

The importance of warm season warming to western U.S. streamflow changes

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[1] Warm season climate warming will be a key driver of annual streamflow changes in four major river basins of the western U.S., as shown by hydrological model simulations using fixed precipitation and idealized seasonal temperature changes based on climate projections with SRES A2 forcing. Warm season (April–September) warming reduces streamflow throughout the year; streamflow declines both immediately and in the subsequent cool season. Cool season (October–March) warming, by contrast, increases streamflow immediately, partially compensating for streamflow reductions during the subsequent warm season. A uniform warm season warming of 3°C drives a wide range of annual flow declines across the basins: 13.3%, 7.2%, 1.8%, and 3.6% in the Colorado, Columbia, Northern and Southern Sierra basins, respectively. The same warming applied during the cool season gives annual declines of only 3.5%, 1.7%, 2.1%, and 3.1%, respectively. **Citation:** Das, T., D. W. Pierce, D. R. Cayan, J. A. Vano, and D. P. Lettenmaier (2011), The importance of warm season warming to western U.S. streamflow changes, *Geophys. Res. Lett.*, 38, L23403, doi:10.1029/2011GL049660.

1. Introduction

[2] Much of the western U.S. is dependent on mountain runoff as a primary supply of water for agriculture, industry, and growing urban populations. Streamflow is likely to decline over much of the region, especially the Southwest, due to anthropogenic climate change [Milly *et al.*, 2005; Seager *et al.*, 2007; Christensen and Lettenmaier, 2007; Cayan *et al.*, 2010]. In some of the region, one mechanism driving this prospective decline is a projected reduction in precipitation due to altered atmospheric circulation. The other major impact arises through the response of land surface processes to climate warming, in particular through changes in evapotranspiration (ET) and sublimation from snow. Land surface processes are the focus of this work.

[3] Previous authors have documented declining annual streamflow in the Pacific Northwest [Luce and Holden, 2009] and a shift towards earlier runoff in the western U.S. in observations and climate change simulations [e.g., Hayhoe *et al.*, 2004; Stewart *et al.*, 2005; Hidalgo *et al.*, 2009; Kim

and Jain, 2010]. Those studies have generally used combined changes in temperature (T) and precipitation (P), which complicates identification of the seasonal response in the T -driven signal. The effects of T and P trends on snowpack were considered separately by Hamlet *et al.* [2005], but they did not examine winter vs. summer changes. Additionally, these earlier works have explored how streamflow changes with realistic year-round warming, but have not determined the response of the land surface to warming in different seasons.

[4] If precipitation is fixed and changes in local storage are small, only changes in evaporation or transpiration from plants can change yearly-averaged streamflow. Increased evaporation can drive increased local or regional precipitation that partially mitigates the drying effects of the evaporation, an effect often called “recycling”. Traditional estimates of recycling suggest only 3–10% of the precipitation over the western U.S. originates as local evaporation [Trenberth, 1998], but moisture budgets suggest higher values [Anderson *et al.*, 2009]. We do not consider moisture recycling, so our results are an upper limit to changes in runoff (see also section 4).

[5] Runoff depends on interacting processes that vary with season, such as snow accumulation in winter, melting in spring, and high evaporative demand in summer. Most water in the western U.S. is delivered to stream channels during the cool season. However, our results show that summer warming affects annual streamflow the most: there is a seasonal asymmetry in the response of streamflow to warming. This asymmetry could be of practical importance, since most global climate models project more summer than winter warming in the western U.S. [e.g., Hayhoe *et al.*, 2004].

2. Methods

2.1. Study Area and Data

[6] The study area consists of four regionally important river basins: a) the Upper Colorado; b) the Columbia; c) California’s Northern Sierra Nevada; d) California’s Southern Sierra Nevada (Figure S1 in Text S1 in the auxiliary material).¹ Geographical and climatic conditions of the basins are shown in Table S1 in Text S1.

[7] Daily gridded observations of precipitation, maximum daily temperature (T_{max}), minimum temperature (T_{min}), and wind speed over the period 1915–2003 at 1/8 degree latitude by longitude spatial resolution across the western U.S. [Hamlet and Lettenmaier, 2005] were used to drive the Variable Infiltration Capacity (VIC) hydrological model

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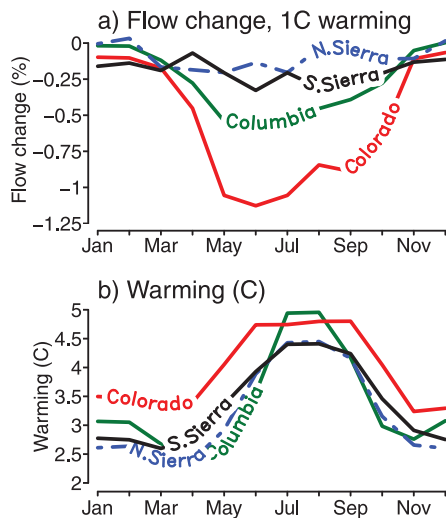


Figure 1. (a) Change in annual streamflow (%) obtained when the temperature in the indicated month is increased by 1°C. (b) Monthly temperature change (2050–2099 average minus 1950–1999 average) projected by an ensemble of 16 global climate models using the SRES A2 climate change scenario and statistical downscaling [Maurer et al., 2007]. Values are averaged over the Colorado, Columbia, North and South Sierra basins.

[Liang et al., 1994; Cherkauer et al., 2003]. Details of the forcing are given in section S2 of Text S1. Humidity and downward solar and longwave radiation are estimated using the algorithms of Kimball et al. [1997] and Thornton and Running [1999], which are based on the daily temperature range and daily average temperature, respectively, as described by Maurer et al. [2002].

[8] Future temperature changes across the western U.S. are based on the average of 16 statistically downscaled global climate model projections using the SRES A2 emissions scenario, obtained from <http://www.engr.scu.edu/~emaurex/data.shtml> [Maurer et al., 2007].

2.2. Models and Analysis

[9] The VIC model was applied over the four basins on a $1/8 \times 1/8$ degree grid [Barnett et al., 2008]. Up to 11 different vegetation classes were allowed in each grid cell. We assumed that movement of the tree line will lag climate warming by some decades, and so kept vegetation geographically fixed (although parameters such as the leaf area index vary monthly). A sensitivity experiment showed that future streamflow declines are greater when the tree line is allowed to migrate to higher elevations with warming, in which case the increased tree cover leads to greater transpiration losses. Runoff is routed [Lohmann et al., 1996] to simulate streamflow in the four river basins. Further model details are given in section S3 of Text S1.

[10] Different land-surface models produce different runoff given the same forcing [e.g., Henderson-Sellers et al., 1993]. Also, different parameterizations of potential evapotranspiration (PET; the ET that would arise with no water limitations) respond differently to warming [Kingston et al., 2009]. Because of this, results from any one model are uncertain. However, we use VIC because 1) it is computationally efficient; 2) it reproduces observed streamflow over

the western U.S. quite well when driven by observed meteorological forcing [e.g., Hidalgo et al., 2009; Hamlet et al., 2005]; 3) it uses the Penman-Montieth formulation of PET, a physically based method well-suited to a range of climates [Allen et al., 1998]; 4) it represents the vapor pressure feedback (closing plant stomata under dry conditions), which is important for the land surface water balance [Koster and Milly, 1997].

[11] When forced with observed P , T_{\max} and T_{\min} over the period 1915–2002, VIC simulated flow compares well with observations, reproducing the strong seasonality (Figure S3 in Text S1), runoff efficiency (Figure S4 in Text S1), and substantial interannual-decadal variability observed in each basin. The difference between modeled and observed (naturalized) mean annual flow is 0.2%, 3.0%, 12%, and -1.2% in the Colorado, Columbia, northern and southern Sierra, respectively, while the correlation between the annual time series of simulated and observed flow is 0.93, 0.87, 0.91, and 0.97 in the same basins. The RMS error in the annual-mean time series, expressed as a percent of the observed mean annual flow, is 11%, 9.5%, 19.5%, and 12.2%, respectively. In sum, VIC provides both a credible first look at the problem and a benchmark for later studies using other models.

[12] In our warming experiments we offset daily temperatures by a constant, leaving downward solar radiation unchanged. This allows us to isolate the effects of warming while imposing daily P and T sequences that are reasonably realistic. VIC determines the form of P (rain or snow) based on T , so the appropriate phase is always applied.

3. Results

[13] To examine the sensitivity of streamflow to warming at different times of the year, we performed 12 experiments where T_{\min} and T_{\max} were increased 1°C in just one month of the year (Jan, Feb, ..., Dec). Water storage in the soil and snowpack impart memory to the system, so we calculated changes in annual streamflow even though the temperature perturbations were confined to one month.

[14] Annual streamflow decreases for a single month's warming are shown in Figure 1a as a function of the month warming is applied. Three of the four basins are more sensitive to warm season warming than to cool season warming. The Colorado River is especially sensitive, with annual streamflow declines in response to a single summer month's 1°C warming of 1.2%, vs. 0.1% for a winter month's 1°C warming. The Northern and Southern Sierra Nevada are the least sensitive (peak streamflow decreases of 0.3%) and show less asymmetry in response to summer vs. winter warming. The Columbia River falls between, with declines in annual streamflow of about 0.6% for a summer month's warming, and near zero for winter warming.

[15] Figure 2 shows the simulated monthly changes in actual (as contrasted with potential) evapotranspiration (AET), runoff, and soil moisture from experiments with 1°C warming applied in January (top row of Figure 2), April, July, and October (bottom row). For brevity, only the Colorado River results are shown. The time evolution of the anomalies depends greatly on which month the warming is applied. In some cases, for example AET and runoff with July and October warming, there is a delayed response the following summer. This shows there is “memory” in the

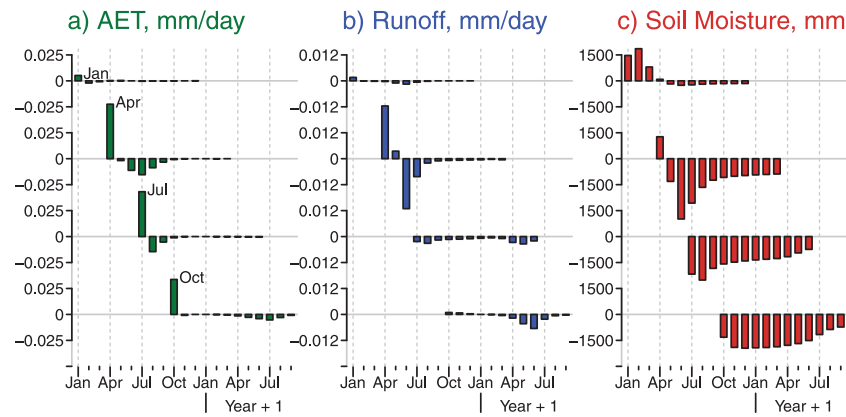


Figure 2. (a) Climatological change in actual evapotranspiration (AET, mm/day), (b) runoff (mm/day), and (c) soil moisture (mm) over the 12-month period following a single month of 1°C warming applied uniformly over the Colorado River basin in the month indicated.

system that resides primarily in the soil moisture (Figure 2c); January warming increases soil moisture as snow melts and precipitation shifts to rain, but the effects drop nearly to zero after two months. Summer warming, in contrast, results in an immediate increase in AET which imparts a soil moisture deficit that persists through the winter into the following summer. In such a case there has been a net loss of water from the land surface, which becomes more arid as a result, despite the fact that P is held constant.

3.1. Year-Round Warming

[16] The single-month warming experiments are useful in diagnosing the response of the hydrological system, but anthropogenic climate warming will almost certainly be experienced in all months. Figure 1b shows that the yearly averaged warming is likely to be at least 3°C in all the basins examined here by late in the current century. We consider the yearly-averaged warming case first, then explore the additional effects of the greater summer than winter warming in section 3.3.

[17] An asymmetry in seasonal streamflow response is still present when a 3°C uniform year-round warming is applied (Figure 3a). All basins exhibit a comparable reduction in warm season streamflow: -14.6%, -11.8%, -11.0%, and -14.3% in the Colorado, Columbia, northern and southern Sierra, respectively. Yet annual streamflow reductions vary by a factor of four: -16.1%, -8.4%, -3.9%, and -6.4% in the same basins. Thus, inter-basin differences in annual response must be determined largely by the disparity in cool season streamflow changes, which range from -1.5% (Colorado) to +7.9% (southern Sierra).

[18] Even with fixed precipitation, warming increases the liquid water input (LWI, the sum of rainfall and snow melt) to the surface during the cool season by changing the rain/snow mix and causing earlier snow melt. The difference in both the quantity and disposition of extra LWI accounts for the differences in cool season streamflow, and hence annual streamflow between basins. The difference in quantity between basins is well understood; 1 degree of warming in the higher, colder basins results in less transition from snow to rain and less snowpack crossing the melting point. However the disposition of that extra LWI also matters, and varies widely across basins. In the Northern Sierra, 41% of

the extra LWI in winter due to warming runs off that same winter, largely offsetting the severe warm season drying in the annual average. The comparable figures for the Southern Sierra and Columbia basins are 25% and 16%, respectively. In the Colorado basin, systematically more arid conditions, even during winter, lead to only 2% of the extra LWI input during winter running off that same winter; the severe summer declines in streamflow are little compensated by increased winter streamflow, and the annual average decline in streamflow is greatest of the four basins.

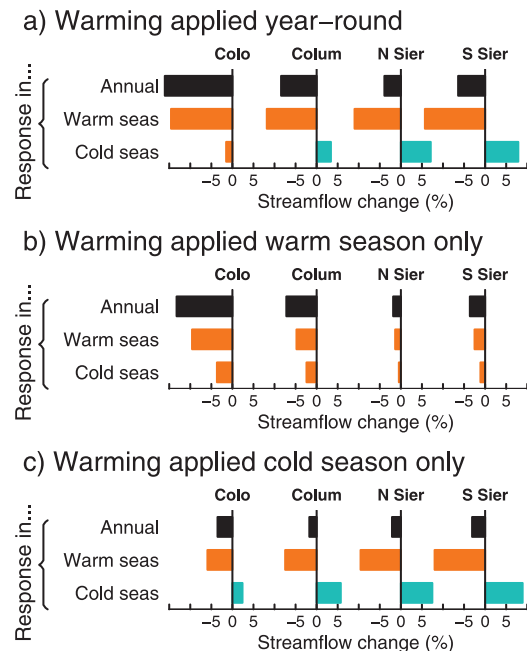


Figure 3. Changes in streamflow (%) in the annual, warm (April–September), and cool (October–March) season averages, for: (a) uniform 3°C year-round warming; (b) 3°C warming in the warm season only; (c) 3°C of warming in the cool season only. All values are given as percent of the annually averaged flow in the historical climate control run, so that the annual change is the sum of the warm and cool season changes. X axis tic interval is 5%.

3.2. Warm Versus Cool-Season Warming

[19] The different behavior by season is even more clearly seen when 3°C of warming is applied only during the warm or cool season, an idealized experiment that probes the hydrological effects of warming. Consider the response to either warm-only or cool-only season warming in the annual mean streamflow (black bars of Figures 3b and 3c). Warm-season warming reduces annual streamflow far more in the Colorado and Columbia Rivers than does cool season warming.

[20] Figure 3 also shows a profound difference between the seasonal contributions to the annual flow reduction. For warm season-only warming (Figure 3b), streamflow is reduced in both the season the warming is applied and in the subsequent cool season. The annual streamflow reduction is therefore larger than in either season separately. By contrast, with cool season-only warming (Figure 3c), streamflow increases during the cool season, but decreases during the subsequent warm season. The contemporaneous increase in streamflow and the subsequent decrease in streamflow partially compensate. The response to year-round warming (Figure 3a) is approximately the sum of the responses to warm-only and cool-only season warming applied separately (Figures 3b and 3c).

3.3. Projections of Future Warming

[21] Downscaled projections from 16 climate models using SRES A2 forcing [Maurer *et al.*, 2007] consistently exhibit more summer warming than winter in this region (Figure 1b). The greatest summer/winter asymmetry is found in the Columbia basin, with about 3°C warming in winter averaged over models but nearly 5°C in summer. The other basins show a pronounced seasonal asymmetry as well. Although the experiments described above with constant year-round warming of 3°C are reasonably consistent with the climate model projections, the climate model results can be more closely approximated by applying more warming in the warm season than in the cool season. To see what effect this has, we compare a “uniform experiment” (with 3°C year-round warming) to an “asymmetric experiment” with 2°C of cool season warming and 4°C of warm season warming (yielding the same annual warming). We find that Colorado River streamflow declines 16.0% in the uniform experiment compared to 18.6% in the asymmetric experiment. Columbia River streamflow has a 8.4% decline for uniform warming compared to a 9.8% decline for asymmetric warming. The expectation of more summer than winter warming combines with the greater sensitivity of streamflow to summer than winter warming to further reduce streamflow in these basins.

4. Discussion and Conclusions

[22] Within the major river basins of the western U.S., VIC simulations show a substantial decrease in annual streamflow in response to warm season (April through September) warming: −13.3%, −7.2%, −1.8%, and −3.6% for a 3°C warming in the Colorado, Columbia, northern and southern Sierra basins, respectively. Cool season warming, by contrast, gives annual changes that are mostly much smaller: −3.5%, −1.7%, −2.1%, and −3.1% respectively. Cool season warming stimulates greater streamflow immediately, which partly compensates for a subsequent decrease

in summer streamflow that happens because less water is available. Summer warming stimulates greater evapotranspiration and diminished soil moisture immediately, with no process during the following winter that can generate a compensating streamflow increase.

[23] The net response to year-round warming arises from the interplay of these seasonal responses, and varies greatly between basins. When 3°C year-round warming is applied to the Northern Sierra Nevada, increased cool season streamflow nearly compensates for the decreased warm season streamflow, yielding a net annual change of only −3.9%. At the opposite extreme, the Colorado River basin is dry enough that additional liquid water input to the system during the winter generates almost no additional streamflow immediately. Instead, the water is stored in the soil and becomes available for later loss via ET, so the net annual streamflow decline is highest of the four basins studied: −16.1% for 3°C of year-round warming. Such a reduction would likely result in water shortages on the Colorado River [Barnett and Pierce, 2009]. The Colorado River basin receives the highest warm season precipitation of the four basins, and 80% of the warm season precipitation is consumed by evapotranspiration, compared to 50% in the Columbia and Sierra Nevada (Figure S4 in Text S1).

[24] In contrast to the substantial variability between the basins’ annual streamflow response to a uniform year-round climate warming, the reduction in warm-season flow is much more consistent: −14.6%, −11.8%, −11.0%, and −14.3% in the Colorado, Columbia, northern and southern Sierra, respectively. This has important implications for a built water system that depends on melting snow to deliver water in the warm season [Barnett *et al.*, 2005]. Streamflow decreases during the season it is needed most will be substantial, and can only be compensated for by increased storage or decreased demand. A strong reduction of summer streamflow might also have significant impacts on the western U.S. ecology. For example, higher moisture deficits in summer might increase forest fire frequency [Westerling *et al.*, 2006].

[25] Previous work has generally focused, implicitly or explicitly, on the effects of winter/spring warming and associated shifts in runoff timing, because the immediate manifestations of warming in those seasons are large. However, these results emphasize the key role of future warm season warming in determining annual streamflow changes. Climate model projections generally show higher warming in the warm than in the cool season over the western U.S., typically by 1.5–2°C by 2100 under SRES A2 forcing scenarios [Maurer *et al.*, 2007]. This could accentuate the reductions that are indicated here, especially in the Colorado, which exhibits a high sensitivity to warming in the warm season. In river beds where warming leads to increased winter streamflow, flooding could become a greater problem [Dettinger *et al.*, 2004; Lundquist *et al.*, 2008; Das *et al.*, 2011]. The interplay of these seasonal responses is crucial to understanding how western U.S. rivers will respond to climate warming, and must be considered in adapting to climate change’s effects there.

[26] Our analysis does not consider precipitation recycling, so our results likely represent an upper bound to changes in runoff, although not necessarily to the seasonal asymmetry in warming’s effect on runoff. However the size of this effect, estimated from our simulated changes in

evaporation combined with recycling values from Anderson *et al.* 2009, suggest that our overall conclusion of the importance of warm season warming to reducing annual runoff is robust. Evaporation increases are greatest in spring, but moisture recycling peaks in summer, when all the basins (except the Columbia) have decreased evaporation, and hence less local recycling. Incremental precipitation also tends to be less efficient at generating runoff during the warm season, when generally the soil is drier and evaporative demand higher.

[27] The results presented here were obtained from one widely applied and well-documented macroscale hydrology model, VIC. Future work no doubt will investigate similar predictions using other land surface models. Also, our experimental design does not examine joint precipitation and temperature changes, or changes in the daily sequencing of *P* and *T*, both of which merit future investigation since climate change will likely alter both.

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