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Snowpack sensitivity to perturbed climate in a cool mid-latitude mountain catchment

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Abstract:

There is great interest in ascertaining the degree of climate change necessary to induce substantial changes in snow accumulation and ablation processes in mountain headwater catchments. Therefore, the response of mountain snow hydrology to changes in air temperature and precipitation was examined by simulating a perturbed climate in Reynolds Mountain East (RME), a headwater catchment with a cool mountain climate in Idaho, USA. The cold regions hydrological model was used to calculate snow accumulation, wind redistribution by blowing snow, interception by forest canopies, sublimation and melt for 25 seasons in RME. The uncalibrated simulations of the highly redistributed snow water equivalent compared well to measurements. Results showed that with concomitant occurrence of warming (5 °C) and precipitation change (±20%) in RME, the peak seasonal snow accumulation decreased by 84–90%, snowmelt decreased 51–79%, rainfall to total precipitation ratio increased from 30% to 78%, and overwinter blowing snow transport and sublimation losses from intercepted snow, the snow surface and blowing snow decreased dramatically. Warming causes an increase in inter-water year snowcover variability but a decrease in spatial snow accumulation variability. When warming exceeded 1 °C and a precipitation increased by less than 20%, the peak snow accumulation declined dramatically. The results contrast with those from further north along the North American Cordillera in Yukon, Canada, where the impacts of similar warming on alpine snow can be partly compensated for by concomitant increases in precipitation of less than 20%.

KEY WORDS Reynolds Creek; snow hydrology; climate change; Cold Regions Hydrological Model; mountain hydrology

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INTRODUCTION

Snow cover extent in the northern hemisphere has shrunk 5.4% over the period from 1972 to 2006 (Déry and Brown, 2007), especially in March (7%) and in April (11%) when loss of snow cover is associated mainly with ongoing warming (Brown and Robinson, 2011). Air temperature increases are predicted to exceed 2 °C by 2040 in regions such as Canada and Eurasia and by 2100 over the globe, compared with the period from 1850 to 1900 (Joshi *et al*., 2011). Nogués-Bravo *et al*. (2007) investigated the climate warming projection on mountain systems for the end of the 21st century and found that these regions are expected to warm even more than other systems, by about 2.8 °C in temperate areas and 5.3 °C in northern latitudes. This will alter the snow dynamics, hydrological mass balance and water flux in mountainous regions. These areas are ecologically and hydrologically important, as they are hotspots for biodiversity due to the strong elevation and temperature gradients and key zones for runoff generation due to orographic precipitation and steep topographic gradients (Beniston, 2003; Bales *et al*., 2006). Alpine snow hydrology generates the majority of river flows from high mountains in much of Europe and North America, and these rivers provide water supply for vast downstream populations (Fang *et al*., 2013; López-Moreno *et al*., 2013). The distinctive cold season nature of alpine climates means that processes of blowing snow redistribution, sublimation, melt, infiltration, evaporation and runoff over and through frozen and unfrozen soils govern the generation of spring and summer flows from non-glacierized high mountain catchments (Pomeroy *et al*., 2012). Climate warming is evident in many alpine catchments and is expected to proceed further and to threaten the ecological and hydrological integrity of these regions (Intergovernmental Panel on Climate Change, IPCC, 2014; Harder *et al*., 2015). Mountains control direct runoff and erosion rates as they are susceptible to rapid weather changes over short distance and time scales (Beniston, 2003). The higher sensitivity of snow and permafrost to climate change in alpine regions makes alpine basins appropriate study areas for investigating

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climate change impacts on the hydrological cycle (Bunbury and Gajewski, 2012).

The conventional approach for investigating hydrological response to climate change is to apply climate model projections under different greenhouse gas emission scenarios. This is challenging in alpine regions, for the following reasons: (1) hydrological state variables respond uncertainly to different sets of the gridded atmospheric driving data in a given mountainous catchment (e.g. Eum *et al.*, 2014); (2) projected changes in regional precipitation are very uncertain, because of the coarse spatial resolution of the atmospheric circulations that are responsible for most precipitation events (Shepherd, 2014); (3) orographic complexity in mountainous terrain adds uncertainty to alpine precipitation and temperature simulations (Barry, 1992); (4) most of the climate models used for future projections, even those that are dynamically downscaled by regional climate models (RCMs) (Fowler *et al.*, 2007), still show large simulation biases when compared with current control conditions (Maraun *et al.*, 2010), even though increased resolution of RCMs output has reduced these biases.

Alternatively, sensitivity analysis can be conducted on different modelling components including observations, parameters and structures, to either identify and quantify the sources of model uncertainties or investigate the response of hydrological processes to the changes. Wilby (2005) recommended that sensitivity analyses can help in quantifying hydrological uncertainty in climate change impact studies. One method for doing this is the change factor method, which examines the change in a variable for a change in air temperature due to global warming (Fowler *et al.*, 2007). A previous study by the authors explored the sensitivity of Wolf Creek Research Basin (WCRB), Yukon, Canada (Rasouli *et al.*, 2014) to changes in air temperature and precipitation and found that snow hydrology was more sensitive than streamflow hydrology to climate warming. Even though increases in temperature reduced the effectiveness of snow redistribution and sublimation processes and accelerated the timing of snowmelt and snow depletion, spring snowmelt runoff was reduced. In this high-latitude environment, concomitant increases in precipitation of 20% could compensate for much of the impact of warming up to 3 °C.

The objectives of this paper are the following: (1) to examine the degree of sensitivity of snow regimes in a mid-latitude mountain basin to perturbed climate and (2) to compare the sensitivity of mountain snow regimes in a cool climate to those in a cold climate. The well-studied Reynolds Creek Experimental Watershed in Idaho, USA, is used as the cool climate basin, and the climate sensitivity of its snow regime is compared and contrasted with the cold climate WCRB. The approach investigates changes in snow accumulation [snow water equivalent (SWE)] using a sensitivity analysis to meteorological forcing inputs. The sensitivity analysis is restricted by plausible ranges of future precipitation and air temperature from recent climate model projections.

METHODS

Study sites and data sources

The primary study area chosen for this research is the Reynolds Mountain East (RME) basin, located in the Owyhee Mountains. It is one of the headwater catchments of the Reynolds Creek Experimental Watershed (RCEW), approximately 80 km southwest of Boise, Idaho, USA (Figure 1). The RME basin is small, with an area of 0.38 km^2, and it has two primary meteorological stations, representing a wind/topographically sheltered and an open area (Reba *et al.*, 2011a). The elevation varies between 2028 and 2137 m above sea level. RME has a seasonally cool and wet mountain climate, with a total mean water year precipitation of 858 mm, of which 70% falls as snow, and a water year mean air temperature of 5.2 °C averaged over 1984–2008. This basin was chosen for a detailed modelling study as it is densely monitored and has well-studied parameters that can be applied to develop a physically based snow model (Hanson, 2001; Hanson *et al.*, 2001; Marks *et al.*, 2001; Seyfried *et al.*, 2001; Slaughter *et al.*, 2001). It is also described in detail and has a published and freely available 25-year modelling data set (see Reba *et al.*, 2011a). It was selected to investigate the climate change impacts on hydrological processes in mountain watersheds with a cool climate. The collected data include hourly air temperature, relative humidity, wind speed, precipitation (corrected), shortwave radiation, and longwave radiation. A snow pillow near the sheltered site measures SWE, which can be used for diagnostic purposes. Snow accumulation is thought to be enhanced at this site because of the impact of topographic and vegetation sheltering on wind redistribution (Reba *et al.*, 2011b). However, Winstral and Marks (2014) showed that over recent snow seasons (2001–2012), detailed ground measurements indicate that snow pillow has represented the basin-wide SWE adequately. Distributions of vegetation, soils and SWE show great variability in the RME (Seyfried *et al.*, 2009; Kumar *et al.*, 2013; Winstral and Marks, 2014). Six main vegetation types are present, ranging from grass to mountain sagebrush through riparian willow, aspen and conifer trees. Basic characteristics of the study area, including dominant land cover, elevation of representative stations and soil type, are defined in the hydrological model.

Potential hydrological responses to warming and precipitation changes from a mountainous subarctic basin with land covers ranging from boreal forest to shrub

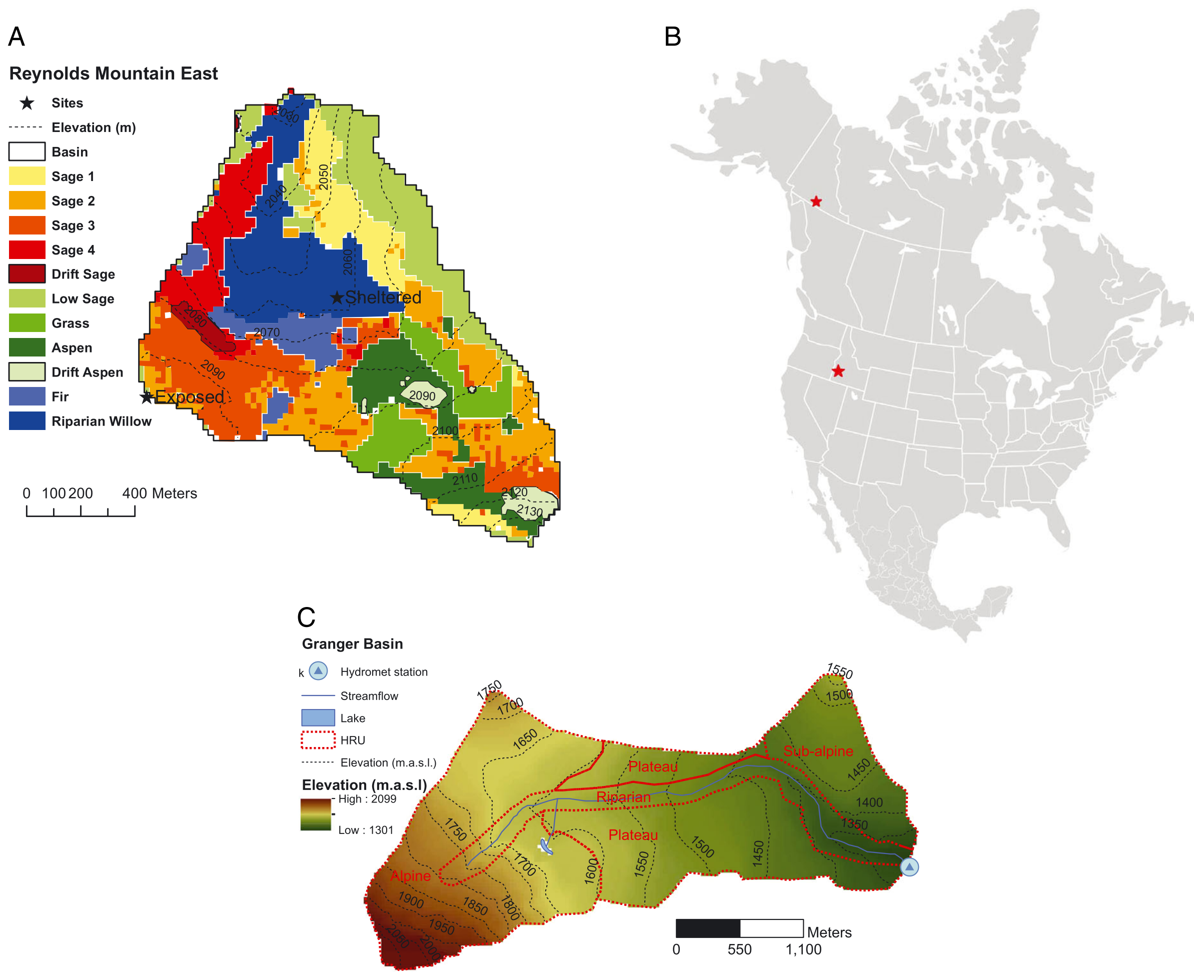

Figure 1. Topography, hydrological response units and location of sheltered and exposed weather stations in Reynolds Mountain East Basin, located in the highlands of the Reynolds Creek Experimental Watershed, Idaho, USA (regenerated from data presented in Newman *et al*., 2014) and topography and hydrological response units (HRUs) in Granger Basin located in the Wolf Creek Research Basin, Yukon, Canada

tundra to tundra were already investigated in a modelling study of Wolf Creek Research Basin (WCRB), Yukon Territory, Canada (Rasouli *et al*., 2014). The climate change impacts on alpine snow processes in central and northern parts of the North American Western Cordillera, RME and Granger Basin (GB), an alpine headwater sub-basin of WCRB with a colder climate, were selected for comparison with RME (Figure 1). GB (60°32.79′N, 135° 11.08′W) in southern Yukon, Canada, has a drainage area of 6.6 km^2 with a very cold subarctic continental climate characterized by a large temperature range and low precipitation (water year precipitation is 260 mm); it is underlain by a permafrost layer of 15 to 20 m thickness (Quinton *et al*., 2005). Elevation ranges between 1300 and 2100 m in this sub-basin and vegetation includes shrub tundra in riparian and alpine tundra in higher elevations. The GB was divided into four HRUs with varying physiography and vegetation including north facing and south facing with alpine tundra, plateau and riparian with shrub tundra land cover. Data sources and basin physiography are described in detail by Rasouli *et al*. (2014).

Hydrological modelling

To assess the impacts of climate change and variability on the cold regions hydrological cycle, models require a full set of physically based representations of hydrological processes. To assess the impacts on the snow regime in alpine regions, these include direct and diffuse radiation to slopes, longwave radiation in complex terrain, intercepted snow, blowing snow, sub-canopy turbulent and radiative transfer, sublimation and energy balance snowmelt. In this research, the alpine snow regime was studied using the Cold Regions Hydrological Modelling (CRHM) platform (Pomeroy *et al*., 2007), which represents all of the aforementioned processes. Modules selected for this study included routines to downscale meteorological inputs (Pomeroy *et al*., 2007); estimate precipitation phase (Harder and Pomeroy, 2013); estimate solar radiation on slopes (Ellis and Pomeroy, 2007);

estimate long-wave radiation (Sicart *et al*., 2006; Pomeroy *et al*., 2009); calculate snow interception and sublimation in forest canopies and sub-canopy radiation and turbulent transfer to underlying snowpacks (Ellis *et al*., 2010); calculate blowing snow transport and sublimation for simulating snow redistribution (Pomeroy and Li, 2000); and calculate energy balance snowmelt (Reba *et al*., 2011b). For more details on each module, refer to Rasouli *et al.* (2014), Fang *et al*. (2013) or Pomeroy *et al*. (2012). The energy and mass balance snowmelt model (Snobal) developed by Marks *et al*. (1999) was employed in the CRHM platform. Figure 2 illustrates all energy and mass fluxes defined in Snobal. This module conceptually divides the snowpack into two layers: surface-active layer and lower layer, and solves for the temperature and equivalent water depth per unit area in both layers. Snowmelt from each layer is estimated when the input energy exceeds the energy required to warm the snow cover temperature to freezing (0 °C).

The temporal resolution of the CRHM model is hourly in this study, whilst the spatial resolution is that of the hydrological response unit (HRU). The model operates on the internally relatively homogeneous HRUs shown in Figure 1, which are spatially segregated based on surface physiographic information relevant for hydrological model parameterization including vegetation cover, topography, soil depth and layers, adapted from Newman *et al*. (2014). The HRUs are developed from a multivariate *a priori* classification, an approach for capturing the variability of snow cover and snow depth within the catchment. This method was found to be superior to approaches in which grids are segregated based on elevation bands and vegetation types without considering sub-grid variability. Mountain sage is the dominant vegetation in RME and because of its higher variability was disaggregated into five HRUs based on a wind-sheltering index. Drift HRUs in aspen and sage vegetation are in topographic depressions downwind of slope breaks where deep snow accumulates (Newman *et al*., 2014). For better parameterization and evaluation of snow accumulation and ablation in RME, HRUs were categorized into four groups to analyse different snow regimes: blowing snow sink and source, and intercepted and sheltered snow. The division of a catchment into source and sink areas for blowing snow based on topographic exposure and vegetation height was proposed by Pomeroy *et al*. (1997) as a way to implement a distributed blowing snow model in a shrub covered arctic catchment. Blowing snow sink HRUs include not only drift HRUs but also riparian and tall sage HRUs. Therefore, the drift aspen, drift sage, willow and sage 4 are considered sink HRUs, and other short vegetation HRUs were grouped as source HRUs. The division of forested landscapes into those that are subject to interception and subsequent sublimation of intercepted snow (evergreens) and those that are cleared or have minimal winter interception capacity was proposed by Pomeroy *et al.* (2002). The fir forest has substantial canopy interception capacity in winter and so is considered a forest HRU with interception, and the aspen forest and gap HRUs have neither blowing snow nor intercepted snow fluxes and so are sheltered HRUs.

Estimation of the parameters in the RME basin was based on previous studies in RME and other headwater basins in RCEW and similar snow-dominated basins. Parameters have been adapted, including those that represent the characteristics of vegetation across the

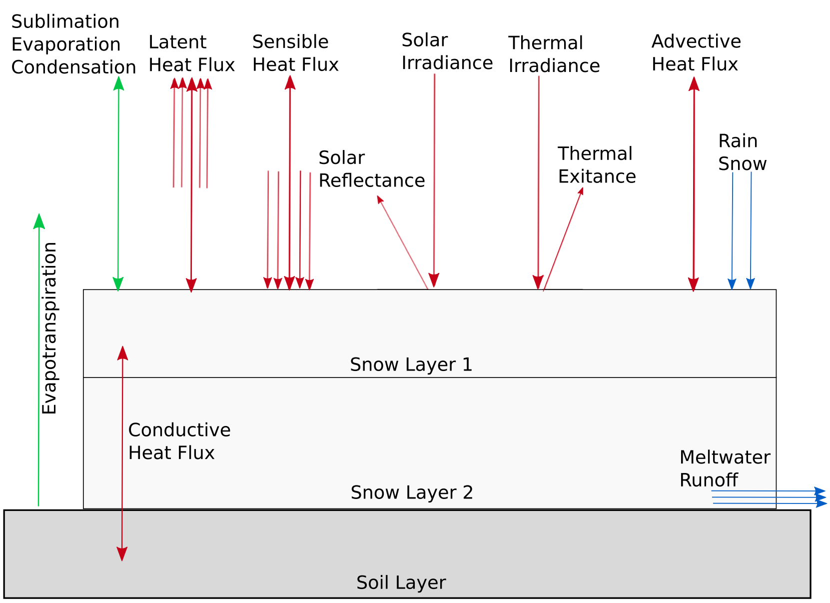

Figure 2. Schematic view of the energy and mass balance snow melt model (Snobal) with all input and output fluxes after Marks *et al.* (1999). Coloured arrows denote flux types; red is energy exchange with snow, green is energy exchange with water and blue is mass exchange

catchment and the parameters of snowmelt and blowing snow modules (Table I). Blowing snow was inhibited for the sheltered HRUs. Initial soil temperature was measured by soil thermocouples prior to the major snowmelt. Thermal conductivity was set to $1.65\,\mathrm{J\,m^{-1}\,s^{-1}\,K^{-1}}$, a value for wet sand taken from Oke (1978). Hourly snow water equivalent (SWE) time series recorded from a pillow in sheltered site (Figure 1) was used for evaluating the CRHM model developed in this study.

Sensitivity analysis

The temporal distributions and means of locally observed forcing data and RCM outputs vary substantially from each other, such that RCM outputs cannot be directly applied in hydrological studies without bias correction and downscaling. However, bias corrections and downscaling modify the spatio-temporal fields of meteorological variables in ways that are not always sufficiently justified and so add uncertainty to the simulations. To avoid this uncertainty, the original long observational datasets were perturbed using changes whose ranges fit within those projected under different emission scenarios (IPCC Special Report on Emission Scenarios, 2014) and to some extent by representative concentration pathways (Moss *et al.*, 2010). These sensitivity experiments were conducted by changing both air temperature and precipitation for control and future periods, whilst holding relative humidity constant. Climate change time series for the experiments were obtained by perturbing historical air temperature time series by 1 °C intervals from 0 to 5 °C (6 states) and precipitation observations by 10% intervals from 80% to 120% of the current precipitation (5 states), a combination of 30 scenarios of perturbed climate. The case of no change in air temperature and precipitation ($\Delta T = 0\,°\mathrm{C}$; $P = 100\%$) represents the historically averaged observed data over the 1984–2008 control period for RME and over the 1993–2011 period for GB. The case of ($\Delta T = 5\,°\mathrm{C}$; $P = 120\%$) indicates a warming of 5 °C and an increase of 20% in averaged precipitation relative to that in the control period. Although the same temperature and precipitation changes were applied to measured meteorological data throughout the year, the dependence of the analysis on the snow-covered period effectively focuses examination of the impact of these changes on winter and spring.

RESULTS

Evaluation of hydrological modelling performance

Figure 3 illustrates the agreement between modelled and observed hourly SWE time series in a forest gap over

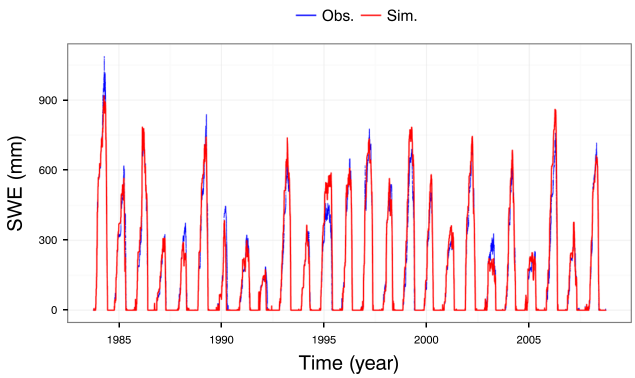

Figure 3. Modelled and observed hourly snow water equivalent (SWE) in the gap hydrological response unit near the sheltered meteorological station in Reynolds Mountain East Basin, Idaho, USA

Table I. Vegetation characteristics of the HRUs and parameters used in CRHM model in Reynolds Mountain East Basin

Parameter	Unit	Aspen	Low sage	Sage 1	Sage 2	Sage 3	Sage 4	Grass	Fir	Riparian willow
Height[a,c]	m	8	0.30	0.50	0.75	1	2	0.15	12	12
Max LAI[b,c]	m^2m^{-2}	5.4	1.2	1.2	1.2	1.2	1.2	4.3	5.9	3.6
Mean LAI[b,c,f]	m^2m^{-2}	1.35	1.1	1.1	1.1	1.1	1.1	1.2	3	1.35
Min LAI[b,c]	m^2m^{-2}	0.4	0.77	0.77	0.77	0.77	0.77	0.77	1.35	0.3
Veg. density[a]	$1/m^2$	0.2	4	2	2	1	1	320	0.2	2
Snow roughness[d]	m	0.006	10^{-4}	10^{-4}	10^{-4}	10^{-4}	10^{-4}	10^{-4}	0.006	0.006
Terrain emissivity[c]	—	0.98	0.98	0.98	0.98	0.98	0.98	0.98	0.98	0.98
Stalk diameter	m	0.45	0.01	0.01	0.01	0.01	0.01	0.01	0.45	0.45
Fetch distance	m	300	300	300	300	300	300	300	300	300
Snow active layer thickness[e]	m	0.15	0.15	0.15	0.15	0.15	0.15	0.15	0.15	0.15
HRU category	—	shelt	sourc	sourc	sourc	sourc	sink	sourc	forest	sink

Hydrological response units (HRUs) are categorized as shelt: sheltered, sourc: source, sink: sink and forest (with interception).
[a] Winstral *et al.*, 2013.
[b] Flerchinger *et al.*, 2012.
[c] Oke, 1978.
[d] Reba *et al.*, 2012.
[e] Reba *et al.*, 2014.
[f] Link *et al.*, 2004.

the period of 1984–2008 in RME for the water years starting on October 1 and ending on September 30. The mean absolute error, root mean square error and normalized root mean square error between the observed and simulated time series are 26 mm, 50 mm, and 0.046, respectively, over 25 years of modelling. These values are in close agreement with those reported by Reba *et al.* (2011b) for a detailed grid-based simulation based on the same data and time period. This is encouraging performance as parameters were not calibrated and the number of HRUs used in this study is much lower than the gridded spatial units used in Reba *et al.* (2011b). The CRHM model for RME performs sufficiently well in capturing the snow accumulation and ablation magnitude and timing to be used for sensitivity analysis to climate.

Sensitivity of snowpack to perturbed climate

Hydrological response units were grouped into the three snow regimes affected by blowing snow or snow interception in forest canopy and one with a sheltered snow regime, which is not affected by wind or interception (e.g. in forest gap) to compare and contrast snow accumulation and ablation amongst sites. The sensitivity of hourly SWE to warming air temperatures and precipitation change (see Sensitivity Analysis) for four snow regimes was simulated using measured and perturbed meteorology in RME (Figure 4). To estimate the probability density function (PDF) of snowpack in Figure 4, the kernel density estimation, a non-parametric approach, was used. This estimation is based on a normal kernel function and a window parameter (bandwidth) that is a function of the length of the time series ($n = 25\,\text{years} \times 365\,\text{days} \times 24\,h$). Figure 4 shows that PDFs have much narrower spreads as temperature warms more than 1 °C and a slight tendency for wider spreads as precipitation increases. Compensation for warming by precipitation increase is notable; a warming of 1 °C can be compensated for by a precipitation increase of 20% for all SWE values in all snow regimes (Figure 4). However, warming of 2 °C or more cannot be compensated by an increase in precipitation of 20%. The temporal frequency distributions of different snowpack regimes show that sheltered and blowing snow sink HRUs are more resilient to impacts of warming and changes in precipitation than blowing snow source and forest HRUs with snow interception. Of particular interest is the disappearance of the rare high SWE values with warming from blowing snow sink and forest interception HRUs with warming. Current medium range snowpacks are expected to become the peak SWEs with warming of 5 °C for any precipitation scenario. The warming impacts peak more than shallower snowpacks in all the snow regimes. These areas are important for early summer runoff generation in some years and so their absence under a warmer climate is expected to have great hydrological significance. However, an increase in precipitation can partially compensate for the impacts of warming and increases highest values of SWE, but not necessarily the low and medium range SWEs (Figure 4). In general, the snowpack regime in RME is more sensitive to warming than to changes in precipitation, a finding supported by the

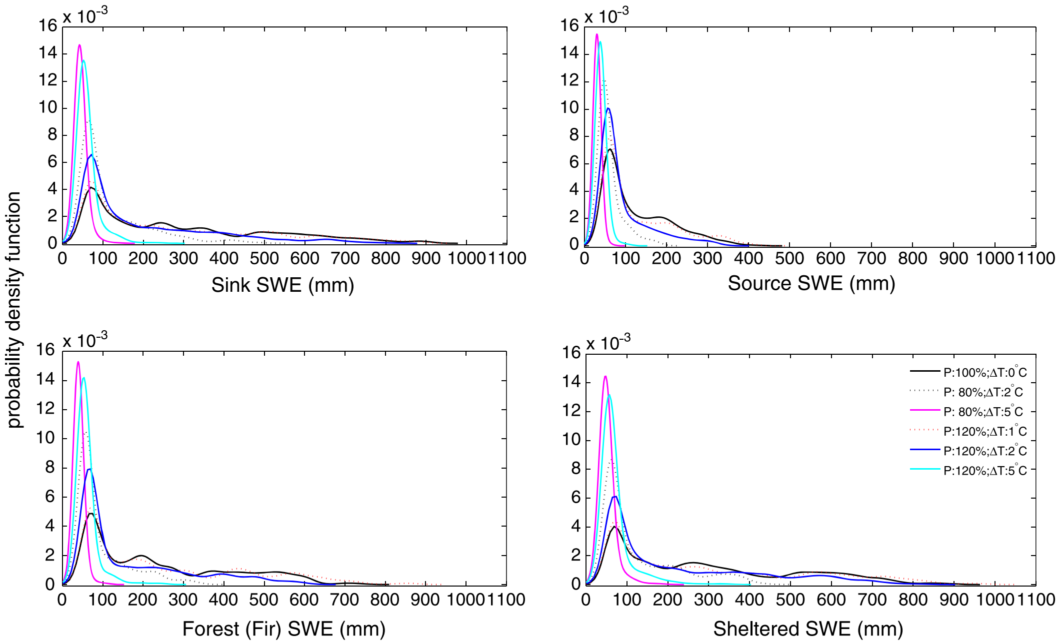

Figure 4. Sensitivity of snow water equivalent (SWE) in four categories of hydrological response units to warming and precipitation change shown as probability density functions for 25 years of simulation

results of Sproles *et al*. (2013) for the nearby but more temperate Cascades Mountains of Oregon, USA.

Snowpack seasonal and inter-water year variability from early accumulation in October to complete ablation in summer is illustrated in Figure 5 for the recent climate (1984–2008) and for warming and precipitation change. Model outputs for low sage, aspen drift, fir and gap HRUs represent snowpack variability in blowing snow source and sink areas, within tall trees that intercept and sublimate snow, and in forest gaps, which are sheltered from snow redistribution. Snow ablation starts at different times of the year in different HRUs: on March 1 for the sage HRU, in mid-March for forest gap HRU and on April 1 for aspen drift and the fir HRUs. The response of snow ablation to a moderate warming (2 °C) with and without change precipitation in RME is relatively similar amongst the different HRUs; the start of snow ablation advances to early March everywhere except for the source HRU (low sage) and advances up to 2 months for 5 °C of warming for all types of vegetation. The snow-free date is sensitive to concomitant warming and precipitation change and advances to before April 1 for warming of 5 °C and a 20% increase in precipitation. Aspen drift and sage drift HRUs in RME are deep, cold and north-facing and, therefore, latest to melt in most years. During colder years, meltwater from drifted snow is available to supply streamflow even as late as July (Figure 5) and is hydrologically important for this catchment (Winstral and Marks, 2002). The inter-water year coefficient of variation (CV) of SWE, defined as the ratio of standard deviation to average of the SWE values over 25 years, shows greater variability in fall and spring seasons (CV > 1) and lesser variability in mid-winter. Warming increases the snowpack inter-water year variability even in winter months, whilst a precipitation increase slightly decreases the variability in winter and significantly decreases it in spring (Figure 5, left panels). The small CV in the gap and sink HRUs from December to May, especially in ablation period relative to other sites, shows how snow regimes in small forest clearings are not representative of the natural temporal variability of snow regimes in source or forest zones that dominate land cover in a mountain basin.

Figure 6 shows the spatial variability of peak SWE in RME calculated using the recent climate (Figure 6a) and different scenarios of warming and change in precipitation (see Sensitivity Analysis). Note that the timing of peak SWE is not synchronized across the basin and differs from HRU to HRU. With warming and change in precipitation, spatial variability of peak snowpack decreases, and the response of shallow snowpack under severe climate conditions in different regimes becomes more similar. With moderate warming (2 °C), peak SWE dropped slightly in riparian willow, aspen, low sage and tall sage HRUs with a precipitation increase of 20% (Figure 6b), and declined dramatically across the basin with a 20% decrease in precipitation (Figure 6c). Severe warming (5 °C) causes peak SWE to drop below 200 mm across the basin with a 20% increase in precipitation (Figure 6d) and below 100 mm (Figure 6e) with a 20% decrease in precipitation.

Figure 7 shows the magnitude and associated percentage changes of mean water year peak SWE with warming and changes in precipitation. Peak SWE shows a strong sensitivity to increases in air temperature and a secondary sensitivity to changes in precipitation. Further, the sensitivity to precipitation change decreases as temperature increases. This suggests that peak snowpack in RME is very sensitive to warming and that increased precipitation

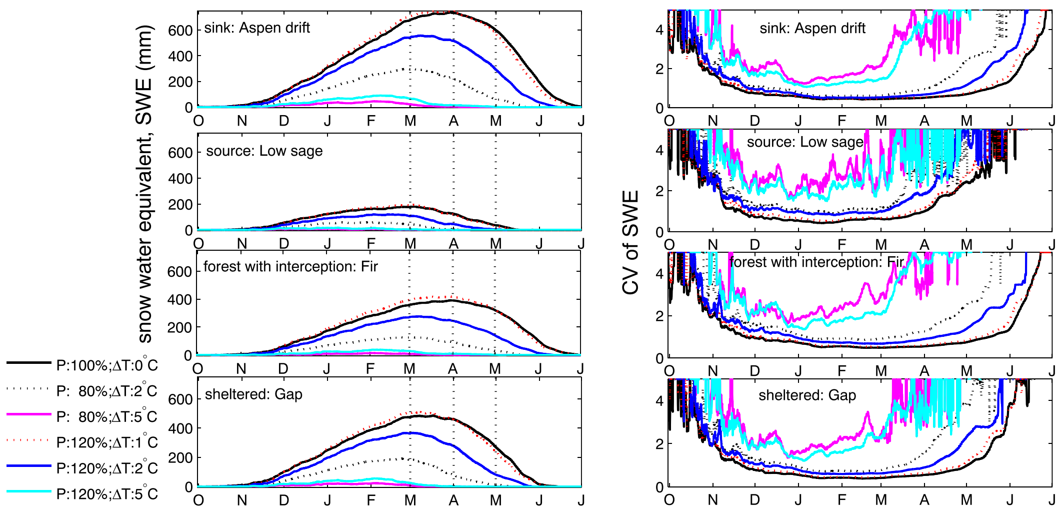

Figure 5. Sensitivity of snow accumulation and ablation [mean snow water equivalent (SWE)] to warming and changes in precipitation during snow season over 25 years of simulation (left panels) and associated coefficient of variability (right panels) for blowing snow sink (aspen drift) and source (low sage), forest (fir) and sheltered (forest gap) hydrological response units in Reynolds Mountain East. Vertical dashed lines represent March 1, April 1 and May 1 SWE values

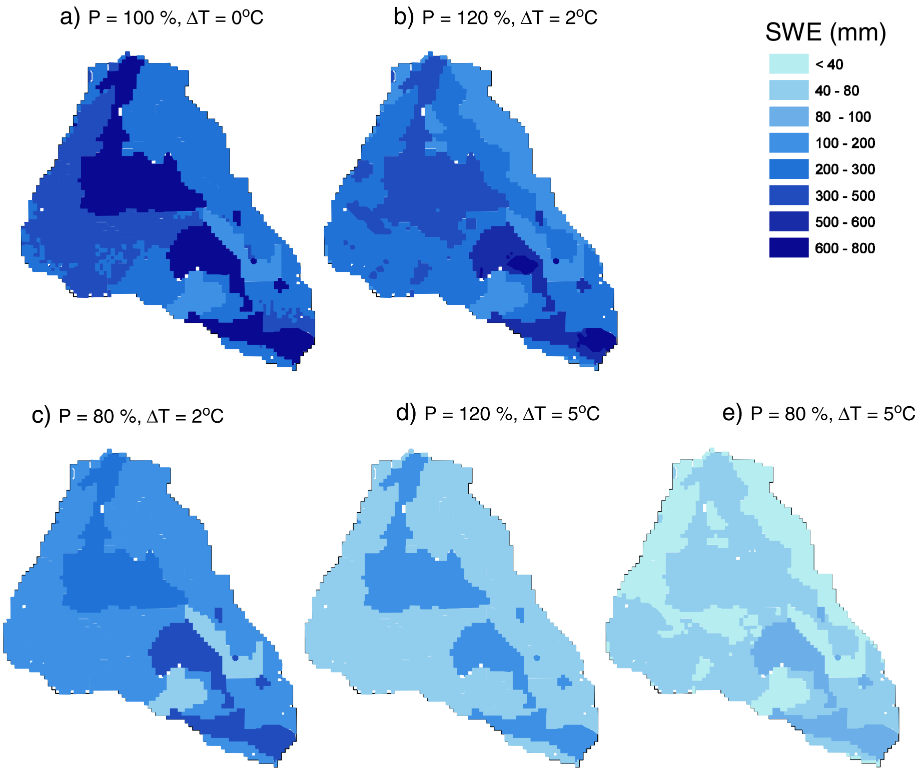

Figure 6. Spatial variability of mean water year peak snow water equivalent (black) in Reynolds Mountain East (a) during control period of 1984–2008 and under (b) increase in precipitation and moderate warming, (c) decrease in precipitation and moderate warming, (d) increase in precipitation and severe warming and (e) decrease in precipitation and severe warming scenarios (Sensitivity Analysis)

cannot compensate for the impacts of warming on SWE when warming exceeds 1 °C. For instance, concomitant warming of 5 °C and precipitation change of ±20% leads to an 84–90% drop in peak SWE. In the most extreme climate change case, a warming of 5 °C and decline in precipitation of 20% causes the peak SWE to decline by 90%, from 570 to 58 mm in blowing snow sink HRUs and from 427 to 39 mm in the HRUs with intercepted snow. Maximum snow accumulation is lower in the blowing snow source HRUs when compared with other snow regimes, and therefore, these drop the least, declining from 250 to 39 mm. The response of snow characteristics to warming and precipitation change is complex and very nonlinear because snow redistribution processes by wind and forest canopy add complexity and spatial variability to snow accumulation. For instance, peak SWE responded variably to a 20% increase of precipitation without warming, increasing by 22% in blowing snow sink HRUs and 32% in the intercepted HRUs. However, the results show that under severe warming and reduction in precipitation, peak snowpack becomes relatively uniform in all of the snow regimes in RME because of the suppression of snow redistribution processes. Climate warming would have to be less than 1 °C and be accompanied by a precipitation increase of at least 20% to allow peak SWE to remain within its historical range. However, all climate model scenarios predict greater warming than this for the 21st Century for western North America including the RME region (Stewart *et al.*, 2004) and substantial warming has already occurred. For instance, Nayak *et al.* (2010) analysed 45 water years of data (1962–2006) from RCEW. They reported for the RME catchment a trend showing an increase in mean water year temperature of +0.5 °C per decade, and a trend showing a reduction of April 1 SWE of 58 mm per decade. Their analysis shows that mean water year temperature in the RME catchment increased from around 4.0 °C in 1962 to 5.8 °C in 2006, and during the same 45-year period, April 1 SWE decreased from around 648 mm in 1962 to 436 mm in 2006. This represents about 18.5% reduction in April 1 SWE per degree of warming and is within the 13–30% reduction in water year peak SWE that is expected per 1 °C of future warming, based on the analysis in this paper. Water year maximum snow accumulation decreases 2–16% per 10% reduction in precipitation with and without warming; however, as air temperature increases above 3 °C, the sensitivity of snowpack to changes in precipitation decreases.

The timing of snowcover initiation, snow-free date, duration of snowmelt period and length of the snow season along with the magnitude of the peak snowpack are five main characteristics that describe the snow regime including the effects of accumulation, redistribution and ablation

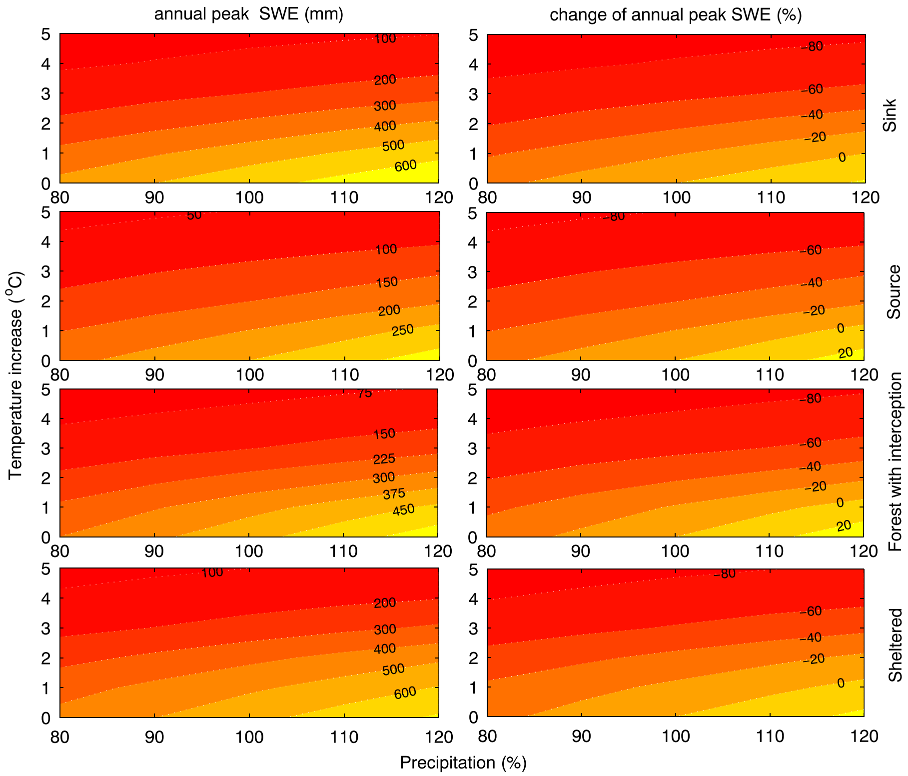

Figure 7. Water year peak SWE magnitude (left panels) and associated percentage (right panels) changes with warming up to 5 °C and precipitation decrease/increase up to 20% in different snow regimes (see Sensitivity Analysis)

processes. Figure 8 illustrates sensitivity of these characteristics to warming and change in precipitation averaged over the basin. The mean timing difference between water year peak SWE and snow-free date indexes the snowmelt period and is very sensitive to warming, but less sensitive to precipitation changes. With concomitant warming of 5 °C and decreasing precipitation (20%), the peak SWE drops 87% from 390 to 47 mm, snow ablates 2 months earlier, and snow season and snowmelt period become, respectively, five months and 48 days shorter than those in the recent climate (Figure 8 and Table II). These results are generally supported by trend analysis by Nayak *et al.* (2010) in RME as they found a 58 mm per decade reduction in peak SWE and a 6.4 day per decade delaying in snow cover initiation over the past 50 years. It is estimated that a 1 °C warming advances the timing of peak SWE by approximately 15 days; this can be compensated for by a 20% increase in precipitation. The date of peak SWE in source HRUs is very sensitive to warming of less than 3 °C, but less sensitive to warming of greater than 3 °C and changes in precipitation. The snow-free date advances from mid-May in the recent climate to early April with a moderate warming of 2 °C (Table II). Similar to peak SWE timing, the snow-free date is also very sensitive to warming but less sensitive to precipitation changes as shown in Table II. With concomitant warming (5 °C) and decreasing precipitation, the snow-free date across the catchment advances by 4 months to January (Table II). As shown in Figure 8, changes in the snow season duration are largely driven by warming and not by precipitation changes. This is because the snowpack is shallow and warm at the beginning and end of the season and shallow warm snow ripens and melts faster than does deep cold snow. Change in precipitation is the secondary factor that changes the magnitude of peak SWE and snowmelt duration. The values in Figure 8 reflect the relatively small influence of precipitation change when compared with the impact of warming. This suggests that peak SWE is primarily affected by warming and to a lesser extent by changes in precipitation. The combination of air temperature increasing by at least 1 °C (mean water year temperature exceeds 6.2 °C) and precipitation increasing by less than 20% (mean water year precipitation less than 1030 mm) results in declining peak SWE and substantial deviation from the historical ranges of snowpack in RME. These temperature and precipitation conditions are considered highly likely in climate model projections.

Sensitivity of snow fluxes to perturbed climate

The spatial variability of precipitation, snow transport, snowmelt and sublimation from various snow sources is illustrated in Figure 9. The positive bars in this figure

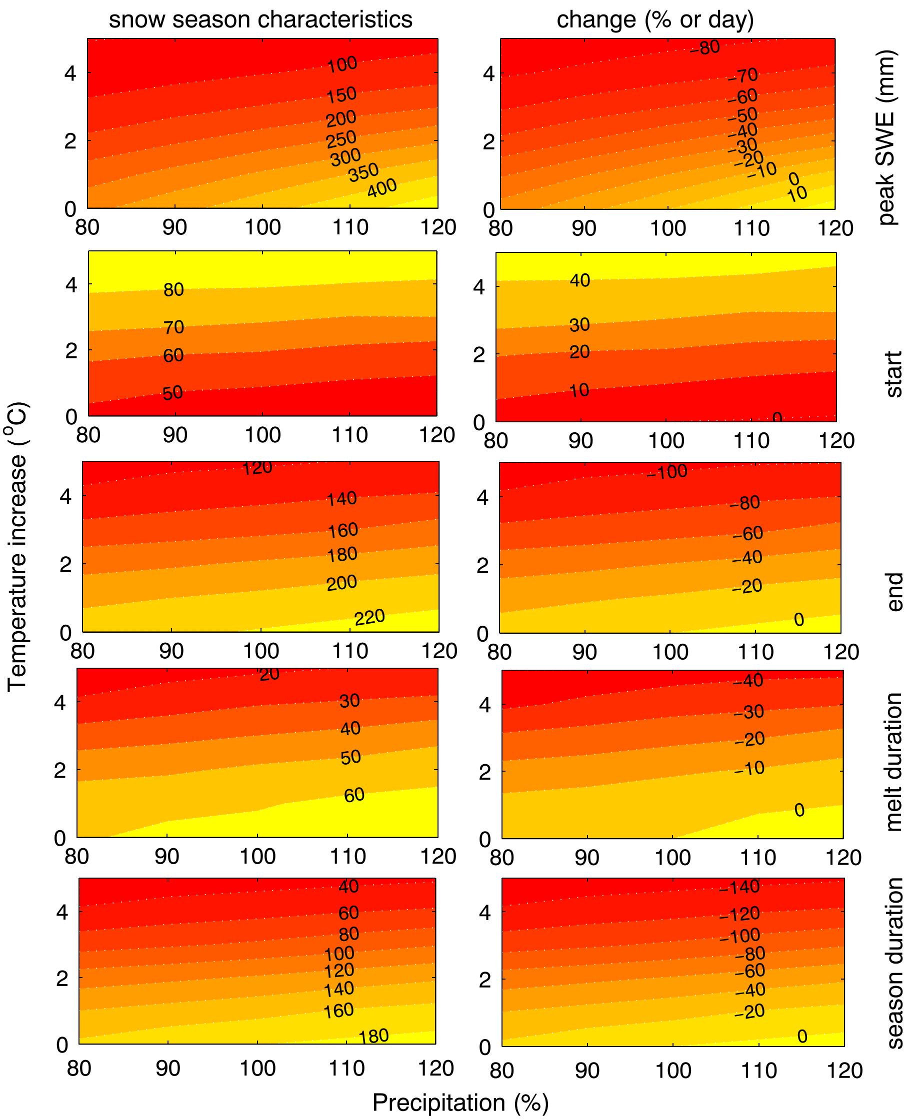

Figure 8. Magnitude and change of water year peak snow water equivalent and the timing shift of the snow season start/end, snowmelt period, and snow season duration with warming up to 5 °C and precipitation change up to 20% in Reynolds Mountain East (see Sensitivity Analysis). Negative and positive values show advancing and delaying dates, respectively. Julian water year dates starting from October 1 are given for the first day of each month at the end of Table II

represent the input to the HRUs, and the negative bars show the loss sources from the HRUs. In general, snow inputs from snowfall and blowing snow are greatest to HRUs that have tall vegetation such as willow, tall sage and sites in topographic depressions leading to snow drifts; in contrast, accumulation is less than cumulative snowfall in topographically exposed, short vegetation HRUs such as grass and short mountain sagebrush shrubs and where an evergreen canopy permits snow interception and subsequent sublimation losses. Water year sublimation loss from some HRUs is massive, by up to 178 mm (in grass HRU) or 20% of total water year precipitation in the catchment over the period of 1984–2008 – it is primarily from blowing snow in sparsely vegetated sites and from intercepted snow by fir canopies. Changes in the various mass budget terms are discussed in the succeeding discussions.

Precipitation. Simulated basin averaged snowfall over 25 years is 602 mm, which under 5 °C of warming drops to 224 mm with a 20% increase in precipitation and to 149 mm with a 20% decrease in precipitation (Figure 9). Precipitation phase is strongly affected by air temperature that varies with elevation and phase showed high variability, from 26% rainfall in aspen HRUs to 33% in tall sage HRU with an average of 30% over the basin (Figure 9). With warming of 5 °C, the rain to total precipitation ratio rises to 78% in the catchment

Table II. Comparison of the snow variables sensitivity to warming and changes in precipitation (see Sensitivity Analysis) in the Reynolds Mountain East Basin with a subarctic research basin, Granger Basin (Rasouli *et al.*, 2014)

Basin	Warming (°C)	0	2	5	0	5	5
	Precipitation (%)	100	100	100	120	80	120
Reynolds Mountain East	Peak SWE (mm)	390	222	63	486	47	80
	Accumulation start (dowy[a])	42[b]	59	86	40	84	84
	Timing of peak SWE (dowy)	158	131	99	162	96	101
	Snow-free date (dowy)	222	184	118	230	111	122
	Length of snow season (day)	180	125	32	189	27	38
Granger	Peak SWE (mm)	148	130	96	184	73	120
	Accumulation start (dowy)[a]	4	4	18	4	21	13
	Timing of peak SWE (dowy)	175	168	154	177	147	158
	Snow-free date (dowy)	249	236	217	252	212	221
	Length of snow season (day)	245	232	199	248	191	208

Julian date (water year)	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep
	1	32	62	93	124	152	183	213	244	274	305	336

[a] The unit dowy denotes the day of the water year, starting October 1, for convenience a guide for the Julian water year date is given for the first day of each month.
[b] For example, number 42 in the table denotes November 11.

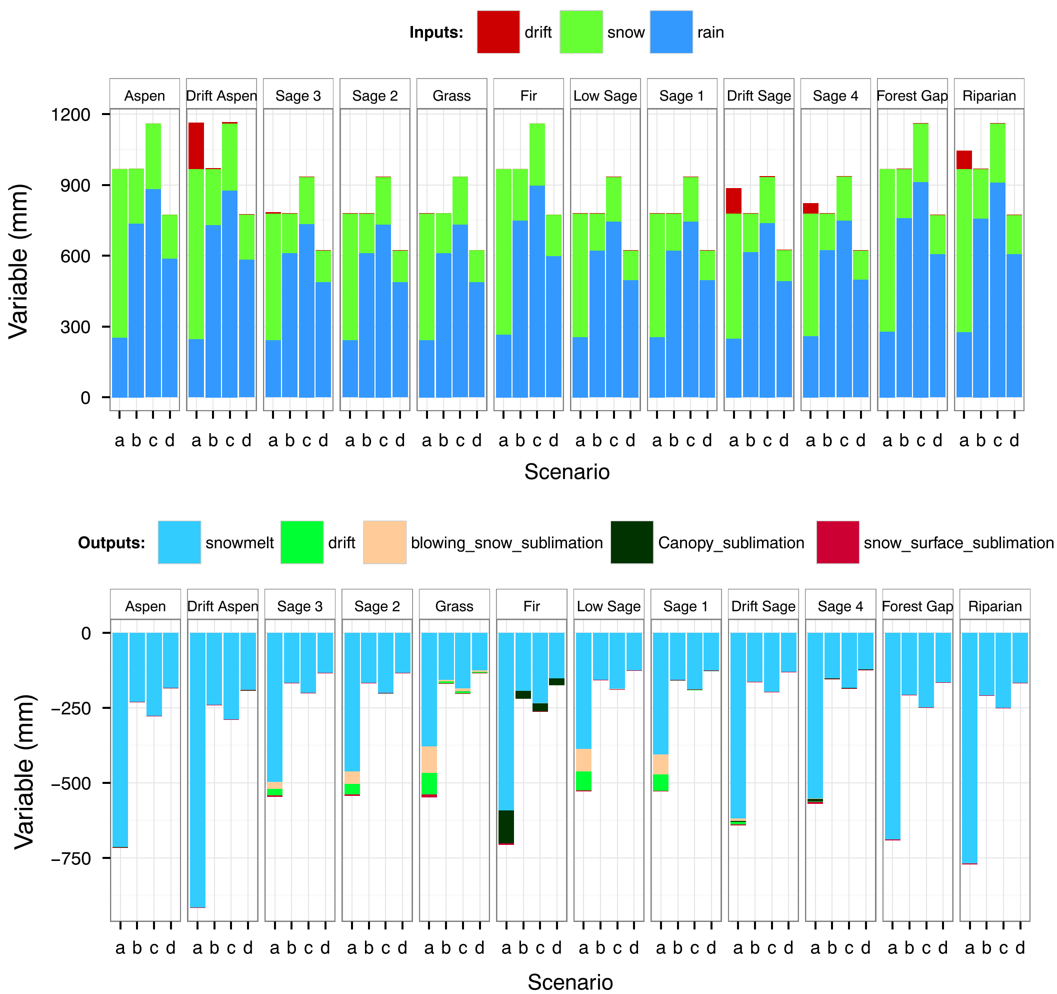

Figure 9. Vertical snow flux inputs (upper panel) and outputs (lower panel) in Reynolds Mountain East catchment for different hydrological response units under the following scenarios: (a) averaged over the control period of 1984–2008, (b) $P = 100\%$, $\Delta T = 5$ °C; (c) $P = 120\%$, $\Delta T = 5$ °C and (d) $P = 80\%$, $\Delta T = 5$ °C

(Figure 10) and is unaffected by precipitation change. Note that RME is an alpine basin, and this ratio is expected to be even more for regions with mid-elevation and low-elevation (Nayak *et al.*, 2010). A snow-dominated basin becomes a rain-dominated one under warming of 5 °C (Figure 9).

Snow transport. Blowing snow is transported from short vegetation HRUs such as grass and mountain sagebrush to tall vegetation and valley bottom HRUs, where snowdrifts may form. Vegetation density and height, which determine the aerodynamic roughness, play the key role in snow transport. HRUs with shorter vegetation such as grass show the greatest snow erosion and transport out. Blowing snow transport into and out of an HRU can be up to 196 and 71 mm, respectively, equivalent to 23% and 8.4% of the average water year precipitation. With 5 °C warming blowing, snow transport drops to ≤8 mm (Figure 9). This sensitivity to warming is because of the increasing bond strength and cohesion of snow as it warms, which raises the threshold wind speed required to initiate saltation (Li and Pomeroy, 1997). This shows that the occurrence of blowing snow transport in RME is very sensitive to warming and almost disappears completely when the basin climate warms by 5 °C.

Snowmelt. Water year snowmelt varies amongst different HRUs depending on the snow transport to or from the HRU and sublimation from surface and intercepted snow. Therefore, it has a nonlinear relationship with snowfall. For the same amount of snowfall, aspen drift has the highest depth of snowmelt because of the strong snow transport to this HRU. In general, sink HRUs release the greatest snowmelt depth, then sheltered HRUs and the smallest depth from the source HRUs, especially grasslands. Under a 5 °C warming, snowmelt is reduced 51–79% across the basin depending on the direction of precipitation change. Without precipitation change and under the same warming condition, it drops from 915 to 240 mm in aspen drift (Figure 9). Under different warming scenarios, the snowmelt rate drops with greater decline in the sink HRUs including aspen drift, sage drift, willow and tall sage and lesser decline in the source HRUs (e.g. grass and short sages). This shows that the spatial variability of the response of snowmelt to climate change is high and depends on the combined impacts of snow transport and sublimation processes and melt induced by processes such as rain-on-snow which increases as the rainfall ratio increases under climate warming. Scenario c in Figure 9 illustrates the impact of a 20% increase in precipitation under warming of 5 °C and shows a substantial increase in snowmelt. The increased incidence of warm precipitation would increase the rain-on-snow contribution to melt in such situations.

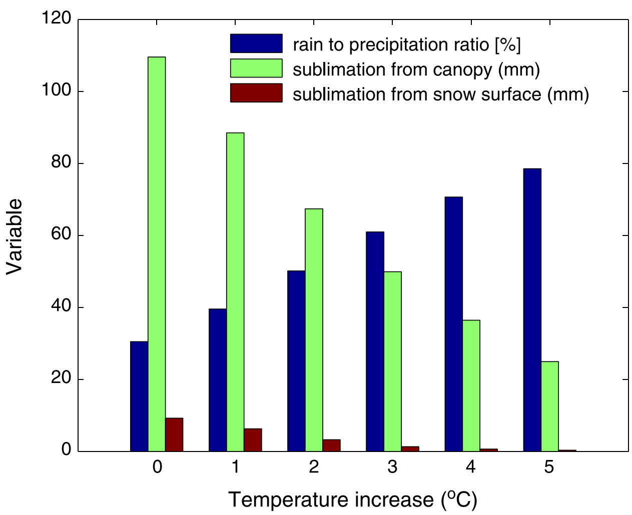

Figure 10. Sensitivity to an increase in air temperature for rainfall to total precipitation ratio, sublimation from intercepted snow in the fir forest hydrological response unit (HRU) and from the snow surface in the grass HRU in Reynolds Mountain East catchment

Sublimation from blowing snow. The spatial variability of sublimation from blowing snow is relatively high in RME, ranging from zero in sink HRUs to 87 mm in a short grass HRU (Figure 9). This indicates that up to 10% of the total precipitation in short-vegetated HRUs such as grass can be sublimated from blowing snow in RME under the recent climate. Because sublimation of blowing snow requires snow transport, with warming of 5 °C and a 20% increase in precipitation, sublimation from blowing snow declines to no more than 9 mm in grass HRU and is negligible in other HRUs and as a basin average (Figure 9).

Sublimation from intercepted snow. Water year mean sublimation from snow intercepted by vegetation canopies varies from 8 mm in the tall sage HRU to 110 mm in the fir HRU (Figure 9). There is a high inter-water year variability in sublimation especially from the fir HRU where water year loss ranges between 75 and 140 mm, or on average, 11.3% of the total precipitation in this HRU (Figure 10). Under warming of 5 °C, water year intercepted snow sublimation from fir drops to 22 mm or less when precipitation decreases 20% and 28 mm or less when precipitation increases 20%. For this warming scenario, sublimation from intercepted snow in other HRUs is negligible (≤2 mm) irrespective of precipitation change (Figure 9). This sensitivity to warming is a result of unloading or melt rather than sublimation of intercepted snow during mid-winter thaws (Pomeroy and Gray, 1995; Gelfan *et al.*, 2004; Ellis *et al.*, 2010). This shows that similar to the blowing snow transport, sublimation from intercepted snow is very sensitive to warming, much more so than to precipitation change. However, intercepted snow sublimation is less sensitive to warming than blowing snow transport and sublimation. Intercepted snow sublimation becomes significant by early December and ceases by the end of April. With 5 °C of warming, the period with sublimation from intercepted snow ends in mid-March, 45 days earlier than under the recent climate. This is because of the earlier precipitation phase transition from snowfall to rainfall under warming and greater unloading and drip of intercepted snow from the canopy. Sublimation declines linearly with warming up to 5 °C, whilst the rainfall to total precipitation ratio increases from 30% to 78% (Figure 10). Change in precipitation does not affect the cumulative seasonal intercepted snow sublimation under warming.

Sublimation from/condensation to snow surface. In comparison with other sublimation terms, sublimation from the snow surface is small and rarely reaches 1% of the total precipitation. The sensitivity of sublimation from the snow surface in warm conditions and condensation in cold and wet conditions (Marks *et al*., 1999) were investigated. As shown in Figures 9 and 10, up to 9 mm (≈1% of total precipitation) in different HRUs is lost due to the net sublimation from snow surface. The sensitivity of the sublimation from snow surface to warming is relatively high. Under concomitant change in precipitation (20%) and warming (5 °C), sublimation is reversed and up to 1 mm of water vapour condenses onto some HRUs.

Comparison of snow regime sensitivity with climate perturbation

In this section, the sensitivity of snow regime characteristics including snowpack formation date, peak SWE date, snow-free date, peak SWE and duration of snow season to warming and precipitation changes is compared with the results from a similar study of GB, an alpine sub-basin of Wolf Creek, Yukon. Here, only snowpack characteristics are compared in both sub-basins to compare the climate changes impacts in north and central parts of the North American Cordillera. There are substantial differences in snow regime sensitivity between the two basins. With 5 °C warming and no changes in precipitation, the onset of winter is delayed 42 days in RME and 2 weeks in GB and the end of winter comes 104 and 32 days earlier in RME and GB, respectively. When compared with historical winters, a 20% increase in precipitation would lengthen the winter season by only a few days in either catchment. When warming is limited to 2 °C, as little as a 10% increase in precipitation was able to compensate for the effect of warming on timing in GB. Peak snowpacks in GB increased slightly under this scenario set, but with warming of 5 °C, peak snowpacks decreased in all scenarios, even with increased precipitation of 20%. The 25-year mean water year peak SWE in RME and GB is 390 and 148 mm, respectively, both occur in early March. With warming of 5 °C and a 20% decrease in precipitation, peak SWE drops by 87% and 55% and timing advances to early January and late February, respectively. With the same warming but 20% greater precipitation, peak SWE declines 79% in RME and 19% in GB and its date advances approximately 2 months in RME and 17 days in GB (Table II). The maximum snowpack increases 25% and 24% in RME and GB without warming and with a 20% increase in precipitation. This clearly shows a strong response of the snow accumulation amount and timing to warming and weaker response to changes in precipitation in RME if warming occurs and a fairly strong sensitivity to warming and mild response to precipitation changes in GB if warming occurs.

DISCUSSION

Perturbations of hourly air temperature and precipitation were used to investigate the sensitivity of the snow regime modelled in RME. The impact of warming of 1 °C on SWE values over the winter and spring seasons can be compensated by a precipitation increase of 20% for almost all SWE values in all snow regimes. However, warming of 2 °C or more cannot be compensated by increases in precipitation of less than 20%. The sensitivity of SWE in the blowing snow source and sink HRUs to warming is higher than that in the forested intercepted snow and sheltered forest gap HRUs which is likely due to the suppression of blowing snow redistribution processes by warming. The low temporal variability in the forest gap and blowing snow sink SWE from December to May relative to other sites shows how snow regimes in small forest clearings and snow drifts are relatively stable and not representative of the natural temporal variability of snow regimes in exposed source or forest zones. The sheltered HRU has insignificant snow redistribution processes and shows the lowest response to warming. Because the locations of USDA SNOTEL sites are usually in forest gaps, this may have implications for the ability of the SNOTEL network to fully represent the dynamics involved in changing basin snow hydrology due to climate change.

The implication of the results is that despite the apparent uniformity of high mountain climates, the alpine snow regime responses to climate change differ substantially across North America and require regional analysis. The great difference between snowpack response in RME and GB implies that warming in cool climates impacts the maximum accumulated snowpack more than it does in cold climates. Warming affects the phase of precipitation, causing a shift from snowfall to rainfall in the spring and fall transition seasons and a shift from March to January for timing of peak snow accumulation. As the rainfall to precipitation ratio increases (Figure 10), advective, sensible and latent heat fluxes associated with rain-on-snow events (Marks *et al*., 1999) facilitate more rapid snowmelt in the cool mountain climate of RME (Figure 9; Scenario c), much more than is found in the cold subarctic climate of GB. Warming and accelerated rain-snow processes can accelerate the initiation of snowmelt in RME, as the melt period is shifted forward into a lower solar irradiance period. Despite this effect, snowmelt ended earlier as temperatures increased and the snow season shortened. The impacts of warming on snowpacks can be partly compensated for by precipitation increase in the cold GB climate but not in the cool RME climate. The snow season is expected to shorten 5 months in the cool RME basin and about 1.5 months in the subarctic GB with concomitant warming and decline in precipitation. This implies that if warming also occurs, the snow hydrology of RME is insensitive to precipitation increases; however, it is very sensitive to warming and precipitation phase change. GB hydrology is sensitive to a 'loss of cold' that is connected to large decreases in snowpack with warming temperatures but also sensitive to changes in the amount of precipitation, especially if warming is minimal.

CONCLUSION

A physically based semi-distributed snow hydrological model using HRU spatial discretization was developed from the cold regions hydrological modelling platform and used to calculate snowpack magnitude and timing along with the other mass balance fluxes for a 25-year period in RME without any calibration of model parameters. The model simulations of SWE accumulation and melt were very acceptable when compared with measurements in the basin and comparable with those obtained using a finely spatially distributed model. The results show that SWE is heavily redistributed from short vegetation and high wind exposure sites to taller vegetation and topographic breaks where drifts may form and that snow interception in fir forest canopies is significant, resulting in greater accumulation in forest gaps than in the surrounding forest. As a proportion of total precipitation over the last 25 years, SWE losses due to blowing snow transport and sublimation are 23%, intercepted snow sublimation losses from evergreen vegetation are 11%, and sublimation from the surface snowpack is about 1%; these fluxes depend on the vegetation cover, exposure to wind and physiography of the HRUs in RME. With warming and change in precipitation, the spatial variability of peak snowpack decreases and the snow regimes in different HRUs become more similar. Warming impacts peak SWEs much more than low and medium range snowpacks in all the snow regimes in RME. However, an increase in precipitation slightly increases peak SWEs which can partially compensate for the impact of warming. With warming of 5 °C, peak SWE in RME drops 84% and advances to early January with little influence from changes in precipitation. The snow season shortens from 6 months to 1 month in RME with 5 °C warming and 20% decline in precipitation. With warming of 5 °C, snow transport almost ceases and sublimation from open areas and to a lesser extent sublimation from the canopy decreases and the rainfall to total precipitation ratio rises to 78%. With a 5 °C warming, snowmelt is reduced 51–79% across the basin depending on the direction of precipitation change over the period of 1984–2008. As a result, the snow dominated RME catchment becomes rain-dominated, with SWE more variable over time and less variable over space.

The results in RME contrast with those from colder GB from the northern part of the North American Cordillera where concomitant precipitation changes control the declining SWE rate as temperature rises. For instance, with a 5 °C warming and a 20% increase in precipitation, peak SWE drops 79% and 19% in RME and GB, respectively. This indicates that precipitation increases can partially compensate for the impact of warming in the cold GB alpine climate but not in the cool RME alpine climate. These results show that the impacts of warming on cold regions hydrological processes in mountain basins vary along the Western Cordillera, sensitivity being very strong to warming and lower to precipitation change in the central part and moderate to warming and precipitation change in the northern part. Therefore, regional responses to warming and changes to precipitation must be considered to evaluate future alpine hydrology. Simulations of future conditions for snow regimes in this paper are in accord with the SWE magnitude and timing trends of the past 50 years in the RME catchment. This indicates that the simulation results are similar to measured results, and therefore, the model developed here can be applied to other alpine regions. Further investigation is needed to consider the impact of temperature and precipitation changes in other alpine regions around the world. In the meantime, the results of this study can inform water resources stakeholders on the vulnerability of alpine headwaters to first-order climate change impacts.

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