI) and dependent variables (Q, ET, D, mean saturation at 100 cm: Sat100).

Using both the event and annual results, we examined (1) whether partitioning of rainfall input differed from that of snowmelt input, and (2) how partitioning was affected by climate, soil texture, and soil depth. For question 1, we tested for differences in event partitioning between input type (rain or snowmelt) and differences in annual partitioning between historical and all rain scenarios using ANOVA. For question 2, we tested for differences in annual partitioning between climate (wet, dry) and soil depth groupings, also using ANOVA. Additionally for question 2, we tested the pairwise difference in linear regression slopes using indicator-variable regression with interaction in JMP (SAS-based statistical software) to determine whether the rate of change between explanatory and response variable differed by climate or soil depth grouping.

3 Results

Simulations for each of the 15 climate scenarios exhibit substantial variability at the annual scale in precipitation (P), runoff (Q), and deep drainage (D) (Figure 3). All regions have a wide range of annual P, but overall the highest P was in the Cascades region and lowest in the Uinta. The wide range of climate conditions simulated allows for an evaluation of climate effects on Q, ET, D, and Sat100 (Table S3). Annual precipitation (P) is positively correlated with runoff (Q, r=0.97), deep drainage (D, r=0.92), and Sat100 (r=0.73) (Table S3). The relationship is linear for Q but nonlinear for D and Sat100. Sat100 plateaus at ~250 cm P with further P partitioned to Q instead of D. Evapotranspiration (ET) has the weakest correlations with P (r=0.08) of all partitioned components. Q/P increases with P up to around 250 cm of P, and D/P increases with P up to around 100 cm (Figure 3). ET/P decreases with precipitation, whereas S/P is unrelated to P. At values of P greater than around 300 cm, all variables have relatively consistent values even as P increases.

3.1 Snowmelt vs rainfall and climatic influences on partitioning

Our first research question asks whether snowmelt and rainfall are partitioned differently. At the event scale, input rates are significantly greater on average for snowmelt than for rainfall in each of the three regions and for the full dataset (ANOVA $p<0.0001$, mean snowmelt 1.1 cm/d, mean rainfall 0.9 cm/d, Figure 4), though rainfall events have a higher maximum input rate (maximum snowmelt 8.0 cm/d, maximum rainfall 14.7 cm/d). Snowmelt events tend to occur on wetter soils, as estimated by antecedent soil moisture storage for the rooting zone (ANOVA $p<0.0001$, mean S for snowmelt 56.6 cm, mean S for rainfall 48.2 cm). Average runoff ratios (Q/P) are higher for snowmelt than for rainfall (ANOVA $p<0.0001$, mean Q/P snowmelt 0.20, mean Q/P rainfall 0.03), whereas ET/P is lower for snowmelt as compared to rainfall (mean snowmelt 0.24, mean rainfall 0.40). Deep drainage responses are affected by longer time scales than single events, so we did not include these in the event analysis.

At the annual scale, input at all sites is a mixture of rain and snowmelt. To examine the importance of snow to partitioning, we used snowmelt fraction, defined as the fraction of snowmelt to total precipitation, and input concentration index (ICI). Snowmelt fraction and snow persistence are generally positively correlated with ICI at dry sites in the Uinta and Sierra, but this correlation declines with wetter sites in the Cascades (Figure S7). This indicates that areas with greater snowmelt tend to have greater input concentration in dry climates. Q/P increases with snowmelt fraction (r=0.41), most noticeably where snowmelt fraction is >0.5 and increases with ICI (r=0.80) (Figure 5). The ranges of Q/P are higher in wet than in dry climates, though dry climates show greater rates of change with increasing snowmelt fraction and input concentration (Table S4). D/P is somewhat correlated with snowmelt fraction (r=0.20) and ICI (r=0.43). D/P ranges are higher in wet than in dry climates, with many dry years not generating D. ET/P is not related to snowmelt fraction and generally declines with ICI (r = -0.75); ranges are lower for wet climates, where greater input is partitioned to Q and D.

We then compared the hypothetical scenarios where we treated all precipitation as rain to snow-dominated historical scenarios. All rain leads to significantly lower Q/P ($p<0.0001$, all rain mean 0.17; historical mean 0.31) for both wet and dry sites (Table 3, Figure 6). This partly relates to lower near-surface saturation in all rain scenarios; the mean fraction of annual runoff from saturation excess is 88% when all input is rain as compared to 97% with historical rain and snow input. All rain also leads to higher ET/P for dry sites ($p<0.0001$, all rain mean 0.95; historical mean 0.83); lower D/P for dry sites (all rain mean 0.01; historical mean 0.03), and higher D/P at wet sites ($p=0.011$, all rain mean 0.14; historical mean 0.12) (Table 3, Figure 6).

Another effect of snow loss can be a decrease in input concentration because snow melt concentrates input in a short period of time. Experimental scenarios with constant P separated into intermittent and concentrated inputs for a wet site (375) and a dry site (559) show that increasing input concentration leads to significantly greater Q/P in the dry site ($p<0.05$, intermittent mean 0.54, concentrated mean 0.68, Table 3, Figure 6) but no significant difference in the wet site. In contrast, D/P is significantly greater ($p<0.0001$) for the concentrated input scenarios for both dry and wet sites, as no deep drainage is produced with intermittent input. ET/P is significantly lower in concentrated input scenarios, with a greater difference in dry climates ($p=0.004$, mean intermittent 0.80 vs. concentrated 0.66) than in wet climates ($p=0.013$, mean intermittent 0.34 vs. concentrated 0.28).

3.2 Soil property influences on partitioning

Soil stores water that may later be partitioned into Q, ET, and D. Using Sat100 as an indicator of soil moisture storage, Figure 7 displays the relationships between Q/P, D/P and ET/P vs Sat100 as separated by climate type, soil texture, and root zone depth. Sat100 has strong relationships with Q/P, D/P, and ET/P for all, wet, and dry sites (Figure 7, Table S5). Q/P is generally low (Figure 7a, <0.3) until Sat100 is greater than >0.5. D/P in the simulations also increases with Sat100, and many simulation years have limited D when Sat100 <0.5. ET/P generally decreases with saturation for Sat100 values >0.5.

When these same relationships are separated by soil texture rather than wet/dry climate (Figure 7b, Table S5), the response patterns are similar between soil types except for the sandy loam profile, which displays higher Q/P and D/P than the loam and sandy clay loam profiles at similar Sat100 levels. Differences between responses by soil texture are more evident at sub-annual time scales (Figure 8a). For the example time period shown in Figure 8a, loam and sandy clay loam profiles wet up each spring during snowmelt prior to the sandy loam profile, and their higher water retention means they can generate deep drainage earlier and longer than sandy loam. However, sandy loam has higher Ks, which allows a greater rate of deep drainage during time periods of saturation. Consequently, the differences in deep drainage between soil textures are limited (Figure 6), except that annual D/P for sandy loam is higher than for sandy clay loam and loam profiles when Sat100 values are low. The latter soils retain more water, so they ultimately reach the highest annual D/P values at higher Sat100 values. More water retention in the sandy clay loam and loam soils can lead to more runoff generation via saturation excess, whereas the drier conditions in

sandy loam can lead to infiltration excess runoff. However, the net differences in annual total runoff are limited between the three soil textures (Figure 6).

To assess the influence of soil profile depths on partitioning, we altered the loam soil profile to be 0.5x, 1.5x and 2x times its original depth (Figure 6, Table 3). For historical input, Q/P and D/P are greatest for the 0.5x depth scenario, and Q/P declines significantly with deeper soils for both dry and wet sites ($p<0.0001$), with the greatest declines between 0.5x and 1x (original) depth. D/P declines significantly between 0.5x and 1x depth, then increases slightly for all sites with subsequent increases in depth to 1.5x and 2x (Figure 6, Table 3). Q/P and D/P differences by depth are significant between 0.5x and 1x depth, but not for all subsequent depth comparisons for all, wet and dry site classifications (Table 3). In pairwise comparisons between depth scenarios Q/P is only significantly different between 0.5x and 1x depth categories ($p <0.0001$). Changes in ET/P with soil depth are not significant according to ANOVA tests.

Figure 8b displays daily time series of surface runoff, deep saturation, deep drainage, and cumulative deep drainage during an example period for the four different soil root zone depth scenarios. The shallowest rooting zone of 0.5x original depth displays the greatest surface runoff as well as cumulative deep drainage throughout the example period. Each depth reaches and remains at saturation for different amounts of time, with the shallowest profile reaching saturation earliest and remaining saturated longest, but also decreasing more rapidly to the lowest ending saturation. The deepest profile takes the longest to increase Sat100, not reaching as high a peak, yet remaining higher at the end of the period. Deep drainage occurs earliest for the shallowest depth scenario, though reaching a lower daily flux than the original depth. Deep drainage from the 1x 1.5x and 2x original depth scenarios lag behind the 0.5x scenario following the same succession as their Sat100 patterns. These patterns in daily Sat100 and deep drainage result in the highest cumulative deep drainage for the shallowest scenario.

4 Discussion

4.1 Snowmelt as an efficient runoff generator and factors accentuating snowmelt efficiency

The initial hypotheses for this study were that runoff and deep drainage would be greater from snowmelt than rainfall. Multiple lines of evidence from our 1-D hydrologic simulations point towards snowmelt as a more efficient driver of runoff, and to a lesser extent deep drainage, than rainfall. Results confirmed that runoff efficiency from snowmelt events was elevated because snowmelt events were 22% greater in input rate, and occurred on 17% wetter soils than rainfall. This stands in agreement with previous studies showing that snowpack development and subsequent melt tend to occur when soils are at elevated moisture contents due to lower ET (Liu et al., 2008; Williams et al., 2009; Bales et al., 2011). Whether input is snowmelt or rainfall becomes less important for hydrologic response at annual times scales, for which the correlation between snowmelt fraction and response variables is weak to moderate (Figure 5, Table S3). When input scenarios are forced into the extreme case of all rain, they show a lower annual Q/P (Dry: 0.13 vs. 0.04; Wet: 0.46 vs. 0.29), corroborating the event results that indicate snowmelt elevates runoff (Table 3, Figure 6). We also hypothesized that the effects of changing snowpacks would

be greatest in dry climates, where soil saturation is less frequent. However, evidence suggests that both wet and dry climates are likely to show reduced surface runoff with declining snow water inputs at the 1-D scale.

We had hypothesized based on prior research (Hunsaker et al., 2012; Langston et al., 2015; Barnhart et al., 2016; Li et al., 2017; Hammond et al., 2018a) that input concentration would be the primary reason for elevated Q and D from snowmelt relative to rainfall. While ICI is correlated with increasing Q/P and D/P (Figure 5, Table S3), snowmelt is not consistently the cause of greater input concentration in the wetter sites (Figure S7). When input is extremely concentrated, as in the hypothetical experiments (Table 3, Figure 6) D/P increases significantly in both wet and dry climates, whereas Q/P only increases significantly with input concentration in dry climates. Therefore, snowmelt likely enhances runoff due to greater input concentration in dry climates, whereas the importance of snow for concentrated input reduces with wetter climates, where greater runoff from snowmelt may relate more to higher antecedent moisture.

The effects of snow loss on D were not as consistent across our simulations as the effects on Q. In general, Q/P was greater than D/P, so Q was more sensitive to changes in input: Q was higher for snowmelt than rainfall events; Q/P decreased in all rain simulations, increased in concentrated input simulations, and increased with both snowmelt fraction and input concentration at the annual time scale. In contrast D/P increased for all rain simulations in wet climates but decreased in dry, increased in concentrated input simulations, and was not strongly correlated to snowmelt fraction. This variability in D/P response as compared to Q/P is likely because S mediates the connection between input and D. In the 1D model Q is affected by infiltration rate and near-surface storage and can more rapidly respond to input changes. In the simulations shown here once subsurface storage is at capacity, D will plateau, and Q will increase with further input due to the saturation excess mechanism.

Soil texture and depth generally do not change partitioning at the annual time scale as much as the varying climate scenarios (Figure 6), although both runoff and deep drainage increase in the shallowest soils. Shorter durations of deep drainage for the coarser sandy loam profile compared to the finer texture soils are offset by higher rates of flux during deep drainage in the coarser profile (Figure 8a). Lower likelihood of surface saturation in the sandy loam soil compared to other soils is offset by greater likelihood of infiltration excess runoff. Altering soil profile depth and the associated root zone to 0.5, 1.5x and 2x the original depth produces the largest effects on Q/P and D/P from 0.5x to 1x depth, and mixed directional response from 1x to 2x depth (Table 3, Figure 6). When soil depths exceed the 1x scenario, the relative amounts of Q and D change (Figure 6). Q gradually declines with greater storage because surface soils do not stay as wet, whereas D gradually increases with greater storage because less water is lost to Q. The responsiveness of fluxes to changes in soil depth from 0.5-1x may relate to storage capacity relative to input. The soil depths ranged from 106-127 cm, which with a porosity of 0.4 gives a storage capacity of 42-51 cm, large enough to store the mean annual precipitation in most watersheds (Figure 3). When this storage is reduced by half to 21-25 cm, it is smaller than the mean annual precipitation at the wetter sites, which would lead to greater likelihood of soil saturation that leads to D and Q. Consequently the change in profile depth from 0.5 m to 1 m represents a shift from annual input greatly exceeding profile storage, to storage approximately accommodating annual input. At

the sites used in this study, mean annual P ranges from 0.8 to 11.3 times the storage of the 1x soil profile, and peak SWE ranges from 0.1 to 5.9 times the storage. Reducing soil depth increases the likelihood that peak SWE will exceed the soil storage capacity, leading to greater surface runoff and deep drainage (Smith et al., 2011). Deeper rooting depths can allow more water to remain in storage and be lost to ET before contributing to surface runoff and deep drainage (Smith et al., 2011).

4.2 Uncertainties

Given the complex nature of soil water movement in heterogeneous mountain topography, this study makes several assumptions and simplifications. The simulations do not include the intricacies of vegetation water use, and the routine chosen assumes a static leaf area index (LAI) with root uptake controlled only by PET and soil moisture. The water balance in hydrologic models can be highly sensitive to the method chosen to represent root uptake and plant water use (Gerten et al., 2004), and hydrologic models generally poorly capture or replicate the interactions between soil, vegetation and atmospheric properties that combine to control plant water use (Gómez-Plaza et al., 2001; Gerten et al., 2004; Zeng et al., 2005). Additionally, simulations are generally wetter than measured water contents; therefore, the representation of partitioning shown here displays relative response between climates and soil profiles rather than absolute quantification of these partitioned components.

Hydrologic response in hillslopes and catchments is not fully captured in the 1-D modelling approach. Sites across elevation and precipitation gradients in this study show different responses within each region with generally lower Q/P and D/P at drier, lower elevations (Figure S6). Within each elevation zone, local variability in microclimate, vegetation, and soil properties can also lead to heterogeneity in input and partitioning. In addition, we did not allow for frozen soils in our simulations, but this can be a strong influence on soil input partitioning in places where snow depth was <50 cm and incapable of insulating the soil (Slater et al., 2017). The 1-D model does not incorporate lateral surface or subsurface flow, which can be redistribute water downslope along the soil snow interface (Webb et al., 2018) and within the shallow subsurface (Kampf et al., 2015). Lateral redistribution of water thus leads to spatially variable patterns of input, storage, runoff generation, and ET at the hillslope to watershed scales (Brooks et al., 2015). While simulating only vertical flow is reasonable for SNOTEL sites located in relatively flat forest openings, 1-D simulations will tend to be biased wet because they do not allow any lateral redistribution. A progression of the work shown here would be to simulate 3-D flow and examine the spatial variability in effects of snow loss. For example, a decline in deep drainage near a ridge line, where flow paths are predominantly vertical could reduce subsurface flow emergence at downslope locations, and this decreased groundwater emergence may reduce ET in areas where vegetation is reliant on the emergence of deeper flow paths. Water partitioned into Q and D in a 1-D model may not represent the same Q and D observed at a stream: Q generated at a point location may reinfiltrate downslope; D may also emerge downslope to supply streamflow rather than remaining in the deep subsurface.

The partitioning of input into the components used here (Q,D,ET,S) is affected by soil depth. The profile depths we simulated represent the minimum likely soil depth, as the collection of the pedon reports was limited by the depth of refusal for sample collection. Shallow soil profiles can also lead to a wet bias in simulations, and this can artificially elevate saturation excess flow leading to our observations of greater Q/P than D/P in most site-years. Saturation excess overland flow has been documented in the elevation bands of many SNOTEL sites (Newman et al., 2004; Eiriksson et al., 2013; Kampf et al., 2015), but it may not occur as frequently as simulated.

Sub-daily dynamics in snow melt and ET are not captured by our use of a daily time step. We chose to model soil water response to rainfall and snowmelt at the daily timestep due to better data quality, but processes affecting partitioning of these inputs take place at sub-daily scales. Comparisons of results from simulations using daily vs hourly input demonstrate similar timing of response, but greater cumulative surface runoff from hourly simulations and greater cumulative deep drainage from daily simulations (Table S2, Figure S2). The short hourly time period allows for higher intensity input, which causes infiltration excess overland flow, whereas daily input is of lower intensity, allowing for greater deep percolation.

The simulations used here only allow for matrix flow, excluding macropore flow, for a simplified representation of soil water movement. Preferential flow though the profile can enhance deep drainage relative to surface runoff, which is another potential reason why soil moisture simulations were biased wet. 60-80% of deep drainage has been shown to occur as preferential rather than interstitial flow (Wood et al., 1997; Jaynes et al., 2001; Sukhija et al., 2003), yet our dry climate simulated annual D/P of ~0.05 is of similar magnitude to that reported prior (Wood et al., 1997). The simulated Q/P (~0.0-0.9) vs snowmelt fraction plots from HYDRUS-1D simulations follow the same general increasing pattern (r = 0.41) as Q/P (~0.0-1.0) vs SP in Hammond et al., 2018a (r = 0.39). This lends confidence to the HYDRUS-1D simulations, as their simulated values are in the same range as observed streamflow.

5 Conclusions

This study helps to explain the mechanisms that lead to greater runoff from snowmelt. At event scale snowmelt generates more runoff because it tends to be greater in input rate and to occur on wetter soils than rainfall; the concentration of input during seasonal snowmelt elevates runoff in dry climates but has less of an influence as in wet climates. Deep drainage can also decline with loss of snow, but it has a weaker response because soil storage buffers the impacts of snow loss. Soil texture modifies short-term timing of soil moisture and runoff generation, but these effects are not large enough to alter the annual response of different soil types to changes in snow. Soil depth can have a greater effect on input partitioning, particularly where soil water storage is less than mean annual precipitation. Soils that are shallower than observed soil depths generate the greatest runoff and deep drainage, indicating that shallow soils may show the largest changes in partitioning as input transitions from snowmelt to rainfall. Increasing soil depth above observed depths gradually reduces surface runoff while increasing deep drainage. The 1-D simulations provide basic hypotheses for hydrologic partitioning under changing snowmelt that

should be further explored in a 2-D or 3-D hydrological models and direct observations. Although more work is necessary to translate these finding to streamflow response, water managers should develop strategies to mitigate impacts of reduced streamflow generation in places that are most at risk for shifts from snow to rain.

Author Contributions

JH, AH and SK designed the experiments and JH and SW carried them out. JH and SW performed the simulations. JH conducted statistical analyses on model outputs. JH prepared the manuscript with contributions from all co-authors.

1. Competing interests

The authors declare that they have no conflict of interest.

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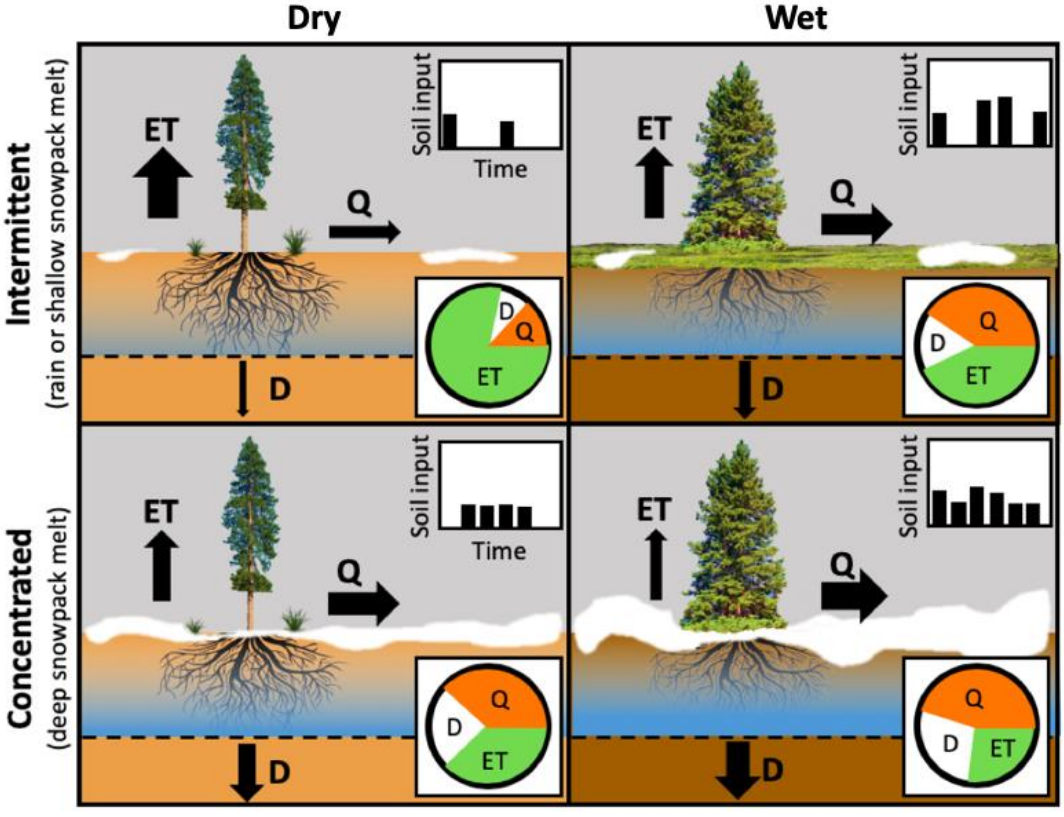

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734 **Figure 1. Conceptual illustration of study hypotheses indicating the importance of concentrated snowmelt**
735 **input (bottom panels) versus intermittent input (top panels) for runoff generation. The wet climate (right-**
736 **hand panels) generates more runoff (Q) and deep drainage (D) and less evapotranspiration (ET) compared to**
737 **the dry climate (left-hand panels). In both climates, concentrated input can increase both Q and D because it**
738 **is more likely to allow soil saturation than intermittent input, which allows ET during periods of drying. The**
739 **concentrated input from snowmelt leads to greater increases in Q and D in the dry climate than in the wet**
740 **climate because snowmelt is the most likely cause of soil saturation in dry climates.**
741

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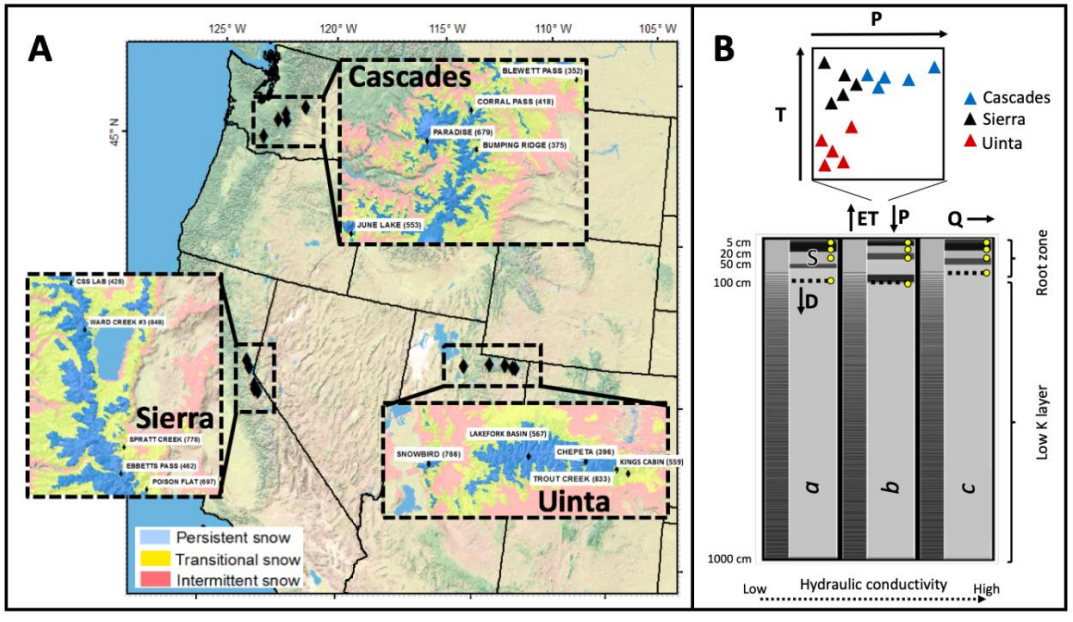

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743 **Figure 2. (A) SNOTEL sites utilized for climate scenarios in this study with insets displaying snow zones**
744 **classified by mean annual snow persistence (Moore et al., 2015). (B) Modeling domain layout with yellow**
745 **points showing 5, 20 and 50 cm depths where volumetric water content time series were used for model**
746 **calibration. Deepest yellow point is the depth where time series were extracted to calculate deep saturation.**
747 **Symbols in the graph above the discretized soil profile represent the range of climate scenarios used plotted**
748 **by mean annual precipitation (P) and mean annual temperature (T), and the three soil profiles below**
749 **represent the soil parameter sets labeled with italicized capital letters (*a*) loam (*b*) sandy clay loam (*c*) sandy**
750 **loam. Different layers in each soil profile are represented as shades of gray.**
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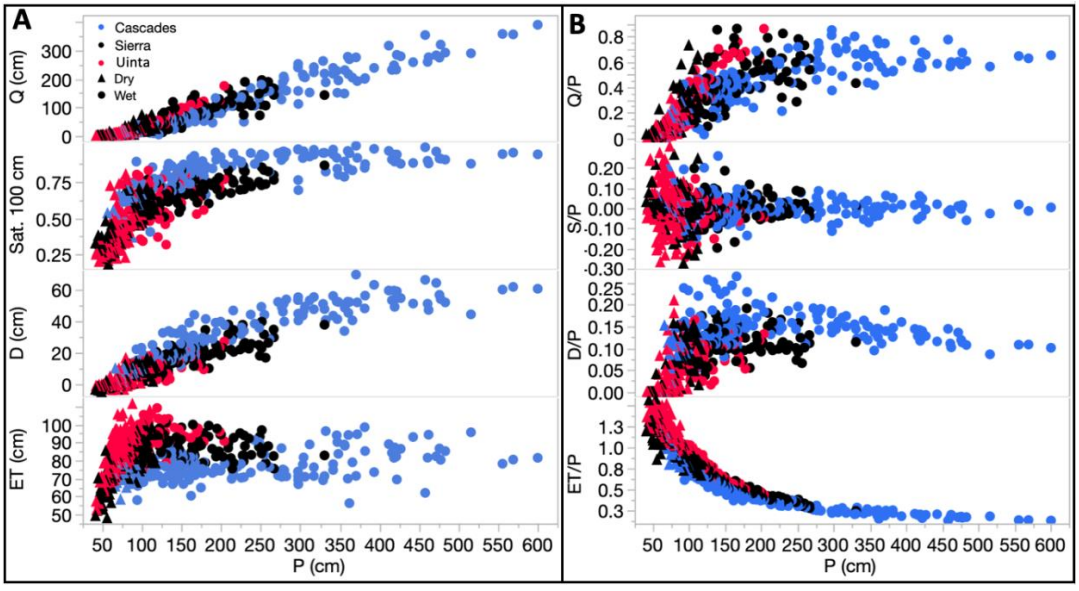

753

754 **Figure 3. A) Annual runoff (Q), mean saturation at 100 cm depth (Sat100), deep drainage (D) and**
755 **evapotranspiration (ET) vs annual precipitation (P) classified by region and climate type. B) Q/P, ΔS/P, D/P**
756 **and ET/P vs P classified by region and climate type. Dry sites P/PET ≤1, Wet P/PET >1. Data from historical**
757 **input scenarios for soil profile 1056, loam.**
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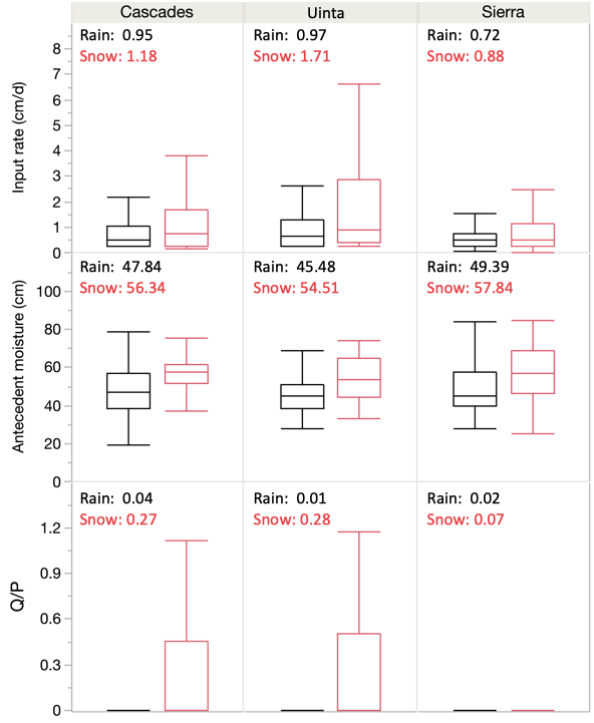

Figure 4. Boxplots of event input rate (cm/d) (top), antecedent soil moisture storage (S, cm) (middle) and event runoff ratio (Q/P, bottom) by region and event type (rain black, snowmelt red). Text in each subplot gives mean values. All ANOVA comparisons between values for rain and snowmelt have p-values <0.0001. Results from historical simulations on loam profile.

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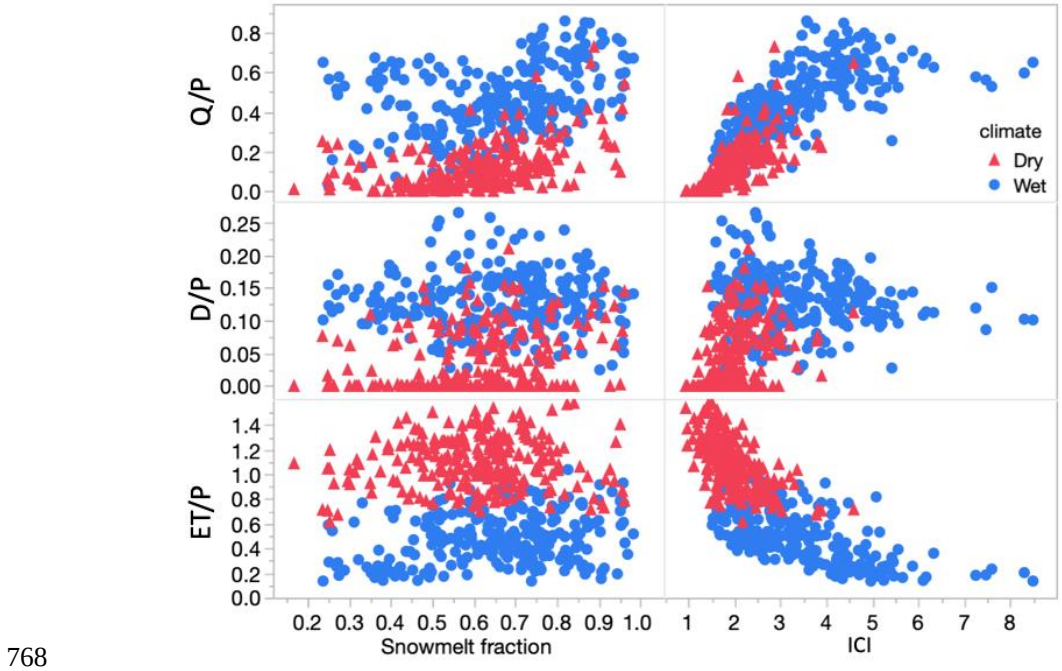

Figure 5. Ratio of runoff (Q), deep drainage (D) and evapotranspiration (ET) to input (P) vs. snowmelt fraction of input and input concentration index (ICI) at the annual time scale. Data from historical simulations on loam profile. Dry sites P/PET <1=, Wet P/PET >1.

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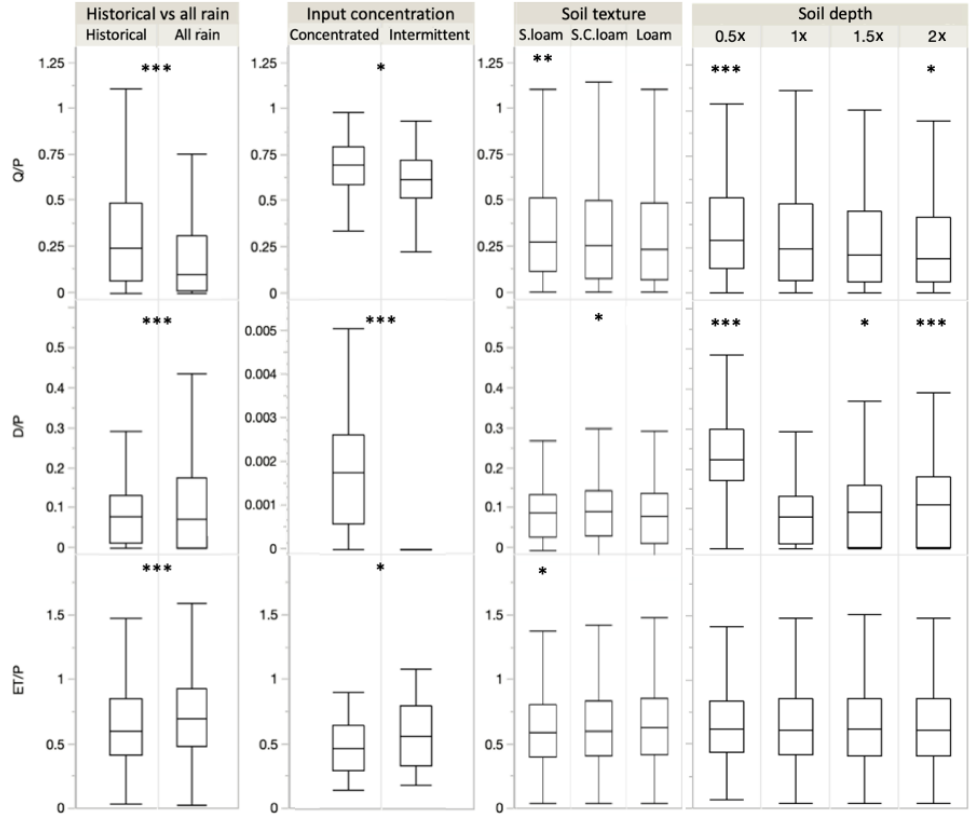

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774 **Figure 6. Boxplots of Q/P, D/P and ET/P for four different experiments: historical vs all rain input on loam**
775 **soil and constant 1x depth, intermittent vs concentrated input on loam soil and constant 1x depth, different**
776 **soil textures with constant 1x depth, and different soil depths all with loam soil texture. Asterisks denote**
777 **significance of ANOVA tests between groupings. P-value of ANOVA, *<0.5, **<0.01, ***<0.001. Boxplots**
778 **correspond with values in Table 3. Soil texture and soil depth scenarios are compared to 1x depth and loam**
779 **texture profile for ANOVAs.**
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781

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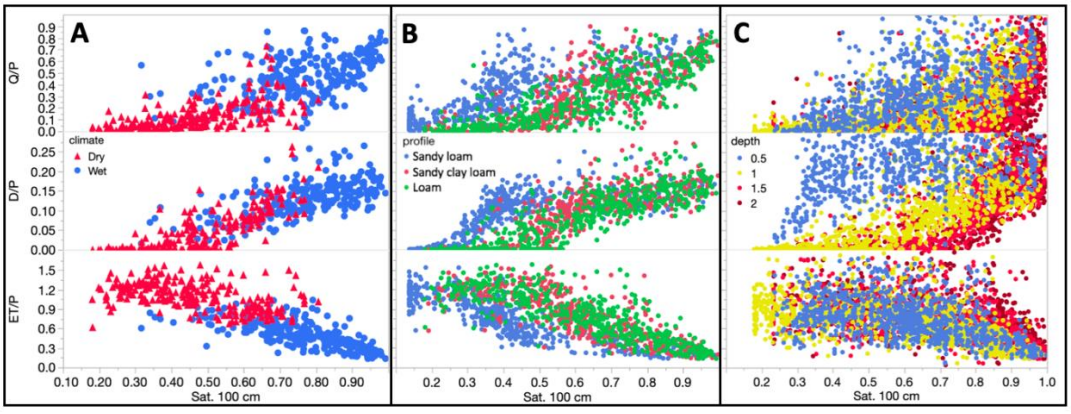

Figure 7. A) Annual surface runoff (Q), deep drainage (D) and evapotranspiration (ET) as a fraction of annual precipitation (P) vs annual mean saturation at 100 cm depth (Sat100) and classified by climate type on the loam profile, Dry sites P/PET <=1, Wet P/PET >1. B) The same variables displayed in A but classified by soil texture on three different soil profiles. C) The same variables in A but classified by root zone depth on four different profiles of differing root zone depth. All simulations use historical input.

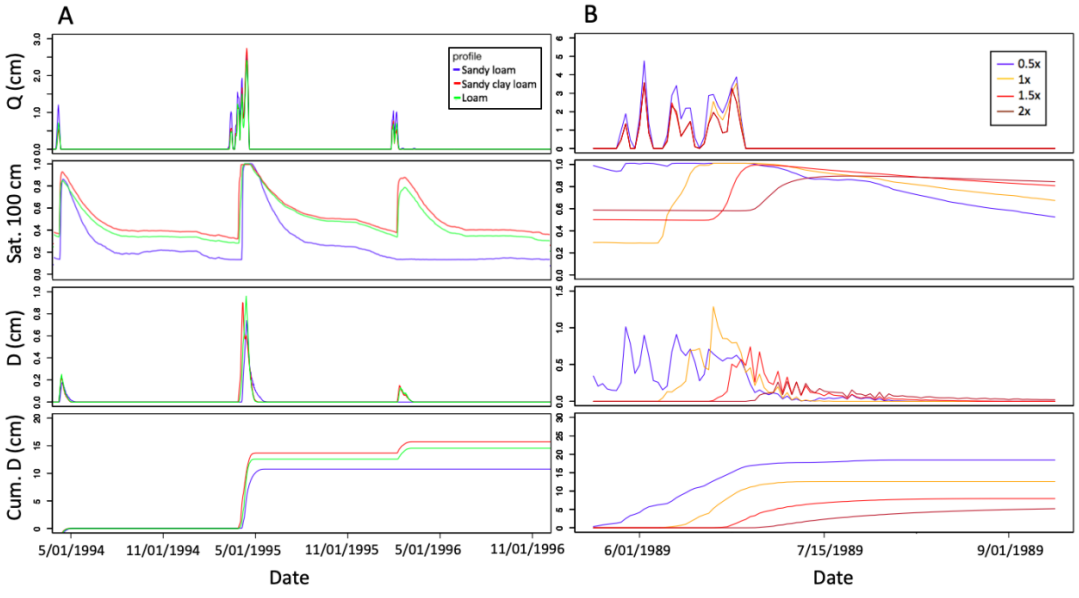

Figure 8. (A) daily time series of runoff (Q), saturation at 100 cm depth (Sat100), deep drainage (D), and cumulative deep drainage for SNOTEL site 698 input on SNOTEL site 515 (sandy loam), 1049 (sandy clay loam) and 1056 (loam) profile. (B) daily series for the same variables plotted for four depth scenarios 0.5x, 1x 1.5x and 2x original rooting zone depth.

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Table 1. SNOTEL station properties including the start and end of data records, site elevation, and mean annual climatic characteristics: precipitation (P), temperature (T), snow persistence (SP, %), and aridity index (P/PET).

SNOTEL ID	Region	State	Start	End	Elevation (m)	P (cm)	T (C)	SP	P/PET
352	Cascades	WA	1981	2015	1292	90	6.3	54	0.8
553	Cascades	WA	1982	2015	1049	433	6.9	65	4.4
375	Cascades	WA	1978	2015	1405	146	4.9	69	1.8
679	Cascades	WA	1980	2015	1564	263	4.8	77	4.9
418	Cascades	WA	1981	2015	1768	158	3.6	83	1.9
778	Sierra	CA	1980	2015	1864	69	8.0	53	0.7
697	Sierra	CA	1980	2015	2358	98	3.8	63	0.6
428	Sierra	CA	1981	2015	2089	180	6.0	72	1.3
848	Sierra	CA	1978	2015	2028	197	5.9	74	1.3
462	Sierra	CA	1978	2015	2672	142	4.0	78	1
559	Uinta	UT	1979	2015	2659	74	1.4	60	0.6
833	Uinta	UT	1979	2015	2901	70	1.5	69	0.7
396	Uinta	UT	1981	2015	3228	81	-0.1	76	0.9
567	Uinta	UT	1980	2015	3342	98	0.0	86	0.9
766	Uinta	UT	1989	2015	2938	157	3.2	87	1.3

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Table 2. Soil profile properties derived from NRCS pedon reports and ROSETTA (Ros.) neural network. Columns are SNOTEL site, soil profile horizon, depth range of horizon, rock percent of sample volume, organic carbon percent of sample volume, sand percent of sample weight, silt percent of sample weight, clay percent of sample weight, Db_{33} bulk density of soil sample desorbed to 33kPa, Θ_{33} volumetric water content at field capacity, Θ_{1500} volumetric water content at wilting point, soil texture, residual volumetric water content Θ_r, saturated volumetric water content Θ_s, pore size distribution parameter α, and K_s saturated hydraulic conductivity. The lowest horizon K_s value was calibrated. Soil textures abbreviated as follows: sandy loam (SL), sand (S), loamy sand (LS), sandy clay loam (SCL), loam (L). SNOTEL 515, Harts Pass, WA, SNOTEL 1049, Forestdale Creek, CA, SNOTEL 1056, Lightning Ridge, UT.

Site	Hor.	Depth (cm)	rock % vol	organic C % vol	sand % wt	silt % wt	clay % wt	Db_{33} g cm^{-3}	Θ_{33}	Θ_{1500}	Text.	Ros. Θ_r	Ros. Θ_s	Ros. α (1/cm)	Ros. K_s (cm/d)
515	A1	0-15	9	9	53.5	35.6	10.9	0.63	0.41	0.14	SL	0.06	0.62	0.009	17.4
515	A2	13-38	8	8	57.6	35.3	7.1	0.64	0.47	0.14	SL	0.05	0.60	0.011	20.5
515	2Bw1	38-61	27	3	73.1	22.1	4.8	0.86	0.3	0.08	SL	0.04	0.55	0.032	15.1
515	2Bw2	61-81	55	1	81	11	8	1.46	0.16	0.09	LS	0.05	0.40	0.036	5.49
515	Cd	81-106	7	1	91.3	4.1	4.6	1.52	0.14	0.05	S	0.05	0.38	0.033	17.4
515	Cd	106-1000	7	1	91.3	4.1	4.6	1.52	0.14	0.05	S	0.05	0.38	0.033	1.74
1049	A	0-9	10	7	52.6	25.2	22.2	0.94	0.40	0.14	SCL	0.08	0.55	0.014	5.17
1049	Bt1	9-20	14	2	48.6	25.4	26	1.13	0.30	0.14	SCL	0.08	0.50	0.014	2.13
1049	Bt2	20-43	14	1	52.9	23.8	23.3	1.24	0.32	0.12	SCL	0.07	0.47	0.016	1.74
1049	Bt3	43-63	21	1	53.4	24	22.6	1.19	0.33	0.13	SCL	0.07	0.48	0.015	2.18
1049	Bt4	63-77	19	1	55.5	25.9	18.6	1.39	0.32	0.12	SL	0.06	0.42	0.017	1.22
1049	Bt5	77-110	11	0	52.4	30.2	17.4	1.21	0.39	0.13	SL	0.06	0.45	0.013	2.22
1049	Bt5	110-1000	11	0	52.4	30.2	17.4	1.21	0.39	0.13	SL	0.06	0.45	0.013	0.22
1056	A	0-10	11	3	36.1	48.8	15.1	1.17	0.30	0.12	L	0.06	0.44	0.010	2.41
1056	A	10-38	7	2	35.3	49.5	15.2	1.27	0.28	0.11	L	0.06	0.41	0.006	1.47
1056	Bt1	38-76	6	2	36	48.6	15.4	1.25	0.30	0.10	L	0.06	0.42	0.006	1.59
1056	Bt2	76-89	16	1	39.3	46	14.7	1.26	0.34	0.09	L	0.06	0.41	0.007	1.54
1056	2B	89-127	6	2	36.3	48.2	15.5	1.18	0.24	0.09	L	0.06	0.44	0.006	2.23
1056	2B	127-1000	6	2	36.3	48.2	15.5	1.18	0.24	0.09	L	0.06	0.44	0.006	0.22

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Table 3. Mean values of unitless response variables Q/P, D/P, and ET/P compared by climate type for four hypothetical scenarios: (1) historical vs all rain input, (2) intermittent vs concentrated input, (3) historical input on sandy loam, sandy clay loam, and loam profiles, (4) historical input on 0.5x, 1x, 1.5x and 2x original rooting zone depth. Dry sites P/PET ≤1, Wet P/PET >1. All scenarios in the table besides those explicitly altering soil texture use the loam profile (1056). Asterisks denote the significance of ANOVA tests between groupings of simulations and arrows show the direction of change relative to the base scenario: historical input on 1x depth profile with loam texture. P-value of ANOVA, *<0.5, **<0.01, *<0.001. Boxplots correspond with values in Table 3.**

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Experiment	Scenario	Climate	Q/P	D/P	ET/P
Historical vs. all rain	Historical	All	0.31	0.09	0.66
		Wet	0.44	0.12	0.51
		Dry	0.13	0.03	0.83
	All rain	All	0.19↓***	0.12↑	0.73↑**
		Wet	0.28↓***	0.14↑*	0.55↑
		Dry	0.04↓***	0.01↓***	0.95↑***
Intermittent vs. concentrated[2]	Intermittent	All	0.59	0.00	0.58
		Wet	0.64	0.00	0.34
		Dry	0.54	0.00	0.80
	Concentrated	All	0.68↑*	0.002↑***	0.48↓*
		Wet	0.68↑	0.002↑***	0.28↓*
		Dry	0.68↑*	0.002↑***	0.66↓**
Soil texture	Loam (L)	0.31	0.09	0.66	0.31
		0.44	0.12	0.51	0.44
		0.13	0.03	0.83	0.13
	Sandy loam (SL)	All	0.35**↑	0.09	0.63↓*
		Wet	0.05↓	0.13↑	0.51↓
		Dry	0.19↑*	0.05↑	1.01↓*[1]
	Sandy clay loam (SCL)	All	0.32↑	0.10↑*	0.65↓
		Wet	0.48↑	0.14↑	0.52↑
		Dry	0.14↑	0.06↑	1.08↓[1]
Soil depth	0.5x	All	0.35↑***	0.25↑***	0.67↑
		Wet	0.54↑***	0.28↑***	0.53↑*
		Dry	0.17↑**	0.22↑***	0.80↓*
	1x	All	0.31	0.09	0.66
		Wet	0.44	0.12	0.51
		Dry	0.13	0.03	0.83
	1.5x	All	0.29↓	0.10↑*	0.67↑
		Wet	0.46↑	0.16↑*	0.51
		Dry	0.09↓	0.03	0.84↑
	2x	All	0.27↓*	0.11↑***	0.66
		Wet	0.44	0.18↑***	0.51
		Dry	0.09↓	0.04↑	0.84↑

[1]Values of ET/P >1 indicate root uptake from soil storage for years with low input (Figure S5).
[2]For a dry site (375) and a wet site (559). Intermittent simulations have annual total input separated into four five-day periods, whereas concentrated input simulations have all input in twenty-day period of high magnitude.