Snow sensitivity to climate change during compound cold-hot and wet-dry seasons in the Pyrenees.

Josep Bonsoms [1], Juan I. López-Moreno [2], Esteban Alonso-González [3]

[1] Department of Geography, Universitat de Barcelona, Barcelona, Spain.

[2] Instituto Pirenaico de Ecología (IPE-CSIC), Campus de Aula Dei, Zaragoza, Spain.

[3] Centre d'Etudes Spatiales de la Biosphère (CESBIO), Université de Toulouse, CNES/CNRS/IRD/UPS, Toulouse, France.

Corresponding author: Juan I. López-Moreno (nlopez@ipe.csic.es)

Abstract. The Mediterranean basin has experienced one of the highest warming rates on earth during the last few decades, and climate projections predict water-scarcity in the future. Mid-latitude Mediterranean mountain areas, such as the Pyrenees, play a key role in the hydrological resources for the highly populated lowland areas. However, there are still large uncertainties about the impact of climate change on snowpack in the high mountain ranges of this region. Here, we perform a sensitivity analysis of the Pyrenean snowpack using five key snow-climatological indicators. We analyzed snow sensitivity to seasons with four different compound extreme weather conditions (cold-dry [CD], cold-wet [CW], warm-dry [WD], and warm-wet [WW]) at low elevations (1500 m), mid-elevations (1800 m), and high elevations (2400 m) in the Pyrenees. In particular, we forced a physically based energy and mass balance snow model (FSM2), with validation by ground-truth data, and applied this model to the entire range, with forcing of perturbed reanalysis climate data for the period 1980 to 2019 as the baseline. The FSM2 model results successfully reproduced the observed snow depth (HS) values (R^2 > 0.8), with relative root-mean square error and mean absolute error values less than 10% of the observed HS values. Overall, the snow sensitivity decreased with elevation and increased towards the eastern Pyrenees. When the temperature increased progressively at 1ºC intervals, the largest seasonal HS decreases from the baseline were at +1ºC (47% at low elevation, 48% at mid-elevation, and 25% at high elevation). A 10% increase of precipitation counterbalanced the temperature increases (≤1ºC) at high elevations during the coldest months, because temperature was far from the isothermal 0ºC conditions. The maximal seasonal HS and peak HS max reductions were during WW seasons, and the minimal reductions were during CD seasons. During WW (CD) seasons, the seasonal HS decline per °C was 37% (28 %) at low elevations, 34% (30%) at mid-elevations, and 27% (22 %) at high elevations. For climate indicators of snow ablation, the largest decreases were during the WD seasons, when the peak HS date is anticipated 10 days and snow duration decreases 12% per ºC of warming. Results suggests snow sensitivity will be similar at other mid-latitude mountain

areas, where snowpack reductions will have major consequences on the nearby ecological and socioeconomic systems.

Keywords: Snow, Climate change, Sensitivity, Alpine, Mediterranean Mountains, Mid-latitude, Pyrenees.

1 Introduction

Snow is a key element of the Earth's climate system (Armstrong and Brun, 1998) because it cools the planet (Serreze and Barry, 2011) by altering the Surface Energy Balance (SEB), decreasing the albedo, and increasing surface and air temperatures (Hall, 2004). Northern-Hemispheric snowpack patterns have changed rapidly during recent decades (Hammond et al., 2018; Hock et al., 2019; Notarnicola et al., 2020). It is crucial to improve our understanding of the timing of snow ablation and snow accumulation due to changing climate conditions because snowpack affects many nearby social and environmental systems. From the hydrological point of view, snow melt controls mountain runoff rate during the spring (Barnett et al., 2005; Adams et al., 2009; Stahl et al., 2010), river flow magnitude and timing (Morán-Tejeda et al., 2014; Sanmiguel-Vallelado et al., 2017), water infiltration and groundwater storage (Gribovszki et al., 2010; Evans et al., 2018), and transpiration rate (Cooper et al., 2020). The presence and duration of snowpack affects terrestrial ecosystem dynamics because snow ablation date affects photosynthesis (Woelber et al., 2018), forest productivity (Barnard et al., 2018), freezing and thawing of the soil (Luetschg et al., 2008; Oliva et al., 2014), and thickness of the active layer in permafrost environments (Hrbáček et al., 2016; Magnin et al., 2017). Snowpack also has remarkable economic impacts. For example, the snowpack at high elevations and surrounding areas determines the economic success of many mountain ski-resorts (Scott et al., 2003; Pons et al., 2015; Gilaberte-Búrdalo et al., 2017). Changes in the snowpack of mountainous regions also influence associated lowland areas because it affects the availability of snow meltwater that is used for water reservoirs, hydropower generation, agriculture, industries, and other applications (e.g., Sturm et al., 2017; Beniston et al., 2018).

Mid-latitude snowpacks have among the highest snow sensitivities worldwide (Brown and Mote, 2009; López-Moreno et al., 2017; 2020b). In regions at high latitudes or high elevations, increasing precipitation can partly counterbalance the effect of increases of temperature on snowpack duration (Brown and Mote, 2009). Climate warming decreases the maximum and seasonal snow depth (HS), the snow water equivalent (SWE) (Trujillo and Molotch, 2014; Alonso-González et al., 2020a; López-Moreno et al., 2013; 2017), and the fraction total precipitation as snowfall (snowfall ratio; e.g., Mote et al., 2005; Lynn et al., 2020; Jeenings and Molotoch, 2020; Marshall et al., 2019), and also delays the snow onset date (Beniston, 2009; Klein et al., 2016). However, warming can slow the early snow ablation rate on the season (Pomeroy et al., 2015; Rasouli et al., 2015; Jennings and Molotch, 2020; Bonsoms et al., 2022; Sanmiguel-Vallelado et al., 2022) because of the earlier HS and SWE peak dates (Alonso-González et al., 2022), which coincide with periods of low solar radiation (Pomeroy et al., 2015; Musselman et al., 2017a).

The Mediterranean basin is a region that is critically affected by climate change (Giorgi, 2006) being densely populated (>500 million inhabitants) and affected by an intense anthropogenic activity. Warming of the Mediterranean basin will accelerate for the next decades, and temperatures will continue to increase in this region during the warm months (Knutti and Sedlacek, 2013; Lionello and Scarascia 2018; Cramer et al., 2018; Evin et al., 2021; Cos et al., 2022), increasing atmospheric evaporative demands (Vicente-Serrano et al., 2020), drought severity (Tramblay et al., 2020), leading to water-scarcity over most of this region (García-Ruiz et al., 2011). Mediterranean mid-latitude mountains, such as the Pyrenees, where this research focuses, are the main runoff generation zones of the downstream areas (Viviroli and Weingartner, 2004) and provide most of the water used by major cities in the lowlands (Morán-Tejeda et al., 2014).

Snow patterns in the Pyrenees have high spatial diversity (Alonso-González et al., 2019), due to internal climate variability of mid-latitude precipitation (Hawkins and Sutton 2010; Deser et al., 2012), high interannual and decadal variability of precipitation in the Iberian Peninsula (Esteban-Parra et al., 1998; Peña-Angulo et al., 2020) as well as the abrupt topography and the different mountain exposure to the main air masses (Bonsoms et al., 2021a). Thus, snow accumulation per season is almost twice as much in the northern slopes as in the southern slopes (Navarro-Serrano and López-Moreno, 2017), and there is a high interannual variability of snow in regions at lower elevations (Alonso-González et al., 2020a) and in the southern and eastern regions of the Pyrenees (Salvador-Franch et al., 2014; Salvador-Franch et al., 2016; Bonsoms et al., 2021b). Since the 1980s, snow ablation has significantly increased in the Pyrenees (Bonsoms et al., 2022), and winter snow days and snow accumulation have non-statically significantly increased (Buisan et al., 2016; Serrano-Notivoli et al., 2018; López-Moreno et al., 2020a; Bonsoms et al., 2021a) due to the increasing frequency of positive west and south-west advections (Buisan et al., 2016).

Climate projections for Pyrenees during the mid-late 21st century anticipate a temperature increase of more than 1°C to 4ºC (relative to 1986–2005), and an increase (decrease) of precipitation by about 10% for the eastern (western) regions during winter and spring (Amblar-Frances et al., 2020). Therefore, changes in snow patterns in regions with high elevations are uncertain because winter snow accumulation is affected by precipitation (López-Moreno et al., 2008) and Mediterranean basin winter precipitation projections have uncertainties up to 80% of the total variance (Evin et al., 2021).

Previous studies in the central Pyrenees (López-Moreno et al., 2013), Iberian Peninsula Mountain ranges (Alonso-González et al., 2020a), and mountain areas that have Mediterranean climates (López-Moreno et al., 2017) demonstrated that snowpack sensitivity is mostly controlled by elevation. Despite the impact of climate warming in mountain hydrological processes, there is limited understanding of the snow sensitivity and seasonality of mid-latitude Mediterranean mountain snowpacks. Some studies reported different snowpack sensitivities during wet and dry years (López-Moreno et al., 2017; Musselman et al., 2017b; Rasouli et al., 2022; Roche et al., 2018). However, the sensitivity of snow during periods when there are seasonal extremes of temperature and precipitation has not yet been analyzed. The high interannual variability of the Pyrenean snowpack, which is expected to increase according to climate projections (López-Moreno et al., 2008), indicates a need to examine snowpack sensitivity focusing on the year-to-year variability, especially during the warm seasons in the Mediterranean basin (Vogel et al., 2019; De Luca et al., 2020) because these are likely to increase in the future (e.g., Meng et al., 2022). Further, the occurrence of different HS trends at mid- and high-elevation areas of this range (López-Moreno et al., 2020a) suggest that elevation and spatial factors contribute to the wide variations of the sensitivity of snow to the climate.

Therefore, the main objective of this research is to quantify snow accumulation, ablation, and timing sensitivity due to climate change during compound temperature and precipitation extremes seasons in the Pyrenees.

2 Geographical area and climate setting

The Pyrenees is a mountain range located in the north of the Iberian Peninsula (south Europe; 42ºN-43ºN to 2ºW-3ºE) that is aligned east-to-west between the Atlantic Ocean and the Mediterranean Sea. The highest elevations are in the central region (Aneto, 3404 m) and elevations decrease towards the west and east (Figure 1). The Mediterranean basin, including the

Pyrenees, is located in a transition area, and is influenced by the continental climate and the subtropical temperate climate. Precipitation is mostly driven by large-scale circulation patterns (Zappa et al., 2015; Borgli et al., 2019), the jet-stream oscillation during winter (Hurell, 1995), and land-sea temperature differences (Tuel and Eltahir, 2020). During the summer, the northward movement of the Azores high pressure region brings stable weather, and precipitation is mainly convective at that time (Xercavins, 1985). Precipitation is highly variable depending on mountain exposure to the main circulation weather types; it ranges from about 1000 mm/year to about 2000 mm/year (in the mountain summits), with lower levels in the east and south (Cuadrat et al., 2007). There is a slight disconnection of the general climate circulation towards the eastern Pyrenees, where the Mediterranean climate and East Atlantic/West Russia (EA-WR) oscillations have greater effects on snow accumulation (Bonsoms et al., 2021a). In the southern, western, and central massifs of the range, the Atlantic climate and the negative North Atlantic Oscillation (NAO) phases regulate snow accumulation (W and SW wet air flows; López-Moreno, 2005; López-Moreno and Vicente-Serrano, 2007; Buisan et al., 2016; Alonso-González et al., 2020b). In the northern slopes, the positive phases of the Western Mediterranean Oscillation (WeMO) linked with NW and N advections trigger the most episodes of snow accumulation (Bonsoms et al., 2021a). The seasonal snow accumulation in the northern slopes is almost double the amount (about 500 cm more) as in the southern slopes at an elevation of about 2000 m (Bonsoms et al., 2021a). The temperature/elevation gradient is about 0.55ºC/100 m (Navarro-Serrano and López-Moreno, 2018) and the annual 0ºC isotherm is at about 2750 to 2950 m (López-Moreno and García-Ruiz, 2004). Net radiation and latent heat flux governs the energy available for snow ablation; the former increases at high elevations and the latter towards east (Bonsoms et al., 2022).

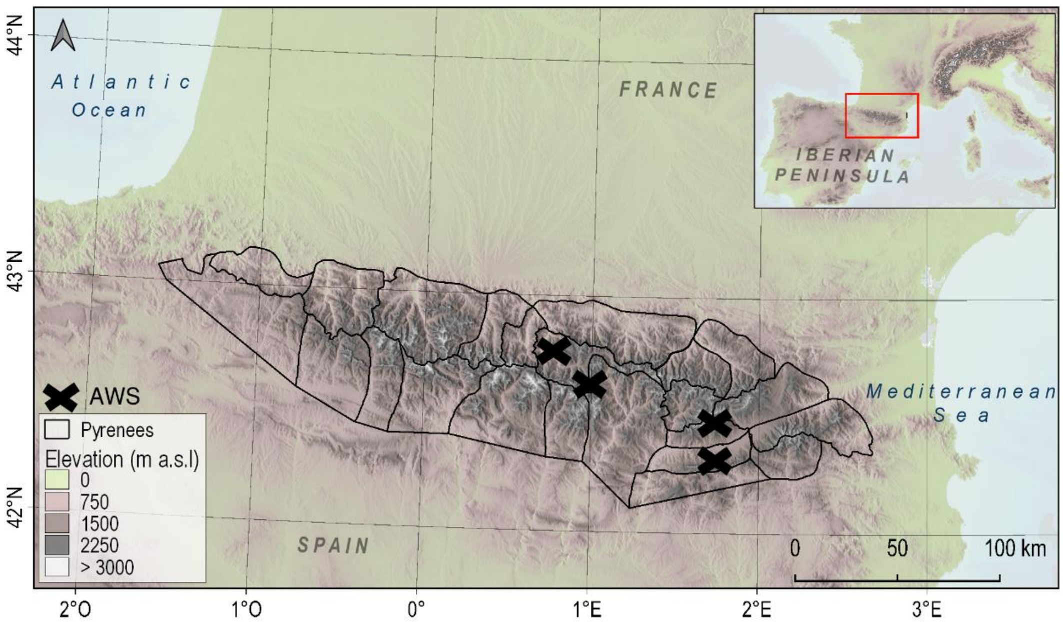

Figure 1. Locations of automatic weather stations (AWS) and geography of the Pyrenean massifs based on the spatial regionalization of the SAFRAN system, which groups massifs according to topographical and meteorological characteristics (modified from Durand et al., 1999).

3 Data and methods

3.1 Snow model

Snowpack was modelled using a physical-based snow model, the Flexible Snow Model (FSM2; Essery, 2015). This model resolves the SEB and mass balance to simulate the state of the snowpack. Previous studies tested the FSM2 (Krinner et al., 2018), and its application in different forest environments (Mazzoti et al., 2021), and hydro-climatological mountain zones such the Andes (Urrutia et al., 2019), Alps (Mazzoti et al., 2020), Colorado (Smyth et al., 2022), Himalayas (Pritchard et al., 2020), Iberian Peninsula Mountains (Alonso-González et al., 2020a; Alonso-González et al., 2022), Lebanese mountains (Alonso-González et al., 2021), providing confidential results. The FSM2 requires forcing data of precipitation, air temperature, relative humidity, surface atmospheric pressure, wind speed, incoming shortwave radiation (SW_{inc}), and incoming long wave radiation (LW_{inc}). We have evaluated different FSM2 model configurations (not shown) without significant differences in the accuracy and performance metrics. Therefore, we selected the most complex FSM2 configuration, except for the snow cover fraction estimation, that is based on a linear function of HS. In detail, albedo is calculated based on a prognostic function, with increases due to snowfall and decreases due to snow age. Atmospheric stability is calculated as function of the Richardson number. Snow density is calculated as a function of viscous compaction by overburden and thermal metamorphism. Snow hydrology is estimated by gravitational drainage, including internal snowpack processes, runoff, refreeze rates, and thermal conductivity.

3.2 Snow model validation

FSM2 configuration was validated by in situ snow records of four automatic weather stations (AWSs) that were at high elevations in the Pyrenees. Precipitation in mountainous and windy regions is usually affected by undercatch (Kochendorfer et al., 2020). Thus, the instrumental records of precipitation were corrected for undercatch by applying an empirical equation validated for the Pyrenees (Buisan et al., 2019). Precipitation type was classified by a threshold method (Musselman et al., 2017b; Corripio et al., 2017): snow when the air temperature was below 1°C and rain when the air temperature was above 1°C, according to previous research in the study area (Corripio et al., 2017). The LW_{inc} heat flux of the AWSs (Table 1) were estimated as previously described (Corripio et al., 2017). Due to the wide instrumental data coverage (99.3% of the total dataset), gap-filling was not performed. The HS records were measured each 30 min using an ultrasonic snow depth sensor. The meteorological data used in the validation process were provided and managed by the local meteorological service of Catalonia (https://www.meteo.cat/wpweb/serveis/formularis/peticio-dinformes-i-dades-meteorologiques/peticio-de-dades-meteorologiques/; data requested: 14/01/2021). Quality-checking of the data was performed using an automatic error filtering process in combination with a climatological, spatial, and internal coherency control defined by the SMC (2011).

Table 1. Characteristics of the four AWSs.

Area	Code	Lat/Lon°	Elevation (m)	Atlantic Ocean, Distance (km)	Mediterranean Sea, Distance (km)	Validation period (years)	Years
Central-Pyrenees, Northern slopes	A1	42.77/0.73	2228	200	190	2004–2020	16
	A2	42.61/0.98	2266	225	170	2001–2020	19
Eastern Pyrenees, Southern slopes	A3	42.46/1.78	2230	295	115	2005–2020	15
Eastern Pre-Pyrenees, Northern slopes	A4	42.29/1.71	2143	300	110	2009–2020	11

Model accuracy was estimated based on the mean absolute error (MAE) and the root mean square error (RMSE), and model performance was estimated by the coefficient of determination (R^2). The MAE and the RMSE indicate the mean differences of the modelled and observed values.

3.3 Atmospheric forcing data

We forced the FSM2 with the reanalysis climate dataset of Vernay et al. (2021), which consists of the modelled values from the SAFRAN meteorological analysis. The FSM2 was run at an hourly resolution for each massif, each elevation range, and each climate baseline and perturbation scenario from 1980 to 2019. The SAFRAN system provides data for homogeneous meteorological and topographical mountain massifs every 300 m, from 0 to 3600 m (Durand et al., 1999; Vernay et al., 2021). We analyzed three elevation bands: low (1500 m), middle (1800 m), and high (2400 m). Precipitation type was classified using the same threshold approach used for model validation. Atmospheric emissivity was derived from the SAFRAN LW_{inc} and air temperature. SAFRAN was forced using numerical weather prediction models (ERA-40 reanalysis data from 1958 to 2002 and ARPEGE from 2002 to 2020). Meteorological data were calibrated, homogenized, and improved by in situ meteorological observations data assimilation

(Vernay et al., 2021). Durand et al. (1999; 2009a; 2009b) provided further technical details of the SAFRAN system. Previous studies used the SAFRAN system for the long-term HS trends (López-Moreno et al., 2020), extreme snowfall (Roux et al., 2021), and snow ablation analysis (Bonsoms et al., 2022). SAFRAN system has been extensively validated for the meteorological modelling of continental Spain (Quintana-Seguí et al., 2017), France (Vidal et al., 2010) or alpine snowpack climate projections (Verfaille et al., 2018), among other works.

3.4 Snow sensitivity analysis

Snow sensitivity was analyzed using a delta-change methodology (López-Moreno et al., 2008; Beniston et al., 2016; Musselman et al., 2017b; Marty et al., 2017; Alonso-González et al., 2020a; Sanmiguel-Vallelado et al., 2022). In this method, temperature and precipitation were perturbed for each massif and elevation range based the historical period (1980–2019). Temperature was increased from 1 to 4ºC at 1ºC intervals, assuming an increase of LW_{inc} accordingly (Jennings and Molotch, 2020). Precipitation was changed from −10% to +10% at 10% intervals, in accordance with climate model uncertainties and the maximum and minimum precipitation projections for the Pyrenees (Amblar-Frances et al., 2020).

3.5 Definitions of compound temperature and precipitation extreme seasons

The snow season was from October 1 to June 30 (inclusive). Snow duration was defined by snow onset and snow ablation dates in situ observations (Bonsoms, 2021a), and results from the baseline scenario snow duration presented in this work. A “compound temperature and precipitation extreme season” (season type) was assessed based on each massif and the elevation historical climate record (1980–2019) using a joint quantile approach (Beniston and Goyette, 2007; Beniston, 2009; López-Moreno et al., 2011a). Season types were defined according to López-Moreno et al. (2011a), based on the seasonal 40^{th} percentiles (T40 for temperature and P40 for precipitation) and the seasonal 60^{th} percentiles (T60 and P60). There were four types of extreme seasons based on seasonal temperature (Tseason) and seasonal precipitation (Pseason) data:

Cold and Dry (CD): Tseason ≤ T40 and Pseason ≤ P40;

Cold and Wet (CW): Tseason ≤ T40 and Pseason ≥ P60;

Warm and Dry (WD): Tseason > T40 and Pseason ≤ P40;

Warm and Wet (WW): Tseason > T60 and Pseason is > P60.

All remaining seasons were classified as having average (Avg) temperature and precipitation. Figure S1 shows the number of season types by elevation and massifs.

3.6 Snow-climatological indicators

Snowpack sensitivity was analyzed using five key indicators: (*i*) seasonal average HS, (*ii*) seasonal maximum absolute HS peak (peak HS max), (*iii*) date of the maximum HS (peak HS date), (*iv*) number of days with HS > 1 cm on the ground (snow duration), and (*v*) daily average snow ablation per season (snow ablation). Snow ablation was calculated as the difference between the maximum daily HS recorded on two consecutive days (Musselman et al., 2017a), and only days with decreases of 1 cm or more were recorded. Seasonal HS and peak HS max are snow accumulation indicators; snow ablation, snow duration, and peak HS date are snow ablation indicators. All indicators were computed according to massif and elevation range.

4. Results

4.1 Snow model validation

Our snow model validation analysis (Figures 2 and 3) confirmed that FSM2 accurately reproduces the observed HS values. On average, the FSM2 had a R^2 greater than 0.83 for all stations. In general, the snow model slightly overestimated the maximum HS values. The highest R^2 values were at A4 and A2 (R^2 = 0.85 in both stations), and the lowest were at A3 and A1 (R^2 = 0.79 and R^2 = 0.82, respectively). The highest accuracy was at A4 (RMSE = 18.5 cm, MAE = 8.9 cm), and the largest errors were at A2 (RMSE = 45.8 cm, MAE = 29.0 cm).

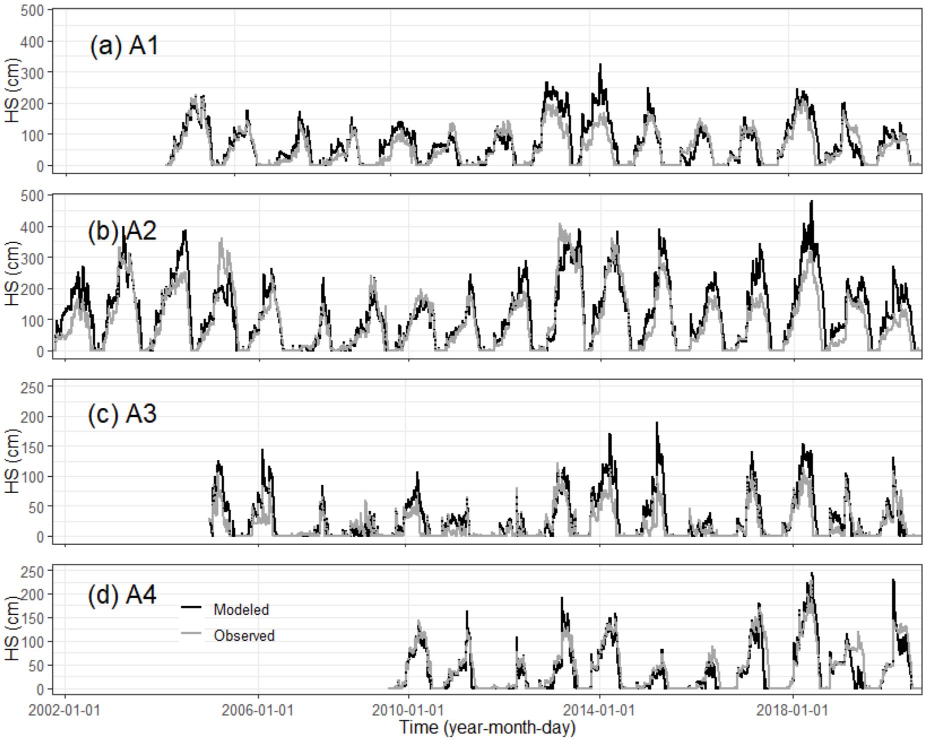

Figure 2. Time series of the observed (gray) and modelled (black) HS values at the four AWSs.

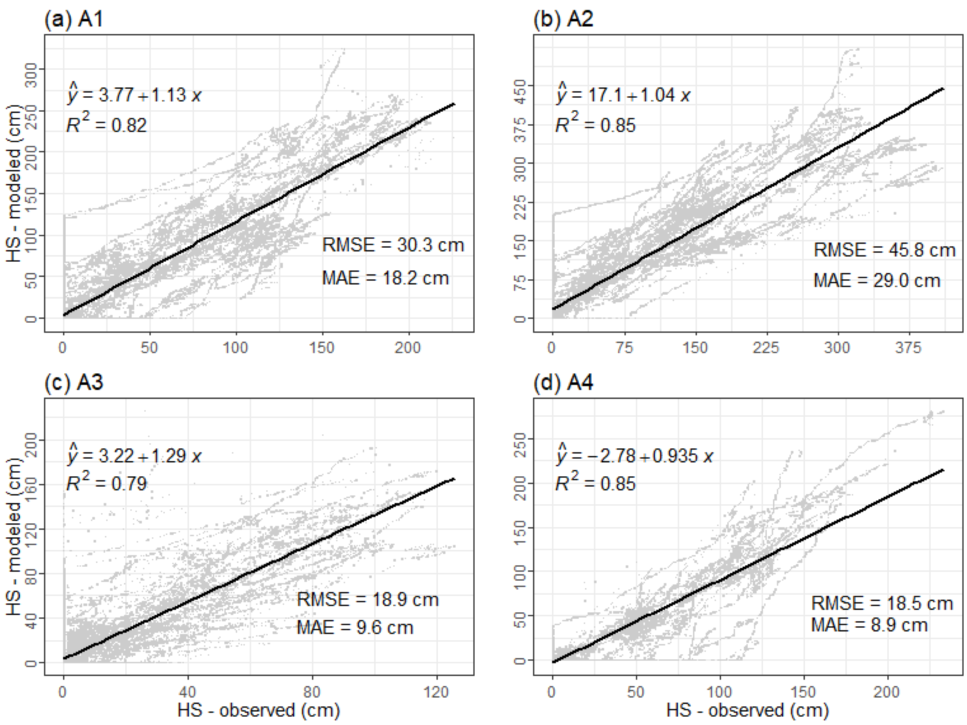

Figure 3. Regression analysis of observed (x-axis) and simulated (y-axis) HS values.

4.2 Snow sensitivity

We then determined seasonal HS profiles for each perturbed climate scenario and compound temperature and precipitation extreme season (Figure 4). The results show a non-linear response between seasonal HS loss and temperature increase. When the temperature increased at 1ºC intervals, the largest relative seasonal HS decrease from the baseline was at + 1ºC for all elevations and all compound temperature and precipitation extreme seasons. High elevation areas had lower season-to-season snow variability than low elevations for all season types (Figure 4). At low elevations, snow was significantly greater during CW seasons than other seasons. All the snowpack-perturbed scenarios indicated that snowpack decreased at low and mid elevations under warming climate scenarios. Snowpack sensitivity depended on the season type (Figures 5 and 6). At low elevations, the seasonal changes in HS ranged from −37% (WW) to −28% (CD) per ºC increase. For mid-elevation ranges, there were no significant differences among season types (Table 2), and the seasonal HS changes ranged from −34% (WW) to −30% (CW) per ºC increase. Low and mid-elevations had greater snowpack reductions than high elevations. In the latter, a 10% increase of precipitation counterbalanced a temperature increase of about 1ºC, and there were no remarkable differences in the seasonal HS from the baseline scenario especially in the coldest months of the season (Figure S2, Figure S3 and Figure S4). The maximum seasonal HS was during WD seasons (27%/1ºC), and the minimum was during CW seasons (−22%/ºC).

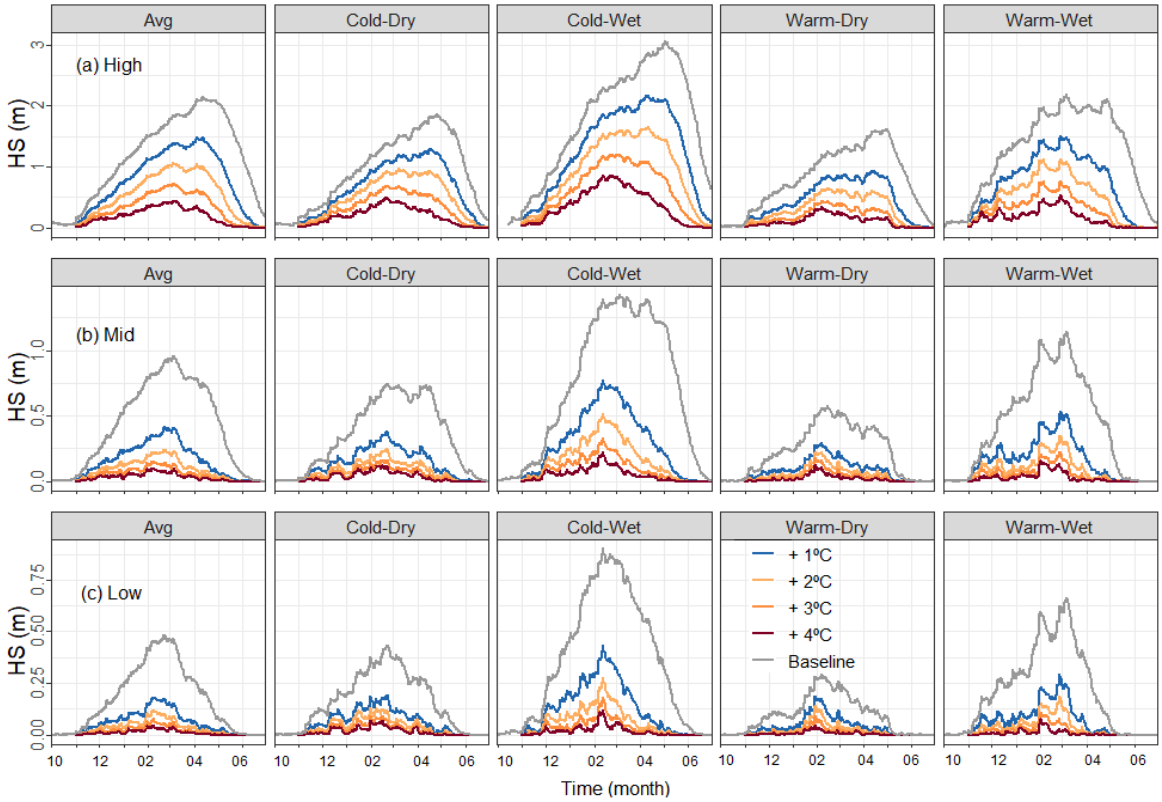

Figure 4. Average daily values for elevation, season type, baseline climate and different temperature increases for (a) high (b) mid and (c) low elevation.

Table 2. Average and seasonal HS and peak HS sensitivity during the four different compound temperature and precipitation extreme seasons at three different elevations.

Compound extreme season	**%HS/ ºC**			**%peak HS max/ºC**		
	Low	**Mid**	**High**	**Low**	**Mid**	**High**
Avg.	−33	−33	−25	−20	−20	−16
CD	−28	−30	−22	−17	−17	−14
CW	−33	−32	−22	−22	−20	−15
WD	−32	−30	−27	−19	−16	−16
WW	−37	−34	−26	−24	−24	−16

At low and mid elevations, the peak HS max was greatest during WW seasons (−24%/ºC) and lowest during the CD and WD seasons (−17%/ºC for both). At high elevations, there were no significant differences in the peak HS max for the different seasons. The maximum peak HS max was during WD seasons (−16%/ºC) and the minimum was during CD seasons (−14%/ºC).

We also determined average seasonal snow duration for each elevation range and season type for different temperature increases (Table 3 and Figure 6). The minimum snow duration was during CW seasons (−13%/ºC at low elevations, −10%/ºC at mid-elevations, −5%/ºC at high elevations). At low elevations, the snow duration was most sensitive during WW seasons (−17%/ºC). On the contrary, at mid-elevations and high elevations, the snow duration was most sensitive during WD seasons (−13%/ºC at mid-elevations and −8%/ºC at high elevations).

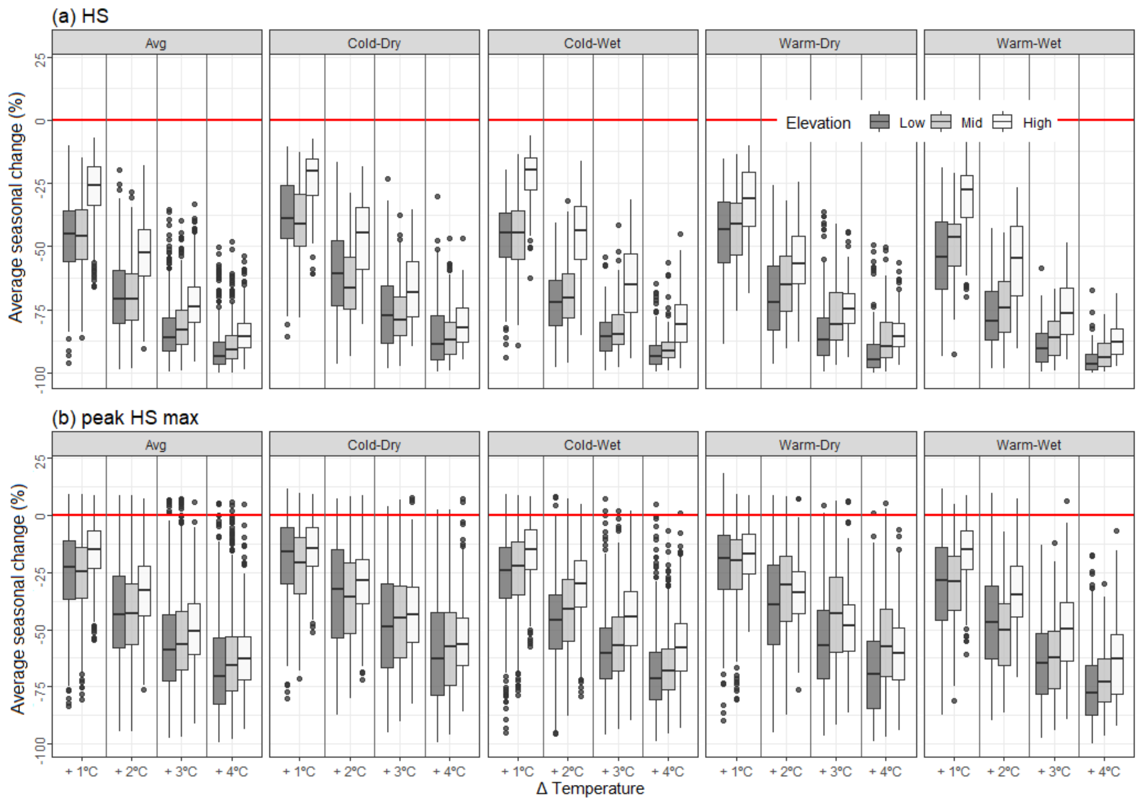

Figure 5. Anomalies of seasonal HS (a) and peak HS max (b) for different temperature increases relative to baseline at three different elevations during the four different temperature and precipitation extreme seasons. The solid black lines within each boxplot are the average. Lower and upper hinges correspond to the 25th and 75th percentiles, respectively. The whisker is a horizontal line at 1.5 interquartile range of the upper quartile and lower quartile, respectively. Dots represent the outliers. Data is grouped by season, season type, increment of temperature, precipitation variation, elevation, and massif.

Overall, the peak HS date occurred earlier due to warming, independently of precipitation changes. During WD seasons, the peak HS date per °C was earlier by 9 days at low elevations, 3 days at mid-elevations, and 17 days at high elevations; during CD seasons, the peak HS date per °C was earlier by 15 days at low elevations, 8 days at mid-elevations, and 1 day at high elevations. In high elevation areas, if the temperature increase was no more than about 1ºC above baseline, there was little change in the peak HS date (Figure 7), except during dry seasons, when the maximum peak HS date sensitivity was found. On the contrary, the peak HS date did not change significantly due to warming during WW seasons (Table 4), and because the snowpack would be scarce at those times, and there were no defined peaks (Figure 4).

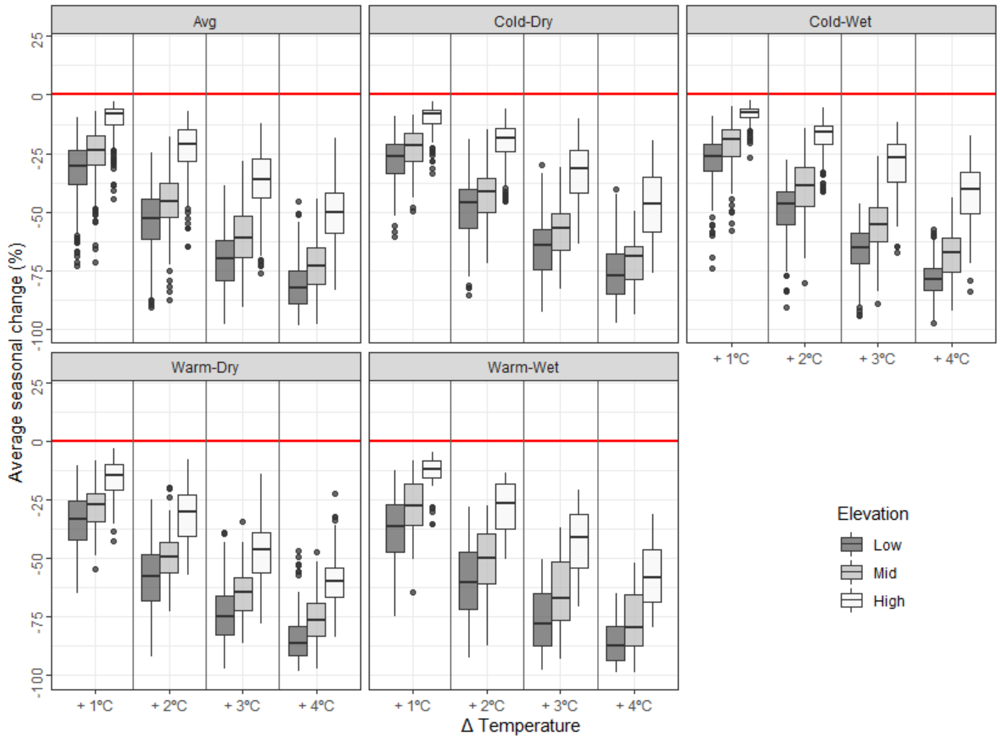

Figure 6. Average and seasonal decrease of snow duration at three different elevations, increments of temperature, and during the four different temperature and precipitation extreme seasons. The solid black lines within each boxplot are the average. Lower and upper hinges correspond to the 25th and 75th percentiles, respectively. The whisker is a horizontal line at 1.5 interquartile range of the upper quartile and lower quartile, respectively. Dots represent the outliers. Data is grouped by season, season type, increment of temperature, precipitation variation, elevation, and massif.

We determined average daily snow ablation in response to different temperature increases at different elevations and during different extreme seasons (Figure 8). The results show there were moderate differences in the average daily snow ablation in a warmer climate. At low elevations, the snow ablation in all four extreme seasons was 12%/ºC (Table 4). At mid-elevations and high elevations, the maximum snow ablation was during dry seasons; WD seasons had a snow ablation of 13%/°C at mid-elevations and 10%/ºC at high elevations. On the other hand, the minimum values for mid-elevations were during WW seasons, when the snow ablation was 8%/ºC; the minimum values at high elevations were during CW seasons, when snow ablation was 5%/°C.

Figure 7. Difference (days) from baseline Peak HS date at three different elevations and during the four different temperature (colors) and precipitation shifts (columns) for each season (boxes). Error bars indicate the maximum and minimum values (mean by massifs and elevation).

Table 4. Snow duration, snow ablation, and peak HS date sensitivity during the four different compound temperature and precipitation extreme seasons.

Extreme season	**Snow duration (%/ºC)**			**Snow ablation (%/ºC)**			**Peak HS date (days/ºC)**		
	Low	Mid	High	Low	Mid	High	Low	Mid	High
Avg.	−15	−12	−6	12	11	7	−2	1	−4
CD	−13	−11	−5	12	13	8	−15	−8	−1
CW	−13	−10	−5	12	10	5	−3	−1	4
WD	−16	−13	−8	12	13	10	−9	−3	−17
WW	−17	−13	−7	12	8	7	−5	8	0

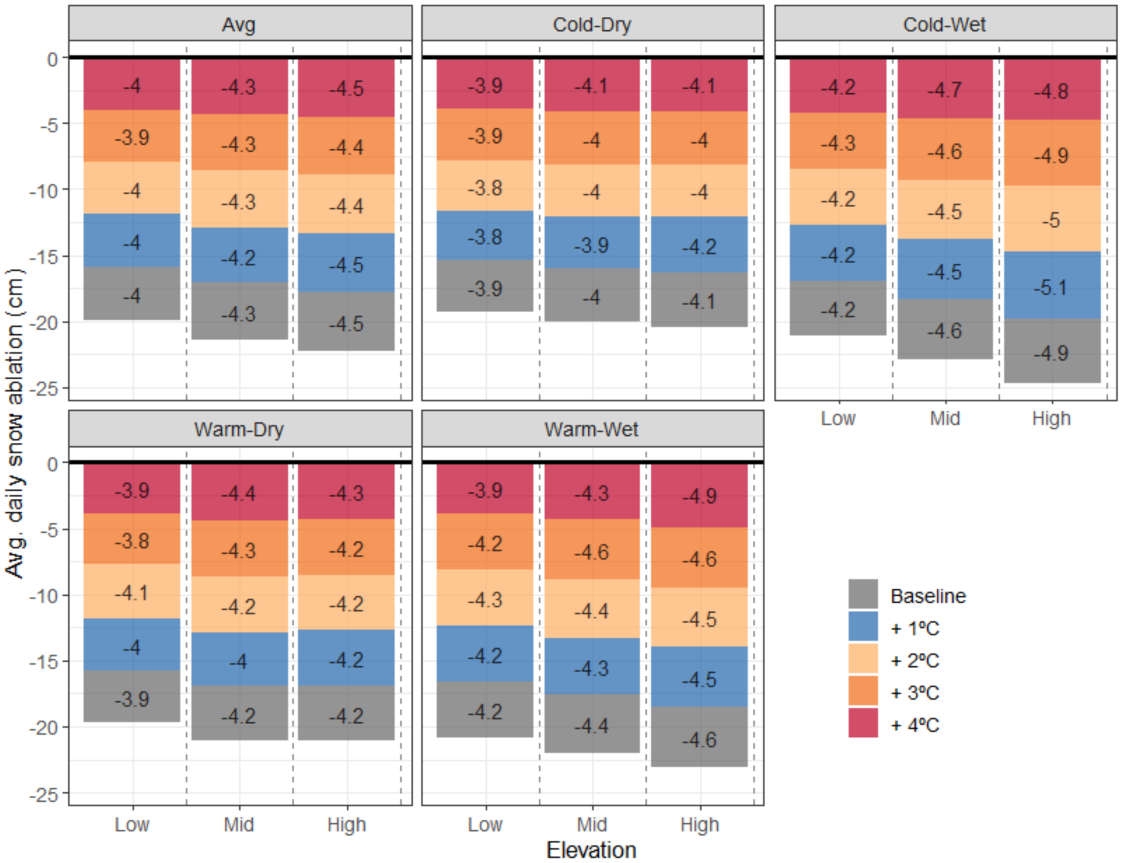

Figure 8. Average daily snow ablation at three different elevations during four different compound temperature and precipitation extreme seasons for different temperature increases above baseline (gray).

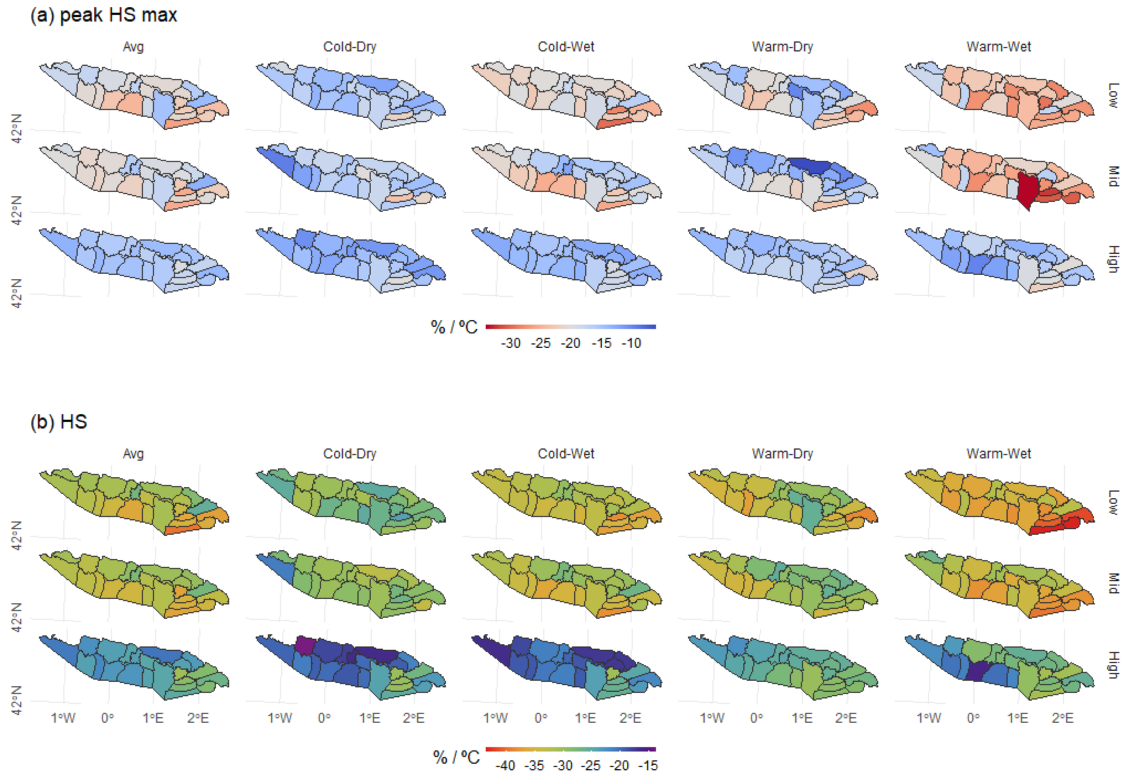

Figure 9. Geographical distribution of seasonal (a) peak HS and (b) HS sensitivity during the four different compound temperature and precipitation extreme seasons.

The sensitivity of HS to climate change also had remarkable spatial patterns. The sensitivity was greater in the eastern massifs, independently of the elevation range and season type (Figures 9 and 10). The greatest sensitivities of HS were at low elevations. Here, the seasonal HS ranged from −20%/°C during CD in the central area to −40%/°C during WW in the southern slopes of the eastern Pyrenees. Similarly, the maximum sensitivity of peak HS max was at mid-elevations in the southern slopes of the eastern Pyrenees, and the change was less than −35%/°C (Figure 9). There was a general tendency for higher sensitivity in the southern slopes than in the northern slopes. Snow duration in the northern slopes and at high elevations had the lowest sensitivity, especially during CD and CW seasons (−5%/ºC). The sensitivity of snow duration increased during WW seasons, and the maximum changes were at the lowest elevations of the southern-eastern Pyrenees (−35%/ºC; Figure 10).

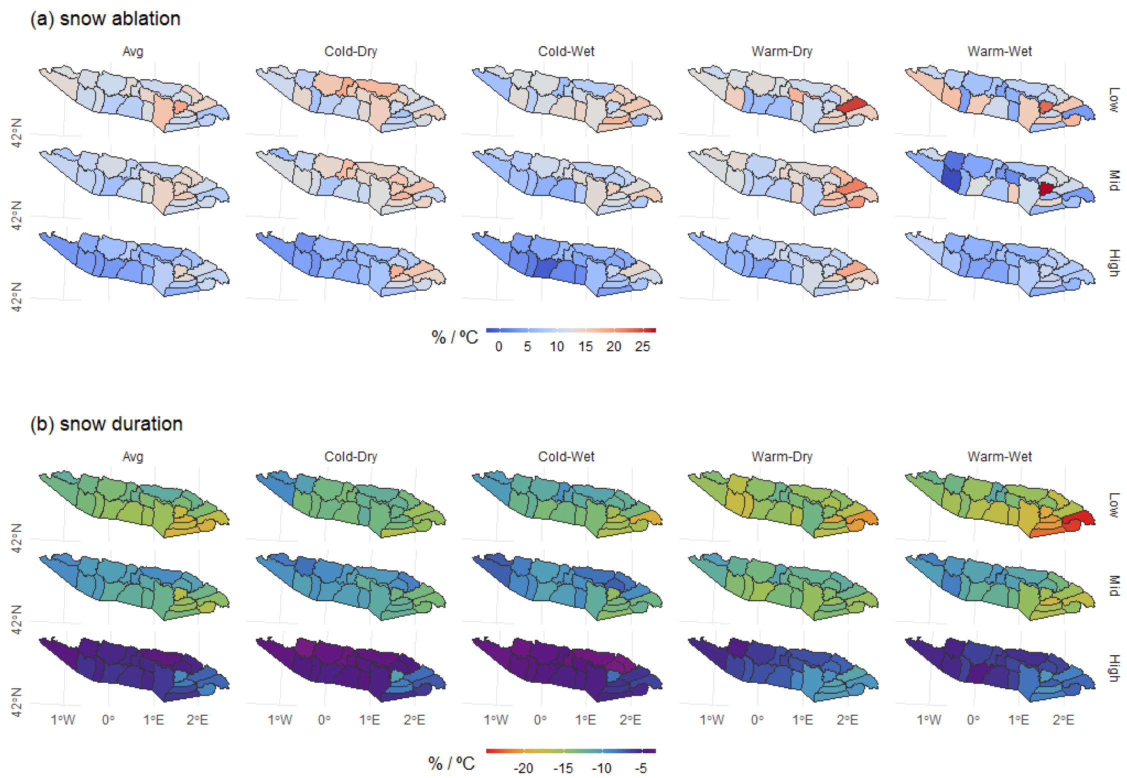

Figure 10. Geographical distribution of seasonal (a) snow ablation and (b) snow duration sensitivity during the four different compound temperature and precipitation extreme seasons.

5. Discussion

The spatial and temporal patterns of snow in the Pyrenees are highly variable, and climate projections indicate that extreme events will likely increase during future decades (Meng et al., 2022). Therefore, we analyzed factors that affect the snowpack sensitivity to gain insight into how future climate changes may affect the snow regime.

5.1 Spatial and elevation factors controlling snow sensitivity to climate change

The sensitivity of snow to different spatial patterns of climate change that we identified here (Figures 9 and 10) are consistent with the snow accumulation and ablation patterns previously reported in this region (Lopez-Moreno, 2005; Navarro-Serrano et al., 2018; Alonso-González et al., 2020a; Bonsoms et al., 2021a). The Atlantic climate has less of an influence in the eastern massifs; in this area, in situ observations indicated there was about half of the seasonal snow accumulation amounts as in northern and western areas at the same elevation (>2000 m; Bonsoms et al., 2021a). The snow in the southern slopes of the eastern Pyrenees is more sensitive to climate change because these massifs are exposed to higher turbulence and radiative heat fluxes (Bonsoms et al., 2022). The results thus show an upward displacement of the snow line due to warming. Previous studies described the sensitivity of the snow pattern to elevation at specific stations of the central Pyrenees (López-Moreno et al., 2013; 2017), Iberian Peninsula mountains (Alonso-González et al., 2020a), and other ranges such as the Cascades (Jefferson, 2011; Sproles et al., 2013), the Alps (Marty et al., 2017), and western USA (Pierce et al., 2013; Musselman et al., 2017b). In these regions, the models suggest larger snowpack reductions due to warming at subalpine sites than at alpine sites (Jennings and Molotch, 2020; Mote et al., 2018). Low elevations are more sensitive simply because the temperatures there are closer to isothermal conditions (Brown and Mote, 2009; Lopez-Moreno et al., 2017).

5.2 Snow sensitivity to climate change and relationship with historical and future snow trends

5.2.1 Snow accumulation phase

The snow losses due warming that we described here are mainly associated with increases in the rain/snowfall ratio (Figure S4), changes in the snow onset and offset dates (Figure 4), and increases in the energy available for snow ablation during the later months of the snow season, as it was previously reported by literature (e.g., Pomeroy et al., 2015; Lynn et al., 2020; Jennings and Molotch, 2020). At high elevations, a trend of increasing precipitation (+10%) could counterbalance temperature increases (<1ºC; Figure S2), consistent with the results previously reported for specific sites of the central Pyrenees (Izas, 2000m; López-Moreno et al., 2008). Rasouli et al. (2014) also found that a 20% increase of precipitation could compensate for 2ºC increase of temperature in subarctic Canada. A climate sensitivity analysis in the western Cascades (western USA) found that increases of precipitation due to warming modulated the snowpack accumulation losses by about 5%/1°C (Minder, 2010). These results are consistent with recent data that examined snow above 1000 m in the Pyrenees, which found that an increase in the frequency of west circulation weather types since the 1980s increased the HS (Serrano-Notivoli et al., 2018; López-Moreno et al., 2020), snow accumulation (Bonsoms et al., 2021a), and changes in winter snow days (Buisan et al., 2016). There are similar trends in the Alps, with an increase of extreme (exceeding the 100-year return level) snowfall above 3000 m during recent decades (Roux et al., 2021) and increases in extreme winter precipitation (Rajczak and Schär, 2017).

5.2.2 Snow ablation phase

Climate warming leads to a cascade of physical changes in the SEB that increase snow ablation near the 0ºC isotherm. On overall, the average daily snow ablation showed moderate to low changes due to warming. Comparison between low and high elevations indicated slightly faster snow ablation at high elevations (Figure 8). This higher rate of snow ablation per season at high elevations (which have deeper snowpacks) are probably because the snow there lasts until late spring, when more energy is available for snow ablation (Bonsoms et al., 2022). Temperature increase does not imply significant changes in the daily snow ablation rate per season because warming decreases the magnitude of the snowpack (seasonal HS and peak HS max) and triggers an earlier onset of snowmelt (Wu et al., 2018). The earlier peak HS date at low and mid elevations (Table 4 and Figure 7) implies lower rates of net shortwave radiation, because snow melting starts earlier in warmer climates (Pomeroy et al., 2015), coinciding with the shorter days and lower solar zenith angle (Lundquist et al., 2013; Sanmiguel-Vallelado et al., 2022). Our results agree with the slow snow melt rates reported in the Northern Hemisphere from 1980 to 2017 (Wu et al., 2018). The results of previous studies were similar for subarctic Canada (Rasouli et al., 2014) and western USA snowpacks (Musselman et al., 2017b), but Arctic sites had faster melt rates (Krogh and Pomeroy, 2019).

5.2.3 Snow sensitivity and snowpack projections

Our results suggest that warming had a non-linear effect on snowpack reduction. Our largest snow losses were for seasonal HS when the temperature increased by 1ºC above baseline. At low and mid elevations, the average seasonal HS decrease was more than 40% for all season types, and the maximum sensitivity was during WW seasons. Previous research in the Pyrenees and other mid-latitude mountain ranges reported similar results. A study in the central Pyrenees reported the peak SWE was 29%/ºC, whereas snow season duration decreased by about 20 to 30 days at about 2000 m (López-Moreno et al., 2013). The average peak HS max at high elevations in the Pyrenees (−16 %/ºC; Figure 6 and Table 2), was similar to the average peak SWE sensitivity (−15%/°C) reported in the Iberian Peninsula mountains at 2500 m (Alonso-González et al., 2020a). These results are also consistent with climate projections for this mountain range. In particular, for a 2ºC or more increase of temperature, the snow season declined by 38% at the lowest ski resorts (~1500 m) in the southern slopes of the eastern Pyrenees (Pons et al., 2015). However, high emission climate scenarios projected an increase in the frequency and intensity of high snowfall at high elevations (López-Moreno et al., 2011b). Climate projections for the mid-late 21st century indicated that sensitivity in the easternmost areas could decline during the winter because of a trend for an increase of about 10% in precipitation in this area (Amblar-Francés et al., 2020). Our projected changes in the Pyrenean snowpack dynamics are similar to the expected snow losses in other mountain ranges. For example, a study of the Atlas Mountains of northern Africa concluded that snowpack decreases were greater in the lowlands and projected seasonal SWE declines of 60% under the RCP4.5 scenario and 80% under the RCP8.5 scenario for the entire range (Tuel et al., 2022). A study in the Washington Cascades (western USA) found that snowpack decline was 19 to 23% per 1ºC (Minder, 2010), similar to the values in the present study at high elevations. A study of the French Alps (Chartreuse, 1500 m) found that seasonal HS decreases on the order of 25% for a 1.5ºC increase and 32% for a 2ºC increase of global temperature above the pre-industrial years (Verfaille et al., 2018). A study of the Swiss Alps reported a snowpack decrease of about 15%/ºC (Beniston, 2003); in the same alpine country, another study predicted the seasonal HS will decrease by more than 70% in massifs below 1000 m in all future climate projections (Marty et al., 2017). The largest snow reductions will likely occur during the periods between seasons (Steger et al., 2013; Marty et al., 2017). Nevertheless, at high elevations, snow climate projections found no significant trend for maximum HS during the mid-late 21st century above 2500 m in the eastern Alps (Willibald et al., 2021), suggesting that internal climate variability is a major source of uncertainty of SWE projections at high elevations (Schirmer et al., 2021).

5.3 Influence of compound temperature and precipitation extreme seasons

We found that the maximum sensitivities of seasonal HS and peak HS max were during WW seasons at low and mid-elevations and during WD seasons at high elevations. Brown and Mote (2008) analyzed the sensitivity of snow to climate changes in the Northern Hemisphere and found maximal SWE sensitivities in mid-latitudinal maritime winter climate areas, and minimal SWE sensitivities in dry and continental zones, consistent with our results. López-Moreno et al. (2017) also found greater decreases of SWE in wet and temperate Mediterranean ranges than in drier regions. Furthermore, Rasouli et al. (2022) studied the northern North American Cordillera and found higher snowpack sensitivities in wet basins than dry basins. Our maximum snow ablation and peak HS date occurred during dry seasons, in accordance with Musselman et al. (2017b), who found a higher snowmelt rate during dry years in the western USA. Low and mid-elevations are highly sensitive to WW seasons because wet conditions favor decreases in the seasonal HS due to advection from sensible heat fluxes. The temperature in the Pyrenees is still cold enough to allow snowfall at high elevations during WW seasons, and for this reason we found maximal sensitivities during WD seasons. Reductions of snowfall in alpine regions can be compensated in a warmer scenario, because warm and wet snow is less susceptible to blowing wind transport and losses from sublimation (Pomeroy and Li, 2000; Pomeroy et al., 2015). During spring, snow runoff could be also greater in wet climates due to rain-on-snow events (Corripio et al., 2017), coinciding with the availability of more energy for snow ablation.

5.4 Environmental and socioeconomic implications

Our results indicated there will be an increase of snow ablation days and imply a disappearance of the typical sequence of snow accumulation and snow ablation seasons. Climate warming triggers the simultaneous occurrence of several periods of snowfall and melting, snow droughts during the winter, and ephemeral snowpacks between seasons. These expected decreases in snow will likely have significant impacts on the ecosystem. During spring, a snow cover cools the soil (Luetschg et al., 2008), delays the initiation of freezing (Oliva et al., 2014), functions as a thick active layer (Hrbáček et al., 2016), and protects alpine rocks from exposure to solar radiation and high air temperatures (Magnin et al., 2017). Due to warming temperatures, the remaining glaciers in the Pyrenees are shrinking and are expected to disappear before the 2050s (Vidaller et al., 2021). The shallower snowpack that we identified in this work will increase the vulnerability of glaciers, because snow has a higher albedo than dark ice and debris-covered glaciers and functions as a protective layer for glaciers (Fujita and Sakai, 2014).

The earlier onset of snowmelt suggested by our results, which is greater at low and mid-elevations during WD seasons, is in line with previous global studies that reported earlier streamflow due to earlier runoff dates (Adam et al., 2009; Stewart, 2009), and with a study of changes in the Iberian Peninsula River flows (Morán-Tejeda et al., 2014). Overall, our results are consistent with the slight decrease of the river peak flows that have occurred in the southern slopes of the Pyrenees since the 1980s (Sanmiguel-Vallelado et al., 2017). The significant reductions of seasonal HS that we identified, which are driven by increases in the rainfall ratio, suggest that snowmelt-dominated stream flows will likely shift to rainfall dominated regimes. Although high elevation meltwater might increase and contribute to earlier groundwater recharging (Evans et al., 2018), the increased evapotranspiration in the lowlands (Bonsoms et al., 2022) could counter this effect, so there is no net change in downstream areas (Stahl et al., 2010). Snow ephemerality triggers lower spring and summer flows (Barnett et al., 2005; Adam et al., 2009; Stahl et al., 2010) and has an impact on the hydrological management strategies. Winter snow accumulation affects hydrological availability during the months when water and hydroelectric demands are higher. This is because reservoirs store water during periods of peak flows (winter and spring), and release water during the driest season in the lowlands (summer) (Morán-Tejeda et al., 2014). Recurrent snow-scarce seasons may intensify these hydrological impacts and lead to competition for water resources among different ecological and socioeconomic systems. The economic viability of mountain ski-resorts in the Pyrenees depends on a regular deep snow cover (Gilaberte-Burdalo et al., 2014; Pons et al., 2015), but this is highly variable, especially at low and mid-elevations. The expected increase in snow-scarce seasons that we identified here is consistent with climate projections for this region, which suggest that no Pyrenean ski resorts will be viable under RCP 8.5 scenario by the end of the 21st century (Spandre et al., 2019).

5.5 Limitations and uncertainties

The meteorological input data that we used to model snow were estimated for flat slopes and the regionalization system we used was based on the SAFRAN system. According to this system, a mountain range is divided into massifs with homogeneous topography. The SAFRAN system has negative biases in shortwave radiation, a temperature precision of about 1 K, and biases in the accumulated monthly precipitation of about 20 kg/m^2 (Vernay et al., 2021). The snow model used in this work (FSM2) is a physics-based model of intermediate complexity, and the estimates of snow densification are simpler than those from more complex models of snowpack; however, a more complex model does not necessarily provide better performance in terms of snowpack and runoff estimation (Magnusson et al., 2015). Biases in the SAFRAN system and biases related to the FSM2 were minimal because we quantified relative changes between a modeled snow scenario (climate baseline) and several perturbed scenarios. Finally, our estimates of snow sensitivity were based on the delta-approach, which considers changes in temperature and precipitation based on climate projections for the Pyrenees (Amblar-Francés et al., 2020), but assumes that the snow patterns of the reference climate period will be constant over time.

6 Conclusions

Our study assessed the impact of climate change on the Pyrenean snowpack during seasons with extreme compound temperature and precipitation using a physical-based snow model that was forced by reanalysis data. We determined the snow sensitivity using five key indicators of snow accumulation and snow ablation. Our results indicated that elevation affected the snow sensitivity. In particular, snowpack losses were greatest during WW seasons at low and mid-elevations and were greatest during WD seasons at high elevations. The lowest snow sensitivity was at high elevations of the western and northern Pyrenees, and sensitivity increased at lower elevations and in the eastern and southern slopes. An increase of 1ºC at low and mid elevation regions led to a significant decrease in the seasonal HS and snow duration. However, at high elevations, precipitation plays a key role, and temperature is far from the isothermal 0ºC during the middle of winter. In this region, a 10% increase of precipitation, as suggested by many climate projections over the eastern regions of this range, could compensate for temperature increases on the order of about 1ºC. The impact of climate warming depends on the combination of temperature and precipitation during extreme seasons. Our analysis of seasonal HS indicated the greatest declines were during WW seasons and the smallest declines were during CD seasons. For snow duration, however, the highest (lowest) sensitivity is found during WD (CW) seasons. Similarly, the peak HS date and snow ablation had slightly greater climate sensitivities during dry seasons, in that snow ablation increased by about 10% and the peak HS date occurred about 10 days earlier per ºC at all elevations. Our findings thus provide evidence that the Pyrenean snowpack is highly sensitive to climate change, and suggest that the snowpacks of other mid-latitude mountain ranges may also show similar response to warming.

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Author contributions

JB analyzed the data and wrote the original draft. JB, JILM and EAG contributed in the manuscript design and draft editing. JB, JILM and EAG read and approved the final manuscript.

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