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Decoupling of warming mountain snowpacks from hydrological regimes

J I López-Moreno[1] , J W Pomeroy[2], E Alonso-González[1], E Morán-Tejeda[3] and J Revuelto[1]

[1] Pyrenean Institute of Ecology, CSIC, Zaragoza 50059, Spain
[2] Centre for Hydrology, University of Saskatchewan, Saskatoon, SK, S7N 1K2, Canada
[3] Department of Geography, University of the Balearic Islands, Palma 07122, Spain

E-mail: nlopez@ipe.csic.es

Abstract

Climate warming will reduce the duration of mountain snowpacks and spring runoff, impacting the timing, volume, reliability, and sources of water supplies to mountain headwaters of rivers that support a large proportion of humanity. It is often assumed that snow hydrology will change in proportion to climate warming, but this oversimplifies the complex non-linear physical processes that drive precipitation phases and snowmelt. In this study, snow hydrology predictions made using a physical process snow hydrology model for 44 mountains areas worldwide enabled analysis of how snow and hydrological regimes will respond and interact under climate warming. The results show a generalized decoupling of mountain river hydrology from headwater snowpack regimes. Consequently, most river hydrological regimes shifted from reflecting the seasonal snowmelt freshet to responding rapidly to winter and spring precipitation. Similar to that already observed in particular regions, this study confirms that the worldwide decline in snow accumulation and snow cover duration with climate warming is substantial and spatially variable, yet highly predictable from air temperature and humidity data. Hydrological regimes showed less sensitivity, and less variability in their sensitivity to warming than did snowpack regimes. The sensitivity of the snowpack to warming provides crucial information for estimating shifts in the timing and contribution of snowmelt to runoff. However, no link was found between the magnitude of changes in the snowpack and changes in annual runoff.

1. Introduction

Temperature increase is the clearest consequence of climate change caused by anthropogenic emissions of greenhouse gases. In the 20th century, mountains warmed more than lowlands because of enhanced longwave emissions from the atmosphere (Vuille *et al* 2003) and albedo feedback (Pepin *et al* 2015). Based on available future projections, this amplified warming is expected to continue at high elevations (Beniston *et al* 2018). Temperature increases have already reduced the water storage capacity provided by the seasonal snowpack in most mountain areas (Harpold *et al* 2012, Ning and Bradley 2015, Bormann *et al* 2018, Mcgowan *et al* 2018), and the IPCC has consistently projected a continuation or intensification of this trend and its impacts on the

subsequent release of meltwater (Musselman *et al* 2017, Simpkins 2018). A warming climate is expected to reduce spring runoff (Barnett *et al* 2005) and will probably also affect the annual water balance (Barnhart *et al* 2016), and there is an assumption that the magnitude of these processes will be linearly related to the intensity of warming (Simpkins 2018). In contrast, recent research has shown marked divergence in the sensitivity of seasonal snow cover regimes to higher temperatures, driven by climatic differences that control snowpack accumulation and ablation regimes, and associated energy budgets (López-Moreno *et al* 2017). The snow hydrology in basins close to the 0 °C isotherm during the cold season is particularly sensitive for two reasons: (i) a sharp increase in the rainfall ratio and a decrease in snowfall as ice-bulb temperatures increase (Harder

and Pomeroy 2014, Sospedra-Alfonso and Merryfield 2017); and (ii) the occurrence of snowpacks closer to 0 °C (isothermal conditions), which will melt faster than cold snowpacks with any increase in incoming energy (Painter *et al* 2010, López-Moreno *et al* 2017). Atmospheric humidity has also been found to control the sensitivity of the snowpack to a warming climate (Harpold and Brooks 2018). A dryer atmosphere tends to favor warmer threshold temperatures for the rainfall–snowfall phase change (Harder and Pomeroy 2013, Harpold *et al* 2017, Jennings *et al* 2018). Low atmospheric humidity levels also reduce emissivity compared with high moisture conditions; under such conditions the snowpack may remain cool even at positive temperatures, and increased latent fluxes may cancel out the sensible fluxes that are mostly driven by temperature (López-Moreno *et al* 2017, Harder *et al* 2017, Harpold and Brooks 2018). Recent studies have reported reduced blowing and intercepted snow redistribution and sublimation under warmer climate conditions (Rasouli *et al* 2019), introducing even more complexity to the response of snow to a warming climate. There has not yet been a global assessment of how snowmelt energy processes mediate the response of snow accumulation and melting to climate change. This is because of the very limited availability of *in-situ* observations, and the need to run physically detailed hydrological process models at very high temporal and spatial resolution to enable sound conclusions to be reached (Pomeroy *et al* 2015).

Previous studies have predicted similar hydrological consequences from climate warming for various snow-dominated basins (Adam *et al* 2009). These include an increase in the role of rain relative to snowmelt in the hydrological processes; an earlier spring freshet, which is often quantified by the temporal advance of the center of mass of the hydrograph; and a loss of storage capacity of the snowpack, with river regimes progressively resembling the seasonal distribution of precipitation (Musselman *et al* 2017). Several recent studies on mountain hydrology have also reported that loss of the water storage capacity of the snowpack affects the seasonal distribution of water releases, but also results in less annual runoff because of the lower efficiency of runoff production under slower melt rates associated with a warming climate (Musselman *et al* 2017). These findings are consistent with the hypothesis that rapid snowmelt enhances runoff (Berghuijs *et al* 2014, Barnhart *et al* 2016). However, annual runoff can also decline with warming as consequence of greater evapotranspiration in spring because of the earlier and longer snow-free period that permits plant transpiration (Rasouli *et al* 2019). This effect could be further enhanced by an increase in the vapor pressure deficit with warmer temperature (Bosson *et al* 2012). Nevertheless, there is a lack of direct evidence that the hydrological impacts of snowpack sensitivity to climate warming

are linearly related to the timing and total availability of water resources in snow-dominated basins and downstream areas (Huntington and Niswonger 2012, Huning and Aghakouchak 2018). The complicating effects of elevation and many other basin characteristics (e.g. geology, topography, soils, and landcover) on hydrological processes and varying pathways of snowmelt water makes it difficult to perform in-depth comparisons of the sensitivity of hydrological processes to increased air temperature amongst different mountain regions. In addition, the scarcity of observational data available to run physically based hydrological models on a wide range of mountain basins makes it difficult to validate conclusions at global scales.

In this study, bias-corrected reanalysis data (Weedon *et al* 2014) were applied to a mountain basin model, created using the Cold Regions Hydrological Modelling platform (CRHM) (Pomeroy *et al* 2007, Rasouli *et al* 2015), to simulate snowpack and streamflow regimes over 33 years in an idealized, 'virtual' basin for 44 mountain regions worldwide. The virtual basin had a typical hypsometry and shape for a high mountain basin and applying the physically based snow hydrology model to a virtual basin permitted calculating the sub-basin variability of snow energetics and dynamical processes, including energy budget snowmelt and snow redistribution and runoff generation in a standard, comparable manner. This set of virtual basins encompassed most of the climate conditions found in snow-dominated mountain headwaters (*SI appendix,* table S1). Each simulation was run for each virtual basin for a control period (1982–2014), and involved a progressive 1 °C increase in temperature to +5 °C. A set of indices were developed from the snowpack and runoff series to: (i) quantify their sensitivity to climate warming and assess the role of temperature and humidity in the observed differences amongst mountains; (ii) to assess if the changes in river regimes resulting from a decline in the snowpacks was proportional to the sensitivity of the snowpack to a warmer climate; and (iii) to assess whether the reduction in snowpacks caused by global warming will result in reduced annual streamflow generation.

2. Data and methods

2.1. Virtual basin approach
Virtual experiments have been used to run hydrological simulations in previous studies (Weiler and Mcdonnell 2004, Bhj *et al* 2011, Armstrong *et al* 2015). These have compared the impact of climate variability and change on the hydrology in various regions, where the effects of varying physiography and land cover on runoff generation have been removed. This was a prerequisite to investigating links between snow and hydrological sensitivity to climate warming in the present study. The virtual

basin approach used relied on analysis of a 'typical' small basin of 5.25 km^2 having 1000 m vertical gradient, where vegetation, soil, and groundwater storage had very limited hydrological effects; this simplification ensured that climate and snow dynamics were the primary factors explaining runoff generation. The virtual basin included seven hydrological response units (HRUs): (i) a summit facing to the west; (ii) a high plateau facing to the west; (iii) and (iv) north and south high slopes; (v) and (vi) north and south low slopes; and (vii) the bottom of the basin facing to the west. The soil depth was set to zero at the summit and increased progressively to 50 cm depth at the outlet of the basin. Short meadow grass was the only vegetation included below the high plateau, and there was no vegetation above (*SI appendix,* figure S1, and table S1 (available online at https://stacks.iop.org/ERL/15/114006/mmedia)).
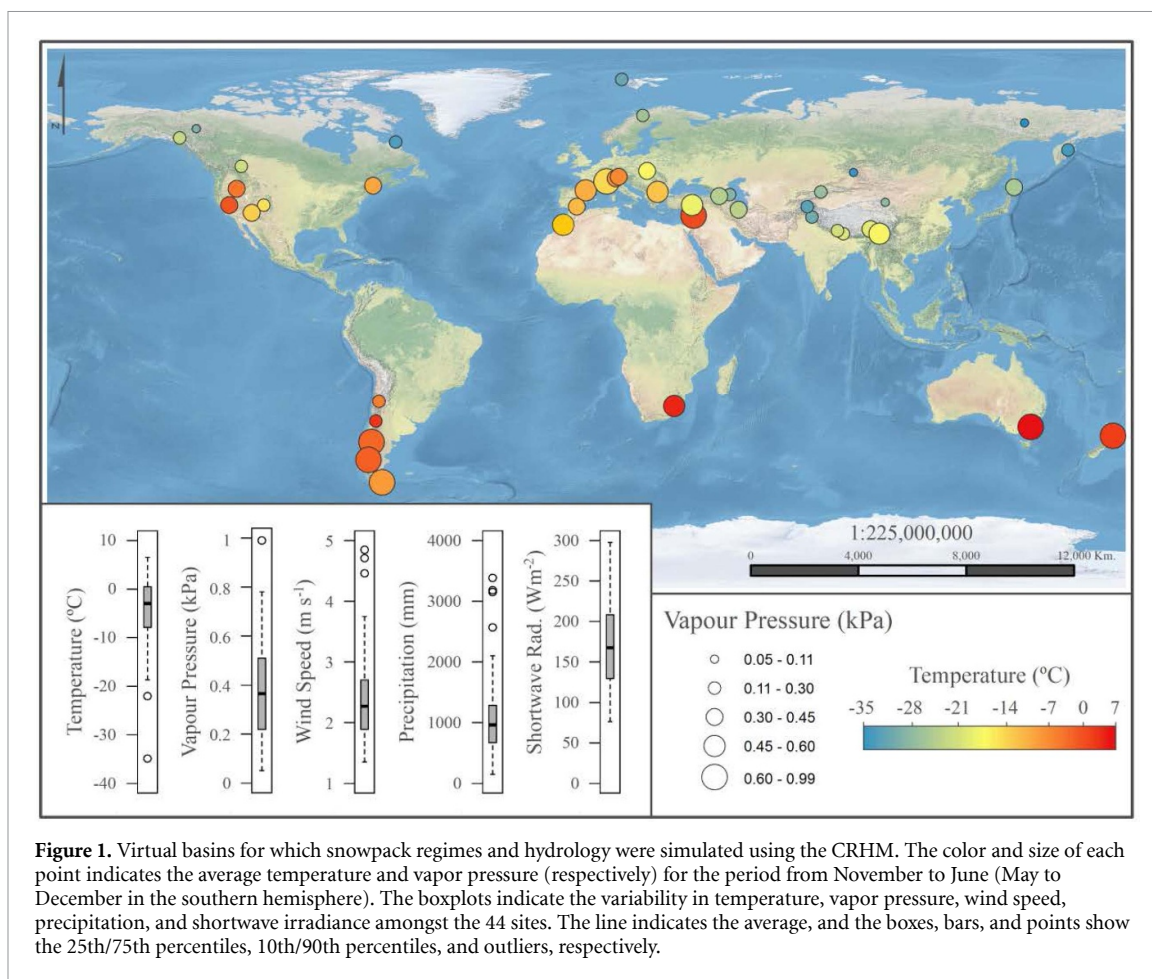
2.2. Cold regions hydrological model and input data

The flexible, modular CRHM platform was used to link various modules representing the main snow and hydrological processes that are characteristic of alpine regions. CRHM considers the full array of physical processes involved in snow redistribution, snowmelt dynamics, and runoff generation, including blowing snow transport, energy balance snowmelt, sublimation, infiltration to frozen soils, subsurface hydrology, evapotranspiration, and flow routing. The model downscales gridded air temperature, humidity, wind speed, radiation, and precipitation to HRUs within the virtual basins. The HRU correspond to slope/aspect and elevation zones within the basin, and mass and energy balance calculations are conducted at both the HRU and virtual basin scales. Use of the CRHM platform enabled incorporation of these physical factors without the need for calibration of parameters; instead these were set from hydrological process studies in mountain basins (Pomeroy *et al* 2012, Fang *et al* 2013). A flowchart for the modules used in the study is shown in *SI appendix,* figure S2. Previous studies have described the application of this model (Rasouli *et al* 2014, Revuelto *et al* 2014), and its performance has been shown to be satisfactory when tested against other models in the SNOWMIP initiatives (Rutter *et al* 2009; Essery and Pomeroy 2004), again without calibration.

Meteorological inputs were elaborated in the WATCH project (www.eu-watch.org) and correspond to the WFDEI dataset (Weedon *et al* 2014), which comprises bias-corrected ERA-40 reanalysis data that encompasses temperature, specific humidity, surface pressure, wind speed, incoming shortwave radiation, and precipitation at three-hour intervals for the period 1979–2012, at 0.5° spatial resolution. Data for the pixel containing the coordinates of the target mountain areas were used to drive the CRHM virtual basin model, which rescaled the radiation,

precipitation, wind speed, temperature, and humidity from the elevation of the WATCH data centroid to the elevation, slope, and aspect of each HRU in the virtual basin. However, the initial 0.5° resolution of the driving meteorological data is likely to underestimate extreme precipitation and lack sufficient information on mountain topographic effects on precipitation, temperature, humidity, radiation and wind fields. This is why the downscaling within CRHM is very important to applying these fields in mountain terrain.

Many of the selected mountain areas (*SI appendix,* table S2) correspond to experimental basins contributing to the INARCH project (the table from www.usask.ca/inarch/is indicated by an asterisk in table S2); the snowpack, micrometeorology, and hydrological processes involved in these basins have been studied extensively. The 44 basins selected were considered to be a good compromise including differing climates, provided a sample size sufficient to enable simple statistical analysis, and a number of case studies where each simulation could be examined to ensure its consistency with the known snow hydrology of the region. The elevation assigned to each basin was subjectively defined after several runs of the snowpack at various elevations at each site. The aim was to have a well-developed seasonal snowpack during the control period, avoiding HRUs where the snowpack did not melt from one year to the next (which would result in the formation of glaciers). Given the deployment of the model to virtual basins, it was not possible to conduct a quantitative evaluation of the CRHM simulations using field observations because the virtual basin does not correspond to any specific gauged basin. It should be noted that there are inherent and assumed uncertainties in the reanalysis data (Weedon *et al* 2014). However, mountain models developed using CRHM and similar to what is applied here have already been developed without resorting to parameter calibration from streamflow and have been successfully tested in many mountain regions of the world such as the US and Canadian Rocky Mountains, Yukon Territory, Patagonia, Spain, Chile, Morocco, California, Germany, Svalbard and high mountain Asia (Fang and Pomeroy 2016, Rasouli *et al* 2014; López-Moreno *et al* 2013, 2016, 2017, Weber *et al* 2016; Zhou *et al* 2014). This lends great confidence that the model, informed by these other CRHM applications, can be applied in many mountain regions for the comparative and diagnostic purposes of this study. In addition, the aim of the simulations was not to reproduce the exact conditions at each mountain site, which cannot be guaranteed using reanalysis data, but to ensure that coherent inputs represented the climates of major snow-dominated basins worldwide, so that the outputs could be used to compare and contrast the snow hydrology under differing climates of the world and the effects of climate warming. Even considering

Figure 1. Virtual basins for which snowpack regimes and hydrology were simulated using the CRHM. The color and size of each point indicates the average temperature and vapor pressure (respectively) for the period from November to June (May to December in the southern hemisphere). The boxplots indicate the variability in temperature, vapor pressure, wind speed, precipitation, and shortwave irradiance amongst the 44 sites. The line indicates the average, and the boxes, bars, and points show the 25th/75th percentiles, 10th/90th percentiles, and outliers, respectively.
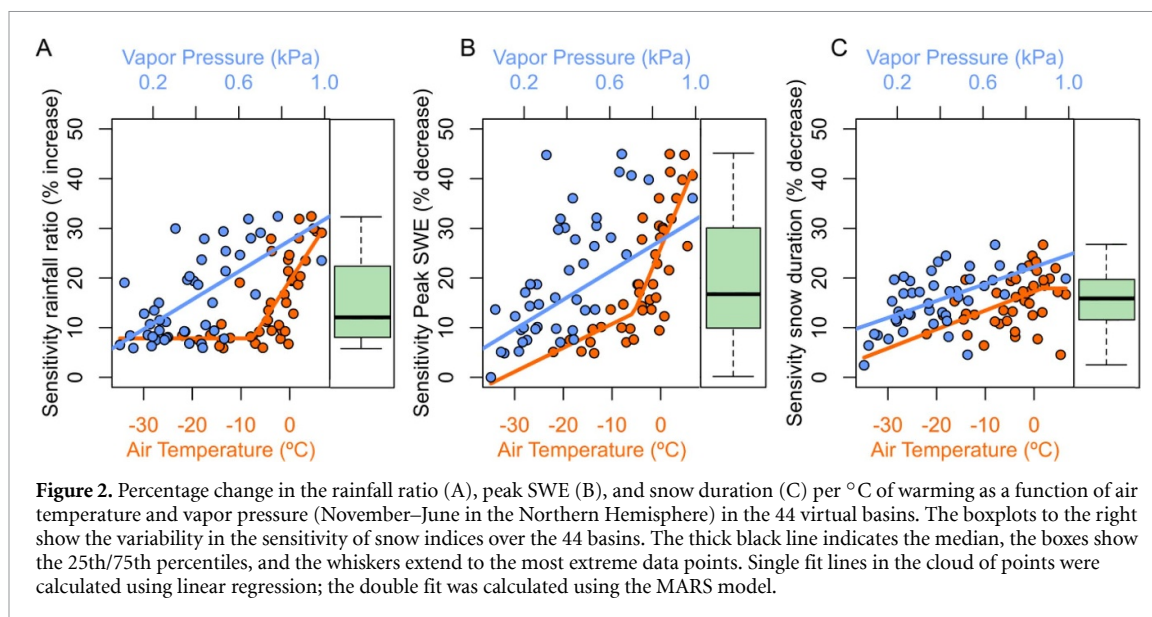
uncertainties contained in WATCH dataset, figure 1 and table S2 show that temperature, precipitation and air humidity show coherent values according the latitude, elevation and continentality of each analyzed mountain area.

2.3. Selected indices and statistical analysis

Several indices were developed characterize the snow and hydrological characteristics of each basin from the simulated series (control) and warmer conditions (*SI appendix,* table S3). The snowpack indices comprised the rainfall ratio, the peak snow water equivalent (SWE), and the duration of the snowpack. The rainfall ratio is the portion of total precipitation falling in a liquid phase. The peak SWE is the long-term average of the maximum annual value of the simulated SWE. The duration of the snowpack was calculated as the long-term average number of days annually having a SWE >5 mm. The runoff series indices comprised the snowmelt ratio, the snow damming (SDam), and D50. The snowmelt ratio is the percentage of annual runoff derived from snowmelt. The SDam index is a measure of the storage of precipitation in the snowpack during the cold season, and so reflects the extent to which the snowpack acted as a natural water storage reservoir. This index was quantified as the difference between the accumulated fraction of precipitation and the accumulated

fraction of runoff. It reflects the increase in synchronicity between the seasonal distribution of precipitation and runoff as climate warms and becomes dominated by rainfall-runoff processes (see example in *SI appendix,* figure S3). D50 identifies the center of mass of the hydrograph (the Julian day when cumulative annual runoff reaches 50%); this index has been widely used to quantify shifts in the occurrence of the spring freshet under changing snow conditions (Whitfield 2013).

First, relationships between the sensitivity of the snowpack, air temperature (Tair), and vapor pressure (Vp) were established using the coefficient of determination (the percentage of the explained variance calculated from the square of the Pearson's correlation coefficient once the Gaussian distribution of all variables in this study was confirmed by Kolmogorov-Smirnov test (Chakravarti *et al* 1967). Tair and Vp were selected as explanatory variables because they exhibited the largest partial correlations in an initial exploratory analysis that included other meteorological variables (results not shown). Furthermore, although these two variables generally show high correlation, they have been previously shown to have a complementary role in explaining the partition of the precipitation phase (Harder and Pomeroy 2013, Harpold *et al* 2017, Jennings *et al* 2018) and in snow ablation, which clearly affect the sensitivity of

Figure 2. Percentage change in the rainfall ratio (A), peak SWE (B), and snow duration (C) per °C of warming as a function of air temperature and vapor pressure (November–June in the Northern Hemisphere) in the 44 virtual basins. The boxplots to the right show the variability in the sensitivity of snow indices over the 44 basins. The thick black line indicates the median, the boxes show the 25th/75th percentiles, and the whiskers extend to the most extreme data points. Single fit lines in the cloud of points were calculated using linear regression; the double fit was calculated using the MARS model.

snowpacks to temperature (see figure 5 in Harpold and Brooks 2018). For snow duration, the joint effect of the predictor variables in explaining the total variance in sensitivity among the basins was analyzed using multiple linear regressions (MLRs). For the rainfall ratio and peak SWE, multivariate adaptive regression splines (MARS; Friedman 1991) were used. MARS automatically models nonlinearities and interactions amongst variables to account for observed changes in the response between sensitivity and air temperature.

Secondly, the coefficient of determination was also used to assess to which extent the sensitivity of snowpack duration and peak SWE drives the sensitivity of the hydrological indices and also annual runoff.

3. Results

3.1. Sensitivity of snowpack indices to climate warming

Figure 2 shows the variability in the sensitivity of the three snow indices (rainfall ratio, peak SWE, and snow cover duration) and their relationships to the average annual temperature and humidity (expressed as vapor pressure) for each virtual basin. figure S4 shows the geographic distribution of simulated snow sensitivities. The increase in the rainfall ratio varied from 5%–32% per °C of temperature increase (figure 2(a)). Sensitivity increased with increasing temperature and air humidity, and showed a positive and statistically significant linear correlation with vapor pressure ($r^2 = 0.42$) and air temperature ($r^2 = 0.45$). However, there was no significant response to temperature increase in the colder basins (those having November–June temperatures colder than −8 °C). In contrast, the response to temperature was very strong in warmer basins. Consequently,

the MARS model was applied using a double regression to fit temperature to changes in the rainfall ratio. The decrease in peak SWE (figure 2(b)) showed even greater variability (0.3%–45% per °C of warming), and exhibited a greater response to vapor pressure ($r^2 = 0.58$) and temperature ($r^2 = 0.72$) than did the rainfall ratio. There was also a clear break point at approximately −8 °C, separating the colder basins, having a gradual increase in sensitivity to climate warming, from the warmer basins, where the increase in sensitivity to climate warming was greater. The decrease in snow duration in the analyzed basins varied (figure 2(c)) from 2%–27% per °C of warming. There was also a significant positive correlation with humidity and temperature, but the coefficient values were lower ($r^2 = 0.31$ and 0.26, respectively). No threshold distinguishing the response of snow duration between colder and warmer basins was found.

Partial correlation analysis revealed that the effects of humidity and temperature on the sensitivity of the three snow indices among basins were independent, and not affected by covariance. The combination of these two independent variables (air temperature and vapor pressure) in predicting the sensitivity of the rainfall ratio and peak SWE using MARS explained 58% and 79% of the total variance, respectively, while the predictability of the sensitivity of snow cover duration in the various basins using MLR was lower (explained variance = 45%).

Maps of the analyzed basins showed relatively similar relationships of air temperature and humidity to the three indices, with the rainfall ratio and peak SWE having similar distributions (*SI appendix*, figures S4, and S5(a); $r^2 = 0.84$). The lowest sensitivities were found at high latitudes in the northern hemisphere and basins at high elevations in dry areas (Tibetan Plateau, central Himalayas, Anatolia, Middle East, and Northern Chile), while the highest

sensitivities were evident at medium latitude areas of both hemispheres (45°N–45°S), including Australia, New Zealand, central Chile, the Pyrenees, the Alps, and the most humid parts of the Himalayas.

The spatial distribution of the sensitivity of snow duration showed a more complex geographic distribution than for the other two indices. The magnitude of sensitivity was closely related to that of peak SWE for those basins showing low or medium sensitivity, but those basins in which the sensitivity of peak SWE was 30% showed a weaker relationship to snow duration (*SI appendix,* figure S5(b); $r^2 = 0.43$). Thus, several basins in the southern hemisphere, the most humid areas of the Himalayas, and mountains of South Africa showed a marked decrease in the sensitivity of snow duration compared with peak SWE or the rainfall ratio. Spatial associations between the sensitivity of the rainfall ratio and snow duration were weak (*SI appendix,* figure S5(c), $r^2 = 0.16$).

3.2. Relationship between hydrological and snowpack sensitivities

The model showed a decrease of 2%–13% per °C of warming in annual runoff derived from snowmelt (snowmelt ratio) in the catchments, and this decrease was highly and linearly related to the sensitivity of peak SWE (figure 3(a); $r^2 = 0.54$), and to a lesser but statistically significant extent with the sensitivity of snow duration (figure 3(d); $r^2 = 0.32$). Partial correlation analysis indicated that this relationship was not affected by differences in the rainfall ratio during the control period (1982–2014) among the analyzed basins.

SDam was almost stable (decrease <3% per °C) in only three of the analyzed basins (figures 3(b) and (e)), all of which are located in very dry and cold environments (figure S5). In the remainder, SDam would have decreased the storage capacity of the snowpack to varying degrees, but never exceeding 20% per °C of warming. There was a high and statistically significant positive correlation between decreasing SDam and the sensitivity of snow duration ($r^2 = 0.71$), and to a lower but statistically significant extent with the sensitivity of peak SWE ($r^2 = 0.36$). As for the snowmelt ratio, the rainfall ratio during the control period does not mediate between the sensitivity of SDam and the sensitivity of the snowpack.

The sensitivity of the center of mass of the hydrograph (D50 index) indicated an advance in timing as climate warmed (Morán-Tejeda *et al* 2014, Sanmiguel-Vallelado *et al* 2017). With the exception of a few basins where the sensitivity of snowpack was very low and the D50 remained almost unchanged, or was slightly delayed, most (>10%) mountain areas showed an advance in the center of mass to dates more than one month earlier per °C of warming (figure 3(b)). The sensitivity of D50 was statistically positively correlated with the sensitivity of snow duration ($r^2 = 0.46$; figure 3(c)), but the correlation
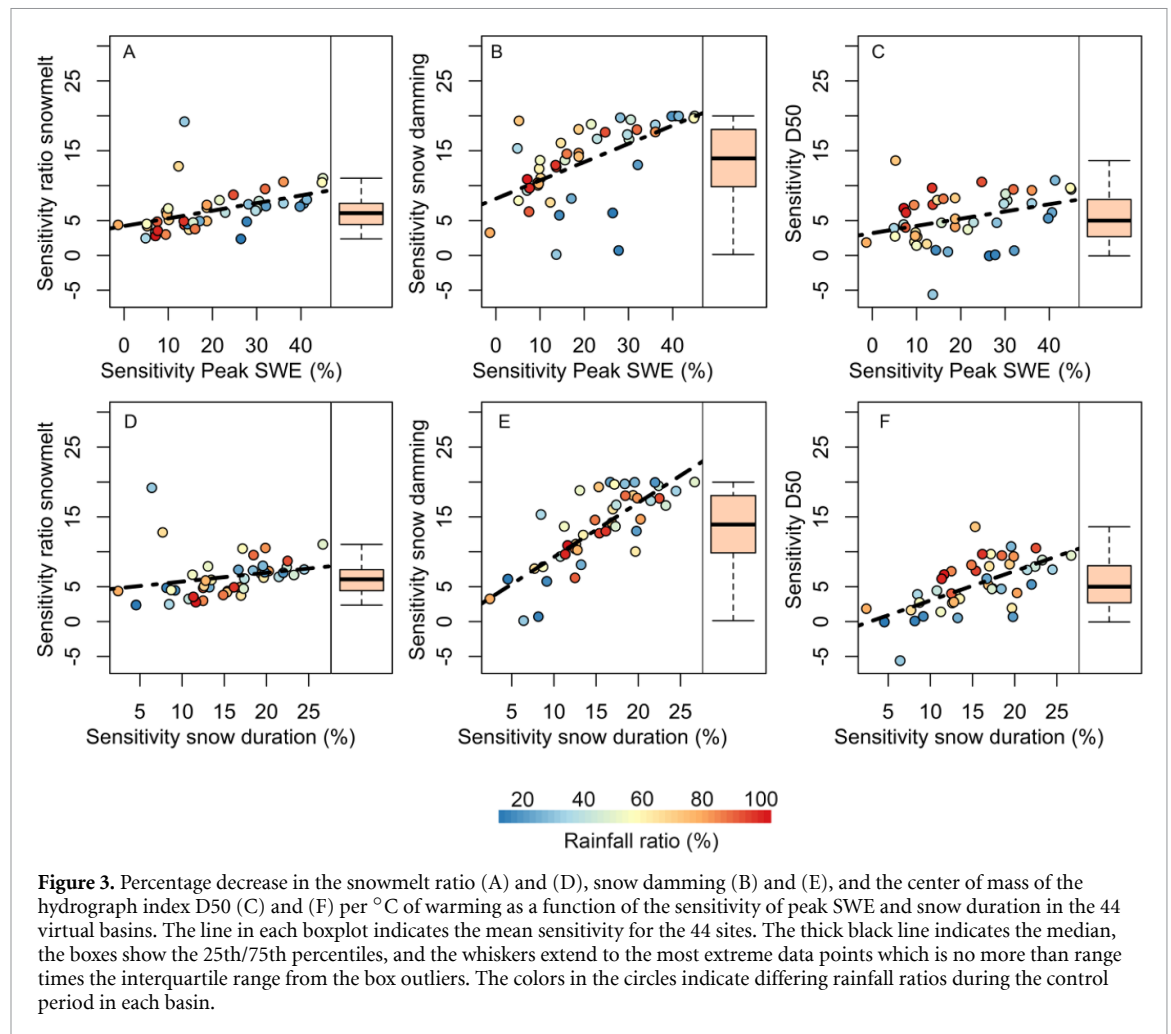
was much lower with the sensitivity of peak SWE ($r^2 = 0.17$; figure 3(f)). The D50 sensitivity differed among basins, with those having the largest rainfall ratios during the control period generally above the regression line. This suggests greater displacement of their hydrographs towards earlier melting, contrasting with those basins where more precipitation occurred as snowfall during the cold season. The combined use of the sensitivity of snow duration and rainfall ratio during the control period in the MLR model explained 63% of the total variance of D50 observed amongst the analyzed basins.

SI appendix and figures S6 and S7 show a high spatial agreement ($r^2 = 0.61$) between the spatial distribution of estimated sensitivities between D50 and SDam, and very low agreement in the distribution of snowmelt ratio to total runoff and D50 and SDam. Figure S5 shows that geographic patterns in the sensitivity of hydrological indices to warming are less evident than those for snow indices.

The sensitivity of annual runoff to climate warming was generally low, and annual runoff did not always decrease with increasing temperature, and showed no clear geographic pattern (figures 4 and *SI appendix,* S8). Sensitivities ranged from a −1.8% (increase) to +3.5% (decrease) per °C of warming; there was no statistical correlation with the sensitivity of peak SWE or snow duration, or with the ratio of snowmelt contribution to total runoff during the control period.

4. Discussion

The results of this study indicate a substantial and almost worldwide decline in snow accumulation and snow cover duration with climate warming, although with large spatial variability in the sensitivity to warming. Snowpacks located in warmer and more humid mountain ranges are likely to be the fastest to respond to temperature increase. Thus, coastal mountains, particularly those located at mid latitudes (including the majority of analyzed basins in the southern hemisphere, the Iberian Peninsula, and humid sectors of the Himalayas) were found to be most impacted by climate warming. Mountains located at high elevations or latitudes, particularly those in continental locations, showed a much more attenuated response to warming, and some were not sensitive to warming of up to 4 °C. This finding is consistent with other regional studies (Pomeroy *et al* 2015, Rasouli *et al* 2015, López-Moreno *et al* 2017, Harpold and Brooks 2018), and strongly suggests that simulated projections of temperature are poor proxies for how snowpacks will evolve in various mountain areas worldwide. However, analysis based on historical information on temperature and atmospheric humidity, readily available from global datasets, enables sound predictions of different levels of snowpack vulnerability to climate warming. This is

Figure 3. Percentage decrease in the snowmelt ratio (A) and (D), snow damming (B) and (E), and the center of mass of the hydrograph index D50 (C) and (F) per °C of warming as a function of the sensitivity of peak SWE and snow duration in the 44 virtual basins. The line in each boxplot indicates the mean sensitivity for the 44 sites. The thick black line indicates the median, the boxes show the 25th/75th percentiles, and the whiskers extend to the most extreme data points which is no more than range times the interquartile range from the box outliers. The colors in the circles indicate differing rainfall ratios during the control period in each basin.
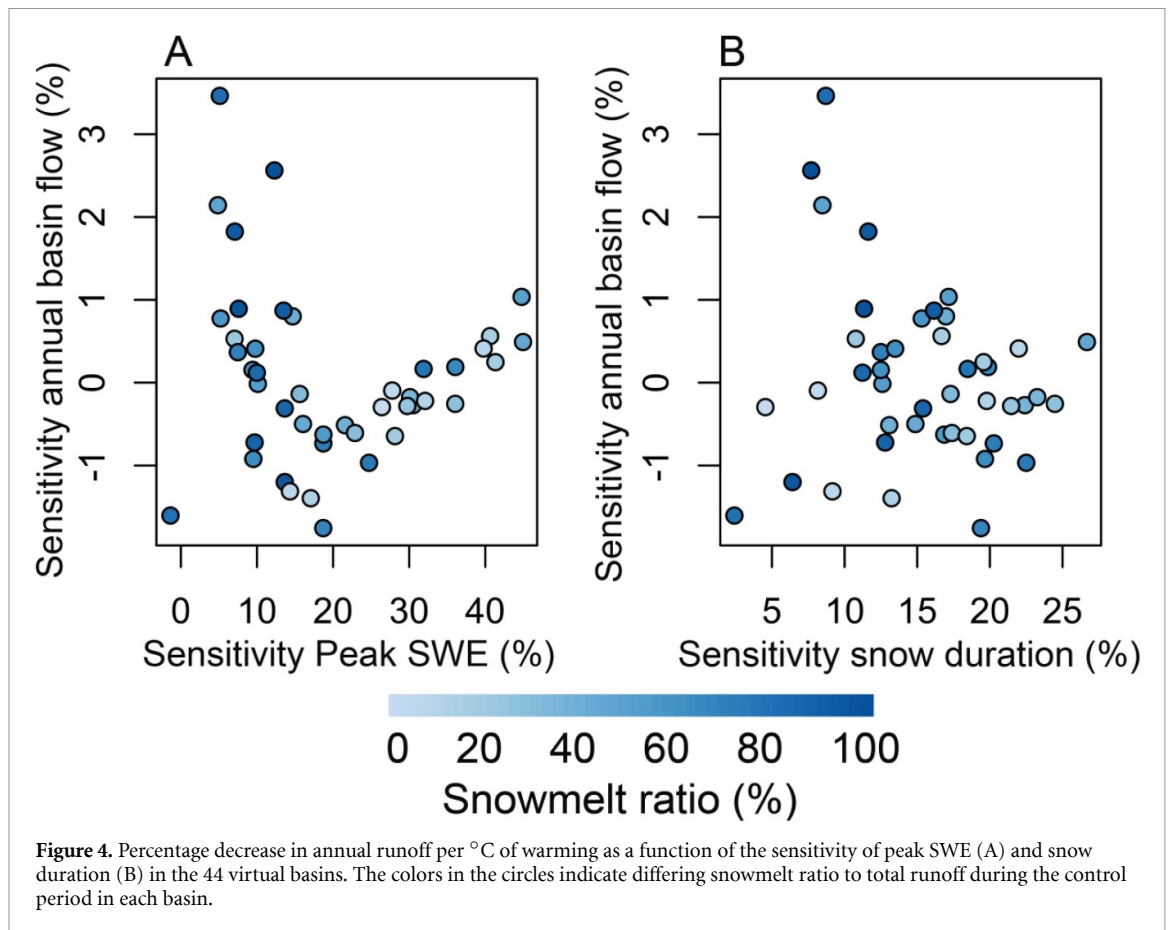
particularly relevant for predicting the response of the rainfall ratio and peak SWE (and probably other snow indices related to accumulation), because air temperature and humidity directly control the precipitation phase (reflected in the psychrometric equation; Harder and Pomeroy 2013), and hence influence seasonal snow accumulation and resulting streamflow hydrology (Harder and Pomeroy 2014). In contrast, predicting the sensitivity of snow cover duration to warming is more complicated because it also depends on the melt rate. Apart from rain-on-snow events, the melt rate is primarily driven by solar radiation rather than turbulent transfer of heat from the atmosphere (Painter *et al* 2010). In the most vulnerable basins for peak SWE, the sensitivity of snow duration was more attenuated. This may be related to the transition from seasonal to ephemeral snowpacks in the warmer and more humid mountain basins, leading to the peak SWE being more linked to isolated snowfall events or mid-winter ablation events, rather than to progressive snow accumulation during the entire cold season (Bilish *et al* 2018).

The hydrological behavior of mountain basins is likely to be less dependent on the snowpack as climate warms, with lesser contributions from snowmelt and river regimes reflecting more the seasonal

distribution of precipitation and rainfall–runoff processes, and consequently an earlier occurrence of the center of mass of the hydrograph. However, the magnitudes of these changes will be very variable amongst mountains having differing climatic conditions and snowpack sensitivities. This variability is more difficult to explain than that for snow indices. The most predictable index was the sensitivity of the snowmelt ratio, which responded almost linearly to the measured sensitivity of peak SWE. However, the sensitivity of the snowmelt ratio was markedly lower than that for peak SWE. This is because under warmer climate conditions the snowpack is thinner and of shorter duration; however, a large contribution of snowmelt to runoff through more numerous cycles of accumulation and melting is likely, even under ephemeral snowpack conditions (Bilish *et al* 2018).

Changes in the water stored by snowpacks (SDam) and shifts toward earlier snowmelt (D50) are more dependent on the sensitivity of snow duration, which in turn is more difficult to predict using simple diagnostic variables. In addition, to reach acceptable levels of predictability of D50 (63% of explained variance), it was necessary to combine the observed snowfall ratio with the sensitivity of snow duration. This finding strongly suggests that the D50 index

Figure 4. Percentage decrease in annual runoff per °C of warming as a function of the sensitivity of peak SWE (A) and snow duration (B) in the 44 virtual basins. The colors in the circles indicate differing snowmelt ratio to total runoff during the control period in each basin.

is not a robust indicator of change in the timing of snowmelt, as it strongly depends on the seasonal distribution of precipitation (Whitfield 2013).

No clear relationship was found between the magnitude of change in the snowpack as climate warms and a decline in annual runoff, as suggested in previous regional studies (Musselman *et al* 2017). In addition, among the virtual basins, those currently more influenced by snowmelt were not necessarily the most vulnerable to a decline in annual runoff as temperature increased, as has been suggested by recent studies based on the Budyko framework (Barnhart *et al* 2016). This difference can be explained by differing partitioning of the water balance components at each site, in particular the percentages of water loss from sublimation and evapotranspiration. Thus, sublimation from surface, intercepted and blowing snow is likely to decrease under a warmer climate (Painter *et al* 2010, Pomeroy *et al* 2015), and evapotranspiration during the snow-free period is expected to increase (Rasouli *et al* 2019); these factors may have opposing effects on the evolution of annual runoff. In addition, the presence or absence of frozen ground in cold and milder basins may have opposing effects on the infiltration of snowmelt water, also leading to contrasting effects on the evolution of annual runoff under climate warming (Iwata *et al* 2010), even under the assumption of slower melt rates in a warmer climate (Musselman *et al* 2017).

The virtual basin modelling approach in this study permitted isolation of the effects of different climates amongst snow-dominated basins in explaining their hydrological response to warming temperatures. The results obtained were based on the specific characteristics and parameters assigned to the virtual basins in the study, which are common to many alpine environments worldwide (steep slopes, shallow soils, little vegetation). However, real basins are much more complex than virtual basins in terms of the hydrological response to warming, because the response depends on many other factors including the size and hypsometry of each catchment, the vegetation cover and growth periods, the soil type and depth, and the role of groundwater storage. Each of these variables could have a profound impact on the sensitivities found in this study. For example, annual runoff in basins having deeper soils and more groundwater recharge could be much more sensitive to climate warming because of increased infiltration of melt water (Berghuijs *et al* 2014).

The predicted changing role of snowpacks as natural reservoirs, and the shifts in runoff timing as climate warms are problematic because the prediction has major economic consequences (Li *et al* 2017). This is especially the case for those sites where most of the precipitation occurs during the cold period (i.e. mountains having a Mediterranean climate), where the snowpack is key to ensuring water supply for

natural vegetation and irrigated crops during the high water demand summer period (López-Moreno *et al* 2017). The results highlight the difficulty of predicting the hydrological vulnerability of snow basins to climate change without developing specific case studies relying on scientifically sound, physically based approaches. This should be a priority for future research given the ecological, economic, and social importance of water generated in snow-fed mountains worldwide.

5. Conclusions

This study provides the first comparison of snowpack and hydrological sensitivity to climate warming across most of the climates that occur in the world's mountain areas. Snowpacks worldwide will be negatively affected by climate warming, and a generalized decoupling of mountain river hydrology from headwater snowpack regimes will occur. However, this study has revealed much complexity behind this generalization because snow and hydrology under the different climatic conditions explored will respond with different magnitudes to the same levels of warming. The variability found for the sensitivity of rainfall ratio and Peak SWE is highly predictable using diagnostic variables such as air temperature and humidity. This allows identification of the most/least vulnerable mountain regions to climate warming. However, the sensitivity of snow duration is more difficult to predict because of the more complex processes affecting the response of snowmelt to climate warming (e.g. radiative fluxes, redistribution and sublimation). This in turn hinders identification of the most vulnerable basins to: (1) losing the precipitation storage capacity of the snowpack during the cold season (quantified by the SDam index); and (2) advancing the snowmelt peakflow (quantified by the center of mass of the hydrograph, D50), because the sensitivity of both, SDam and D50, is mainly driven by the sensitivity of snow duration.

The sensitivity of the snowmelt ratio and of Peak SWE were closely related; but the former was substantially lower, decreasing mostly between 5% and 10% per °C, than the latter, which decreased mostly between 10% and 30% per °C. This important difference in magnitude has been associated with the shift from the important hydrological role of snowpacks with sustained accumulation and melting periods, to less important roles for ephemeral snowpacks, characterized by multiple accumulation and melting events during the snow season. In addition, no association was found between the magnitude of change of either Peak SWE or snow duration, and the change in annual runoff as climate warms.

The virtual basin approach has been proved useful for isolating the effects of differing climatic conditions on the response of snow and hydrology to climate warming. However, interpretation of the results needs to consider that the magnitudes of the sensitivities found are dependent on the specific parametrization imposed to the virtual basin such as basin shape and elevation range. Thus, sensitivity studies to basin configuration need to be developed using physically based approaches if accurate regional assessments of climate change impacts on water resources availability in snow dominated regions are to be achieved.

Data availability statement

All data that support the findings of this study are included within the article (and any supplementary information files).

ORCID iD

J I López-Moreno ⬤ https://orcid.org/0000-0002-7270-9313

References

Adam J C, Hamlet A F and Lettenmaier D P 2009 Implications of global climate change for snowmelt hydrology in the twenty-first century *Hydrol. Process* **23** 962–72

Armstrong R N, Pomeroy J W and Martz L W 2015 Variability in evaporation across the Canadian Prairie region during drought and non-drought periods *J. Hydrol.* **521** 182–95

Barnett T P, Adam J C and Lettenmaier D P 2005 Potential impacts of a warming climate on water availability in snow-dominated regions *Nature* **438** 303

Barnhart T B, Molotch N P, Livneh B, Harpold A A, Knowles J F and Schneider D 2016 Snowmelt rate dictates streamflow *Geophys. Res. Lett.* **43** 8006–16

Beniston M *et al* 2018 The European mountain cryosphere: A review of its current state, trends, and future challenges *Cryosphere* **12** 759–74

Berghuijs W R, Woods R A and Hrachowitz M 2014 A precipitation shift from snow towards rain leads to a decrease in streamflow *Nat. Clim. Change* **4** 583–6

Bhj V N, Antoine M, Wyseure G and Govers G 2011 Pattern-process relationships in surface hydrology: hydrological connectivity expressed in landscape metrics *Hydrol. Process* **25** 3760–73

Bilish S P, Mcgowan H and Callow J N 2018 Energy balance and snowmelt drivers of a marginal subalpine snowpack *Hyd. Process* **31** 3837–51

Bormann K J, Brown R D, Derksen C and Painter T H 2018 Estimating snow-cover trends from space *Nat. Clim. Change* **8** 924–8

Bosson E, Sabel U, Gustafsson L, Sassner M and Destouni G 2012 Influences of shifts in climate, landscape, and permafrost on terrestrial hydrology *J. Geophys. Res.* **117** D05120

Chakravarti R G, Laha G and Roy J 1967 *Handbook of Methods of Applied Statistics* vol I pp 392–4 (New York: Wiley)

Essery R and Pomeroy J 2004 Vegetation and topographic control of wind-blown snow distributions in distributed and aggregated simulations for an Arctic Tundra Basin *J. Hydrometeorol.* **5** 735–44

Fang X and Pomeroy J W 2016 Impact of antecedent conditions on simulations of a flood in a mountain headwater basin *Hydrol. Earth Syst. Sci.* **21** 3783–3506

Fang X, Pomeroy J W, Ellis C R, Macdonald M K, Debeer C M and Brown T 2013 Multi-variable evaluation of hydrological model predictions for a headwater basin in the Canadian Rocky Mountains *Hydrol. Earth Syst. Sci.* **17** 1635–59

Friedman J H 1991 Multivariate adaptive regression splines *Ann. Stat.* **19** 1–67

Harder P and Pomeroy J W 2014 Hydrological model uncertainty due to precipitation-phase partitioning methods *Hydrol. Process* **28** 4311–27

Harder P, Pomeroy J W and Helgason W 2017 Local-scale advection of sensible and latent heat during snowmelt *Geophys. Res. Lett.* **44** 9769–77

Harder P and Pomeroy J 2013 Estimating precipitation phase using a psychrometric energy balance method *Hydrol. Process* **27** 1901–14

Harpold A A and Brooks P D 2018 Humidity determines snowpack ablation under a warming climate *Proc. Natl. Acad. Sci. U. S. A* **115** 1215–20

Harpold A A, Rajagopal S, Crews J B, Winchell T and Schumer R 2017 Relative humidity has uneven effects on shifts from snow to rain over the Western U.S *Geophys. Res. Lett.* **44** 9742–50

Harpold A, Brooks P, Rajagopal S, Heidbuchel I, Jardine A and Stielstra C 2012 Changes in snowpack accumulation and ablation in the intermountain west *Water Resour. Res.* **48**

Huning L S and Aghakouchak A 2018 Mountain snowpack response to different levels of warming *Proc. Natl. Acad. Sci.* **115** 10932 LP–10937

Huntington J L and Niswonger R G 2012 Role of surface-water and groundwater interactions on projected summertime streamflow in snow dominated regions: an integrated modeling approach *Water Resour. Res.* **48**

Iwata Y, Hayashi M, Suzuki S, Hirota T and Hasegawa S 2010 Effects of snow cover on soil freezing, water movement, and snowmelt infiltration: A paired plot experiment *Water Resour. Res.* **46**

Jennings K S, Winchell T S, Livneh B and Molotch N P 2018 Spatial variation of the rain–snow temperature threshold across the Northern Hemisphere *Nat. Commun.* **9** 1148

López-Moreno J I, Pomeroy J, Revuelto J and Vicente-Serrano S M 2013 Response of snow processes to climate change: spatial variability in a small basin in the Spanish Pyrenees *Hydrol. Process.* **27** 2637–50

López-Moreno J I *et al* 2017 Different sensitivities of snowpacks to warming in Mediterranean climate mountain areas *Environ. Res. Lett.* **12**

López-Moreno J I, Boike J, Sanchez-Lorenzo A and Pomeroy J W 2016 Impact of climate warming on snow processes in Ny-Ålesund, a polar maritime site at Svalbard *Global Planet. Change* **146** 10–21

Mcgowan H, Callow J N, Soderholm J, Mcgrath G, Campbell M and Zhao J 2018 Global warming in the context of 2000 years of Australian alpine temperature and snow cover *Sci. Rep.* **8** 4394

Morán-Tejeda E, Lorenzo-Lacruz J, Lo´pez-moreno J I, Rahman K and Beniston M 2014 Streamflow timing of mountain rivers

in Spain: recent changes and future projections *J. Hydrol.* **517**

Musselman K N, Clark M P, Liu C, Ikeda K and Rasmussen R 2017 Slower snowmelt in a warmer world *Nat. Clim. Change* **7** 214–9

Ning L and Bradley R S 2015 Snow occurrence changes over the central and eastern United States under future warming scenarios *Sci. Rep.* **5** 17073

Painter T H, Deems J S, Belnap J, Hamlet A F, Landry C C and Udall B 2010 Response of Colorado river runoff to dust radiative forcing in snow *Proc. Natl. Acad. Sci. U. S. A* **107** 17125–30

Pepin N *et al* 2015 Elevation-dependent warming in mountain regions of the world *Nat. Clim. Change* **5** 424–30

Pomeroy J W, Gray D M, Brown T, Hedstrom N R, Quinton W L, Granger R J and Carey S K 2007 The cold regions hydrological model: A platform for basing process representation and model structure on physical evidence *Hydrol. Process* **21** 2650–67

Pomeroy J, Bernhardt M and Marks D 2015 Research network to track alpine water *Nature* **521** 32

Pomeroy J, Fang X and Ellis C 2012 Sensitivity of snowmelt hydrology in Marmot Creek, Alberta, to forest cover disturbance *Hydrol. Process* **26** 1891–904

Rasouli K, Pomeroy J W, Janowicz J R, Carey S K and Williams T J 2014 Hydrological sensitivity of a northern mountain basin to climate change *Hydrol. Process* **28** 4191–208

Rasouli K, Pomeroy J W and Marks D G 2015 Snowpack sensitivity to perturbed climate in a cool mid-latitude mountain catchment *Hydrol. Process* **29** 3925–40

Rasouli K, Pomeroy J W and Whitfield P H 2019 Are the effects of vegetation and soil changes as important as climate change impacts on hydrological processes? *Hydrol. Earth Syst. Sci.* **23** 4933–54

Revuelto J, López-Moreno J I, Azorin-Molina C and Vicente-Serrano S M 2014 Topographic control of snowpack distribution in a small catchment in the central Spanish Pyrenees: intra- and inter-annual persistence *Cryosphere* **8**

Rutter N *et al* 2009 Evaluation of forest snow processes models (SnowMIP2) *J. Geophys. Res.* **114** D06111

Sanmiguel-Vallelado A, Morán-Tejeda E, Alonso-González E and López-Moreno J I 2017 Effect of snow on mountain river regimes: an example from the Pyrenees *Front. Earth Sci.* **11**

Simpkins G 2018 Snow-related water woes *Nat. Clim. Change* **8** 945

Sospedra-Alfonso R and Merryfield W J 2017 Influences of temperature and precipitation on historical and future snowpack variability over the Northern Hemisphere in the second generation Canadian Earth System Model *J. Clim.* **30** 4633–56

Vuille M, Bradley R S, Werner M and Keimig F 2003 20th century climate change in the tropical Andes: observations and model results *Clim. Change* **59** 75–99

Weber M, Bernhardt M, Pomeroy J W, Fang X, Härer S and Schulz K 2016 Description of current and future snow processes in a small basin in the Bavarian Alps *Environ. Earth Scie.* **75** 1–18

Weedon G P, Balsamo G, Bellouin N, Gomes S, Best M J and Viterbo P 2014 The WFDEI meteorological forcing data set: WATCH Forcing Data methodology applied to ERA-Interim reanalysis data *Water Resour. Res.* **50** 7505–14

Weiler M and Mcdonnell J 2004 Virtual experiments: a new approach for improving process conceptualization in hillslope hydrology *J. Hydrol.* **285** 3–18

Whitfield P H 2013 Is 'Centre of Volume' a robust indicator of changes in snowmelt timing? *Hydrol. Process* **27** 2691–8

Zhou J, Pomeroy J W, Zhang W, Cheng G, Wang G X and Chen C 2014 Simulating cold regions hydrological processes using a modular model in the west of China *J. Hydrol.* **509** 13–24