HYDROLOGICAL PROCESSES
Hydrol. Process. **28**, 5048–5064 (2014)
Published online 24 August 2013 in Wiley Online Library
(wileyonlinelibrary.com) DOI: 10.1002/hyp.9943

Effects of changes in winter snowpacks on summer low flows: case studies in the Sierra Nevada, California, USA

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Abstract:

Seasonal low flows are important for sustaining ecosystems and for supplying human needs during the dry season. In California's Sierra Nevada mountains, low flows are primarily sustained by groundwater that is recharged during snowmelt. As the climate warms over the next century, the volume of the annual Sierra Nevada snowpack is expected to decrease by ~40–90%. In eight snow-dominated catchments in the Sierra Nevada, we analysed records of snow water equivalent (SWE) and unimpaired streamflow records spanning 10–33 years. Linear extrapolations of historical SWE/streamflow relationships suggest that annual minimum flows in some catchments could decrease to zero if peak SWE is reduced to roughly half of its historical average. For every 10% decrease in peak SWE, annual minimum flows decrease 9–22% and occur 3–7 days earlier in the year. In two of the study catchments, Sagehen and Pitman Creeks, seasonal low flows are significantly correlated with the previous year's snowpack as well as the current year's snowpack. We explore how future warming could affect the relationship between winter snowpacks and summer low flows, using a distributed hydrologic model Regional Hydro-ecologic Ecosystem Simulation System (RHESSys) to simulate the response of two study catchments. Model results suggest that a 10% decrease in peak SWE will lead to a 1–8% decrease in low flows. The modelled streams do not dry up completely, because the effects of reduced SWE are partly offset by increased fall or winter net gains in storage, and by shifts in the timing of peak evapotranspiration. We consider how groundwater storage, snowmelt and evapotranspiration rates, and precipitation phase (snow *vs* rain) influence catchment response to warming.

KEY WORDS snowpack; low flow; climate change; baseflow; SWE; mountain

Received 7 January 2013; Accepted 18 June 2013

INTRODUCTION

Flows during dry periods between storms, often called base flows or low flows, are important for sustaining aquatic ecosystems and meeting human needs. Low flows are particularly important in Mediterranean climates like California's, in which a ~6-month dry season coincides with peak water demand. For example, in California, base flows are critical to sustaining agricultural production during the rainless summer growing season. Fish also depend on base flows remaining high enough to supply in-stream habitat refugia with cool, oxygenated water (e.g. May and Lee, 2004). Sufficient low flows are also required to prevent saltwater intrusion into freshwater intake pumps in the Sacramento River Delta (Knowles and Cayan, 2002; Hayhoe *et al*., 2004). Seasonal low flows are sustained by the release of water stored as groundwater, as snowpacks, or as impoundments behind dams. In California's Sierra Nevada mountains, a substantial fraction of winter precipitation is typically stored aboveground in seasonal snowpacks that persist beyond the end of the winter precipitation season (Hayhoe *et al*., 2004). These snowpacks usually melt in late spring or early summer, depending on altitude, aspect, shading, and other factors (Lundquist *et al*., 2004). Snowmelt sustains flows through the spring and early summer, and infiltrates into the ground to recharge groundwater. Some of this stored groundwater then slowly feeds low flows later in the season (Panagoulia and Dimou, 1996). In semi-arid and Mediterranean regions like California, groundwater as deep as 2–6 m below the ground surface also directly supplies water for transpiration (e.g. White, 1932; Nichols, 1994). Because seasonal low flows are important for human and ecosystem needs, and because they are dependent on stored subsurface water and evapotranspiration losses, it is important to understand how low flows respond to changes in the temporal distribution of net recharge and evapotranspiration.

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Both evapotranspiration and the temporal distribution of net changes in storage are expected to respond to climate warming. Evapotranspiration losses may change as plants experience more widespread water stress in a warmer climate (Bates *et al.*, 2008, Table 3.2), or as leaf-level water use efficiency increases under higher atmospheric CO_2 concentrations (Bates *et al.*, 2008). In snow-dominated regions such as the mountainous western USA, the temporal distribution of net changes in storage will largely depend on the phase of precipitation (i.e. the fraction falling as snow *versus* rain), and the timing and volume of snowmelt or rainfall (Earman *et al.*, 2006; Winograd *et al.*, 1998). Winter rainstorms and spring/summer snowmelt can be expected to yield different amounts of net recharge because (1) the timing and intensity of the arrival of the liquid (infiltrating) phase differ (Kingsmill *et al.*, 2006; Lundquist *et al.*, 2009), (2) the antecedent soil moisture – and thus the conductivity and infiltration capacity – differs (Brady and Weil, 2008; Perkins and Jones, 2008), and (3) the immediate losses of near-surface water to evapotranspiration differ (Christensen *et al.*, 2008).

Climate change is expected to affect the volume and timing of snowmelt across the western USA (e.g. Hayhoe *et al.*, 2004; Mote *et al.*, 2005; Bates *et al.*, 2008). Compared with historical averages from the mid- to late-20th century, Sierra Nevada snowpack volumes are expected to decrease 40–90% by 2100 (Leung and Wigmosta, 1999; Knowles and Cayan, 2002; Hanson *et al.*, 2004; Hayhoe *et al.*, 2004). The projected decrease in snowpack volume may be due to (1) more frequent melt events throughout the winter (Mote *et al.*, 2005), (2) warmer temperatures that shift the phase of winter precipitation from snow to rain (Lettenmaier and Gan, 1990; Lettenmaier and Sheer, 1991; Cayan *et al.*, 1993; Gleick and Chalecki, 1999; Lettenmaier *et al.*, 1999; Dettinger and Cayan, 2003; Leung *et al.*, 2004; Knowles *et al.*, 2006), or (3) lower total precipitation (e.g. Dettinger *et al.*, 2004). Current downscaled global climate model predictions for the Sierra Nevada suggest that total precipitation may increase or decrease by 10% or less (e.g. Leung and Wigmosta, 1999; Dettinger *et al.*, 2004; Christensen *et al.*, 2007), so we do not focus on changes in total precipitation in this study. Instead, we focus on the anticipated changes in the phase of precipitation and their impacts on the temporal distribution of net changes in storage. We also consider the effect of the anticipated earlier melt-out of the entire snowpack (Cayan *et al.*, 2001a,2001b; Mote *et al.*, 2005). We recognize that these effects are expected to vary with elevation, aspect, and shading (Gleick, 1987; Lundquist and Flint, 2006); elevation is considered explicitly in our modelling work.

Here, we explore the primary controls on catchment low flow response to changes in climate by examining historical trends and model predictions of responses to potential warmer temperatures. We ask, *how will warming change the phase and timing of precipitation? How will these changes affect the temporal distribution of net changes in storage and subsequent seasonal low flows? How important is the effect of warming on evapotranspiration and, in turn, on seasonal low flows? Are the amount and timing of low flows controlled more by changes in precipitation or changes in evapotranspiration?* Specifically, we explore the potential impact of significant reductions in snowpack volume, and the anticipated changes in the temporal distribution of net recharge and evapotranspiration, on low flows in Sierra Nevada streams and rivers. We estimate the historical sensitivity of streams to net changes in storage as indicated by changes in snowpack volume and snowmelt timing. We also use a coupled eco-hydrologic model, Regional Hydro-ecologic Ecosystem Simulation System (RHESSys; Tague and Band, 2004), to examine the effects of changes in precipitation phase and evapotranspiration due to climate warming.

HISTORICAL RELATIONSHIPS BETWEEN SNOW, SUBSURFACE STORAGE, AND LOW FLOWS

Site description and data collection

Sierra Nevada hydrology is dominated by California's Mediterranean climate, in which precipitation falls predominantly in the winter, snowmelt generates a broad peak of streamflow in late spring or early summer, and flows decline to annual minima in the late summer and autumn (e.g. Figure 1). We quantified the relationship between snowpack volume and stream flow at eight snowmelt-dominated Sierra Nevada catchments with unimpaired flows (i.e. free of dams and diversions). It is worth noting that because many streams in the Sierra Nevada are gauged only below dams or other impairments, it was the availability of unimpaired stream gauges, rather than snow pillows, that limited the number of catchments that could be used in this analysis. For our analysis, we included all catchments with at least a 10-year overlap of daily streamflow from USGS gauges and SNOTEL snowpack information. In catchments with multiple snow sensors, we used the sensor with the longest continuous record. At the eight sites that met our criteria for analysis in the Sierra Nevada (Table I and Figure 2), the elevations of the snow pillows vary between 2022 and 3547 m, and the stream gauges are located at elevations between 607 and 2184 m. Drainage areas range from 25 to 1373 km^2 (median = 118 km^2). Average annual runoff varies from 8 to 86 cm/year (median = 52 cm/year) over the available flow records.

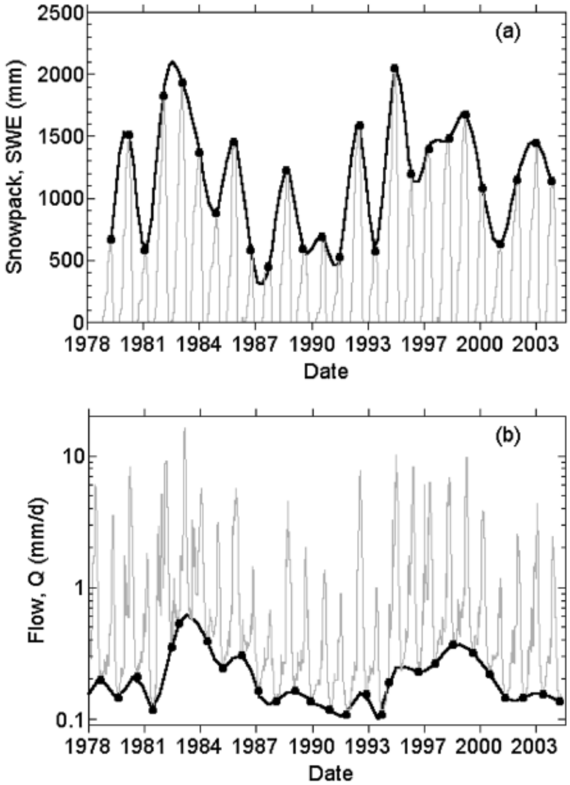

Figure 1. Grey lines are time series plots of (a) daily 15-day running median snow water equivalent (SWE) at the Independence Lake Snow Telemetry (SNOTEL) site located on the divide between the Sagehen Creek and Independence Lake basins, and (b) daily 15-day running median flow at Sagehen Creek (Q, log scale). The minimum flow in Sagehen Creek varies from year to year, partly in response to changes in peak SWE, as indicated by the black cubic spline curves in (a) and (b)

Flow–snowpack relationships

For each of our study sites, we calculated the 15-day running median daily flow and snow water equivalent (SWE). We used 15-day running medians to minimize the effect of individual, potentially spurious, values in the raw time series. To ensure that the windowing had no effect on the results, we also repeated the analyses using individual daily values and windowed medians of 3 and 7 days. Results for different window sizes never diverged by more than 10% from the reported values (and typically by much less than their reported uncertainties), so results for only the 15-day running medians are reported here. For each year at each site, we calculated the annual low flow as the minimum of the 15-day running median flow following the spring snowmelt. We also calculated the annual peak SWE as the maximum of the 15-day running median SWE for each water year (i.e. the maximum SWE that occurred in the winter or spring preceding the low flow for a given calendar year). We then normalized all annual low flows and peak SWEs by their average for all years at each site. For example, for a given site, each year's minimum flow was divided by the average of all years' minimum flows.

At all of the study catchments, normalized annual minimum flow is a roughly linear function of normalized annual maximum SWE (Figure 3). The slopes of these linear relationships estimate the percentage change in low flows that can be expected from a given percentage change in maximum SWE. Low flows are relatively sensitive to changes in SWE at six of the eight study sites (Pitman, Trout, and Ward Creeks, and South Fork Kern, Upper Truckee, and South Fork Mokelumne Rivers) with regression slopes that are steeper than one by at least one standard error (Table I, $p < 0.0001$ for all sites except the Upper Merced with $p = 0.0027$). Extrapolations of the linear relationships shown in Figure 3 imply that minimum flows at these six gauge locations could decrease to zero if peak winter SWEs were reduced to 20–61% of their historical averages (Table I). These linear extrapolations may or may not be representative of future low flows, because low SWE years in the historical data reflect a combination of lower total precipitation as well as rain/snow partitioning. Plots of normalized total precipitation *versus* normalized minimum flows (not shown) for the same time period, at sites where rainfall data were also available, generally exhibit similar patterns to those represented in Figure 3 and Table I. Trout Creek's low flows appear to be slightly more sensitive to changes in total precipitation than to changes in peak SWE. The similarity in patterns across most sites reflects the fact that snow has historically dominated the total annual precipitation. Future climate may reflect a larger shift in rain/snow partitioning than is represented by the historical data. If the partitioning differs in the historical record and a warmer future climate, low-flow responses will depend on subsurface storage characteristics. Later in this paper, we explore whether these linear extrapolations accurately predict low flows in a warmer climate and discuss possible low-flow responses to warming-induced precipitation phase change.

The slopes of the best-fit lines relating normalized SWE and minimum flow are significantly steeper than one at these six sites, implying that these sites exhibit a more-than-proportional relationship between maximum SWE and minimum flow. For these catchments, a 10% decrease in peak SWE results in an ~12(±1) to 25(±4)% decrease in minimum flow (Table I and Figure 3). Low flows at the other two sites (Sagehen Creek and the Upper Merced River) respond proportionally or less-than-proportionally to changes in maximum SWE. In these catchments, a 10% decrease in peak SWE corresponds to an ~8(±2) to 11(±3)% decrease in low flows.

The more-than-proportional relationship between minimum flows and maximum SWE can be explained by the seasonal dynamics of snowmelt, evapotranspiration, and streamflow recession. Following winters with lower peak SWE, recharge from snowmelt will end earlier in the year. Thus, more time will elapse between melt-out and the onset

Table I. Site information for the eight study catchments in the Sierra Nevada

Name		Altitude (m)				Snowpack *versus* Q (% of normal)	
Snow pillow	Stream gauge	Snow pillow	Stream gauge	Drainage area (km^2)	Years of overlapping record	Regression slope (± S.E.)	x-intercept (± S.E.)
Tamarack Summit (TMR)	Pitman Creek below Tamarack Creek	2349	2184	59	22	2.55 (0.47)	61 (23)
Ward Creek 3 (WC3)	Ward Creek at Hwy 89 near Tahoe Pines	2100	1948	25	23	2.08 (0.30)	52 (17)
Black Springs (BLS)	SF Mokelumne near West Point	2022	607	195	24	2.02 (0.36)	51 (21)
Upper Tyndall Creek (UTY)	SF Kern River near Onyx	3547	902	1373	33	1.74 (0.19)	42 (13)
Echo Peak 5 (EP5)	Upper Truckee River at South Lake Tahoe	2427	1938	142	22	1.69 (0.26)	41 (18)
Heavenly Valley (HVN)	Trout Creek near Tahoe Valley	2738	1942	95	24	1.25 (0.10)	20 (9)
Ostrander Lake (STR)	Upper Merced River at Happy Isles Bridge near Yosemite	2551	1250	469	15	1.12 (0.30)	10 (30)
Independence Lake (IDP)	Sagehen Creek near Truckee	2629	1966	27	24	0.84 (0.16)	−19 (22)

The last two columns indicate the best-fit slope and x-intercept of the minimum annual 15-day running median flow *versus* the maximum annual 15-day running median snow water equivalent (SWE), as shown in Figure 3. The x-intercept value indicates the percent of normal peak SWE at which low flows cease. Best-fit non-linear parameter estimates at Pitman were slightly better than linear ones but depended strongly on one influential outlier; Pitman Creek's best-fit linear slope excluding this outlier was 1.38 ± 0.33. The squared fit is slightly better than the linear fit at SF Mokelumne, but for ease of comparison with other sites, the linear parameters are included in this table.

of autumn rains, which typically end the low-flow period. The longer period of streamflow recession will result in lower low flows (Tague and Grant, 2009). Earlier snowmelt will also imply earlier emergence of seasonal grasses and forbs, and therefore more rapid losses of subsurface water to evapotranspiration.

In years with high peak SWE, the peak in snowmelt will occur later and the period of groundwater recession will be shorter before the onset of autumn rains. In some catchments, high-elevation snowpacks may persist through the summer in the wettest years, continuing to recharge groundwater throughout the growing season. The briefer period of groundwater recession and the closer timing of melt and peak evapotranspiration demand imply that low flows in wet years will be higher than one would expect solely from the difference in total precipitation.

In catchments where evapotranspiration is primarily derived from groundwater, a less-than-proportional relationship between low flows and peak SWE could potentially arise. In dry years, the groundwater table could drop below the typical rooting depth, forcing a reduction in evapotranspiration rates and thus limiting the decrease in low flows. Conversely, in wet years, the water table could rise into the more densely rooted zone, raising evapotranspiration rates and limiting the increase in low flows. However, the evidence for less-than-proportional relationships between low flows and peak SWE among our study sites is equivocal at best; even the shallowest slope in Figure 3 (Sagehen Creek) is indistinguishable from 1 (proportional response) within error.

The observed flow–snowpack response could also depend on the location of the snow measurement relative to the overall basin hypsometry. Precipitation varies strongly with elevation in the Sierra Nevada (e.g. Daly *et al.*, 1994). Therefore, if much of the basin lies at higher elevations that receive much more precipitation than measured at the SNOTEL site, then low-flow responses could be amplified by the 'unmeasured' snowpack. Evapotranspiration can also vary with elevation (Christensen *et al.*, 2008) and can amplify or moderate the streamflow response to precipitation. However, a comparison of Table I and Figure 2b shows that basins with SNOTEL sites located relatively high in the catchment exhibit a wide range of responses (e.g. Sagehen, SF Mokelumne, and SF Kern), as do basins with SNOTEL sites located in the mid-range of catchment elevations (e.g. Upper Merced and Pitman).

Melt-out and low-flow timing

Low-flow discharge and timing depend on both melt-out timing and the volume of peak SWE. In dry years, low flows are smaller and occur earlier (Figure 4). To understand the

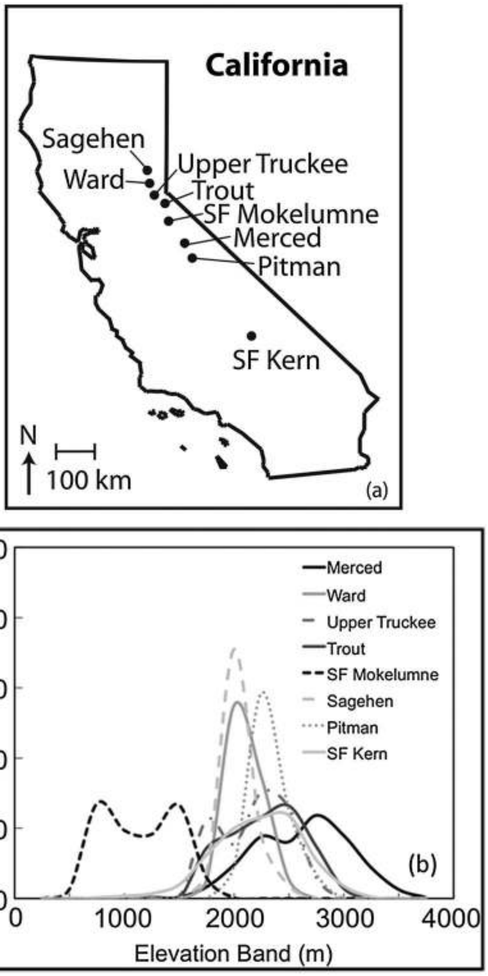

Figure 2. (a) Map of stream gauge and snow pillow locations selected for this study. (b) Hypsometric plots for each of the sites. See Table I for additional site information

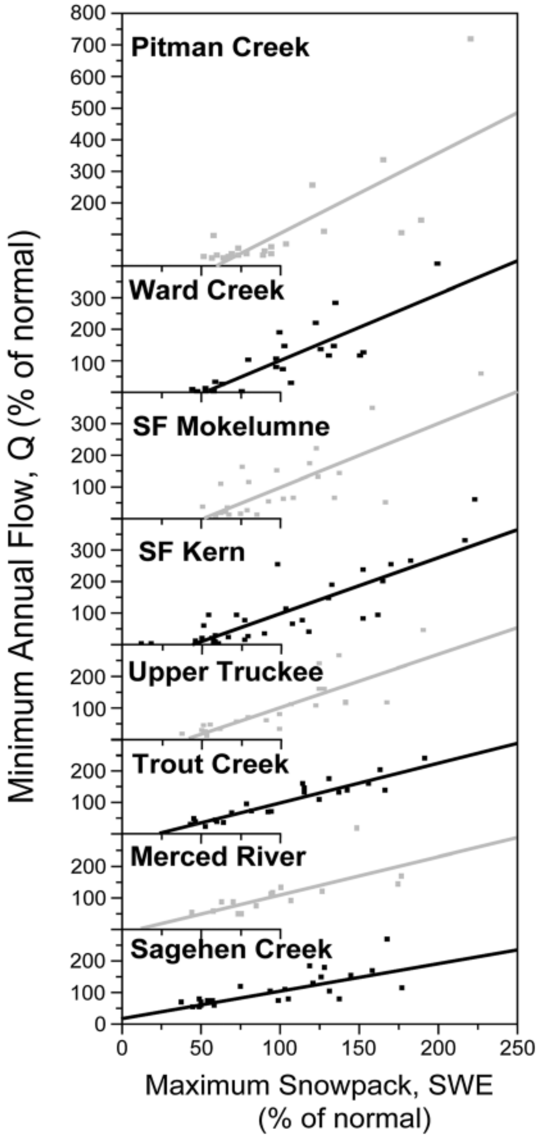

Figure 3. Relative minimum runoff (*Q*) as a function of relative maximum snow water equivalent (SWE) for each study catchment. The solid lines indicate the best-fit regression lines for each catchment. In most cases, the solid line has a slope that is significantly steeper than 1, indicating a more-than-proportional runoff response to changes in snowpack. The *x*-intercepts (shown on the truncated *x*-axes) of the best-fit lines also indicate that streams may run dry with relatively small decreases (~45% or more) from current average peak SWE. Data points alternate between black and grey solely to visually distinguish sites from one another

role of melt timing, we recorded the first day on which the SWE was zero – the day of melt-out – at each snow sensor location for each year of record. We also defined the low-flow period as the range of days that discharge remains continuously below the 25th percentile of all historical flows. We chose this range because at most sites, it encompasses at least 1 day in most years, while also excluding most snowmelt and autumn storm days in drought years. In a Mediterranean climate with a long dry period, the recession limb of the hydrograph is relatively flat near the end of the dry season, so low flows may occur for several days or weeks, depending on site conditions and the onset of autumn or winter storms. With this in mind, we also noted the first, mean, and last day on which the low-flow period occurred (in Julian days). As shown in Figure 5, there is a strong correlation between maximum snowpack and the mean day of melt-out. Melt-out occurs 3–7 days earlier for each 10% decrease in peak SWE (Figure 5a). Based on the best-fit linear relationship shown in Figure 5b, the mean day of the low-flow period also occurs 3–7 days earlier for each 10% decrease in peak snowpack. Most of the shift occurs because the low-flow period starts earlier, not because of earlier autumn rainfall. The mean day of the low-flow period varies by as much as ~3 months across all locations.

The scatter in the relationship between low-flow timing and maximum SWE (Figure 5b) is larger than the scatter in the relationship between melt-out timing and maximum SWE (Figure 5a). Melt-out is typically expected to be a

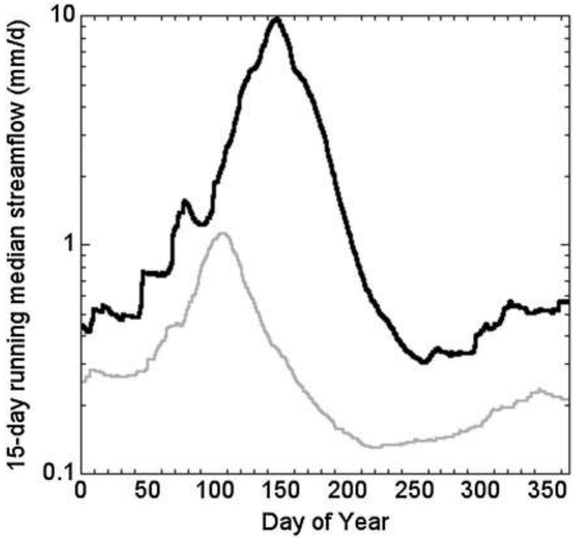

Figure 4. Average of the annual hydrograph for the wettest (black) and driest (grey) 5 years of record at Sagehen Creek. Low flows are lower and reach a minimum value earlier in dry years than in wet years. The peak flow also occurs substantially earlier in dry years relative to wet ones

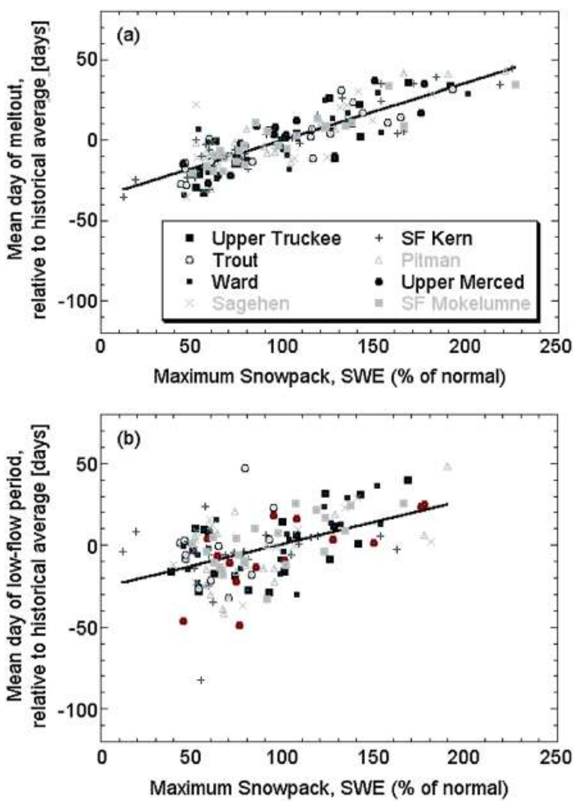

Figure 5. The timing of melt-out and the middle of the minimum flow period (as day of year) for all study catchments, plotted as functions of the relative maximum snowpack (SWE) with overall best-fit line shown. The upper graph (a) shows deviations from the mean first day of zero snowpack. The lower graph (b) shows deviations from the mean day of the low-flow period. The overall trends indicate that a 10% decrease in maximum SWE will result in snowmelt and the low-flow period occurring ~3–7 days earlier in the year. The low-flow period is defined as the range of days with flow less than the 25th percentile of flow (see text)

function of elevation, aspect, and temperature (Lundquist *et al.*, 2004). Low-flow timing is expected to depend on additional factors, including geologically mediated groundwater storage and depletion rates (Jefferson *et al.*, 2008), leading to more scatter in its dependence on peak SWE (Figure 5b). Low-flow recession characteristics often reflect both geologic and vegetative controls on water movement (Hall, 1968; Singh, 1968; Tallaksen, 1995 & references in each), although sometimes baseflow recessions are independent of evapotranspiration signals (Post and Jakeman, 1996) or are interpreted as reflecting a combination of snowmelt timing and geological controls (Tague and Grant, 2009). At all eight sites, we found that the start date of the low-flow period is significantly correlated with melt-out timing and that fall rainstorms that quickly raise streamflows often mark the end of the low-flow period. About 56% of the time across all of the sites in this study, the end of the low-flow period is clearly defined by a fall storm event. In about 6% of cases, the low-flow period ends without a clear connection to a storm event. The remaining 38% of cases are unclear. Combined with observations that fall has become wetter across the western USA (e.g. Lettenmaier *et al.*, 1994), this suggests that additional work looking at storminess in the fall and winter would be worthwhile. End dates of the low-flow period in any given year rarely coincide across sites (consistent with increased scatter in the mean day of the low-flow period seen in Figure 5b), implying that local-scale storms may play a large role in determining the length of the low-flow period.

Memory

At some locations, low flows exhibit a ‘memory effect’ in which they depend not only on the current year's snowpack but also on the previous year's snowpack. Sagehen Creek shows this memory effect most clearly (Figure 6a). We divided the snowpack and low flow data at Sagehen Creek into two groups: years for which the previous year's snowpack was above average (closed symbols) and years for which it was below average (open symbols). Low flows are more sensitive to a given year's snowpack if the previous year's snowpack was above average (as seen in the steeper slope of the best-fit line for the solid symbols in Figure 6a). Thus, a wet year following a wet year produces higher flows than a wet year following a dry year. Note that low flows in dry years at Sagehen are approximately the same, regardless of whether the previous year was wet or dry. Risbey and Entekhabi (1996) showed that streamflow is less responsive to precipitation after a drought year at the larger Sacramento Basin scale. They attributed this ‘drought memory’ to atmospheric, geologic, and vegetative effects, but did not explore which tributaries to the Sacramento might be more likely to exhibit a memory effect. At each of our sites, we performed a multiple regression of low flows against peak

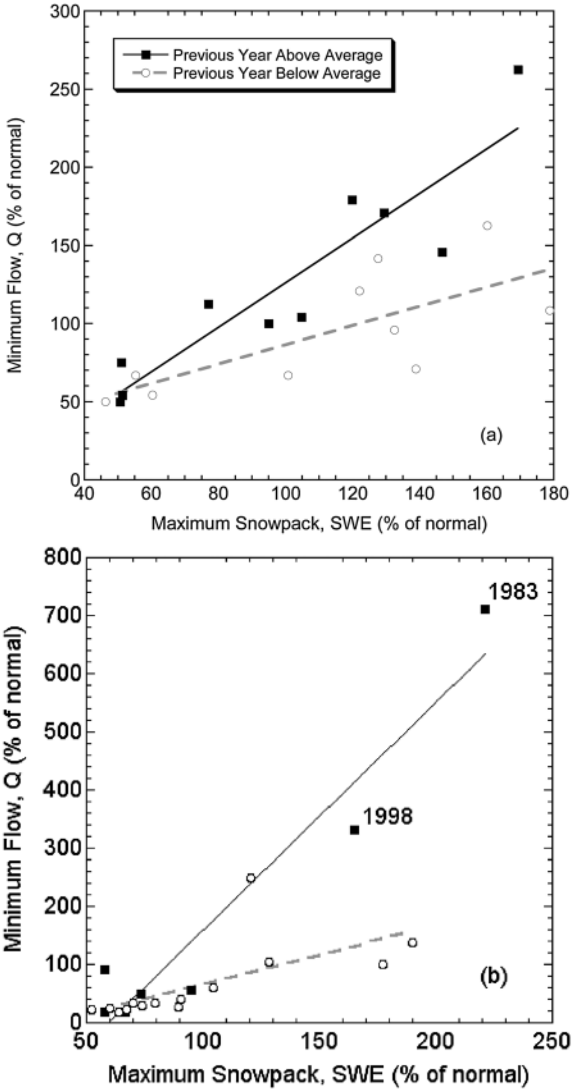

Figure 6. The annual minimum flow at (a) Sagehen Creek and (b) Pitman Creek depends not only on the current year's snowpack but also on the snowpack of the previous year as measured at the Independence Lake or Tamarack Summit SNOTEL sites. Relative minimum flow (*Q*) is plotted as a function of the current year's relative maximum snowpack (SWE), normalized as described in the text and Figure 3. Two subsets of data are distinguished: the closed symbols indicate when the previous year's SWE was above average, and the open symbols indicate when the previous year's SWE was below average. When the previous year's SWE is above average, minimum *Q* is more sensitive to the current year's maximum SWE. Two high-flow years (1983 and 1998) in Pitman are indicated in (b); without these years, there is no statistically significant memory effect

SWE for up to three previous winters to determine the persistence and significance of the memory effect. The only two sites with statistically significant memory effects were Sagehen Creek (Figure 6a) and Pitman Creek (Figure 6b) where the memory effect persisted for the two previous winters. Pitman Creek is predominately underlain by granite and granodiorite, whereas Sagehen Creek is underlain by volcanics. At Pitman Creek, the statistical significance of the memory effect depends on the above-average SWE years in 1983 and 1998. These are the only two years at Pitman Creek in which an above-average snowpack follows an above-average snowpack, and thus, it is possible that the Pitman memory effect is an artifact of a limited record.

At Sagehen Creek, the persistent effect of past snowpacks on low flows likely reflects the catchment hydrogeology. Unlike the other sites in this study, Sagehen is underlain by a layer of volcanics, including pyroxene and basaltic andesite (Sylvester *et al*., 2007). The other study sites are generally underlain by granites, which usually have more limited groundwater storage capacity (Kakue and Kishi, 2003; Price, 2009). At least 14 springs are found within the Sagehen catchment boundaries, and 15- to 28-year-old groundwater contributes nearly all the streamwater at low flows and up to 70% of streamwater at high flows (Erman and Erman, 1995; Rademacher *et al*., 2005). Shun and Duffy (1999) reported that inter-annual signals indicating long-term memory are strengthened at sites in the Great Salt Lake basin where groundwater dominates streamflow. Similar memory effects are associated with groundwater dominance in streams in the US Pacific Northwest (Jefferson *et al*., 2008; Tague and Grant 2009; Mayer and Naman, 2011). Thus, sites with a strong memory effect may reflect an important groundwater contribution to streamflow. Conversely, sites without a strong memory effect may have a smaller groundwater signal in streamflow. Memory effects and groundwater contributions also alter the sensitivity of summer flow to climate warming. Although flows are generally higher in groundwater-dominated streams, summer streamflows decrease by approximately four times more in groundwater-dominated streams than in other streams, per unit decrease in recharge, given similar timing of peak snowmelt (Tague *et al*., 2008; Tague and Grant, 2009). Although greater groundwater contributions may buffer changes in high-flow periods (winter and spring), greater groundwater storage essentially smooths the hydrograph and may in fact extend the impact of changes in the timing of recharge further into the summer season (and for streams like Sagehen into following years), and thus accentuate the impact of changing snowmelt storage on low flows.

As discussed in the introduction, climate model projections for California suggest that warmer temperatures will result in smaller snowpacks that melt earlier in the year. The relationships between changes in snowpack volume and melt timing for the eight sites in the Sierra Nevada (Figure 5) suggest that in many streams, one would expect to see a significant decrease in flows and a shift to earlier arrivals of low flows. Historically, higher peak SWE generally corresponds to wetter years and lower peak SWE to drier years because most precipitation falls as snow. In the future, a larger fraction of precipitation is projected to fall as rain, and it is difficult to know whether the historical relationships between peak

SWE and low flows will accurately describe catchment behaviour. Extending the approach used in Figure 3, we performed multiple regressions of historical low flows against three potentially explanatory factors: peak SWE, total rainfall during snow-covered periods, and total rainfall during snow-free periods. Rainfall, either during snow-covered or snow-free periods, is not significantly correlated with low flows at any of our study sites, except at Sagehen Creek. At Sagehen, only rainfall during snow-free periods led to a significant difference in low-flow responses: a 10% increase in rain falling during snow-free periods coupled with a 10% decrease in peak SWE would lead to a an ~4.1% decrease in low flows, whereas without the additional rainfall, low flows would be expected to decrease by ~8% for a 10% decrease in peak SWE. However, low flows were not significantly correlated with rain falling during the snow season at any of our sites, implying that the impact of shrinking snowpacks on low flows would not be offset by increased winter rainfall.

FUTURE RELATIONSHIPS BETWEEN PRECIPITATION, SUBSURFACE STORAGE, AND LOW FLOWS

In the rest of this paper, we discuss the results from model simulations exploring how warming may alter low flows by changing precipitation phase, subsurface storage, and evapotranspiration demands. We use the RHESSys model to evaluate the sensitivity of low flow/snowpack relationships to a range of warming conditions (outlined later) and compare the model results with the historical low flow/snowpack relationships discussed earlier.

Site, model, and description of sensitivity analysis

We modelled two sites in the Sierra Nevada: Sagehen Creek, in the northern part of the range, and the Upper Merced River, which flows through Yosemite National Park in the central part of the range (Figure 2). The catchments differ substantially: Sagehen is ~27 km^2, with a peak elevation of 2672 m, volcanic geology, and nearly complete vegetative cover, whereas the Upper Merced is ~469 km^2 with a peak elevation of 3997 m, granitic geology, and vegetation covering ~75% of the catchment area. We examine each component of the water budget to compare the two catchments' responses to climate change.

We use RHESSys, a spatially distributed watershed hydrologic model (see detailed explanation of subcomponents in Tague and Band, 2004), to model both sites. Modelled hydrologic processes include interception, snow accumulation and melt, infiltration, evaporation, transpiration, and vertical drainage between unsaturated and saturated stores, as well as lateral redistribution of shallow groundwater and drainage to deeper groundwater stores. Snowmelt is estimated by combining an energy budget approach for radiation-driven melt with a temperature-index based approach for latent-heat-driven melt processes. Precipitation is distributed spatially based on an isohyet multiplier, according to the following equation:

$$P_{\text{patch}} = P_{\text{base}} + \left(1 + k\left(e_{\text{patch}}\text{-}e_{\text{base}}\right)\right) \qquad (1)$$

where $k = 0.0002$ and 0.0003 per metre above the base station at Sagehen and the Upper Merced, respectively, e_{patch} and P_{patch} are the elevation and precipitation of any given patch, and e_{base} and P_{base} are the elevation and precipitation at the meteorological station. When no additional spatial information is available, the multiplier default is set to 1. Precipitation is partitioned between snow and rain based on a linear transition from snow to rain defined by $T_{\text{min\_rain}} = -2$ °C and $T_{\text{max\_snow}} = 2$ °C. Air temperature and dewpoint are adjusted for elevation according to specified lapse rates (0.0064 and 0.0015 °C/m, respectively), and variation in meteorological parameters according to topography follows MT-CLIM (Running *et al*., 1987). For a detailed evaluation of the elevation effects of warming on transpiration as represented by the RHESSys model, see Christensen *et al*. (2008). A complete description of RHESSys implementation and calibration for Sagehen Creek and Upper Merced can be found in Tague and Grant (2009).

We compare modelled daily streamflow with the historical record for the Upper Merced and Sagehen. The Upper Merced simulations have a Nash–Sutcliffe efficiency of 0.58 (and an R^2 of 0.80 for log-transformed flow) over the 43-year record, whereas the Sagehen Creek simulations have a Nash–Sutcliffe efficiency of 0.48 (and an R^2 of 0.57 for log-transformed flow) over a 41-year record, with a bias toward overpredicting annual streamflow amounts. Although these results suggest that the model could still be improved, the major deviations between model and data occur during high flows. The low-flow behaviour is robust (seasonally windowed Nash–Sutcliffe efficiencies up to ~0.95), and therefore, we believe the model results are useful for the purposes of this paper.

We model the sensitivity of each catchment's response to warming by comparing a range of scenario simulations: (1) a no-forcing base case with the historic temperature and precipitation regimes, (2) two warming cases (with temperature increases of 2 and 4 °C) in which precipitation is partitioned between snow and rain based on temperature, but no evapotranspiration changes or increased melt rates due to warming are permitted, and (3) two warming cases (with the same temperature increases of 2 and 4 °C) in which precipitation is partitioned between snow and rain based on temperature, and snowmelt and evapotranspiration also depend on temperature. These cases encompass the range of temperature increases that are expected in the region over the coming century (e.g. Hayhoe *et al.*, 2004) but are far less complex than a downscaled global climate model for the region. We deliberately excluded possible changes in the amount of precipitation in order to focus on quantifying the effects of changing the temporal distribution of melt and rainfall. The same precipitation record is used in all modelling efforts for a particular catchment (i.e. no change in total precipitation), but because RHESSys partitions precipitation between snow and rain based on the temperature at each location within the catchment, the proportions of snow *versus* rain differ depending on the modelled temperature. Scaling of transpiration to stand and watershed scales is based on leaf area index, which is estimated from the Normalized Difference Vegetation Index, derived from summer Thematic Mapper Remote Sensing imagery at each site (using the approach outlined by White *et al.*, 1997). Although shifts in vegetation patterns may be important in shaping catchments' responses to climate change (Alo and Wang, 2008), little information is available to constrain the pace and pattern of possible vegetation shifts at our sites, and we did not include them in our modelling efforts.

We also focus here on subsurface storage changes because as aboveground storage in snowpacks decreases, the ability of the subsurface to store and release water will grow in importance. Net changes in storage (excluding storage in the snowpack itself) are calculated using a mass balance approach where

$$\text{net change in storage} = \text{melt} + \text{rain–evapotranspiration–streamflow} \tag{2}$$

Melt is assumed to be equal to the decrease in measured daily SWE on days when SWE decreases, which will slightly underestimate total melt in these catchments because snow falling and melting within the same day will instead be categorized as rain. Evapotranspiration also includes canopy interception losses in which precipitation falls to the canopy and evaporates without ever reaching the ground. Transpiration varies in all the model runs based on the availability of soil water and potential evapotranspiration, which is a function of temperature and other factors. To evaluate the sensitivity to changes in potential evapotranspiration, potential transpiration remains the same as in the base case in case (2) and is allowed to vary with temperature in case (3). Actual transpiration may decrease if insufficient water remains available, or it may increase if it is not limited by water availability. The 2 and 4 °C warming cases yield qualitatively similar results for cases (2) and (3), differing only in magnitude, so for clarity, we display only the base and 4 °C results.

Modelled snowpack–flow relationships

We first test whether the historical relationships between peak SWE and subsequent low flows seen in Figure 3 are consistent with model simulations of future climates in which more precipitation falls as rain. In particular, we explore whether the decrease in snowpack storage, and the shift in timing of recharge due to the decreased importance of melt-out, affects low flow responses.

According to our modelling results, the historical relationships provide insight into the future hydrology of Sierran streams, although the uncertainty in predicted flow responses increases as snowpacks decrease. The relationships between peak SWE and low flows are similar in the base and warming scenarios for the Upper Merced (Figure 7a), but with warmer temperatures, the data points typically occupy a smaller region in the lower left corner of the low flow/peak SWE plot. At Sagehen, when peak SWE is less than about 50% of the base-case average, low flows are more variable in the warming scenario than in the average current climate (Figure 7b). The best-fit low-flow/snowpack slopes for the warming case with vegetation response remain within two standard errors of the base case at both sites (Figure 7). Thus, we infer that the historical relationships may indicate future low-flow responses to changes in snowpack, albeit with less certainty in flow responses at very low peak SWE.

Shifts in timing of the low-flow period are consistent between the historical record and the RHESSys model results presented here at the Upper Merced, but not at Sagehen. The historical records indicate that the mean day of the low-flow period advances by ~3–7 days for each 10% decrease in maximum SWE (Figure 5b). Peak SWE decreases by an average of 35% between the modelled base case and the 4 °C warming case with vegetation response at the Upper Merced, leading to a modelled average timing shift of ~20 days, within the range of ~10–24 days expected by the historical record. At Sagehen, the low-flow timing shifts by just 2 days from the base case to the 4 °C warming case with vegetation response, in which average peak SWE decreases by 77% from the base case (Table II). This small 2-day timing shift is much less than the ~6- to 22-day shift predicted by the historical relationship based on this large loss in SWE (Figure 5b); the difference may indicate that when the peak snowpack volume drops below a certain threshold, rainfall rather than snowpack dynamics control stream low flows. If most Sierra Nevada basins behave like the Upper Merced rather than Sagehen, and peak SWE decreases by ~70% over the next century (Leung and Wigmosta, 1999), the middle of the low-flow period should advance by ~20–50 days. However, if basins throughout the Sierra Nevada behave more like Sagehen than like the Upper Merced, a much smaller change in the timing of low flows would be expected.

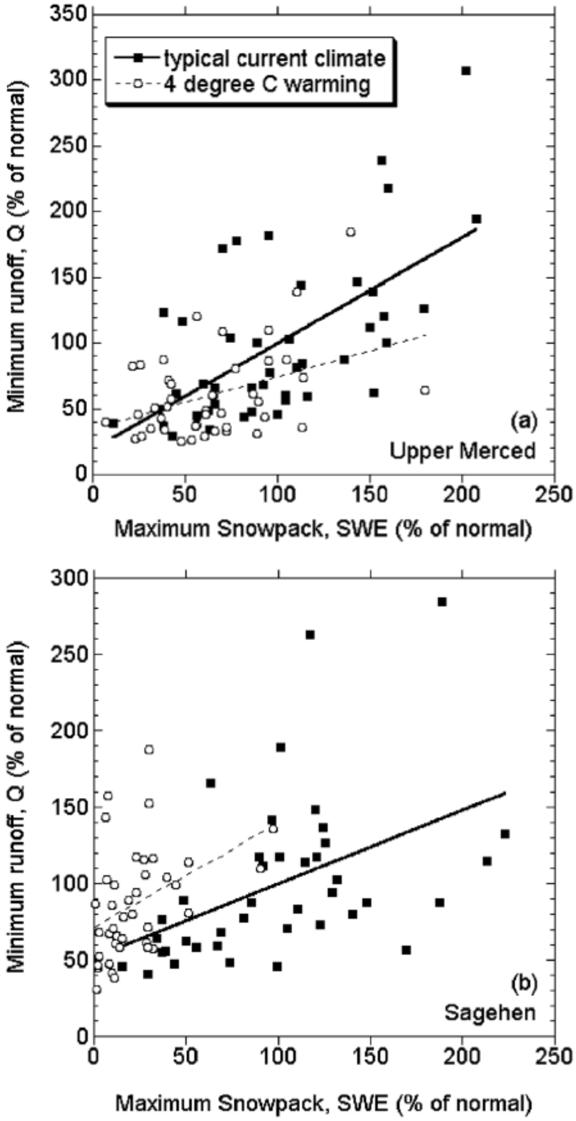

Figure 7. Relative minimum runoff (*Q*) *versus* relative maximum snowpack (SWE) in RHESSys model results for the (a) Upper Merced and (b) Sagehen catchments under the base case of no warming (black squares) and under the 4 °C warming case (with changes in vegetation and melt response, open circles). Slopes are significant in all cases ($p < 0.009$ or lower), but differences in the slopes are not statistically significant. Note that the warming case exhibits a similar trend to the base case for the Upper Merced but is limited to a smaller range of flows, telescoping down to the lower left quadrant of the plots. The warming case at Sagehen is also in the lower left quadrant, but low flows exhibit more variability when snowpacks are small than in the base case

Modelled water balance and warming-induced changes in precipitation phase change, evapotranspiration, or melt

Here, we systematically compare the effects of changes in precipitation phase *versus* changes in melt rates and evapotranspiration on key components of the catchment water balance to identify controls on catchment warming response (Figure 8). We look at both average annual and seasonal changes. We define the seasons as four approximately equal (91–92 day) periods. For convenience, we refer to them by their approximate seasonal names (spring = Julian days 61 to 152, summer = Julian days 153 to 245, autumn/fall = Julian days 246 to 334, and winter = Julian days 335 to 60). We compare the 4 °C warming case in which only precipitation phase changes in response to warming (case 2, Figure 8, dotted lines) to the 4 °C warming results (case 3, Figure 8, solid lines) in which all hydrologic processes (precipitation phase, melt rates, and evapotranspiration) respond to warmer temperatures. We discuss how changes in precipitation phase, melt timing, and evapotranspiration affect both sites and highlight the differences between the sites.

Snow, rain, and flow. Unsurprisingly, more precipitation falls as rain in the warming scenarios than in the base case (Figure 8a and f, Table II) regardless of whether melt and transpiration rates change. Peak snowpack water content also drops (Figure 8b and g, Table II). In both warming cases, melt-out occurs earlier in the year compared with the base case (Figure 8b and g, Table II). Low flows decrease and occur slightly earlier in the year (Figure 8c and h, Table II).

Warming-induced changes in the phase of precipitation can be distinguished from changes due to warming effects on melt rates and evapotranspiration demand (case 2 *vs* 3, as outlined earlier). Precipitation phase change dominates the decrease in peak SWE (Figure 8b and g, Table II) but has a minor effect on the timing of melt-out. In contrast, temperature-driven melt rates have a dominant effect on the timing of melt-out, but a minor impact on peak SWE (Figure 8b and g, Table II). The shift in low-flow timing is due to the change in precipitation phase (Figure 9a, Table II); no additional shift is observed when melt or evapotranspiration rates change with temperature.

The models for the two sites do not respond identically to warmer temperatures. Shifts in the phase of precipitation from snow to rain, in the absence of changes in the total amount of precipitation, melt rates, or evapotranspiration, affect the timing and magnitude of flow much more at Sagehen than at the Upper Merced (Figure 8c and h). At Sagehen, low flows decrease by ~10% with a change only in precipitation phase and decrease by a smaller amount if melt and evapotranspiration rates also respond to warmer temperatures. At the Upper Merced, most of the change in the timing and magnitude of low flows is due to changing melt and evapotranspiration rates, and only a small amount is attributable to changing the phase of precipitation (Figure 9a, Table II). This is likely because the Upper Merced basin is generally higher in elevation than the Sagehen Creek basin (Figure 2b).

Table II. Summary of changes observed in different components of the water budgets modelled by RHESSys for the base current climate case (case 1), a 4 °C warming case that includes only changes in the phase of precipitation due to warming (case 2) and a 4 °C warming case that also includes melt and evapotranspiration changes resulting from warmer temperatures (case 3) for two sites in the Sierra Nevada mountains

	Sagehen			Upper Merced		
	Case 1	Case 2	Case 3	Case 1	Case 2	Case 3
Annual average streamflow [mm/year]	317	321	274	799	815	815
Average annual rainfall [mm/year][a]	562	859	862	278	481	505
Average annual evapotranspiration [mm/yr]	693	691	738	314	299	299
Excess flows into storage[b] [mm/year]	585	694	734	195	228	245
Annual average 15-day running median peak SWE [mm]	392	133	88	729	562	473
Annual average 15-day running median low flow [mm/day]	0.20	0.18	0.17	0.11	0.10	0.07
Std dev peak SWE [mm]	199	101	84	344	279	253
Std dev low flow [mm/day]	0.11	0.08	0.07	0.07	0.07	0.04
Winter net change in storage (NCIS)[c] [mm/season]	112	232	276	7	8	18
Spring NCIS[c] [mm/season]	167	-5	-85	53	29	6
Summer NCIS[c] [mm/season]	-297	-297	-272	-83	-81	-75
Fall NCIS[c] [mm/season]	23	74	86	23	45	52
Winter evapotranspiration (ET)[c,d] [mm/season]	37	40	64	34	34	39
Spring ET[c,d] [mm/season]	196	201	256	110	108	131
Summer ET[c,d] [mm/season]	321	313	292	134	120	88
Fall ET[c,d] [mm/season]	123	119	116	38	38	43
Std dev winter NCIS [mm/day]	1.3	1.9	2.0	0.2	0.3	0.4
Std dev spring NCIS [mm/day]	1.3	1.9	2.3	0.3	0.3	0.5
Std dev summer NCIS [mm/day]	0.8	0.7	0.8	0.4	0.4	0.5
Std dev fall NCIS [mm/day]	1.9	2.4	2.4	0.6	0.8	0.9
Timing [Julian day]						
Melt-out	191	163	125	214	209	184
Peak 15-day running median flow	148	50	75	144	144	119
Minimum 15-day running median flow	284	282	282	279	259	259
Peak 15-day running median net change in storage	121	321	321	138	139	84
Peak 15-day running median evapotranspiration rate	175	170	156	162	162	143

The lower portion of the table indicates shifts in timing in units of days from the base case. Superscripts: a = differences in total annual rainfall between the two warming cases are due to variations in melting of the snowpack on days with precipitation, b = excess inflows into storage are defined as daily inflows to storage (melt + rainfall) that exceed daily outflows (ET + streamflow), c = seasons are defined by Julian day (see text), and d = 15-day running median ET.
NCIS, net change in storage; ET, evapotranspiration; SWE, snow water equivalent.

Evapotranspiration response. Warming also shifts the timing of evapotranspiration demand, which can, in turn, alter low flows. Unsurprisingly, the timing of peak evapotranspiration shifts noticeably only when we permit evapotranspiration rates to respond to warming (Figure 8d and i, Table II). Total evapotranspiration over the period of record changes little (~1–6%) from the base case to either warming case (Figure 9b, Table II), but substantial seasonal changes do occur (Figures 8d and i, and 9b). When evapotranspiration rates respond to warming, spring evapotranspiration at both sites is higher than in the base case (Figure 9b, dark grey bars), and summer evapotranspiration is lower (solid black line in Figure 8d and i, light grey bars in Figure 9b, Table II). Peak 15-day running median evapotranspiration also occurs ~3–4 weeks earlier in an average year. At both sites, the shift in evapotranspiration timing allows low flows to remain higher than they would be if the growing season were to lengthen and total evapotranspiration were to increase.

The coincident shift in timing of evapotranspiration and low flows suggests that it is important to understand when light or water may limit transpiration and under which conditions the growing season might lengthen. Historically, the timing of peak flows is approximately in phase with peak evapotranspiration and radiative fluxes. As warming occurs, peak evapotranspiration and flows shift earlier in the year and are increasingly out of phase with peak light availability. For some plants, light limitations may affect transpiration rates and net primary productivity. Studies of energy limitations along elevation gradients have suggested that the mid-elevation bands may be the most sensitive to warming because of water limitations (e.g. Christensen *et al*., 2008; Tague *et al*., 2009; Lundquist and Loheide, 2011; Trujillo *et al*., 2012).

Storage response changes. The temporal distribution of flows into and out of subsurface storage (Eq. 2) is affected by changes in precipitation phase, evapotranspiration, and melt because of warmer temperatures. The timing and variability of net changes in storage shift within the year (Figures 8e and j, and 9a and c, Table II). In all cases, there is a large decrease in spring net changes in storage, shifting the regime from a positive change in net storage to a near-zero or negative change in net storage (dark grey bars in Figure 9c). The losses from storage in spring are partially compensated by increases in fall and winter because of precipitation phase change (black and white bars in Figure 9c, Table II). These net changes in storage effectively sustain early season evapotranspiration and low flows by contributing subsurface water to surface flows during the spring, partially replacing meltwater from the smaller snowpacks.

Past studies have shown that precipitation falling as rain is less effective than snowmelt at recharging groundwater stores (Winograd *et al*., 1998; Earman *et al*., 2006). We observe that net changes in storage increase by up to 26% in the warming cases as the fraction of rain increases. The discrepancy between our model results and this previous work may be partially explained by the differences in the studied catchments and in the timing of the rainfall. For example, Winograd *et al*. (1998) found that summer rain in Nevada, which often falls in intense storms when potential evapotranspiration is very high, contributes proportionally less to groundwater storage than does snowmelt in winter and spring. Lower-intensity winter rain in the Sierra Nevada likely contributes more to subsurface storage than summer rain does. (RHESSys also assumes that rain falls evenly throughout the day unless sub-daily duration time series are available, which may lead to overestimates of net changes in storage during rainfall in all model runs. However, we do not expect that rainfall intensities would regularly be high enough to exceed the infiltration capacity at the study sites, and therefore, we expect that RHESSys's assumption of evenly distributed rainfall introduces minimal error in the net change in storage estimates.) Thus, more frequent, low-intensity, cool-season rainfall can lead to increased fluxes into and out of subsurface storage.

Because the timing of fluxes into and out of subsurface storage influences low flows, we also examined differences in timing of subsurface fluxes at Sagehen and the Upper Merced. When only the phase of precipitation changes in response to warming, spring and winter net changes in storage decrease by a larger amount at Sagehen than at the Upper Merced (Figure 9c, Table II). Peak storage fluxes occur earlier in the year during late autumn at Sagehen in all warming cases (Figure 9a, Table II) and shift by approximately 2 months at the Upper Merced only when melt and evapotranspiration rates respond to warming. There is a 1-day shift in the timing of peak net changes in storage at the Upper Merced when only the phase of precipitation changes (Table II, Figure 9a). Warming increases the variability of fluxes into and out of storage. For example, the seasonal timing shifts noted earlier are larger at Sagehen in part because a larger fraction of the total annual precipitation falls as rain under either warming case. Seasonal standard deviations in net changes in storage increase because of warming by up to ~50% in fall and winter with only phase changes permitted and can as much as double when melt and transpiration rates also respond to warmer temperatures (Table II). Because there is no long-term change in modelled groundwater net storage, if subsurface storage volumes remain constant, the increase in excess inflows and outflows

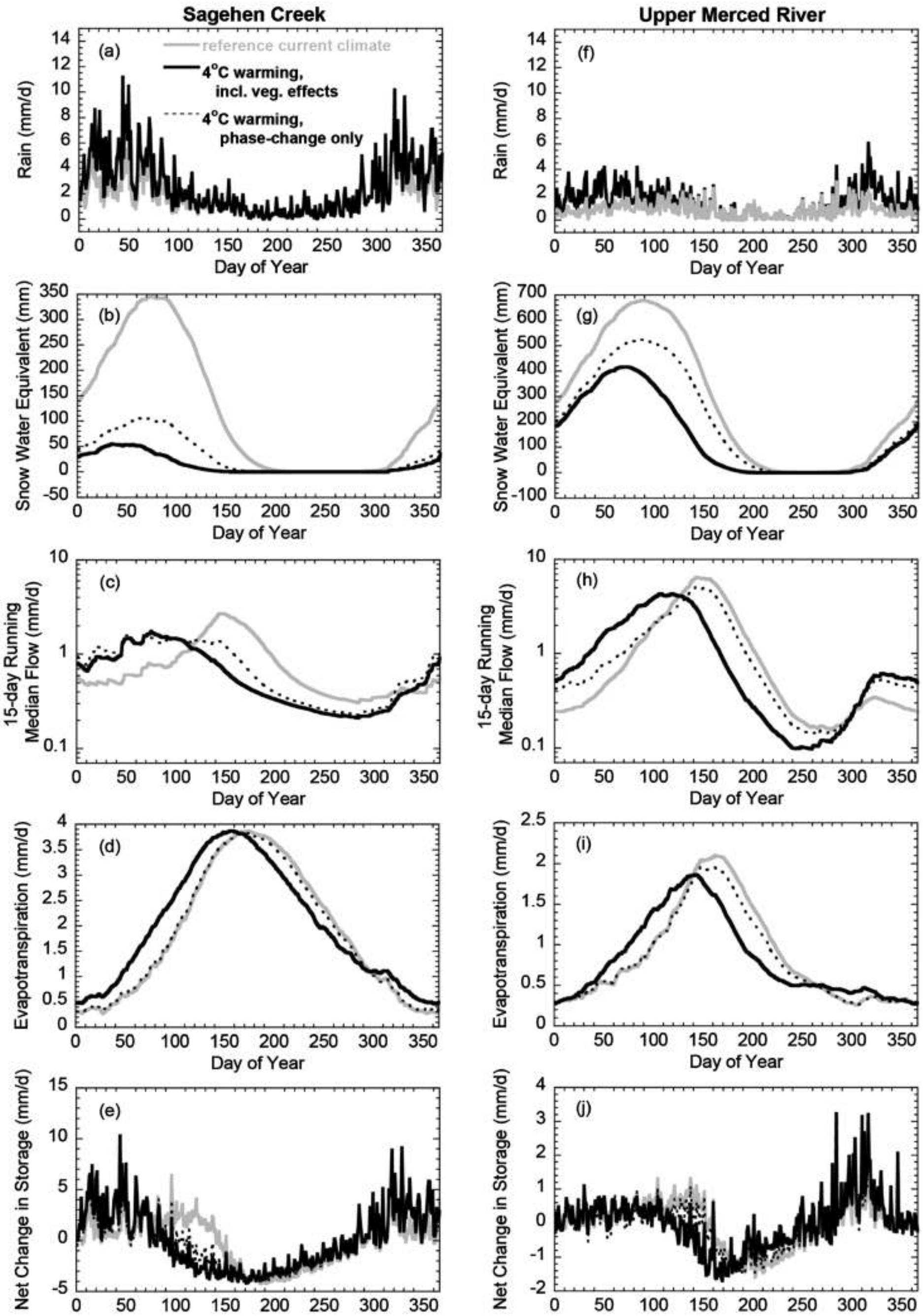

Figure 8. 4 °C warming case time series for different elements of the water budget for Sagehen Creek (left) and the Upper Merced River (right). In all panels, the solid black line reflects responses to 4 °C warming including evapotranspiration demand responses, whereas the black dotted line indicates 4 °C warming with no shift in evapotranspiration demand. Grey lines are the no-forcing case, representing the reference or current climate conditions. Shown are mean values for the given day of year across the entire record for (a and f) rainfall (mm/d), (b and g) 15-day running median snow water equivalent,(SWE) (mm), (c and h) 15-day running median flow (mm/day), note logarithmic scale on *y*-axis, (d and i) 15-day running median evapotranspiration (mm/day), and (e and j) net change in storage (mm/day)

will lead to shorter and more variable residence times of water in a warmer climate.

Sagehen and the Upper Merced basins have different recession characteristics (Tague and Grant, 2009); Sagehen tends to be more groundwater-dominated (Rademacher *et al*., 2005) than the Upper Merced (Conklin and Liu, 2008). Thus, a similar ~10% drop in low flows at both sites when precipitation phase change only is considered may seem surprising (Figure 8c and h, Table II). However, both drainage rates and the timing of melt influence low-flow responses to changes in peak SWE. Low flows are sustained in Sagehen by larger

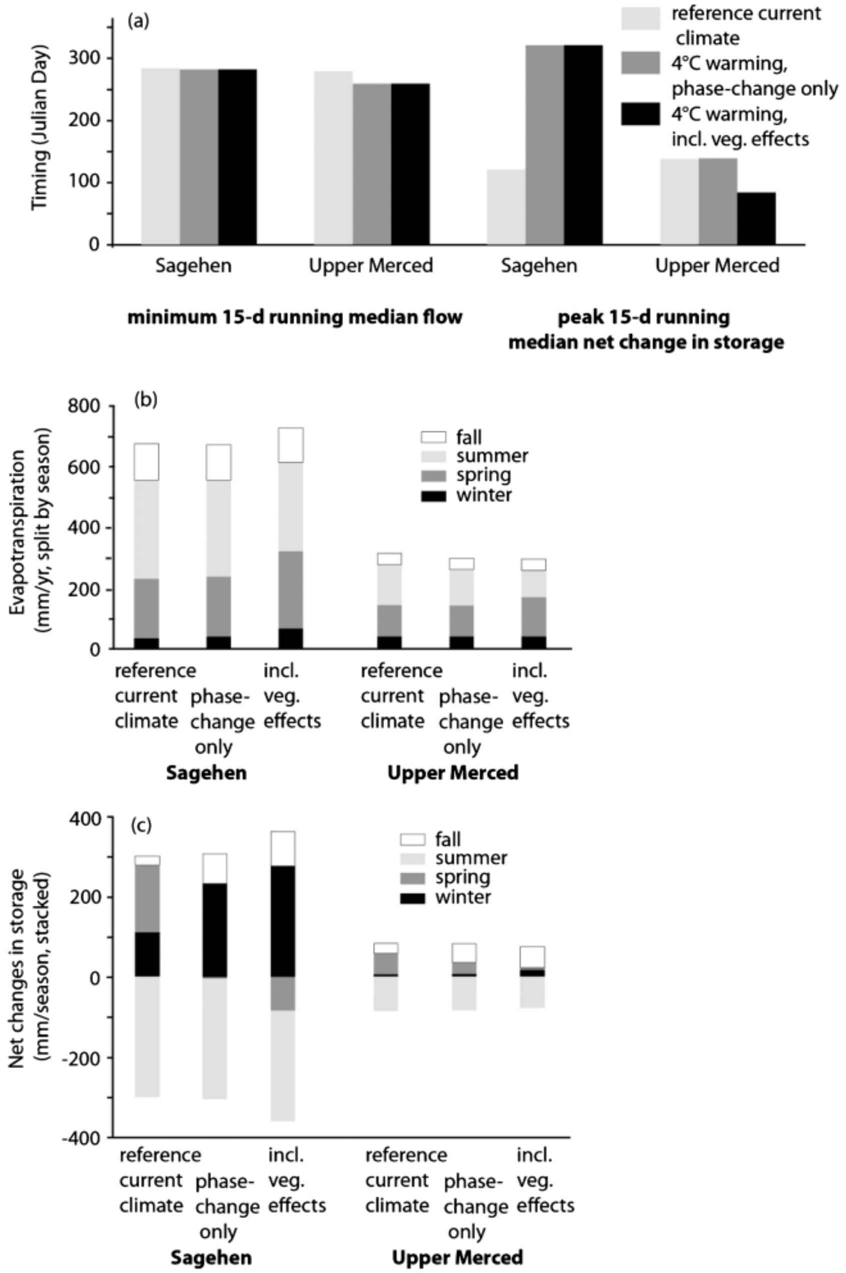

Figure 9. Changes observed in the (a) timing or (b and c) magnitude of different components of the water budgets modelled by RHESSys for the base case of the average current climate (case 1), a 4 °C warming applied uniformly to all years that includes only changes in the phase of precipitation due to warming (case 2) and a 4 °C warming applied uniformly to all years that also includes melt and evapotranspiration changes resulting from warmer temperatures (case 3), for two sites in the Sierra Nevada mountains, Sagehen Creek and the Upper Merced River. Net changes in storage are defined as the difference between inflows (melt + rainfall) and outflows (evapotranspiration + streamflow). Seasons are defined by Julian day (see text for details)

groundwater stores. Low flows in the Upper Merced are reached following the recession from peak flow during melt-out so that the small shift in melt-out timing relative to the base case leads to a small drop in low flows. When the full vegetation response to warming is considered, low flows decease more at the Upper Merced than at Sagehen. That is, the surface-water-dominated site responds more strongly than the groundwater-dominated site to vegetation water use changes. Jefferson *et al.* (2008) found the opposite result; they found that in a Cascade stream with a large groundwater component, low flows decrease in response to decreasing precipitation by more than low flows in a nearby stream with a smaller groundwater component. Our work differs from Jefferson *et al.* (2008) because we compare changes in the phase of precipitation and do not alter the total amount of precipitation. In addition, differences in the permeability and thickness of the volcanic bedrock underlying the two regions may influence stream flow response. Thus, subsurface characteristics as well as precipitation magnitude, timing, and phase changes all influence low-flow response.

Brief summary of catchment responses to warming and future work. Changes in the magnitude of low flows are due to a combination of changes in precipitation phase, melt, subsurface storage, and the timing of evapotranspiration losses. Modelled changes in low-flow timing occur only because of changes in the phase of precipitation (Figure 9a, Table II). Net changes in storage and shifts in the timing of melt-out and peak evapotranspiration are due to precipitation phase change as well as melt and vegetation responses to warming (Figures 8 and 9).

Responses at Sagehen Creek and the Upper Merced River may represent other catchments throughout the Sierra Nevada mountains. Approximately 46% of the Sierra Nevada is underlain by volcanic lithologies, whereas the balance is underlain primarily by granitic rocks. Thus, we may expect that low-elevation to mid-elevation catchments underlain by volcanic rocks, similar to Sagehen, may experience a shift in timing of storage fluxes into and out of the subsurface as more rain falls instead of snow, and that seasonal evapotranspiration and net changes in storage will shift at these sites, but that low flows may not change very much. On the other hand, higher-elevation sites underlain by granitic rocks, similar to the Upper Merced, may experience a larger decrease in low flows compared with the Sagehen-like catchments, as well as a decrease in evapotranspiration and smaller net changes in storage. Further warming would presumably lead to larger changes at high-elevation sites because the modelled 4 °C increase in temperatures does not lead to rain-dominated regimes.

Future work focusing on *in situ* measurements of evapotranspiration rates and water table depths across the rain–snow line, especially if coupled with tracer experiments to determine the depth of water sources, would provide crucial mechanistic testing of the model results. Understanding whether the modelled shift in the temporal distribution of groundwater storage applies to a range of groundwater-dominated catchments is also an important future project. Furthermore, we did not examine how changes in growth phenology, frost damage, vegetation die-off, and species composition could affect evapotranspiration rates; these changes could produce large effects not accounted for in our model. Monitoring studies, especially those conducted near the snow–rain boundary, should measure precipitation and unimpaired streamflows to understand how low flows in these systems respond to shifts in precipitation and temperature. Some streams' low flows will be very sensitive to such changes, whereas other streams will be more robust to changes in climate.

CONCLUSION

Low flows are important to human and ecological systems. We present historical data demonstrating that changes in snowpack volume affect subsequent summer and fall low flows in the Sierra Nevada of California. At all eight of our study catchments, summer and autumn low flows are strongly correlated with annual peak SWE, and in six of the eight catchments, low flows vary more-than-proportionally with variations in SWE from year to year. In these six catchments, linear extrapolations of the historical low-flow/SWE relationships suggest that low flows could drop to zero if peak SWE decreases by roughly 50% from historical norms. At two sites (Pitman Creek and Sagehen Creek), low flows depend on both the current year's snowpack and the previous year's snowpack. At these sites, streamflow is more sensitive to the current year's snowpack in years for which the previous year's snowpack was above average.

RHESSys model results at two sites (Sagehen Creek and the Upper Merced River) indicate that when air temperatures increase by 2 to 4°, the more-than-proportional relationship between maximum SWE and low flows still holds; however, the modelled future warming peak SWE/low flow relationship is more variable than the historical relationship, indicating that increasing rainfall may alter the hydrological processes in these watersheds. At both modelled sites, increasing temperatures that shift the phase of precipitation from snow to rain increase positive net changes in storage in autumn (and in winter at Sagehen). This phase change alters the temporal distribution of net changes in storage that in turn affect low flows. Sites with more volcanic geology and greater potential for subsurface storage may buffer the loss of snowpack more easily than sites with fewer volcanics and shallower soils, but the buffering capacity appears to be limited. Total evapotranspiration changes relatively little in the modelled warming cases, but the timing of peak evapotranspiration shifts slightly earlier to better coincide with the earlier peak water availability in a warmer climate. Our modelling results suggest that critical gaps in our knowledge include understanding the timing of groundwater recharge and melt, the spatiotemporal distribution of rain and snow, and the response of vegetation to warming in mid-elevation catchments. All of these aspects will affect the resilience of catchment low flows to climate warming.

ACKNOWLEDGEMENTS

Funding from the Berkeley Water Center, the National Science Foundation Graduate Research Fellowship Program, and the NSF Idaho EPSCoR Program under award number EPS-0814387 supported this research. The US

Geological Survey, Natural Resources Conservation Service, and the California Department of Water Resources provided essential data.

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