Forest ecohydrological response to bimodal precipitation during contrasting winter to summer transitions

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ABSTRACT

The ecohydrological response of forested landscapes in the Southwestern United States depends on the relative magnitude of precipitation during the winter and summer seasons. Prior climate studies have identified an inverse relation in the bimodal regime such that wet winter periods are followed by dry summers and vice versa. Despite this key prediction, the impact of contrasting winter to summer transitions on hillslope hydrology has not been investigated. In this study, we use the Triangulated Irregular Network-based Real-time Integrated Basin Simulator distributed hydrologic model with modifications in the snow module to generate a consistent set of high-resolution hydrologic estimates in a ponderosa pine hillslope in northern New Mexico, United States. The model is evaluated against available observations of snow depth, soil moisture and runoff over two water years yielding reliable spatial distributions during the winter to summer transitions. We find that a wet winter followed by a dry summer promotes evapotranspiration losses that gradually dry the soil and disconnect lateral fluxes in the forested hillslope, leading to soil moisture patterns resembling vegetation patches. Conversely, a dry winter prior to a wet summer results in soil moisture increases that promote lateral connectivity and soil moisture patterns with the signature of terrain curvature. An opposing temporal switch between infiltration and saturation excess runoff is also identified. These contrasting responses indicate that the inverse relation has significant consequences on hillslope water availability and its spatial distribution with implications on other ecohydrological processes including vegetation phenology, groundwater recharge and geomorphic development.

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KEY WORDS forest hydrology; hillslope; soil moisture; snow depth; distributed hydrologic model; Southwestern United States

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INTRODUCTION

Precipitation in the Southwestern United States is characterized by a bimodal regime consisting of winter frontal storms and summer convective rainfall (e.g. Mock, 1996; Sheppard et al., 2002). The relative amount and timing of precipitation in each season vary from year to year in response to several synoptic-scale mechanisms that control winter-time Pacific storms and the summer-time North American monsoon (e.g. Cayan, 1996; Higgins and Shi, 2000). The proportion of annual precipitation falling within each season also varies geographically with a general increase of winter precipitation at higher latitudes and elevations (e.g. Douglas et al., 1993; Vivoni et al., 2008a; Forzieri et al., 2011). Interestingly, several studies have identified an inverse relation between the relative magnitudes of precipitation in each season and explored the potential mechanisms or teleconnections underlying the observations (e.g. Gutzler and Preston, 1997; Gutzler, 2000; Ellis and Hawkins, 2001; Zhu et al., 2005; McCabe and Clark, 2006; Mo, 2008; Notaro and Zarrin, 2011). In these studies, the snow amount and duration in the Rocky Mountains are considered to alter surface albedo, temperature and moisture, which affect the land–ocean thermal gradient with the East Pacific and the strength of the North American monsoon.

As noted by Notaro and Zarrin (2011), the existence of and predictability derived from the snow–monsoon inverse relation have broad implications for water resources and ecosystem productivity. Essentially, when the spatial extent and duration of winter snow cover are high, the amount of summer precipitation is lower, and vice versa, thus providing a prediction on summer conditions a few months in advance (Gutzler and Preston, 1997). To our knowledge, limited attention has been paid to the ecohydrological consequences of the inverse relation by analysing how landscapes respond to contrasting sequences of winter and
summer precipitations. Most studies have focused on precipitation and its associated ecohydrological response within single seasons (e.g. Seth et al., 1999; Kurc and Small, 2004; Vivoni et al., 2008b; Molotch et al., 2009). Despite that, Notaro et al. (2010) attributed the bimodal growth of ecosystems in the Southwestern United States to seasonal precipitation inputs, whereas Ogle and Reynolds (2004) indicated the need to assess the effects of winter and summer precipitations on plant responses across the region. An example of these interactions was discussed by Jenerette et al. (2010) who reported that increases in winter precipitation negatively impacted maximum vegetation growth in the summer.

In addition to effects on ecosystem productivity, relative amounts of winter and summer precipitations have important consequences on hydrologic systems in the Southwestern United States because (1) soil moisture is replenished through snowmelt inputs and summer convective storms (e.g. Ogle and Reynolds, 2004), (2) streamflow generation can have the signature of spring snowmelt and summer storms (e.g. Newman et al., 2006) and (3) evapotranspiration is affected by storm frequency through cloud cover (e.g. Mahmood and Vivoni, 2011a). Thus, a wet winter followed by a dry summer is hypothesized to lead to the drying of a hydrologic system as greater energy is added during the winter to summer transition. Prior studies have shown how wet conditions from snowmelt inputs are followed by gradual drying during the summer when evapotranspiration is higher (e.g. Newman et al., 1998; Molotch et al., 2009; Bales et al., 2011). Fewer efforts have documented the hydrologic dynamics occurring when a dry winter is followed by a wet summer as a lower amount of energy is inputted during the transition (Newman et al., 1998). A particularly interesting outcome of the inverse relation is the potential to alter soil moisture patterns and their underlying local and nonlocal controls as described by Grayson et al. (1997).

Detecting the ecohydrological consequences of the inverse relation in a comprehensive fashion from field studies alone is difficult because of the need for coordinated winter and summer observations. Distributed hydrologic models evaluated against field data can be a useful means to depict ecohydrological processes, to extrapolate limited field data over broader spatio-temporal scales and to track the underlying physical mechanisms of the winter to summer transition. To do so, a numerical model should represent cold and warm season processes, and their interaction, in a continuous and reliable manner relative to site observations. Here, we use the Triangulated Irregular Network (TIN)-based Real-time Integrated Basin Simulator (tRIBS) in a ponderosa pine field site to study the hydrologic dynamics of the winter to summer transition during two contrasting water years that correspond well to the inverse relation. To apply the model effectively to the forested site, we performed a series of modifications to the original formulation of Rinehart et al. (2008). Because the site is representative of ponderosa pine hillslope areas in Southwestern United States (Brandes and Wilcox, 2000) and the selected water years depict the snow–monsoon inverse relation (McCabe and Clark, 2006), the ecohydrological insights gained from the modelling study are relevant to forests throughout the region and for other years. Moreover, our focus on spatial patterns during the winter to summer transition sheds light on the controlling factors and thresholds for runoff production and lateral connectivity in forested hillslopes.

METHODS

Study area and hydrologic observations

The study area, a ponderosa pine (Pinus ponderosa) hillslope (35°53′N, 106°17′W, elevation of ~2315 m) is located in the Los Alamos National Laboratory (LANL), New Mexico, United States (Figure 1). A series of studies at the site between 1993 and 1998 collected a set of hydrologic datasets, including snow depth, soil moisture...
and surface runoff (Newman et al., 1997, 1998; Wilcox et al., 1997). Snow depth (cm) and volumetric soil moisture (m$^3$ m$^{-3}$) estimates are available at 11 sampling locations within the hillslope at variable intervals in time. Soil moisture was measured manually using a neutron probe (NP) at a 22-cm soil depth through access tubes, whereas snow depth was visually read from metre sticks at these sites. In addition, snow depth was inspected at 28 snow metre posts distributed throughout the hillslope. Here, we used the snow depth, soil moisture at 22 cm and hillslope outlet runoff (mm h$^{-1}$) obtained during the 1992–1993 and 1993–1994 water years (October to September) that constituted the most complete dataset for studying the winter to summer transition. Additional data from the site include hourly meteorological data (i.e. precipitation, air temperature, relative humidity, wind speed and solar radiation), which were applied as spatially uniform forcing at an hourly resolution. Unfortunately, there were large meteorological data gaps from October 1992 to February 1993, requiring the use of data from the TA-6 site at LANL, ~2 km to the east and at a lower elevation (2263 m).

In addition to hydrologic observations in the hillslope, we obtained a high-resolution (0.305-m) digital elevation model (DEM) and canopy height model from an aircraft survey using Light Detection And Ranging (LiDAR). These landscape datasets allowed deriving the hillslope domain (~1280 m$^2$ in size), including the boundary upstream of the runoff trench, the flow network and the local slope, aspect and curvature (Mahmood and Vivoni, 2011a). Overall, the hillslope has a low relief (~6 m) and a gentle slope from west to east following the general features of the Pajarito Plateau. The canopy height model allowed identifying the spatial locations of the open ponderosa pine stands (1–20 m in height, Figure 2) and their intercanopy grasses (0–1 m), which are characteristic of this region (e.g. McDowell et al., 2008). Soil stratigraphy is characterized as A and Bw horizons (loess deposit and sandy loam texture), a Bt horizon (alluvium), a clay-rich CB horizon (weathered tuff and clay texture) and an R horizon of Bandelier Tuff, with soil hydraulic properties and their variation with depth available at one location (Wilcox et al., 1997).

As an example of the distributed observations, Figure 2 presents the variation of snow depth for the five available dates in the 1992–1993 water year at the NP tubes (snow metre sticks). The spatial patterns show differences in snow accumulation and ablation between northern (1607–1611) and southern (1601–1606) sites in the hillslope. Southern sites are in intercanopy grasses and receive more snow with delayed ablation, whereas northern locations receive less snow and exhibit earlier ablation. One northern site (1608) is located underneath a ponderosa pine and accumulates almost no snow. Thus, the distributed observations provide insights into the processes leading to spatial variations in snow depth. Unfortunately, there were no equivalent data collected for the 1993–1994 water year at the NP tubes. The snow depths at the 28 snow metre posts (Figure 1) during both water years lacked information relating the site location to the observed values, restricting their use as a spatially averaged quantity.


![Figure 2. Spatio-temporal dynamics of observed snow depth during the 1992–1993 winter period.](image-url)
Distributed hydrologic modelling with a modified snow component

We used the tRIBS model for the hillslope simulations. tRIBS is a distributed hydrologic model with a snow component (Ivanov et al., 2004a, b; Vivoni et al., 2007; Rinehart et al., 2008). The physical processes in the model include rainfall interception, infiltration, evapotranspiration, groundwater fluctuations, lateral subsurface transport and runoff production and routing. Runoff generation occurs via four mechanisms (infiltration excess, saturation excess, groundwater exfiltration and perched return flow) that have been described in detail by Vivoni et al. (2007). Cold-season processes include snow interception and unloading, sublimation of intercepted and on-the-ground snow, snow accumulation and ablation, and infiltration of melt water (Rinehart et al., 2008). The distributed model incorporates hillslope descriptors of topography, vegetation and soil properties in the simulation at high resolution, when available. Here, we utilized the LiDAR DEM to generate a TIN at the finest possible resolution (0.305 m). Although coarser resolutions are possible without significant loss of hydrologic information (Vivoni et al., 2005; Mahmood and Vivoni, 2011b), this selection was made to represent the sampling sites with high fidelity.

Ivanov et al. (2004a, b) and Vivoni et al. (2007, 2011) presented a detailed description of the model domain, physical processes, parametrization and initialization, as well as the model capabilities to produce spatio-temporal estimates of hydrologic variables. Furthermore, the model application to the ponderosa pine hillslope has been fully documented by Mahmood and Vivoni (2011a, 2011b) for summer conditions, with a focus on reproducing observed soil moisture and runoff for three periods (1996–1998). As a result, we limit the following discussion of the model to the cold-season processes initially described by Rinehart et al. (2008) and updated in this study. The snow component is a single-layer, coupled energy and mass balance approach that accounts for direct and diffuse solar (shortwave) radiation and the age-dependent albedo effects of snow; incoming and outgoing longwave radiation; precipitation heat flux; and latent and sensible heat flux from the snowpack, including sublimation (e.g. Tarboton and Luce, 1996; Wigmosta et al., 1994; Wilson and Gallant, 2000). Incoming precipitation is linearly partitioned between liquid and solid phases using air temperature (Wigmosta et al., 1994) and used to estimate precipitation heat flux. When falling in solid form, vegetation intercepts snow on the basis of the leaf area index, unloads snow in relation to air temperature and sublimates snow using the absorbed shortwave radiation and relative humidity (Pomeroy et al., 1998; Liston and Elder, 2006).

The snowpack internal energy at each time step (U) determines the snow temperature (Tsn) and how liquid and solid phases are partitioned within the snowpack (see Rinehart et al., 2008, for details). We modified the original model to account for the latent heat leaving the snowpack upon melt by adding a term (Rinehart et al., 2008, Equation A2),

\[ L_m = \lambda_i j \rho_j M_{ij} \]  

(1)

where \( L_m \) is the latent heat flux, \( \lambda_i \) is the latent of freezing, \( \rho_j \) is the density of water and \( M_{ij} \) is the amount of phase change from ice (i) to liquid water (j). Overall, a positive U indicates the presence of liquid water in the snowpack, which can be held internally up to a fraction of 0.35 of the snow water equivalent (SWE) (modified from a fraction of 0.06 in Rinehart et al., 2008), with the remaining routed to the soil surface. We modified the fraction of retained liquid water to account for the shallower snowpack and the higher frequency of rain-on-snow events at the site as compared with those in Rinehart et al. (2008). Currently, the snow model neglects (1) shallow ground heat flux, (2) wind redistribution of snow, (3) local differences in meteorology from wind sheltering, (4) the effects of unstable temperature profiles on turbulent exchanges and (5) the effects of vegetation or topography on the incoming longwave radiation.

Rinehart et al. (2008) also described the model approaches to account for the impacts of topography and vegetation on the incoming solar radiation for both direct and diffuse terms. Topographic effects include both local controls of slope, aspect and plant canopies as well as remote controls from the surrounding landscape (e.g. distant mountains and their shading and reflection of light). For the hillslope application, we redefined the remote controls by using a simpler approach for the remote sky-view factor (vremote) following Dozier (1981),

\[ v_{remote} = \frac{1}{m} \sum_{d=1}^{m} \cos HA_m \]  

(2)

where \( HA_m \) is the horizon angle measured from the vertical in the azimuth direction \( m \) (16 total directions), instead of that in Rinehart et al. (2008, Equation 2). Furthermore, we utilized the LiDAR canopy height model to determine the remote controls, as the gentle relief of the hillslope made the effects of distant mountains negligible. This allowed for a more detailed treatment of vegetation effects on sky-view for the ponderosa pine hillslope. Other aspects of the shortwave radiation, including the treatment of albedo effects, remain as reported in Rinehart et al. (2008).

Distributed model application in ponderosa pine hillslope

Simulations in the ponderosa pine hillslope are based on a model domain with 12 755 Voronoi polygons, each characterized by elevation, soil and vegetation properties (Figure 3). By using the LiDAR-derived canopy heights, four static vegetation classes were mapped: grassland (< 1-m height) and short (1–5 m), medium (5–10 m) and tall ponderosa pines (10–20 m). Mahmood and Vivoni
(2011a) calibrated vegetation parameters for summer conditions in each class, finding a good match between simulated and observed soil moisture during recession periods. Because of a lack of distributed data, soil hydraulic properties and depth were assumed to be spatially uniform in the hillslope (sandy loam texture and 1.06-m depth) following Wilcox et al. (1997). Certain soil properties were also adjusted to match the observed soil moisture at distributed locations. Table I lists the vegetation and soil parameter values and describes whether these were obtained from field measurements, literature values or model calibration. Here, we relied on the calibration and testing of Mahmood and Vivoni (2011a) and did not further alter soil or vegetation parameters with one exception: The grassland vegetation fraction and height were reduced when covered by snow. This important change allowed capturing of the appropriate heat fluxes when a snowpack developed in the intercanopy grassland areas during the two winter to summer transitions.

The distributed simulations for the two water years were conducted as a single model run (October 1992 to September 1994) at an hourly time step. A lack of soil moisture observations prior to the simulation period prevented a distributed initialization as in Mahmood and Vivoni (2011a). Thus, a moderately wet, spatially uniform initial condition was assumed for 1 October 1992 as this followed the summer season. Any potential errors introduced by this assumption were minimized by the fall season evapotranspiration that reduced soil moisture in the hillslope prior to the onset of winter snowfall. Over the simulation period, spatially uniform forcing was applied for air temperature, relative humidity, wind speed and solar radiation above the canopy (after which modifications were made because of local and remote shading. During winter, however, the differential

Table I. Vegetation and soil parameter values from field observations, literature or manual calibration, including percentage of hillslope area (Area), throughfall coefficient (p), albedo (A) (Iziomon and Mayer, 2002), canopy water storage capacity (S), drainage rate from canopy (K), drainage exponential parameter (g) (Rutter et al., 1971), vegetation height (Hv), optical transmission coefficient (Kt) (Zou et al., 2007), minimum stomatal resistance (rs) (Karlson and Assmann, 1990; McDowell et al., 2008), vegetation fraction (v), hydraulic conductivity at surface (Ks), saturated (θs) and residual (θr) soil moisture (Rawls et al., 1983), soil moisture stress threshold (θ*), pore size distribution index (λ), air entry bubbling pressure (ψ), conductivity decay parameter (f) and soil anisotropy ratio (As).

<table>
<thead>
<tr>
<th>Vegetation units</th>
<th>Area (%)</th>
<th>p</th>
<th>S</th>
<th>K</th>
<th>g</th>
<th>A</th>
<th>Hv</th>
<th>Ks</th>
<th>rs</th>
<th>λ</th>
<th>ψ</th>
<th>f</th>
<th>As</th>
</tr>
</thead>
<tbody>
<tr>
<td>Grassland (0–1 m)</td>
<td>52</td>
<td>0.1a</td>
<td>1.0a</td>
<td>0.12a</td>
<td>4.7a</td>
<td>0.28a</td>
<td>1b</td>
<td>0.9a</td>
<td>40b</td>
<td>0.8c</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Short ponderosa pine (1–5 m)</td>
<td>20</td>
<td>0.4a</td>
<td>1.5a</td>
<td>0.12a</td>
<td>4.7a</td>
<td>0.2a</td>
<td>5b</td>
<td>0.5a</td>
<td>10b</td>
<td>0.85c</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Medium ponderosa pine (6–10 m)</td>
<td>15</td>
<td>0.4a</td>
<td>1.5a</td>
<td>0.12a</td>
<td>4.7a</td>
<td>0.1a</td>
<td>10b</td>
<td>0.5a</td>
<td>10b</td>
<td>0.95c</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tall ponderosa pine (10–20 m)</td>
<td>13</td>
<td>0.4a</td>
<td>1.5a</td>
<td>0.12a</td>
<td>4.7a</td>
<td>0.1a</td>
<td>20b</td>
<td>0.5a</td>
<td>10b</td>
<td>0.95c</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Soil unit</th>
<th>Ks</th>
<th>θs</th>
<th>θr</th>
<th>θ*</th>
<th>λ</th>
<th>ψ</th>
<th>f</th>
<th>As</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sandy loam</td>
<td>0.29b</td>
<td>0.45b</td>
<td>0.01a</td>
<td>0.18c</td>
<td>1.9c</td>
<td>-250</td>
<td>0.007b</td>
<td>40c</td>
</tr>
</tbody>
</table>

* Values from literature.

b Values from field observations.

c Values from manual calibration.
accumulation of snow in the northern and southern grassland areas of the hillslope (Figure 2) could not be simulated with uniform forcing. Thus, we explored several alternative hypotheses (e.g. spatial variation of air temperature; Karlsson, 2000) and found evidence for the effect of wind redistribution of snow, as shown in Figure 4a. During snowfall events, winds in the hillslope were primarily from the southerly direction, leading to higher accumulation in the open southern sites and tree-induced sheltering in the northern sites.

Because wind-driven snow redistribution is not currently simulated (Rinehart et al., 2008), we developed an obstruction map derived from the LiDAR canopy height model. For each pixel, we estimated the presence of an obstruction to wind in the eight surrounding directions by comparing the elevation of the selected pixel and its neighbours within a 3-m radial distance. If the elevation difference was greater than 5 m, we assigned 1 for that pixel and 0 otherwise, as

\[
O_{ij}^d = \begin{cases} 
1, & \text{if } (z^d_r < 3 \text{ m} - z_{i,j}) > 5 \text{ m} \\
0, & \text{if } (z^d_r < 3 \text{ m} - z_{i,j}) < 5 \text{ m} 
\end{cases}
\]

(3)

where \(O_{ij}^d\) is the obstruction index at site \(i, j\) in a direction \(d\). \(z^d_r<3\text{m}\) are all pixel elevations within a 3-m radius \((r)\) and \(z_{i,j}\) is the given pixel elevation. Each binary map for a direction \(d\) was then multiplied by the fraction of time that wind blows from that direction \(p_d\) during winter storms (Figure 4a) as \(B_{ij}^d = p_d O_{ij}^d\). The summation of \(B_{ij}^d\) over all directions leads to the final obstruction fraction map shown in Figure 4b with values ranging from 0 (no obstructions) to 1 (high obstructions) in the dominant wind direction. The sensitivity of the approach to changes in the radius and threshold elevation was minimal.

Thus, to account for wind redistribution of snow during storm events, we moved snow precipitation from sheltered areas (e.g. north of the ponderosa pine patches in Figure 4b) into open areas (e.g. south of the pine patches) following the obstruction fraction, while conserving the total snow mass input to the hillslope. The spatially variable precipitation forcing was utilized only during snowfall events in winter and early spring.

Two additional changes were made to the model to account for the patchy ponderosa stands and the relatively thin snowpacks. First, the minimum snow temperature \(T_{\text{min}}\) in the single-layer model (Rinehart et al., 2008) was replaced by the snow temperature \(T_{\text{sn}}\) occurring for SWE equal to 10 cm when the internal energy \((U)\) was less than zero. This allowed for a more stable \(T_{\text{sn}}\) during a rapidly melting snowpack. Second, the snowmelt water in the hillslope was retained on the soil surface until the snowpack melted within each melt period (i.e. SWE returned to zero during snow ablation). Subsequently, snowmelt was allowed to slowly infiltrate into the soil at a rate of 0.25 mm h\(^{-1}\), allowing for more soil absorption rather than runoff generation, consistent with the low-runoff observations in the trench during winter and the potential effects of frozen soil (Newman et al., 2004).

RESULTS AND DISCUSSION

In the following, we first describe an evaluation of the snow model at the Quemazon SNOTEL site to provide confidence in the modifications introduced. Then, we discuss the simulated ponderosa pine hillslope response to the two contrasting winter to summer transition periods. Comparisons with snow, soil moisture and runoff observations allow for a detailed evaluation of the modelled spatio-temporal patterns that reveal stark differences between the two water years. Finally, a detailed analysis tracks how precipitation during each season influences the hillslope patterns of ecohydrological response and how these are linked to landscape properties including terrain curvature and vegetation (Figure 3).

Snow model evaluation at the Quemazon SNOTEL site

To test the revised snow model physics, we carried out simulations at the Quemazon SNOTEL station (35°55′N,
106°24′W, Los Alamos, New Mexico) for multiple winter seasons with nearly complete meteorological and snow data (2004–2005, 2006–2007, 2008–2009 and 2009–2010). The Quemazon location is ~7 km northwest of the ponderosa pine hillslope and at a higher elevation of ~2900 m (Rinehart et al., 2008). The site is in a small meadow that is sheltered from winds by the surrounding forest such that snow interception processes or wind-induced undercatch of snow is considered negligible. Available data included hourly air temperature, precipitation and SWE measured by a snow pillow. We found that the cumulative precipitation data from the site weighing gauge could introduce some uncertainty to the total water input, in particular for the 2006–2007 winter season. Other meteorological variables, such as solar radiation, relative humidity and pressure, were obtained from the Los Posos weather station, located ~2 km west of Quemazon at a similar elevation, an improvement over the model forcing used by Rinehart et al. (2008) from a farther and lower site.

Figure 5 compares the observed SWE at the Quemazon SNOTEL site with simulated values at the co-located Voronoi polygon with the station for each winter season. Because a small domain was constructed around the station (1427 m² with 1495 Voronoi polygons derived from the LiDAR DEM), we are able to present the spatial variability around the station through the grey shading (±1 standard deviation of eight neighbouring polygons). Good model agreement is obtained for all winter seasons across a range of different precipitation and temperature conditions (Table II). The model captures the snow accumulation and ablation processes for each season and performs well in mimicking the peak SWE, ranging from ~20 to 38 cm. Overall, the root mean square error (RMSE) of the SWE is low (Table II), and the performance is superior to that of Rinehart et al. (2008, their Figure 4). This is attributed primarily to an improved model forcing and the effects of the revised snow physics. As expected, the seasonal weather plays a dominant role in the snow accumulation and ablation for each period, with more snow in 2004–2005 and 2009–2010 owing to the higher precipitation and lower mean air temperatures. For these wetter seasons, the snow duration extends into late May, whereas the drier winters have snow cover until late April. The snow model performance for multiple seasons builds confidence in the simulated processes for conducting spatially explicit snow simulations, as discussed in the following.

**Distributed snow conditions for contrasting winter seasons**

Figure 6 presents the observed and simulated snow depths at a selected number of snow meter sticks (at NP tube locations) for the 1992–1993 winter season. Visual measurements for five dates (black circles) are compared with the continuous snow depth simulations, presented as an average of the co-located Voronoi polygon with the site and its neighbouring polygons (shading represents ±1 standard deviation among the polygons). The sites depict the general behaviour in the southern portion of the hillslope (1602, 1604 and 1606) with higher snow...
accumulation (peak snow depth of 25–70 cm) and the northern grassland areas (1607, 1609 and 1611) with a thinner snowpack (peak depths of 20–30 cm). Note that the model is able to reproduce the snow observations well at the selected sites, with an RMSE ranging from 2·3 to 16·4 cm. The mean RMSE of snow depth in the hillslope across all sites was 18·3 cm ($R^2 = 0·5$). Large spatial variations of snow depth (e.g. shading in 1602 and 1606) are found for sites at the edge of a ponderosa pine patch because of the impact of snow interception by the canopy. Model performance at other locations is also comparable, including for 1608 under a ponderosa pine where little snow accumulates, except at 1601 (RMSE = 31·5 cm) where the simulated snow depth underestimates the wind-driven redistribution at this exposed (unobstructed) site.

The simulated snow accumulation and ablation at the southern and northern sites help to interpret the contrasting field observations. At the southern locations (1602, 1604 and 1606), snow depth peaks in January and persists for 60 days owing to the constant input of wind-redistributed snow. Snow ablation in southern grassland sites occurs rapidly in March because of increases in air temperature and the relatively high amount of incoming shortwave radiation at these exposed sites. In contrast, the northern grassland sites (1607, 1609 and 1611) exhibit a lower maximum snow depth owing to the obstructed nature of these locations (note the lower precipitation input) that persists for a shorter time, ranging from 30 to 60 days. The snow ablation characteristics are fundamentally different among the sites, with a more gradual decrease in snow depth at northern locations that are less exposed to incoming solar radiation.

Because of data limitations, the winter simulations in the two water years are compared only in terms of the spatially averaged snow depth across all available sites (both snow metre sticks and snow posts). Figure 7 presents the mean observed (black circles) and simulated (black lines) snow depths, along with measures of the spatial variability ($±1$ spatial...
standard deviation) at the sampling sites. In addition, the spatially averaged snow depth simulated over the entire hillslope is presented (grey lines). The similarity between the spatial averages of snow depth over the hillslope and sampling sites indicates that these are representative of the entire domain. Clearly, the model is able to reproduce well the contrasting snowpack development in each winter season, with an RMSE of 18·3 cm ($R^2 = 0·5$) and 9·0 cm ($R^2 = 0·54$) across all sites in the hillslope. In 1992–1993, a sequence of snow storms under cooler weather during November leads to snow accumulation throughout the winter, resulting in a continuous and thick snowpack with high sublimation (47% of snowfall sublimates). In contrast, the infrequent precipitation events in 1993–1994 arrived during February under warmer conditions, leading to a late snowpack development that was thinner and discontinuous. Under these drier and warmer conditions, the snowpack has less exposed time with the atmosphere with lower sublimation (28% of snowfall sublimates) and also was subject to numerous snowmelt periods.

To further compare the winter seasons, Figure 8 presents the spatial distribution of time-averaged snow depth, snow cover duration ($T_d$), total snowmelt and total canopy sublimation ($S$). As expected, snow depth is higher in open grassland areas as compared with that in the ponderosa pine patches, although the spatial variations are minimal for 1993–1994. $T_d$ resembles the spatial pattern of snow depth, with longer periods in intercanopy grassland sites, in particular for 1992–1993. The northern sampling sites can be distinguished well from other grassland areas by a lower $T_d$ for both seasons because of the effects of tree sheltering. Differences between each season are pronounced in terms of the magnitudes of snowmelt and canopy sublimation, although the spatial patterns are similar in each winter. Interestingly, the snowmelt delivered to the soil surface is higher for the drier and warmer 1993–1994 period. This can be attributed to (1) the lower canopy sublimation due to the shorter $T_d$ and (2) the warmer temperatures that lead to more frequent snowmelt periods. In contrast, the wetter and colder 1992–1993 winter has a higher $S$ because of a greater snow cover duration promoting losses to the atmosphere rather than snowmelt. Note that the more complex spatial patterns of snowmelt, as compared with sublimation, indicate that this flux is dependent on several interacting processes, leading to spatially variable water inputs to the hillslope soil surface, as discussed in the following.

**Distributed soil moisture and runoff generation in contrasting water years**

Figure 9 compares observed and simulated soil moisture at selected southern and northern sites during the two water years. Overall, the model captures well the soil moisture dynamics during winter, spring and summer seasons in 1992–1993 (RMSE from 0·04 to 0·15 m$^3$ m$^{-3}$, with an RMSE of 0·09 m$^3$ m$^{-3}$, $R^2 = 0·5$, across all sites), despite no further calibration beyond that of Mahmood and Vivoni (2011a). This is complemented with a comparison of
spatially averaged soil moisture at all sites and the hillslope domain in Figure 7. Note that there were no soil moisture observations in the 1993–1994 water year, limiting the capacity to further test the model. Despite this, the model helps to identify the hillslope response to the variable snowmelt and rainfall inputs during the contrasting periods. Note that the soil moisture initial condition is quickly dissipated by the first winter. In 1992–1993, soil moisture exhibits a strong seasonality with a wet winter followed by drying during the spring because of elevated evapotranspiration (ET) caused by high radiation and air temperatures. In the summer, the drying trend is briefly interrupted by small rainfall pulses that rapidly increase soil moisture, but these amounts quickly recede because of high ET (Mahmood and Vivoni, 2011a). In contrast, the 1993–1994 water year consists of drier winter soils that experience brief episodes of wetting from snowmelt inputs that are triggered by warmer temperatures. The spring and summer experience a sequence of frequent, large storms that induce higher soil moistures, in spite of the high ET rates. As a result, a strong seasonality is observed with wetter soils during the North American monsoon.

Contrasts between the two water years are shown in terms of the spatial distribution of soil moisture during winter, spring and summer seasons in Figure 10. Soil moisture is a good indicator of the hillslope ecohydrological processes as it responds to atmospheric inputs and losses. For example, the winter soil moisture in 1992–1993 has wet grassland areas and dry ponderosa pine patches induced by differences in snowmelt. In the spring,
Grasslands dry at a faster rate than do ponderosa pines, leading to a nearly uniform and dry summer soil. Spatial maps of the temporal standard deviations in 1992–1993 show higher variability in grassland areas, consistent with the patterns described by Mahmood and Vivoni (2011a) for the drier summers of 1996 and 1998. In contrast, a uniformly dry soil condition is observed during the winter in 1993–1994, with low temporal variations throughout the hillslope. This is followed by a wetting period during the spring and summer, resulting in a progressively wetter soil moisture distribution that resembles the vegetation pattern, with wetter grassland sites as compared with the ponderosa pine patches. The spatial distribution of the mean and standard deviation in the summer of 1993–1994 obtains the signature of the terrain curvature, as shown by Mahmood and Vivoni (2011a) for the wetter summer of 1997. As a result, differences in the winter to summer transition can lead to substantially different controls on hillslope soil moisture patterns.

Snowmelt or rainfall events can induce runoff generation in the hillslope as shown by Wilcox et al. (1997) and Newman et al. (2004) at the site. Figure 11 compares trench runoff observations with simulated values at the hillslope outlet without additional calibration. The model is able to capture the major events during the 1993–1994 summer season (see Guan et al., 2010, for a similar comparison). However, the model overestimates runoff during the 1992–1993 water year (with the potential reason being inaccurate runoff data transcription). The model also provides insight into the fraction of hillslope runoff...
contributed by different runoff mechanisms during each event. In 1992–1993, winter and early spring runoff is dominated by the saturation excess mechanism due to the wet soil moisture condition. As the hillslope dries out in spring and summer, the dominant mechanism switches to infiltration excess runoff. Thus, a transition in runoff generation is detected in the model when a wet winter is followed by a dry summer. The opposite behaviour is simulated during the 1993–1994 water year, with a transition from infiltration excess runoff in the early spring to saturation excess runoff in the summer. As a result, the relative wetness conditions in each season and their sequencing lead to significantly different switching of runoff generation mechanisms at the hillslope scale.

**Contrasting ecohydrological responses at site to hillslope scales**

A detailed analysis of the hillslope ecohydrological response for the contrasting water years can reveal how the bimodal precipitation regime influences the underlying mechanisms at scales ranging from single sites to the entire domain. Figure 12 presents the hydrologic dynamics at southern (1604) and northern (1611) grassland sites, including snow depth, soil moisture, latent heat flux from the land surface ($ET$) and snow surface (sublimation), runoff, lateral subsurface flow and depth to groundwater. The southern grassland site receives more snow during both winters, although the difference with the northern site is greater for 1992–1993. Snowmelt from the southern site leads to a saturated soil profile and groundwater depth near the surface for the wet winter in 1992–1993, which is not present at the northern site. Wetter soils at the southern grassland also induce more runoff generation through the saturation excess mechanism during the winter and spring of 1992–1993. In contrast, winter wetting at the northern site elevates soil moisture more moderately, thus reducing runoff generation and promoting lateral flow away from the site. During the drier 1993–1994 winter season, marked spatial differences in hydrologic dynamics are not observed. The reduced snowpacks at both sites have lower sublimation, whereas the warmer temperatures in the spring lead to snowmelt-induced soil moisture increases and lateral flow (or subsurface flux). Soil water availability in the 1993–1994 water year is also attributed to lower evapotranspiration from more frequent cloud cover in the spring and summer.

Figure 13 presents a summary of the hydrologic contrasts between the two water years based on the spatially averaged water balance, $\Delta S/\Delta t = P - ET - Q$ for each season, where $\Delta S$...
is the total change in storage (snow and soil moisture), $P$ is the total precipitation (snow and rain), $ET$ is the total losses to the atmosphere (sublimation and evapotranspiration) and $Q$ is the total runoff (omitted owing to its low values for these two water years and at the seasonal scale). In all cases, the water balance components are presented as mean seasonal quantities (symbols) with their ±1 spatial standard deviations (vertical bars). Clearly, the two water years exhibit opposing behaviours during the winter to summer transition: (1) 1992–1993 has a decreasing $\Delta S$ in time resulting from an increasing $ET$ that exceeds $P$, thus depleting both the snowpack and the soil water storage, and (2) 1993–1994 has an increase in $\Delta S$ in time, primarily because of soil water availability from periods of higher $P$ than $ET$. This summary highlights how the differential sequencing of precipitation during winter and summer seasons (wet-to-dry or dry-to-wet) impacts the ecohydrological response of a forested hillslope, leading to either land surface water depletions to the atmosphere or water inputs from the

Figure 13. Seasonality in spatially averaged responses during 1992–1993 and 1993–1994 for winter, spring and summer seasons, including precipitation ($P$), evapotranspiration ($ET$), storage change ($\Delta S$), snow depth and soil moisture. Vertical bars depict ±1 spatial standard deviation.

atmosphere that remain in storage and may be redistributed internally as lateral subsurface flow or runoff.

To investigate if the winter to summer transitions impact lateral transport, we derived an index of hydrologic connectivity. This dimensionless index was obtained as the hillslope areal fraction with root zone (top 1 m) soil moisture above a certain threshold. Three thresholds were based on the work of Newman et al. (2004) who found that beyond a root zone moisture of 0.33 m$^3$m$^{-3}$, a lateral connection was established in subsurface macropores (represented in the model by the anisotropy of saturated hydraulic conductivity, Mahmood and Vivoni, 2011a). Two other thresholds (0.28 and 0.38 m$^3$m$^{-3}$) are used to test the sensitivity of the connectivity, which ranges from 0 (disconnected) to 1 (fully connected). Figure 14 presents the hydrologic connectivity for each water year, along with spatially averaged hillslope conditions for reference. In 1992–1993, the hillslope is well connected (index values greater than 0.5 for all thresholds) during the winter season because of moderately wet initial conditions and snowmelt-induced infiltration. During spring, lateral connectivity remains constant followed by a sudden decrease (index values fall to zero) in early summer due to the coincident rise in evapotranspiration. This is consistent with the soil moisture pattern resembling the vegetation distribution for the dry summer. A contrasting behaviour is observed in the hydrologic connectivity of the 1993–1994 water year. During the dry winter, the hillslope is disconnected in terms of lateral fluxes for all thresholds. As hillslope wets up with a series of rain and snow events in the spring, connectivity increases to values above 0.4 for all thresholds. Consistent with prior analyses, the wet summer in 1993–1994 increases the lateral connectivity because of high rainfall and low evapotranspiration, leading to soil moisture patterns that resemble the terrain curvature distribution (Mahmood and Vivoni, 2011a).

SYNTHESIS AND CONCLUSIONS

Bimodal precipitation in the winter and summer seasons is an important climate feature of the Southwestern United States. Precipitation amounts in each season have been hypothesized to be linked physically through several proposed pathways (e.g. Gutzler and Preston, 1997; Gutzler, 2000; Zhu et al., 2005; McCabe and Clark, 2006; Mo, 2008; Notaro and Zarrin, 2011). In this study, we analyse the ecohydrological response in a ponderosa pine hillslope during two contrasting winter to summer transitions: a wet winter followed by a dry summer (1992–1993) and a dry winter followed by a wet summer (1993–1994). These water years represent well the inverse relation between winter and summer precipitations found in prior studies. As a result, the outcomes obtained for these years are considered robust for similar winter to summer transitions in the historical record. We use a distributed hydrologic model tested against field observations to provide spatio-temporal estimates of hillslope states and fluxes, including snow cover, soil moisture and runoff. Used in this way, the model is a tool for interpreting the plausible physical mechanisms that underlie the contrasting responses to the seasonal precipitation and as a means for generating a consistent set of spatially distributed hydrologic estimates (see Vivoni, 2012, for further discussion). Although conducted at the hillslope scale, the study results shed light upon the forest ecohydrological response in ponderosa pine ecosystems that are commonly found throughout the Southwestern United States (Brandes and Wilcox, 2000).

Winter to summer season simulations at the hillslope scale are challenging because of the high number of process representations and the hydrologic variations found over short distances. For example, accounting for the role of wind redistribution on snow through simplified tree sheltering was essential for reproducing the available snow depth measurements (Wilcox et al., 1997). For wet winters, spatial differences in snow depth induced by tree sheltering propagated to melt water infiltration and soil moisture patterns during the spring. For dry winters, spatial variations are muted and the role of vegetation on soil moisture patterns is limited to effects on summer interception and evapotranspiration. For the purposes outlined earlier, the model performance is considered to be good when compared with the snow depth, soil moisture and runoff observations, in particular in light of the number of interacting processes. The spatio-temporal simulations for the two water years are consistent with summer simulations (1996–1998) conducted by Mahmood and Vivoni (2011a) in several ways: (1) The same set of model parameter reproduced available data, (2) similar vegetation and terrain curvature controls were identified on soil moisture patterns, and (3) model performance relative to observations was similar. As a result, including the winter period in the continuous simulation allowed a detailed investigation of the winter to summer transition. Other winter to summer season transitions could also be analysed within this same framework. For example, the site observations indicate that the dry summer in 1996 was preceded by a wet winter (160 mm), whereas the wet summer in 1997 had a dry (14 mm) antecedent winter. Thus, the inverse relation could be applicable over other years, as shown by McCabe and Clark (2006), and form the basis for a long-term modelling study.

The impact of the contrasting winter to summer transitions on the hillslope ecohydrological response can be summarized as follows. (1) Wet winters lead to a substantial snowpack in ponderosa pine forests, which increases snowmelt input into soils throughout the winter and spring, despite the high sublimation losses. When followed by a dry summer with low cloud cover (22% of day time), evapotranspiration increases substantially,
leading to drier soils; a switch occurs from saturation excess to infiltration excess runoff mechanisms; lateral subsurface flow diminishes such that hydrologic connectivity is reduced after snowmelt; and the drier soil moisture pattern in the summer resembles vegetation patches. (2) Dry winters lead to a reduced snowpack in the forested landscape that is exposed to less sublimation and yields higher proportional snowmelt inputs into the soil during the spring. When followed by a wet summer with frequent cloud cover (36% of day time), soil moisture is preserved in the subsurface because of precipitation exceeding evapotranspiration rates \((P > ET)\); runoff generation progressively favours the saturation excess mechanism; and a lateral connection is established in the hillslope such that the wetter soil moisture patterns contain the signature of the terrain curvature distribution.

To our knowledge, this is the first attempt to compare the ecohydrological consequences of contrasting winter to summer transitions at hillslope or watershed scales in the Southwestern United States. At regional scales, however, several studies have used hydrologic models to study the land influence on the inverse relation (Zhu et al., 2005, 2007; Notaro and Zarrin, 2011). Our modelling approach complements these regional efforts by (1) providing spatial details on snow and soil moisture distributions and the influence of vegetation and topography (Rinehart et al., 2008), (2) revealing the underlying physical processes involved in the link between winter and summer seasons, and (3) predicting the consequences on runoff production and hydrologic connectivity in the subsurface. In particular, the high-resolution modelling approach allowed distinguishing of the land surface states and implications on vegetation phenology (Jenerette et al., 2010). The contrast between winter and summer transitions should have implications on vegetation phenology (Jenerette et al., 2010), groundwater recharge (Small, 2005) and geomorphic development (Etheredge et al., 2004), which merit additional attention.

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