Understanding the role of asymmetrical warming on streamflow changes in the Western U.S.

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Key Points:
- Streamflow responses to warm versus cool season warming across four major basins are mostly consistent across four land surface models.
- Where inter-model sensitivities exist, they are mostly attributable to the differences in their ET algorithms.
- Basins' different runoff responses to warm vs. cool season warming are linked to seasonal incoming water ratios and temperature differences.

Abstract

Climate models project stronger warming in the warm season than in the cool season over much of the Western U.S. Across much of this region, the widely-accepted hydrologic signature of climate change is reduced winter snow accumulation and earlier streamflow peaks, which are mainly caused by cool season (October-March) warming. However, the warm season (April-September) warming effect receives far less attention. One of the few studies that investigated the warm season warming signature across the region (using a single land surface model) showed that while runoff timing shifts are associated with cool season warming, runoff volume changes are mostly attributable to warm season warming. However, what remains unclear are the
mechanisms controlling runoff responses to asymmetrical (warm/cool season) warming, as well as the extent to which the previous results are model-dependent. To answer these questions, we expand the earlier work to include experiments with four land surface models, and focus on the reasons for differences in annual and seasonal streamflow response to asymmetric seasonal warming. Our results show that: (i) the general features of seasonal and annual streamflow responses to asymmetric warming are consistent across models, although the magnitudes vary; and (ii) basins with higher ratios of warm to cool season gross incoming water, cooler summers, and colder winters have the strongest relative annual streamflow decreases for warm season warming relative to cool season warming. This is a consequence mostly of evapotranspiration change in response to temperature-related albedo changes and temperature sensitivity of the slope of the saturated vapor pressure curve.

1.0 Introduction

The Western U.S., one of the most rapidly growing parts of the country, is also its driest region. Much of the water consumed in the region comes from snowmelt (Li et al., 2017), a source experiencing significant volume and timing fluctuations as the climate of the region warms (Mote et al., 2005, 2018). These fluctuations in snowpack are usually accompanied by changes in the region’s annual streamflow, motivating efforts to understand the mechanisms dominating the region’s streamflow volumetric response to climate warming.

Many previous studies have examined both historical changes and future projections of the region’s streamflow (e.g., Stewart et al., 2005; Hayhoe et al., 2004; Kim and Jain, 2010; Gergel et al., 2017). These studies have generally found that warmer temperatures will lead to reduced winter snowfall and increased rainfall, earlier snowmelt, and hence earlier seasonal peak runoff in the year. These phenomena are primarily a consequence of cool season warming
The effect of warm season warming has been given less attention because of its perceived small(er) immediate effect on streamflow (Das et al., 2011). However, evaporative demands are larger in the warm season than in the cool season, and temperature increases predicted from climate model projections mostly indicate more warming in the warm season than in the cool season over much of the Western U.S. (Hayhoe et al., 2004) (see also supplement section S1 for more recent projection results using the CMIP5 multi-model ensemble archive). These facts raise questions as to the warm season warming effect on the region’s hydrology, especially annual streamflow volume changes.

Das et al. (2011) investigated the annual streamflow sensitivity of major river basins across the Western U.S. to cool and warm season warming using a single model simulation (Variable Infiltration Capacity model: VIC, Liang et al., 1994). They showed that warm season warming generally leads to larger annual streamflow decreases than cool season warming in four major Western U.S. river basins. This conclusion highlighted the importance of both seasons’ warming effects on the sensitivity of Western U.S. streamflow to climate warming. Two other recent papers (Vano and Lettenmaier, 2014; Vano et al., 2015) investigated the sensitivity of Western U.S. streamflow to temperature change on a seasonal scale, again finding substantial negative streamflow changes in response to warming. Like Das et al. (2011), they performed simulations with a single model (also VIC) to study how streamflow responds to warm and cool season warming. Furthermore, they focused more on “how much” than on “why”. Taken together, these three papers prompt two motivating questions for our work here: first, whether or not “how” the region’s streamflow will change in a warming climate is model-dependent, and second, “why” the streamflow responses to seasonal warming show the seasonally-dependent sensitivities documented in Das et al. (2011).
To address the two questions, we first investigate the streamflow response patterns to seasonal warming over the four river basins using VIC and three other hydrology models (described in sections 3 and 4.1). Then we identify the dominant factors controlling annual streamflow responses to seasonal warming using a water balance framework (section 4.2). The mechanisms controlling those dominant factors of streamflow response are also discussed in sections 4.3-4.4.

2.0 Study area

Our study domain consists of the same four major Western U.S. river basins as in Das et al. (2011): the Columbia basin; the Upper Colorado basin (hereafter Colorado); California’s Northern Sierra Nevada (N. Sierra); and California’s Southern Sierra Nevada (S. Sierra). For the Columbia basin, we consider the naturalized flows above the Dalles, OR, obtained from the Bonneville Power Administration; for the Colorado, we consider the naturalized flows of the Colorado River above Lees Ferry, AZ, obtained from the U.S. Bureau of Reclamation (USBR); for the N. Sierra, we use the sum of flows from gauges designated SBB, FTO, and YRS; for the S. Sierra, we use the sum of flows from gauges named MRC, SJF, SNS, and TLG. We obtained both N. Sierra and S. Sierra’s naturalized flow data from the California Data exchange center (http://cdec.water.ca.gov/dynamicapp/QueryWY). Details of the gauges’ full names and data sources are in Table 1.

Table 1. Stream Gauge (or location) names and data sources.

<table>
<thead>
<tr>
<th>Basin</th>
<th>Gauge full names</th>
<th>data source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Columbia</td>
<td>The Dalles</td>
<td>Bonneville Power Administration (BPA): <a href="https://www.bpa.gov/p/Power-Products/Historical-Streamflow-Data/Pages/No-Regulation-No-Irrigation-Data.aspx">https://www.bpa.gov/p/Power-Products/Historical-Streamflow-Data/Pages/No-Regulation-No-Irrigation-Data.aspx</a></td>
</tr>
<tr>
<td>Region</td>
<td>Station</td>
<td>Flow Details</td>
</tr>
<tr>
<td>--------</td>
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</tr>
<tr>
<td>Colorado</td>
<td>Lees Ferry</td>
<td>The sum of the Sacramento River above Bend Bridge (SBB), the Feather River at Oroville (FTO), and the Yuba River near Smartville (YRS)</td>
</tr>
<tr>
<td>N. Sierra</td>
<td></td>
<td>The sum of the Stanislaus River at Goodwin (SNS), The Merced River near Merced Falls (MRC), The Tuolumne River at LA Grange Dam (TLG), and the San Joaquin River Below Friant Dam (SJF)</td>
</tr>
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Based on the location of stations providing the naturalized flow, we delineated the four basin masks using a river channel network data set (Wu et al., 2012). Comparisons between model-simulated monthly hydrographs and observations are provided in supplement S2.

### 3.0 Data and approach

#### 3.1 Forcing and models

We used the Livneh (L13) meteorological forcings (Livneh et al., 2013) to drive VIC version 4.1.2 (Liang et al., 1994), the community Noah LSM with multi-parameterization options (hereafter Noah-MP) (Niu et al., 2011; Yang et al., 2011), the Sacramento Soil Moisture Accounting model (SAC-SMA) (Burnash et al., 1973) and the Catchment model (Koster et al., 2000). The L13 forcing was extended through 2018 (available to be downloaded from ftp://livnehpublicstorage.colorado.edu/public/sulu), so we had available to us consistent space-time records from 1915 through 2018 (all at 1/16 lat-long degree spatial resolution) over the entire conterminous U.S. (CONUS) domain. The L13 data contain four primary time-varying forcing fields: daily precipitation, maximum daily temperature (Tmax), minimum daily
temperature (Tmin), and surface wind speed. The first three are interpolated from daily weather station observations, while surface wind speed is interpolated from atmospheric reanalysis data (Kalnay et al., 1996; Livneh et al., 2013). Other variables required to force the models (e.g., downward solar and longwave radiation; humidity) are archived in L13 based on algorithms described in Bohn et al. (2013).

We applied all four models over the four basins for the period 1916 to 2018, using the water year 1915 as spin-up (10 times). As in Das et al. (2011), we ran VIC in water balance mode, which means that surface temperature is set to the surface air temperature for purposes of computing surface energy fluxes. We used VIC parameters from L13 to ensure consistency between model parameters and forcings, therefore we did not perform additional parameter estimation/calibration for VIC. We disaggregated the daily forcings (to hourly and three-hourly) using the Mountain Microclimate Simulation Model (MTCLIM) algorithms incorporated in VIC to estimate downward shortwave and longwave radiation, relative humidity, near-surface vapor pressure deficit, and surface atmospheric pressure. We also used the hourly disaggregated forcings as inputs to the Noah-MP model, and 3-hourly disaggregated forcings as inputs to the Catchment and SAC-SMA models. We computed potential evapotranspiration (PET, a required input to SAC-SMA) from the 3-hourly VIC disaggregated forcing variables using the Penman-Monteith algorithm (to produce reference ET, or ET₀, which was used as a surrogate for PET) based on the modified IGBP 20-category vegetation classification scheme (description: https://ral.ucar.edu/sites/default/files/public/product-tool/unified-noah-lsm/parameters/vegparm.html, data: https://ral.ucar.edu/sites/default/files/public/product-tool/unified-noah-lsm/parameters/VEGPARM.TBL). We then calibrated the PET adjustment parameter (PEADJ) in SAC-SMA, to make the output long-term climatological hydrographs...
approach the observed ones (Figure S2). Additionally, because the models cannot represent the horizontal movement of glaciers, in VIC, we reset SWE to zero on September 1st at every grid cell to avoid the continuous accumulation of SWE in some cold grid cells (mostly in the North Cascades), which are usually covered by glaciers in the real world. We only reset SWE for VIC because VIC tends to cause more pixels to have abnormally high SWE accumulation than other models, while other models’ over-year accumulated SWE only happens at very few pixels and had a negligible effect on our basin-scale discussion.

3.2 Temperature warming scenarios

We adopted the same seasonal warming scenarios as in Das et al. (2011). We applied 3°C warming to all historic values of both Tmax and Tmin. This assures that the downward solar radiation produced by the MTCLIM model, which is based on the daily temperature range, is hardly changed. For each model and basin, we applied four warming scenarios: a) 3°C warming year-round, b) 3°C warming only in the warm season, c) 3°C warming only in the cool season and d) baseline (no warming). We kept the precipitation unchanged in all model runs. Downward and emitted longwave radiation do change, though, as they are temperature-related.

4.0 Results and discussion

4.1 Multi-model streamflow response to seasonal warming

Multi-model simulations of streamflow responses to seasonal warming using the four models are shown in Figure 1. The results for VIC (red) are very nearly the same as in Das et al. (2011). In response to year-around warming, three out of four models (Catchment is the exception) show a reduction in streamflow in all four basins (Figure 1, top), with reductions ranging from about 4% to about 18%. The annual runoff response to year-around warming by the
Catchment model is quite small, ranging from about a 3% reduction in the Colorado basin to about a 3% increase in the N. Sierra. The annual flow response to year-around warming is the net result of differing seasonal responses, with reductions in warm season streamflow and increases in cool season streamflow, which is mostly less than the former (warm season flow reductions).

Interestingly, warm season-only warming produces reductions in flow in the warm season and, to some extent, in the cool season for all four models and four basins. In contrast, cool season-only warming produces increased streamflow in the cool season and diminished streamflow in the warm season.
Figure 1. Multi-model streamflow changes in the annual, warm season (Apr-Sep) and cool season (Oct-Mar) averages, in response to the annual, warm season, and cool season warming scenarios (in order, top to bottom). All values are percent changes relative to the average annual flow for the historical baseline run (1916-2018 water year), so the annual response equals the sum of warm and cool season responses. The black bars show the multi-model mean responses, and the colored lollipops are each model’s results.

Figure 1 shows that, for the Columbia basin and Colorado basin, the annual streamflow decreases are mostly associated with warming in the warm season, but for the California basins, the decreases are more associated with cool season warming in VIC and SAC-SMA, whereas Noah-MP shows slightly higher decreases under warm season warming than under cool season warming (Figure 1). SAC-SMA performs differently than other models in the Colorado, with a somewhat smaller annual streamflow decrease under warm season warming than under cool season warming. This is because, under warm season warming, SAC-SMA has a much smaller warm season evapotranspiration increase (Figure 2) compared with Noah-MP and VIC. As we will discuss later, warm season evapotranspiration change dominates annual evapotranspiration change under warm season warming. As a consequence, this smaller warm season evapotranspiration yields a smaller annual evapotranspiration increase, and thus a much smaller annual streamflow decrease under warm season warming than under cool season warming. We confirmed that the insensitivity of warm season evapotranspiration to warm season warming in SAC-SMA is caused by the constraining effect of the small increase of PET (a forcing for SAC-SMA) under warm season warming. Unlike the other models, PET is not computed internally by the model, and thus may not be well coupled with the variation of hydrological variables within SAC-SMA.
Figure 2. Multi-model 1916-2018 warm season average AET change under warm season warming, compared to the baseline run, using the Colorado River basin as an example. Figure 1 also shows that Catchment differs from the other models in its positive annual streamflow responses to cool season warming, which are negative for all the other models. This is because for cool season warming, MTCLIM produces increased specific humidity (Qair), which has a strong suppressive effect on evapotranspiration loss in the Catchment algorithm. We confirmed this by investigating eight sample pixels with negative cool season evapotranspiration change (both snow-free and snow-covered) under cool season warming: if we keep Qair fixed at baseline value, while using other variables from MTCLIM output under cool season warming, the cool season evapotranspiration in those pixels increases (similar to the other models). This suppressive effect of Qair on evapotranspiration causes Catchment to have a smaller evapotranspiration increase under cool season warming (Figure 3), and yields large cool season streamflow increases, which lead to a positive annual streamflow response. Therefore, despite the fact that Catchment has small warm season evapotranspiration increases for warm season warming (Figure 2), it still has stronger streamflow decreases under warm season warming than under cool season warming. We also confirmed the reason for the small warm season evapotranspiration increase in Catchment: under warm season warming, snow in the spring transition months (e.g., April) melts faster, leaving less soil moisture for the following warm months. With lower soil moisture, the fraction of area with less efficient evapotranspiration
generation (not-saturated, even wilting in some cases) increases, while the fraction of area with highly efficient evapotranspiration generation decreases (Koster et al., 2000). Therefore, the total evapotranspiration generation efficiency decreases under warm season warming in Catchment.

Figure 3. Multi-model 1916-2018 cool season average AET change under cool season warming, compared to baseline run, for the Colorado River basin.

Despite the discrepancies, the features of streamflow response to asymmetrical warming in the three models not used in Das et al. (2011) (Noah-MP, Catchment, SAC-SMA), as well as the multi-model mean (black bars, Figure 1) are overall consistent with Das et al. (2011), in which S. Sierra’s annual streamflow experienced similar decreases under both warm (3.6%) and cool (3.1%) season warming; N. Sierra had slightly stronger decreases under cool season warming (2.1% vs. 1.8%), and the Colorado and Columbia basins had larger annual streamflow decreases under warm season than cool season warming. This spatially and temporally heterogeneous pattern across basins motivates questions as to the mechanisms that are responsible for the differences.

4.2 Dominant factors controlling responses of annual streamflow to seasonal warming

In the following sections, we use the symbols (see also Table S2): ET for evapotranspiration, P for precipitation, T for temperature, SM for soil moisture, SWE for snow water equivalent. For a term with the form $X_{a,w3d}$, the first subscript indicates the period ($a$: annual, $w$: warm season, $c$: cool season) and the second subscript is the warming scenario ($w3d$:
warm season warming 3°C, c3d: cool season warming 3°C or b: baseline).

\[ \partial X, \ \Delta X \] represent the value of variable X under warming scenarios minus the baseline value. \( \bar{X} \) represents the long-term climatology of variable X averaged across the historical period (1916-2018).

According to the water balance equation (Eq. 1),

\[ P - ET - Q = \Delta SM + \Delta SWE \quad (\text{Eq. 1}) \]

when precipitation (P) is fixed (as in our experiments), and when the change of long-term soil moisture (\( \Delta SM \)) and snow water equivalent (\( \Delta SWE \)) are negligible over the multi-decade period of analysis, the change of annual actual evapotranspiration (ET) equals the change of annual streamflow (Q) (Eq. 2).

\[ \Delta ET = -\Delta Q \quad (\text{Eq. 2}) \]

Therefore, the source of differential annual Q response to seasonal warming can be derived from the source of differential annual ET response to seasonal warming. It is easier to understand the annual Q response from the aspect of ET than from the complicated precipitation-snowmelt-infiltration-saturation process of Q generation, since ET is a more “instant” and “local” process with no time-lag and no spatial movement, and has a cleaner seasonal partition than Q. Here, we quantify the relative strength of annual ET response to warm vs. cool season warming using a term \( \text{Pref}_{ET} \). \( \text{Pref}_{ET} \) (Eq. 3) is defined as annual ET change under warm season warming \((\overline{ET_{a,w3d}} - \overline{ET_{a,b}})\), divided by the annual ET change under cool season warming \((\overline{ET_{a,c3d}} - \overline{ET_{a,b}})\).

\[ \text{Pref}_{ET} = \frac{\overline{ET_{a,w3d}} - \overline{ET_{a,b}}}{\overline{ET_{a,c3d}} - \overline{ET_{a,b}}} \quad (\text{Eq. 3}) \]

\( \text{Pref}_{ET} > 1 \) indicates that basin has a larger annual ET change under warm season warming than under cool season warming, and \( 0 < \text{Pref}_{ET} < 1 \) indicates the opposite. In short, when \( \text{Pref}_{ET} \) is
positive, the larger Pref$_{ET}$ is, the stronger “preference” for a basin to have stronger annual ET increase (streamflow decrease) under warm vs. cool season warming.

Similarly, we can define a basin’s “preferred” seasonal tendency for annual streamflow change under warm vs. cool season warming, Pref$_{Q}$ (Eq. 4).

$$\text{Pref}_{Q} = \frac{Q_{a,w3d} - Q_{a,b}}{Q_{a,c3d} - Q_{a,b}}$$

(Eq. 4)

Pref$_{ET}$ should (approximately) equal Pref$_{Q}$ since we expect the sum of annual ET and annual Q to be constant (annual precipitation) under different warming scenarios. To check this, we plot the Pref$_{ET}$ and Pref$_{Q}$ from all four basins in Figure 4.

![Figure 4](image)

Figure 4. Pref$_{ET}$ and Pref$_{Q}$ across four basins estimated from outputs from the four models.

Figure 4 confirms that Pref$_{ET}$ and Pref$_{Q}$ are nearly the same across basins, with small differences between the two, except for larger differences in the sac results, especially in the Columbia basin. The large difference is probably due to the external computation of PET which is prescribed to the SAC ET algorithm. The small differences can be caused by multiple reasons.
that lead to the difference between $P_a$ and $\text{ET}_a+Q_a$ in different warming scenarios: removal of
SWE on September 1st in VIC; closing error of all four models within error limits; and SM and
SWE differences between the beginning year and ending year. Nonetheless, small discrepancies
between Pref$_{\text{ET}}$ and Pref$_Q$ are acceptable in the following discussions since they do not affect the
relative order of the basins’ response to seasonal warming (Figure 4). Therefore, to be brief,
hereafter we just refer to Pref$_Q$, and everything is the same for Pref$_{\text{ET}}$.

Figure 4 also shows that Pref$_Q$ is always positive except for in Catchment. For the other
three models, Columbia and Colorado are always the basins with first and second largest Pref$_Q$,
while the relative orders between the two basins with lower Pref$_Q$ (N. Sierra and S. Sierra) differ,
which is probably within the uncertainty of model behaviors. As discussed earlier, Catchment’s
negative Pref$_Q$ occurs because cool season warming causes ET$_c$ to decrease in Catchment,
yielding decreasing annual ET under cool season warming (which is related to the Qair
suppressive effect on ET in Catchment). For example, for the Colorado, Catchment has Pref$_Q$ of
around -6, which means warm season warming causes annual Q to have six times larger response
(decreasing) in the opposite direction to the response to cool season warming (increasing). This
means that the annual Q decrease or ET loss in Catchment can only be caused by warm season
warming. Therefore, it is not very meaningful to discuss the relative response of streamflow to
warm vs. cool season warming using Pref$_Q$ for the Catchment model. To simplify the discussion
of Pref$_Q$, hereafter we only consider the models whose basin-average Pref$_Q$ values are larger than
zero (i.e., VIC, SAC-SMA, and Noah-MP).

Across the models with positive Pref$_Q$, the relative order of the basins’ Pref$_Q$ values are
similar, while the magnitudes differ (Figure 4). These magnitude differences are mostly
attributable to differences in the models’ ET algorithms, but it would be arbitrary to say any one
of the algorithms is better than the others. Therefore, with the aim of treating each model with equal weight and focusing on the hydrological characteristics of the basins rather than model particulars, in the following section, we focus mostly on multi-model mean results.

So far we have discussed the features of annual ET responses under seasonal warming. Here, we discuss the factors that contribute most to those features. Note that seasonal warming causes both energy changes in the warm season and water availability changes in the following (cool) season. For example, cool season warming decreases snowfall in favor of rainfall and enhances snowmelt, leaving less meltwater for the following warm season. Warm season warming enhances evapotranspiration, increasing the soil moisture deficit in the warm season, leaving less soil moisture for the following cool season. Despite both energy and water availability changes under seasonal warming, the ET change for a particular season is usually a consequence (mostly) of a single factor (i.e., either energy or water availability change) under seasonal warming. Therefore, we plot the seasonal ET response to seasonal warming in Figure 5, to see whether one season dominates the annual ET change under seasonal warming scenarios.

Figure 5. Multi-model ET changes in the annual, warm season (Apr-Sep) and cool season (Oct-Mar) averaged across 1916-2018 water year, in response to warm season and cool
season warming (in order, top to bottom). The layout is the same as in Figure 1, but for ET. Note the bars in Figure 5 are ET change relative to baseline ET, so they are different in magnitude from the bars in Figure 1, which show Q changes relative to baseline Q.

Figure 5 indicates the season contributing most to annual ET change for cool season and warm season warming. Our interpretation is as follows:

(a) Under warm season warming, most of the ET increase occurs in the warm season, while the cool season ET increase is negligible (Figure 5, first row, column 1-4), which is also confirmed by monthly ET changes under warm season warming in Figure S7-S10 (multi-model results, each plot's first row, red lines). Note that the negligible cool season ET (ET) change under warm season warming does not conflict with the simultaneous reduction in cool season streamflow (Figure 1, middle row), because a greater portion of winter precipitation is used to satisfy a greater SM deficit at the end of warm season, which significantly reduces runoff but not ET. (Figure S7-S10, each plot’s second row, red lines). Therefore, the annual ET change is dominated by warm season ET (ET) change.

(b) Under cool season warming, snowfall decreases in favor of increased rainfall, less snow is accumulated, runoff and snowmelt are accelerated, so less snow remains when the warm season comes. In this case, the annual ET is the result of two often opposing changes: ET change (usually increases) in warmer winters (Figure 5, second row, lowest bars), and ET decreases due to the balance between increased evaporative demand and increased evaporative resistance (a consequence of drier soils with less snowmelt supply in the summer) (Figure 5, second row, middle bars). However, the increase in ET is usually much larger compared with the ET decrease (Figure 5, second row), and dominates the annual ET change under cool season warming.
Considering both (a) and (b), we expect to see that the annual ET response to warm vs. cool season warming is dominated by two contributors: the warm season ET response under warm season warming, and cool season ET response under cool season warming. Therefore, the relative orders of \( \text{Pref}_Q \) across basins should mostly follow the order of the ratio of \( \Delta ET_w \) change under warm season warming to \( \Delta ET_c \) change under cool season warming (Eq. 5).

\[
\text{Pref}_{ET} \sim \frac{ET_{w,3d} - ET_{w,b}}{ET_{c,3d} - ET_{c,b}}
\] (Eq. 5)

To check this, we plotted the \( \text{Pref}_Q \) and the ET change ratios (Eq. 5: \( \frac{ET_{w,3d} - ET_{w,b}}{ET_{c,3d} - ET_{c,b}} \) termed \( \frac{\Delta ET_w}{\Delta ET_c} \)) for the three basins for Noah-MP, SAC-SMA, and VIC as well as the multi-model mean (across the four models) in Figure 6.

**Figure 6.** Bar plots of \( \text{Pref}_Q \) and ET change ratio (\( \frac{\Delta ET_w}{\Delta ET_c} \), the second row) across four basins for Noah-MP, SAC-SMA, VIC and the multi-model mean.

The bar plots in Figure 6 confirm that the relative orders of \( \text{Pref}_Q \) (Figure 6, first row) are highly consistent with the ET change ratio (\( \frac{\Delta ET_w}{\Delta ET_c} \)) (Figure 6, second row). We also further...
confirmed the consistency of the spatial pattern of $\text{Pref}_Q$ and $\frac{\Delta \text{ET}_w}{\Delta \text{ET}_c}$ by plotting long-term climatological maps across four basins and four models, as shown in Figure S11a-d.

4.3 Dominant factors controlling ET temperature sensitivity in warm and cool season

We have confirmed in section 4.2 that the ET change ratio $\Delta \text{ET}_w/\Delta \text{ET}_c$ is the factor dominating the ratio of annual streamflow response to warm season vs. cool season warming. Now we attempt to determine whether changing water supply or evaporative demand is the main factor dominating $\Delta \text{ET}_w/\Delta \text{ET}_c$. In the numerator of $\Delta \text{ET}_w/\Delta \text{ET}_c$, $\Delta \text{ET}_w$ is warm season ET change under warm season-only warming. In this case, because cool season temperature remains unchanged, the moisture provided by the cool season to the warm season (i.e., SM and SWE remaining at the beginning of the warm season, April 1st) is nearly unchanged (Figure S3-S6, subplot a & b, across four models). Together with the unchanged $P_w$, the $\Delta \text{ET}_w$ is mainly due to the increasing evaporative demand under warming temperature in the warm season.

Similarly, the denominator, $\Delta \text{ET}_c$ is the change of cool season ET under cool season-only warming. In this case, because warm season temperature remains unchanged, the moisture provided by the warm season to the cool season (i.e., mainly the SM at the beginning of the cool season, Oct 1st) is also nearly unchanged (Figure S3-S6, subplot e, across four models). Together with the unchanged $P_c$, the $\Delta \text{ET}_c$ is mainly caused by the increasing evaporative demand associated with warmer temperatures in the cool season. Therefore, both $\Delta \text{ET}_w$ and $\Delta \text{ET}_c$ are most related to temperature increases, rather than water availability changes.

Considering the two aspects above, if we can find the factors linking the basins’ warm vs. cool season ET change to temperature, we can explain why the $\text{Pref}_Q$ values differ, and finally why the relative response strength of streamflow to warm vs. cool season warming differs across
basins. To do this, we investigate the pattern of ET-temperature (ET-T) sensitivity in this section, with a focus on the key factors that cause ET-T sensitivities to vary across basins. We do so using multi-model mean results, and link them back to the PrefQ patterns we identified earlier.

We start from four forcing variables that might strongly affect ET: vapor pressure deficit (VPD), air temperature (Tair), net radiation (Rnet), and wind speed (u). Because in the warm season, snowmelt is also a dominant source of water for evapotranspiration, we add a fifth term: gross incoming water (GIW). GIW to the cool season (GIW_c) is defined as the sum of soil moisture (SM) and snow water storage (SWE) at the beginning of the cool season (defined as Oct 1st), plus the cool season precipitation. GIW to the warm season (GIW_w) is defined as the sum of soil moisture (SM) and snow water storage (SWE) at the beginning of the warm season (defined as April 1st), plus warm season precipitation. In Figure 7, we show (as scatterplots) the relationship of multi-model mean ET with each of the five variables, on a basin-averaged, annual basis. The red points in Figure 7 represent the warm season ET-T sensitivity together with each of the five variables under warm season warming, and the blue points are for cool season warming. We evaluate correlations based on normalized values of the variables using the Pearson correlation method, and linear regressions are fit to the points. The correlation and the p-value of the correlations are in the legend for each plot (Figure 7a-e), where we take p=5% as the significance level.
Figure 7. Scatterplots of the relationship between ET-T sensitivity and VPD, Tair, Rnet, gross incoming water (GIW) and wind speed. ET is obtained from the multi-model means. The forcing variables are the same for each of the four models.

Among the five variables, for VPD (Figure 7, first row), the correlation values are all negative. Warm season correlations are all significant, but two of the cool season correlations are not (for p<5% significance). The warm season has larger correlations, as well as generally higher slopes of dET/dT vs VPD linear regression lines as compared with the cool season. These patterns indicate that decreases in the warm season VPD are associated with increases in ET-T sensitivity in the warm season, while the cool season VPD does not have a strong relationship with the ET-T sensitivity.

For Tair (Figure 7, second row), the correlation values in the warm season are all negative and significant. The correlation values in the cool season are mostly significant, and the significant ones are all positive. The warm season generally has higher slopes for dET/dT vs Tair linear regression lines than for the cool season. These patterns indicate that decreases in the warm season Tair and increases in the cool season Tair are associated with strong ET-T sensitivities.

For Rnet (Figure 7, third row), the correlations in the warm season are all negative and significant, this suggests that there are clear negative correlations between Rnet and ET-T sensitivity in the warm season, which may well be reflecting the relationship between the decreasing ET-T sensitivity and stronger warm season Tair (usually with higher Rnet). In the cool season, the significant correlations (for dET/dT vs Rnet) shift from negative (Colorado) to positive (Columbia and N. Sierra), not consistent across basins. This is suggesting that the
relationship between Rnet and ET-T sensitivity may be influenced by the other variables’ yearly change.

For GIW (Figure 7, fourth row), the correlation values in both warm and cool seasons are positive. In the warm season, all of the correlations are significant; in the cool season, the correlation in the Colorado and Columbia basins are significant. This suggests that in both seasons, increases in the GIW are generally associated with stronger ET-T sensitivity.

For wind speed (Figure 7, fifth row), almost all of the correlations are not significant. Therefore, we ignore the effect of wind speed on ET-T sensitivity in our subsequent discussion.

Considering all of the above, we summarize as follows: in the cool season, ET-T sensitivities increase with GIW and Tair; in the warm season, ET-T sensitivity increases with GIW, but decreases with Tair, VPD, and Rnet. Wind speed has no significant relationship with ET-T sensitivity in either season.
Figure 8. Boxplots of seasonal dET/dT, VPD, Tair, Rnet and GIW for the four basins’ yearly basin-average value based on multi-model mean results. The forcing variables are from MTCLIM output, which serves as the forcings for all four models. The upper and lower edges of the boxes define the Inner-Quartile Range (IQR: 25th percentile (Q1) ~ 75th percentile (Q3)), and the horizontal line is the median (Q2). The red points are outliers, defined as the values smaller than Q1-1.5*IQR or larger than Q3+1.5*IQR. The upper and lower limits of the vertical bars are the maximum (Q3+1.5*IQR) and minimum (Q1-1.5*IQR). This plot was generated using function geom_boxplot in R software.

Figure 8 shows that, in the cool season, Tair values have the following order: N. Sierra > S. Sierra > Columbia > Colorado; and GIW has the order: S. Sierra > N. Sierra > Colorado > Columbia. The dET/dT has the order: S. Sierra > N. Sierra > Colorado > Columbia. These reflect that overall, dET/dT in the cool season follows positive correlations with GIW and Tair as suggested in Figure 7. More specifically, taken across basins, dET/dT tends to have a stronger dependence on precipitation when Tair is warmer (i.e., in N. Sierra and S. Sierra), and a stronger dependence on Tair when Tair is lower (i.e., in Colorado and Columbia basins).

In the warm season, the order of dET/dT generally follows the order of GIW. The relative order of VPD, Tair, and Rnet are also generally consistent with their negative correlations with dET/dT as suggested in Figure 7 for the warm season: Columbia basin has the smallest VPD, Tair, and Rnet, and has the largest dET/dT; the other three basins have larger VPD, Tair and Rnet, which is consistent with their smaller dET/dT as compared with the Columbia basin. Since VPD and Tair are highly correlated, and Tair usually increases with Rnet, we can simplify the pattern in the warm season to be: ET-T sensitivity generally decreases at higher Tair, when Tair is similar, the basin with more available water (i.e. higher GIW) tends to have higher ET-T
Overall, water availability dominates the ET-T sensitivity in the warm season across basins.

In summary, the ET-T sensitivity is closely associated with the basins’ water availability (reflected by precipitation and GIW in the cool and warm season) and Tair. ET-T sensitivity generally increases with water availability; this tendency is stronger especially when Tair is high. When Tair is low, the ET-T sensitivity order mainly increases with Tair, when Tair is high, ET-T sensitivity decreases with Tair. This pattern is also confirmed by pixel-by-pixel scatterplots across all four basins (Figure S13).

4.4 Understanding the relationship between Tair and ET-T sensitivity

In the previous section, we confirmed that Tair and water availability (GIW) are the two variables that control ET-T sensitivity. It is easy to understand that in general, when energy does not limit evapotranspiration, higher water availability favors higher seasonal ET-T sensitivity. What remains to understand is, what causes the shift from positive dET/dT vs. T relationships to negative when Tair increases from cool to warm?

Figure S12 shows the strong positive correlation between Tair and VPD, which suggests that the shift from lower to higher Tair is usually accompanied by an increase in VPD. As we will see later, this increase in VPD is linked with the shift in the sign of the dET/dT vs Tair-correlation.

We explain this shift using the Penman-Monteith reference ET₀ framework (Eq. 6) (Monteith, 1965; Allen et al., 1998).

\[
ET₀ = \frac{0.408 \Delta (R_{\text{net}} - G) + \frac{208}{r_a} \frac{900}{T+273} VPD}{\Delta + \frac{\Delta}{r_a}} \quad (\text{Eq. 6})
\]
In Eq. 6, ET₀ is the reference evapotranspiration flux, Rₚ is net radiation (fixed at 10 MJ/(m²*d), which is a typical value of clear sky conditions in the warm season). G is the soil heat flux (we assume it to be 0), γ is the psychrometric constant≈0.0674 kPa °C⁻¹, T is the mean daily air temperature at 2m height (Tair), u is wind speed at 2m height, (eₛ - eₐ) is the saturation vapor pressure minus actual vapor pressure, which equals vapor pressure deficit of the air (VPD, kPa), ρₐ is the mean air density at a constant pressure, Δ is the slope of the saturation vapor pressure-temperature relationship, which follows the form of Eq. 7, rₛ and rₐ are the bulk surface resistance and aerodynamic resistance, in this reference framework, rₛ=70s/m, here we assume rₛ and wind speed follows the relationship rₐ= 208/u.

\[
\Delta = \frac{4098 \cdot 0.6108 \cdot e^{17.27T/(237.3+T)}}{(237.3+T)^2} \tag{Eq. 7}
\]

From Eq. 6, we can see that ET₀ is composed of two parts: a radiation-driven part (Part 1), and a vapor pressure deficit part (Part 2).

Part 1 takes the form of \( \frac{0.408 \Delta (Rₚ-G)}{\Delta + \gamma (1+\frac{rₛ}{rₐ})} = \frac{0.408 (Rₚ-G)}{1 + \frac{rₛ}{rₐ}} \). When other variables in Part 1 are fixed, the derivative of Part 1 w.r.t Tair takes the form of \( \frac{0.408 (Rₚ-G) \gamma (1+\frac{rₛ}{rₐ})}{(\Delta + (1+\frac{rₛ}{rₐ}) \gamma)^2} \cdot \frac{d\Delta}{dT} \). This sensitivity of Part 1 to Tair is positive and mostly increases with Tair (Figure S15a). This is because the warmer Tair leads to a larger slope of saturation water vapor vs Tair (i.e., Δ) and Δ increases faster under warmer Tair. It could be possible that beyond a certain Tair threshold (not plotted), higher Δ in the denominator starts to play a larger role and the sensitivity of Part 1 to temperature could decrease. Part 1 indicates the effect of the higher moisture holding capacity on the moisture flux.
The sensitivity of Part 2 mostly decreases with Tair (Figure S15b). This is mainly because warmer Tair leads to a larger $\Delta$ in the denominator, and $\Delta$ increases more quickly for warmer Tair, thus reducing Part 2.

Combining Part 1 and Part 2, when VPD becomes larger (transitioning from a cool environment to a warm environment), the relative contribution of Part 2 to ET is gradually magnified compared with that of Part 1, therefore, the response of ET-T sensitivity shifts from increasing with Tair to decreasing with Tair.

Our discussion above assumes that Rnet is fixed for a warmer temperature. However, Rnet does change (mainly due to upward shortwave change, see Figure S16, fourth row based on VIC results). This is because the effective albedo decreases in a warmer climate due to the transition of snow cover to snow-free conditions, especially from late winter through early summer. We further explored the relationship between warming-induced Rnet change and ET change in warm and cool season warming using the VIC model results.

First of all, we derive the relationship between the change of upward shortwave and albedo between the baseline and warming scenarios, as shown in Eqs. 8 and 9:

$$
\Delta \alpha_{ttl} = \Delta f_s (\alpha_s - \alpha_{non-s}) + f'_s \Delta \alpha_s + (1 - f'_s) \Delta \alpha_{non-s} \quad \text{(Eq. 8)}
$$

$$
\Delta SW_{up} = SW_{down} * \Delta \alpha_{ttl} \quad \text{(Eq. 9)}
$$

Where $\Delta$ is the baseline value minus the warmer climate value, $\alpha_{ttl}$ is albedo averaged across each basin, considering both snow-covered and snow-free regions, $\alpha_s$ is snow albedo, $\alpha_{non-s}$ is non-snow region albedo, $f_s$ is baseline snow cover fraction (as a ratio of basin area), $f'_s$ is snow cover fraction under warmed scenarios. From Eq. 8, Eq. 9 and our results from the VIC model, we found that, under cool season warming, multiple factors affect the relative change of the cool season Rnet compared to other basins, and consequently the cool season ET-T...
sensitivity. First, basins with warmer baseline cool season temperatures or thinner winter snowpack (e.g. Colorado and N. Sierra, Figure S17, S22) tend to have greater snow coverage reductions (Figure S19a, cool season months). Second, when snow cover reductions are similar, the basin with smaller non-snow regional albedo (usually means less grass and more trees) and smaller f_s tends to have larger albedo changes (e.g. S. Sierra compared with Colorado and Columbia, Figure S19a, S20a, S23a, and S24a, all see January, cool3d lines). Third, when albedo changes are similar, the basin with a stronger downward shortwave tends to have stronger Rnet increases (e.g. S. Sierra compared with Columbia, Figure S20a, S18, S21a, all see cool season months), thus stronger cool season ET-T sensitivity due to Rnet change (Figure 8a, cool season). All of these three tendencies that lead to stronger cool season ET response to cool season temperature warming via Rnet change are more prone to occur in basins with warmer winters, synergistically supporting the smaller ET-T sensitivity at lower Tair in the cool season as discussed when Rnet is fixed.

Second, under warm season warming, similar factors are functioning but snow cover change becomes the most influential factor influencing Rnet change and the ET-T sensitivity. Basins with cooler summer (e.g., Columbia and S. Sierra in comparison to Colorado and N. Sierra, Figure S17) tend to have larger snow-covered fractions, which will generally lead to greater snow cover reductions (Figure S19b, warm season months) for the same temperature increment (3°C) accumulated across the entire warm season. In this way, more SW absorption (Figure S21b, warm season months) will occur due to the associated albedo change, which leads to a greater warm season ET increase (Figure 8a, warm season). Contrastingly, basins with warmer summers in the baseline (i.e., Colorado and N. Sierra, Figure S17) tend to have higher snow lines. They may have thinner snowpack and therefore may have larger snow coverage
reductions in April than other basins (Figure S19b), but in the following summer months, they generally have smaller snow coverage reductions (Figure S19b). These basins, therefore, have smaller SW absorption changes (Figure S21b) and smaller ET-T sensitivity in the warm season (Figure 8a, warm season). These tendencies also synergistically support the larger ET-T sensitivity at lower Tair in the warm season as discussed when Rnet is fixed.

5 Conclusions

Our multi-model simulation of streamflow response to warm and cool season warming confirms that the response pattern is generally consistent across models. Moreover, our multi-model averages are consistent with the results of Das et al. (2011) for a single model (VIC). Compared with the impact of cool season warming, warm season warming always has the largest relative impact on annual streamflow change in the Columbia basin, and is progressively less for the Colorado basin, and smaller still for the N. Sierra and S. Sierra basins.

We applied a water balance framework that links the evapotranspiration-temperature response with the streamflow response, from which we found that the ratio of ET-T sensitivity in warm vs. cool season warming is the factor that dominates the relative response of annual ET and streamflow to warm vs. cool seasonal warming (i.e., Pref_{ET} and Pref_{Q}).

We further evaluated the dominant factors affecting the ET-T sensitivity, and discovered that it is closely associated with the basins’ water availability (i.e., gross incoming water GIW in the warm and the cool season) and Tair. The ET-T sensitivity generally increases with water availability, and this tendency is strongest when Tair is high. When Tair is low, the ET-T sensitivity mainly increases with Tair, whereas when Tair is high, the ET-T sensitivity decreases with Tair.
We first explored the mechanisms controlling the shift in the relationship between Tair and ET-T sensitivity in cool and warm environments when Rnet is fixed. When Tair warms, the slope of the saturation vapor pressure (Δ) vs. Tair increases. The increase in Δ with Tair implies that the ET-T sensitivity increases when evapotranspiration is dominated by the radiation part of the Penman-Monteith equation (i.e., VPD is small, in which case Tair is usually low, especially in the cool season), or decreases when evapotranspiration is dominated by the vapor deficit part (i.e., VPD is large, in which case Tair is usually high, especially in the warm season).

When the change of Rnet in response to temperature warming is considered, we found synergistic relationships with the fixed Rnet case: under cool season warming, in basins with higher Tair, upward shortwave generally decreases more with reduced albedo due to snowmelt, favoring higher ET-T sensitivity. Under warm season warming, in basins with higher Tair, higher snow lines lead to less available snow-soil/snow-meltwater transition, thus less albedo decrease and less upward shortwave decrease, which favors lower ET-T sensitivity.

With all above considered, we suggest that basins with colder winters and cooler summers, and higher warm season vs. cool season precipitation ratios (e.g., the Columbia basin and the Colorado basin), have a stronger tendency to have a larger response to warm season warming than to cool season warming. (More accurately, the relationship can be described in terms of higher warm season GIW vs. cool season GIW ratio). In any event, cool basins tend to have a larger response to warm season relative to cool season warming in terms of decreases in annual runoff volumes.

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All data used in our paper are publicly accessible online at the following URLs: Livneh meteorological forcings from 1915 to 2011
Livneh meteorological forcings 2012-2018, extended by Lu Su in Land Surface Hydrology Research Group in UCLA.

All four hydrologic models are open source and can be downloaded online: VIC-412g (https://github.com/UW-Hydro/VIC/releases/tag/VIC.4.1.2.g), Noah-MP hrldas version 3.9 (https://github.com/NCAR/hrldas-release/tree/master/HRLDAS), Catchment source code is available as a part of the GEOS source code under the NASA Open-Source agreement (http://opensource.gsfc.nasa.gov/projects/GEOS-5). Anyone interested in the SAC-SMA source code can either request via dist_hydro_mod@infolist.nws.noaa.gov or write a request email to nws.nwc.ops@noaa.gov and the code would be available through FTP for the requestor to download. Other representative data related to the results in this paper (and supplements) are uploaded to the link https://doi.org/10.6084/m9.figshare.11605818.v1.

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