How evaporative water losses vary between wet and dry water years as a function of
elevation in the Sierra Nevada, California and critical factors for modeling

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Abstract

High altitude basins in the Sierra Nevada, California have negligible summer precipitation and very little groundwater storage, making them ideal laboratories for indirectly monitoring changes in evaporative losses between wet and dry years. Dry years typically have greater potential evapotranspiration (ET), due to warmer June and July air temperatures, warmer summer water/soil temperatures, greater solar radiation exposure due to less frequent cloud cover, greater vapor pressure deficit, and longer growing seasons. However, dry years also have limited moisture availability compared to wetter years, and thus actual evapotranspiration is much less than the potential in dry years. The balance of these factors varies with elevation. Here, we use gridded temperature, precipitation, and snow data, along with historic streamflow records in two nested basins of the Merced River, California and a simple model to determine the following: Annual ET increases in wetter years at mid-elevations (2100-2600 m), but this pattern can only be represented in model simulations that include some representation of water transfer between higher and lower elevation soil reservoirs. At higher elevations (>2600 m), greater water availability in wet years is offset by shorter growing seasons due to longer snow cover duration. These results suggest that models seeking to represent changes in ET in mountainous terrain must, at a minimum, include both hillslope processes (water transfer down steep slopes) and snow processes (timing of water and energy supply).
1. Introduction

Research has shown that temperatures in the western U.S. have been warming [Bonfils et al. 2008; Dettinger and Cayan 1995] and that snowmelt timing has shifted earlier in the season [Stewart et al. 2005]. Both trends are likely to continue [Stewart et al. 2004]. These changes beg the question if, and how, evapotranspiration (ET), which makes up approximately 30 to 50% of the annual water budget in the Sierra [e.g., Jeton et al. 2004], is likely to change under these circumstances. For example, recent modeling studies for projected climate impacts in the Colorado River basin show a range of climate impacts on Colorado River runoff ranging from 5% reduction in flows by 2050 [Christensen and Lettenmaier 2007] to 20% reduction in flows by 2050 [Milly et al. 2005], with one of the major differences being due to how ET and its relation to snow cover is handled in the models [Hoerling et al. 2009]. ET in cold regions, specifically how ET relates to snow and ice processes, was highlighted by Shuttleworth [2007] as an area with substantial need for more research and understanding.

Overall, changes in ET depend critically on the balance between energy and moisture available for evaporation [Monteith and Unsworth 1973; Brutsaert 1982; Jeton et al. 1996]. Warmer temperatures increase potential ET, and an earlier disappearance of the snowpack reduces albedo, warms soil and air temperatures, and increases potential ET. However, as soils dry out earlier [e.g., Hamlet et al. 2007], plants close their stomata and reduce transpiration [e.g., Goldstein et al. 2000], resulting in actual ET that is much less than potential.

The mountains of the Western United States, and particularly the Sierra Nevada of California, provide an interesting test case. Here, the majority of precipitation falls in the winter, in the form of snow, and plants depend on stored moisture from snowmelt during the summer growing season. A larger snowpack serves both to potentially decrease ET (through a longer...
period of snow cover, low temperatures, high albedo, and plant dormancy) and to potentially increase ET (through a greater summer water supply).

It is still uncertain whether ET in this region is likely to increase, decrease, or remain the same with warmer temperatures and earlier snowmelt. Fortunately, interannual climate variations in the western U.S. span a range of precipitation larger than any projections of future mean change [Cayan et al. 2008]. Risbey and Entekhabi [1996] examined data from the Sacramento Basin, CA and found that annual streamflow responded to changes in annual precipitation and not to changes in annual temperature. Thus, any climate impact on ET in this region is likely to propagate from the temperature impact on snowpack. Therefore, we can examine how ET varied in wet (late melt) vs. dry (early melt) water years in the past to gain insight regarding future change and/or variability.

Tague et al. [2009] and Christensen et al. [2008] both looked at the differences in ET with elevation in the Merced River above Happy Isles as represented by the RHESSys model [Tague and Band 2004] under current and future climates. They found that transpiration at middle elevations (1800-2150 m) was primarily water-limited, with the most transpiration in wet years, whereas the highest elevations (2600-4000 m) were more energy-limited and responded to temperature. Thus, ET increased under future warming scenarios for only the higher elevations, and decreased for the lower elevations, resulting in negligible changes over the watershed as a whole.

Here, we use gridded temperature, precipitation, and snow data, along with historic streamflow records in two nested basins of the Merced River, California and a simple model to address two key questions: 1) How does ET vary across elevations and years in the Sierra
Nevada? and 2) What key processes are essential to include in any model that strives to represent variability in ET across elevations and years in snowy mountain terrain?

2. Observational Data and Methods

2.1 Study Area

We focus our study on the Merced River Basin within Yosemite National Park, California, which has been gauged at two locations, Happy Isles (USGS gage 11264500) and Pohono (USGS gage 11266500) since 1916 (Figure 1a). Elevations range from about 1050 m at the Pohono gage to nearly 4000 m at the top of Mt. Lyell. The region receives most of its precipitation from October through May in the form of snow. The basin is underlain by granodiorite, and soils are shallow [Huber 1987]. Because of its long streamflow record and unimpaired flows, the Merced River has been the subject of many prior modeling studies and analyses [e.g., Clow et al. 1996; Dettinger et al. 2004; Christensen et al. 2008].

The nesting of the Happy Isles basin within the Pohono basin provides a unique opportunity to parse out different processes occurring at high vs. low elevations. As seen in Figure 1b, all of the contributions to the Pohono gage from elevations above 3200 m come from the Happy Isles sub-basin. If we consider the contributing areas above and below the Happy Isles gage separately, we see that the lower basin has the middle 50% of its contributing area between 2170 and 2580 m, and the upper basin has the middle 50% of its area between 2420 and 3050 m elevation (Fig. 1b). The area-weighted mean of the lower basin is 2320 m and of the upper basin is 2750 m. The different elevations of these two basins control where water at the Pohono gage originates from at different times of year (Fig. 2). Winter streamflow peaks originate primarily from the lower basin, where a greater fraction of precipitation falls as rain...
than as snow. Late summer streamflow originates almost entirely from the upper basin, the only
region with remaining late-lying snow (Fig. 2). Note that the entire system has limited deep
groundwater flow, so that streamflow is closely linked to rain and snowmelt processes.

2.2 Data

Annual runoff was calculated as the sum of daily USGS streamflow records for Happy
Isles and Pohono, from 1 October to 30 September each water year from 1981 to 2007. Both of
these gages have a quality classification of “good” (<10% error) [Rockwell et al. 1997]. To
emphasize differences between elevations, we subtracted the streamflow measured at Happy
Isles from the streamflow measured at Pohono to estimate the contribution from the lower
elevations (Fig. 1). From here on, we will refer to the “upper basin” as the areas contributing to
the Merced River gauge at Happy Isles, and the “lower basin” as the areas contributing to the
Merced River gauge at Pohono with the Happy Isles contribution removed.

Precipitation data were obtained from the Parameter-elevation Regressions on
Independent Slopes Model’s (PRISM, available at http://www.prism.oregonstate.edu) monthly
precipitation maps for water years 1981 to 2007 at 4-km resolution [Daly et al., 1994, 2008].
PRISM uses empirical relationships to account for the influence of elevation, rain shadows, and
costal proximity when interpolating between existing measurement stations. The data from
PRISM have been quality-controlled and peer-reviewed, and are widely used in modeling
applications [Daly et al., 2002]. This time period (post-1980) was chosen because high altitude
snow pillows, which determine snow water equivalent by weighing the snowpack, were first
installed in the region around 1980, providing guidance on upper elevation precipitation
distributions. For water-year analysis, all of the PRISM monthly maps from October to the
following September were summed. The 4-km maps were then sub-sampled at a 100-m resolution, and all grid cells falling within a basin’s boundaries were summed to represent the total annual precipitation that fell within that watershed. Precipitation in the upper and lower Merced basins was calculated separately. Snow-covered area (SCA) and its rate of change with time were calculated from daily 500-m gridded MODIS maps of fractional SCA for water years 2002 to 2007, which were created using the MODSCAG algorithm [Dozier et al. 2008; Dozier and Frew 2009; Painter et al. 2009]. Similar to treatment of the PRISM gridded data, these daily grids were sub-sampled to 100 m and clipped to represent SCA in the upper and lower Merced basins, respectively. Within each of the two watersheds, fractional SCA was then calculated separately for each 200-m elevation band, ranging from 1200 m to 4000 m. Meteorological data for the 2002-2007 period were obtained from the Dana Meadows Snow Pillow Site (at 3000 m, Fig. 1) just to the north of the two watersheds. Temperature, relative humidity, wind speed, and incoming solar radiation were measured hourly and converted to daily averages and daily maximum and minimum temperatures. Daily precipitation data was measured at the Tuolumne Meadows Snow Pillow Site (2600 m, Fig. 1).

**2.3 Methods: Water balance approach to estimating ET**

Over longer time-periods, such as the annual average, basin storage in steep basins with shallow soils can be considered negligible, such that

\[ P = R + ET \]  

(1)

where \( P \) is total annual precipitation, \( R \) is total annual runoff, and \( ET \) is total water lost to evapotranspiration and sublimation [Adam et al. 2006]. Because precipitation and runoff are
both easier to measure than ET, ET is typically estimated as the residual of equation (1). The catchment water balance method of calculating ET agrees within 10% of upscaled direct eddy-covariance flux measurements of ET [Miller et al. 2007, at AmeriFlux sites; Wilson et al. 2001, at a 97.5 ha watershed in Tennessee], although errors are potentially larger in more extensive areas of complex terrain. To compare interannual variations in ET at different elevations, we constructed annual water balance estimates for the upper and lower basins of the Merced River (Fig. 1) for 1981 to 2007, using PRISM maps of precipitation and USGS streamflow records, as outlined in Section 2.2. Assuming no net changes in water storage between one year and the next, annual ET was calculated as the difference between precipitation and runoff.

3 Observational Results

Fig. 3 illustrates 27 water years (October-September) of area-normalized annual basin precipitation (from PRISM), discharge (from USGS gages), and ET (from the difference between the prior two) for (a) the upper basin (above Happy Isles) and (b) the lower basin (between Happy Isles and Pohono). Area-normalized precipitation varies from 0.6 to 2 meters, and discharge varies from 0.3 to 1.5 m. ET varies much less than either precipitation or discharge, but ranges from 0.3 to 0.5 m in the Upper Basin, and from 0.4 to 0.8 m in the Lower Basin.

To better illustrate the dependence of ET on wet vs. dry years, Fig. 4 plots the annual ET shown in Fig. 3 as a function of basin-average annual precipitation. In the Lower Basin, ET increases in wetter years, with a significant correlation (p<0.01) and a best-fit slope of 13%. This suggests that the 2170 - 2580 m elevation range that dominates this Lower Basin is water-limited, with ET responding to increases in precipitation to a greater extent than to changes in temperature. These observations agree with other Sierra studies of ET in this elevation range.
[e.g., Risbey and Entekhabi 1996; Jeton et al 1996; Miller et al. 2007; Christensen et al. 2008].

ET in the Upper Basin, which has the middle 50% of its area between 2420 and 3050 m elevation, has no significant trend, although ET appears to decrease in both the wettest and driest years on record, suggesting that it is sometimes energy-limited and sometimes water-limited. To better understand what controls these patterns and what parameters are essential to model them, we create a simple model of basin ET, detailed in Section 4.

4 Basin ET Model

4.1 Model Set-up

Model representations of ET and water availability vary widely, ranging from relatively simple physical parameterizations in global climate models [Cornwell and Harvey 2007], to extensively parameterized stomatal responses in vegetation canopy models [e.g., Running and Coughlan 1988; Tague and Band 2004].

We start with the commonly-used basis for ET modeling, the Penman-Monteith equation [e.g., Shuttleworth, 1993; Allen et al., 1998]:

\[
ET = \frac{\Delta (R_{net} - G) + \rho_{air} c_p (e_s - e) / r_s}{\lambda \rho_{water} (\Delta + \gamma (1 + r_s / r_a))}
\]

where \(\Delta\) is the gradient of the saturated vapor pressure with temperature, \(de/dT\) (kPa °C⁻¹), \(R_{net}\) is the net radiation (W m⁻²), \(G\) is the ground heat flux (W m⁻²), \(\rho_{air}\) is the density of air (kg m⁻³), \(c_p\) is the heat capacity of air (J kg⁻¹ °C⁻¹), \(e_s\) is saturated vapor pressure (kPa), \(e\) is vapor pressure (kPa), \(\gamma\) is the psychometric constant (kPa °C⁻¹), \(\lambda\) is the latent heat of vaporization of water (J kg⁻¹), \(\rho_{water}\) is the density of water (kg m⁻³), \(r_s\) is the surface resistance (m s⁻¹), \(r_a\) is the aerodynamic resistance (m s⁻¹), and \(P\) is atmospheric pressure (kPa).
We used the approximations of $r_s$ and $r_a$ laid out by the FAO (Food and Agriculture Organization of the United Nations) to adjust the Penman-Monteith equation to reference (potential) evapotranspiration [Allen et al. 1998]. We divided the basin into 200-m elevation bands, which were weighted by area, according to basin hypsometry. Daily maximum and minimum temperatures were distributed to the center elevation of each zone assuming a 6.5°C km$^{-1}$ lapse rate, the average lapse rate for the region found by Lundquist and Cayan [2007]. Wind speed and solar radiation were assumed constant with elevation. Atmospheric pressure was a function of elevation [Allen et al. 1998, their equation 7] but was fixed in time. Measured relative humidity at the Dana meteorological station (Fig. 1) was used to determine the mixing ratio (g kg$^{-1}$ of water in the air), which was assumed constant with elevation. Albedo was set to 0.15, a typical value for open forests [Jarvis et al. 1976; Brandes and Wilcox 2000]. We neglected ground heat flux, which is much smaller than net radiation and can generally be neglected over daily time scales [Allen et al. 1998]. This is especially true in the Sierra Nevada, where soils do not freeze but rather, remain at or near 0°C throughout the snow season [Lundquist and Lott 2008].

Actual ET was assumed to deviate from potential ET due to three main causes. First, we assumed that ET did not occur at elevations above tree line (>3000 m) when fractional snow covered area (SCA) exceeded a set threshold, where snow cover was represented by direct MODIS observations of fractional SCA (see Section 2.2). Second, we assumed that no ET occurred at any elevation on days when minimum temperatures at that elevation fell below a set threshold, indicating a hard freeze, based on studies showing that low soil and air temperatures inhibit plant metabolic activity [e.g., Troeng and Linder 1982; Tanja et al. 2003; Mellander et al.].
The formulation of actual ET follows a “bucket” approach, similar to that used by Manabe [1969], Koster and Milly [1997], Rodriguez-Iturbe et al. [1999] and Gedney et al. [2000], among many others. Actual ET for each elevation band was derived by scaling potential ET as a function of soil moisture:

\[ ET_{actual} = f(\theta)ET_{potential} \]  

[following Feddes et al. 1978; Laio et al. 2001; D’Odorico and Porporato 2006], where \( \theta \) is the volumetric soil moisture. \( f(\theta) \) is set to 1 above a critical soil moisture (\( \theta_c \), which depends on both soil and vegetation properties), to 0 below the wilting point (\( \theta_{wp} \)), and varies linearly in between these values; this functional form was found to work best for forests by Brandes and Wilcox [2000]. The model set-up for determining soil moisture is illustrated in Figure 5. Soil water in each elevation band is represented by a simple bucket with root zone depth (\( Z_r \)), set porosity (\( n \)), rate of soil drainage (\( K \)), and field capacity (\( \theta_{fc} \)). Water is added to the bucket through liquid precipitation (measured at the Tuolumne Meadows snow pillow, Fig. 1, and determined to be liquid if \( T_{mean} > 1^\circ C \) and through snowmelt, which is calculated as the product of mean daily air temperature times a melt factor, which follows the formulation in Snow-17 [Anderson 1973], using calibration parameters determined by Shamir and Georgakakos [2005], on all days when \( T_{mean} > 0^\circ C \) and when snow covered area (SCA) greater than 10\% is observed at that elevation by MODIS. Soil water in excess of porosity is assumed to run off via overland flow and be lost to the soil water system. Soil water less than porosity but in excess of field capacity drains from soil storage by the following equation from Rodriguez-Iturbe et al. [1999]:
246 \[ \text{transfer}_{\text{out}} = \frac{K(\theta - \theta_{fc})}{(n - \theta_{fc})} \]  

(4)

In our model, we use this to not only represent vertical drainage from the soil column, but also lateral drainage through interflow and hillslope processes. Thus \( K \), as used here, describes rates of both vertical water movement and downslope lateral transfer, and thus represents a rate of drainage rather than a true hydraulic conductivity. Water ceases draining as soon as soil moisture reaches field capacity. Some portion of this water transferred down slope will enter streams and be removed from the watershed, but some portion will be transferred from the soil zone in one elevation band to that in the next lower elevation band. The percentage of water that is transferred out and that reaches the soil in the next lower elevation band is set by a transfer coefficient \( (tc) \). The remaining percentage \( (1 - tc) \) is presumed to join a stream or river and be exported from the system, unavailable for ET. Thus, soil moisture in one elevation band is defined as follows:

\[ \Delta \theta = \frac{P + \text{Melt} + \text{transfer}_{\text{in}} - ET - \text{transfer}_{\text{out}}}{Z_r} \]  

(5)

where the transfer in is equal to the transfer coefficient times the leak out at the next higher elevation band. Note that the total efficiency of water transferred for the basin as a whole is a function of the number of elevation bands, and thus \( tc \) would change as the resolution of elevation bands changes.

4.2 Design of Modeling Experiments

To better understand parameter sensitivity and importance, we ran the model for 5 water years from 2003-2007. We chose this period because a) this period has complete, quality-controlled meteorological data at high elevation to calculate potential ET; b) this period has
quality-controlled daily MODIS information on SCA; and c) this period contains 2 years in the highest quartile of the 91 years of Merced river streamflow (2005 and 2006) and 2 years in the lowest quartile (2004 and 2007), which aid in comparing model sensitivity to wet vs. dry years.

We examined which model parameters and assumptions could adequately represent the mean and inter-annual variations in soil moisture, as a function of elevation, in the Sierra Nevada by assessing our model’s ability to fit the following values as determined from the water balance calculations: 1) mean annual ET (2003-2007) for the upper basin (36 cm); 2) mean annual ET (2003-2007) for the lower basin (54 cm); 3) mean annual difference between dry years (2004 and 2007) and wet years (2005 and 2006) for the upper basin (3 cm) and 4) mean annual difference between dry years (2004 and 2007) and wet years (2005 and 2006) for the lower basin (-12 cm).

Individual values for each component of the water balance for these years are presented in Figure 3 and in Table 2.

For these runs, we held the following parameters constant: The soil moisture where ET begins to be limited was set at 0.25, except in cases where field capacity ($\theta_{fc}$) was less than 0.25, when it was set to equal field capacity. The wilting point was set at 0.10. Porosity was set at 0.4 based on basin soil maps [USDA NRCS 2006]. The temperature threshold at which ET stopped was $T_{min} = -1^\circ$C. The fraction of SCA above which ET could not occur above treeline was set to 0%. Sensitivity to each of the above parameters was investigated, but was found to be less than the sensitivity to the varied parameters discussed below.

A sensitivity analysis was performed on the remaining parameters by modeling a range of reasonable values in all reasonable combinations (determined based on field measurements and basin soil maps, USDA NRCS 2006) to investigate which best fit the water balance observations. The rate of soil drainage, $K$, varied by orders of magnitude from 0.001 to 1 m day$^{-1}$. Field
capacity varied from 0.15 to 0.4. The transfer efficiency of water from soil in one elevation band to the next varied from 0 to 75%, and the rooting depth, which is the same as the soil depth in these shallow Sierra soils, varied from 0.25 to 1.5 m.

4.3 Results

The model accurately simulated mean annual ET for both the upper basin and the lower basin with a wide variety of parameter sets (lower basin shown in Fig. 6; upper basin patterns qualitatively similar, not shown). When soil drainage is slow ($K \leq 0.01 \text{ m day}^{-1}$), mean ET in the upper basin is insensitive to field capacity (see horizontal lines in top rows of Fig. 6) because soil moisture stays above field capacity for much of the growing season. Rather, annual ET is most sensitive to the rooting depth ($Z_r$), because this represents the total water available. As the rate of soil drainage ($K$) increases, other parameters become more important. When drainage is fast (e.g., bottom row in Fig 6, $K = 1 \text{ m day}^{-1}$), soil moisture quickly drops to field capacity, so $\theta_c$ and $Z_r$ both influence the mean annual ET. With no water transfer ($t_c=0$), ET increases with both increasing $\theta_c$ and increasing $Z_r$.

As the transfer of water from upper to lower elevations increases, the relationship to $\theta_c$ becomes more complex, with a local maximum ET near $\theta_c=0.35$, with lower ET for both higher and lower field capacities. This occurs because of the interplay between field capacity, porosity, and water transfer in the model, which jointly control drainage and downslope water transfer. Porosity ($n$) was held fixed at $n=0.4$ for all model simulations. Soil water in excess of porosity is assumed to run off the surface to rivers and thus is unavailable for ET at lower elevations. Thus, only soil water content both greater than $\theta_c$ and less than $n$ is available for subsurface transfer to lower elevation soils. As $\theta_c$ increases, more water is held locally, which acts to increase ET.
However, as $\theta_c$ approaches 0.4 (the soil porosity), less water is available for transfer to lower elevations, up to the limit of $\theta_c = 0.4$, where no water is transferred, regardless of what the transfer coefficient is. (Notice that the ET values at $\theta_c = 0.4$ are identical for all values of $t_c$ (Fig. 6).) Because transfer of water from higher to lower elevations acts to increase annual ET, this decrease in water transfer offsets the local increase in water availability for field capacities greater than 0.35.

For soil rooting depths $\leq 1$ m, only model simulations with 50-75% transfer efficiency and relatively fast water movement through the soil calculated annual average ET as high as the observed value (54 cm). Based on soil surveys [USDA NRCS 2006], only isolated meadows and wetlands in the region have soils deeper than 1 m, and most basin areas (forested slopes) have soils shallower than 1 m. These results imply that transfer of water from upslope (higher elevations) to downslope (lower elevations) is essential to provide enough water for the observed ET values in the lower basin.

Requiring the model to match the observed differences between wet and dry years further constrains the acceptable parameter sets. In both basins (Fig. 7 and Fig. 8), when drainage was slow ($K \leq 0.01$ m day$^{-1}$, top two rows), there was little difference in annual ET between dry and wet years. This corresponded to an over-prediction of 12 cm for the wet-dry ET difference in the lower basin and an under-prediction of 2.5 cm for the wet-dry ET difference in the upper basin. With slow drainage, the water available each year is set primarily by the rooting depth and does not vary between years.

A wide range of parameter sets led to simulations with ET that was relatively constant between wet and dry years. However, the model only simulated greater ET in wet years than in dry years, as observed in the lower basin, when it was parameterized with a high transfer...
coefficient, high $K$, and mid-range $\theta_c$ (Fig. 7). With these parameters, the model transferred more water to lower elevations to support ET in wetter years (when snow lasts much longer at higher elevations) than drier years. As discussed with the mean annual ET, larger values of $\theta_c$ that approach porosity result in less overall transfer between elevations, offsetting this effect. Note that the parameters that best fit the observed difference in the lower basin do not best fit the observed difference in the upper basin (Fig. 8). This difference between parameterization of the upper and lower basins is not an unexpected result given large differences in vegetation (Table 3, also see Lutz et al. 2010), slopes, and soils across this elevation gradient.

Figure 9 illustrates the processes leading to the model results summarized in Fig’s 6 to 8 for parameters that fit the lower basin well ($tc = 0.75; K = 1 \text{ m day}^{-1}; Z_r = 0.75 \text{ m}; \theta_c = 0.3$). For a fast draining soil ($K = 1 \text{ m day}^{-1}$), soil moisture stays close to field capacity so long as local snowmelt, or snowmelt transferred in from higher elevations, sustains it (Fig. 9 a, b, c, d, e, f). The rate of snowmelt (so long as snow is available to melt), exceeds the rate of ET by almost a factor of 10 (Fig. 9 c, d, g, h), so Sierra soils are saturated under snowmelt conditions (see Flint et al. 2008 for an example of this in the Merced Basin). In this situation, the primary factors limiting ET are the date ET begins (limited by SCA at elevations above treeline and by minimum temperatures at all other elevations), the date snowmelt stops (both local SCA<10% and combined effects of higher elevation snowmelt and transfer efficiency), the depth of the local soil reservoir ($Z_r$), and the ability of the soil to hold onto water once snowmelt input ceases ($K$ and $\theta_c$). For this parameter set, more ET occurs in the lower basin (example, 2300 m elevation band) in the wetter year (2006, Fig. 9g) than the drier year (2004, Fig. 9h) because moisture limitation begins in July in 2004 (Fig. 9e) but not until mid-August in 2006 (Fig. 9f).
The lowest elevations, where more precipitation falls as rain, are moisture-limited for all reasonable parameter sets, while the highest elevations, above treeline where snowcover lasts late into the summer, are energy-limited for all reasonable parameter sets (Fig. 10a). Only intermediate elevations (2000-3000 m) are strongly sensitive to the parameter set chosen. However, as shown in Fig. 1b, these same intermediate elevations are the ones that dominate the runoff in both the upper and lower watersheds examined. In Fig. 10, these intermediate elevations tend towards water limitation (more ET in wet years), but for other parameter sets (not shown) can tend towards an even balance or towards energy limitation. The majority of parameters result in annual ET at these elevations that stays relatively fixed from one year to the next.

With the parameters of $t_c = 0.75; K = 1 \text{ m day}^{-1}; Z_r = 0.75 \text{ m};$ and $\theta_k = 0.3$, the model matches the observed mean and variability for the lower basin well (Fig. 10b). However, while it approximates the mean ET in the upper basin, it does not well-represent the upper basin’s interannual variability. This may be due to error in PRISM precipitation variations at the highest elevations due to sparse observations above 3000 m (which would bias the observed patterns) or to high elevation vegetation behavior differing from the parameters set here.

5 Summary and Discussion

In the Sierra Nevada, California, annual ET is limited more by moisture availability than energy availability, and thus, annual ET is unlikely to be strongly affected by warming temperatures and earlier snowmelt, with the exception of the highest elevations above treeline, which currently remain snow-covered throughout most of the growing season. These observed water balance results agree qualitatively with earlier modeling studies [e.g., Christensen et al.]
The new insight we have gained relates to which model components matter most in recreating these results. Mean annual ET can vary by as much as 30 cm depending on the basin soil properties (Fig. 6). When soil properties are not well known, this results in a poorly-constrained parameter space that can be set in a variety of configurations to match observed precipitation and runoff. Fortunately for studies of change in ET between years (such as climate change experiments), the interannual variability in ET is less sensitive to the soil parameterization (Fig. 7 and Fig. 8), with about a 10 cm range in the calculated difference between dry and wet years depending on the parameter set chosen. For this region, with about 100 cm annual average precipitation, a 10 cm range corresponds to approximately 10% uncertainty.

However, our work here demonstrates that this 10 cm range is primarily controlled by one parameter. Specifically, water transfer is the most important parameter required to model more ET in wetter years in mid to low snow elevations in the Central Sierra. This transfer represents movement of water from higher, cooler locations with snow to lower, warmer elevations with transpiring vegetation, which is an important feature of steep terrain. This downslope transfer of water in our model is intended to represent several physical mechanisms, including the following: interflow along the soil-bedrock interface, perched groundwater flow in glacial moraines, overland flow particularly in regions with exposed granitic domes, groundwater flow through the bedrock fractures, and streamflow which can provide water to downstream riparian areas. In our basin, most of this transfer is likely due to interflow, which is partitioned into baseflow to rivers and shallow down-slope perched groundwater flow. Observations and modeling of groundwater levels in the nearby Tuolumne Meadows indicate that these hillslope water fluxes are an essential component of the meadow groundwater budget [Lowry et al. 2010]. While we
have modeled this water movement as the gross downslope movement of water from one
elevation band to the next, each individual process is much more complex, operating at scales
ranging from a single hillslope, for processes such as interflow, to basin-wide, for processes such
as flow in the bedrock aquifer. Our model demonstrates that while it may not be necessary to
simulate all of the complexities of these spatially variable transfer mechanisms, at least a crude
representation of downslope water transfer is required to simulate the observed variability in ET.

When interpreting our results, we must be clear that estimating ET from the water balance is
subject to uncertainty in runoff, to uncertainty in the measurements and distribution of
precipitation, and to uncertainty in the assumption that storage does not change between years.
As stated previously, the USGS streamflow measurements used here are among the most
accurate in the network, with relatively small uncertainty on an annual basis. Characterizing the
uncertainty in the precipitation distribution across complex terrain is more difficult, and we
believe this to be the largest source of error in our estimates of ET (see Lundquist et al. 2010 for
further discussion). To check the robustness of our results to potential errors in basin-wide
precipitation, we also employed an alternative approach to estimating ET using precipitation data
collected at the Yosemite Valley Park Headquarters and streamflow at our two USGS sites from
1916 to 2000. (Reliable precipitation data was not available at this location after 2000.) To map
point precipitation to each basin’s areal-average precipitation, we used the mean annual areal
precipitation calculated by PRISM divided by the mean annual precipitation measured at the
gage for the period of overlap (water years 1981-2000) to calculate a constant weight function
for each basin. Despite the assumption of a constant precipitation distribution each year in this
alternate estimation of precipitation, the same general pattern emerged as in our 1980-2007
dataset. Namely, more ET occurred in wet years than in dry years in the lower basin, while ET
remained relatively constant between wet years and dry years in the upper basin. The pattern of the upper basin reaching a maximum ET in average years, with less ET for both wetter and drier years (as suggested by visual inspection of Figure 4), did not appear in the historic dataset, suggesting that this pattern is likely less robust. Changes in the average annual precipitation at the highest elevations, where there are no gages, would change the average ET estimated by the water balance but would not change the year to year variability.

Our analysis assumes that there is limited water storage from one year to the next, and annual changes in storage would introduce error into our estimates of ET. We assumed annual changes in storage are small because the Merced basin is primarily granite overlain with shallow soils, and this is supported by our model with the best-fit parameter set illustrated in Figures 9 and 10. With the exception of 2007, which had a substantial late-summer rain event [Lundquist and Roche 2008], all elevations below 2800 m had soil moisture reduced to the wilting point each year (e.g., Figure 9). Elevations between 2800 and 3400 m had the greatest difference in end-of-water-year soil storage between a dry year (2004) and a wet year (2005), with an average of 15 cm more soil water at the end of the wet year. Above 3400 m, there was on average of 2 cm more soil water at the end of the wet year. Weighted by basin area, these changes in soil water represented 6 cm more water stored in 2005/2006 than in 2003/2004 in the upper basin, but only 1 cm more water in the lower basin. Therefore, the differences in simulations and observations for the upper basin (Figure 10) may be partially explained by changes in storage in these upper elevations. However, our results for changing ET in the lower basin are not strongly influenced by changes in storage and are robust.

Our results demonstrating the importance of water transfer to variations in ET are specific to the seasonal snow zone. This differs from prior work showing ET increases with increased
precipitation in rain-dominated regions [e.g., Hickel and Zhang 2006] because water transfer across elevations is less critical when rain falls during the growing season and when precipitation and melt rates do not have strong elevational gradients. Such water transfer is included as part of hillslope processes in some form in most distributed hydrologic models, e.g., DHSVM [Wigmosta et al. 1994; see Lowry et al. 2010 for a discussion of the importance of subsurface water transfer to maintain meadow water levels in steep terrain] or RHESSys [Tague and Band 2004], which documented similar ET patterns to those presented here [Christensen et al. 2008].

However, most climate predictions are made with coarser models that neglect this key process [e.g. Koster and Milly 1997; Gedney et al. 2000; Cornwell and Harvey 2007]. Our work suggests that such models will not accurately represent changes in ET in response to changes in precipitation and snowpack timing in steep, snow-fed terrain, and thus, should not be used for this purpose. Specifically, as illustrated in Figure 6 for the lower basin, for a parameter set of K = 1 m day$^{-1}$; Z$_r$ = 0.75 m; $\theta_{fc}$ = 0.3, a simulation with no water transfer (t$_c$=0) had 25 cm less mean annual ET than a simulation with substantial water transfer (t$_c$=75%). This would result in a 25 cm overestimation of the amount of water available for runoff, which is substantial compared to actual area-normalized runoff (30 to 150 cm). As illustrated in Figure 7, the simulation with no water transfer showed no change in ET between wet and dry years in the lower basin, while the simulation with water transfer (t$_c$=75%) showed 10 cm more ET in wetter years. This would result in error in the model’s estimate of how runoff varies with changing precipitation. While mean errors in ET would likely be compensated for during model calibration, the model without water transfer would not exhibit the correct sensitivity to climate change. Based on these results, a hydrologic model with at least a crude representation of
hillslope processes, i.e., lateral moisture transfer, is essential in hydroclimatic modeling to adequately represent ET in complex terrain with seasonal snow.
6. Acknowledgements

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8. Figure Captions

Figure 1. (a) Locations of gages (Merced River at Happy Isles and Merced River at Pohono) and of meteorological stations (Tuolumne and Dana) in Yosemite National Park, California. (b) Area as a function of elevation for the basin above Pohono gauge, the basin above the Happy Isles gauge (Upper Basin), and the basin area below the Happy Isles gauge that drains to the Pohono gauge (Lower Basin).

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Figure 3. Annual, basin-averaged water balance components for (a) Merced River above Happy Isles (Upper Basin) and (b) for the Merced River above Pohono Bridge and below Happy Isles (Lower Basin) for 1992 to 2000. Evapotranspiration (ET) is estimated as the residual of annual basin precipitation minus runoff.

Figure 4. Scatterplot of ET in the upper and lower basins as a function of annual basin-average precipitation. For the lower basin, the best fit line has a slope of 13%, and the correlation is significant (p<0.01).
Figure 5. Illustration of soil water accounting for determining actual ET in the model. Model details are in main text.

Figure 6. Model simulation performance relative to lower basin average annual ET for 2003-2007, which is 54 cm, based on water balance calculations. Contour intervals are each 2.5 cm, and a value of zero indicates a parameter set that precisely predicts the mean ET of the lower basin. The y-axis of each plot shows the rooting depth ($Z_r$). The x-axis shows the soil field capacity ($\theta_{fc}$). Columns indicate the transfer coefficient ($tc$), representing the percent of soil water transferred between one elevation band and the next. Rows indicate the value of $K$. Other model parameters were held fixed as follows: Porosity, $n$, is 0.4. Melt ceases when SCA drops below 10%, and ET begins at elevations about 3000 m when SCA reaches 0. Plant water stress begins at $\theta_c = 0.25$, and the wilting point occurs at $\theta_{wp} = 0.10$.

Figure 7. Mismatch between modeled and observed mean ET differences between dry years (2004 and 2007) and wet years (2005 and 2006) for the lower basin. The observed mean ET differences between dry years (2004 and 2007) and wet years (2005 and 2006) for the lower basin is -12 cm based on the water budget calculations (Table 2). The parameter values are the same as in Fig 6.

Figure 8. Mismatch between modeled and observed mean ET differences between dry years (2004 and 2007) and wet years (2005 and 2006) for the upper basin. The observed mean ET differences between dry years (2004 and 2007) and wet years (2005 and 2006) for the upper
basin is 3 cm based on the water budget calculations (Table 2). The parameter values are the same as in Fig 6.

Figure 9. Timeseries for a dry year (2004, left) and a wet year (2006, right) for (a, b) fractional snow covered area, (c,d) snowmelt, (e,f) soil moisture, and (g,h) actual ET as measured by MODIS and calculated by the model for a parameter set with \( t_c = 75\% \), \( K = 1 \text{ m day}^{-1} \), \( \theta_c = 0.3 \), and \( Z_r = 0.75 \). The horizontal dashed lines in (e,f) identify the field capacity and wilting point.

Figure 10. (a) Model-simulated annual ET for individual elevation bands and (b) model-simulated and observed ET for upper and lower Merced Basins for the same parameter set as in Fig. 9. Years are ranked from driest to wettest, where 2007 and 2004 were dry, 2003 was intermediate, and 2005 and 2006 were wet (see Table 2).
### Table 1. Parameter Values Used in Model Simulation Experiments

<table>
<thead>
<tr>
<th>Parameter</th>
<th>$\theta_c$</th>
<th>$\theta_e$</th>
<th>$\theta_{wp}$</th>
<th>$Z_r$</th>
<th>$n$</th>
<th>$K$</th>
<th>tc</th>
<th>SCA-Melt cutoff</th>
<th>ETstart at el&gt;3000 m</th>
<th>Freeze cutoff</th>
</tr>
</thead>
<tbody>
<tr>
<td>Values Used</td>
<td>0.15-0.4</td>
<td>0.25</td>
<td>0.10</td>
<td>0.25-1.5 m</td>
<td>0.4</td>
<td>1, 0.1, 0.01, 0.001</td>
<td>0-0.75</td>
<td>0.1</td>
<td>SCA=0</td>
<td>-1°C</td>
</tr>
</tbody>
</table>
Table 2. Annual values from water balance (all in cm). HI indicates the upper basin, and LO indicated the lower basin. Lowest row is difference between dry and wet years.

<table>
<thead>
<tr>
<th></th>
<th>HI</th>
<th>HI</th>
<th>LO</th>
<th>LO</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Precipitation</td>
<td>Discharge</td>
<td>ET</td>
<td>Precipitation</td>
</tr>
<tr>
<td>2003</td>
<td>114</td>
<td>69</td>
<td>45</td>
<td>118</td>
</tr>
<tr>
<td>2004</td>
<td>92</td>
<td>49</td>
<td>43</td>
<td>96</td>
</tr>
<tr>
<td>2005</td>
<td>142</td>
<td>112</td>
<td>30</td>
<td>152</td>
</tr>
<tr>
<td>2006</td>
<td>158</td>
<td>121</td>
<td>36</td>
<td>166</td>
</tr>
<tr>
<td>2007</td>
<td>60</td>
<td>32</td>
<td>28</td>
<td>67</td>
</tr>
<tr>
<td><strong>mean</strong></td>
<td>113</td>
<td>77</td>
<td>36</td>
<td>120</td>
</tr>
</tbody>
</table>

\[
\text{mean(2004 \\
& 2007)} - \text{mean(2005 \\
& 2006)} = -74 \quad -76 \quad 3 \quad -78 \quad -65 \quad -12
\]
Table 3. Vegetation fractions for different elevation zones of upper and lower basins, based on Yosemite National Park vegetation surveys.

<table>
<thead>
<tr>
<th>Elevation Range (m)</th>
<th>Conifers (%)</th>
<th>Broadleaf (%)</th>
<th>No data (%)</th>
<th>Meadow (%)</th>
<th>Barren (%)</th>
<th>Lakes (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Upper Basin</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1200-1800</td>
<td>63.3</td>
<td>11.5</td>
<td>0.0</td>
<td>0.0</td>
<td>25.3</td>
<td>0.0</td>
</tr>
<tr>
<td>1800-2600</td>
<td>88.3</td>
<td>3.5</td>
<td>0.0</td>
<td>0.2</td>
<td>7.7</td>
<td>0.3</td>
</tr>
<tr>
<td>2600-4000</td>
<td>59.3</td>
<td>0.6</td>
<td>0.1</td>
<td>4.8</td>
<td>34.3</td>
<td>0.9</td>
</tr>
<tr>
<td><strong>Lower Basin</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1200-1800</td>
<td>59.3</td>
<td>22.7</td>
<td>0.0</td>
<td>1.7</td>
<td>16.2</td>
<td>0.1</td>
</tr>
<tr>
<td>1800-2600</td>
<td>91.7</td>
<td>2.0</td>
<td>0.0</td>
<td>0.5</td>
<td>5.5</td>
<td>0.3</td>
</tr>
<tr>
<td>2600-4000</td>
<td>77.0</td>
<td>1.4</td>
<td>0.0</td>
<td>2.2</td>
<td>18.8</td>
<td>0.6</td>
</tr>
</tbody>
</table>
10. Figures

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