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Small soil storage capacity limits benefit of winter snowpack to upland vegetation

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Abstract:

In the western United States, the mountain snowpack is an important natural reservoir that extends spring and summer water delivery to downstream users and ecosystems. The importance of winter snow accumulation to upland ecosystems is not as clearly defined. This study investigates the relative contribution of winter precipitation to upland spring and summer soil moisture storage and availability in a semi-arid mountainous watershed. At this site, coarse soil textures and shallow soil depths limit soil storage capacity to 6–16 cm. Winter precipitation exceeds soil storage capacity by 2.5 times. Accordingly, soil moisture profiles at most locations in the watershed reach field capacity in early winter. With soil storage near capacity, water released by snowmelt primarily contributes to deep drainage and makes a limited contribution to the soil moisture reservoir. Water that is retained by the soil after the snowpack melts is lost to evapotranspiration in as little as 10 days. In contrast, spring precipitation extends moist soil conditions by up to 90 days into the warm season, when ecological water demand is highest. These field observations suggest that changes in spring precipitation, not winter snowpack, may have the greater impact on upland ecosystems in this environment. Furthermore, because winter precipitation is in excess compared to the soil storage capacity, soil moisture availability may be fairly insensitive to climate change-induced transitions from snow to rain. Copyright © 2011 John Wiley & Sons, Ltd.

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KEY WORDS soils; storage; snow; semi-arid; western United States; climate change

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INTRODUCTION

Two potential water storage reservoirs, the snowpack and the soil profile, regulate hydrologic and ecological functions in semi-arid upland ecosystems, which are defined here as the non-riparian plant communities. In this paper, we examine the capacity of the soil profile to store winter precipitation, primarily from melted snowpack, into the summer growing season. Ongoing and impending climate change is reducing winter snow accumulation across the western United States (Mote *et al.*, 2005; Knowles *et al.*, 2008; Clow, 2010) with potentially profound implications for closely linked hydrologic and ecological systems. There is strong evidence that diminishing snowpacks are reducing summer stream flows (Barnett *et al.*, 2008; Luce and Holden, 2009; Nayak *et al.*, 2010) and producing earlier peak stream flow (Nayak *et al.*, 2010). These changes are altering both aquatic ecosystems (Rieman *et al.*, 2007) and downstream water availability (Barnett *et al.*, 2005).

The impact of changing snowpack on upland ecosystems, the landscape that hosts the majority of the snowpack, is less certain. A number of studies have documented a strong statistical linkage between higher fire activity/severity and both declining and earlier snowmelt (Westerling *et al.*, 2006; Lutz *et al.*, 2009; Morgan *et al.*, 2008; Heyerdahl *et al.*, 2008). These observations suggest that changes in snowpack alter summer plant water availability and subsequent fire susceptibility. In a multi-year study, Concilio *et al.* (2009) linked the size of the winter snowpack and the subsequent magnitude of summer soil respiration, a surrogate for ecological activity. This study also observed limited dependency on summer precipitation. In contrast, Hamlet *et al.* (2007) and Litaor *et al.* (2008) demonstrated that summer precipitation can offset changes in winter snowpack accumulation.

At least some of the observed differences in the importance of winter snowpack may reflect the fact that ecosystems do not directly use the snowpack as a water source, but rather extract stored water from the soil profile. In these environments, there is an important spatiotemporal disconnect between the arrival of precipitation as snow and its availability to plants for transpiration. In semi-arid upland ecosystems, water availability is often controlled by soil moisture (Rodriguez-Iturbe *et al.*, 2001). Groundwater levels are generally sufficiently deep to limit access of many plant types, and even vegetation that is able to reach the groundwater can be inhibited by dry soils during seed

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germination (Larson and Schubert, 1969; Bai *et al*., 1995). Accordingly, the soil reservoir is the pathway by which most snowpack water reaches upland ecosystems.

While snow accumulates during winter when ecological growth is suppressed by low temperatures (Saxe *et al*., 2001), ecological water stress is typically most pronounced during the warm, and often dry, summers that characterize much of the western United States (Henderson-Sellers and Robinson, 1986; Viola *et al*., 2008). Therefore, the upland ecological value of the snowpack is determined, in large part, by the ability of soil to retain that water into spring and summer (Geroy *et al*., 2011). Water stored as snow can, in some instances, remain as snow well into the growing season (Litaor *et al*., 2008), or, alternatively, snowmelt can transfer the water to storage in the soil profile (McNamara *et al*., 2005). While differences in the magnitude of the snowpack and melt timing have been shown to impact spring and summer soil moisture, evapotranspiration, and vegetation (Hamlet *et al*., 2007; Williams *et al*., 2009), such influences can be modulated by the seasonal distribution of precipitation (Hamlet *et al*., 2007; Litaor *et al*., 2008); spring or summer precipitation can provide moisture after snowmelt is lost from the system.

While it is plausible that climate change-induced declines in snowpack will negatively influence soil moisture storage, empirical evidence supporting that assumption is limited. With the notable exception of Litaor *et al*. (2008), there is a paucity of data directly linking snowmelt and soil water storage dynamics. The objective of this study was to determine the relative importance of snowpack and snowmelt to upland soil moisture and storage in a semi-arid ecosystem. We have used a mass balance approach to demonstrate that the potential soil water storage capacity is small compared to both the annual precipitation and winter snow accumulation; observed trends in seasonal soil moisture highlight the limited impact of snow melt on spring and summer soil moisture levels. Alternatively, our results illustrate the important role spring precipitation plays in influencing soil moisture storage into the spring and summer months when vegetation is most active.

Field site description

The Dry Creek Experimental Watershed (DCEW) is a 27 km^2 semi-arid basin extending from 1100 to 2200 m elevation near Boise, Idaho, USA (Figure 1) (McNamara *et al*., 2005). The watershed is instrumented with four meteorological stations, seven stream gages, and multiple soil moisture monitoring stations (http://earth.boisestate.edu/drycreek/) (Aishlin and McNamara, 2011). A meteorological station (Bogus Basin) is maintained by the United States Department of Agriculture (USDA), Agricultural Research Service at an elevation of 1932 m, just north of the DCEW upper boundary.

The bedrock of DCEW consists of granodiorite of the Idaho Batholith. Soils on hillslopes are typically shallow (< 2 m deep) gravelly loams to gravelly sands (USDA,

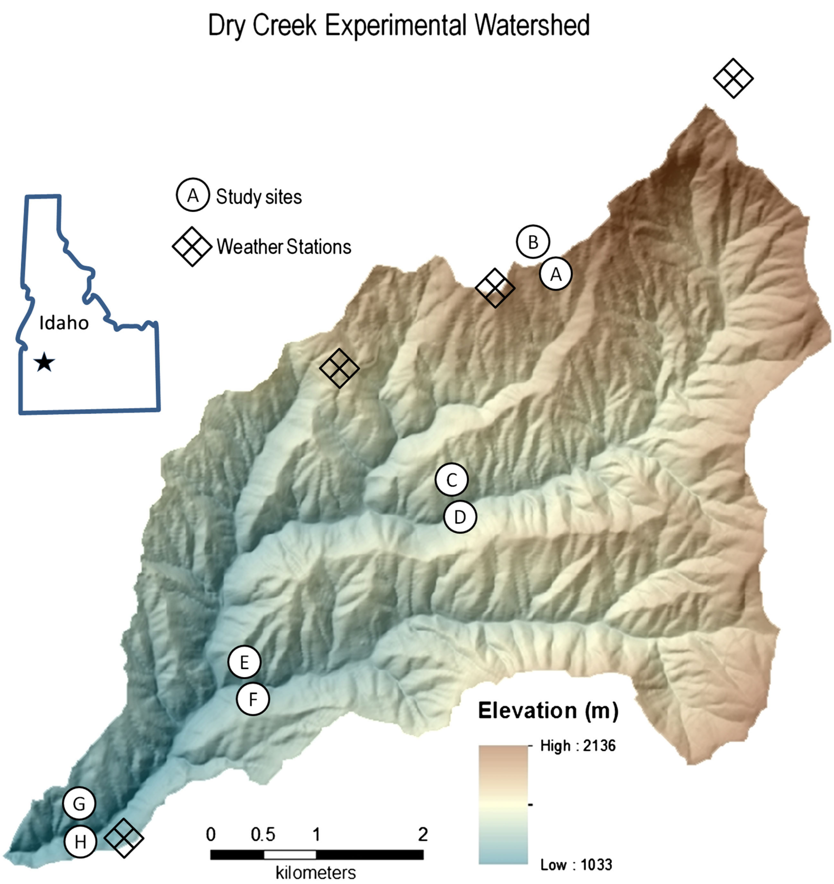

Figure 1. Dry Creek Experimental Watershed site map

1997; Gribb *et al.*, 2009; Tesfa *et al.*, 2009). At its lower elevations, the DCEW is characterized by grass and sagebrush shrublands, while higher elevations support forest vegetation including fir and pine (McNamara *et al.*, 2005; Williams, 2005). Like most semi-arid environments in the American west, the watershed has been historically affected by fire and invasive species.

The climate of the DCEW is characterized by cold, wet winters and hot, dry summers, as well as an orographic effect of increasing precipitation and decreasing temperature with increasing elevation. The lower DCEW is classified as a *steppe summer dry climate*, and the upper DCEW as a *moist continental climate with dry summers*, using the Köppen climate classification system (Henderson-Sellers and Robinson, 1986). Annual air temperatures range from −15 °C to 33 °C at lower elevations, and from −14 °C to 26 °C at higher elevations. Precipitation and temperature are out-of-phase, with most precipitation falling during the cold winter months (Figure 2). At the top of the watershed, approximately 77 % of annual precipitation falls as snow, while at the bottom of the watershed, precipitation is dominantly rain, with less than 33% falling as snow.

METHODS

Eight study sites were located at four elevations, with sites on north-facing and south-facing aspects at each elevation (Figure 1, Table I). Estimates of mean annual precipitation and mean annual air temperature were calculated for the elevation of each study site in this

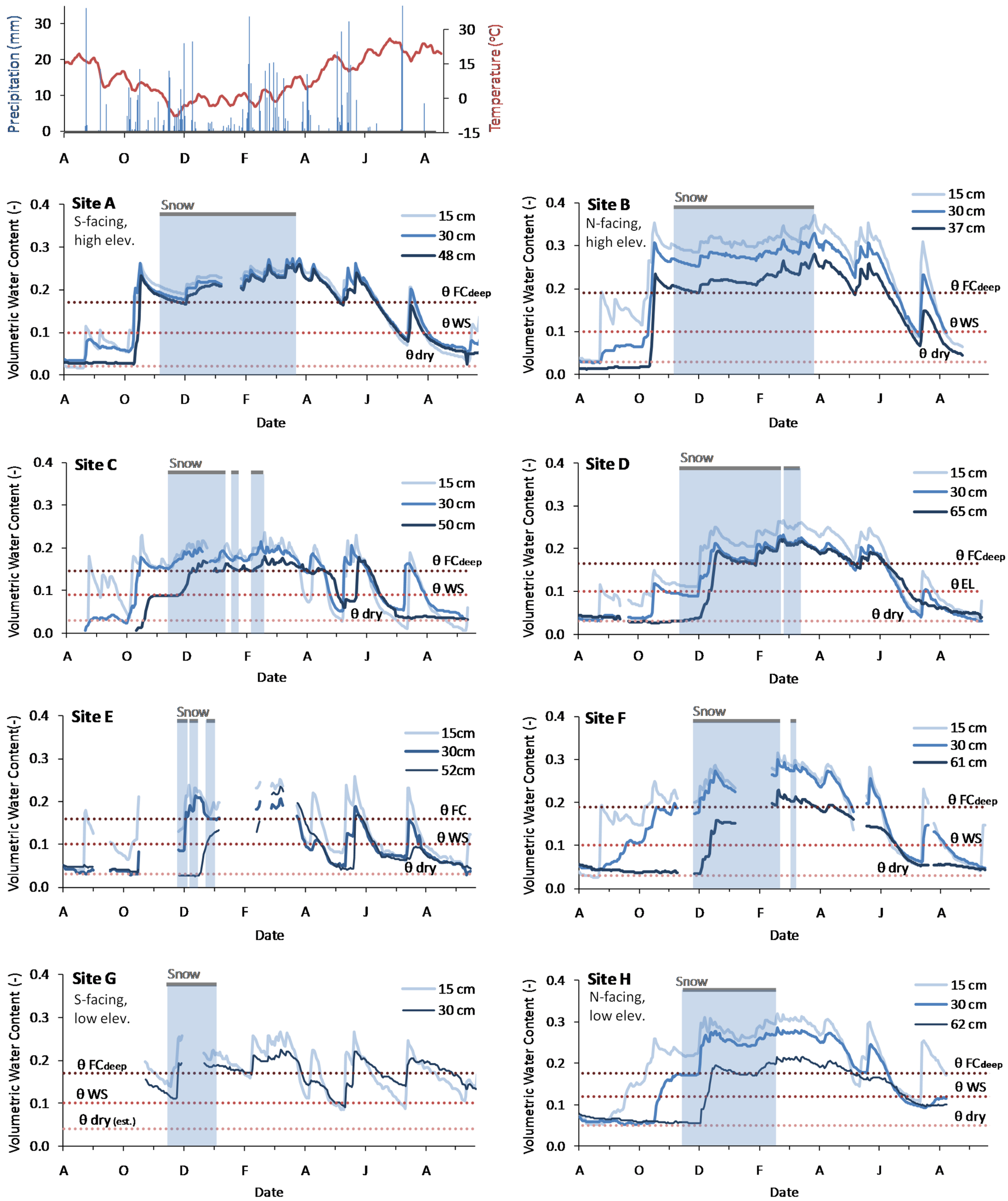

Figure 2. Temporal soil moisture data for the eight study sites August 2008 to September 2009. Precipitation and air temperature data are from the weather station at the 1610 m elevation for the same time period. The terms *dry*, FC_{deep}, and *WS* are defined in Methods. Depths of moisture data is below ground surface

project using lapse rates based on three weather stations in the watershed. At each site, soils were characterized and soil moisture sensors were installed in replicate at multiple depths. Soil moisture data were collected for one year. Measured values of soil temperature at 2 cm soil depth were used to infer dates of snow cover formation and disappearance at each site. Periods of snow cover were interpreted as the period during which diurnal fluctuations of shallow soil temperature were strongly dampened at approximately 0 °C. The resulting inferences of snow cover periods agreed well with snow depth data recorded using Judd Ultrasonic Depth Sensors (Judd Communications, Salt Lake City, UT) from weather stations located at elevations in the watershed similar to those of the soil moisture monitoring sites. At these weather stations, both snow and rain rates are determined gravimetrically with precipitation occurring while temperature <0 °C was assumed to be snow. This assumption is periodically confirmed with changes in snow depth measured by acoustic sensors and in-field manual snow water equivalent measurements.

Soil characterization

The depth to the soil–bedrock interface was measured in four pits at each site (Figure 1). The soil–bedrock interface was identified as the depth at which the soil became dominated by gravelly, decomposed granite. Soil textural analysis was performed on soil samples of 300–500 g collected at up to four depths in each pit. Soil texture was determined using two particle size analysis methods: laser diffraction and mechanical (sieve/hydrometer). The two methods of particle size analysis differ in how they measure particle size distributions, but results from the methods are linearly correlated to one another (Konert and Vandenberghe, 1997; Beuselinck *et al.*, 1998; Arriaga *et al.*, 2006; Malvern Instruments Ltd., 2009). Laser diffractometry determines particle size fractions on a volume basis, while hydrometer and sieve analyses determine particle size fractions on a mass basis (Syvitski, 1991; Beuselinck *et al.*, 1998). The laser diffraction method allows rapid analysis of many samples, but is not directly comparable to most soil texture values in the literature (Konert and Vandenberghe, 1997; Arriaga *et al.*, 2006). In our approach, the sub-2 mm fraction from all soil samples was first analyzed by laser diffraction. A subset of samples was also analyzed mechanically, and a site-specific linear relationship was established between the two data sets. This relationship was then used to convert all the data into hydrometer equivalent values. The results were classified using the USDA soil classification system (USDA, 1999).

Soil volumetric water content, soil temperature

Soil moisture (measured as volumetric water content) and soil temperature were monitored at each site (Figure 1, Table I). Each site consisted of four replicate soil profiles of three to four sensors each (ECH$_2$O EC-TM, Decagon Devices, Inc., Pullman, WA) producing 121 soil moisture and soil temperature time series. Each site was selected at the mid-slope position (approximately halfway between the ridge crest and the drainage bottom) in a location representative of vegetation characteristics of the hillslope; forested sites were selected in mature-stand timber that had not been recently logged. Each of the four replicate profiles (2 to 6 m apart) at each site consisted of sensors at 2, 15, and 30 cm depth below the mineral soil surface; in areas where the soil was deeper than 30 cm, a fourth sensor was installed above the soil–bedrock interface. Data were collected at 10-min intervals. Presented data represent the average value for each depth in the four replicate profiles at each site; data have also been averaged to a daily interval. The datasets from the 15 cm, 30 cm, and deepest ($>$30 cm) sensors are plotted in Figure 2. The standard deviations of the replicate datasets demonstrate that the sites exhibit observable differences and can be considered representative (Figure S1).

Soil water retention characteristics

To better characterize soil water behavior over time, we have empirically identified three soil water availability conditions which we refer to as: *field capacity*, *soil dry*, and *water stress point* conditions. These points are defined empirically by examination of the temporal soil moisture data over the water year. *Field capacity* is an estimate of the state when the soil moisture content is no longer freely

Table I. Site and soil characteristics

Site	elevation	Aspect	Annual precipitation	Soil depth	Grain size distribution[1] Sand	Silt	Clay	Field capacity	Water stress point	Soil dry condition	Soil storage capacity
	m.a.s.l.		cm	cm	wt %	wt %	wt %	θ_{FC}	θ_{WS}	θ_{dry}	cm
A	1835	S-facing	69	73	74	17	9	0.18	0.10	0.02	11.7
B	1812	N-facing	68	66	71	20	9	0.25	0.10	0.03	14.5
C	1457	S-facing	53	76	82	11	8	0.17	0.09	0.03	10.6
D	1472	N-facing	53	92	66	25	9	0.18	0.10	0.03	13.8
E	1298	S-facing	46	38	78	14	8	0.16	0.10	0.03	4.9
F	1288	N-facing	45	87	69	22	9	0.19	0.10	0.03	13.9
G	1139	S-facing	39	34	80	12	8	0.17	0.10	0.05	4.1
H	1120	N-facing	38	66	52	37	11	0.23	0.12	0.04	12.5
Average			**51**	**67**	**72**	**20**	**9**	**0.19**	**0.10**	**0.03**	**10.8**

[1] wt.% values are calculated on the sub 2 mm size fraction.

draining and is used in soil storage capacity calculations as the condition of a full soil reservoir. This approach follows that employed by McNamara *et al*. (2005). *Soil dry* is an estimate of the point at which water is no longer being extracted from the soil, similar to the plant extraction limit described by Seyfried *et al*. (2009). In this application, we consider this an estimate of the point below which evapotranspiration no longer extracts appreciable moisture from the soil. The *soil dry* value is used in storage calculations as the indicator of an empty soil reservoir with respect to plant availability. In addition to these more widely used characteristic points on the soil moisture dry-down curve, we also identify a point that we refer to as the *water stress point*. This value is an estimate of the water content during soil dry down when the rate of water extraction from the soil starts to become limited by the soil rather than climate (Eagleson, 1978). The *water stress point* is identified theoretically by Rodriguez-Iturbe *et al*. (2001) as the point at which plant water stress is assumed to commence (Nilsen and Orcutt, 1996). All of these empirically defined characteristics of the dry-down curve are influenced by the specific grain-size distribution of the soil and, therefore, can vary with depth and location.

Field capacity moisture content (θ_{FC}) values were identified for each site as an estimate of the moisture retention capacity following soil drainage. In this watershed, there is effectively no capillary contribution of water from a shallow water table, so vertical and lateral redistributions are the most important sources for soil moisture at the soil–bedrock interface (McNamara *et al*., 2005). In this previous study, soil moisture time series exhibit near-maximum moisture content at all depths in early winter with the onset of fall precipitation. Following the early winter peak in moisture content, soil moisture values at all depths exhibit an exponential decline. Because temperatures and light availability are low during this time, evapotranspiration is assumed to be negligible, and the observed decline in moisture content is wholly attributed to soil drainage. Field capacity is commonly identified as the moisture content at which the rate of soil moisture decline during drainage approaches horizontality, or drainage ceases (Brady and Weil, 2002). This near-horizontal behavior can be observed in most of the moisture time series following the early winter drainage period. Even small differences in grain-size distribution in the soil profile produce different field capacity values with depth at many of the sites. Accordingly, a depth-averaged field capacity value was used for soil storage calculations (Table I). For the sake of graphical simplicity, only the estimated field capacity value for the deepest location for each site is plotted in Figure 2. The occurrence of field capacity at the bottom of the profile during the period of minimal evapotranspiration suggests that the entire soil profile is at capacity.

The *soil dry* condition (θ_{dry}) was identified in the temporal soil moisture data as the point following summer dry down where the curve becomes nearly horizontal, indicating that water is no longer being extracted from the soil. In practice, all depths at all sites did not reach *soil dry* conditions. Accordingly, a single *soil dry* value was identified for each site based on those depths where soil moisture trends reached horizontality.

The *water stress point* (θ_{WS}) is determined by identifying a distinct inflection point, or change in slope, in the field-observed dry-down curve. The early dry-down curve is characterized by a constant, negative slope; this section of the dry-down curve is assumed to reflect the period during which water availability is not limiting the rate of water extraction. During this period, soil water pressures are becoming more negative. The absence of a change in the rate of decline, despite this increasing resistance, indicates that soil moisture retention is not inhibiting extraction. The water stress point is identified as the location on the dry-down curve where the rate of extraction begins to decline. This location is characterized by a decline in the slope of the dry-down curve. Because this point is used to denote the onset of limited water availability, it is dictated by the first curve (of the multiple depths) that exhibits this distinct change in slope.

Soil storage capacity is defined as the amount of extractable water available in the soil profile following drainage. This value was calculated by first calculating the available water content (field capacity moisture content (θ_{FC}) − soil dry moisture content (θ_{dry})) and multiplying that value by the soil depth. The resulting value is reported in cm of water and represents the longer term potential storage capacity of the soil at each location.

RESULTS

Soil characteristics

Soil moisture storage and drainage characteristics are influenced by both soil texture, which governs soil retention capacity, and soil depth. Soil pit profiles indicate that soils are shallow at all sites in the watershed, averaging approximately 60 cm deep and ranging from 30 to 110 cm (Table I). The deeper soils are found on north-facing slopes with the deepest at mid-elevations, generally consistent with the observations of Tesfa *et al*. (2009). Textural analysis indicates that the soils across the watershed are coarse grained (Table I), and the sub-2 mm fractions contain 52–82% sand, 11–37%, silt and 8–11% clay. Gravel (>2 mm) content exceeded 15% in many samples, and the soils classify as loam to gravelly loamy sands (USDA, 1999). The primary grain size difference between sites is that the north-facing sites have approximately twice the silt content. This increased silt content comes at the expense of the sand fraction. The clay fraction is always low, but is slightly higher on north-facing slopes. There are limited systematic trends in grain-size distribution with depth, although the soil at the bedrock interface often contains slightly less fine-grained material.

Soil moisture retention

Empirically derived field capacity values (θ_{FC}) range from a moisture content of 0.16 to 0.25 with an average value of 0.19. There was no distinct trend in these values with elevation; however, the highest values were observed on the north-facing slopes. This trend was also observed by Geroy *et al.* (2011) and may be attributable to the higher soil carbon contents on these aspects (Kunkel *et al.*, 2011). *Soil dry* conditions for all the soils were similar, ranging from 0.02 to 0.05 with an average value of 0.03; no distinct trends with elevation or aspect were evident. The initiation of *water stress* conditions was similar throughout the watershed with a range from 0.09 to 0.12 and an average water content of 0.10; no trends with elevation or aspect were evident.

Soil storage capacity was calculated for each site using soil depth and soil field capacity (θ_{FC}) and *soil dry* (θ_{FC}) moisture content values. Soil moisture storage capacity ranged from 4 to 14 cm with an average storage capacity of 11 cm for all sites (Table I, Figure 3). Due to the relatively narrow range in field capacity and *soil dry* values between sites, the soil storage capacity differences closely follow soil depth differences, with the highest soil storage capacity on mid-elevation, north-facing slopes. Total winter (Dec–Mar) precipitation, which includes snow and rain, averages 27 cm and is approximately 2.5 times the soil storage capacity. Spring (Apr–Jun) precipitation averages 12 cm, and is roughly equivalent to the soil storage capacity at most elevations.

Snow cover

Snow accumulation began in late November or early December at all sites but differed dramatically with elevation (Figure 2). At the highest elevation sites, the ground was snow covered for approximately 130 days, while at the lower elevations, snow cover was more intermittent and the sites were snow covered for as little as 20 days. Subsequently, lower elevation sites became snow free as early as February, while the upper elevation sites were snow covered until early April.

Temporal soil moisture trends

Temporal soil moisture values for both the 15 cm, 30 cm, and bedrock interface depths exhibit generally similar trends (Figure 2). Soil moisture contents rise in November, driven by fall precipitation coupled with cooling temperatures. The soils remain wet through the winter months before declining in the spring as precipitation deceases and temperatures rise. The period of winter-moist conditions is more sustained at higher elevations (compare Site A with Site G) and on northern aspects (compare Site E with Site F). While north-facing high elevation sites (B and D) appear to remain at or above field capacity for the entire winter, south-facing, low elevation sites (E and G) exhibit varying degrees of saturation throughout the winter months. While the deeper soil moisture generally tracks with the 15 cm observations at all sites, north-facing locations generally exhibit more divergence over the profile, likely reflective of the greater soil depths on this slope aspect. At all sites, the 15 cm depth generally maintains a higher water content than the greater depth(s). This trend may, in part, reflect small differences in soil texture; soils coarsen somewhat at the bedrock interface; it may also reflect periods of active recharge occurring at soil surface. Soil moisture at the shallowest 2-cm depths (data not shown) tends to be the wettest during precipitation events, but this portion of the soil profile loses water quickly when precipitation ceases.

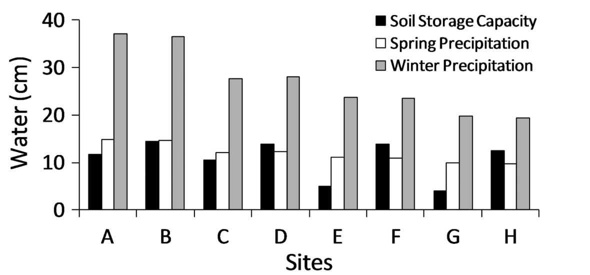

Figure 3. Comparison of the soil storage capacity for each of the eight study sites with estimated average ($n = 10$) annual winter (Dec–Mar) and spring (Apr–Jun) precipitation

At all sites, the timing of snowmelt does not correspond to dramatic change in soil moisture content; at nearly all sites, the soil profile is already at or near saturation when snowmelt occurs. At nearly all sites, spring dry down is initiated shortly after snowmelt and is characterized by a stepwise descent over a 30- to 120-day period during which soil moisture briefly rises in response to spring rain events. With some rain events, the entire soil reservoir refills, as evidenced by the response of the deeper soil moisture sensor (e.g. Site A, May and July rain events). With other rain events, the soil moisture increases at 15 cm and limited, or no, response is seen at greater depths. In some cases, these differential responses may be due to differences in soil thickness; in others, they may reflect variation in precipitation between sites. Two distinct spring dry-down periods were observed, one in May and a second in July; they were separated by a major rain event. During the May dry down, the lower elevation, south-facing sites dried to or nearly to the *water stress* point (θ_{WS}), while the higher elevation and north-facing sites did not dry to that extent. During the July dry down, nearly all the sites at all depths reached the water stress point (θ_{WS}).

The duration of wet conditions following the loss of snow cover varied with elevation and aspect. However, soil moisture remained above the *water stress* point (θ_{WS}) for approximately 60–120 days after snowmelt. During periods of no precipitation, the rate of moisture declines from field capacity to the *water stress* point averaged approximately 20 days and occurred in as few as 10 days. This rapid loss of moisture calls attention to two characteristics of the study setting: a relatively high potential evapotranspiration rate with the onset of summer, and the small storage capacity of the relatively shallow and coarse-grained soils. There is a dramatic difference between the observed extended moist period following snowmelt (60–120 days) and the rapid dry down in the absence of precipitation (20 days). It is noteworthy that even though annual spring precipitation is significantly less than winter precipitation, the timing of the precipitation appears to have a disproportionately large impact on the duration of wet conditions in spring and summer. Because the winter precipitation arrives when temperatures are low, its contribution to spring and summer soil water availability is limited to helping establish the initial moisture conditions at the beginning of the growing season but prior to the onset of the spring dry down. In other words, fall and winter precipitation may determine the initial conditions, but make a limited contribution to spring and summer soil water availability in this system.

DISCUSSION

The results indicate that the impact of winter snow accumulation on spring and summer soil moisture in the DCEW is constrained by limited storage capacity. The coarse-grained, shallow soils, which are already near-storage capacity when snowmelt occurs, can store only a fraction of the snowmelt water; water that is stored is rapidly lost to evapotranspiration when spring arrives. These results are generally consistent with previous observations that the period during which soil water is mobile in these ecosystems is limited to a few weeks during snowmelt (McNamara *et al*., 2005). Because the snowpack melts and drains prior to elevated summer temperatures, the snowpack water has been largely lost from the upland soil reservoir before the peak in the potential growing season. Spring rains appear to extend wet soil conditions by up to 90 days beyond that provided by snowmelt alone. An important corollary to these observations is that spring soil moisture status may be fairly insensitive to the phase change of winter precipitation from snow to rain, an anticipated outcome of impending climate change. In fact, this dataset indicates that the fall soil 'wet-up' is driven by rain, not snow. This prediction will, of course, be made more complicated by the fact that such a phase transition may be associated with warmer conditions, which may extend the period of active evapotranspiration into the winter months.

The data suggest that this system may be more sensitive to changes in precipitation timing than amount. Under the constraints of limited soil storage capacity, maintaining an extended period of moist soil conditions during the period of active evapotransporation requires delivery of additional precipitation. Therefore, the system may be particularly vulnerable to changes in the timing of spring precipitation.

Extension of these results to other semi-arid ecosystems is bounded by a number of constraining variables. First, this semi-arid system is characterized by a summer water limitation; summers are warm, precipitation is limited, and the soil reservoir is typically dry by mid-summer. In systems where water is not, on the average, limiting in the summer, the loss of stored winter precipitation from soils may not be as rapid. At higher elevations, lower temperatures can also constrain ecological activity, and the loss of soil moisture by transpiration may be more gradual. Alternatively, in areas where the depth of the winter snowpack is greater, it may take longer for snow to melt, extending the moisture benefits later into the spring and summer (Flint *et al*., 2008). Further, in more heavily forested areas with substantial winter snowpack, vegetation canopies may also shelter the snowpack from incoming solar radiation driving melt. Finally, in systems with deeper soils, or soils containing a larger fine-grained and/or organic carbon fraction, the soil reservoir would be larger or have greater capacity for moisture retention, and this may result in greater storage of snowmelt, thus extending water availability longer into the summer.

Soil storage dynamics in semi-arid ecosystems strongly influence ecological function. This study highlights the value of simple mass balance calculations for constraining soil storage capacity relative to the temporal precipitation regime. Specifically, the ratio of precipitation amount to soil storage capacity may be a useful metric for defining ecohydrologic vulnerability to climate change. This metric can be indicative of the degree to which local precipitation climatology, on average, can satisfy the soil water deficit: The higher the ratio, the lower the sensitivity to changes in precipitation amount. At this study site, the winter precipitation-soil storage capacity ratio is approximately 3, suggesting soil moisture storage will exhibit limited sensitivity to changes in winter precipitation. In contrast, the spring precipitation-soil storage capacity ratio is near 1 for this site, suggesting even relatively minor changes in spring precipitation may strongly influence soil moisture storage.

CONCLUSIONS

Winter snowpack has little impact on spring and summer soil moisture in the DCEW because of limited soil water storage capacity. Furthermore, growing season soil moisture status may be fairly insensitive to climate change-induced transitions from snow to rain. These results highlight the need to consider the impact of climate change on upland ecosystems through a different lens from that applied to riparian or downstream water uses. While there has been considerable effort expended to document and develop predictions of climate change-induced snowpack changes, there has been limited attention given to explicitly quantifying or predicting changes in spring precipitation; global climate models currently exhibit significant uncertainty with respect to changes in spring precipitation in North America (Dai, 2006). The limited ability of shallow, coarse-grained soils to store water from snowmelt highlights the potential importance of spring and early summer precipitation; changes in spring precipitation may have a profound impact on upland water availability in these environments. Finally, this work does not suggest that these upland ecosystems will be less sensitive to climate change, but rather that changes to winter snowpack may not be the primary vector of impact.

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