Environmental Research Letters

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To cite this article: Lauren M Foster *et al* 2016 *Environ. Res. Lett.* **11** 044015

Environmental Research Letters

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OPEN ACCESS

RECEIVED
26 October 2015

REVISED
23 February 2016

ACCEPTED FOR PUBLICATION
3 March 2016

PUBLISHED
13 April 2016

LETTER

Energy budget increases reduce mean streamflow more than snow–rain transitions: using integrated modeling to isolate climate change impacts on Rocky Mountain hydrology

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Keywords: Rocky Mountains, climate change, integrated modeling, snow to rain transitions, energy budget, hydrology, precipitation phase

Supplementary material for this article is available online

Abstract

In snow-dominated mountain regions, a warming climate is expected to alter two drivers of hydrology: (1) decrease the fraction of precipitation falling as snow; and (2) increase surface energy available to drive evapotranspiration. This study uses a novel integrated modeling approach to explicitly separate energy budget increases via warming from precipitation phase transitions from snow to rain in two mountain headwaters transects of the central Rocky Mountains. Both phase transitions and energy increases had significant, though unique, impacts on semi-arid mountain hydrology in our simulations. A complete shift in precipitation from snow to rain reduced streamflow between 11% and 18%, while 4 °C of uniform warming reduced streamflow between 19% and 23%, suggesting that changes in energy-driven evaporative loss, between 27% and 29% for these uniform warming scenarios, may be the dominant driver of annual mean streamflow in a warming climate. Phase changes induced a flashier system, making water availability more susceptible to precipitation variability and eliminating the runoff signature characteristic of snowmelt-dominated systems. The impact of a phase change on mean streamflow was reduced as aridity increased from west to east of the continental divide.

1. Introduction

More than one-sixth of the world's population depends on surface water supplies from snowmelt-dominated systems (Barnett *et al* 2005). These systems are complex, leading to inadequate observational networks to inform research on local micro-climatological processes that feed into large-scale flows (Barry 1994, Beniston *et al* 1997). As the climate warms, possibly more rapidly in these regions (Rangwala *et al* 2013), temperature increases will affect local hydrology. Impacts include a timing shift in snowmelt and peak flows (Stewart *et al* 2004a), decreases in the extent of snow cover (Cayan *et al* 2001, Regonda and Rajagopalan 2004), changes in peak soil moisture timing (Harpold and Molotch 2015), and increases in summer evapotranspiration (Christensen and Lettenmaier 2007, Goulden and Bales 2014). There are two main hydrologic drivers of these changes that are affected by temperature increases: (1) phase of precipitation will shift from snow to rain; and (2) available energy at the land surface increases. Teasing out the effects of phase versus energy changes within complex mountain systems is challenging, especially in observational and statistical studies where natural variability obscures individual drivers of change. Nonetheless it is critical to understand the impact of these two drivers on hydrologic partitioning in order to prepare for climate changes in snow-dominated regions.

Some recent efforts have focused on isolating these drivers in mountain regions. Berghuijs *et al* (2014a, 2014b), used a statistical analysis within a Budyko framework, quantifying moisture and energy limitations on evapotranspiration, to understand the impact of a snow to rain transition on mean streamflow, finding that this precipitation shift would decrease annual streamflow, which is in agreement with a site-specific, mechanistic study in Sweden (Bosson *et al* 2012). Previous work in this area has assumed (Laternser and Schneebeli 2003, Hamlet *et al* 2005, Mote *et al* 2005, Solomon 2007) or shown (Williams *et al* 2012) that phase will impact only the timing of runoff, not the quantity. Given that most studies of phase transitions have been conducted using different locations as proxies of environmental change, there are calls for work investigating the physical processes behind a phase change (Berghuijs *et al* 2014b, Pelletier *et al* 2015). Energy driven changes have also been shown to impact streamflow, with some studies demonstrating that reductions in streamflow are driven by increases in summer evapotranspiration (ET) more than by decreases in the total volume of precipitation (Christensen and Lettenmaier 2007, Goulden and Bales 2014) and that energy available for evapotranspiration is a critical component to water partitioning (Zapata-Rios *et al* 2015a, 2015b).

Despite conflicting results in previous studies, and the call for work on the physical processes resulting from snow–rain transitions, no modeling study has yet isolated the impact of energy-driven changes from phase changes. Integrated modeling provides a unique opportunity to probe the physical mechanisms behind these two climate impacts via controlled, hypothetical experiments (Weiler and McDonnell 2004, Maxwell and Kollet 2008, Kumar *et al* 2009, Ferguson and Maxwell 2010, Sulis *et al* 2010). The objective of this study is to identify the relative hydrologic sensitivity of continental mountain regions to changes in precipitation phase from snow to rain versus increases in surface energy fluxes due to warming. Using observed atmospheric forcing data, we isolate a phase transition from snow to rain by increasing temperatures only during precipitation events and compare results to uniform warming scenarios. This work is the first to explicitly separate these two hydrologic drivers using integrated modeling. The experiment provides insight into contradicting literature regarding the effects of a phase shift from snow to rain as well as to Colorado water managers about potential climate impacts east and west of the continental divide.

2. Model and experiments

2.1. Model construction

A two-dimensional hillslope was modeled using the platform ParFlow (PF) coupled with the Common Land Model (CLM). PF is an integrated hydrologic model that solves the 3D Richards' equation for unsaturated and saturated flow in the subsurface, and Manning's equation for overland flow (Kollet and Maxwell 2006). CLM, version 3.0 with additional updates, solves the energy budget at the land surface, resolving snow processes, vegetation processes, and evaporation (Dai *et al* 2003, Maxwell and Miller 2005, Kollet and Maxwell 2008, Ferguson *et al* 2016, Jefferson and Maxwell 2015, Jefferson *et al* 2015). Previous research has validated the physics of PF-CLM using site-specific observational data (Maxwell and Miller 2005, Maxwell and Kollet 2008, Atchley and Maxwell 2011, Ajami *et al* 2014, Condon and Maxwell 2014, Shrestha *et al* 2014, Maxwell *et al* 2015).

A 2D, idealized domain was simulated to represent a mountain hillslope transect. The hillslope is 500 m in the x-direction, 250 m in the y-direction, 60 m in the subsurface and discretized in 5 m cells in x (slope of 22°), one cell in y, and the subsurface is variably discretized into ten layers with higher resolution at the surface, decreasing with depth (figure 1). Subsurface characteristics are based on a review of hydraulic conductivity in mountain regions by Welch and Allen (2014). The layers are divided into soil, weathered bedrock, and fractured bedrock; hydraulic conductivity decreases with depth. Parameters for the model are listed in table 1.

2.2. Scenarios

We conducted 22 warming experiments (table 2) using observed atmospheric data from two locations. The first is compiled from the North American Land Data Assimilation System (NLDAS) at Pennsylvania Gulch, Colorado, west of the continental divide (elevation 3000 m) for September 2007 through August 2008 (Mikkelson *et al* 2013). The second location was chosen at a similar elevation on the North Fork of the Big Thompson River, east of the continental divide and compiled from NLDAS for the same period. The climate in the 2008 water year is typical of an average year for both regions, allowing for comparisons at similar elevations east and west of the continental divide. Atmospheric forcing was applied uniformly across each domain.

At each location 11 scenarios were run from combinations of four unique temperature alterations applied over three different seasons of warming. CLM uses a set threshold to determine the phase of precipitation (called t-critical), which was used to determine warming seasons. Below 2.5 °C, precipitation falls as snow in the model, above as rain. Though this is a simplification, in natural systems there are many mixed phase forms of precipitation, it provides a setting in which a modeling experiment separating rain and snow can take place. We defined three seasons in which to apply temperature increases: (1) full year warming, (2) snow season warming- the season in which weekly average temperatures fall below

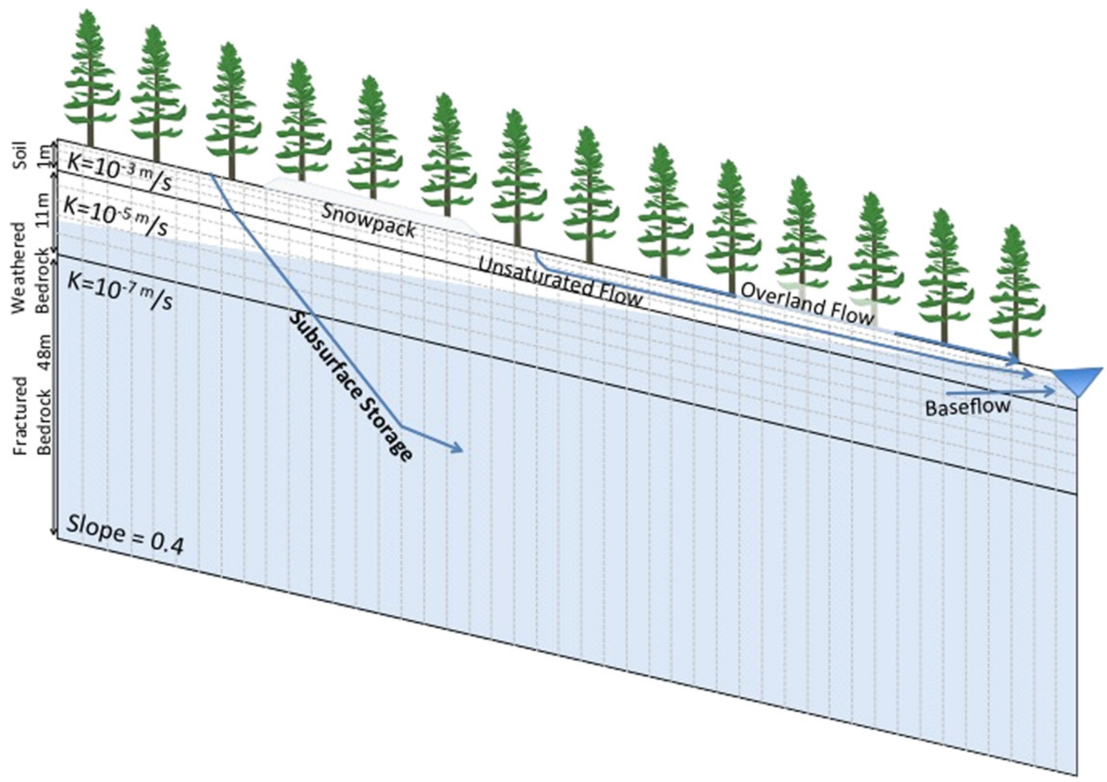

Figure 1. Conceptual model of 2D hypothetical hillslope.

Table 1. Model parameters.

Parameter	Value
Soil K	10^{-3} m s^{-1}
Weathered bedrock K	10^{-5} m s^{-1}
Fractured bedrock K	10^{-7} m s^{-1}
Porosity	0.399
Mannings coefficient	0.020 s m$^{-1/3}$
van Genuchten: α	1.12 m^{-1}
van Genuchten: n	2.48
Specific storage	10^{-5} m^{-1}

t-critical, and (3) transitional season warming- spring and fall transitions from first snowfall to mid-winter. The first two temperature alterations were uniform 2.5 °C and 4 °C increases applied in all three seasons. The second two alterations could only exist in a modeling environment and allow a detailed comparison of the physical processes that would result from a phase change- snow to rain- with no late-summer energy increases. In order to maintain typical winter behavior in these Phase Change experiments, code in the model was altered to prevent build-up of ice in the soil. These hypothetical scenarios augment observational studies attempting to separate phase and energy impacts in real systems. In the 'warm to t-critical' scenarios the temperature was increased to 2.5 °C during winter and transitional seasons, causing precipitation to fall as rain in the model. In order to better isolate phase changes from energy increases, the last alteration was to increase temperatures to *t*-critical only while precipitation was occurring (table 2). Integrated modeling provides the opportunity to hold many complex variables constant, in this case using a simplified domain and altering only air temperature, in order to tease out the climate mechanisms driving specific changes to the hydrologic cycle.

Out of our analysis of all 22 scenarios, two main temperature alterations from the Pennsylvania Gulch forcing, west of the divide, will be discussed in depth, then compared with the same two scenarios in the Big Thompson, east of the continental divide. We focus on the snow season scenario in which warming was applied only during precipitation events, hereafter referred to as the *Phase Change* scenario, and the full year uniform warming of 4 °C, hereafter referred to as the *Warming* scenario, because these scenarios are most effective at separating a phase change from snow to rain and an increase in surface energy input due to warming. Figure 2(a) shows that increasing the temperature during precipitation events only, the *Phase Change* scenario, raises the temperature by an average of 0.4 °C, but that the temperature change is minimized relative to the 4 °C increase in the *Warming* scenario. Figure 2(b) shows that the *Warming* scenario impacts phase, with snowfall decreasing from 70% of precipitation to 63%, but again this is minimized

Table 2. Ten warming scenarios and a baseline scenario. Each temperature alteration was applied to two separate forcing data sets, one to the west of the continental divide at Pennsylvania Gulch and one to the east at the North Fork of the Big Thompson River. The bold scenarios will be discussed in depth in the paper.

		Warming methods			
	Baseline scenario	Warming +2.5 °C	Warming +4 °C	Warm to rain forcing	Warm only while precipitating
Season	Full Year	'2.5 °C FullYear'	**'4 °C Full Year'**		
	Snow (12 October–26th May)	'2.5 °C Snow'	'4 °C Snow'	'Warm Snow'	**'Rain snow'**
	Transitional (12 October–1st December and 1st March–26th May)	'2.5 °C Transition'	'4 °C Transition'	'Warm transition'	'Rain transition'

relative to the *Phase Change* scenario, which reduces snowfall from 70% of precipitation to 3%.

3. Results and discussion

The *Baseline* scenarios demonstrate expected dynamics of a watershed in snowmelt-dominated, mountainous terrain in a manner also observed by Berghuijs *et al* (2014a). Streamflow (Q) is relatively stable until an early summer peak during snowmelt. The overall pattern of ET follows temperature trends-high in the summer, low in the winter and about equal in the spring and fall- and also increases during precipitation events. Through the fall and winter, Q is sustained by baseflow, so the storage change (ΔS) is negative. There is a long period of recharge from the start of snowmelt in the spring into late summer, which is reflected in the positive ΔS during this period (figure 3).

3.1. Shifts in seasonal water budget behavior—Pennsylvania Gulch

Given that climate change has been shown to impact components of the water balance differently within each season (Jasechko *et al* 2014), figure 4 shows volumetric components of the seasonal water balance at the Pennsylvania Gulch domain. Streamflow is calculated using Mannings equation for overland flow at the outlet of the domain, storage includes all saturated and unsaturated water in the subsurface, and ET includes evaporation from the canopy and ground surfaces, transpiration, and sublimation. Precipitation is the same for all three scenarios as only temperature was altered in this experiment. Comparing the *Baseline* (figure 4(a)) to the *Phase Change* (figure 4(b)), we see that the progression of net seasonal storage changes are shifted one season earlier. In the *Baseline* scenario storage decreases through the fall and winter, while baseflow sustains streamflow, and in the spring and summer storage increases as snowmelt infiltrates into the subsurface. With no snowpack in the *Phase Change* scenario, recharge occurs mostly through the winter, when ET is low and more precipitation falls. Recharge decreases through the spring, switching to net storage reductions in the summer- one season earlier than in the *Baseline* scenario. Without snowpack storage, Q, ΔS, and ET all respond quickly to the amount of water available from precipitation in a given season.

In the *Warming* scenario (figure 4(c)), instead of a shift in the ΔS pattern, simulations exhibit reduced Q, increased ET, and a more dynamic storage response. In the *Baseline* scenario (figure 4(a)), seasonal balances demonstrate a typical mountain hydrograph in which snowmelt increases storage from the spring into the summer. ET reaches a maximum during the summer, but enough water is stored in the system to simultaneously provide streamflow and positive ΔS until the fall season. In the *Warming* scenario, figure 4(c), storage only increases in the spring. This reduction in storage implies that streamflow is sustained by baseflow for three full seasons. In conjunction with the increase in ET, this reduces Q throughout the year. Summertime ET is nearly twice the magnitude of summertime Q in the *Warming* scenario. The ubiquitous increase in ET is a main driver of the 29% reduction in yearly mean streamflow in the *Warming* scenario, a result that is consistent with predicted streamflow declines in the 21st century across the Western United States (Milly *et al* 2005).

3.2. Temporal patterns of climate impact scenarios-Pennsylvania Gulch

Figure 3 shows the same three components of the water balance, Q, ET, and ΔS, as a weekly-averaged time series for scenarios west and east of the divide. Consistent with the seasonal water balance analysis, it is clear that the *Phase Change* scenario shifts the entire pattern of Q and ΔS to a flashy system where signals from single storms move through and out of the system quickly. The *Warming* scenario follows a typical snowmelt-dominated hydrologic pattern, but with peak discharge occurring 28 days earlier and a 16% increase in amplitude of peak flow, figure 3(d). This timing shift is within the range published by Stewart *et al* (2004a, 2004b) but higher than predicted for elevations of 3000 m in other historical analysis (Nash and Gleick 1991, Christensen *et al* 2004, Regonda and Rajagopalan 2004). An earlier and larger release of snowpack storage reduces summer and fall Q when drought risks are already high, with

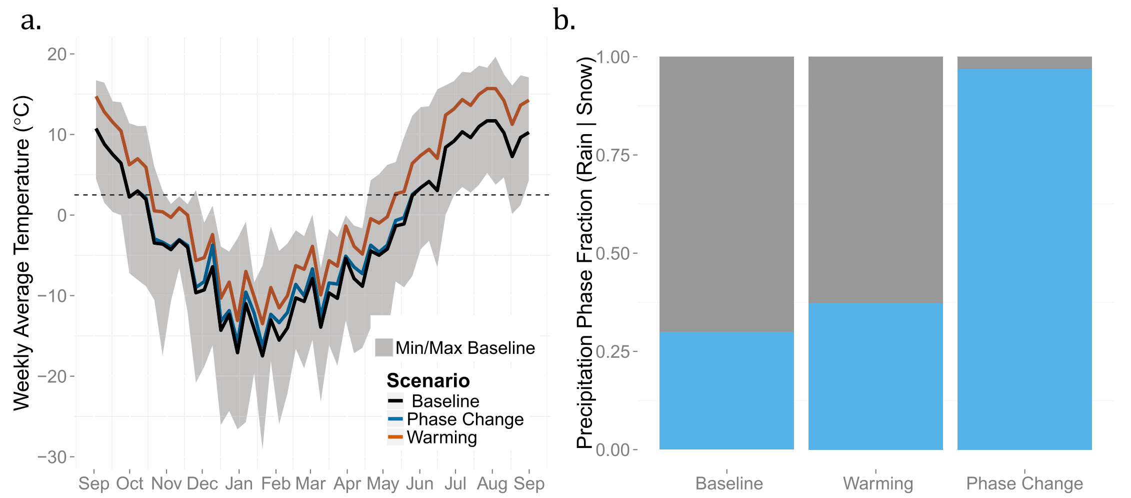

Figure 2. Temperature alterations (a) and precipitation phase impacts (b) for scenarios at the Pennsylvania Gulch site for 2007–2008 in degrees Celsius. Dashed line at 2.5 °C in (a) shows the temperature below which CLM interprets precipitation as snow (*t*-critical). Energy fluxes at the land surface due to temperature increases (*Warming* scenario) are isolated from snow-rain transitions (*Phase Change* scenario) due to warming.

implications for total water supply as well as potentially increased flood risks due to the higher amplitude of peak flow in early spring.

Both the *Phase Change* and *Warming* scenarios reduce mean streamflow through the year on the west slope. The initial reduction in streamflow is due to reduced baseflow from the steady state storage capacity, which is 2.3% less than *Baseline* in both *Phase Change* and *Warming*. Despite similar losses in steady state storage, *Warming* has higher ET throughout the year, driving further reductions in streamflow. Mountain regions depend on two major storage components, subsurface storage and snowpack. In the *Phase Change* scenario it is critical to track changes in subsurface storage as groundwater becomes the only store of water to buffer supply from seasonality in precipitation input. As the climate shifts to a new steady state, impacts cannot be understood by partitioning water into ET and *Q* alone. Changes in the subsurface storage component must also be considered to explain long-term shifts in streamflow. Reductions in stored subsurface water, such as those simulated here, will reduce the system's ability to maintain levels of baseflow during drought seasons.

In the ET time-series, figure 3(e), the *Phase Change* and *Warming* scenarios demonstrate regional moisture-limitation. At the beginning of the simulation, September 1st, ET is the same in *Baseline* and *Phase Change*, but the higher temperatures increase ET in *Warming* by approximately 15%. Moving into the snow season, ET from *Phase Change* is greater than in the other two scenarios. This is due to most precipitation falling as rain, minimizing surface albedo and retaining moisture in liquid form. The ET pattern is dominated by canopy evaporation given that the domain is fully vegetated, but a closer look at the components of ground evaporation, sublimation, and transpiration (figure S1) demonstrates the shift in seasonality in the *Phase Change* scenario from *Baseline* and *Warming* that drives the increase in total ET through the winter. When snowmelt begins in the *Warming* scenario in early May, ET crosses the *Phase Change* scenario and remains approximately 30% higher than *Baseline* until July. From May through July the difference between *Baseline* and *Warming* ET is double what it was in the fall. This result corroborates previous research finding that summer ET is more significant for total streamflow volumes than reduced precipitation input (Christensen and Lettenmaier 2007) as well as demonstrating the dominance of ET after snowmelt, when more moisture is available, than other times of the year. Previous research has found that warming reduces ET in mountains (Barnett *et al* 2005, Zapata-Rios *et al* 2015b); in these studies the moisture limitation overcomes higher energy input from warming temperatures. All of our scenarios increased ET, showing that both moisture and energy are available, in varying amounts, throughout the year.

The ΔS time-series, figure 3(f), exhibits similar patterns to the streamflow time-series, but with changes almost one order of magnitude higher than streamflow. For this reason it is imperative to diagnose subsurface behavior in addition to ET and *Q* partitioning in understanding the mountain water balance. In the *Phase Change* scenario, storage oscillates regularly between gains and losses based on the precipitation input. It is clear from comparing precipitation (figure 3(a)) to ΔS (figure 3(f)) that even the subsurface system exhibits flashy responses to precipitation in *Phase Change*. This volatile storage signature provides little capacity to buffer yearly water availability from interannual variability. The *Warming* scenario leaves a narrow window for recharge to occur by shortening the time of snowmelt recharge by 1.5 weeks

Figure 3. Precipitation (mm d^{-1}) and temperature (°C) for the *Baseline* scenarios (a), (b), (g), (h), snow-water equivalent (mm) and weekly-averaged time-series of streamflow (Q), evapotranspiration (ET), and change in storage (ΔS) all in mm d^{-1}.

as compared to *Baseline*, as well as shifting peak recharge 2 weeks earlier in the year.

3.3. Regional differences: west versus east of the continental divide

Figure 3 demonstrates the driving climatic differences and resulting water balance impacts between regions west (a)–(f) and east (g)–(l) of the continental divide. On the east side, total precipitation is reduced by 12%, the snow–rain ratio is reduced by 15%, and temperature is, on average, 4 °C warmer when compared to the west forcing data. The *Phase Change* and *Warming* scenarios are able to isolate snow–rain transitions from temperature increased energy fluxes on the east slope with *Phase Change* increasing temperature by only 0.5 °C, similar to the effect on the west side. The

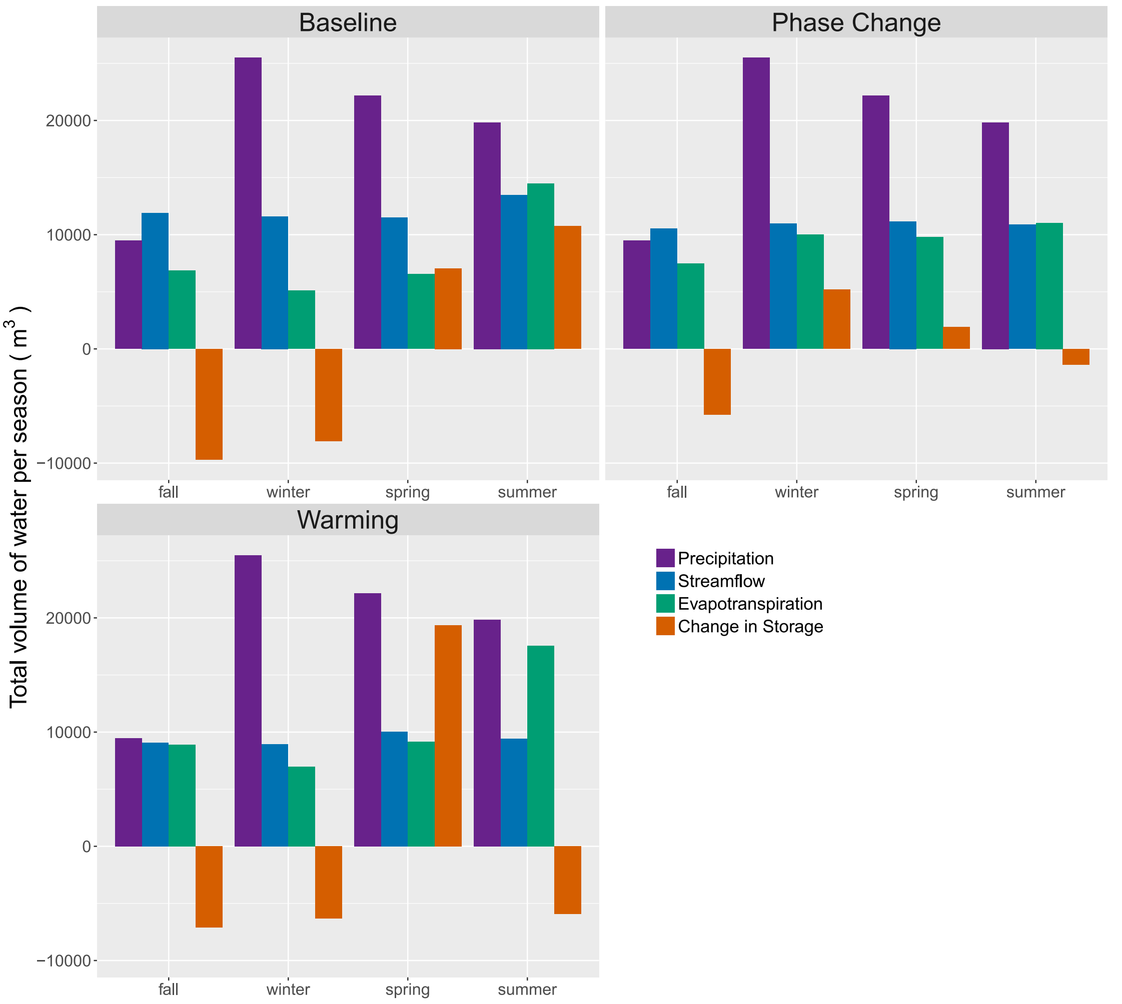

Figure 4. Net water balance components: precipitation, streamflow (Q), evapotranspiration (ET), and change in storage (ΔS) plotted seasonally for each scenario at the Pennsylvania Gulch site. As compared to *Baseline* (a), the *Phase Change* scenario (b) alters the pattern of storage fluctuations. The *Warming* scenario (c) exhibits an extreme peak of recharge during a short snowmelt, but the summer demonstrates increased drought risk with lower Q, very high ET, and significant storage losses.

Warming scenario reduces snowfall to 35% of precipitation from the 56% in *Baseline*. This reduction is 14% greater on the east side than the west. Due to reduced precipitation input, all components of the *Baseline* water balance on the east slope exhibit lower magnitudes throughout the year (figures 3(h)–(j)). In Colorado the majority of the population resides east of the divide, and much of the water supply is transferred from the western slope where more precipitation occurs (Carlson 1975). Understanding impacts on each side of the divide has implications for management of water systems that will behave differently under similar climatic changes.

Comparing total mean streamflow (Q) from west to east of the divide (figures 3(d) and (j)), the *Phase Change* difference from *Baseline* is reduced from 18% on the west side to 11% on the east side. In the *Warming* scenario the difference from *Baseline* is reduced half as much, from 23% to 19% west to east, because ET remains high both west and east at 29% and 27% respectively. The streamflow reduction in the *Phase Change* scenario is due to much higher winter ET on the west slope, leading to a 26% increase in yearly mean ET, as opposed to the 16% ET increase on the east side In both regions the *Phase Change* Q patterns follow precipitation input, with streamflow responding quickly to storm events, instead of following a traditional snowmelt dominated hydrograph. Only one study that includes Rocky Mountain climates found that phase changes will reduce streamflow (Berghuijs *et al* 2014b), while others have found or assumed that phase changes will affect timing but not total mean streamflow (Laternser and Schneebeli 2003, Hamlet *et al* 2005, Mote *et al* 2005, Solomon 2007, Williams *et al* 2012). Our results demonstrate both timing changes and reduced streamflow. More notably they show that, as aridity increases from west to east, precipitation phase has less impact on mean streamflow, though it affects timing in both scenarios.

Similar to the 4-week shift in peak streamflow on the west slope during the *Warming* scenario, in the east the peak is shifted 3 weeks earlier (figure 3(j)). However, the eastern Q in the *Warming* scenario resembles a combination of *Phase Change* and *Baseline* patterns, with reduced amplitude in the main peak, and more significant secondary peaks. This is likely due to the east forcing having a reduced snow–rain ratio of 15%, combined with less precipitation occurring during winter. On the west side (figure 3(d)) the scenarios isolate phase changes more completely due to consistently colder temperatures. Importantly, the west side *Warming* scenario demonstrates a 16% increase in peak amplitude of flow from the warmer spring and summer temperatures, but on the east side peak amplitude is reduced by 33%. In the colder, wetter climate of the west side, warming increases amplitude of peak flow and potential flood risks, while these risks are mitigated by higher aridity, less snowpack storage and a greater reduction in snowfall from the *Warming* scenario, on the east side of the divide. The ΔS *Warming* pattern (figure 3(l)) shows a similar transition from *Baseline* patterns to *Phase Change* patterns due to combined impacts of phase and warming on the east side of the divide. The high amplitude recharge peak on the west side, discussed in section 3.2, is, like Q, reduced on the east slope.

The ET signal west of the divide (figure 3(e)) follows the precipitation input (figure 3(a)) closely, indicating that it is a mostly moisture, not energy, limited system because ET responds more quickly to moisture inputs than temperature increases. On the east side there is less precipitation input, which reduces moisture available for ET. The ET pattern demonstrated by scenarios on the west side is repeated on the east side (figures 3(e) and (k)) where *Warming* exhibits approximately 15% higher ET until snowfall begins. Unlike the west side, on the east side the *Phase Change* scenario does not dominate as much in the winter, due to the limited and more variable precipitation input in a more arid climate. The spring and summer patterns are similar on both sides of the divide, with the *Warming* scenario dominating from the start of snowmelt onward. A notable difference on the arid, east side however, is that the summer ET peak is 18% lower than on the west side, even with the significantly increased temperatures, because ET is limited by reduced soil moisture available on the east slope.

The scenarios used here are specific to semi-arid, continental mountain ranges, but they can also serve as proxies for other snow-dominated mountain regions; providing insight about the range of impacts observed in complex terrain despite significant variability. Simulations of phase changes from snow to rain can be compared to hydrologic patterns of mountainous regions at lower elevations where a phase transition from snow to rain is occurring, while energy budget changes can be compared to arid climates and high mountain zones in which phase changes are likely to be minimal in the coming century. For example, in the *Warming* scenario east of the divide, nearest of our scenarios to the snow–rain transition, there are 41 fewer days of snow cover than in the *Baseline*. This is comparable to a 36 day reduction modeled in the Pacific Northwest near the snow–rain transition line for 2 degree of warming (Sproles *et al* 2013). This may indicate that the east slope *Warming* scenario could be compared to some climate impacts in other ranges near the transition line. However, a maritime climate is characterized by higher humidity and different weather patterns, limiting the ability to extrapolate fully to that region. These scenarios also serve as proxies for seasonal warming impacts in mountains. Warm summers will reflect changes due to an energy budget increase as they will have a minimal impact on snow–rain ratios, while warm winters will likely decrease snow–rain ratios and exhibit dominant impacts from a phase change.

4. Conclusions

Our results suggest that land-energy changes have a larger effect on total water available for use in the mountain hydrologic system, both in surface and groundwater, than a phase change from snow to rain. This reduction in usable water is mostly driven by an increase in summer evapotranspiration due to warming. A phase change introduces a flashier system, minimizing the ability to buffer water availability from interannual variability and completely changing the hydrologic pattern. Given that most reservoirs in snow-dominated basins depend on snowmelt timing and snowpack storage for replenishment, phase changes may require more involved reservoir management to compensate for the loss of snow storage (Dettinger and Anderson 2015). The impact on streamflow from a phase change is reduced from a 18% to 11% decrease from the west side of the continental divide to the east side, suggesting that phase change impact on streamflow depends on moisture and energy being available simultaneously. 4 °C warming in the system reduces streamflow by 23% west of the divide and by 19% east of the divide, demonstrating that energy budget has a large impact on streamflow in both regions. Warming intensifies summer evapotranspiration enough that baseflow must sustain streamflow for an additional season throughout the year, shifting the transition to groundwater fed streams from fall to summer.

The magnitude of these two changes will vary with location and climate so it is critical to separate energy driven impacts from phase driven impacts in order to understand and prepare water management and policy for environmental change. Future modeling work will test these scenarios at the watershed scale and in real domains. At the same time, signatures from our numerical scenarios should be compared to observational data to analyze the effectiveness of using these scenarios as proxies for seasonal warming and different mountain regions across the globe. This experiment is able to isolate hydrologic drivers and minimize noise from complex natural systems using controlled scenarios, possible only in a modeling environment.

Instead of resolving small scale feedbacks in the system, such as energy budget differences between north and south facing slopes (Hinckley *et al* 2014), these results serve as endmembers in efforts to tease out causes and effects of climate signals in continental mountain regions.

Acknowledgments

This study was funded by a National Science Foundation IGERT (Integrated Graduate Education and Research Traineeship) under award DGE-1069333 (to L Foster and R Maxwell) and Dean's Fellowship from the Colorado School of Mines (to L Foster).

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