

Climate regime and soil storage capacity interact to effect ET

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Climate regime and soil storage capacity interact to effect evapotranspiration in western United States mountain catchments

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Abstract

In the winter-wet, summer-dry forests of the western United States (US), total annual evapotranspiration (ET) is largely a function of three separate but interacting properties: (1) climate, especially magnitude of precipitation, its partitioning into rain or snow, and snowmelt timing; (2) soil characteristics, including soil water holding capacity and rates of drainage; and (3) the total biomass where larger, more abundant vegetation is directly proportional to greater ET. Understanding how these controls influence ET in Mediterranean mountain environments is complicated by shifts between water and energy limitations both within the year and between years. We use a physically based process model to evaluate the strength of climate controls and soil properties in predicting ET in three snow-dominated, mountainous catchments in the western US. As we expect, statistical analysis shows that annual precipitation is a primary control of annual ET across all catchments. However, secondary climate controls vary across catchments. Further, the sensitivity of annual ET to precipitation and other climatic controls varies with soil characteristics. In the drier, more snow-dominated catchments ET is also controlled by spring temperature through its influence on the timing of snowmelt and the synchronicity between seasonal water availability and demand. In wetter catchments that receive a large fraction of winter precipitation as rainfall, the sensitivity to ET is also strongly influenced by soil water holding capacity. We show that in all catchments, soil characteristics affect the sensitivity of annual ET to climatic drivers. Estimates of annual ET become more sensitive to climatic drivers at low soil water holding capacities in the catchments with the stronger decoupling between precipitation and growing season demands.

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1 Introduction

Accurate estimation of evapotranspiration (ET) in the western US is complicated by interactions between the timing and magnitude of climatic drivers, landscape topographic features and vegetation distribution (Hamlet et al., 2007; Viola et al., 2008). In this region, forest ET is estimated to account for 30–80 % of the annual water budget (Baron and Denning, 1992; Jeton et al., 1996; Waichler et al., 2005) and water availability is considered to be its primary control (Boisvenue and Running, 2006). However, in the mountainous headwater catchments, much of the seasonal water availability is temporarily disconnected from demands. This region's mountain watersheds receive a large fraction of their annual precipitation (P) as snow in the winter. The magnitude and duration of snowpack varies geographically with climatic drivers such as gross precipitation and regional temperature patterns (Hamlet et al., 2005; Mote, 2006) and vegetation (Molotch et al., 2009; Varhola et al., 2010) and its melt is the primary source of water to headwater forest catchments and downstream municipalities in spring and summer (Bales et al., 2006). The steep topography within these catchments strongly controls P , temperature (T), and their interaction (Daly et al., 1994) leading to variability in snowpack magnitude and duration and forest productivity across an elevation gradient (Bales et al., 2011; Goulden et al., 2012; Littell et al., 2008). This spatiotemporal variability in climatic interactions influencing water availability complicates our ability to understand how climate change will impact forest ET.

Warmer temperatures in the western US have affected hydrology through reductions in snowpack accumulation, due in part to precipitation falling as rain instead of snow (Knowles et al., 2006), and earlier occurrence of peak snowpack (Mote et al., 2005). These changes to winter precipitation shift seasonal water availability. Spring streamflow pulses are occurring up to four weeks earlier in the season (Stewart et al., 2005) which reduces plant water availability by drying out soils in the late growing season (Hamlet et al., 2007). Increased temperatures also effect plant responses through plant phenology shifting to earlier in the year (Cayan et al., 2001) which, in turn, may

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increase springtime productivity and extend the growing season. Though early season productivity may increase, the growing season is lengthened which can lead to soil water deficits later in the season. Soil dry-down reduces stomatal conductance and increases respiration costs which reduce overall productivity (Goldstein et al., 2000; Hu et al., 2010). These changes in water and energy inputs are expected to intensify (Ashfaq et al., 2013), and whether their interactions will increase or decrease ET remains uncertain.

Soil moisture characteristics have been shown to mitigate the effects of climate interactions on ET to varying degrees depending on precipitation regime (Hamlet et al., 2007; Porporato et al., 2004). In western US mountains, soil affects the timing of winter water availability into the growing season by two mechanisms: redistribution and storage. Steep topography extends plant water availability at lower elevations later into the growing season through lateral redistribution of snowmelt that recharges soil stores. Thus subsurface drainage can provide moisture to points on the landscape (i.e., riparian zones and swales) during seasonal summer drought, contributing to vegetation presence and enhancing ET (Hamlet et al., 2007; Hwang et al., 2011; Voepel et al., 2011). Modeling studies have highlighted that by including lateral redistribution of soil moisture, ET is more accurately captured in the Sierra Nevada (Lundquist, 2011; Tague and Peng, 2013) and in the Colorado Rockies (Litaor et al., 2008). Like snowpack, soil has the potential to act as a water reservoir, storing winter precipitation into the growing season (Geroy et al., 2011). However the amount of snowmelt captured and held by soil is strongly controlled by precipitation regime, snowpack dynamics, and soil characteristics. For example, Litaor et al. (2008) showed a positive correlation between amount of snowfall and soil moisture, with summer P offsetting soil moisture depletion. In contrast, shallow soils in Idaho's Dry Creek Experimental Watershed did not have the capacity to capture enough snowmelt to have a large effect on ET (Smith et al., 2011). Because substantial heterogeneity in soil characteristics exists not just across, but also within mountain watersheds (Dahlgren et al., 1997; Denning et al., 1991; McGuire et al., 2007), we hypothesize that there is a resulting variability in soil

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Water availability is influenced by both the magnitude of annual precipitation and the ability of a landscape to retain water, through soil water holding capacity and lateral moisture redistribution. The magnitude of annual precipitation, its seasonal distribution and, especially in the western US, its partitioning to rain or snow strongly influence water supply. A key factor in meeting PET demands is the timing of water availability relative to timing of demand. Mediterranean climates are often characterized by a disconnect between seasonal water availability and growing season water demand when, for example, precipitation falls as rain in winter or snowmelt recharge occurs prior to onset of the growing season water demand.

Our conceptual model defines two distinct types of controls, climate and geology, that can directly alter this disconnect between water availability and demand. Firstly, climate varies the timing of precipitation events, transitions between snow and rain, and the timing and rate of snowmelt all of which can all alter the timing of soil moisture recharge. At the same time, climate (specifically temperature) can also alter water demand by extending the period over which high water demand occurs. Secondly, geologic and topographic properties of the landscape determine how much recharge is stored, and potentially allow water that falls earlier in the season to be available for ET at a later date. Soil available water capacity (AWC) is a landscape feature that is relatively constant through time. Larger AWC indicates that more water can be held in the soil. These soil moisture stores are replenished in the winter and drained by subsurface flow and ET. We expect that in Mediterranean environments ET will typically increase with higher AWC because more of the recharge will infiltrate into soil stores and be available for plants instead of being lost as runoff. Soil interacts with climate to extend plant water availability by capturing snowmelt, one of the primary sources of water for forest ET. If the rate of snowmelt allows for soil stores to be replenished later into the growing season, more of the winter precipitation is made available for plant water use. If, however, soils are too shallow to capture a significant amount of runoff or if the rate of rain or snowmelt inputs exceeds the rate of infiltration, then soil will have little ability to

extend water availability. Therefore, soil potentially alters the effect of climate on ET, but the degree of that effect depends on regional climate drivers relative to soil properties.

Though this conceptual model relies on uncontroversial drivers of ET, it focuses on the importance of their interactions in space and time. Because water availability limits ET, regional precipitation regime is a primary control of these interactions. Regional variation in the fraction of precipitation falling as snow can act as a first order effect on climate interactions with ET. Soil potentially changes climatic interactions to extend water availability and though its characteristics do not change over time, it is spatially variable both between and within watersheds. In order to address the strength of these interactions, and their variability in different geoclimatic conditions, we use an eco-hydrologic model that represents the relevant processes in our conceptual model.

2.2 Model

We use a physically based model, the Regional Hydro–Ecological Simulation System (RHESSys), to calculate vertical water, carbon and nitrogen fluxes in three watersheds (Tague and Band, 2004). It is a spatially explicit model that partitions the landscape into units representative of the different hydro-ecological processes modeled (Band et al., 2000). The watershed delineates the largest unit and is divided into hillslopes, which drain to the stream via explicit subsurface routing. The smallest spatial unit, the patch, is the scale at which vertical soil moisture fluxes and carbon and nitrogen cycling occurs. The model is more fully described in Tague and Band (2004), but as it is under continuous development we note that version 5.15 is used in this work.

RHESSys requires data describing spatial landscape characteristics and climate forcing; a digital elevation model (DEM), soil and vegetation maps are used to represent the topographic, geologic, carbon and nitrogen characteristics within a watershed. MTN-CLM algorithms are used for extrapolation of climate processes (Running and Nemani, 1987) from point station measurements. Hydrologic processes modeled in RHESSys include interception, evapotranspiration, infiltration, vertical and lateral subsurface drainage, and snow accumulation and melt. The Penman–Monteith formula

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(Monteith, 1965) is used to calculate evaporation of canopy interception, snow sublimation, evaporation from soil and litter stores and transpiration by leaves. A model of stomatal conductance allows transpiration to vary with soil water availability, vapor pressure deficit, atmospheric CO₂ concentration, and radiation and temperature (Jarvis, 1976). A radiation transfer scheme that accounts for canopy overstory and understory, as well as sunlit and shaded leaves, controls energy available for transpiration. RHESys accounts for changes in vapor pressure deficit for fractions of days that rain occurs (wet vs. dry periods). Plant canopy interception and ET are also a function of leaf area index (LAI) and gappiness of the canopy such that as LAI increases and gap size decreases, plant interception capacity and transpiration potential increases. Soil water availability varies as a function of infiltration and water loss through transpiration, evaporation and drainage. RHESys also routes laterally and thus patches can receive additional moisture inputs as either re-infiltration of surface flow or through shallow subsurface flow from upslope contributing areas. Lateral subsurface drainage routes subsurface and surface water between spatial units and is a function of topography, surface, soil and saprolite drainage characteristics. Deep groundwater stores are drained to the stream using a simple linear reservoir representation.

Carbon and nitrogen cycling in RHESys was modified from BIOME-BGC (Thornton, 1998) to account for dynamic rooting depth, sunlit and shaded leaves, multiple canopy layers, variable carbon allocation strategies and drought stress mortality. The Farquhar equation is used to calculate gross primary productivity (GPP) (Farquhar et al., 1980). Plant respiration costs include both growth and maintenance respiration and are influenced by temperature following Ryan (1991). Net primary productivity (NPP) is calculated by subtracting total respiration costs from GPP.

In our three study sites (described in more detail below), RHESys is driven with daily records of precipitation and maximum and minimum temperature. Each basin is calibrated for seven soil parameters that characterize subsurface drainage properties. Soil drainage rates are controlled by saturated hydraulic conductivity (K) and its decay with depth (m). Soil air-entry pressure (φ_{ae}), pore size index (b), and rooting depth (Z_r)

2013; Winkler et al., 2014). We note that future work will study the behavior of ET following disturbance that changes canopy structure on the landscape. Of the climatic metrics from our conceptual model, we focus on those that are most affected by climate change. We first examine the sensitivity of ET to climate drivers predicted to have increasing variability, specifically precipitation and temperature. We then examine how soil water storage mediates the sensitivity of ET to climate drivers.

We used statistical analysis to identify a season when temperature most strongly influences estimates of annual ET modeled using historic climate. We performed linear regressions between total annual ET and 1 and 3 month averages of daily maximum (T_{\max}), minimum (T_{\min}) and average temperatures ($T_{\text{avg}} = (T_{\max} + T_{\min})/2$) for all watersheds and for all months of the year. We test the correlation significance with a p value and set a threshold at 0.05, i.e., a p value greater than 0.05 is not significant. Our analysis found a three-month average of daily T_{avg} in April, May and June (T_{AMJ}) to have the greatest explanatory power as a temperature variable for estimating interannual variation in annual ET under historic climate variability across our three study watersheds (results not shown). We note that the p value for T_{AMJ} in our warmer watershed (OR-CAS) was greater than 0.05 so it is not reported as a significant result.

In these seasonally water-limited basins, we use total annual precipitation (P) as a metric of gross climatic water input. Annual precipitation P is summed over a water year (1 October to 30 September of the following calendar year). To assess the impact of timing of soil moisture recharge (as influenced either by year to year variation in the precipitation timing, snowmelt or rain-snow partitioning) we calculate R_{75} , the day of water year by which 75% of the total annual recharge has occurred. Recharge is defined as liquid water (e.g. rain throughfall or snowmelt) that reaches the soil surface. For this metric, we do not differentiate between water that, upon reaching the soil surface becomes runoff, and water that infiltrates into the soil.

We examine the role of storage through soil available water holding capacity (AWC), calculated as the difference between soil field capacity and the plant wilting point. AWC represents the water stored in the soil after gravity drainage (field capacity) that can be

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extracted by plant root suction (wilting point), and is thus still viable for plant water use (Dingman, 1994, p. 236). We calculate AWC as:

$$AWC = (\theta_{fc} - \theta_{wp})Z_r \quad (1)$$

where θ_{fc} represents the soil's field capacity per unit depth, θ_{wp} the soil's characteristic wilting point also per unit depth, and AWC is scaled by vegetation rooting depth, Z_r , a model calibration parameter. The field capacity and wilting point are calculated, respectively, as

$$\theta_{fc} = \phi(\varphi_{ae}/0.033)^b \quad (2)$$

$$\theta_{wp} = \phi(\varphi_{ae}/\psi_v)^{1/b} \quad (3)$$

where ϕ is soil porosity, φ_{ae} represents the soil air-entry pressure (in meters), b is a pore size distribution index that describes the soil moisture-characteristic curve, and ψ_v describes the pressure at which the plants' stomata close. Variables φ_{ae} and b are also model calibration parameters. The range in soil parameter values is bounded by literature values specific to each site's soil properties (Dingman, 1994). This means all values of AWC calculated in calibration represent physically feasible soils for each watershed.

2.3.2 Use of process model to test conceptual model

The strength of these metrics of water availability and demand as controls on annual ET is quantified using linear regression analysis. We use RHESSys to calculate total annual ET in each basin over the entire available climate record (25–45 yr). We examine the controls on inter-annual variability of ET by regressing our climatic indicators – P , T_{AMJ} , and R_{75} – against total annual ET. We continue to use a threshold value of p values less than 0.05 to determine significance. We then investigate how long-term mean ET and its relationship with these climatic indicators are influenced by AWC. As noted above, all calibration parameter sets are physically viable representations of soils

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total annual P across basins. For example, OR-CAS receives the most precipitation and also supports stands of large, old-growth forests; its LAI is more than twice that of either CO-ROC or CA-SIER. As presented in the model description (Sect. 2.1), we use a stable, climatic optimum for vegetation biomass for all analyses in this paper. Garcia et al. (2013) and Tague et al. (2013) describe the soil and climate data, model vegetation and soil carbon store spin-up and calibration used for model implementations of OR-CAS and CA-SIER, respectively. Though RHESSys has previously been used in CO-ROC (Baron et al., 2000), we have made significant updates in RHESSys since that time, so we re-implemented the model as described in the next section.

RHESSys model development for CO-ROC

In CO-ROC, landscape topographic characteristics including elevation, slope and aspect were derived from a digital elevation model (DEM) downloaded from the US Geologic Survey (USGS) National Elevation Data set at 1/3 arc second resolution <http://datagateway.nrcs.usda.gov/>; accessed June 2011). A stream network was then derived so to accumulate surface and subsurface flow at USGS gage #06733000. Subcatchments were delineated using GRASS GIS's watershed basin analysis program, *r.watershed*. Terrestrial data was aggregated such that the average size of the patch units, the smallest spatial units for calculation of vertical model processes, was 3600 m². Soil classification data was downloaded from the Soil Survey Geographic database (SSURGO) and aggregated to four primary soil types: gravelly loam, sandy loam, loamy sandy and rock <http://datagateway.nrcs.usda.gov/>; accessed June 2010). Parameter values associated with these soil types are based on literature values (Dingman, 1994; Flock, 1978) and adjusted using model calibration, as described above. Vegetation land cover from the National Land Cover Database (NLCD) was aggregated to four primary vegetation types: subalpine conifer, aspen, shrubland and meadows (Homer et al., 2007). Because a shift in climatic regimes occurs at approximately 2700 m, we use daily records of precipitation, T_{\max} , and T_{\min} from two points within the watershed. RHESSys then interpolates data from these points based on

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MTN-CLM (Running and Nemani, 1987) to provide spatial estimates of temperature, precipitation and other meteorologic drivers for each patch. Climate data from 1980–2008 was downloaded from the DAYMET system for two locations – one at elevation 2460 m (latitude 40.35389, longitude –105.58361) and the second at 3448 m (latitude 40.33769, longitude –105.70315) (Thornton and others, 2012).

Plant carbon (C) and nitrogen (N) stores were initialized by converting remote-sensing derived LAI to leaf, stem and woody carbon and nitrogen values using allometric equations appropriate to the vegetation type <http://daac.ornl.gov/MODIS/MOD15A2> Collection 5; accessed June 2011). In order to stabilize soil C and N stores relative to the LAI-derived plant C and N, we run the model repeatedly over the basin's climate record until the change in stores stabilizes (Thornton and Rosenbloom, 2005). After stabilizing soil biogeochemical processes, we remove vegetation C and N stores and then dynamically “regrow” them using daily allocation equations (Landsberg and Waring, 1997) for 160 yr in order to stabilize plant and soil C and N stores with model climate drivers. For all three basins, an optimum maximum size for each vegetation type was determined using published, field-derived estimates of LAI and aboveground and total annual NPP.

3 Results

3.1 Annual P vs. ET

Not surprisingly, in all watersheds higher P results in greater total annual ET (Fig. 3a). This is a significant relationship in the drier watersheds (CO-ROC and CA-SIER, Table 3) where the years of highest annual P are correlated with the years of greatest annual ET. Of the three basins, CO-ROC's annual ET shows the greatest sensitivity to P , having the steepest slope (0.7). Annual P is the strongest explanatory variable of annual ET in both CO-ROC ($r^2 = 0.88$) and CA-SIER ($r^2 = 0.72$) (Table 4, Fig. 3a). For CO-ROC, annual P has a greater influence (steeper slope) in the dry years when

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P is less than 1000 mm (Fig. 3a). OR-CAS lacks a significant relationship between P and ET on an annual scale. OR-CAS is a relatively wet basin and on average receives more than twice the amount of winter (January–March) precipitation than CA-SIER or CO-ROC. High annual P in OR-CAS most years likely diminishes the sensitivity of ET to magnitude of P . However, OR-CAS only receives 21 % of its annual P during the growing season, thus it is still water-limited during this period. The growing season is assumed to extend from 1 May to 1 October all watersheds. For OR-CAS, we find a significant relationship between ET and the precipitation that falls between July–September (Fig. 3b). A positive, but not significant, relationship between summer precipitation and ET also occurs in CO-ROC where summer monsoonal pulses provide significant growing season soil moisture. CA-SIER has the smallest percentage of precipitation received during the growing season (19 % Table 2) and does not show a relationship between growing season precipitation and ET.

3.2 Timing of recharge vs. ET

For all three catchments, later R_{75} has a significant positive correlation with ET (Fig. 4). In OR-CAS and CA-SIER, R_{75} is between February and May. There is more scatter in the predictive power of R_{75} for annual ET when R_{75} is early in the water year. The earliest R_{75} are in OR-CAS, where a greater fraction of winter precipitation falls as rain. CA-SIER and CO-ROC are more sensitive to the timing of recharge than OR-CAS. Summer monsoonal pulses in CO-ROC push R_{75} to later in the water year as compared to OR-CAS or CA-SIER. The explanatory power of R_{75} for ET is greatest in CO-ROC where later days of recharge, due to later snowmelt and summer monsoons, can combine to provide soil moisture during peak growing season.

3.3 Spring temperature vs. ET

Warmer spring temperature (T_{AMJ}) in all basins generally reduces annual ET (Fig. 5a), which is somewhat counter-intuitive. Among the predicted consequences of increased

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(Table 4). In CA-SIER there is substantially more scatter in the relationship between AWC and ET. We note that in this figure, each point refers to a separate simulation with a unique set of soil parameters impacting both local AWC for each patch and vertical and lateral drainage rates. In some cases, parameter sets may give the same AWC but result in different drainage rates, affecting downslope lateral redistribution of moisture. For OR-CAS and CO-ROC these differences in soil parameters for a given AWC do not substantially alter ET. However in CA-SIER, we see greater sensitivity of ET to drainage parameters, even for the same AWC.

Lateral redistribution occurs for all three watersheds. However the topography of CA-SIER results in many swale-like features that concentrate drainage from upslope areas. We calculate the topographic wetness index (TWI) using a 30 m resolution DEM for each watershed (Moore et al., 1991) (Table 2). The TWI reflects the propensity of a location to develop saturated conditions using the assumption that topography controls water flow. Higher TWI values represent flatter, converging terrain and lower values reflect steep topography. The mean TWI for CA-SIER is greater than, and significantly different from (Welch's t test), the mean TWI for CO-ROC and OR-CAS. Particularly for CA-SIER, changing soil parameters associated with drainage rates can alter the timing of flow into areas that concentrate flow and subsequently alter their ET rates.

3.5 Sensitivity of ET to climate drivers with soil AWC

We analyze the sensitivity of ET relationships with climate drivers to soil properties by plotting the slope of linear regressions between ET and P , R_{75} , and T_{AMJ} across all soil parameters sets in Figs. 7–9, respectively. We note that AWC has been normalized in these plots to facilitate cross-site comparison. Because AWC is calculated as a function of soil storage characteristics and rooting depth (Eq. 2), soils with normalized AWC greater than one represent soils that can hold more water relative to the basin average value.

3.5.1 Sensitivity to P with AWC

Of the climate drivers explored, annual precipitation P has the greatest robustness across soil parameter sets, as suggested by number of sets that show a statistically significant relationship between annual P and annual ET (Fig. 7). As expected, slopes are positive between P and ET across all basins. Only the drier basins CO-ROC and CA-SIER have p values less than 0.001, highlighting the strength of P as a climatic driver in these drier basins, as discussed above. The response in slope sensitivity across AWC is similar in OR-CAS and CA-SIER where ET's sensitivity to P is highest at low AWC and decreases with increased soil AWC. OR-CAS has a much smaller range in sensitivities (slope varies from 0.2–0.6) compared to CA-SIER (slope varies from 0.0–0.8). Thus in CA-SIER for low values of AWC, year-to-year variation in P becomes a greater control on year-to-year variation in ET. For both OR-CAS and CA-SIER increasing AWC becomes less important at higher values of AWC. Higher scatter in slope of annual P vs. ET relationship for CA-SIER also reflects the greater sensitivity of ET to soil parameters that influence lateral drainage as discussed above (Sect. 3.1). The variation of ET response to P across AWC in CO-ROC is noteworthy for two reasons. First, CO-ROC has the narrowest range of slopes across AWC and has the highest values (0.6–0.8), which again reflects the consistency of annual P as a control on inter-annual variation in ET in this basin. Second, unlike OR-CAS and CA-SIER, increasing AWC does not substantially reduce that sensitivity to P (i.e., slope). We note that CO-ROC has a seasonal precipitation regime where a significant fraction of its annual precipitation is received later in the growing season as summer monsoonal pulses. When precipitation occurs during the growing season, the water available for ET is less likely to be limited by storage capacity. Instead ET is limited by the amount or intensity of precipitation. Water that does recharge the system is used relatively quickly, making variation in storage (or AWC) less important as a control on how much P can be used in CO-ROC.

3.5.2 Sensitivity to R_{75} with AWC

After precipitation, the timing of recharge (R_{75}) most significantly correlates with increased ET across all AWC and all basins (Fig. 8). There are several similarities in the response of ET's sensitivity to R_{75} across AWC when compared to sensitivity to P (Fig. 7). For example, the dry basins CO-ROC and CA-SIER have the highest degree of sensitivity (significant slopes > 1.0) as compared to OR-CAS (slopes < 1.0) and CA-SIER has the greatest variability in its sensitivity to AWC with slopes ranging from 1.0–3.0 across variation in soil parameters. CO-ROC once again has the least variability in the ET vs. R_{75} relationship, with consistently high (2.0–2.5) slopes unaffected by AWC.

3.5.3 Sensitivity to T_{AMJ} with AWC

Finally, T_{AMJ} has the fewest soil parameter sets with significant correlation with ET. None of the linear regressions of ET on T_{AMJ} have statistical significance less than 0.001 (Fig. 9). The slopes are always negative because earlier melt results in less ET. For all basins, the sensitivity of ET to T_{AMJ} is greatest at the lowest values of AWC, though CO-ROC once again demonstrates the least variability in slopes across the entire range of AWC (–0.2 to –0.3). At OR-CAS, T_{AMJ} is only significant for the lower AWC values. We suggest this is in part due to the small fraction of P that falls as snow. Because T_{AMJ} 's largest effect is through timing of snowmelt (Fig. 5), AWC interacts with T_{AMJ} to modulate the melt response. With relatively less snowmelt in OR-CAS, only the soils with the smallest capacities will have a significant negative interaction effect with AWC.

4 Discussion

Though forest ET in western US basins is controlled by water availability, it does not scale linearly with gross precipitation. Instead, water availability and forest ET in these

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systems vary with climatic interactions that influence the timing of water delivery to forests relative to water demand and soil characteristics that control the storage of available water. We use a mechanistic model to calculate ET and linear regression models to quantify how separate geoclimatic metrics relating to water availability in these regions – springtime temperature, total precipitation, and soil moisture capacity and recharge timing – affect forest ET under historic climate variability.

Our analysis uses an ecohydrologic model that has been validated by comparison against diverse observations in mountain environments, including streamflow and carbon flux. The model implements our current understanding of key energy, temperature and moisture controls on evaporative fluxes and their spatial-temporal patterns in complex topography of mountain environments. The model provides a simplified representation of these interactions, ignoring many additional complexities. For example, this study does not include key ecosystem characteristics such as species distributions or disturbances (i.e., fire, drought or insect outbreaks) that would influence productivity, evapotranspiration and their relationship with climate (Loudermilk et al., 2013). Our goal here is to focus on geophysical and climate effects on ET and use this analysis to highlight key sensitivities and reasons for differences in these sensitivities across the western US. Other top-down studies have adjusted Budyko curve relationships to address seasonality in P and T (Donohue et al., 2012; Feng et al., 2012), but do not account for the dynamics of snowmelt, soil storage and their interaction.

In our three western US headwater catchments, water availability, unsurprisingly, was a first order control of ET. In addition to total precipitation received, our analysis highlights the timing of water availability (R_{75}) as a key predictor of ET, showing that the strength of climatic controls influencing water availability changes with precipitation and temperature regimes. Key climatic differences in these catchments include total precipitation received, the fraction of precipitation that falls as snow, and the timing of soil moisture recharge (through snowmelt or summer precipitation pulses) relative to growing season water demands. Total annual P is the first order control of ET in the two drier watersheds, CO-ROC and CA-SIER. OR-CAS, alternatively, does not have

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variation in ET associated with across (and possibly within) site AWC is on the same order of magnitude as inter-annual variation in ET with P . For example, ET in CO-ROC ranges from 380–600 mm across annual P variation, and across all calibrated soil parameters long-term average ET ranges from 450–600 mm. Similarly, in CA-SIER ET ranges from 400–800 mm across the P record and across all soil parameters its long-term ranges from 700–1000 mm. The nonlinear relationship between ET and AWC suggests that below a threshold point in each basin (195–265 mm of AWC), long-term average ET is more sensitive to AWC and above these threshold values the effect of climate on ET is greater than an increase in soil storage.

An implicit control of ET in our soil representation is subsurface redistribution potential. Only a few studies have emphasized the role of lateral redistribution (Barnard et al., 2010; Tague and Peng, 2013). We argue, however, that modeling subsurface moisture redistribution makes larger soil AWC more effective in increasing ET because (1) captured recharge from snowmelt will be available to lower elevation forests and (2) high elevation areas that redistribute soil moisture are depleted and able to capture recharge that would have been lost (i.e., through overland flow).

We show that soil can also substantially change the sensitivity of ET to year-to-year variability in climate drivers. CO-ROC has a high sensitivity to climate drivers across all values of AWC. We suggest that a strong summer P signal in CO-ROC explains the negligible change in ET's sensitivity to climate drivers across values of AWC, similar to other studies that show that summer P can offset the dependence of ET on soil replenishment or winter snowpack (Hamlet et al., 2007; Litaor et al., 2008). However, the two basins receiving the smallest fraction of annual P in the summer (CA-SIER and OR-CAS) have high sensitivities in ET's response to P , R_{75} , and T_{AMJ} at lower values of soil water content (AWC) but this sensitivity substantially decreases for higher AWC. The slope of the relationship between P and ET decreases by more than 50% for both CA-SIER and OR-CAS at higher AWC. These results have important implications for understanding and ultimately predicting the spatial variability of ET within a basin. There is often little to no information about a basin's soil beyond coarse description of

5 general characteristics; we have used that imperfect knowledge of soil characteristics to calculate a range of physically feasible values of AWC for each basin. Because there is substantial heterogeneity in soil characteristics within each basin (Dahlgren et al., 1997; Denning et al., 1991; McGuire et al., 2007) we might expect that the full
10 range of AWCs can be observed when we look at soils across individual forest stands within a basin. The effect of AWC is also as large as cross-basin differences in ET sensitivity. By normalizing slopes between climate drivers and ET across the range of AWC, we show that a similar sensitivity to P and T_{AMJ} can be achieved for all basins by varying soil AWC. For example, OR-CAS's stands with the smallest AWC are similarly
15 sensitive (slope of 0.6) to inter-annual variation precipitation as stands in CO-ROC (Fig. 7). This indicates potential for a high degree of ET heterogeneity within the basin due to soil properties, in addition to other landscape (elevation, aspect) and climatic drivers. It furthermore suggests that forests without a strong summer P signal and with low AWC are more sensitive to climate drivers but the soils with higher AWC may buffer
20 forest response to climate. In our example watersheds, OR-CAS and CA-SIER show the greatest sensitivity to climate perturbations in soils with lowest AWC.

As western US forests show increases in disturbance and mortality (Allen et al., 2010; Hicke et al., 2012), it becomes more important to understand these ecosystems' responses to primary climate drivers. The responses vary geographically with regional
25 climate patterns, and our work also suggests that spatial heterogeneity in soils vary the strength of these drivers within and across watersheds as well. This understanding of how climate and soils combine to control ET can enhance our understanding of forest water stress related to increased mortality (van Mantgem et al., 2009). These findings may also extend to understanding of forest disturbances, providing mechanistic explanations for previous work showing that the area burned by wildfire can be correlated to ecosystems and their climates (Littell et al., 2009). Our process based modeling results call out specific climate measures that interact to reduce forest ET. These results emphasize the importance of interactions between several physical drivers and

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accentuate the need for introducing more field data to examine forest response and feedback in the face of climate change (Tague, 2009).

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Table 1. Explanatory variables.

Abbreviation	Definition
P	Total annual precipitation
T_{AMJ}	Average daily temperature for April, May, June
R_{75}	Day of water year that 75 % of soil water recharge occurs
AWC	Available water capacity of soil (field capacity-wilting point)

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Table 2. Basin topography, geology, vegetation and climate characteristics. Climate descriptions are averaged over total available climate record (duration noted in table).

Watershed	CO-ROC	OR-CAS	CA-SIER
Location	Colorado	Oregon	California
US Geological Survey gage number	06733000	14161500	10343500
Geology	Holocene glacial till, rock; Precambrian gneiss, granite	Western Cascade basalt	Sierra granite, with Miocene andesite cap
Elevation range (m)	1470–4345	410–1630	1800–2650
Drainage Area (km ²)	350	64	26
Topographic Wetness Index- Mean (Std Dev)	7.0 (1.9)	6.6 (1.7)	7.9 (1.8)
Climate record	1980–2008	1958–2008	1960–2000
Mean Annual Precipitation (mm)	1000	2250	850
Annual Precipitation as snow (%)	64	29	55
Precipitation received in Growing Season (%)	46	21	19
Min/Max winter <i>T</i> (JFM) (oC)	−12.1/ −0.02	−0.9/5.2	−9.5/3.7
Min/Max spring <i>T</i> (AMJ) (oC)	−2.7/10.9	4.0/14.0	−2.5/13.8
Vegetation	Subalpine fir, aspen, meadows, shrub	Douglas-fir, Western Hemlock	Mixed Conifer, Jeffrey and Lodgepole Pine
Mean basin LAI	3.5	9.0	4.1
Annual NPP range for calibration (gCm ^{−2} yr ^{−1})	280–520	620–1100	450–800
Literature sources used to bound annual NPP range	Arthur and Fahey (1992), Bradford et al. (2008)	Grier and Logan (1977) Gholz (1982)	Hudiburg et al. (2009), Goulden et al. (2012)*

* Values reported as gross primary productivity, converted to NPP using RHESSys calculated values of respiration.

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Table 3. Calibrated average basin soil parameters used for linear regression analyses between annual ET and P , R_{75} , and T_{AMJ} .

Watershed	m (m^{-1})	K (md^{-1})	b (–)	φ_{ae} ($m H_2O$)	gw_1 (–)	gw_2 (md^{-1})	Z_r (m)	AWC (mm)
CO-ROC	0.05	9500	0.2	0.05	0.2	0.7	1.4	400
OR-CAS	0.1	300	0.1	0.01	0.15	0.3	1.0	230
CA-SIER	0.5	2300	0.2	0.007	0.3	0.3	1.7	270

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Table 4. Statistics for ET predictors based on linear regression models.

Watershed		CO-ROC	OR-CAS	CA-SIER
Precipitation (P)	p value	< 0.001	> 0.05	< 0.001
	r^2	0.88	0.24	0.72
	slope	0.7	0.2	0.5
Timing (R_{75})	p value	0.001	> 0.05	0.001
	r^2	0.43	0.21	0.36
	slope	2.2	0.3	1.8
Temperature T_{AMJ}	p value	< 0.01	0.05	0.01
	r^2	0.42	0.12	0.20
	slope	−0.2	−0.3	−0.3
Soil capacity (AWC)	p value	0.001	0.001	0.001
	r^2	0.43	0.53	0.11
	slope	0.1	0.2	0.1

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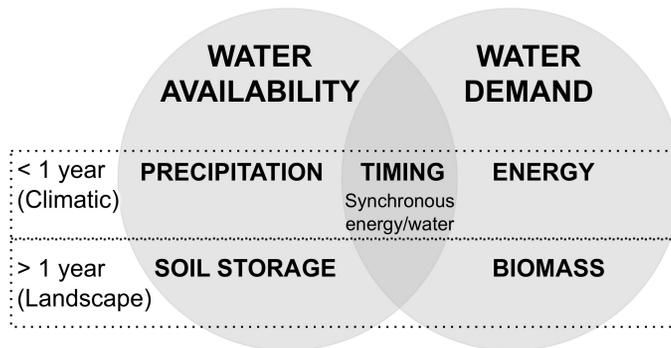


Fig. 1. Conceptual model of evapotranspiration (ET) groups its controls into categories of water availability and demand. These controls influence ET sub-annually and at multi-annual timescales. The duration of overlap between climatic demand and availability is an important control of ET in Western US catchments.

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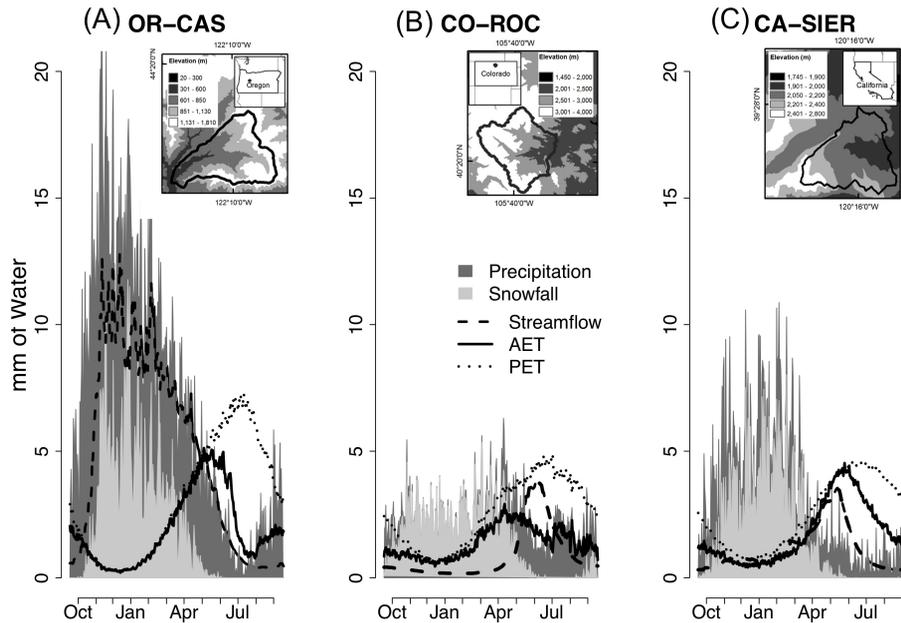


Fig. 2. Locations and average daily water fluxes averaged over 1980–2000 for three study watersheds.

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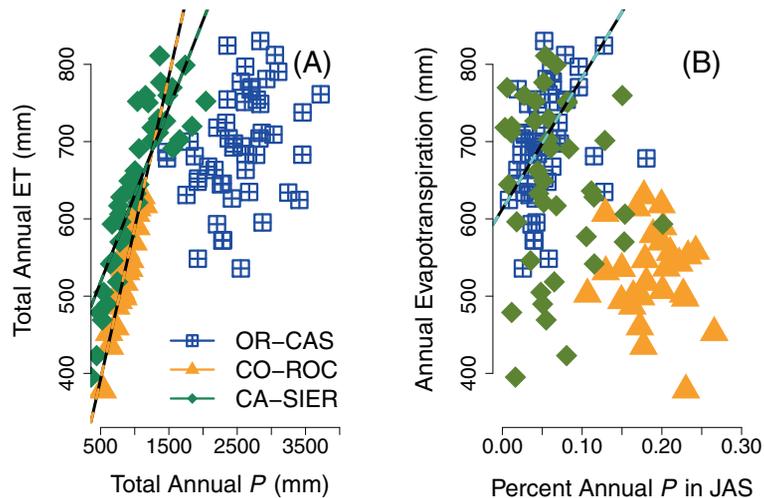
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Fig. 3. (A) Total annual ET increases with total annual precipitation. (B) Though annual P is not significantly related to annual ET in OR-CAS, a significant relationship exists between summer (JAS) precipitation and annual ET.

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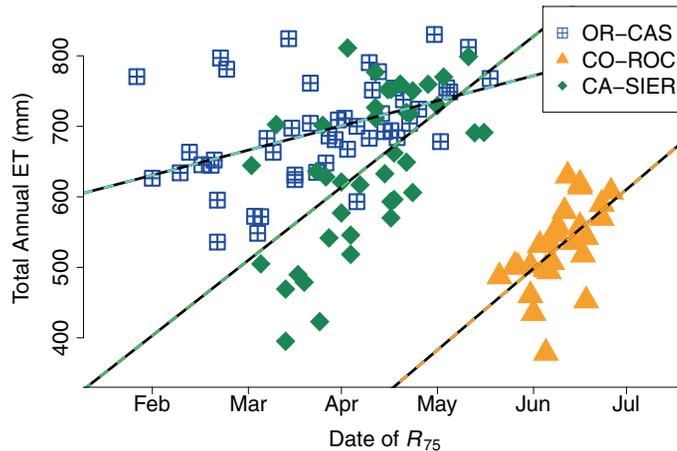
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Fig. 4. Later occurrence of soil moisture recharge (R_{75}) is significantly correlated with increased annual ET in all study watersheds.

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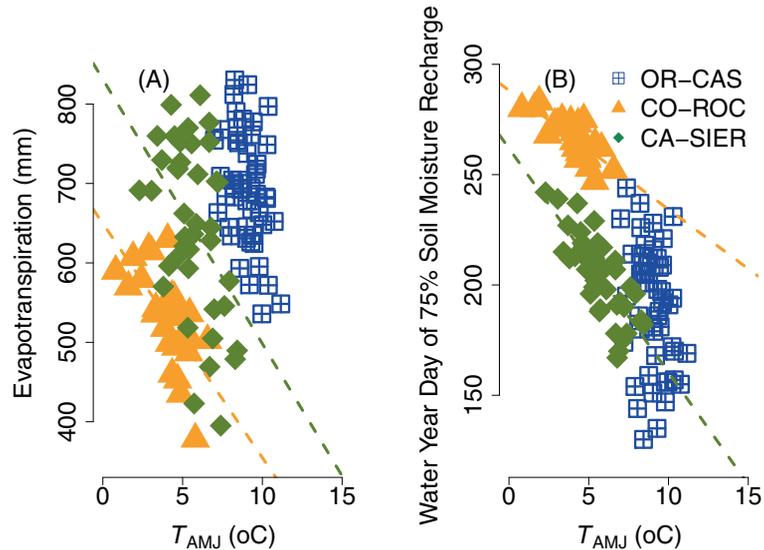


Fig. 5. (A) Warmer spring temperatures are correlated with lower total annual ET and (B) earlier occurrence of soil moisture recharge in the two snow-dominated watersheds.

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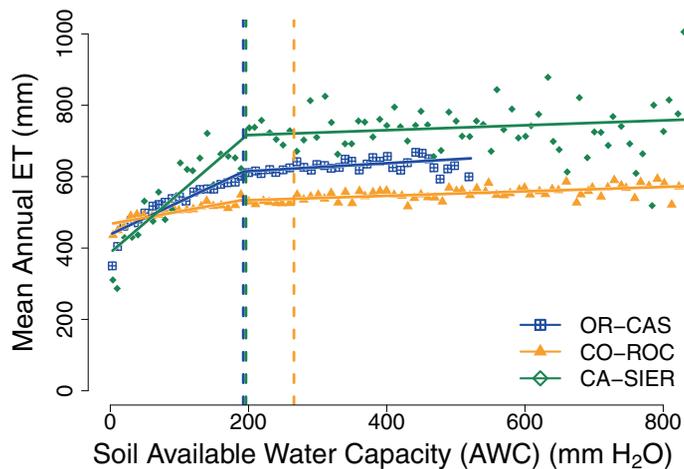
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Fig. 6. Each point represents mean annual ET from WY 1985–2000 for a physically feasible mean basin soil available water capacity (AWC). Vertical lines represent the calculated break-point in the nonlinear relationship between long-term ET and AWC for each basin.

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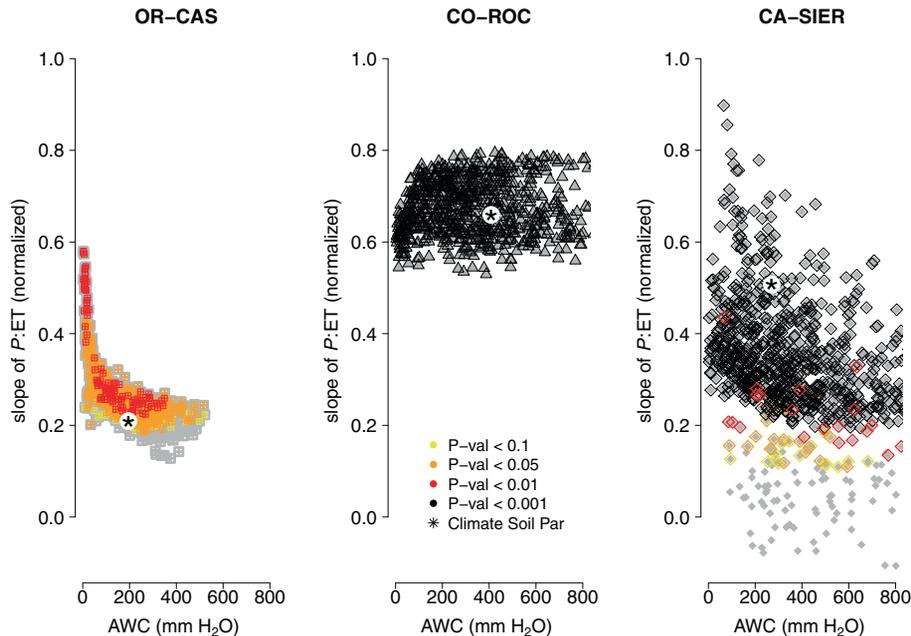


Fig. 7. Slopes of the linear regression model calculated for P against ET across all watersheds' soil AWCs. Note that the slopes are normalized to facilitate inter-basin comparison. “Climate Soil Par” represents the soil parameter set used for model results presented in Figs. 3–5 and Sects. 3.1–3.3.

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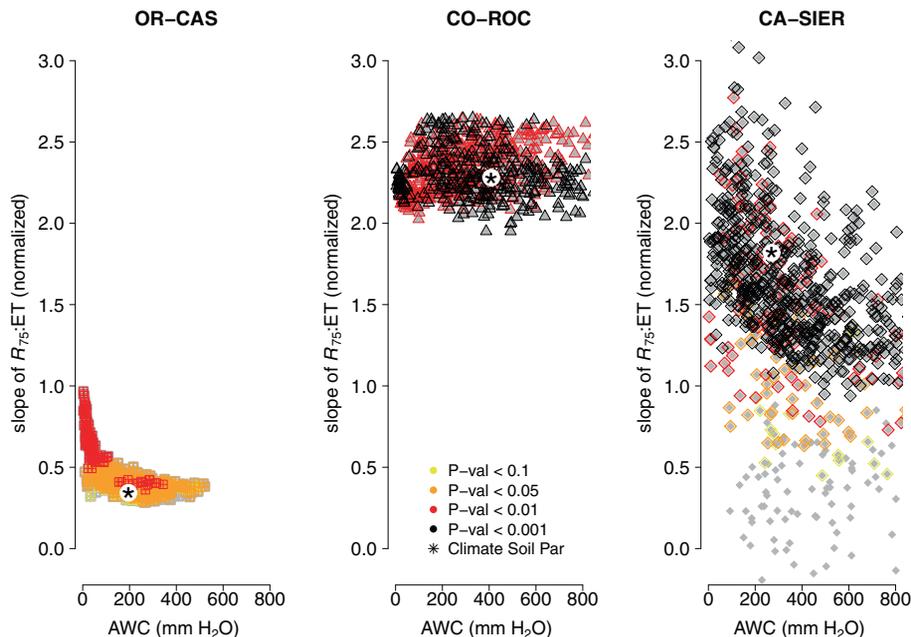


Fig. 8. Slopes of the linear regression model calculated for R_{75} against ET across all watersheds' soil AWCs. Note that the slopes are normalized to facilitate inter-basin comparison. "Climate Soil Par" represents the soil parameter set used for model results presented in Figs. 3–5 and Sects. 3.1–3.3.

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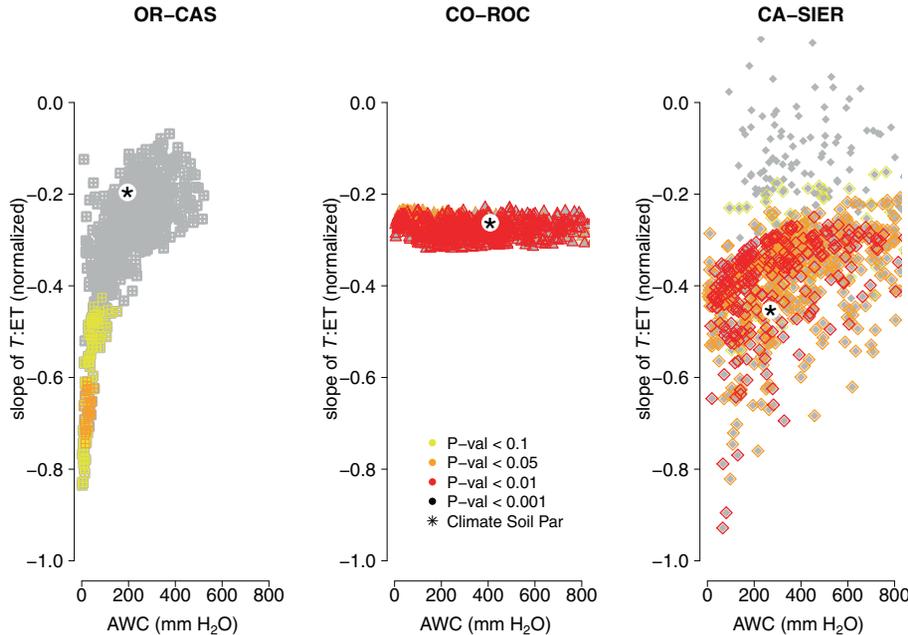


Fig. 9. Slopes of the linear regression model calculated for T_{AMJ} against ET across all watersheds' soil AWCs. Note that the slopes are normalized to facilitate inter-basin comparison. "Climate Soil Par" represents the soil parameter set used for model results presented in Figs. 3–5 and Sects. 3.1–3.3.

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