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Research papers

The sensitivity of snow hydrology to changes in air temperature and precipitation in three North American headwater basins

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ABSTRACT

Whether or not the impact of warming on mountain snow and runoff can be offset by precipitation increases has not been well examined, but it is crucially important for future downstream water supply. Using the physically based Cold Regions Hydrological Modelling Platform (CRHM), elasticity (percent change in runoff divided by change in a climate forcing) and the sensitivity of snow regimes to perturbations were investigated in three well-instrumented mountain research basins spanning the northern North American Cordillera. Hourly meteorological observations were perturbed using air temperature and precipitation changes and were then used to force hydrological models for each basin. In all three basins, lower temperature sensitivities of annual runoff volume ($\leq$ 6% $^\circ\mathrm{C}^{-1}$) and higher sensitivities of peak snowpack (-17% $^\circ\mathrm{C}^{-1}$) showed that annual runoff was far less sensitive to temperature than the snow regime. Higher and lower precipitation elasticities of annual runoff (1.5 – 2.1) and peak snowpack (0.7 – 1.1) indicated that the runoff change is primarily attributed to precipitation change and, secondarily, to warming. A low discrepancy between observed and simulated precipitation elasticities showed that the model results are reliable, and one can conduct sensitivity analysis. The air temperature elasticities, however, must be interpreted with care as the projected warmings range beyond the observed temperatures and, hence, it is not possible to test their reliability. Simulations using multiple elevations showed that the timing of peak snowpack was most sensitive to temperature. For the range of warming expected from North American climate model simulations, the impacts of warming on annual runoff, but not on peak snowpack, can be offset by the size of precipitation increases projected for the near-future period 2041–2070. To offset the impact of 2 $^\circ$C warming on annual runoff, precipitation would need to increase by less than 5% in all three basins. To offset the impact of 2 $^\circ$C warming on peak snowpack, however, precipitation would need to increase by 12% in Wolf Creek in Yukon Territory, 18% in Marmot Creek in the Canadian Rockies, and an amount greater than the maximum projected at Reynolds Mountain in Idaho. The role of increased precipitation as a compensator for the impact of warming on snowpack is more effective at the highest elevations and higher latitudes. Increased precipitation leads to resilient and strongly coupled snow and runoff regimes, contrasting sharply with the sensitive and weakly coupled regimes at low elevations and in temperate climate zones.

1. Introduction

High elevation mountain headwater basins are hydrologically important, as they store water in the form of snowpack during winter and release it in spring and summer (Barry, 1992; Bales et al., 2006). They are also ecologically important, as they are key zones for biodiversity due to steep gradients of air temperature, precipitation, and topography (Beniston, 2003). Mountain snowpacks are sensitive to warming (Minder, 2010). Air temperature changes exert important controls on the hydrology of basins, where snowmelt is the dominant hydrological process (Marks et al., 1998; Pederson et al., 2011; Sospedra-Alfonso et al., 2015). The contribution of mountain headwaters to the downstream discharge of rivers ranges from 35% in cold and humid river basins to 90% in hot and arid basins (Viviroli and Weingartner, 2004). Mountains cover 25% of the earth's land surface (Diaz et al., 2003), and 26% of the world's population lives in high-elevation areas (Meybeck et al., 2001). The origin of discharges from 50% of the world's rivers are mountain headwaters (Beniston, 2003). Snowmelt volume and

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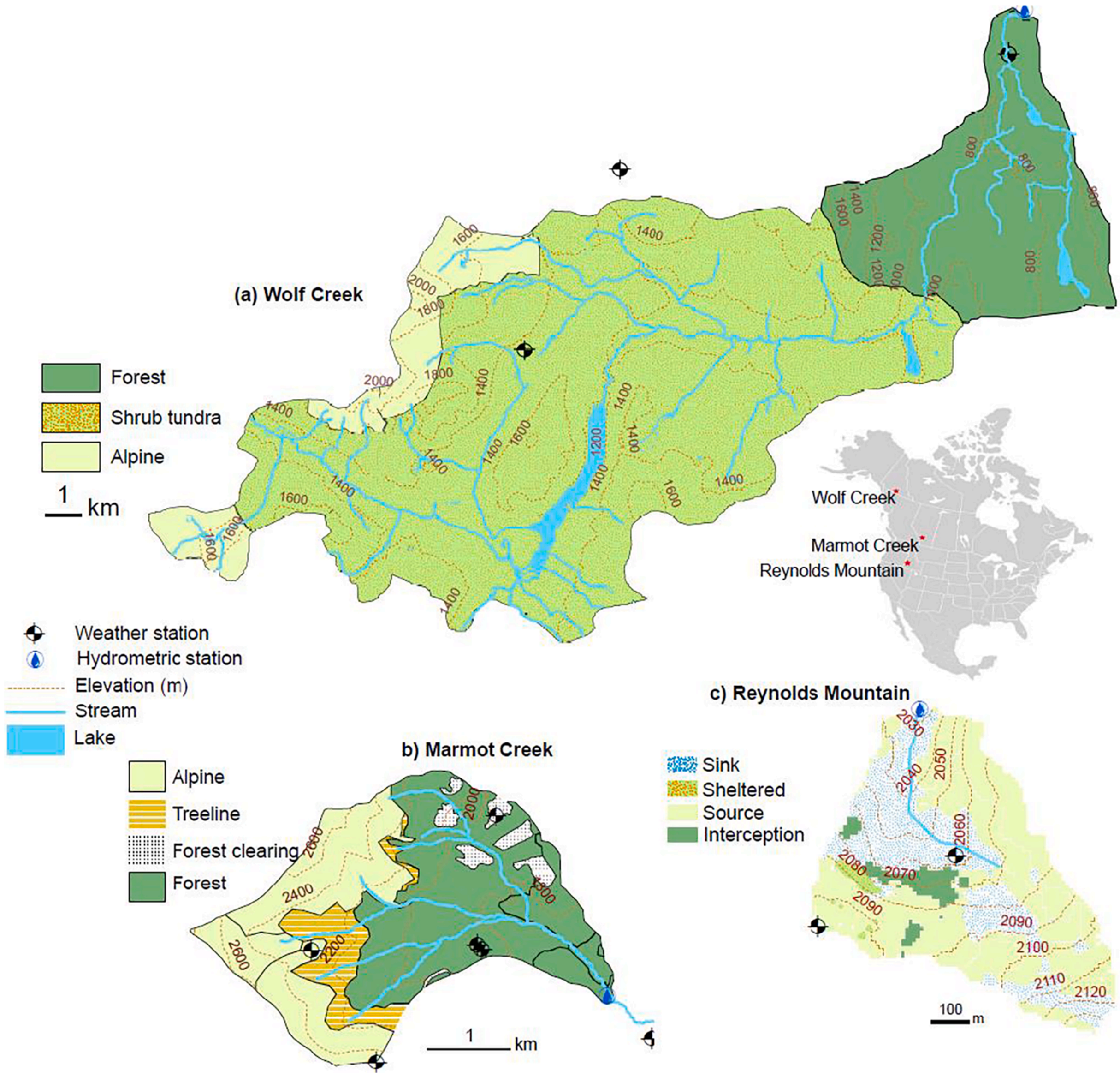

Fig. 1. Three headwater basins across the northern North American Cordillera: Wolf Creek, Yukon Territory; Marmot Creek, Alberta; and Reynolds Mountain East (Reynolds Mountain) catchment within Reynolds Creek Experimental Watershed, Idaho (USA).

timing play a key role in freshwater availability, flood control, and ecological sustainability of cold region mountain environments (Stewart et al., 2004; Semmens and Ramage, 2013).

The higher sensitivity of snow and frozen soils to warming (Negm et al., 2021) makes cold region mountain basins – those with mean annual air temperatures near 0 °C – suitable study areas for investigating climate change impacts on the hydrological cycle (Barry, 1992; Bunbury and Gajewski, 2012). Climate warming effects have been studied in some mountain headwater basins (e.g., Cayan, 1996; Stewart et al., 2004; Bales et al., 2006), and warming is expected to continue to threaten the ecological and hydrological integrity of these regions (Malmqvist and Rundle, 2002). Fyfe and Flato (1999) showed that elevation becomes important to the pattern of climate change over western North America only when significant continental-scale warming dominates, and it is not detectable in the early stages of climate change. Late winter and spring temperatures have a key role in the responsiveness of mountain basins to a warming climate and snowmelt runoff timing in regions with near-freezing air temperatures (Stewart et al., 2004; McCabe and Clark, 2005; Rasouli et al., 2019a; Rasouli et al., 2020). Mote et al. (2005) reported that climatic trends, and not changes in land use and forest canopy, affect snowpack in western North America. A significant increasing temperature trend, especially in minimum temperature, led to a reduction in the number of soil freeze days, earlier occurrence of plant-water stress, and a strong seasonal shift in streamflow in the high-elevation Reynolds Creek Experimental Watershed, USA (Nayak, 2008).

A common approach for investigating the hydrological response to climate change is to apply climate model projections under different greenhouse gas emission scenarios and to downscale regional atmospheric circulations obtained from the climate models to variables at local scales using statistical or dynamical methods (e.g., Jasper et al., 2004; Fowler et al., 2007; MacDonald et al., 2010). Mountain hydrometeorology, however, poses challenges to statistical and dynamical downscaling methods. The assumption in statistical downscaling that the predictor–predictand relationship is stationary and future relationships will be the same as those in the past (Wilby and Wigley, 1997) does not guarantee that statistical downscaling approaches will perform better than the delta change method (Hay et al., 2000; Fowler et al., 2007; Kay et al., 2009; Sunyer et al., 2012). Dynamical models driven by an ensemble of multiple boundary conditions have high computational costs at the fine-scale resolutions needed in mountains and so usually neglect uncertainty and also require bias correction to provide reasonable forcings (Fang and Pomeroy, 2020). A realistic downscaling of atmospheric variables shows a high sensitivity to the choice of downscaling methods (Wilby et al., 2000). These limitations make consideration of an alternative solution necessary for mountainous regions.

As biases due to scale and parametrization issues have not yet been resolved by statistical and dynamical downscaling methods, an alternative perturbation method known as the delta change factor method (e.g., Stockton and Boggess, 1979; Semadeni-Davies et al., 2008; Kawase et al., 2009) can produce plausible hydroclimatological changes for the future. The perturbation method represents changes in climatology between current and future climates for variables such as precipitation and air temperature (Stockton and Boggess, 1979; Pomeroy et al., 2015; Rasouli et al., 2019a; Rasouli et al., 2019b; López-Moreno et al., 2020). The method retains the main hydrometeorological processes present in historical measurements whilst minimizing computational resources. The perturbation method has been widely used; however, its application has been limited to air temperature changes factors (e.g., $\Delta T = \pm 2^{\circ}$C in Nayak, 2008; Pomeroy et al., 2015) and precipitation change factors (e.g., $\Delta P = \pm 25\%$ in López-Moreno et al., 2016).

Hydrological elasticity is a metric that is used to quantify the climatic sensitivity of a hydrological variable (Schaake, 1990; Cooper et al., 2018). Runoff elasticity is defined as the fractional change in runoff for a fractional change in climatic forcing variables such as precipitation or air temperature. Barrera et al. (2020) studied the climate elasticity of runoff to precipitation and warming in the Andes and found that the higher snow-dominated elevations exhibit lower elasticity. Tang et al. (2020) assessed precipitation and potential evapotranspiration elasticities of runoff. The sensitivity of snow processes to warming were studied in the Canadian Rockies (Pomeroy et al., 2015) and the Owyhee Mountains of Idaho, USA (Rasouli et al., 2015). Snow and runoff sensitivities to precipitation change and warming were studied in the Coast Mountains of Yukon Territory, Canada (Rasouli et al., 2014).

This study applies elasticity and sensitivity analyses to both snow and runoff regimes, in a comparable way, using physically based models designed using the Cold Regions Hydrological Modelling platform (CRHM), representing three data-rich headwater basins that span the northern North American Cordillera. This research investigates how the magnitude and timing of peak snowpack and annual runoff respond to combinations of temperature increases (0 °C to 5 °C) and precipitation changes (−20, −10, 0, +10, and + 20%). By considering 30 combinations, applied to each of the three mountain headwater basins, the elasticity and sensitivity of hydrological responses to changes in forcings can be compared. This increases the understanding of the relationships between changes in forcing and model response in these basins. The main question addressed is whether the impact of warming on mountain snow and annual runoff can be offset by precipitation increases. This has not been resolved in the literature (e.g., Arnell, 1999; Prowse et al., 2006; Luo et al., 2008). The specific objective for this sensitivity analysis is, therefore, to quantify the response of simulated mountain hydrological processes to changes in air temperature and precipitation associated with future climate change.

2. Data sources

Three well-instrumented and, therefore, data-rich headwater research basins located in the northern North American Cordillera were used to simulate potential hydrological responses to warming and precipitation changes; Wolf Creek, Yukon Territory, Canada; Marmot Creek, Alberta, Canada; and Reynolds Mountain East (hereafter Reynolds Mountain), Idaho, USA (Fig. 1). The availability of long-term data from multiple hydrometeorological stations at different elevations in each basin makes these basins uniquely suitable case studies for conducting sensitivity analysis on hydrological processes. High quality measurements of hourly air temperature, relative humidity, wind speed, incoming shortwave radiation, precipitation, and streamflow discharge for each basin were used. Consistent with the Ideal Gas Law, relative humidity was held constant to allow water vapour pressure to change with temperature. The forcing data are publicly available (from 1993 to 2011 for Wolf Creek: Rasouli et al., 2019c; from 2006 to 2013 for Marmot Creek: Fang et al., 2019; from 1984 to 2008 for Reynolds Mountain: Reba et al., 2011).

All three basins are snow dominated under the current climate and are partially covered by coniferous forests at lower altitudes: Wolf Creek has spruce and pine forests; Marmot Creek has larch, fir, spruce, and pine forests; and Reynolds Mountain has fir, pine, and aspen forests. All elevations in Wolf Creek and the high elevations in Marmot Creek are very cold. A cold snow season with high precipitation leads to a nine-month snow season at high elevations in Marmot Creek. High wind speeds in the alpine zones of all three basins redistribute snow by wind transport and result in substantive blowing snow sublimation losses. Needleleaf canopy snow interception and sublimation losses from intercepted snow are important in all three basins. Air temperatures of the forested elevations in Wolf Creek are lower than for the alpine zone in winter, due to strong inversions in the Yukon River Valley. In contrast, Reynolds Mountain and lower elevations in Marmot Creek have warmer air temperatures with fewer freezing days, making these more sensitive to warming.

3. Methods

3.1. Hydrological Modelling

Snow and runoff regimes are simulated using models created with the physically based CRHM platform (Pomeroy et al., 2007). Models for each basin were developed based on elevation, ranges of slope and aspect, and vegetation type. Modules that describe the major hydrological mechanisms in cold regions and used in this study include: (1) Radiation module that computes shortwave and longwave irradiance and adjusts the shortwave radiation on slopes; (2) Albedo module that estimates snow and snow-free surface albedos; (3) Canopy module that estimates the sub-canopy short- and longwave irradiance and turbulent transfer to snow and the interception of snowfall and rainfall and its subsequent drip, unloading, sublimation and/or evaporation from the forest canopy; (4) Blowing snow module that simulates the wind redistribution of blowing snow from one hydrological response unit (HRU) to another including blowing snow sublimation losses; (5) Energy balance snowmelt and surface sublimation module that estimates snowmelt, net radiation, sensible heat, latent heat, ground heat, advection from rainfall and in internal energy changes; (6) Infiltration module that estimates snowmelt and rainfall infiltration into frozen and unfrozen soils and surface runoff if snowmelt or rainfall exceeds the infiltration rate; (7) Evaporation module that estimates actual evapotranspiration from unsaturated surfaces using an energy balance and extension of Penman's equation to the unsaturated case and from wetted surfaces including lakes and stream channels (Priestley and Taylor, 1972); (8) Soil and hillslope module that estimates soil moisture, depressional storage, surface and sub-surface flows and moisture withdraws by roots in two soil layers and groundwater discharge in a saturated groundwater layer. Horizontal and vertical drainage from the soil and groundwater layers and hydraulic conductivity in unsaturated zone are estimated using Darcy (Hubbert, 1956) and Brooks and Corey (1964) relationships; and (9) Routing module that estimates runoff lag and route timing.

The CRHM models for each mountain basin were developed and evaluated with basin observations as per Fang et al. (2013), Rasouli et al. (2014); Rasouli et al. (2015) and Rasouli (2017) and were then used to assess the elasticity and sensitivity of the hydrological response to climate change by perturbing the model forcings. The models were run using HRUs that are spatially discretized based on vegetation type and height, land slope and aspect and soil depth. The spatial resolution of the

CRHM models was that of the HRUs, which had different sizes ranging from less than 1 km^2 to greater than 20 km^2, and the temporal resolution was hourly. HRUs were aggregated into three groups of alpine, shrub tundra and forest for Wolf Creek, four groups of alpine, treeline, forest, and forest clearings for Marmot Creek, and four groups of blowing snow source, sink, sublimation, and sheltered from the wind for Reynolds Mountain. These divisions were based on topographic exposure and vegetation height (Pomeroy et al., 1997). Blowing snow sink HRUs included drift, riparian, and tall vegetation HRUs. The forested landscapes were divided into those that are subject to interception and subsequent sublimation (evergreen trees) and those that are cleared (gaps) or have minimal winter interception capacity (sheltered from the wind, Pomeroy et al., 2002).

The sensitivity experiments use the CRHM basin models, driven with perturbed forcings, to simulate outputs such as snowpack and the timing and magnitude of runoff to assess the precipitation and air temperature elasticity of the snowpack and runoff. The sensitivities of interest are the hydrological responses to increases in air temperature and changes in precipitation that use the observed time series of air temperature and precipitation perturbed changes in the ranges projected by climate models under the Special Report on Emissions Scenarios (SRES) A2 (business-as-usual) and the Representative Concentration Pathways (RCPs), and are consistent with the recently defined Shared Socioeconomic Pathways (SSPs) of global change for these three basins. Rather than simulations based upon individual climate models, this linear sensitivity analysis and associated elasticities provide an assessment of the scale of alteration of the hydrological cycle in mountain basins within the range estimated by climate change predictions. This approach illustrates how both the individual and the combination of changes in air temperature and precipitation might induce hydrological changes in these basins. Knowing how combinations of warming and precipitation changes induce future hydrological change in mountain basins from northern to mid-latitudes can be used to assess the possible impacts of climate change.

3.2. Climate perturbation sensitivity

A climate perturbation sensitivity method is introduced here, in which the current climate is perturbed based on projected future climatological changes. With this method, a climate perturbation signal of the future atmosphere is added to high-resolution baseline hourly observations. The general perturbation approach, and the method used here, has two main assumptions: (i) GCM outputs for current and future climates show relative changes rather than absolute changes in climate; and (ii) the number of precipitation events is constant in current and future climates (Semadeni-Davies et al., 2008). The perturbation method only modifies the observed past and does not consider future changes in frequency and intensity of weather patterns. The assumption of linear scaling used for temperature in the perturbation method may introduce uncertainties for non-linear variables such as precipitation, particularly for extremes (Kay et al., 2009). It is also assumed that the basin vegetation and, in the case of Wolf Creek, permafrost (Williams et al., 2015), will remain unchanged.

The range of annual perturbations in precipitation and warming considered for this study is based on the atmospheric changes estimated by the SRES A2 scenario and two RCPs, which are consistent with the recent SSP5 8.5. The climate dataset used for the SRES A2 scenario was obtained from the North American Regional Climate Change Assessment Program (NARCCAP). These simulations provide climate data for regional climate models driven with GCM boundary conditions (Mearns et al., 2007). The range of temperature and precipitation perturbations was chosen based on the average climate changes that were obtained for RCPs and for the eleven NARCCAP regional climate models for the periods 2041–2070 minus 1971–2000. The climate dataset used for RCP scenarios was adapted from the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (Barros et al., 2014), which is consistent with the SSP scenarios provided in the recently published Six Assessment Report (IPCC, 2021). These two RCPs corresponding to specific radiative forcing values of 2.6 W/m^2 and 8.5 W/m^2 were used as a basis for long-term and near-term modelling experiments in climate change studies. The SSP5 8.5 scenario represents a fossil-fueled development with very high CO_2 emissions that will roughly double from current levels by 2050. Table 1 summarizes the expected change for air temperature and precipitation in the southern Yukon Territory (Wolf Creek), Canadian Rockies (Marmot Creek) and northwestern USA (Reynolds Mountain). Most modelled scenarios project the future climate to be wetter, but some SRES scenarios (Moss et al., 2010) show regional decreases in annual precipitation of up to 15% for the 2080 s. Rather than following any specific SRESs, RCP or SSP, the sensitivity analysis spans potential changes in air temperature and precipitation from all SRESs, RCP, and SSP; perturbing air temperature by 0 °C to 5 °C in 1 °C intervals and precipitation by –20% to + 20% in

Table 1
Expected changes in the study areas based on SRES A2 (business-as-usual) scenario, obtained from NARCCAP simulations (Mearns et al., 2007) and Representative Concentration Pathways (RCP, IPCC, 2013) covered by ranges of the recent Shared Socioeconomic Pathway (SSP5-8.5, IPCC, 2021).

Basin	Scenario	Warming (°C)	Precipitation change (%)
Wolf Creek	SRES A2	2.6	16.3
	RCP2.6	2	10
	RCP8.5	5	20
	SSP5 8.5	2.7 – 7.4	7.5 – 21.5
Marmot Creek	SRES A2	2.2	6.6
	RCP2.6	1	5
	RCP8.5	5	10
	SSP5 8.5	1.9 – 5.2	7.5 – 21.5
Reynolds Mountain	SRES A2	2.4	2.3
	RCP2.6	1	5
	RCP8.5	5	10
	SSP5 8.5	1.9 – 5.2	2.7 – 6.8

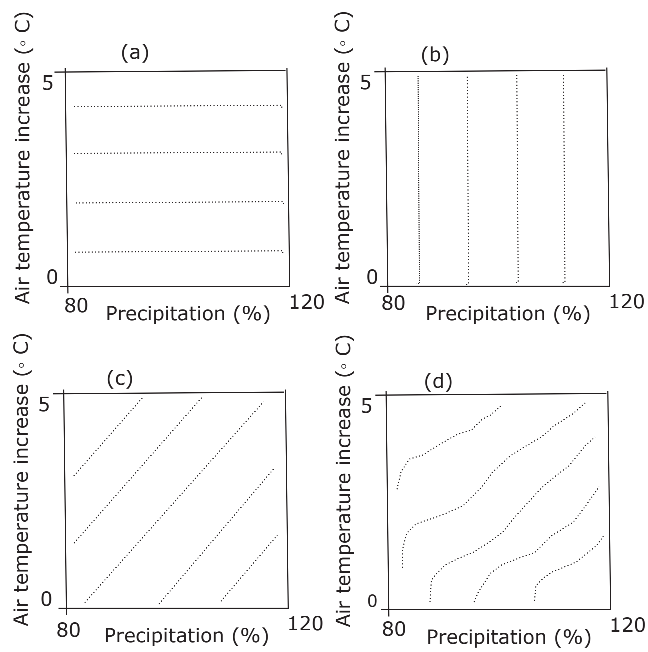

Fig. 2. Guide to interpreting hydrological sensitivity to air temperature increase (0 °C to 5 °C) and to changes in precipitation (−20% to + 20%). The shape and slope of contours represent the sensitivity of a hydrological variable to a) air temperature increase, b) precipitation change, c) a linear interaction of air temperature and precipitation changes, and d) a complex interaction of air temperature and precipitation changes.

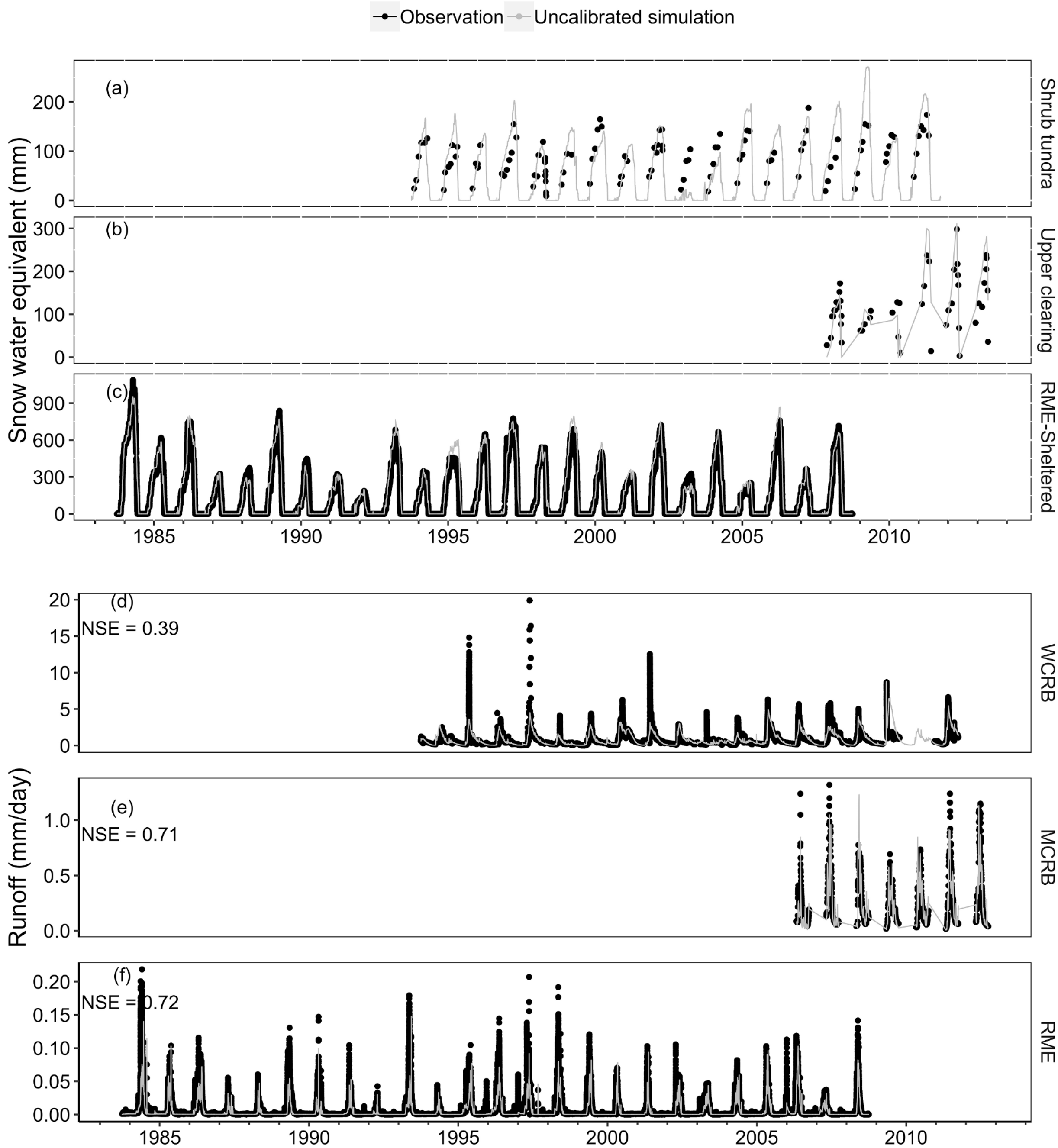

Fig. 3. Performance of the developed Cold Region Hydrological Models in capturing snow water equivalent at the a) shrub tundra station in Wolf Creek, b) Upper Clearing station in Marmot Creek, c) sheltered station in Reynolds Mountain, and runoff at the outlets of d) Wolf Creek, e) Marmot Creek and f) Reynolds Mountain.

10% intervals. These changes were applied to observations from all three basins.

The degree of hydrological sensitivity to climatic changes is evident in the resulting shape and slope of contours of change in a variable (Fig. 2). The contours were estimated by linear interpolation between the mean responses to the 30 combinations of warming (0 °C to 5 °C in intervals of 1 °C) and precipitation change (−20% to + 20% in intervals of 10%). When elasticity of a hydrological variable is high and it is more sensitive to air temperature increase or precipitation change, the contour line is perpendicular to that axis (Temperature, Fig. 2a; Precipitation, Fig. 2b). When the variable is sensitive to a linear interaction of air temperature and precipitation changes, there will be a slope in the contour line (Fig. 2c). If the interaction is complex, the slope and the contours will not be straight lines (Fig. 2d). Applying the same ranges of change in air temperature and precipitation to each of the three basins allows direct comparison of the responses of the simulation model for each basin. Different combinations of warming and precipitation change make it possible to estimate how much additional precipitation is needed to offset the impacts of a specific air temperature increase on annual runoff and peak snowpack. The additional precipitation increases needed to offset the temperature increases were estimated based on the interpolation of the two contour lines above and below the present climate values.

Precipitation elasticity (unitless) is the percent change in peak snowpack or runoff per percent change in precipitation, and air temperature elasticity (% $°C^{-1}$) is the percent change in peak snowpack and runoff per 1 °C warming. These were estimated by averaging responses to the 30 combinations of warming and precipitation change and

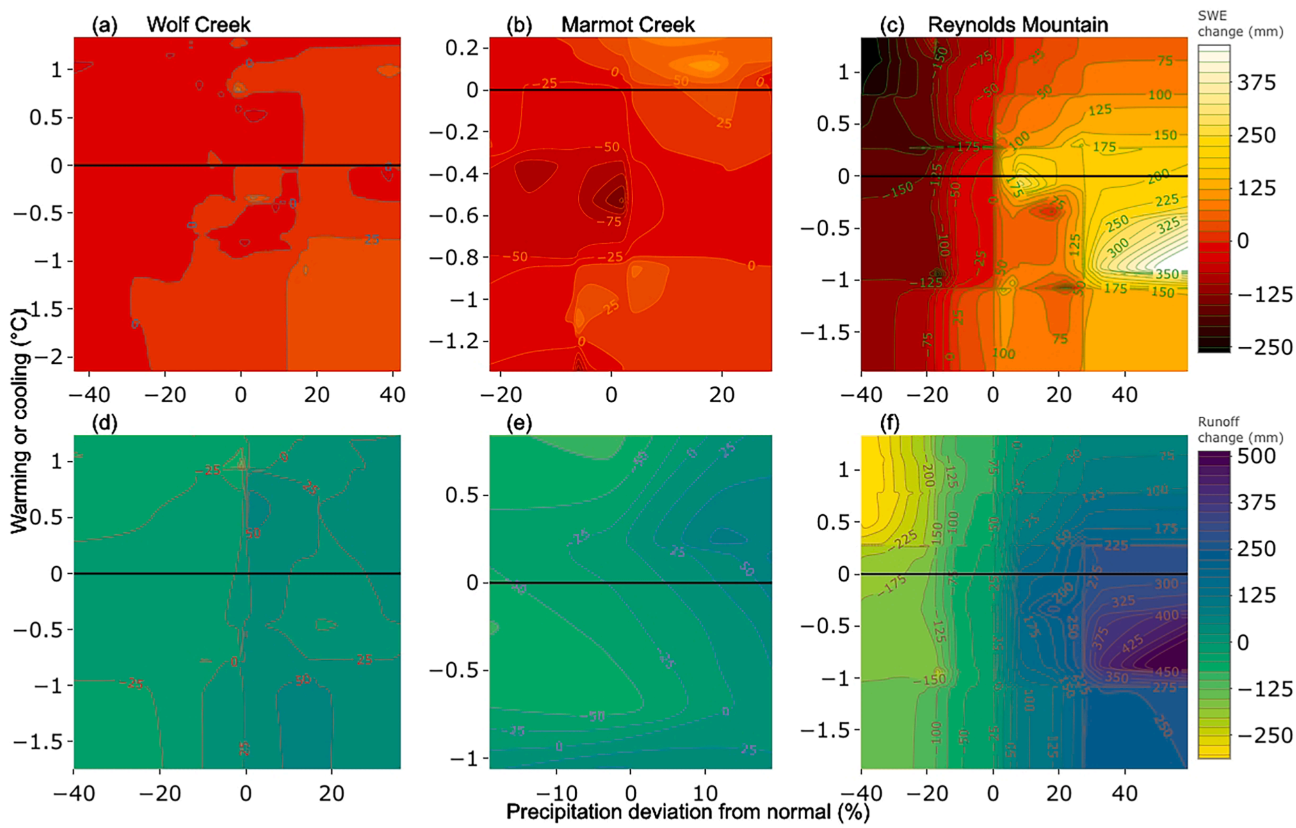

Fig. 4. Observed sensitivity of annual peak snow water equivalent (SWE) to warming at the a) shrub tundra station in Wolf Creek, b) Upper Clearing in Marmot Creek and c) sheltered station in Reynolds Mountain, and precipitation elasticity of annual runoff volume at the outlets of d) Wolf Creek, e) Marmot Creek and f) Reynolds Mountain. The black horizonal line separates warmer and colder than long-term average years.

obtaining the ratio of runoff and precipitation changes $\Delta Q/\Delta P$ and the ratio of runoff and temperature changes $\Delta Q/\Delta T$ in the three mountain basins. The precipitation elasticity of runoff can be reliably estimated by calculating the slope of the linear fit to the scatter plots (Hoerling et al., 2019).

Sensitivities of five main characteristics that describe a basin snow regime were investigated. These characteristics are the timing of snowcover initiation (snow season start), snow-free date (snow season end), duration of the snow season, duration of snowmelt period, and magnitude of the peak snowpack. The duration of the snow season is the difference between the date of snowcover initiation and the date the basin becomes snow free. The duration of the melt period is the difference between the date of peak snow water equivalent (SWE) and the date the basin becomes snow free. The mean annual peak snow accumulation is defined as the average maximum SWE over the hydrological year and occurs in March or April in these basins.

4. Results

The performance of the models developed to study the sensitivities of annual runoff and snow regimes to warming and precipitation elasticity was evaluated against actual observations. The models represented observed SWE and runoff at the outlet of the basins reasonably well, based on a Nash–Sutcliffe efficiency score of 0.72 for runoff (Fig. 3) and the fact that the models were physically based with no calibration for snow simulations and a minimal calibration for routing parameters. The robustness of the model in representing snowpack is more apparent at the station in the sheltered HRU in Reynolds Mountain, where there is a snow pillow with continuous snow records (Fig. 3c). Relatively low performance of the model in capturing peak runoffs in Wolf Creek (Fig. 3d) is due to the streamflow outburst associated with rapid ice breakup at the outlet of Coal Lake within this basin (Fig. 1a), which generates high flows in the spring melt season and is due to a mechanism which is not represented in CRHM.

The inter-annual anomalies relative to the long-term averages were assessed for observed air temperature, precipitation, annual peak snowpack, and annual runoff in the three basins (Fig. 4). The elasticities and sensitivity of observed peak SWE and annual runoff to warming anomaly and precipitation changes were determined. Mean annual air temperature in all three basins has less inter-annual variability than precipitation, which makes it challenging to assess impacts of future warming on snow and runoff using observed analogies (Fig. 4). In contrast, the variability of the precipitation anomalies, which were obtained from subtracting annual precipitation from the long-term average, were sufficiently large to allow assessment of future changes using observed inter-annual variability of precipitation. Observed precipitation and temperature elasticities showed that peak SWE (Fig. 4a) and annual runoff (Fig. 4d) were both sensitive to precipitation change and warming in Wolf Creek. Precipitation elasticity showed that peak SWE (Fig. 4b) and annual runoff (Fig. 4e) were more sensitive to precipitation change in Marmot Creek and, because the maximum warming anomaly observed is less than 1 °C, it is difficult to project warming effect on snow and runoff regimes using observed data. A modest observed warming of 1 °C decreased peak SWE by 50 mm (Fig. 4c) and annual runoff by 75 mm in Reynolds Mountain (Fig. 4f), and the slope of the contours showed that there was a complex interaction of observed air temperature and precipitation changes in this basin. The models developed in this study were physically based and most parameters were uncalibrated, and their performance in representing observations made them suitable for studying future conditions and conducting sensitivity analyses with high confidence. The modelling uncertainties in representing observed high flows and hydrological sensitivities to changes in

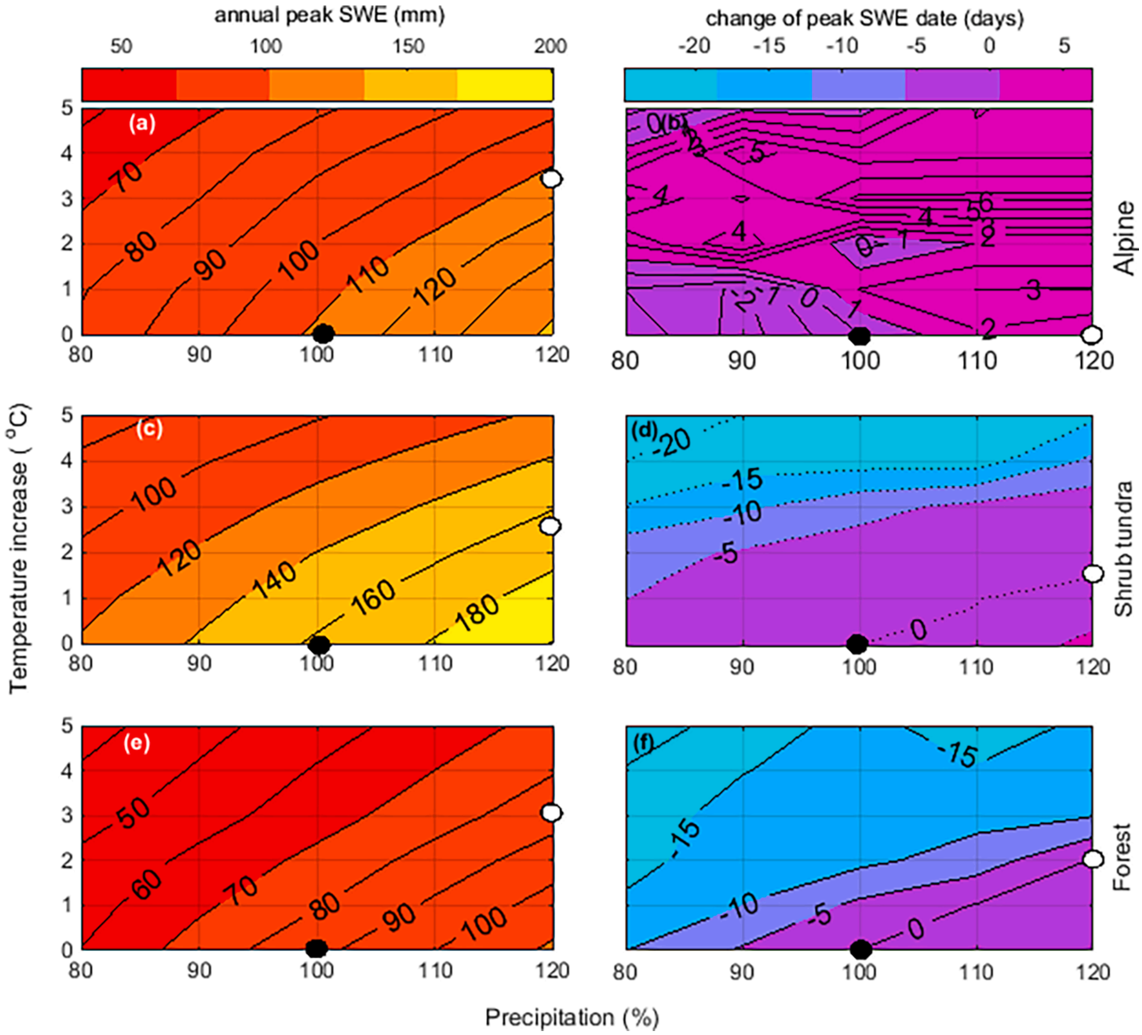

Fig. 5. Sensitivity of mean annual peak snow water equivalent (SWE) (left column) and timing of peak SWE (right column) in Wolf Creek with warming of 0 °C to 5 °C and changes in precipitation from −20 to + 20% in the three zones. The black dot indicates the present climate, and the white dot indicates the temperature increase that a 20% increase in precipitation can offset.

air temperature and precipitation (Fig. 4) suggested that these uncertainties could propagate to the simulations for sensitivity analyses. Therefore, one needs to interpret these temperature results with care. These uncertainties were partly unavoidable, as the analogies of future conditions were limited in historical data, especially for warmings greater than 1 °C.

Similar to the observed sensitivities (Fig. 4a), the magnitude and timing of annual peak snowpack are sensitive to both air temperature and precipitation changes in Wolf Creek, as shown in Fig. 5. The interaction between air temperature and precipitation affecting peak SWE is evident in the curvature and slope of the contours; the interaction is complex in the alpine and shrub tundra (curved contour lines, Fig. 5a and c) but less so in the forest (Fig. 5e). The sensitivity of peak SWE to precipitation is somewhat higher in the high elevation alpine zone (contours have higher slope) and its sensitivity to temperature is somewhat higher in the lower elevation shrub tundra and forest zones (contours have lower slope). The peak SWE in the shrub tundra zone is very sensitive to a decrease in precipitation with warming due to precipitation phase change and suppression of blowing snow redistribution from the alpine zone under warmer air temperatures (Rasouli et al., 2014) and drops from 162 mm to 75 mm (87 mm reduction, Fig. 5c) with 80% of precipitation and +5 °C of warming. The sensitivity of peak SWE to increasing precipitation in the shrub tundra zone declines as the temperature warms. The peak SWE in the forested zone is slightly less sensitive to temperature than shrub tundra because unloading of intercepted snow from the canopy, where it is prone to sublimation, increases with winter air temperatures and moderates the impact of declining snowfall with rising temperature. Whether 20% additional precipitation can offset the effect of warming on snowpacks in Wolf Creek is illustrated in Fig. 5 by comparing the black dot, indicating no change in air temperature or precipitation, to the white dot, indicating the degree of warming that can be offset by a 20% increase in precipitation. A 20% increase in precipitation can offset the effect of warming of 3.5 °C in the alpine zone (Fig. 5a), 2.7 °C in the shrub tundra (Fig. 5c), and 3 °C in the forest zone (Fig. 5e) on peak SWE.

There is no clear pattern to the small changes, of less than six days, in the timing of peak SWE in the Wolf Creek alpine zone with air temperature and precipitation changes (Fig. 5b). This is likely due to the persistently colder temperatures during winter at high elevations in the subarctic (Fig. 5b). In the shrub tundra and forest, the mean annual peak SWE occurs 25 and 20 days earlier, respectively, with 5 °C of warming and 20% reduced precipitation (Fig. 5d and f).

In cold continental Marmot Creek, peak SWE in all zones is influenced by changes in both air temperature and precipitation but responds more strongly to air temperature than in subarctic Wolf Creek (Fig. 6). Peak SWE is more influenced by warming temperature at lower elevations due to the influence of lapse rates on precipitation phase and other factors. Because of reduced blowing snow inputs from the alpine zone,

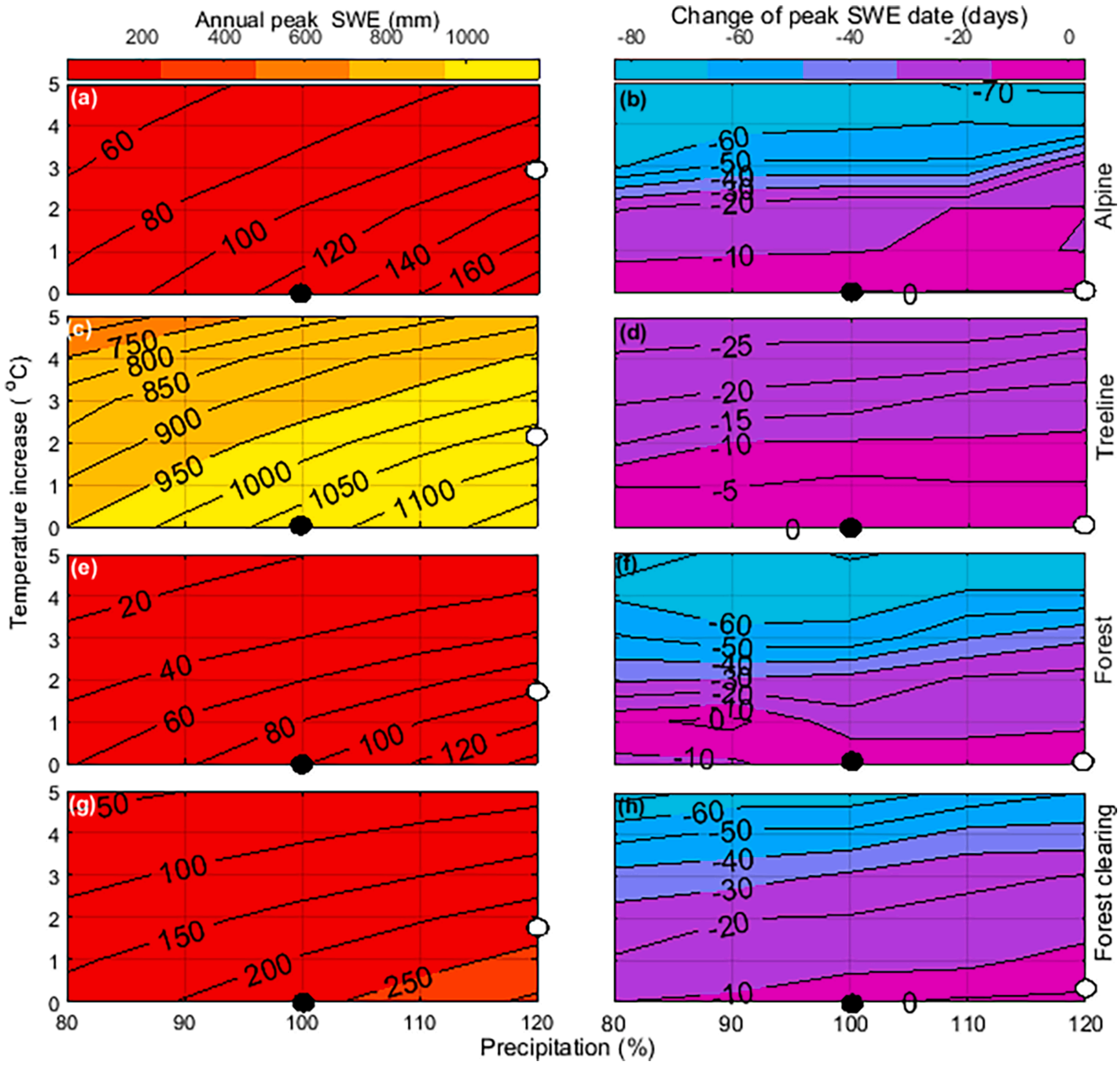

Fig. 6. Sensitivity of mean annual peak snow water equivalent (SWE) (left column) and timing of peak SWE (right column) in Marmot Creek with warming of 0 °C to 5 °C and changes in precipitation from −20 to + 20% in four zones. The black dot indicates the present climate, and the white dot indicates the temperature increase that a 20% increase in precipitation can offset.

the treeline forest zone loses the most snow (−422 mm under 5 °C of warming and 20% less precipitation, Fig. 6c), but because it has the highest snow accumulation, snow is still deep and its proportional change with temperature was not substantially different from the other zones. In contrast, almost all SWE is lost in the Marmot Creek forest zone under this scenario, suggesting a high sensitivity of snowpacks to warming and drying in this elevation zone. The response of the peak SWE to warming and precipitation changes shows that an increase in precipitation of 20%, slightly greater than the maximum indicated by climate models, can offset the effect on peak SWE of warming in the alpine of 2.9 °C (Fig. 6a), in the treeline forests of 2.1 °C (Fig. 6c) and in the forest and forest clearing of 1.8 °C (Fig. 6c and g). The peak snowpack in Marmot Creek is more sensitive to warming, and so increased precipitation can offset less of a temperature increase than in Wolf Creek.

The changes in the simulated timing of peak SWE in Marmot Creek are substantial and complex. Timing responded much more to warming than to precipitation change and precipitation increases could not compensate for any degree of warming at any elevation (Fig. 6b, d, f, and h). In the alpine, forest, and forest clearing zones, peak SWE advanced between 19 and 28 days for 2 °C of warming, and between 60 and 70 days for 5 °C. In contrast, the treeline forest peak SWE timing advanced only 10 and 27 days for 2 °C and 5 °C of warming, its lower sensitivity (range of contours) due to the high snow accumulation in this zone associated with continued redistribution of snow from the alpine (Fig. 6d).

In Reynolds Mountain, annual peak SWE is very sensitive to increases in air temperature and much less sensitive to changes in precipitation (Fig. 7a, c, e, and g). The slope and curvature of the annual peak SWE contours show the sensitivity to precipitation change decreases as temperature increases. This suggests that the effects of warming on SWE cannot be easily offset by increased precipitation; a precipitation increase of +20% can offset warming up to from 1.2 °C to 1.5 °C depending on location. The warmest and driest scenario (+5°C and –20% precipitation) resulted in peak SWE declines in all zones, e.g., from 570 mm to 58 mm in the sink (Fig. 7a) and from 427 mm to 39 mm in the interception zones (Fig. 7e). The blowing snow sink zone lost more snow with warming and drying than other zones due to the suppression of blowing snow transport from the source zone (Fig. 7a).

The response of the timing of annual peak SWE is much more sensitive to warming than to precipitation change in all zones in Reynolds Mountain (Fig. 7b, d, f, and h). The timing changes in Reynolds Mountain are the largest amongst the three basins, with the change in peak SWE date being between 50 and 70 days earlier for the maximum 5 °C

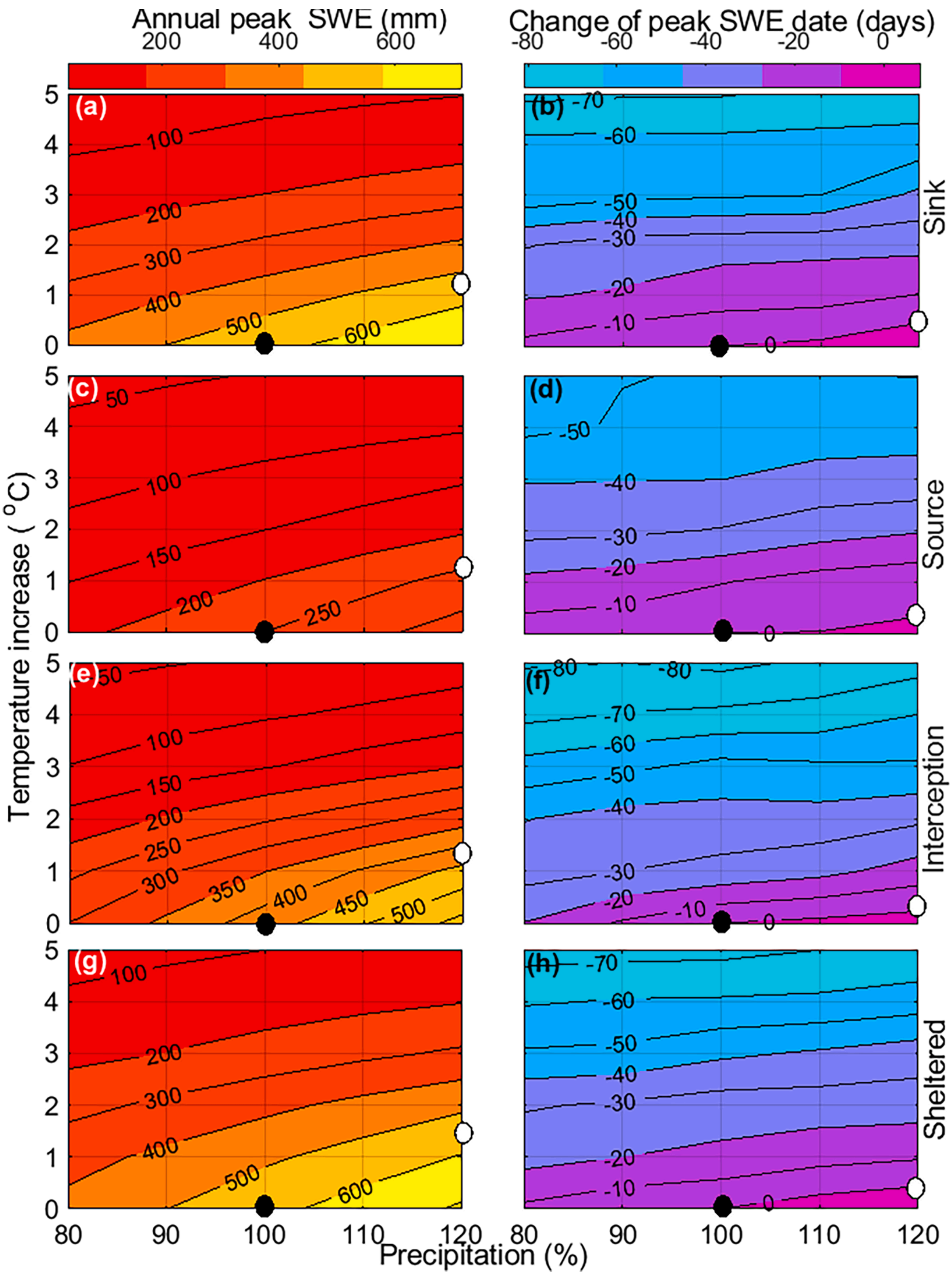

Fig. 7. Sensitivity of mean annual peak snow water equivalent (SWE) (left column) and timing of peak SWE (right column) in Reynolds Mountain with warming of 0 °C to 5 °C and changes in precipitation from −20 to + 20% in four zones. The black dot indicates the present climate, and the white dot indicates the temperature increase that a 20% increase in precipitation can offset.

warming. Additional precipitation of 20% can only offset the effect of 0.5 °C of warming on peak SWE date (Fig. 7b, d, f, and h).

The rate of change in the simulated snowpacks can be estimated in relation to air temperature. Air temperature elasticity of peak SWE is 8% $°C^{-1}$ in Wolf Creek, 10% $°C^{-1}$ in Marmot Creek, and 17% $°C^{-1}$ in Reynolds Mountain (Table 2). Since the air temperature elasticity is defined as percent change of peak SWE per degree (not percent) of warming, it has higher values than the precipitation elasticity, which is unitless and defined as percent change of peak SWE per percent change of precipitation. The loss of snowpack with warming is reflected in the reduction in the snowcover duration of 11 days in Wolf Creek, 18 days in Marmot Creek, and 30 days in Reynolds Mountain per degree of warming (Table 2). The duration of snowmelt declines between 0 and 9 days per degree of warming in all basins, much less than the snowcover duration,

Table 2
Sensitivity and elasticity of snow and runoff regimes per 1 °C warming in the three mountain basins.

	unit	Wolf Creek	Marmot Creek	Reynolds Mountain
ΔPeak SWE	[mm (%)]	−10 (−8)	−22 (−10)	−66 (−17)
ΔAnnual runoff	[mm (%)]	−5 (−3)	−3 (−1)	−22 (−6)
ΔSnow initiation	[day]	3	5	9
ΔPeak SWE timing	[day]	−3	−13	−12
ΔSnow-free timing	[day]	−7	−13	−21
ΔSnowcover duration	[day]	−11	−18	−30
ΔMelt duration	[day]	−5	0	−9
Simulation period	[year]	18	9	25

and smaller than the advance in the timing of snow disappearance, which ranges from 7 (Wolf Creek) to 13 (Marmot Creek) to 21 (Reynolds Mountain) days per degree of warming. Snow melts more slowly as the melt season advances in some of these simulations, which partly offsets the impact of the decrease in peak snowpack on snowmelt period duration.

The sensitivity of five main characteristics of basin snow regimes to warming and change in precipitation averaged over Wolf Creek shows that both changes in precipitation and warming affect the magnitude of the peak SWE (Fig. 8a). Precipitation increases of 20% can offset a 3 °C temperature increase in Wolf Creek peak SWE. Delay in the initiation (start) of snow accumulation is sensitive to warming rates above 3 °C regardless of precipitation changes (Fig. 8b). The snow-free (snow season end) date advances from June 28 in the recent climate to June 11 with a warming of 2 °C (Fig. 8c, Table 3). The snow-free date is also sensitive to warming and almost insensitive to precipitation changes (Fig. 8c). The snow season duration in Wolf Creek is also driven by warming and not by precipitation changes (Fig. 8d). The snowmelt period, the timing difference between peak SWE and the snow-free date, is sensitive to warming and almost insensitive to precipitation changes (Fig. 8e).

In Marmot Creek, the peak SWE drops from 220 mm to 92 mm under a warming of 5 °C and decreasing precipitation (−20%, Fig. 8f, Table 3). The timing of the start of snow accumulation is not affected substantially by either warming or precipitation (Fig. 8g), but increased temperatures have a large effect on the end date (Fig. 8h) and snow season duration (Fig. 8i); this is reduced with increased precipitation. The duration of the melt season is also not affected (Fig. 8j). In contrast to Wolf Creek, the initiation date of snow accumulation is sensitive to precipitation changes and would advance if warming rates are below 2 °C and precipitation increases. The snow-free date advances from early June in the recent climate to late May with a warming of 2 °C (Fig. 8h). Similar to the ablation period, the snow accumulation start date is sensitive to precipitation changes and, to a lesser extent, to warming. With concomitant warming (5 °C) and decreasing precipitation, the snow-free date across the basin advances by 77 days to late March (Fig. 8h). As shown in Fig. 8, the snow-free date is sensitive to warming but insensitive to precipitation changes in Marmot Creek, and snow season length is affected by both warming and precipitation changes. Similar to Wolf Creek, the combination of air temperature increasing by at least 2 °C and precipitation increasing by less than 20% results in declining peak SWE and deviation from the historical ranges of snowpack in Marmot Creek.

In Reynolds Mountain, with warming of 5 °C and decreasing precipitation of 20%, the mean annual peak SWE decreases from 390 mm to 47 mm (Fig. 8k, Table 3), and snow accumulation starts later (Fig. 8l) and ends earlier (Fig. 8m). The duration of the snow season (Fig. 8n) and duration of the melt period snow season (Fig. 8o) become much shorter than in the present climate (Table 3). A 1 °C warming advances the timing of peak SWE by approximately 15 days (Table 2). The magnitude of peak SWE is more sensitive to temperature than precipitation (Fig. 8k); the timing of the snow regime is sensitive to temperature and less so to precipitation (Fig. 8l-o).

The peak SWE is 136 mm in Wolf Creek, 220 mm in Marmot Creek, and 390 mm in Reynolds Mountain; Wolf Creek and Reynolds peak SWE occur in early March, and in Marmot Creek it occurs in late April (Table 3). With a 20% decline in precipitation and a warming of 5 °C, peak SWE declines to 61 mm (55% decrease) in Wolf Creek, to 92 mm (58%) in Marmot Creek, and to 47 mm (88% decrease) in Reynolds Mountain. With a 20% increase in precipitation and no warming, peak SWE increases to 169 mm (24%) in Wolf Creek, to 281 mm in Marmot Creek (28%), and to 486 mm (25%) in Reynolds Mountain. With 5 °C warming and no changes in precipitation, the onset of snow accumulation is delayed 17 days in Wolf Creek, 23 days in Marmot Creek, and 42 days in Reynolds Mountain, and the end of winter comes earlier by 37 days in Wolf Creek, 67 days in Marmot Creek, and 104 days in Reynolds Mountain. When compared to no changes (Table 3), a 20% increase in precipitation would lengthen the snowcover duration by 5 to 20 days.

The simulated air temperature elasticities and sensitivities of simulated annual runoff (ΔQ) and peak snowpack (ΔSWE) changes to air temperature change (ΔT) were compared against the observed elasticities in the present climate for the three basins (Fig. 9). Observed and simulated changes were obtained by subtracting annual magnitudes from the long-term averages. The air temperature elasticities of runoff, the slopes of $\frac{\Delta Q}{\Delta T}$ in the observed and simulated scenarios, matched well in Wolf Creek and Marmot Creek, with less than 6% discrepancy (Fig. 9a and 9b). The discrepancy between observed and simulated precipitation elasticity of runoff, however, was relatively high (38% °C^{-1}) in Reynolds Mountain. The observed and simulated air temperature elasticities matched well in Wolf Creek (Fig. 9d). But, because of the shorter observation record in Marmot Creek, the inter-annual variability of observed air temperatures did not cover the range of future warming projected by climate models (Fig. 9e). The simulated air temperature elasticity of peak SWE was underestimated in Reynolds Mountain, as twice the observed elasticity (Fig. 9f). The observed slopes of $\frac{\Delta Q}{\Delta P}$, which represent the precipitation elasticity of runoff for the present climate, matched well with the projected changes in annual runoff, and the discrepancy between observed and simulated elasticities was less than 0.8 (Fig. 10a–c). The projected precipitation elasticity of annual runoff are 1.7 in Wolf Creek (Fig. 10a), 1.5 in Marmot Creek (Fig. 10b) and 2.1 in Reynolds Mountain (Fig. 10c). The observed slopes of the $\frac{\Delta SWE}{\Delta P}$, which represent observed precipitation elasticity of peak snowpack in the present climate, also matched well with the projected precipitation elasticity of peak snowpack. The projected precipitation elasticity of peak SWE, $\frac{\Delta SWE}{\Delta P}$ is from 0.7 to 1.1, and the observed precipitation elasticity of peak snowpack is from 0.4 to 1.9 in the three basins (Fig. 10d and f), with discrepancies ranging between 0.1 and 1.2 (Table 4). The lower discrepancy between observed and simulated precipitation elasticities of runoff and peak SWE (Fig. 10) showed that the results of sensitivity analysis are reliable and have less uncertainty in all three basins. The higher discrepancy between observed and simulated air temperature elasticities of runoff in Reynolds Mountain (Fig. 9c) and peak SWE in Marmot Creek (Fig. 9e) showed that the results of sensitivity analysis are less reliable in these two basins, and one must interpret the results with care.

The simulations show that changes in snow regime in these mountain basins result in smaller changes in mean annual runoff than peak SWE. Unlike peak SWE, mean annual runoff is more sensitive to changes in precipitation than air temperature (Fig. 11). The near vertical lines in Fig. 11a and b indicate that changes in mean annual runoff are driven predominately by precipitation in Wolf Creek and Marmot Creek, whilst temperature more strongly impacts runoff in Reynolds Mountain. A 1 °C warming in Wolf Creek resulted in a 3% decrease in the annual runoff (Table 2); total decreases rise to ~14% for a 5 °C warming (171 to 147 mm, Table 3, Fig. 11a). The most extreme scenario of climate warming and decreased precipitation caused larger declines in runoff but, if

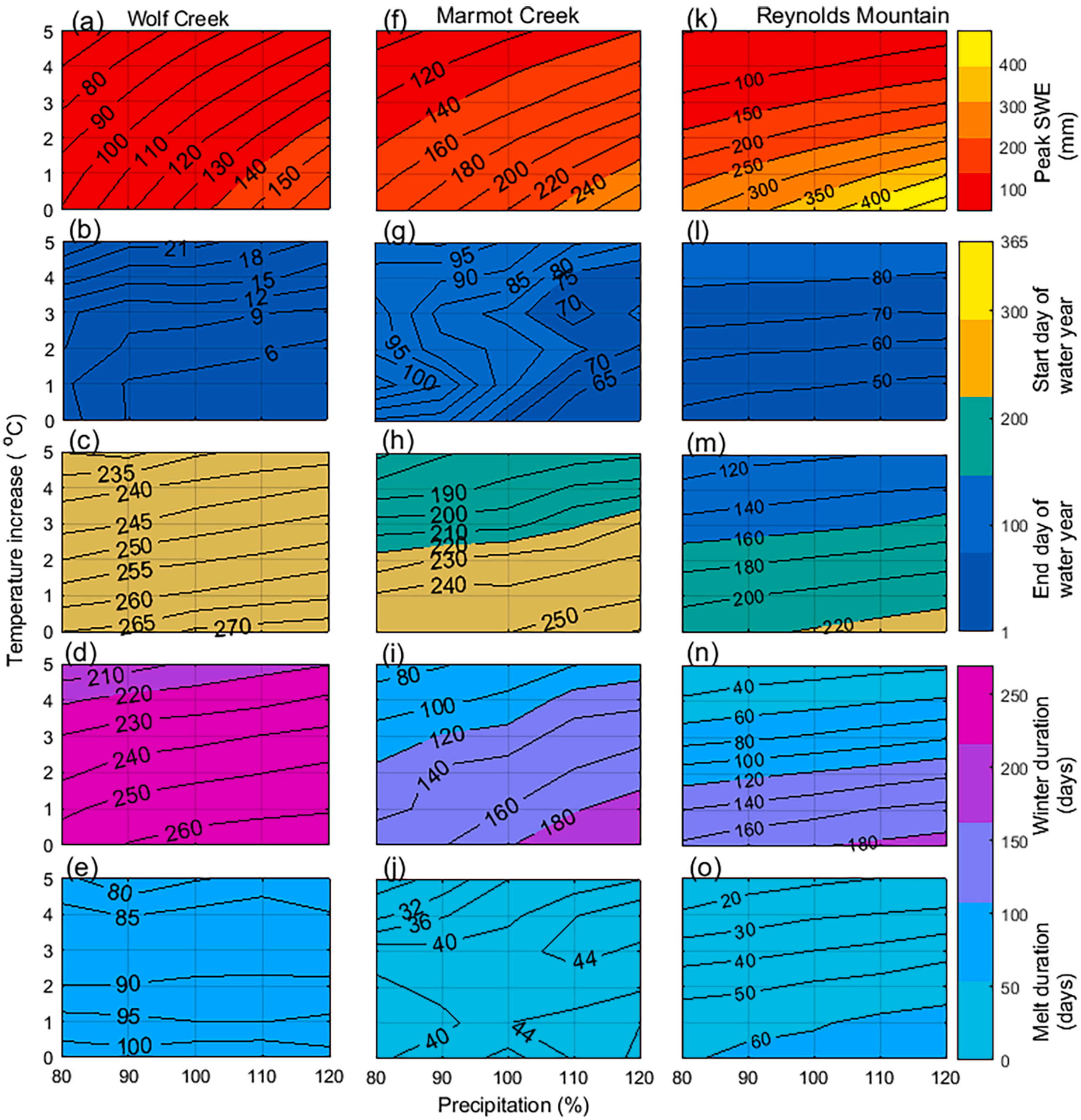

Fig. 8. Magnitude and change of mean annual peak snow water equivalent (SWE), and the timing shift of the snow season start/end, snow season duration, and snowmelt period in the three basins with warming up to 5 °C and precipitation change up to ± 20%. Day of water year for the first day of each month are: 1:Oct, 32: Nov, 62:Dec, 93:Jan, 124:Feb, 152:Mar, 183:Apr, 213:May, 244:Jun, 274:Jul, 305:Aug, 336:Sep.

precipitation increases, there is strong compensation. For instance, if precipitation increases by 20%, annual runoff increases by 35 mm (from 171 to 206 mm) with 5 °C of warming. Mean annual runoff is more sensitive than snow regime to precipitation change in Wolf Creek. Similarly, in Marmot Creek, a 5 °C increase in air temperature results in a 4% decrease in the mean annual runoff (402 to 384 mm Table 3, Fig. 11b). The combination of 5 °C of warming and 20% decreased precipitation reduces mean annual runoff by 34% (135 mm from 402 to 267 mm, Table 3, Fig. 11b). In Reynolds Mountain, mean annual runoff has a stronger temperature sensitivity than Wolf Creek or Marmot Creek (Fig. 11). A 5 °C increase in temperature results in a 29% (371 to 263 mm, Table 3) decrease in the mean annual runoff. The combination of 5 °C of warming and 20% decrease in precipitation reduces annual runoff by 43%, (371 to 161 mm, Table 3).

Change in mean annual runoff (Fig. 11) contrasts with the change in mean annual peak SWE (Figs. 5-7) in that mean annual runoff is more sensitive to precipitation than temperature. The sensitivity of annual runoff to temperature increase in Reynolds Mountain is because of the longer snow-free season and an increased growing season and energy flux for evapotranspiration with increasing temperature (Fig. 7), whilst runoff responds to both precipitation change and warming (Fig. 11c). In contrast to the sensitivity of snowpack to warming in Reynolds Mountain, annual runoff is less sensitive, and the impact of warming on annual runoff can be partly offset by an increase in precipitation in Reynolds Mountain.

Annual runoff changes are given in Table 3 under different scenarios

Table 3
Sensitivity of the snow and runoff variables to warming and changes in precipitation in three mountain basins.

Variable	No change	warm -	warm -	warm dry	- wet	warm wet
Warming [°C]	0	2	5	5	0	5
Precipitation [%]	100	100	100	80	120	120
(1) Wolf Creek						
Peak SWE [mm]	136	117	85	61	169	107
Annual runoff [mm]	171	160	147	96	236	206
Initiation [date]	Oct 05	Oct 07	Oct 22	Oct 27	Oct 04	Oct 18
Peak SWE [date]	Mar 16	Mar 12	Mar 03	Feb 25	Mar 21	Mar 04
Snow-free [date]	Jun 28	Jun 11	May 22	May 18	Jun 30	May 25
Snowcover duration [days]	265	248	212	202	269	219
(2) Marmot Creek						
Peak SWE [mm]	220	176	115	92	281	140
Annual runoff [mm]	402	397	384	270	527	518
Initiation [date]	Dec 10	Dec 23	Jan 02	Jan 08	Dec 01	Dec 20
Peak SWE [date]	Apr 30	Apr 10	Feb 24	Feb 18	Apr 30	Feb 25
Snow-free [date]	Jun 07	May 22	Apr 01	Mar 22	Jun 18	Apr 07
Snowcover duration [days]	179	150	89	73	199	108
(3) Reynolds Mountain						
Peak SWE [mm]	390	222	63	47	486	80
Annual runoff [mm]	371	331	263	161	533	415
Initiation [date]	Nov 11	Nov 28	Dec 25	Dec 23	Nov 09	Dec 23
Peak SWE [date]	Mar 07	Feb 08	Jan 07	Jan 04	Mar 11	Jan 09
Snow-free [date]	May 10	Apr 02	Jan 26	Jan 19	May 18	Jan 30
Snowcover duration [days]	180	125	32	27	189	38

of warming and changes in precipitation. Annual runoff responds strongly to precipitation changes in Wolf Creek and Marmot Creek, and to both warming and precipitation changes in Reynolds Mountain. Annual runoff is the most resilient to warming in Marmot Creek and most sensitive to warming in Reynolds Mountain. Under 5 °C and a 20% increased precipitation, annual runoff increases from 171 mm to 206 mm (20%) in Wolf Creek and increases from 402 mm to 518 mm (29%) in Marmot Creek and from 371 mm to 415 mm (12%) in Reynolds Mountain (Table 3). This shows that increased precipitation with warming increases the runoff in Marmot Creek more than the other two basins. This is due to the very cold alpine snowpack at Marmot Creek, which is relatively unaffected by warming, in contrast to the warm snowpacks at Reynolds Mountain, which become ephemeral with warming.

The amount of additional precipitation needed to offset the effect of increased temperature on peak SWE and annual runoff under future climate can be estimated from the sensitivity analysis. The largest increases in precipitation projected by the SRES and RCPs used here and NARCCAP RCM–GCMs are 34% for Wolf Creek, 18% for Marmot Creek, and 16% for Reynolds Mountain. This also covers the range from SSP5 8.5. Similarly, in Wolf Creek, when warming is limited to 1 °C, increased precipitation of 4% can offset the effect of warming on peak SWE (Fig. 12a) but, with warming of 5 °C, an increase in precipitation of 34% would be required to offset the effect of warming. In Marmot Creek, the effect of 1 °C warming on peak SWE can be offset by 8% increase in precipitation; however, the effect of 5 °C warming on peak SWE would require precipitation increases that are greater than expected from RCP scenarios and NARCCAP simulations. In Reynolds Mountain, the impact of 1 °C warming on peak SWE can be offset by 16% increase in precipitation, but the offset required for more than 2 °C warming exceeds the projected maximum precipitation increases.

Annual runoff is less sensitive than peak snowpack to warming, and smaller precipitation increases are required to offset the effects of warming simulated here. These differences are due to differences in the fraction of snowfall converted to rainfall in each basin under a warmer climate. The additional precipitation needed to offset the impact of warming on runoff varies with elevation range, precipitation regime and latitude; offsetting the effect of warming of 5 °C on annual runoff would require precipitation increases of 8% in Wolf Creek (Fig. 12a), 3% in Marmot Creek (Fig. 12b), and 14% in Reynolds Mountain (Fig. 12c). Rasouli et al. (2019a) showed that the reduced snow sublimation loss offset reduced snowfall amounts, and increased evapotranspiration loss offset increased rainfall amounts. The evapotranspiration to precipitation ratio did not substantially change under a moderate warming of 2.2–2.6 °C in the three basins (Table 5), and the increased evapotranspiration was largely due to enhanced wetness as a result of increased precipitation and to less extent due to warming. Here, an energy balance method to calculate actual evapotranspiration, restrained by continuity in surface and soil moisture availability was used, and relative humidity rather than absolute humidity was held constant with warming, so the overestimation of evapotranspiration with climate warming that is a feature of empirical and potential evapotranspiration schemes, was avoided with the CRHM models (Milly and Dunne, 2016).

5. Discussion

Air temperatures in late winter and spring have a key role in determining the sensitivity of snowpack in mountain basins to warming (Lettenmaier and Gan, 1990; Stewart et al., 2004; McCabe and Clark, 2005). The results presented above demonstrate and quantify the sensitivity of annual peak snowpack and its timing, and of annual runoff, to air temperature and precipitation changes and their interaction in the three basins. Sensitivity of annual peak snowpack timing to air temperature and precipitation changes in the three basins shows that the sensitivity of peak SWE timing to precipitation changes is greater in the colder climate conditions: Reynolds Mountain responds to warming only (Fig. 7), Marmot Creek responds to warming and, to a lesser extent, to precipitation (Fig. 6); and Wolf Creek responds to a complex interaction of warming and precipitation change (Fig. 5). The potential for precipitation to counteract the effect of warming on the magnitude of the annual peak snowpack becomes smaller as latitude decreases. Therefore, regional responses to warming and changes to precipitation must be considered (Bower et al., 2004), particularly when evaluating future mountain hydrology (Roche et al., 2018; Sultana and Choi, 2018). This is because the snowpack is shallow and warm at the beginning and end of the season; shallow warm snow ripens and melts faster than deep cold snow, as it requires less energy input to overcome cold content and fill its liquid water holding capacity (Colbeck, 1976).

Simulations of future conditions for snow regimes in Reynolds Mountain are in accord with the SWE magnitude and timing trajectories of the past 50 years (Nayak et al., 2010). The future simulations of peak SWE and runoff elasticities are compared with observed analogies (Figs. 9 and 10). Precipitation and air temperature elasticities show minimal discrepancies, except for air temperature elasticity of peak SWE in Marmot Creek (Fig. 9e) and runoff in Reynolds Mountain (Fig. 9c). Higher discrepancies in the air temperature elasticities in these two basins are likely due to a wider warming range used in the sensitivity analyses (0 °C to 5 °C), which is not covered in the inter-annual variability of the observation period (less than 1.5 °C, Fig. 4). Higher rates of warming and increased precipitation are projected by RCMs in the

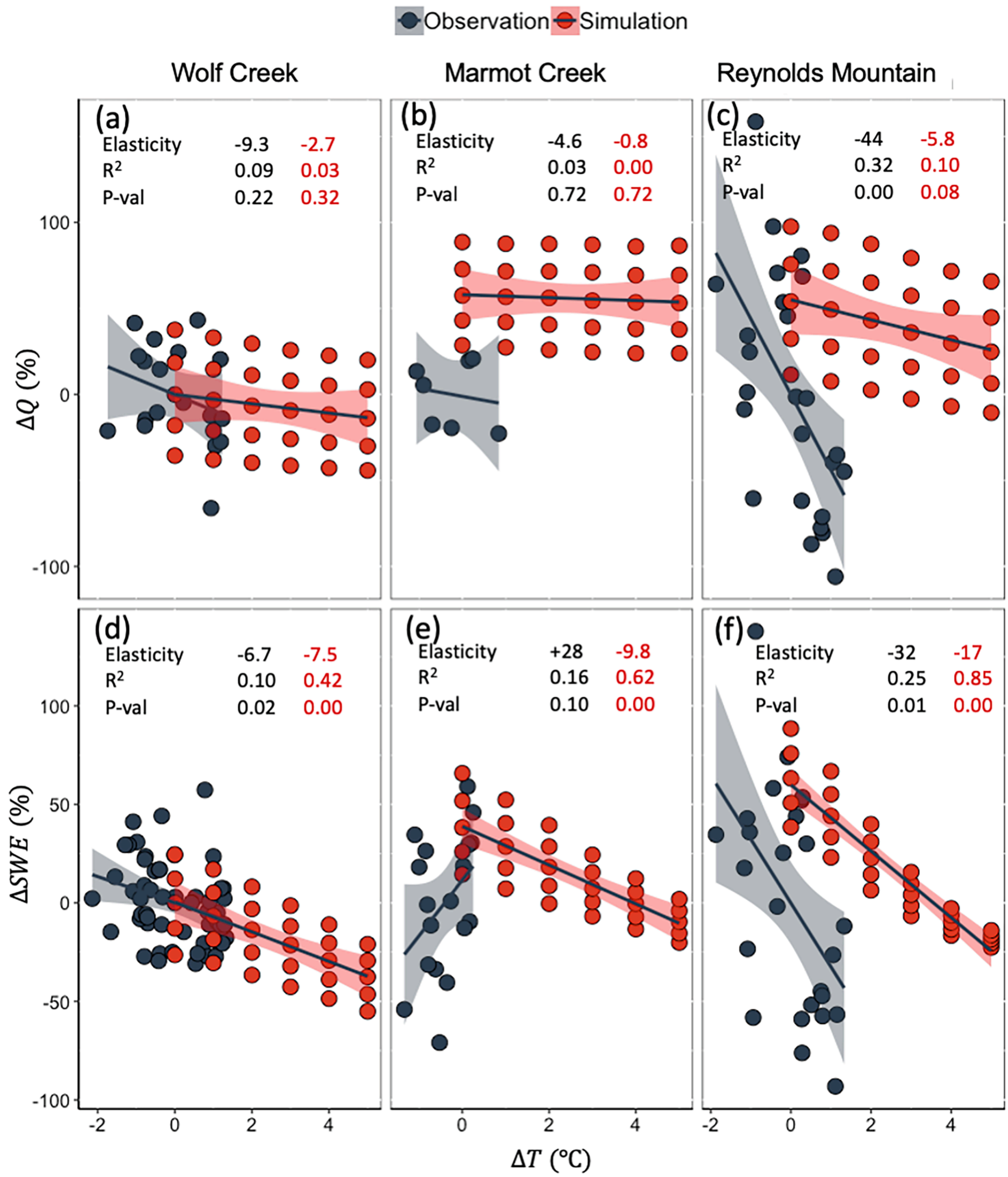

Fig. 9. Observed and simulated sensitivities of annual runoff (ΔQ) and peak snow water equivalent (ΔSWE) changes to air temperature change (ΔT) in the three basins. The black circles show the observations and the red circles represents 30 sensitivity scenarios. The shaded areas show the confidence intervals for the fitted lines. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

northern latitudes (Mearns et al., 2007). Latitudinal change in the role of precipitation increase in offsetting the effect of warming on cold regions hydrology implies that, even though northern latitudes will warm more (Graversen et al., 2008), they may also have more precipitation. Therefore, the precipitation increase may offset the impact of warming on snow and hydrological regimes in northern basins. It is also expected that the response of hydrological processes in different latitudes to the same climatic change will differ.

Although the snow regime in Marmot Creek (Fig. 6) is as sensitive as in Wolf Creek (Fig. 5) to warming and a decrease in precipitation, its runoff regime is less sensitive than that in Wolf Creek, partly due to a seasonal maxima of precipitation in spring, during the seasonal snowmelt, in Marmot Creek (Figs. 8 and 9a and b). These results are consistent with findings and projections for other mountain areas (Sultana and Choi, 2018; Roche et al., 2018; Jennings and Molotch, 2019; López-Moreno et al., 2020). The relatively lower sensitivity of the forest zone peak snow in Wolf Creek (Fig. 5e) is because of the increased unloading and, hence, reduced sublimation of intercepted snow at warmer air temperatures that counteracts the reduced snowfall (Pomeroy et al., 2015). The higher resiliency of the Marmot Creek snowpack is due to smaller changes at high elevations and in the blowing snow sink zone of the treeline forest, in which a deep snowpack is deposited that remains until mid-summer (MacDonald et al., 2010; Harder et al., 2015; Rasouli et al., 2019a). The high elevation and high latitude basins are more resilient to warming because their temperatures are currently well below that required to shift the precipitation phase (Bavay et al., 2013;

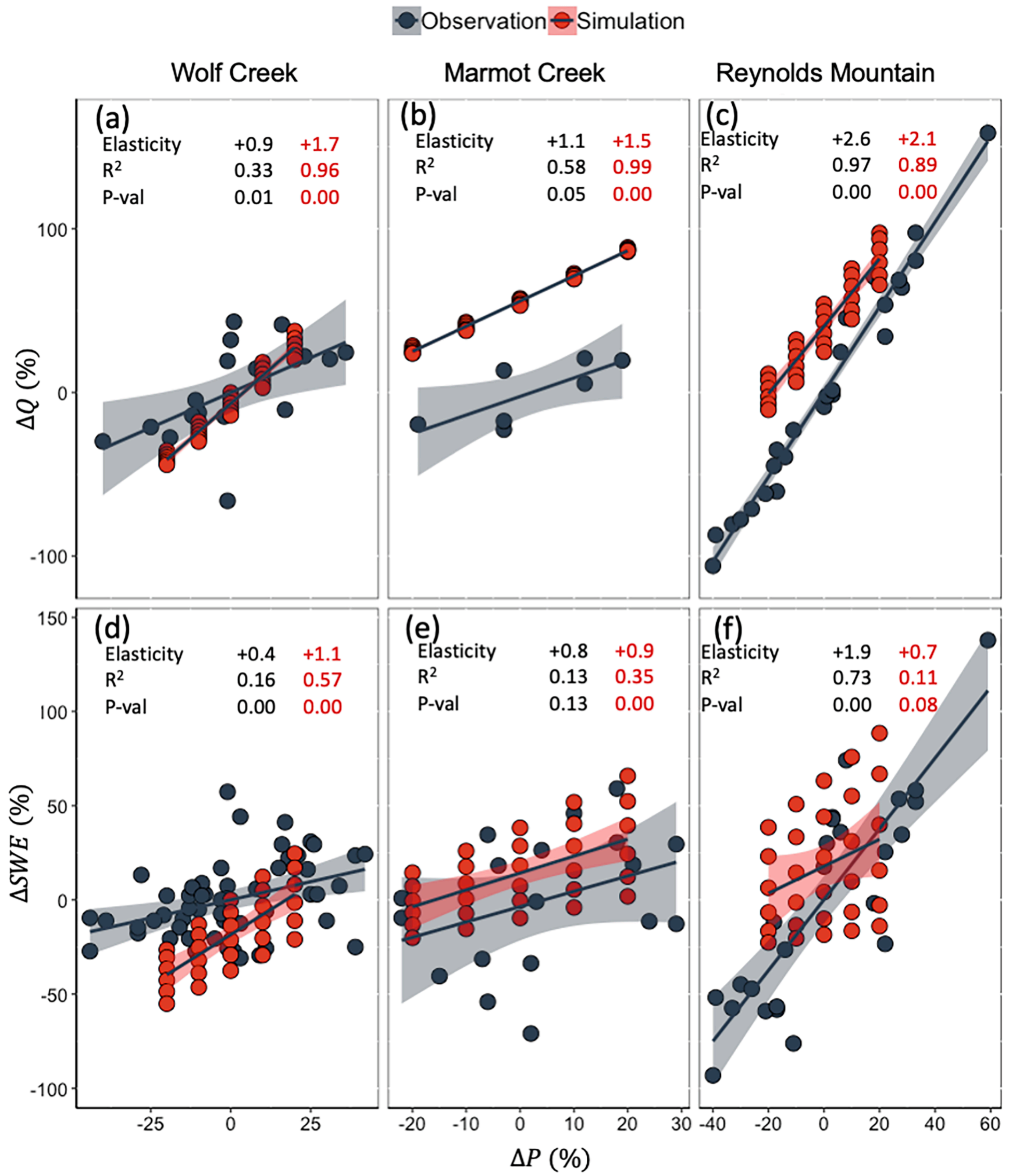

Fig. 10. Observed and simulated precipitation elasticity of annual runoff ($\frac{\Delta Q}{\Delta P}$) and peak snow water equivalent ($\frac{\Delta SWE}{\Delta P}$) in the three basins. The black circles show the observations and the red circles represents 30 sensitivity scenarios. The shaded areas show the confidence intervals for the fitted lines. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Jennings and Molotch, 2019). The snowpack lasts longer on the ground at high elevations in Marmot Creek, which moderates the impact of snow loss at low elevations on runoff (Rasouli et al., 2019a; López-Moreno et al., 2020). A high elevation band with air temperatures similar to that in low elevations in Wolf Creek and a rainy environment in the spring and early summer peak runoff period (Pomeroy et al., 2016) explain why the drop in peak snow accumulation (Fig. 9e) is not reflected by a proportional drop in annual runoff (Fig. 9b) in Marmot Creek. This highlights the role of spatial redistribution of snow on heterogeneous hydrological responses at different elevations in Marmot Creek. The snow and runoff regimes are the most sensitive to warming in Reynolds Mountain because of the (i) higher annual mean air temperature, (ii) near-freezing air temperatures in winter, and (iii) fewer number of days with freezing temperatures (120 days a year, Rasouli et al., 2019a) compared with the other basins. Rasouli et al. (2019a) found that under a moderate warming and increased precipitation of 7% and 2% in Marmot Creek and Reynolds Mountain, respectively, the annual runoff remained unchanged due to the offsetting effect of increased precipitation on increased evapotranspiration and offsetting effect of decreased sublimation on reduced snowfall (Rasouli et al., 2019a). Less sensitivity of annual runoff to warming relative to snowpacks suggests that warming mountain snowpacks can be decoupled from hydrological regimes (López-Moreno et al., 2020). Snowpack regimes in Reynolds Mountain are more sensitive to warming than to changes in precipitation, similar to the Cascade Mountains of Oregon, USA (Sproles et al., 2013).

Table 4
Observed (present climate) and simulated (future climate) precipitation elasticities of peak snow water equivalent and annual runoff and their differences in the three mountain basins. The elasticity is defined as percent change in runoff or peak snowpack per one percent change in precipitation.

	unit	Wolf Creek	Marmot Creek	Reynolds Mountain
Observed				
ΔPeak SWE	[mm (%)]	+0.5 (+0.4)	+2 (+0.8)	+7 (+1.9)
ΔAnnual runoff	[mm (%)]	+1 (+0.9)	+5 (+1.1)	+10 (+2.6)
Simulated				
ΔPeak SWE	[mm (%)]	+1.0 (+1.1)	+2 (+0.9)	+3 (+0.7)
ΔAnnual runoff	[mm (%)]	+3 (+1.7)	+6 (+1.5)	+8 (+2.1)
Modelling uncertainty (observed minus simulated)				
ΔPeak SWE	[mm (%)]	+0.5 (+0.6)	+0.1 (+0.1)	+4 (+1.2)
ΔAnnual runoff	[mm (%)]	+2 (+0.8)	+1 (+0.4)	+2 (+0.5)

The air temperature elasticity of snowpack was found to be −8% °C^{-1} in Wolf Creek, −10% °C^{-1} in Marmot Creek, and −17% °C^{-1} in Reynolds Mountain (Table 2). In Wolf Creek, these results are similar to reductions per degree warming observed in the Svalbard Archipelago (79° N, López-Moreno et al., 2016). Snowpack loss in Marmot Creek is in the range of 11–20% °C^{-1} reduction reported for the Pyrenees (López-Moreno et al., 2013; López-Moreno et al., 2014) and comparable to a 15% °C^{-1} reduction reported for the Swiss Alps (Beniston et al., 2003). Snow loss in Reynolds Mountain is similar to a 20% °C^{-1} reduction reported for the Washington Cascades (Casola et al., 2009). The results here are consistent with other basins with similar climates and that climate change affects snowpack in mountain basins across the globe with large reductions at mid-latitudes and relatively small reductions at high latitudes (Roche et al., 2018; Sultana and Choi, 2018).

Under the severe warming of 5 °C and 20% increased precipitation, annual runoff increases in all three basins (Table 3; Fig. 10a–c) because of the increasing importance of rain in warmer climates, suggesting that precipitation increase has a primary role in changing annual total runoff, and there is a large shift in the runoff mechanism from being snowmelt-driven to rainfall-driven. This shift may result in reduced runoff (Berghuijs et al., 2014) if precipitation does not increase (Table 3). It might also alter forest vegetation over time by making it more prone to wildfire and disease.

The air temperature and precipitation elasticities of annual runoff and annual peak snowpack increase from the northern basin to the southern basin (Figs. 9 and 10). The high slope of $\frac{\Delta Q}{\Delta T}$ and variability around the fitted line for Reynolds Mountain indicate a strong sensitivity of annual runoff to warming and precipitation change (Fig. 9c), suggesting that changes in annual runoff are attributable to both warming and precipitation change in the southern basin. The changes in annual runoff are not attributable to warming in the colder Wolf Creek and Marmot Creek (Fig. 9a and 10b) as demonstrated by the minimal variability around the fitted $\frac{\Delta Q}{\Delta P}$ lines (Fig. 10a and 10b) and relatively mild slope of $\frac{\Delta Q}{\Delta T}$ (Fig. 9a and 10b). The high slope of $\frac{\Delta SWE}{\Delta T}$ (Fig. 9d–f), the low slope of $\frac{\Delta SWE}{\Delta P}$(Fig. 10d–f) and high variability around the fitted lines show that the changes in annual peak snowpack are primarily attributable to warming and, secondarily, to precipitation change in all three basins, and particularly in Reynolds Mountain.

5.1. Offsetting temperature increases

The impact of warming of 1 °C on SWE values over the winter and spring seasons can be offset by a precipitation increase of 20% for almost

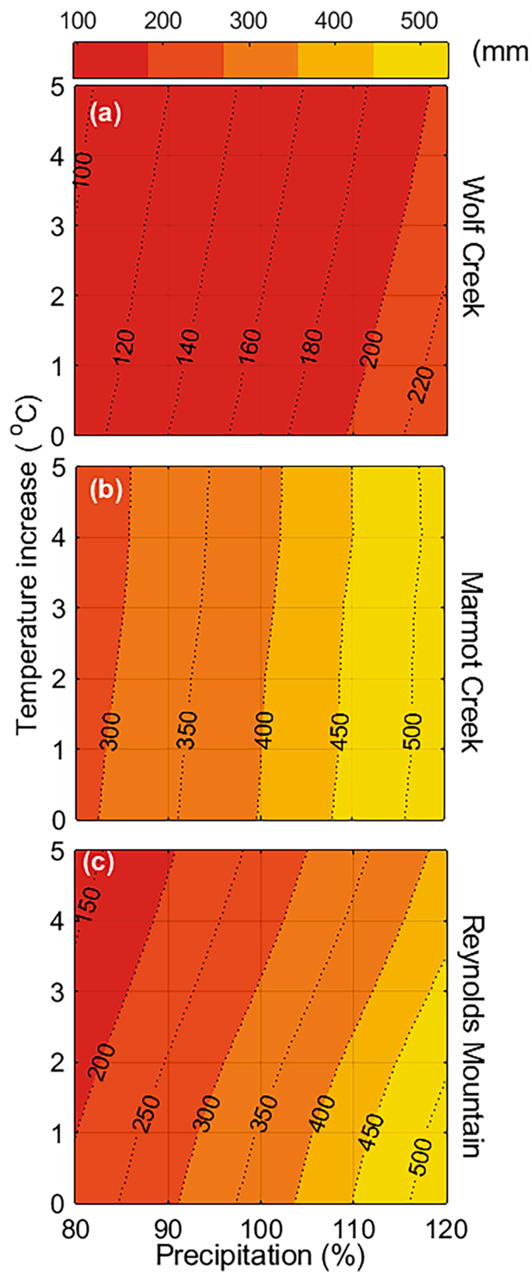

Fig. 11. Sensitivity of mean annual runoff to increases in air temperature and changes in precipitation in (a) Wolf Creek at the Alaska Highway, (b) outlet of Marmot Creek, and (c) outlet of Reynolds Mountain.

all SWE values in all snow regimes in Reynolds Mountain; however, warming of 2 °C or more cannot be offset 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; this is due to suppression of blowing snow redistribution processes by warming. In Wolf Creek (≈ 61° N), not only more warming but also an increase in precipitation is expected (Graversen et al., 2008), which indicates that precipitation increases could partly offset the effect of warming on cold regions hydrology. Despite the uniformity of high mountain climates and similar response per degree warming, the implication of these results is that mountain snow regime responses to climate change can differ substantially (López-Moreno et al., 2020), as noted for the three basins across North America studied here; therefore, regional analysis is required. The large difference between snowpack response in Reynolds Mountain and

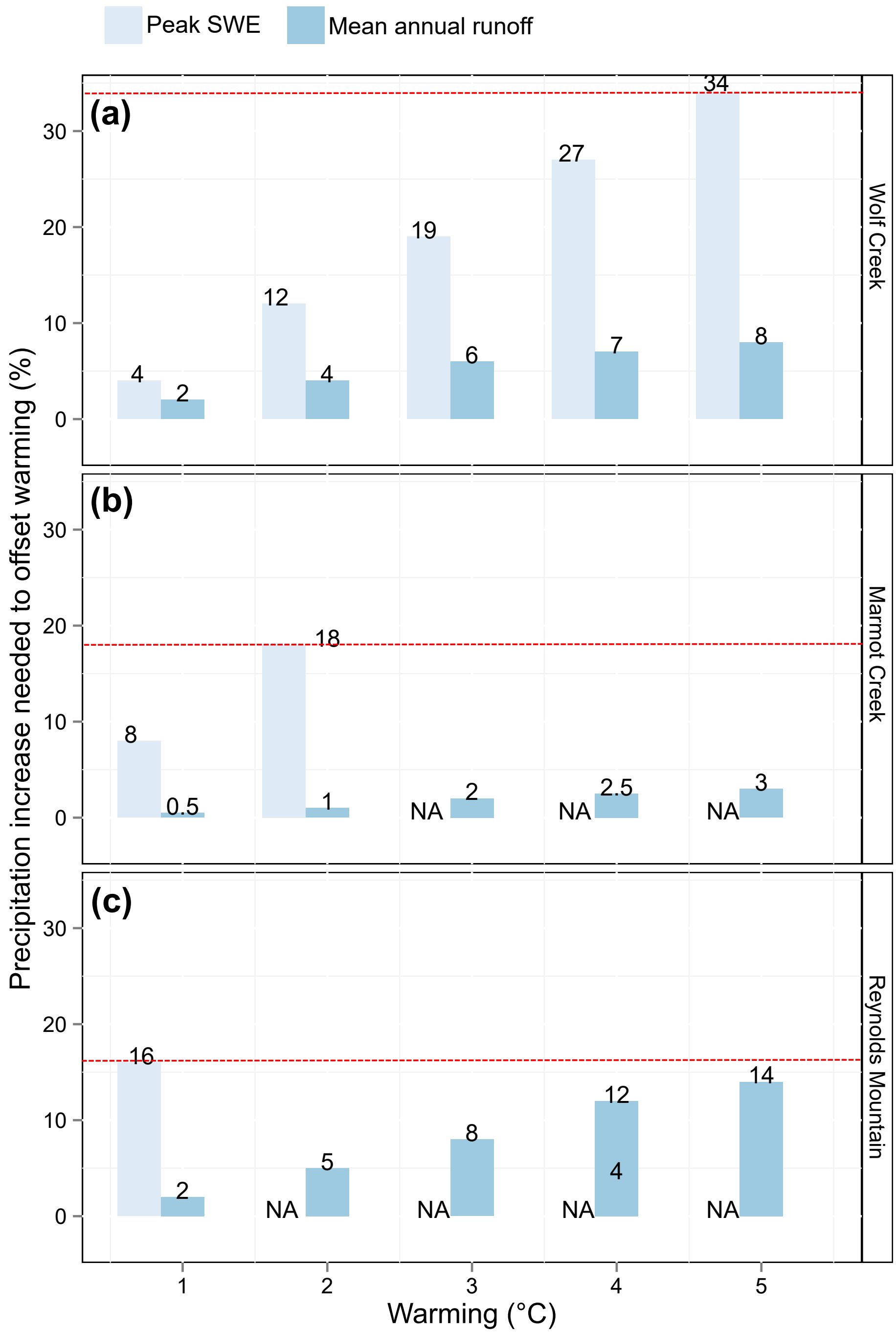

Fig. 12. The percentage of precipitation change required to offset the effect of warming by 1–5 °C on peak snow water equivalent (SWE) or mean annual runoff. NA is assigned to cases where the amount of precipitation required to offset the air temperature increase is greater than the increased precipitation projected by RCP and SSP scenarios and NARCCAP simulations (red horizontal lines) forced with the SRES A2 scenario (business-as-usual scenario) for the period 2041–2070. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Table 5
Annual precipitation (P), evapotranspiration (ET), and their ratio (ET/P) under present and future climates in the three basins.

Basin	Climate	P	ET	ET/P
Wolf Creek	Current	380	130	0.34
	Future ($\Delta T = 2.6$ °C, $\Delta P = +16\%$)	444	157	0.35
Marmot Creek	Current	1011	392	0.39
	Future ($\Delta T = 2.2$ °C, $\Delta P = +6.6\%$)	1062	447	0.42
Reynolds Mountain	Current	845	427	0.51
	Future ($\Delta T = 2.4$ °C, $\Delta P = +2.3\%$)	897	504	0.56

Wolf Creek implies that warming in cool climates affects 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 (Poulin et al., 2011; Whitfield and Shook, 2020) and a shift from March to January in Reynolds Mountain, April to February in Marmot Creek and less than a month in Wolf Creek for the timing of peak snow accumulation (Table 3). The impacts of warming on snowpacks can be partly offset by a precipitation increase in the cold Wolf Creek and Marmot Creek climates but not in the cool Reynolds Mountain climate. The snow season is expected to shorten by about two months in the subarctic Wolf Creek (from 9 to 7 months), three and half months in the cold Marmot Creek (from 6 to 4 months), and five months in the cool Reynolds Mountain basin (from 6 to 1 month) with concomitant warming and a decline in precipitation (Table 3). This implies that, under climate warming, the response of the snow hydrology to a precipitation increase changes with latitude, from very little in Reynolds Mountain to very large in Wolf Creek. Snow hydrology is more sensitive to warming and precipitation phase change in the southern basin and is more resilient in Marmot Creek.

The sensitivity of annual runoff to warming in the three basins increases from north to south, and it is in the range of –1% $^{\circ}C^{-1}$ to –6% $^{\circ}C^{-1}$ (Table 2), while projected precipitation elasticity is in the range of 1.5 to 2.1 (Table 4), indicating that the runoff change is primarily attributed to precipitation change and, secondarily, to warming. Hoerling et al. (2019) separated the effects of temperature and precipitation changes on Upper Colorado River streamflow and found that two-thirds of the climate change signal effect on declining streamflow over the past century was attributable to precipitation change. Warming had the secondary effect, and runoff showed a modest sensitivity of –2.5% $^{\circ}C^{-1}$. In contrast to annual runoff, the projected temperature elasticity of annual peak snowpack is in the range of –8% $^{\circ}C^{-1}$ to –17% $^{\circ}C^{-1}$, lesser in Wolf Creek and higher in Reynolds Mountain (Table 2), while precipitation elasticity is in the range of 0.7 to 1.1 (Table 4), which indicates that changes in the annual peak snowpack is primarily attributed to warming and, secondarily, to precipitation change.

6. Conclusions

Annual perturbations of observed hourly air temperature and precipitation were used to drive physically based cold regions hydrological models of the elasticity and sensitivity of snow and runoff regimes in well-instrumented mountain research basins that span the northern North American Cordillera. Peak snowpack is sensitive to both warming and precipitation change in Wolf Creek in the subarctic Yukon and is more sensitive to temperature in Reynolds Mountain in temperate Idaho. Peak snowpack is most sensitive to warming in the sheltered site in Reynolds Mountain and to both warming and precipitation change in the blowing snow sink regime in Reynolds Mountain, at lower elevations in Marmot Creek, and the shrub tundra zone in Wolf Creek. Peak snowpack timing is more sensitive to warming in Marmot Creek and Reynolds Mountain but, in Wolf Creek, precipitation more strongly affects the timing of peak SWE, as temperatures remain largely below zero. Snow season start, end, and duration were found to be sensitive to warming in temperate Idaho and subarctic Yukon and to both warming and precipitation change in the continental Canadian Rockies (Marmot Creek).

The scenario with severe climate warming and decreased precipitation in all three basins caused dramatic declines in SWE, a shortened snow-covered period, and decreases in annual runoff. The decreases in snowpack depth and advances in the timing of peak snowpack are weakly reflected in changes to runoff regime in each basin. The large changes in snowpack found here do not result in similar magnitude changes in annual runoff.

If precipitation decreases with warming, the impacts on snowpacks are amplified, with major implications for ecology, winter transportation, and hydrology. Smaller snowpacks and warmer weather would cause an increase in the snow-free period, which also would lengthen the evapotranspiration season, increasing the annual evapotranspiration loss. The importance of rainfall–runoff mechanisms in these basins increases while snowmelt decreases. Under warmer and drier climatic conditions, annual runoff decreases.

The projected precipitation elasticity of annual runoff increases from 1.7 in Wolf Creek to 2.1 in Reynolds Mountain, and precipitation elasticity of snowpack increases from 0.7 in Reynolds Mountain to 1.1 in Wolf Creek, indicating that increased precipitation would increase annual runoff more in the wet basin in the south than in the dry basin in the north. Increased precipitation resulted in more increased snowpack in the cold basin in the north than in the cool basin in the south for the same degree of warming. The magnitude of the air temperature elasticity of annual runoff increases from –0.8% $^{\circ}C^{-1}$ in Marmot Creek to –5.8% $^{\circ}C^{-1}$ in Reynolds Mountain, and the magnitude of air temperature elasticity of snowpack increases from –8% $^{\circ}C^{-1}$ in Wolf Creek to –17% $^{\circ}C^{-1}$ in Reynolds Mountain. This shows that increased warming will decrease annual runoff and snowpack more in the wet basin in the south than the dry basin in the north, under the same precipitation conditions. The observed and simulated precipitation and air temperature elasticities of peak snowpack and runoff showed minimal discrepancies, except for the air temperature elasticity of peak SWE in Marmot Creek and of runoff in Reynolds Mountain. The changes in annual runoff in the colder basins and elevations, however, are more attributable to precipitation change, while the changes in peak snowpack are more attributable to warming. In basins with cool climates and higher elevations, changes in annual runoff are attributable to both precipitation change and warming.

Increased precipitation, expected from many climate projections, can partially offset the effect of warming on snowpack and annual runoff. The role of precipitation as a compensator for the impact of warming on peak snowpack and annual runoff is most effective in the colder high elevations and high latitudes, and its effectiveness is reduced where snow regimes currently depend on blowing snow transport and redistribution, which is very sensitive to temperature. With increased precipitation, high elevation and high latitude basin snow and hydrological regimes can be resilient to warming. At lower elevations and latitudes, however, the impact of warming cannot be offset by any projected maximum precipitation increases in future climates. The coupling of snow regimes to streamflow hydrology will remain strong in northern Canada but weaker in the mountains of Idaho and Alberta as the climate warms in a manner consistent with the global decoupling noted by López-Moreno et al. (2020).

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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