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Response of snow processes to climate change: spatial variability in a small basin in the Spanish Pyrenees

J. I. López-Moreno,[1]* J. W Pomeroy,[2] J. Revuelto[1] and S. M. Vicente-Serrano[1]

[1] *Pyrenean Institute of Ecology (CSIC), Department of Geoenvironmental Processes and Global Change, Zaragoza, Spain*
[2] *Centre for Hydrology, University of Saskatchewan, Saskatoon, Canada*

Abstract:

In this study, the Cold Regions Hydrological Modelling platform was used to create an alpine snow model including wind redistribution of snow and energy balance snowmelt to simulate the snowpack over the period 1996–2009 in a small (33 ha) snow-dominated basin in the Spanish Pyrenees. The basin was divided into three hydrological response units (HRUs), based on contrasting physiographic and aerodynamic characteristics. A sensitivity analysis was conducted to calculate the snow water equivalent regime for various combinations of temperature and precipitation that differed from observed conditions. The results show that there was large inter-annual variability in the snowpack in this region of the Pyrenees because of its marked sensitivity to climatic conditions. Although the basin is small and quite homogeneous, snowpack seasonality and inter-annual evolution of the snowpack varied in each HRU. Snow accumulation change in relation to temperature change was approximately 20% for every 1 °C, and the duration of the snowpack was reduced by 20–30 days per °C. Melting rates decreased with increased temperature, and wind redistribution of snow was higher with decreased temperature. The magnitude and sign of changes in precipitation may markedly affect the response of the snowpack to changes in temperature. There was a non-linear response of snow to individual and combined changes in temperature and precipitation, with respect to both the magnitude and sign of the change. This was a consequence of the complex interactions among climate, topography and blowing snow in the study basin.

KEY WORDS snow; cold regions hydrological model (CRHM); climate change; alpine; Pyrenees

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INTRODUCTION

In Mediterranean mountains, the abundance and extent of snow cover affects many aspects of plant and fauna phenology, the availability of water resources, and the feasibility of (and risks associated with) a diverse range of economic activities in mountainous and high latitude regions (Breiling and Charamza, 1999; Beniston *et al.*, 2003; López-Moreno and García-Ruiz, 2004; De Jong *et al.*, 2009). It is widely recognized that mountain snowpacks exhibit large spatial and inter-annual variability at different scales, as a consequence of the controlling influence of topography on various components of the snow energy balance, and the marked sensitivity of snow to prevailing climatic conditions. Thus, the complexity of the terrain can affect temperature gradients over short distances, cause irregular distribution of precipitation, impact on incoming solar radiation, and cause local effects on wind direction and wind speed. The combination of temperature and precipitation can explain annual and decadal variations of snowpack at the regional scale (López-Moreno, 2005), but substantial variation in these parameters can occur at the local scale. Such differences may alter the impacts of climate variability and change on the ecology, hydrology and economy of particular mountain areas. For example, Uhlmann *et al.* (2009) evaluated the impact of climate change on skiing in numerous Swiss ski resorts and concluded that local factors including slope aspect or angle may determine the profitability of many ski resorts under the climate change scenarios for the 21st century. Differences in snowpack behaviour in neighbouring areas may also have implications for interpreting the evolution of snowpack based on a single site (e.g. automatic snow depth and SWE sensor, or snow course), as the physiographic characteristics of the site may cause a specific response of snow to the regional climate variability (Neumann *et al.*, 2006).

In recent years, there has been increasing interest in understanding and quantifying how mountain snowpack is affected by natural climate variability and under projected future climates (Minder, 2010). For analyses of the sensitivity of snowpack to altered climate, it is common to infer the impact of observed climatic trends on snowpack (Casola *et al.*, 2009) and to simulate future snow processes using climate projections from global and regional circulation models (GCMs and RCMs) (Rasmus *et al.*, 2004; Dankers and Christensen, 2005; Keller *et al.*, 2005; Merritt *et al.*, 2006; Hantel and Hirtl-Wielke, 2007; Mellander *et al.*, 2007; Uhlmann *et al.*, 2009).

Most studies have reported a generalized decline in snowpack in recent decades, as a consequence of the global rise in temperature (Brown and Mote, 2009). Deviations from this general trend have only been

*Correspondence to: Juan Ignacio López Moreno, Campus de Aula Dei. Avda. Montañana 1005, Zaragoza 50.059, Spain.
E-mail: nlopez@ipe.csic.es

reported in areas where precipitation has increased and mean winter temperatures are substantially below 0 °C, which results in a thicker snowpack (Raisänen, 2007). Climate models indicate that temperatures will continue to increase in coming decades (Solomon *et al.*, 2007), and mountain areas are expected to undergo particularly high rates of warming (Nogués-Bravo *et al.*, 2007). Thus, the observed decline in snowpack is likely to accelerate in the future (Adam *et al.*, 2009). However, the magnitude of the observed and projected changes may differ depending on the altitude (Mote *et al.*, 2005; López-Moreno *et al.*, 2009), the physiographic characteristics of the basin, and the sign of precipitation change, which can be highly variable even in adjacent areas (Raisänen, 2007) and may enhance or attenuate the impact of warmer temperatures on snowpack.

The Mediterranean mountains are subject to the above trends and projections for the future. The increase in average temperature during the 20th century (Bethoux *et al.*, 1998; Repapis and Philastras, 2004; Camuffo *et al.*, 2010) led to less snowfall and snow accumulation in the Italian Alps (Valt *et al.*, 2005), Slovakia (Vojtek *et al.*, 2003), northern Greece (Baltas, 2007) and the Pyrenees (López-Moreno, 2005). All models project a consistent trend toward warmer conditions in the coming decades (Gibelin and Deque, 2003; Giorgi, 2006; Goubanova and Li, 2007; Sanchez-Gomez *et al.*, 2009). The predicted magnitude of change for the 21st century varies from 1 °C to 6 °C, depending on the model and the greenhouse gas emission scenario used (Nogués-Bravo *et al.*, 2007). The magnitude of the temperature increase may severely affect snow accumulation and melting processes in all mountain ranges within the Mediterranean region (López-Moreno *et al.*, 2011).

In a previous study, an analysis of the sensitivity of the annual snow properties (duration and magnitude of the snowpack) was conducted in the Izas catchment (2056 m a.s.l., 0.33 km^2), which is located in the subalpine belt of the Central Spanish Pyrenees. This enabled assessment of the response of the snow water equivalent (SWE) to changes in climatic variables at a particular site. In the same study, the SWE series were simulated at the station level as a function of changes in temperature, precipitation, solar radiation, wind speed, and relative humidity projected by a set of RCMs for the end of the 21st century (López-Moreno *et al.*, 2008a). The results, which were based in a one-dimensional snow energy balance (ground energy balance for natural surfaces, Keller *et al.*, 2005), revealed the marked sensitivity of snowpack to climate change at that altitude. However, there was no assessment of how local conditions within the basin (including slope angle and aspect, snow redistribution by wind, and exposure to wind) might affect the sensitivity of snow processes or any investigation of the combined effects of changes in precipitation and temperature on snow sensitivity.

In the present study, the snowpack in the Izas basin was simulated for the period 1996–2009 using an alpine hydrology model created using CRHM (Cold Regions Hydrological Modelling platform). The model has been documented in general by Pomeroy *et al*. (2007), and its application to alpine and cold environments was described in detail by MacDonald *et al*. (2009, 2010); Ellis *et al.* (2010) and DeBeer and Pomeroy (2010). Using the various modules in the CRHM, and because of its spatially distributed nature (see methods section), it was possible to obtain the temporal SWE regimes for various hydrological response units (HRUs) reflecting the contrasting physiographic and aerodynamic characteristics of the basin. A sensitivity analysis was conducted, using SWE series calculated from combinations of temperature and precipitation that varied from observed conditions. The sensitivity analysis enabled assessment of a range of factors including the following: (i) how the response of snowpack to climatic conditions in each HRU might change at different altitudes (assuming that an increase or decrease of temperature represents a shift in temperature, and hence in the snow line, according to a given lapse rate); (ii) how snowpack in the basin might respond to warming scenarios projected for the region by various RCMs; and (iii) how changes in precipitation in the region, which are highly uncertain (López-Moreno *et al.*, 2008b), might enhance, or compensate, for the effects of higher temperatures on snowpack.

STUDY AREA: THE IZAS BASIN AND DEFINITION OF HYDROLOGICAL RESPONSE UNITS

The study area (the Izas experimental station; 42°44 N, 0°25 W) is located at 2056 m a.s.l. in Spain, near the main divide of the Pyrenees in the headwaters of the Gallego River, and very close to the border with France (Figure 1). As the Izas basin lies in a transition zone between the Atlantic Ocean and the Mediterranean Sea, it is subject to mixed climatic influences and conditions. The mean annual temperature is 3 °C, and there are, on average, 130 days each year when the mean temperature is colder than 0 °C (López-Moreno *et al.*, 2008a). The mean annual precipitation is approximately 2000 mm, of which, more than half falls as snow (Del Barrio *et al.*, 1997; Anderton *et al.*, 2004). Although the mean winter temperature is colder than 0 °C, the area is subject to intense warm spells during winter, which trigger melting events and major metamorphism of the snowpack throughout the snow season. The wind direction is predominantly NW – SE, and as wind drives blowing snow events, this has a major influence on the spatial distribution of snow within the basin (López-Moreno *et al.*, 2010).

The basin area is 0.33 km^2, and its altitude ranges from 2056 to 2280 m a.s.l. The bedrock is composed of densely fractured carboniferous slate. It presents a typical subalpine pasture landscape, there are no trees in the catchment, and land cover consists of high mountain meadows and rocky outcrops in steeper areas. With the exception of cliffs beneath some ridges, the landscape is rolling mountain topography with a mean slope of 16º.

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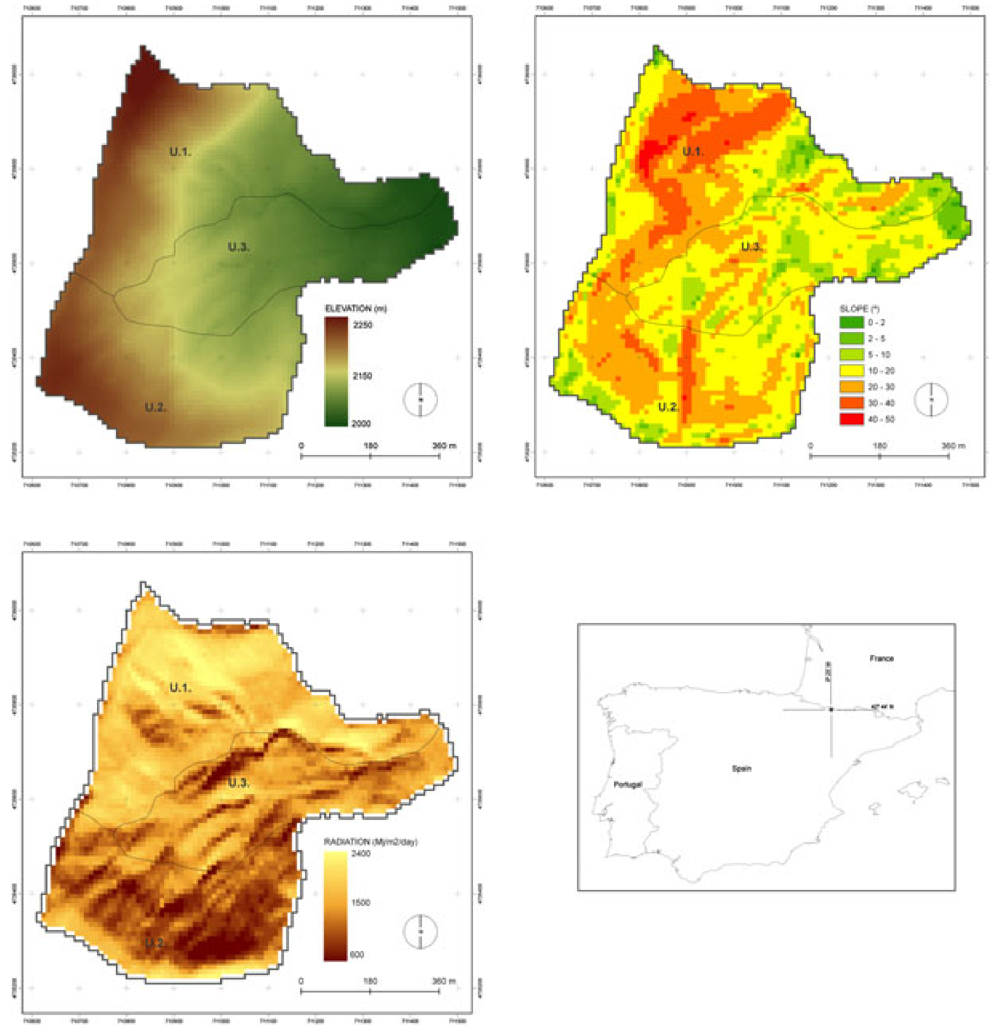

Figure 1. Location of the Izas basin (bottom right), and maps showing the elevation (upper left), slope (upper right) and incoming solar radiation (bottom left) for the area. Each map shows the limits of the three identified hydrological response units

DATA AND METHODS

From October 1996 to September 2009 (1996–2008 water years) information on climate and snow depth was recorded on an hourly basis at the Izas station. This included air temperature (T_{air}), precipitation (P), relative humidity (R_h), dew point temperature (T_d), wind speed (W_s), incoming ($K\downarrow$) and reflected ($K\uparrow$) solar radiation, and snow depth. The measuring devices used included a Vaisala HMP 35A probe for T_{air} and R_h, a Qualimetrics 6011 B tipping bucket rain gauge and Geonor T200B precipitation gauges, a RM Young 03002 anemometer, a Swissteco pyranometer for $K\downarrow$ and $K\uparrow$, and a Campbell SR50 ultrasonic sensor for snow depth. There have been also available 3 years with snow density data provided by a snow pillow (Sommer 1.5 × 1.5 m). Some data (e.g. precipitation, snow depth) were compared with that from other manual meteorological stations located in nearby locations (10–15 km away from Izas catchment) to detect potential errors in the series. Snow depth measurements were made using the Campbell SR50 ultrasonic sensor, which has an accuracy of ± 1 cm under ideal conditions. Depending on the different sensors, measurement reads were made and stored every 5–10 min and then computed into hourly date. The largest uncertainty in data comes from measurements of solid precipitation. These was derived from measured changes in snow depth (assuming a fresh snow density of 100 kg m^{-3}) and validated using measurements of water in an accumulative rain gauge and SWE data from a 1.5 × 1.5 m snow pillow installed in 1995. Installation of the Geonor T200B rain gauge in 2007 improved the direct estimation of winter precipitation and demonstrated the adequacy of the method for deriving snowfall quantity from snow depth variations.

Data for the period 1996–2008 were used as the input to the CRHM platform (Pomeroy *et al.*, 2007). This is a flexible object-oriented modeling system devised to connect algorithms for simulating the hydrological cycle in cold areas over small-sized to medium-sized basins. Landscape elements in the CRHM can be linked episodically in process-specific cascades including blowing snow transport and sublimation, overland flow, organic layer subsurface flow, mineral interflow, groundwater flow, and streamflow. Because there is a high level of confidence in the representation of cold regions processes in the modules, and good flexibility in the

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model structure, there is less need for calibration of parameters to streamflow observations for discharge simulations (Pomeroy *et al.*, 2012). Calibration can often be limited to streamflow routing and baseflow aspects of the model or omitted completely; thus, the model can be used for both prediction, diagnosis, and understanding of the hydrological processes. The CRHM uses a modular modeling object-oriented structure (Leavesley *et al.*, 1996) to develop, support, and apply dynamic model routines. Relative to other hydrological models, the modules associated with the CRHM can address the complete range of processes that characterize mountainous and cold regions (e.g. blowing snow, intercepted snow, energy balance snowmelt, infiltration to frozen soils) and can incorporate a wide range of process descriptions, from the conceptual to the physically based. The CRHM has been successfully applied to a wide variety of environments including prairies, forests, subalpine and alpine environments, and arctic basins (Pomeroy *et al.*, 2007; Dornes *et al.*, 2008; Essery *et al.*, 2009; DeBeer and Pomeroy, 2010; Ellis *et al.*, 2010; Fang *et al.*, 2010; MacDonald *et al.*, 2010; Knox *et al.*, in press). In the frame of Snowmip2 initiative, the model performed quite adequately in Snomip2 blind tests in Japan, Switzerland, Finland, Canada, and USA (Rutter *et al.*, 2009; Ellis *et al.*, 2010). This study is the first to apply the CRHM to a Mediterranean mountain environment.

Selection of the CRHM modules was based on data availability and the characteristics of the Izas basin. For evaporation, the Penman-based unsaturated evapotranspiration routine of Granger and Pomeroy (1997) was used, and for snowmelt, the energy balance snowmelt model (EBSM) developed by Gray and Landine (1988) was used. EBSM does not need any parametrization (Dornes *et al.*, 2008). Temperature thresholds of +2 and 0 °C were used to consider all precipitation as rain or snow, respectively. Routines for short wave direct and diffuse algorithms, slope correction (Garnier and Ohmura, 1970), snow albedo decay from a value of 0.85 for fresh snow (Gray and Landine, 1988), long wave radiation (Sicart *et al.*, 2004), and net radiation (Granger and Gray, 1990) were also selected. Blowing snow transport and sublimation were simulated according to the equations developed by Pomeroy and Li (2000) with adjustment to wind speed variations over complex terrain using the Walmsley *et al.* (1989) parametric solution for a boundary layer windflow model.

For application of the CRHM alpine model, the basin was divided into HRUs on the basis of its physiographic and aerodynamic characteristics. Despite using a subjective criteria for identifying different HRUs in the basin, we used a principal component analysis (PCA) to remove redundancy among variables and to classify them among a reduced number of classes. PCA enables common features and specific local characteristics to be identified (Richman, 1986). The terrain characteristics used to classify the surface in the HRUs were elevation, slope angle, and potential incoming radiation (R↓), all of which were derived from a digital elevation model with a grid cell size of 5 × 5 m (see López-Moreno *et al.*, 2010, for details of R↓ calculation). The PCA analyses reduced a large number of interrelated cases (3868 grid cells in this study) to a small number of independent principal components (PCs) that capture much of the variance of the original dataset (Hair *et al.*, 1998). In this study, three PCs summarized 87% of the total variance. The PCs were rotated (Varimax method) to redistribute the final explained variance and to obtain more stable, physically robust patterns (Richman, 1986). The spatial classification was carried out using the factorial loadings for each component, grouping the grid cells by the maximum loading rule. Most of the pixels classified in each class were contiguous in space, which enabled the limits of the three HRUs considered in the study to be determined.

Table I shows the main characteristics of each HRU. HRU1 has the steepest slopes and is the most exposed to solar radiation, having a general SE orientation. HRU2 represents the NW slopes of the basin, receives less radiation than HRU1, and based on the general slope aspect of the basin, is highly exposed to wind drift processes. HRU3 occupies the bottom of the basin, and has intermediate levels of incoming solar radiation and a low slope angle (Figure 1). Based on its position in the basin and the dominant wind direction, it is well protected from wind drift and could accumulate snow blown from other parts of the basin or from neighbouring areas.

The point station data were distributed to the 3 HRUs. Temperatures were lapsed to the HRU elevation from the station elevation using a lapse rate of 6.5 °C per 1000 m. Temperatures determined precipitation phase, sensible and latent heat fluxes, and incoming longwave radiation fluxes as well as many follow on effects. Shortwave radiation was adjusted for slope and aspect, whereas longwave radiation was adjusted for terrain view. Wind speeds were distributed using the Walmsley wind algorithm, which is based on the Mason and Sykes (1979) boundary layer wind flow model. The Mason and Sykes windflow model is a computational routine for estimating windflow over three-dimensional topography

Table I. Main characteristics of each hydrological response unit

Elevation (m)				**Slope**	**Bare soil**	**Radiation**	**Soil depth**
	Min	Mean	Max	(°)	%	(Mj/m^{-2}/day^{-1})	(cm)
HRU 1	2062	2171	2288	21.93	25.57	1864.24	51.66
HRU 2	2117	2194	2261	18.86	8.65	1297.82	50.22
HRU 3	2056	2119	2196	13.93	30.08	1516.06	48.81

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and is based on Fourier transform techniques and has a division of the inner and outer flow regions. It is linearized and, thus, only applies to low hills; the model assumes neutral thermal stratification and uniform surface roughness within the simulation region. Walmsley *et al.* (1989) derived a simple parametric version of Mason and Sykes' model for estimating wind speed variation induced by small-scale topographic features.

Selected outputs from the model were the hourly snow water equivalent (SWE), snow depth, melt, and blowing snow (snow drift) for each HRU. Hourly series were converted into daily data by taking the SWE simulated at noon and the accumulated drift during each day, and several snow indices were derived for each HRU for further analysis: (i) the accumulated snow water equivalent (ASWE in mm m^{-2} per season), which is the annual sum of daily positive differences in SWE; (ii) the maximum snow water equivalent (MSWE in mm m^{-2}), which is the maximum annual amount of SWE recorded in each HRU; (iii) the day of MSWE (DMSWE), which is the day of the year when the MSWE was observed, indicating commencement of the dominance of the melting conditions (1 October = day 1); (iv) the duration of the snowpack (DSP), which is the number of days with a snowpack thicker than 5 cm on the ground; (v) the last day with snow, which is the day on which the seasonal snowpack disappeared, that is thinner than 5 cm (1 October = day 1); (vi) the simulated snow transport by wind (DRIFT in mm m^{-2}), which is the eroded during each snow season; and (vii) the melting rate (MELT) in mmday^{-1}, which was obtained by dividing the MSWE by the number of days between the day of the MSWE and the last day with snow. It is an indicator of the rate at which the snowpack melted.

The sensitivity analysis was conducted, using SWE series calculated from combinations of temperature and precipitation that varied from observed conditions. The observed temperature was varied from −2 °C to +4 °C, at 1 °C intervals, whereas the precipitation was varied from −30% to +30%, at 10% intervals. This approach facilitated analysis of how climatic conditions drive the inter-annual variability of snowpack in the basin and assessment of how the physiographic characteristics of the basin affect various snow processes in each HRU.

RESULTS

Verification

Figure 2 shows the ability of the CRHM to reproduce the snowpack observed in the Izas basin. This includes plots of the daily snow depth simulated in HRU3 and observed at the meteorological station, which is located in HRU3 (Figure 2A). As the simulated snow depth is representative of an area, but the observed series represents only one point, caution in comparisons and interpretation is necessary. Despite some logical biases, CRHM reproduced the main patterns in the annual cycle

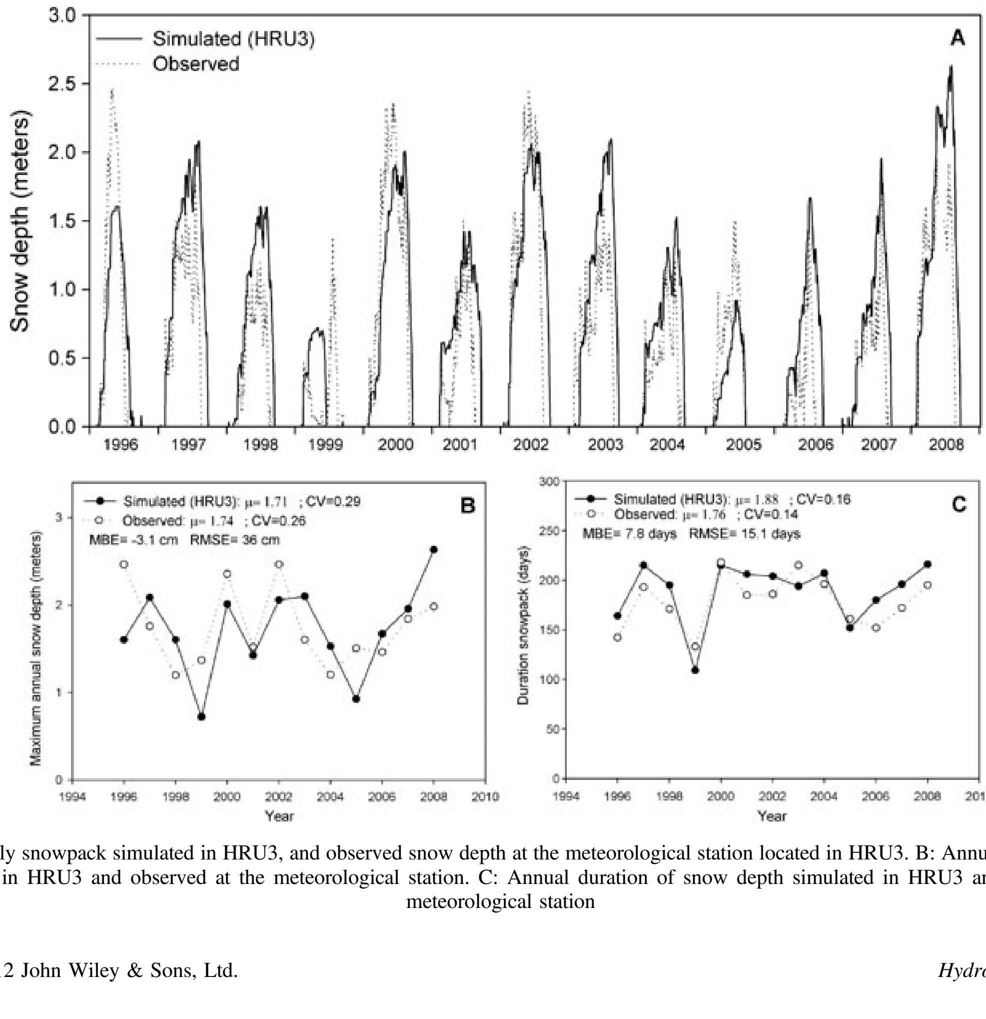

Figure 2. A: Daily snowpack simulated in HRU3, and observed snow depth at the meteorological station located in HRU3. B: Annual maximum snow depth simulated in HRU3 and observed at the meteorological station. C: Annual duration of snow depth simulated in HRU3 and observed at the meteorological station

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(accumulation and melting) as well as the inter-annual variability in the snowpack. Because both accumulation and ablation over time are well simulated and there is a quite good match in the timing of the melting onset, it means that CRHM is not only computing adequately the snow depth and SWE time series but also the cumulative energy balance and the bulk temperature of the snow pack, which drives the melting onset. Figure 2B and 2C shows the annual maximum accumulation and the annual duration of snowpack, respectively. In both cases, the simulated and observed values were similar for most of the study years. Thus, for the period 1996–2008, the observed and simulated mean maximum snowpack depths were 1.74 and 1.71 m, respectively, and the snowpack duration was 176 days and 188 days, respectively. Mean bias error (MBE) and root mean squared error (RMSE) are 7.8 and 15.1 days, and −3.1 and 36 cm for snow duration and maximum snowpack depth, respectively. In addition, the inter-annual coefficients of variation matched adequately, with 0.26 and 0.29 for maximum observed and simulated accumulation, respectively, and 0.14 and 0.16 for snowpack duration, respectively. Moreover, Figure 3 shows the mean monthly observed and simulated snow density in HRU3 for the three available years with snow pillow data. During the 3 years, the model is able to reproduce the variability of snow density along the snow season most of the time, particularly during March and April, when the peak SWE normally occurs. Unfortunately, there is no available quantitative information on snow processes in the two other HRUs. However, field observations have confirmed the reliability of different outputs of CRHM especially related to the marked differences in snow duration in the three units and the contrasting melting patterns as a consequence of the different exposure to solar radiation.

Observed inter-annual variability of snow indices and differences among HRUs

Figure 4 shows the mean monthly simulated snow water equivalent and melt during the period 1996–2009 for each HRU. The mean accumulation in the three HRUs was similar until February, after which accumulation was noticeably lower in HRU1; this HRU receives more radiation, which results in faster melting than in HRUs 2 and 3 (Figure 4B). Melting from February to May was greater in HRU3 (bottom of the catchment) than HRU2 (NW slope), but accumulation was very similar. The similar accumulation in both HRUs despite differing melting rates can only have been caused by wind redistribution, resulting in snow drifting from HRUs 1 and 2 to HRU3.

Figure 5 shows an example of the simulated evolution of daily SWE for three consecutive years with contrasting snow accumulation and duration. The evolution of SWE during this period illustrates that snow in the catchment exhibited high inter-annual variability and that the snowpack in each HRU behaved differently. Sub-basin differences showed marked year-to-year variation. Thus,

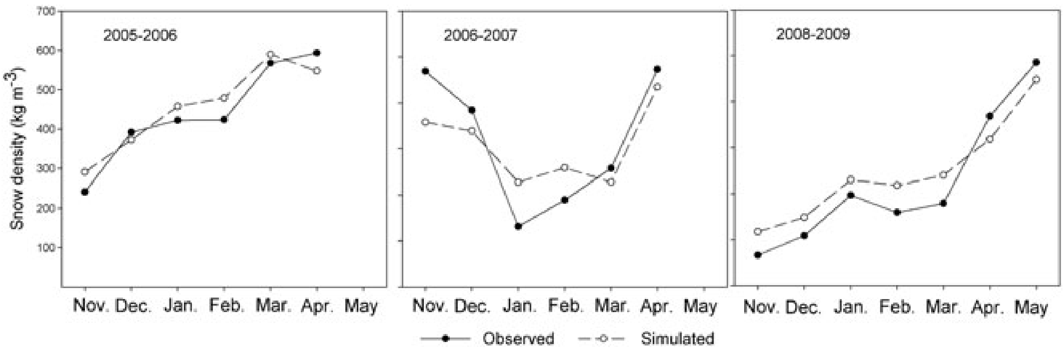

Figure 3. Mean monthly observed and simulated snow density in HRU3 for the three available years with snow pillow data

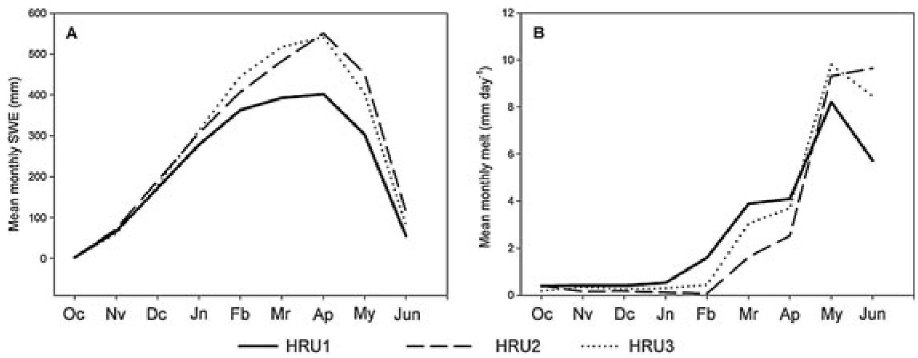

Figure 4. Mean monthly snow water equivalent and melt for each HRU

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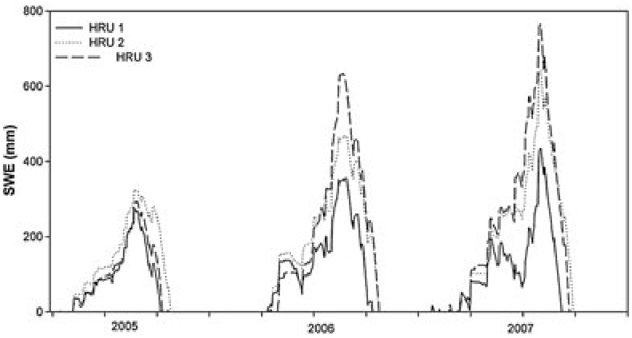

Figure 5. Simulated daily snow water equivalent in each HRU for three consecutive years

during the 2005 snow season, snow accumulation and melting were very similar in the three HRUs, but the snowpack was slightly deeper and lasted longer in HRU1. Simulations for 2006 showed large differences among the HRUs. HRU3 accumulated the deepest snowpack, and it lasted for a very similar period to that of the snowpack in HRU2. The model indicated that blowing snow accumulated at the bottom of the basin, where it was transported from HRU1 and HRU2. Melting was more rapid in HRU3 than HRU2, as consequence of the greater solar radiation in the former. Snow drift accumulation and high solar radiation in HRU1 resulted in markedly less snow. HRUs 2 and 3 had similar amounts of snow during 2007, which was again much less than in HRU3.

The boxplots in Figure 6 show the inter-annual variability of selected snow indices for each HRU and confirm that snowpack in this area was highly variable over the study years. It also shows that the climatic characteristics in the region may have caused contrasting impacts at the local scale, with marked differences apparent within the study basin. The area comprising the bottom of the basin (HRU3) showed the greatest variability in MSWE and ASWE. The average accumulation and the 75th percentile were very similar to those of HRU2, but higher values were found in the upper part of the distribution. In some years, HRU3 had the lowest MSWE and ASWE, although HRU1 had the lowest average MSWE and ASWE, mainly because of the large differences in the snow rich winters. Thus, the 95th percentile for MSWE in HRU1 was similar in magnitude to the 75th percentiles in HRU2 and HRU3. The melting rate between the day of MSWE occurrence and the last day with snow cover was lower in HRU 1 than in the other two HRUs, which exhibited similar melting rates. In several years of the record, the MSWE occurred on the same date, but in some years, the maximum accumulation occurred earlier in HRU3, followed by HRU1. Maximum accumu-

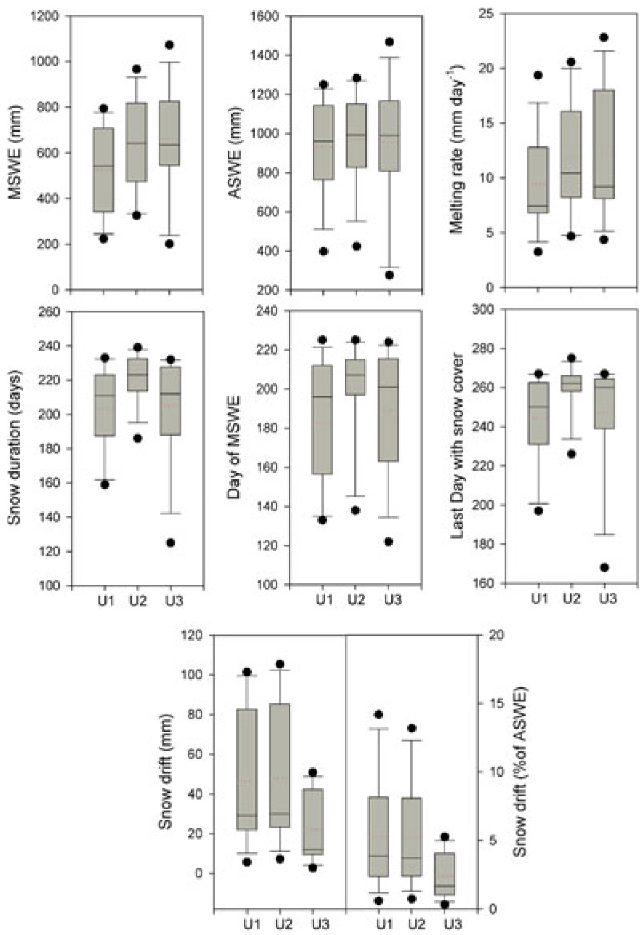

Figure 6. Inter-annual variability of selected snow indices in the three HRUs. The dotted line indicate the average, the solid line is the 50th percentile, the upper and lower parts of the box indicate the 75th and 25th percentiles (respectively), the whiskers are the 90th and 10th percentiles, and the dots indicate the 95th and 5th percentiles

lation in HRU2, which was at higher altitude and had lower incoming radiation, usually occurred later than those in the other two HRUs. In all HRUs, there was very high variability in the date of MSWE, which varied up to 3 months during the analysis period. As snowfall usually commences at the same time throughout the basin, the box-plot of snow duration was very similar to the box-plot of the last day with snow The period over which snow remained on the ground was much longer in HRU2 than in the other HRUs. Throughout most of the year, the snow lasted longer in HRU3 than in HRU1 (see Figure 4), but there were several years where the snow disappeared earlier from HRU3 than from HRU1. Wind transport of snow was a noteworthy process in HRUs 1 and 2, where approximately 15% of the ASWE blew from these areas. HRU3 lost only a 5% of ASWE because of drifting, and based on the ASWE, it was evident that in some years, the bottom of the basin trapped large amounts of wind-blown snow, which explains why the ASWE in HRU3 exceeds that in HRU2 by more than 100 mm, despite HRU2 being at a higher altitude and receiving less solar radiation than the former HRU.

The large inter-annual variability in the snow indices was mainly controlled by the dominant climatic conditions during the year. However, it is evident that each HRU responded differently to specific climatic conditions. Thus, Table II shows the correlations of MSWE and ASWE with mean temperature and precipitation from December to April, the duration of the snowpack in relation to temperature during the most important melting period (April – May), and the MSWE accumulated each winter. The accumulation indices (MSWE and ASWE) were statistically correlated ($p < 0.05$) with temperature and precipitation. However, the correlation was greater in HRU3 than in HRU2 or HRU1, where other factors including wind redistribution and solar radiation may have caused substantial local variations. The DSP was negatively correlated with temperature in April – May, particularly in HRU2. However, MSWE was the main factor controlling inter-annual variability of the DSP, especially in HRU3.

Sensitivity of snow indices to changes in precipitation and temperature

Figure 7 shows the sensitivity of selected snow indices to changes in temperature and precipitation in each HRU. Table III shows the response (% of change with respect the inter-annual average) of snow indices to different combinations of temperature and precipitation change. All indices and HRUs were highly sensitivity to shifts in climatic conditions, but there were marked differences in the magnitude and characteristics of the response. Snow accumulation was very sensitive to both precipitation and temperature. In the case of temperature, for all three HRUs, the MSWE decreased at a rate of approximately 20% per °C for the first two degree increase in temperature and at a slightly lower rate for greater changes (i.e. around 15% of change in MSWE from 2 to 3 °C). The opposite occurred if the temperature fell, but in this case, HRU2 exhibit a slightly lower response than HRU1 and HRU3. Similar changes were found for ASWE. Depending on the sign of change in precipitation, the response of snow to temperature was attenuated or enhanced. Thus, if temperature increased by 1 °C but precipitation also increased, a 20% snow accumulation remained practically unchanged in the three HRUs. However, if precipitation declined by the same amount, the effect of warming was doubled. When snowpack was simulated for a temperature increase of 1.5 °C, the snow indices decreased even if precipitation increased to 30%. For example, Table II shows that the MSWE could decrease more than 20% if temperature increased by 2 °C and precipitation decreased by 20%. However, this number was clearly exceeded if precipitation remained stationary or decreased by 20%, when MSWE might decrease more than 40% and 55%, respectively, in the three HRUs. It is noteworthy that under colder conditions, a change in precipitation could introduce greater differences among the HRUs. Thus, if temperature decreased by 1 °C and precipitation increased by 20%, the MSWE in HRU1 increased by 62%, whereas the increase in HRU2 was only 45%.

The DSP was sensitive to both temperature and precipitation, but the gradient of change associated with temperature was much greater than that for precipitation. Thus, under unchanged precipitation conditions, the DSP decreased by 15% (30 days) and 10% (20 days) for every °C positive change in HRU1 and HRU2, respectively, and an intermediate rate of change in HRU3. Under colder conditions, the rate of increase in DSP was slightly lower than the decrease under warmer conditions, Thus, DSP

Table II. Correlation of selected snow indices [maximum snow water equivalent, MSWE, accumulated snow water equivalent (ASWE), and duration of snowpack, (DSP)] with dominant climatic conditions and the preceding snow conditions. The correlations were statistically significant ($p < 0.05$) when $-0.4 > r$ value > 0.4

		Temperature December-April			**Precipitation December-April**			**Temperature and precipitation**		
		HRU1	**HRU2**	**HRU3**	**HRU1**	**HRU2**	**HRU3**	**HRU1**	**HRU2**	**HRU3**
MSWE	Correlation coefficient	0.53	0.49	0.64	0.35	0.46	0.56	0.62	0.63	0.76
ASWE	Correlation coefficient	0.52	0.49	0.62	0.39	0.43	0.48	0.53	0.59	0.71
		Temperature April-May			**MSWE**			**Temperature and MSWE**		
DSP	Correlation coefficient	−0.32	−0.49	−0.41	0.65	0.76	0.85	0.71	0.82	0.87

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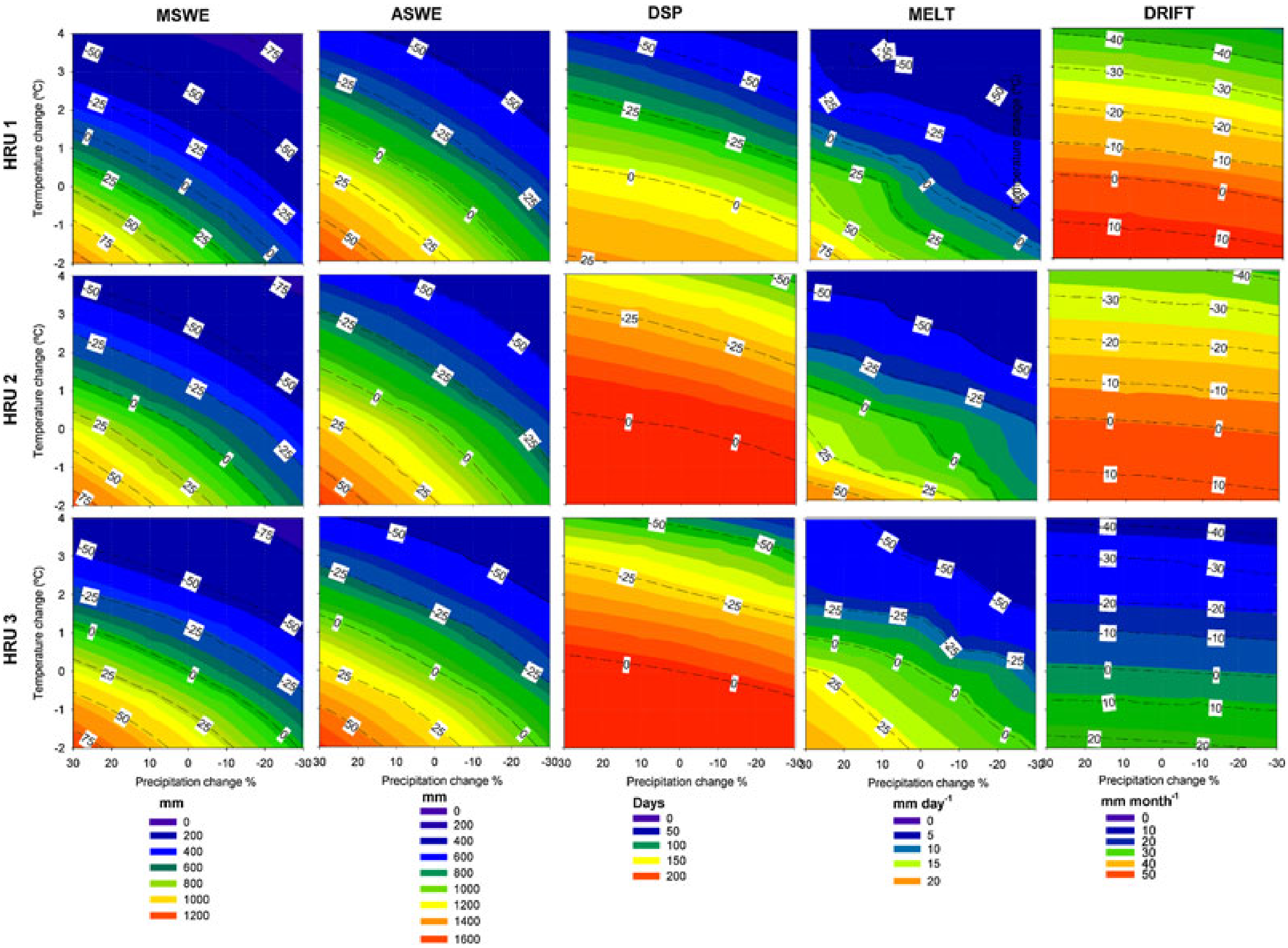

Figure 7. Sensitivity of snowpack to combined temperature and precipitation changes. The colors indicate snow index values for various pairs of temperature and precipitation change, and the isolines indicate the percentage change

would increase approximately an 11% per °C in HRU1 and 8% in each of HRU2 and HRU3. The slight downward trend in the horizontal isolines reflects the relatively low impact of precipitation on this parameter. For example, with an increase of 2 °C in temperature, the DSP in HRU3 would decrease by 18−31%, assuming a change in precipitation ranging from −20 to +20%. For the same area, but assuming a cooling of 2 °C and the same range of precipitation change, the DSP would increase by 16−23%.

The melt rate after the MSWE date was sensitive to both precipitation and temperature. The general simulation pattern was higher snowpack melting rates as precipitation increased and lower melting rates with warmer temperatures. This is because of warmer conditions forcing melt to occur earlier in the season, under lower insolation conditions and wetter conditions forcing melt to occur later in the season under higher isolation conditions. There were marked differences in sensitivity to melt rates among the three HRUs, especially in simulations with colder temperatures. As an HRU received more solar radiation, its response to temperature change was more pronounced. Thus, with a 2 °C decrease in temperature in HRU1, the melt rate increased by 70% and 18% when precipitation changed from +20% to −20% respectively, whereas in HRU2, the rate changed by 59% and −2%, respectively. The melt rate in HRU1 was the most sensitive under warm temperatures, but the differences among the HRUs were less than under cool conditions.

Snow drift was also very sensitive to temperature and precipitation changes, with percentage changes similar to those observed for MSWE. Changes in snow drift are mainly controlled by temperature, whereas precipitation plays a secondary role. Thus, under cold and wet conditions, the percentage of drifted snow would be higher, and the opposite would occur under a warmer and dryer climate because of the effect of warmer temperatures and shallower snow in restricting snow transportability.

Figure 8 shows the inter-annual variability of various snow indices under selected combinations of temperature and precipitation change. It has previously been shown that the inter-annual average of the various indices can change substantially in response to altered precipitation and temperature. Thus, assuming a temperature increase of 2 °C during snow-rich years (90th and 95th percentiles), the accumulation of snow (MSWE and ASWE) would never exceed the mean accumulation under observed climate conditions, and the mean accumulation in a warmer climate would fall below the 25th percentile simulated with observed condition. If temperature were to

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Table III. Response (% of change with respect the inter-annual average) of snow indices to different combinations of temperature and precipitation change

		MSWE			ASWE			DSP			Melt			Drift		
Change		1	2	3	1	2	3	1	2	3	1	2	3	1	2	3
Temperature	Precipitation	524.4	641.7	660.2	932.3	970.9	954.6	203.2	221.2	205.1	9.4	11.9	12.2	46.4	47.9	22.2
T-2°C	**P-20%**	10	3	13	3	0	8	15	13	16	18	−2	7	13	12	20
	P0%	49	36	42	31	26	33	21	17	20	43	36	23	15	14	22
	P + 20%	90	69	72	59	53	58	25	20	23	70	59	40	16	15	23
T-1 °C	**P-20%**	−11	−11	−3	−11	−12	−4	4	4	7	−4	−11	−6	6	6 11	
	P0%	25	17	24	15	13	19	12	8	12	16	5	13	8	7	12
	P + 20%	62	45	50	40	37	41	16	11	15	47	22	30	9	8	13
T0°C	**P-20%**	−29	−25	−23	−23	−22	−20	−9	−4	−5	−26	−16	−11	−3	−1	−1
	P0%	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
	P + 20%	30	25	22	23	22	19	5	3	3	39	17	21	2	1	1
T + 1 °C	**P-20%**	−44	−41	−40	−37	−36	−32	−26	−17	−16	−31	−36	−39	−17	−12	−11
	P0%	−22	−21	−21	−18	−17	−16	−15	−11	−10	−17	−25	−14	−13	−10	−10
	P + 20%	2	−1	−3	1	1	1	−9	−7	−6	−2	−8	−7	−10	−9	−9
T + 2 °C	**P-20%**	−58	−55	−57	−48	−46	−46	−41	−25	−31	−54	−57	−52	−26	−20	−25
	P0%	−41	−39	−43	−32	−31	−32	−31	−21	−24	−37	−37	−38	−23	−19	−23
	P + 20%	−23	−23	−29	−16	−15	−19	−23	−16	−18	−27	−28	−35	−21	−18	−22
T + 3 °C	**P-20%**	−70	−65	−70	−58	−56	−57	−54	−35	−42	−49	−65	−62	−37	−30	−33
	P0%	−56	−54	−60	−45	−43	−46	−44	−29	−36	−52	−58	−52	−33	−28	−30
	P + 20%	−43	−41	−50	−32	−30	−35	−36	−25	−29	−54	−45	−43	−30	−27	−30

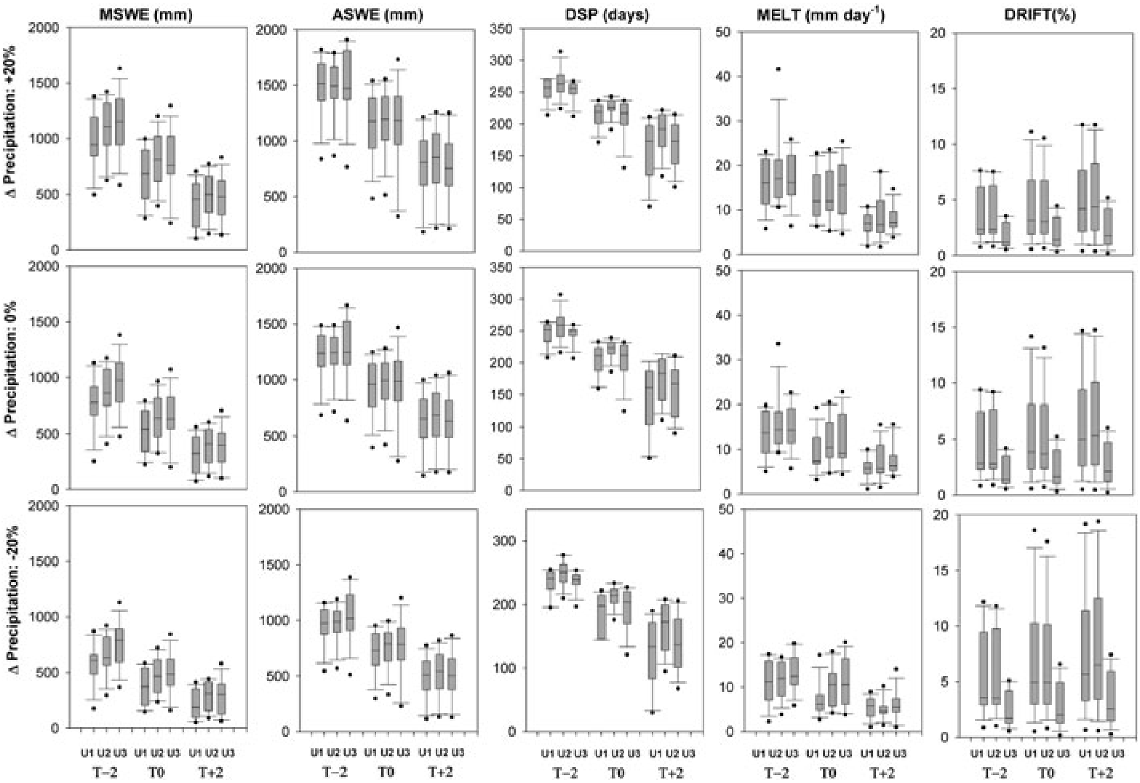

Figure 8. Inter-annual variability of various snow indices under selected combinations of temperature and precipitation change

increase 2º and precipitation decline by 20%, the mean simulated accumulation (MSWE and ASWE) would be below the 10th percentile registered under observed climatic conditions. It was also evident that the range of inter-annual accumulation differed in each HRU under various combinations of precipitation and temperature change. For instance, the boxplots of ASWE for HRUs 1 and 2 are very similar for a temperature 2 °C colder or 20% less precipitation but are quite different if precipitation is 20% less but the temperature is 2 °C warmer. The DSP may be strongly affected by warmer temperatures and less precipitation. In HRU 1 there could be years with less than 2 months of snow cover, although snow could remain for at least for 3 months in HRUs 2 and 3, even under such limiting conditions. A decrease in temperature would lead to a longer DSP in HRU 2 relative to HRU1 and HRU 3. The boxplots of the annual melt rate illustrates the increasing melt rate with decreasing temperature and increasing precipitation. The inter-annual variability of drifted snow increased markedly in HRUs 1 and 2 when snow accumulation decreased under warmer and dryer conditions. Under such conditions, there were alternate years with practically no snow drift, with others in which practically a 20% of accumulated snow is left out from both HRUs. Such variability decreased with greater precipitation and colder temperatures.

DISCUSSION

Snowpack exhibits large temporal and spatial variability in the Izas basin. Inter-annual variability of snowpack is a common feature of snow-covered areas, which is explained by the marked sensitivity of snowpack to the dominant climatic conditions (Brown and Goodison, 1996; Fassnacht, 2006; Kapnik and Hall, 2010). The main variables thought to explain this variability are precipitation and temperature (DeWalle and Rango, 2008; Ghatak *et al.*, 2010). In the Izas basin, both variables have also been shown to have a statistically significant correlation with indices of snowpack accumulation, melting, and duration. Nonetheless, in the study area, we observed that under the same climatic conditions, the correlations with precipitation and temperature differed in the three HRUs and that the HRUs also differed in sensitivity to each climate parameter. The results provide evidence that incoming solar radiation (mainly determined by slope angle), the redistribution of snow by wind, and small differences in altitude in the basin may explain the different responses of each HRU to climate conditions.

The accumulation period in the basin was strongly related to the combination of temperature and precipitation, which determines the amount of snowfall over the basin and the disposition of precipitation into snowfall or rainfall. However, during late winter and early spring, the temperature is often close to or above freezing, which leads to a progressive warming of the snow profile. Thus, in those areas, more exposed to solar radiation, the snowpack receives additional energy, which may be sufficient to initiate the thawing process, whereas other areas may not reach isothermal conditions. This explains why in some years, melting first commenced in HRU1, which has the most sun-exposed slopes, or in HRU3,

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which receives less solar radiation but is at a slightly lower altitude. In contrast, melting generally occurred last in HRU2 because of its NW-facing slopes. This explains why most previous snow studies point to slope aspect, or exposure to radiation, as the main factors underpinning the spatial distribution of snowpack (Anderton *et al.*, 2004; Molotch *et al.*, 2005; López-Moreno *et al.*, 2010). Moreover, wind can cause snow redistribution in a given basin, with snow drifting from wind exposed slopes and accumulating on the lee side or in concave areas. In some cases, wind may be the main factor explaining variability in snow accumulation (Essery and Pomeroy, 2004; Iacozza and Barber, 2010; Mott and Lehning, 2010). In the Izas basin, we observed that during certain years, HRU 3 accumulated the maximum snow water equivalent, despite being at the lowest altitude and receiving most solar radiation. This can only be explained by snow being blown from the other HRUs. After maximum accumulation has occurred, melting rates are determined not only by temperature, but also by topography and the amount of previously accumulated snow. A surprising result was that melting rates tended to increase under colder conditions. This is a consequence of the relationship between temperature and solar radiation in determining melting rates. With cold temperatures, solar radiation at the end of winter is not sufficient to trigger melting, and the snow profile remains cold later into the season. The snowpack eventually reaches isothermal conditions when the solar radiation is more intense and the days are longer. Thus, melting rates are accelerated. A similar rationale explains the negative relationship between snow accumulation and melting rate. As more snow accumulates, more time is needed for complete thawing, with the result that snow remains until the temperature is higher and the solar radiation is greater.

The results of the study confirm the sensitivity of snow processes to climate (temperature and precipitation) in the subalpine belt of the Pyrenees. The response of snow accumulation to temperature was approximately 20% per °C. This is the same as the rate reported by Casola *et al.* (2009) for the Washington Cascades area (20% decrease per °C) and similar to that estimated by Beniston *et al.* (2003) for the Swiss Alps (15% decrease per °C). Changes in precipitation will also determine the percentage change in SWE caused by temperature. Thus, a 20% increase in precipitation in the basin may compensate for a decrease in ASWE associated with a warming of 1 °C. However, a similar decrease in precipitation may double the effect of a temperature change. The importance of changes in precipitation for estimating the impact of climate change in a particular area has been highlighted in previous studies (Dyer and Mote, 2006; Raisänen, 2007; Brown and Mote, 2009). Indeed, this may explain why for some areas, an increase in snowpack has occurred despite a positive evolution of temperature (Hyvärinen, 2003). This highlights the difficulty of assessing the future evolution of snowpack in mountainous areas, where simulations of precipitation responses to anthropogenic forcing remain unpredictable (Räisänen, 2006; López-Moreno *et al.*, 2008a). The results also suggest that wind transport of snow increases with colder temperature because warm temperatures after snowfall events reduce blowing snow, as wet snow is not readily blown relative to dry snow (Li and Pomeroy, 1997).

This study also highlights the non-linearity of the sensitivity of snow parameters to temperature and precipitation, with respect to both the magnitude and sign of climatic change. For instance, the rate of decrease in snow accumulation declined as considered warming from current conditions increased. Similarly, the DSP differed in sensitivity for equivalent increases or decreases in precipitation. Moreover, although the sensitivity was similar in all three HRUs, some differences were evident. In general, as the exposure of each HRU to solar radiation increased, greater sensitivity to temperature and precipitation changes was observed. The non-linearity and spatial differences in the sensitivity of snow processes to climate change were a consequence of the complex interactions among climate, topography and snow remobilization by wind. Such interactions explain why each HRU can respond differently to particular climatic characteristics in any year. These results have clear implications for assessing the impact of future scenarios of climate change on snowpack. Thus, a regional increase in temperature may have different effects at the local scale, depending on altitude, topography, and exposure to wind (Lopez-Moreno *et al.*, 2009; Uhlmann *et al.*, 2009). Moreover, precipitation largely determines the magnitude of the impact of increasing temperature on snowpack. Spatially distributed and semi-distributed approaches to the modeling of energy balances and the hydrology of snowpack (Pomeroy *et al.*, 2007; Magnusson *et al.*, 2010) are appropriate tools for assessing the local response of snowpack to global environmental change.

CONCLUSIONS

A large inter-annual variability in snowpack was found in a small basin in the Pyrenees, resulting from the marked sensitivity of snowpack to climatic condition. Thus, an increase in temperature could decrease the accumulation of snow by 20% and reduce the duration of snowpack by 20–30 days. However, the sign and magnitude of precipitation changes may attenuate or enhance the effects of a warmer climate. Even in small basins with low altitudinal gradients, such as the one involved in this study (0.33 km^{-2}; altitudinal range approximately 200 m), snow processes may be variable because of differing climatic conditions within the basin, arising from particular topographic and aerodynamic characteristics. In 1 year, an HRU (e.g. HRU3) might exhibit the highest MSWE or the longest DSP, but in other years have the lowest accumulation and DSP. This is a consequence of the non-linear response of each HRU to various combinations of temperature and precipitation change. The non-linear response is explained by the complex

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interactions among climatic conditions, topography and wind redistribution of snow, which markedly affect the mass and energy balance of snowpack.

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