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Mountain runoff vulnerability to increased evapotranspiration with vegetation expansion

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Climate change has the potential to reduce surface-water supply by expanding the activity, density, or coverage of upland vegetation, although the likelihood and severity of this effect are poorly known. We quantified the extent to which vegetation and evapotranspiration (ET) are presently cold-limited in California’s upper Kings River basin and used a space-for-time substitution to calculate the sensitivity of riverflow to vegetation expansion. We found that runoff is highly sensitive to vegetation migration; warming projected for 2100 could increase average basin-wide ET by 28% and decrease riverflow by 26%. Kings River basin ET currently peaks at mid-elevation and declines at higher elevation, creating a cold-limited zone above 2,400 m that is disproportionately important for runoff generation. Climate projections for 2085–2100 indicate as much as 4.1 °C warming in California’s Sierra Nevada, which would expand high rates of ET 700-m upslope if vegetation maintains its current correlation with temperature. Moreover, we observed that the relationship between basin-wide ET and temperature is similar across the entire western slope of California’s Sierra Nevada, implying that the risk of increasing montane ET with warming is widespread.

Roughly 4 billion people globally and 20 million people in the state of California rely on mountain runoff for freshwater, and there is growing concern these water resources will prove vulnerable to climate change (1–6). River flow (Q) is a function of precipitation (P) minus evapotranspiration (ET) (P–ET); increased montane ET with warming, either because of the direct effect of temperature on evaporative demand or the indirect effect of warming on vegetation density and distribution, would reduce Q (5, 7–9). However, hydrologic model projections for California’s Sierra Nevada have discounted this possibility, indicating little or no effect of warming on annual ET (10–13). This result appears linked to two model assumptions: (i) models have often assumed the properties of montane vegetation will remain static, and (ii) models have often implicitly assumed that current annual montane ET is almost entirely limited by water availability and that warming will simply hasten the beginning and end of the growing season.

Recent evidence calls both of these assumptions into question. Widespread increases in subalpine tree growth, tree-line altitude, and species distribution with elevation have been reported with recent climate trends in California and elsewhere, implying that rapid vegetation shifts are possible (14–16). Time series of Sierra Nevada forest greenness indicate a transition from water limitation at low elevation to cold limitation at high altitude, implying that upper elevation ET is sensitive to warming (17). Nonetheless, the extent to which annual montane ET is currently temperature-limited, as well as the sensitivity of large-scale ET to vegetation redistribution, remain largely unquantified.

We used the upper Kings River basin in California’s Sierra Nevada as a case study of the sensitivity of runoff to increased ET with warming. We found that Kings River flow is highly sensitive to vegetation expansion; warming projected for 2100 could increase ET across the Kings River watershed by 28% and decrease riverflow by 26%. Moreover, we found a consistent relationship between watershed ET and temperature across the Sierra Nevada; this consistency implies a potential widespread reduction in water supply with warming, with important implications for California’s economy and environment.


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The P–ET increase at high altitude is mainly (∼80%) attributable to reduced ET (Fig. 1A). We quantified the limitations imposed by cold and moisture stress on eddy covariance gross primary production (GPP) (21) by analyzing the seasonal patterns of CO2 uptake. GPP at 405 m was limited strongly by summer moisture stress; GPP at 1,160 and 2,015 m was limited by neither moisture stress nor cold; GPP at 2,700 m was limited strongly by winter cold (Fig. 2A). ET is well correlated with GPP across the elevation gradient (21), and the two fluxes are mechanistically linked through leaf gas exchange and plant phenology, implying that the same processes determine the altitudinal patterns ET. An analysis of the spatial correlation between remotely sensed ET and climate yielded a similar result. ET was systematically lower at locations with colder temperatures and less precipitation (Fig. S4); the combined effect of temperature and precipitation explained 66% of the ET variation across the basin (Figs. S2 and S5). The temperature and precipitation regressions were used to further partition the controls on ET with elevation. Moisture stress limitation decreased with increasing elevation to ∼1,000 m; the midelevation zone was relatively unaffected by moisture or cold; cold limitation increased with elevation above ∼2,000 m (17) (Fig. 2B and Fig. S6).

We examined output from the ensemble runs of the Community Climate System Model version 4 (CCSM-4) prepared using the representative concentration pathways (RCP) and historic experiments in the Coupled Model Intercomparison Project Phase 5 (CMIP-5). The mean 2085–2100 temperature increase in the atmosphere’s lower 4 km above central California ranged from 1.3 °C for RCP 2.6 to 4.1 °C for RCP 8.5 (Fig. S7).

We used the ET regressions against temperature and precipitation (Fig. S4) to estimate the effect of 2085–2100 warming (Figs. S7–S9) on basin water balance. ET below ∼2,000 m was unaffected by warming alone; ET above ∼2,000 m increased in proportion to warming (Fig. 3). The low emission RCP 2.6 expanded ET 200-m upslope, which increased basin ET by 10% (Fig. 4A). The high emission RCP 8.5 expanded ET 700-m upslope, which increased basin ET by 28%. RCP 2.6 with constant P decreased P–ET by 9%; RCP 8.5 with constant P decreased P–ET by 26%.

Precipitation projections for 2100 remain uncertain, with considerable model-to-model and run-to-run variability. Previous analyses have indicated future drying in the southwestern United States (6), but some of the CCSM-4 CMIP-5 ensemble runs indicate a wetter Sierran climate. A recent hydrologic assessment estimated an ~5% mean precipitation decline for the region (22), and we adopted this value for comparison. A 5% P reduction alone decreased total-basin P–ET by 8% (Fig. 4A). A 5% P decrease and RCP 2.6 warming decreased P–ET by 17%. A 5% P decrease and RCP 8.5 warming decreased P–ET by 33%.

We tested our analysis by comparing the ET for 11 major rivers on the western slope of the Sierra Nevada against the corresponding mean temperature. ET was estimated for 1981–2010 by subtracting the observed full natural river flow (Q) for each basin from the corresponding spatially integrated P. The basin mean elevations ranged from 1,210 to 2,330 m, and mean daily maximum temperatures from 18.7 °C to 12.8 °C. Basin P – Q was well correlated with temperature (Fig. 4B) (R2 = 0.716), with a sensitivity of 44.6 mm °C−1; this sensitivity is consistent with that derived in a completely independent way for the Kings River basin (Fig. 4A) (31.8 mm °C−1).

Discussion and Conclusions

We draw three conclusions. First, Sierran ET peaks at mid-elevation and declines above ∼2,000 m; this result is supported by the eddy covariance and remote-sensing observations (Fig. 1A), and also by P – Q comparisons within (19, 23) and between river basins (Fig. 4B). Second, reduced ET at higher elevation...
The second issue is less obvious and involves the effects of moisture stress on phenology. Previous analyses of Sierran ET have often assumed montane tree roots are restricted to the shallow surface soil and forests have a short growing season constrained by moisture access in summer (28). This assumption has led to predictions that warming will hasten both the onset of high rates of ET in spring and the depletion of soil moisture and decline of ET in summer, leading to a quantitative phenological offset that minimizes the effect of warming on annual ET (2, 10, 12). More recently, the importance of deep montane rooting and access to moisture in the underlying fractured bedrock has been recognized; this deep rooting buffers trees from seasonal shifts in precipitation and evaporative demand and leads to a year-round growing season at mid elevation (e.g., refs. 21, 29, and 30). In turn, forest access to large stores of belowground moisture leads to the conclusion that declining summer ET is unlikely to offset increasing winter and spring ET, and forms the basis for our focus on the annual, rather than seasonal, effects of warming on ET.

In fact, we see both of these scenarios playing out along the elevation gradient. At lower elevation, where water availability is currently limiting, we expect a phenological offset that quantitatively offsets accelerated winter ET by decreased summer ET (Figs. 2 and 3). At upper elevation, where ample P and deep rooting allow year-round moisture access, we do not expect a compensatory summer ET decline, but rather increased annual ET. The effect of warming on basin-wide ET is expected to be dominated by the higher elevations, where high P and deep
rooting reduces the likelihood of a phenological offset, leading to increased annual ET. Rooting depth and deep moisture access will therefore play central roles in mediating the impact of climate change on the Sierra Nevada, and further work is needed to improve the representation of rooting depth in models of Sierran hydrology and ecology.

Both individual plants and entire ecosystems show strong functional convergence with climate, raising the possibility of widespread upward ET expansions with warming. For example, the inverse relationship between thermal and water limitation with elevation (Fig. 2) has been previously described in temperate semiarid regions (28); consistent relationships between elevation and properties, such as biodiversity, are widely recognized (31); and a similar relationship between temperature and ET holds across the entire western slope of the Sierra Nevada (Fig. 4B). However, this does not mean the discharge from all rivers will prove equally vulnerable to warming. The Kings River basin is comparatively cold as a result of a high mean altitude (Fig. 4A), and runoff from warmer basins, where the current mean ET is less limited by winter cold, may prove less sensitive to warming.

Our analysis relies on a simple empirical approach, whereas it is likely that models with much more mechanistic detail will ultimately provide the most reliable hydrologic forecasts, especially for novel conditions and locations. Nonetheless, our approach is well justified, given current understanding as well as inherent advantages offered by empirical strategies. Uncertainty over phenomenon, such as the effect of rooting depth on hydrology and montane plant phenology, currently limits the development of process-based models. Moreover, inputs at fine scales are often inadequate to drive detailed models of montane hydrology. Additionally, simple approaches may offer inherent advantages over mechanistic strategies. For example, emergent ecosystem properties associated with resource optimization (32) strengthen the correlation between remotely sensed indices and ET beyond that based solely on the biophysical controls on ET. Hence, our spatial extrapolation of annual ET was ultimately founded on both the effect of Leaf Area Index on ET, and the feedback of annual ET and GPP to Leaf Area Index (Fig. S1). This bi-directional relationship allowed estimates of ET that were superior to those based on a more explicit consideration of the unidirectional biophysical controls on ET (21). There is widespread agreement that mechanistic models will ultimately outperform simpler approaches, but there is little evidence that understanding of Sierran hydrology has reached the point where this is the case.

The simplicity of our approach allows us to clearly identify the underlying assumptions and limitations; several additional caveats require emphasis. Our study relied on a space-for-time approach, and assumed that climate is the main controller of the current ET distribution (Fig. S4). Our analysis will overstate the impact of climate on ET if nonclimatic factors that covary with elevation help explain the spatial patterns of ET. Additionally, we did not consider mechanisms that may mitigate vulnerability, including lags in vegetation and ecosystem-type migration (33), the effect of rising atmospheric CO2 on hydrology (34), and the possible use of forest management to suppress ET (35). For example, upstroke vegetation migration may be delayed by dispersal, establishment, or edaphic conditions (36). The lack of deeply weathered regolith at higher elevations may be especially important, possibly slowing or preventing the upward movement of vegetation with climate change and limiting the impact on river flow (37).

We view our analysis as a first step that establishes the vulnerability of montane ET and P–ET to upstroke vegetation redistribution. The outstanding question is no longer whether warming has the potential to accelerate montane ET and reduce runoff, but how rapidly canopy density, plant species composition, and regolith porosity can redistribute with climate change. Further work is needed to better quantify the risk: prognostic models that couple biogeography, geomorphology, and hydrology will ultimately be needed to forecast the impact of climate change on montane hydrology.

Materials and Methods

Ground-Based Measurements. For ground-based measurements (Figs. 1A and 2A) we installed four eddy-covariance towers at ~800 m altitude intervals in and around the upper Kings River basin (21). All sites were on granite and had vegetation that was typical for the elevation and that had not been disturbed recently. Observations from six additional towers in Southern California were used to establish a relationship between Normalized Difference Vegetation Index (NDVI) and annual ET (Fig. S1).

We analyzed the seasonal patterns of gross ecosystem exchange (GEE) to quantify the limitations imposed by summer moisture stress and winter cold. We divided each year into three elevation-dependent intervals based on meteorological conditions and fluxes. We used the GEE observations during the peak growing season to determine a best-fit rectangular hyperbola against light for periods that were neither cold nor water limited. We then ran the entire time series of observed light through the corresponding peak growing season regression to calculate a time series of GEE that would be expected in the absence of cold or moisture limitation (the unlimited GEE). We then summed unlimited GEE and also the observed GEE for the three intervals. Finally, we calculated the fractional reduction in GEE for each interval as the observed GEE divided by the unlimited GEE. We attributed the fractional GEE reduction in winter to cold limitation and the GEE reduction in late summer to water limitation.

Spatially Gridded P, ET, and P–ET. We extrapolated P, ET, and P–ET to the upper Kings River basin (21) (Figs. 1 and 2B). Elevation was taken from the Shuttle Radar Topography Mission (http://earthexplorer.usgs.gov). P was obtained for 1981–2010 (http://prism.oregonstate.edu). ET was calculated from NDVI measured by the Moderate Resolution Imaging Spectroradiometer (MODIS) MYD13Q1 Collection 5 Aqua satellite and averaged for snow- and cloud-free periods (http://daac.ornl.gov/MODIS). NDVI data were averaged for each water year, and ET was calculated based on a regression across 46 site years in 10 California ecosystems (Fig. S1). Gridded P–ET was calculated, and P, ET, and P–ET sorted into 100-m elevation bins and averaged.

We analyzed the current spatial relationships between the NDVI-based ETs and the corresponding 30-year climate normals (Fig. S4). We created a coregistered data stack of elevation, ET, and normal maximum air T (Tmax) and analyzed the relationship between ET and Tmax for all pixels that were not P-limited (pixels with P < 900 mm yr−1), and between ET and P for all pixels that were not T-limited (pixels with Tmax > 12 °C). We fit separate sigmoidal regressions between ET and Tmax for the non-P-limited dataset (Fig. S4A), and between ET and P for the non-T–Tmax limited dataset (Fig. S4B).

We used the sigmoidal equations to separately calculate the ETs that would be expected for each pixel based on local P and Tmax. We then calculated the ET for each pixel, P, and analyzed the relationship between ET and P for the non-P-limited (pixels with Tmax > 12 °C). We fit separate sigmoidal regressions between ET and Tmax for the non-P-limited dataset (Fig. S4A), and between ET and P for the non-T–Tmax limited dataset (Fig. S4B).

We used the sigmoidal equations to separately calculate the ETs that would be expected for each pixel based on local P and Tmax. We then calculated the ET for each pixel, P, and analyzed the relationship between ET and P for the non-P-limited dataset (Fig. S4A), and between ET and P for the non-T–Tmax limited dataset (Fig. S4B).

Climate Projections. For climate projections (Figs. 3 and 4A) we examined output from the CCSM-4 for the RCP and historic experiments in the CMIP-5. We downloaded monthly near-surface air temperature, air temperature, geopotential height, and precipitation for each of the five or six ensemble runs (www.earthsystemgrid.org/dataset/uacrg.cdg.ccsm4.c mip5.output.html). We averaged temperature across 1985–2010 and 1950–2005. We output the projections for three grid cells immediately upwind (west) of our study region. We averaged across the ensemble runs (Fig. S7) and interpolated by altitude the 2085–2100 increase in air temperature over the historical mean for each 0.002083° resolution pixel in the Kings River basin. We also examined the RCP precipitation output for grid cells in the Sierra Nevada.

We combined the RCP temperature projections for each pixel in the Kings basin with the T- and P-based sigmoidal regressions (i.e., Fig. S4). We added the projected warming to the 1981–2010 PRISM temperature for each pixel, and calculated the ET that would be expected as the minimum of the two sigmoidal regressions. We then calculated the P–ET for each pixel based on the 1981–2010 precipitation and also assuming a 5% P reduction. We binned the resulting ETs at 100-m elevation intervals and averaged, and also averaged across all pixels in the watershed.
Comparison Across River Basins. For comparison across river basins (Fig. 4B) we compared the water balance for 11 large river basins draining the western slope of the Sierra Nevada with mean basin temperature. The historic monthly full natural flow was downloaded for each river basin (http://dec.water.ca.gov/) and annual flow summed and averaged for 1981–2010. The river basins were demarcated by the US Geologic Survey 8 Digit Watershed Boundary Dataset (http://datagateway.nrcs.usda.gov). The river basin boundaries were then combined with the PRISM 1981–2010 normals and the mean Tmax, P, and elevation calculated. The observed riverflow (Q) and P were normalized by basin area and subtracted to estimate basin-average ET, which was compared with the mean basin elevation and Tmax.

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**Supporting Information**

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**SI Text**

**Strategy.** We used a regression-based space-for-time strategy to infer the possible effect of climate change on montane ET. Our analysis is subject to the weaknesses, shortcomings, and caveats inherent in the space-for-time approach (e.g., ref. 1). Our choice of strategy reflects the current state of understanding and modeling of Sierra Nevada evapotranspiration (ET), which is at the point where additional field observations are needed to guide the development of process-based models. Researchers are only beginning to understand the seasonality and phenology of montane plant activity, the effects of winter cold and summer moisture stress on montane plant activity, the relationship between elevation (climate) and ET, and the rates that ecosystem properties can change with climate variability (2–5). Process-based models that reliably simulate all these patterns are still being developed and tested, and field experiments to address these issues have not been attempted in the Sierra Nevada.

**Ground-Based Measurements (Fig. 1A).** Our ground-based measurements were described in detail previously (5, 6). We installed four eddy-covariance towers along a west to east transect at ~800-m altitude intervals in and around the upper Kings River basin. The sites were all on soil developed from granite, and had vegetation that was typical for the elevation and that had not been disturbed recently. The measurements were made from meteorological towers that extended 5–10 m above the trees. The half-hour eddy covariance fluxes of CO₂ (net ecosystem CO₂ exchange, NEE) and water vapor (ET) were calculated from observations of wind made with a sonic anemometer and CO₂ and water-vapor density made with a closed-path infraRed gas analyzer. Air temperature and other meteorological conditions were measured and averaged at half-hour intervals. Observations from six additional sites along a climate gradient in Southern California were used to establish a relationship between the Normalized Difference Vegetation Index (NDVI) and annual ET (Fig. S1).

**Water and Cold Limitation in Local (Tower-Based) Observations (Fig. 2A).** We used the seasonal patterns of gross ecosystem exchange (GEE; calculated by subtracting nocturnal whole-ecosystem respiration from NEE) to quantify the limitations imposed by summer moisture stress and winter cold on gross primary production (GPP; calculated by annually integrating GEE). We divided each year into three elevation-dependent intervals based on the half-hourly meteorological conditions and fluxes: (i) Winter: when soil moisture was abundant and temperatures were comparatively cold (for example, 10/27/08–4/24/09 at 2,015 m); (ii) Peak growing season: in spring and early summer, when soil moisture was abundant and temperatures were warm (for example, 4/24/09–8/22/09 at 2,015 m); and (iii) Late summer: when soil moisture was depleted and temperatures were warm (for example, 8/22/09–10/26/09 at 2,015 m).

We used the GEE observations during the peak growing season for each site and year to determine a best-fit rectangular hyperbola against light for periods that were neither cold- nor water-limited. We then ran the entire time series of observed light for that site and year through the corresponding peak growing season rectangular hyperbola to calculate a time series of GEE that would be expected in the absence of cold or moisture stress limitation (the unlimited GEE). We then summed unlimited GEE and also the observed GEE for the three intervals. Finally, we calculated the fractional reduction in GEE for each interval as the observed GEE divided by the unlimited GEE. We attributed the fractional GEE reduction in winter to cold limitation and the GEE reduction in late summer to water limitation.

**Spatially Gridded Precipitation, ET, and P minus ET (Fig. 1).** We extrapolated precipitation (P), ET, and P minus ET (P–ET) to the entire upper Kings River basin as described in ref. 5. The upper Kings River basin was demarcated by the US Geologic Survey 8 Digit Watershed Boundary Dataset (http://datagateway.nrcs.usda.gov); downloaded March 2013). Elevation was taken from the Shuttle Radar Topography Mission (SRTM) Final dataset (http://earthexplorer.usgs.gov; downloaded March 2013).

Gridded P was obtained for 1981–2010 at 30-arcsec (0.0083°) resolution (7) (Parameter-elevation Regressions on Independent Slopes Model, PRISM, Climate Group, Oregon State University, http://prism.oregonstate.edu, downloaded March 2013). Gridded ET was calculated from NDVI measured by the Moderate Resolution Imaging Spectroradiometer (MODIS) Aqua satellite and averaged for snow- and cloud-free periods. MODIS NDVI for the Kings basin was obtained from the Oak Ridge National Laboratory Distributed Active Archive Center for Biogeochemical Dynamics (http://daacornl.gov/MODIS; MYD13Q1 Collection 5; downloaded March 2013) in geographic coordinates at 0.002083° resolution. NDVI observations were filtered to include only the highest quality data (pixel reliability = 0) for day of year (DOY) 0–201 and DOY 265–365, and a pixel reliability of 0 or 1 for DOY 201–265. NDVI data were then averaged for each water year, and ET was calculated using a regression between annual average NDVI and annual ET across 46 site years in 10 diverse California ecosystems, including the 4 in and around the Kings (Fig. S1) [ET(mm yr⁻¹) = 10.3247 × exp(2.8599 × NDVI); R² = 0.914]. Gridded P–ET was then calculated at 0.002083° resolution by subtraction. Finally, gridded P, ET, and P–ET were sorted into 100-m elevation bins and averaged.

**Water and Cold Limitation in Basin-Wide (Remote-Sensing-Based) Observations (Fig. 2B).** We analyzed the current spatial relationships between annual ET calculated from NDVI (Fig. S2) and the corresponding 30-yr climate normals to better understand the limits on ET (Fig. S4). This approach parallels that used by Nemani et al. (8) to partition the climate controls on plant primary production globally, and makes use of simple, universally accepted relationships between climate and primary production (9) and primary production and ET (10). We created a coregistered data stack of elevation, ET, and 30-yr normal air T (Tmax) and P at 0.002083° resolution. We then analyzed the relationship between ET and Tmax for all pixels that were not P-limited (pixels with a P above 900 mm yr⁻¹) (Fig. S4D), and between ET and P for all pixels that were not T-limited (pixels with a Tmax above 12°C) (Fig. S4A). We then fit separate sigmoidal regressions between ET and Tmax for the non-P-limited dataset [ET= 792.7/1 + exp(–(Tmax – 9.36)/2.50)]; adjusted R² 0.697; n = 62,927], and ET and P for the non–T-limited dataset [ET= 677.7/1 + exp(–(P – 284.2)/242.0)]; adjusted R² 0.150; n = 45,558].

We used the resulting sigmoidal equations to separately calculate the ETs that would be expected for each pixel based on local P and Tmax climatology. We then estimated the ET for each pixel as the minimum ET from the two sigmoidal equations (Fig. S5) (9) and compared the resulting values with those from the original NDVI approach (i.e., Fig. S5 vs. Fig. S2) [NDVI based ET (mm) = modeled ET (mm) × 1.037–13.0; adjusted R² 0.665;
We then used the two regressions to flag each pixel as P-limited, $T_{\text{max}}$-limited, or unlimited, and to calculate the fractional ET limitation imposed by P or $T_{\text{max}}$. We calculated the fraction of maximum ET that was realized for each pixel by subtracting the fractional water and cold limitation from 1. Finally, we binned all of the pixels at 100-m elevation intervals and averaged the fractional limitation imposed by P or $T_{\text{max}}$.

We compared the mean NDVI as a function of elevation between warmer, more-xeric, southern-exposed slopes and cooler, more-mesic, northern-exposed slopes. We calculated the mean NDVI for each pixel using the same pixel reliability criteria used for the ET calculation. We calculated the slope and aspect of each pixel from the digital elevation model, and then excluded relatively flat locations with a slope less than 15°. We sorted the remaining pixels into locations facing south-southwest (aspect from 135° to 270°, where 360° is due north) and those facing north-northwest (315°–90°). Finally, we sorted the mean NDVI into 100-m elevation intervals and averaged the NDVI within each bin (Fig. S6).

**Climate Projections (Figs. 3 and 4A).** We examined output from the Community Climate System Model Version 4 (CCSM-4) for the representative concentration pathways (RCP) and historic experiments in the Coupled Model Intercomparison Project Phase 5 (CMIP-5). We downloaded monthly near-surface air temperature, air temperature, geopotential height, and P for each of the five or six ensemble runs (www.earthsystemgrid.org/dataset/ucar.cgd.casm4.cmp5.output.html; downloaded May through July 2012). We then extracted the data for a 10-by-10 grid-cell region centered over California, and averaged the RCP run temperatures across 2085–2100, and the historic run temperatures across 1950–2005. We then output the projections for three grid cells in California’s Central Valley that were immediately upwind (west) of our study region (cells centered on 36.2827° N, 120° W; 37.2251° N, 121.25° W; 38.1675° N, 121.25° W). Finally, we averaged across the ensemble runs (Fig. S7), and interpolated the altitude-specific (11) 2085–2100 increase in air temperature over the historical mean for each 0.002083° resolution pixel in the Kings River basin. We also examined the RCP P output for grid cells in the Sierra Nevada.

We combined the RCP temperature projections for each pixel in the Kings River basin with the T- and P-based sigmoidal regressions (i.e., Fig. S4) to estimate the effect of climate change on ET. We added the RCP-associated warming to the 1981–2010 Parameter-elevation Regressions on Independent Slopes Model (PRISM) temperature maximum normal for each pixel, and calculated the ET that would be expected as the minimum of the two sigmoidal regressions. We also calculated the P–ET for each pixel based on the 1981–2010 precipitation, and also assuming a 5% mean precipitation reduction. Finally, we binned the resulting ETs at 100-m elevation intervals and averaged, and also averaged across all pixels in the watershed.

We justify the direct extrapolation of upper-air temperatures over central California to near-surface temperature in the Kings River basin based on an examination of the relationship between the air temperatures measured at surface stations in the Sierra Nevada and the simultaneous air temperatures measured by radiosonde in coastal California. We compared the air temperature measured by radiosonde at the Vandenberg Air Force Base interpolated to 2050-m elevation with the simultaneous air temperature measured at the top of the 2015-m tower from September 2008 to July 2009 (Fig. S8). Vandenberg Air Force Base is ~260 km southwest of the tower site. The radiosonde observations were obtained from the University of Wyoming Atmospheric Soundings database (http://weather.uwyo.edu/upperair/sounding.html; downloaded July 2009).

We further justified our approach by examining the interannual 850-mb summer (June, July, and August) temperature anomalies reported for Marine Corps Air Station Miramar in the Radiosonde Atmospheric Temperature Products for Assessing Climate (RATPAC-B) dataset with the temperature anomalies reported for surface stations across the Sierra region by the California Climate Tracker website (Fig. S9). Miramar Air station is 300–800 km south of the Sierra Nevada. The RATPAC-B dataset was obtained from the National Climate Data Center (www.ncdc.noaa.gov/oa/climate/ratpac/; downloaded August 2009). The California Climate Tracker dataset was obtained from the Western Regional Climate Center (www.wrcc.dri.edu/monitor/cal-mon/; downloaded August 2009).

**Comparison Across River Basins (Fig. 4B).** We compared the water balance for 11 large river basins draining the western slope of the Sierra Nevada with mean basin temperature. The historical monthly full natural flow was downloaded for each river basin from the California Data Exchange Center (http://cdec.water.ca.gov/; downloaded March 2013), and the flow summed annually and averaged for 1981–2010. The river basins were demarcated by the US Geological Survey 8-Digit Watershed Boundary Dataset (http://datagateway.nrcs.usda.gov/; downloaded March 2013). The location of the relevant long-term gauging station was determined for each basin, and the relevant eight digit subbasins were merged, such that a single polygon was created for each basin that corresponded to the watershed above the gauge. The river basin boundaries were then combined with the PRISM 1981–2010 $T_{\text{max}}$ and P normals and the SRTM elevation to calculate basin-wide means. Finally, the observed riverflow ($Q$) was normalized by basin area and subtracted from P to estimate the basin-average ET, which was then compared with the mean basin elevation and $T_{\text{max}}$. We also compared the 1981–2010 mean $Q$ – P for each basin with the corresponding NDVI-based annual ET for 2003–2012 (Fig. S3).

Fig. S1. Annual total ET (mm yr$^{-1}$) across 10 sites and 46 site years as a function of annual mean NDVI (dimensionless). Figure is updated version of figure 2b from ref. 5. Ecosystem types are indicated by symbols and the water year of measurement is indicated by color. The year 2007 was much drier than normal and 2010 and 2011 were wetter than normal. The best-fit regression for all years was $ET = 10.3247 \times \exp(2.8599 \times NDVI)$; adj $R^2 = 0.914$.

Fig. S2. Relationship between elevation (meters above sea level) and ET (mm yr$^{-1}$; 2003–2012 average) determined from MODIS NDVI for all pixels in the upper Kings River basin at 0.002083° resolution.
Fig. S3. Relationship between basin-average P − Q and basin average NDVI-based ET for 11 major river basins on the western slope of the Sierra Nevada. P − Q was calculated as the 1981–2010 mean P for the basin based on the PRISM 30-y normal minus the corresponding annual mean full natural river flow for 1981–2010 divided by the area of the basin. NDVI-based ET is the mean annual NDVI for 2003–2012 calculated by averaging the results of the NDVI-base regression. The best-fit line is NDVI ET (mm yr⁻¹) = −117 + 1.25 (R² = 0.75), and the regression does not differ significantly from the 1:1 line.
Fig. S4. (A) Relationship between 30-y normal maximum air $T$ ($T_{\text{max}}$; 1981–2010 normals from PRISM) and annual ET determined from MODIS NDVI for all pixels in the upper Kings River basin with a $P$ above 900 mm yr$^{-1}$ (i.e., pixels that are not considered water-limited). (B) Relationship between 30-y normal $P$ ($P$; 1971–2000 normals from PRISM) and annual ET determined from MODIS NDVI for all pixels in the upper Kings River basin with a $T_{\text{max}}$ above 12 °C (i.e., pixels that are not considered cold-limited). The sigmoidal regression between ET and $T_{\text{max}}$ for the non–$P$-limited dataset was \[\text{ET} = \frac{792.6561}{1 + \exp\left(-\left(T_{\text{max}} - 9.3567\right)/2.5029\right)}; \text{adjusted } R^2 0.697\]. The sigmoidal regression between ET and $P$ for the non–$T$-limited dataset was \[\text{ET} = \frac{699.4783}{1 + \exp\left(-\left(P - 535.8093\right)/112.1342\right)}; \text{adjusted } R^2 0.194\].

Fig. S5. Relationship between elevation (meters above sea level) and ET predicted from the combined regressions in Fig. S4. The sigmoidal equations in Fig. S4 were used to separately calculate the ETs that would be expected for each pixel based on the local PRISM-based $P$ and $T_{\text{max}}$ climatology, and the predicted ET for each pixel determined as the minimum of the two values. A pixel-by-pixel comparison of the MODIS-based ET estimate (Fig. S2) and the regression-based estimate (Fig. S5) was [NDVI based ET (mm) = modeled ET (mm) $\times$ 1.033–9.9770; adjusted $R^2$ 0.690].
Fig. S6. Relationship between mean NDVI (2003–2012 calculated using the same quality flags used for the NDVI-based ET calculation) and local aspect at 100-m elevation intervals for the upper Kings River basin. South-southwest pixels are those with a slope of at least 15° and an aspect from 135° to 270°, where 360° is north. North-northwest pixels are those with a slope of at least 15° and an aspect from 315° to 90°.

Fig. S7. Relationship between elevation (meters above sea level) and projected air warming based on output from the CCSM-4 prepared for the CMIP-5. Lines connect points that show the mean warming projected for 2085–2100 relative to the historical temperature modeled for 1950–2005 based on four RCP. Projections were averaged across five or six CCSM-4 ensemble runs and for three grid cells over California’s central valley and immediately upwind (west) of the Kings River basin.
Fig. S8. Time series from September 2008 to July 2009 of air temperature measured at the top of the 2,015-m tower (at an absolute elevation of ∼2,050 m) and air temperature simultaneously measured above Vandenberg Air Force Base using radiosondes for operational meteorological forecasting. Each Vandenberg sounding was interpolated linearly to an altitude of 2,050 m. Vandenberg Air Force Base is ∼260 km southwest of the 2015-m tower. The radiosonde observations are at 12- or 24-h intervals, and the tower data at 30-min intervals.

Fig. S9. Time series of mean summer (June, July, and August) air temperature anomaly based on surface meteorological observations across the Sierra Nevada (solid line connecting means values) and the 850-mb radiosonde temperatures reported for Marine Corps Air Station Miramar (dashed line connecting means). Miramar Air station is 300–800 km south of the Sierra Nevada. The surface anomaly observations were provided by the California Climate Tracker dataset from the Western Regional Climate Center. The Miramar radiosonde observations were taken from the RATPAC-B dataset.