https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

Rain-on-snow response to a warmer Pyrenees

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Abstract. Climate warming is changing the magnitude, timing, and spatial patterns of mountain snowpacks. A warmer atmosphere may also lead to precipitation phase shifts, with decreased snowfall fraction (Sf). The combination of Sf and snowpack decreases directly affects the frequency and intensity of rain-on-snow (ROS) events, a common cause of flash-flood events in snow dominated regions. In this work we examine the ROS patterns and sensitivity to temperature and precipitation change (delta-change) in the Pyrenees using a physical-based snow model forced with reanalysis climate data perturbed following 21st century climate projections for this mountain range. ROS patterns are characteritzed by their frequency, rainfall quantity and snow ablation. The highest ROS fr for the baseline climate period (1980 – 2019) are found in South-West high-elevations sectors of the Pyrenees (17 days/year). Maximum ROS rain is detected in South-East mid-elevations areas (45 mm/day, autumn), whereas the highest ROS ablation is found in North-West high-elevations zones (- 10 cm/day, summer). When air temperature is increased from 1ºC to 4ºC, ROS rain and frequency increase at a constant rate during winter and early spring for all elevation zones. For the rest of the seasons, non-linear responses of the ROS frequency and ablation to warming are found. Overall, ROS frequency decreases in the shoulders of the season across eastern low-elevated zones due to snow cover depletion. However, ROS increases in cold, high-elevated zones where long-lasting snow cover exists until late spring. Similarly, warming triggers fast ROS ablation (+ 10% per ºC) during the coldest months of the season, high-elevations, and northern sectors where the deepest snow depths are found. On the contrary, slow, and non-changes in ROS ablation are expected for warm and marginal snowpacks. These results highlight the different ROS responses to warming across the mountain range, suggest similar ROS sensitivities in near mid-latitude zones, and will help anticipate future ROS impacts in hydrological, environmental, and socioeconomic mountain systems.

Keywords: Snow, Rain-on-snow, Climate warming, Snow sensitivity, Mountain snowpack, Pyrenees.

1 Introduction

Mountain snowpacks supply large hydrological resources to the lowlands (García-Ruiz et al., 2015; Viviroli et al., 2011), with important implications in the ecological (Wipf and Rixen, 2010), hydrological (Barnett, 2005;

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

Immerzeel et al., 2020) and socioeconomic systems by providing hydroelectricity (Beniston et al., 2018) or guaranteeing winter tourism activities (Spandre et al., 2019). Climate warming, however, is modifying mountain snowfall patterns (IPCC, 2022), through temperature-induced precipitation changes from snowfall to rainfall (Lynn et al., 2020), leading in some cases to rain-on-snow (ROS) events. The upward high-latitude temperature and precipitation trends (Bintanja and Andry, 2017) and mountain elevation-dependent warming (Pepin et al., 2022) will likely change future ROS frequency (ROS fr) in snow-dominated regions (López-Moreno et al., 2021). To date, research has been focused on the ROS predictability (Corripio and López-Moreno, 2017), detection and validation methods through remote sensing (Bartsch et al., 2010) and models (Serreze et al., 2021). Several works have examined ROS from the climatological point of view, by analyzing ROS spatial-temporal patterns for Alaska (Crawford et al., 2020), Japan (Ohba and Kawase, 2020), Norway (Pall et al., 2019; Mooney and Li, 2021) or the Iberian Peninsula mountains (Morán-Tejeda et al., 2019). ROS events have also been linked with Northern-Hemisphere and Arctic low-frequency climate modes of variability (Rennert et al., 2009; Cohen et al., 2015) as well as synoptic weather types (Ohba and Kawase, 2020). Further, several works in mountain catchments of Switzerland (Würzer et al., 2016), Germany (Garvelmann et al., 2014a), United-States (Marks et al., 1992), Canadian Rockies (Pomeroy et al., 2016) or Spain (Corripio and López-Moreno, 2017), have portioned the contribution of Surface Energy Balance (SEB) components during ROS events. ROS alters snow and soil conditions, since the liquid water percolation creates ice layers and could alter the snowpack stability (Rennert et al., 2009). In severe ROS events, water percolation reaches the ground, and the subsequent water freezing causes latent heat releases, leading to soil and permafrost warming (Westermann et al., 2011). Positive heat fluxes during ROS events enhance snow runoff (Corripio and López-Moreno, 2017), especially in warm and wet snowpacks (Würzer et al., 2016). ROS can also trigger a snow avalanche in mountain zones (Conway and Raymond, 1993), flash flood events (Surfleet and Tullos, 2013), impacts in tundra ecosystems (Hansen et al., 2013) and herbivore populations such as reindeers (Kohler and Aanes, 2004).

Different ROS fr trends have been found since the last half of the 20st century. In the western United-States and from 1949 to 2003 (Mccabe et al., 2007) found a general ROS fr decrease in low elevations but an increase in high elevations. Similarly, the analysis of six major German basins from 1990 to 2011, reveals an upward (downward) ROS fr trend during winter (spring) at low and high elevations (Freudiger et al., 2014). On the contrary, from 1979 to 2014, no winter ROS fr trends were found across the entire Northern-Hemisphere (Cohen et al., 2015). ROS projections for the end of the 21st century suggest a general ROS fr increase in cold regions. This is projected for Alaska (Bieniek et al., 2018), Norway (Mooney and Li, 2021), western United-States (Musselman et al., 2018), Canada (il Jeong and Sushama, 2018) or Japan (Ohba and Kawase, 2020). In European mid-latitude mountain ranges, such as the Alps, ROS fr is expected to increase (decrease) in high (low) elevation sectors (Beniston and Stoffel, 2016; Morán-Tejeda et al., 2016). López-Moreno et al. (2021) compared the ROS sensitivity to climate warming across 40 global basins and detected the highest ROS fr decreases in low-elevated and warm Mediterranean mountain sites. Despite the increasing understanding of ROS spatio-temporal past and future trends, little is known about the ROS sensitivity to climate warming

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

across southern European mountain ranges, such as the Pyrenees.

This work examines the ROS sensitivity to temperature and precipitation change (delta-change) for low (1500 m), mid (1800 m) and high (2400 m) elevations of the Pyrenees. ROS delta-change is analyzed using a physically based snow model, forced with reanalysis climate data perturbed according to 21^{st} century climate projections spread for range (Amblar-Francés et al., 2020). Previous studies in alpine zones have shown different ROS response to warming depending on the area and month of the season (e.g., Morán-Tejeda et al. 2016). For this reason, results are focused on these two factors. First, we analyze ROS drivers, namely height of snow (HS) and snowfall fraction (Sf) (López-Moreno et al., 2021), sensitivity to temperature and precipitation. Next, we examine ROS patterns and their response to warming by three key ROS indicators, namely:

(a) Number of ROS days for a season (ROS fr).
(b) Average rainfall quantity during a ROS day (ROS rain).
(c) Average daily snow ablation during a ROS day (ROS ablation).

The study area is presented in Section 2. Section 3 describes the data and methods. Section 4 presents the results. We finally discuss the anticipated ROS spatio-temporal changes, their socio-environmental impacts and hazards in Section 5.

2 Regional setting

The Pyrenees mountain range is located between the Atlantic Ocean (West) and the Mediterranean Sea (East), and is the largest (~ 450 km) mountain range of the Iberian Peninsula. Elevation increases towards the central massifs, where the highest peak is found (Aneto, 3,404 m asl). Glaciers expanded during the Little Ice Age and nowadays are located in the highest mountain summits (Vidaller et al., 2021). The regional annual 0 °C isotherm is at ca. 2700 m (Del Barrio et al., 1990), and at ca. 1600 m during the cold season (López-Moreno and Vicente-Serrano, 2011). The elevation lapse-rate is ca. 0.6º/100 m, being slightly lower during winter (Navarro-Serrano and López-Moreno, 2017). Annual precipitation is ca. 1000 mm/year (ca. 1500 m); maximum values are found in the northern-western massifs (around 2000 mm/year), decreasing towards the southern-eastern (SE) area (Lemus-Canovas et al., 2019). Precipitation is predominantly (> 90%) solid above 1600 m from November to May (López-Moreno, 2005). Due to the mountain alignement, relief configuration, and the distance to the Atlantic Ocean, seasonal snow accumulations in the northern slopes (ca. 500 cm/season), almost doubles the recorded in the SE area for the same elevation (ca. 2000 m) (Bonsoms et al., 2021b). In the western and central area of the southern slopes of the range (SW sector, Figure 1), snow accumulation is ruled by Atlantic wet and mild flows, which are linked with negative North Atlantic Oscillation (NAO) phases (SW and W synoptic weather types) (López-Moreno, 2005; Alonso-González et al., 2020b; Bonsoms et al., 2021a). Positive Western Mediterranean Oscillation (WeMO) phases (NW and NE synoptic weather types) control the

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

snow patterns in the northern-eastern (NE) slopes of the range (Bonsoms et al., 2021a). Snow ablation starts in February (May) in low (high) elevations.The energy available for snow ablation is controlled by net radiation (55 %, over the total), latent (32 %) and sensible (13 %) heat fluxes (Bonsoms et al., 2022a).

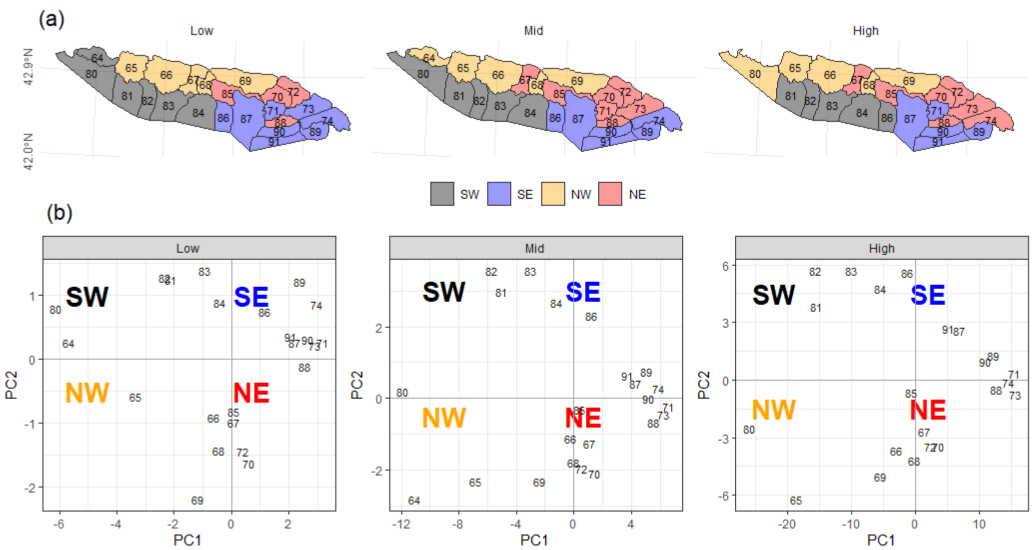

Figure 1. (a) Pyrenean massifs sectors (colors) for low, mid and high elevation. (b) Principle Component Analysis (PCA) scores of each massif for low, mid and high elevation. The black numbers are the SAFRAN massif's identity numbers defined by Vernay et al. (2022). Note that high elevation does not include massif number 64 since this massif does not reach 2400 m.

3 Data and methods

3.1 Snow model description

Snowpack is modeled using the energy and mass balance snow model FSM2 (Essery, 2015). The FSM2 was forced at hourly resolution for each massif and elevation range (c.f. Sect. 3.3) for the baseline climate (1980 – 2019) and several climate perturbed scenarios (c.f. Sect. 3.4). Sf was quantified using a threshold-approach. Precipitation was snowfall when temperature was < 1 ºC according to previous ROS research in the study zone (Corripio and López-Moreno, 2017) and the average rain-snow temperature threshold for the Pyrenees (Jennings et al., 2018). Snow cover is calculated by a linear function of snow depth, snow albedo is estimated based on a prognostic function with the new snowfall. Snow thermal conductivity is estimated based on snow density. Liquid water percolation is calculated based on a gravitational drainage. Compaction rate is simulated from overburden and thermal metamorphism. The atmospheric stability is estimated through the Richardson number stability functions to simulate latent and sensible heat fluxes. The selected FSM2 configuration

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

includes three snow layers and four soil layers. The detailed FSM2 physical parameters and Fortran compilation numbers are shown in Table S1. The FSM2 model and configuration was previously validated in the Pyrenees at Bonsoms et al. (2022b). FSM2 has been successfully used in snow model sensitivity studies in alpine zones (Günther et al., 2019). FSM2 has been implemented in a wide range of alpine conditions, such as for the Iberian Peninsula mountains (Alonso-González et al., 2019), Spanish Sierra Nevada (Collados-Lara et al., 2020) or swiss forest environments (Mazzotti et al., 2020) snowpack modeling. FMS2 has been integrated in snow data-assimilation schemes in combination with in-situ (Smyth et al., 2022) and remote-sensing data (Alonso-González et al., 2022).

3.2 Atmospheric forcing data

The FSM2 was forced with the SAFRAN meteorological system reanalysis dataset for flat slopes (Vernay et al., 2022). The SAFRAN meteorological system integrates meteorological simulations, remote-sensing cloud cover data, and instrumental records through data-assimilation. SAFRAN is forced with a combination of homogenized ERA-40 reanalysis (1958 to 2002) and the numerical weather prediction model ARPEGE (2002 to 2020). SAFRAN system was firstly designed for avalanche monitoring (Durand et al., 1999, 2009), but the accurate results obtained enhanced the diffusion of the meteorological system and its integration in the French hydrometeorological modelling system by the local weather service, Meteo-France (Habets et al., 2008). SAFRAN performance has been extensively validated. For instance, in long-term and high-resolution climate analysis (Devers et al., 2021), seasonal forecasting (Ćeron et al., 2010) or the meteorological modelling of continental France (Quintana-Seguí et al., 2008) and Spain (Quintana-Seguí et al., 2017). SAFRAN system has been used as meteorological forcing data for the snow modeling in complex alpine terrain (Revuelto et al., 2018; Deschamps-Berger et al., 2022), to study long-term snow evolution (Réveillet et al., 2022), avalanche hazard forecasting (Morin et al., 2020), snow climate projections (Verfaillie et al., 2018), snow depth (López-Moreno et al., 2020) and energy heat fluxes spatio-temporal trends (Bonsoms et al., 2022a). SAFRAN meteorological system exhibit and accuracy of around 1 ºC in air temperature and around 20 mm in the monthly cumulative precipitation (Vernay et al., 2022).

3.3 Spatial areas

SAFRAN system provides data at hourly resolution from 0 to 3600 m, by steps of 300 m, grouped by massifs. The SAFRAN massifs (polygons of Figure 1) were chosen for their relative topographical and climatological similarities (Durand et al., 1999). We selected the 1500 m (low), 1800 m (mid), and 2400 m (high) specific elevation bands of the Pyrenees. In order to retain the main spatial differences across the mountain range, reduce data dimensionality and include the maximum variance, massifs with similar interannual snow characteristics were grouped into sectors by performing a Principal Component Analysis (PCA). PCA is an extensively applied statistical method for climatological and snow spatial regionalization (i.e., López-Moreno and Vicente-Serrano, 2007; Schöner et al., 2019; Alonso-González et al., 2020a; Matiu et al., 2021; Bonsoms

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

et al., 2022a). A PCA was applied over HS data for all months and years of the baseline climate. Massifs were grouped into four groups depending on the maximum correlation to the first (PC1) and second (PC2) scores. Pyrenean sectors were named South-West (SW), South-East (SE), North-West (NW) and North-East (NE) due to their geographical position. Figure 1 shows the resulting Pyrenean regionalization for low, mid and high elevation as well as the SAFRAN massifs PC1 and PC2.

3.4 Sensitivity analysis

ROS season extension was defined according to ROS occurrence during the baseline climate period. For the purposes of this research, seasons are classified as follows: October and November (Autumn); December, January, and February (Winter); March, April, May, and June (Spring); and July (Summer). August and September are not included due to the absence of regular snow cover. ROS sensitivity to precipitation, air temperature, increasing incoming longwave radiation (LWin) accordingly, was performed though a delta-change approach. This method has been successfully applied and validated for analyzing the snow sensitivity to temperature and precipitation changes in many mountains, such as the Pyrenees (e.g., López-Moreno et al., 2013), the Iberian-Peninsula mountain areas outside the Pyrenees (Alonso-González et al., 2020a), Alps (Marty et al., 2017), Canadian basins (Pomeroy et al., 2015; Rasouli et al., 2019), or western United-States (Musselman et al., 2017b), among other works. Delta-change has also been also performed in global ROS sensitivity to temperature change studies (López-Moreno et al., 2021). SAFRAN reanalysis climate data was perturbed according to Spanish Meteorological Agency climate change scenarios projected for the 21st Century in the Pyrenees (Amblar-Francés et al., 2020). Precipitation was increased (+10%), left unchanged (0 %) and decreased (- 10%). Air temperature was perturbed from +1ºC to +4ºC by +1ºC. LWin was increased due to warming, by applying the Stefan-Boltzmann law, using the Stefan-Boltzmann constant, and the hourly atmospheric emissivity derived from SAFRAN air temperature and LWin.

3.5 HS, Sf and ROS climate indicators

The average HS and Sf delta-change (expressed in % per ºC) is the average seasonal HS and Sf anomalies under the baseline climate and divided by degree of warming. Days are classified as ROS days when daily rainfall amount was >= 10 mm and HS >= 0.1 m, according to previous works (Musselman et al., 2018; López-Moreno et al., 2021). ROS fr are the number of ROS days. ROS rain is the average daily rainfall (mm) during a ROS day. ROS ablation is the average daily snow ablation (cm) during a ROS day. The average daily snow ablation is the daily average HS difference between two consecutive days (Musselman et al., 2017a). Only the days when a negative HS difference occured were selected. ROS exposure is the relation between ROS rain (y-axis) and ROS fr (x-axis) differences from the baseline climate scenario for the massifs were ROS fr is recorded for all increments of temeperature.

4 Results

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

We provide an analysis of ROS drivers, near-present ROS patterns and their response to warming. ROS spatio-temporal dynamics are analyzed by frequency, rainfall quantity and snow ablation. Since we have detected a non-linear and counter-intuitive ROS sensitvity to temperature, ROS indicators values are shown for each increment of temperature, grouped by elevation and sectors, namely SW, SE, NW and NE.

4.1 ROS drivers

HS and Sf delta-change is shown in Figure 2. Seasonal HS and Sf delta-change variability is mostly controlled by the increment of temperature, season, elevation, and spatial sector (Figure S1). The role of precipitation variability in the seasonal HS evolution is moderate to low (Figure S2 to S4). Only in high elevation an upward trend of precipitation (at least > 10%) can counterbalance small increments of temperature (< 1ºC, over the baseline climate) from December to February (Figure S4). For this reason, precipitation was excluded to further analysis. Snow in low and mid elevations during summer is rarely observed, however, marginal snow cover in high elevation can last until June and July, especially in the wettest sectors of the range (NW and SW). Seasonal HS and Sf delta-change show an elevation-dependent pattern to warming and large seasonality. The average HS decrease per ºC ranges from 39 %, 37 % and 28 % per ºC, for low, mid and high elevations, respectively. However, relevant differences are found depending on the season and degree of warming (Figure 2). Maximum HS and Sf reductions are found in low and mid elevations during the shoulders of the season (autumn and spring), coinciding with the time when ROS events are more frequent for the baseline climate (Figure 3). In these elevations, maximum HS decreases (52 % over the baseline climate) are modeled for spring when temperature is + 1ºC. The greatest HS decreases in high elevation areas are modeled for summer (54 % HS decrease for 1ºC). If temperature reaches maximum values (+ 4 ºC), seasonal HS is reduced 92 %, 89 %, and 79 % for low, mid, and high elevations, respectively (Figure S5).

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

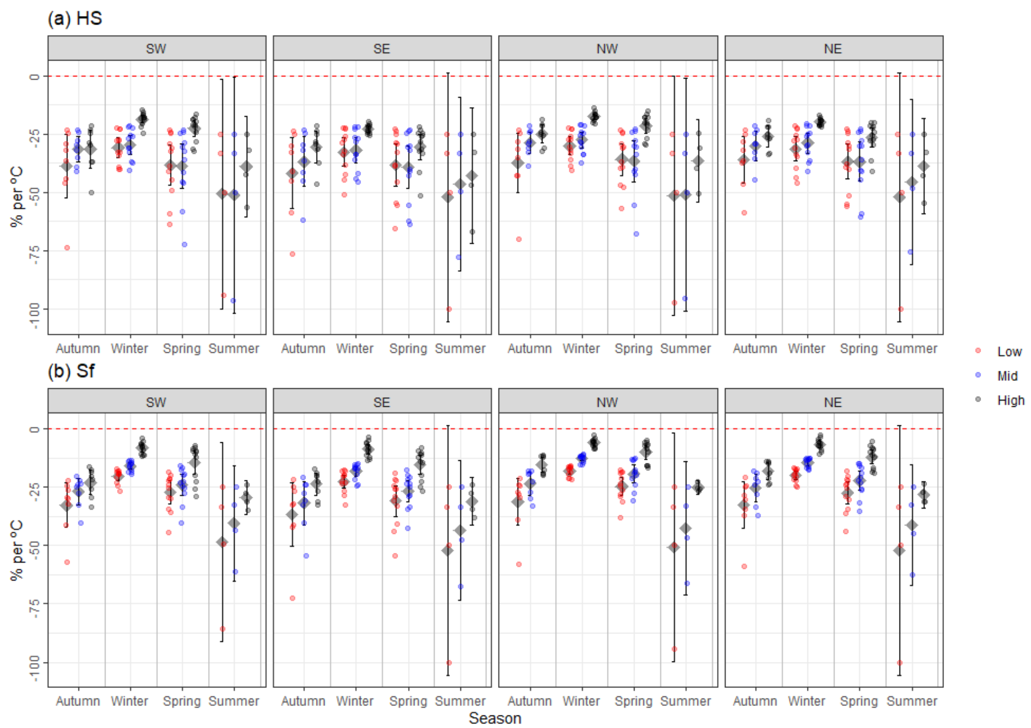

Figure 2. Seasonal (a) HS and (b) Sf anomalies over the baseline climate. Data are shown by elevation (colors), season (x-axis) and sectors (boxes). Points represent the average seasonal HS and Sf anomalies grouped by month of the season and increment of temperature (from 1ºC to 4ºC). The black diamond point indicates the mean, whereas the upper and lower error bars show the Gaussian confidence based on the normal distribution.

Sf shows lower sensitivity to warming than HS and maximum reductions in autumn. On average, Sf decreases by 29%, 22 %, and 12 % per ºC for low, mid, and high elevations, respectively. An increase of 4ºC supposes Sf reductions of 80 %, 69 % and 49 % for low, mid, and high elevations. HS and Sf delta-change shows also different sensitivities across the range. Independently of the elevation band and season, the SE exhibit the greatest HS and Sf decreases (41 % and 35 % per ºC, respectively). On the contrary, minimum reductions are expected in the northern slopes (NW and NE).

4.2 ROS frequency

Low elevation annual ROS fr for the baseline climate is 17, 8, 10 and 7 days/year for SW, SE, NW, NE sectors, respectively (Figure 3). The highest annual ROS fr is however observed at mid elevation. Here, annual ROS fr is 17, 9, 12 and 9 for SW, SE, NW, NE sectors. Within these elevations, the maximum ROS fr is detected in SW during winter and spring (7 days/season, for both elevations and seasons). The eastern Pyrenees follow a

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

similar seasonality. Maximum ROS fr in low elevation is found in winter (4 and 3 days/season, SE and NE, respectively), and during spring in mid elevation (4 and 3 days, SE and NE, respectively). ROS is rarely observed in SE during the latest month of spring (May), which contrast with the modeled values for SW (2 and 3 days/month, for low and mid elevations, respectively). High elevation shows the minimum ROS fr. Here, comparisons between seasons reveal maximum ROS fr during summer, especially in SW (7 days/season), followed by NW (6 days/season), and NE (2 days/season).

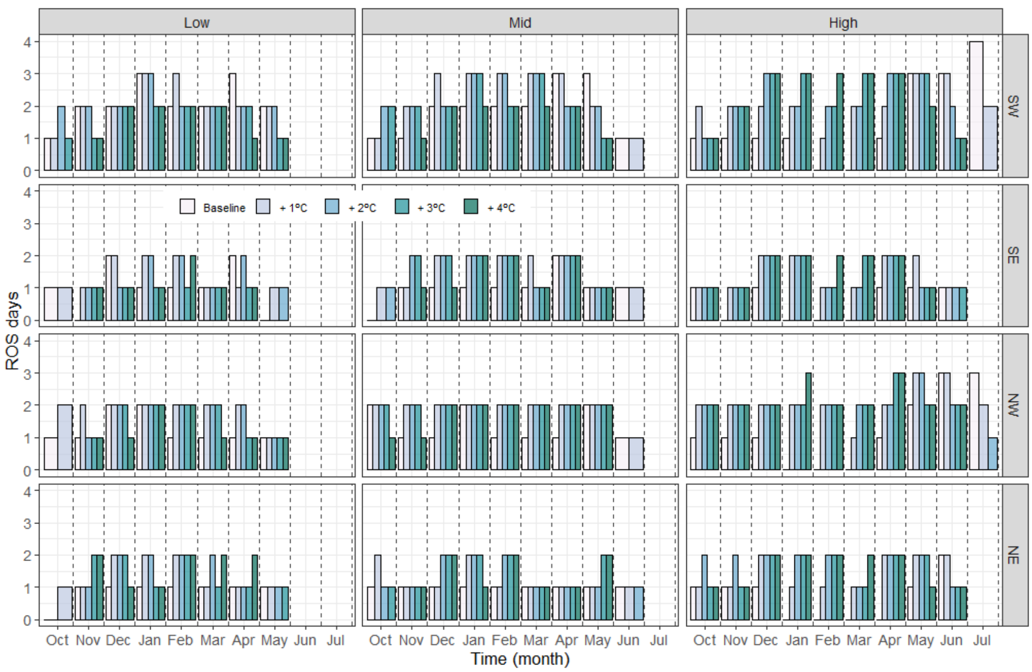

Figure 3. ROS fr for baseline climate period and increments of temperature, grouped by months (x-axis), sector (rows) and elevation (columns).

ROS fr response to warming vary depending on the month, increment of temperature, elevation, and sector. ROS tends to disappear in October for low elevations except in SW (Figure 3). The highest increases are seen during the winter for increments temperature lower than 3ºC, particularly in NE, where ROS fr increases 1 day per month over the baseline scenario for + 1ºC. In mid elevation, ROS fr increases in all regions from November to February (around 1 day per month, for + 1ºC up to + 3ºC). Similar increases are expected in NW and SW during the earliest months of spring and for low to moderate increments of temperature. The contrary is observed during the latest months of spring in SW, where warming reduces ROS events. A slight ROS fr increase is found during spring for the rest of the sectors (Figure 4). ROS events in June are expected to disappear for temperature increases higher than 1ºC. Finally, high elevation shows the largest ROS fr variations (around 1 day/month for + 1ºC). Maximum ROS fr increases (3 days/month) are found in SW for more than +

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

3ºC. ROS fr progressively increases in March and April for all sectors but tends to decrease in May (for + 3ºC), June and July (for + 1ºC).

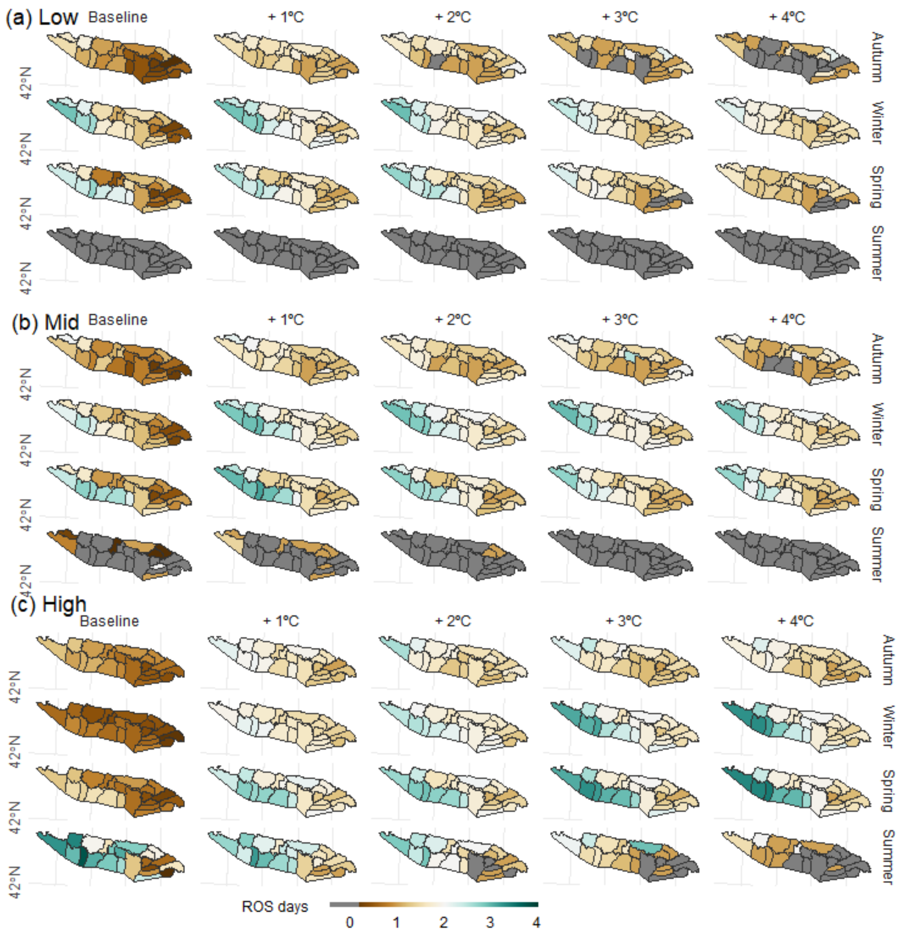

Figure 4. Average ROS fr (days) for a season for (a) low, (b) mid and (c) high elevation. Data are shown for the baseline climate period and increment of temperature (left to right).

4.3 ROS rain

The spatial and temporal distribution of ROS rain is presented in Figure 5 and 6. The average low elevation ROS rain by year is 23, 28, 21, and 20 mm/day for SW, SE, NW, NE sectors, respectively. Similarly, the highest values in mid elevation are found in SE (29 mm/day, respectively). SE sector experiences the highest ROS rain during autumn and summer (around 40 mm/day in low and mid elevations). High elevation maximum ROS rain values are however found in the western Pyrenees during the onset and offset snow season. Here, the

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

largest ROS rain spatial and seasonal distribution ranges from SW (29 mm/day, autumn), NW (28 mm/day, summer), SE (24 mm/day, autumn) to NE (23 mm/day, autumn).

Figure 5. ROS rain (mm) temporal evolution for baseline climate and increment of warming (colors), grouped by elevation (columns) and sector (rows).

ROS rain progressevly increases due to warming (4%, 4%, and 5% per ºC for low, mid, and high elevations, respectively; Table S2). Small differences are found by elevation and sector. Low elevation ROS rain increases until + 3°C, and generally decreases for + 4°C during the earliest (October to December) and latest (April and May) months of the snow season. Similar patterns are found in mid elevation. ROS rain increases up to + 4°C, except in the SE sector for specific months (Figure 5). The lattest sector shows also maximum ROS rain values in autumn due to torrential rainfall. High elevation ROS rain increase at a constant rate of around 5 % per ºC. Yet, maximum increases are modeled in SW during summer, when ROS rain almost doubles the baseline climate (+ 40% for + 4°C).

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

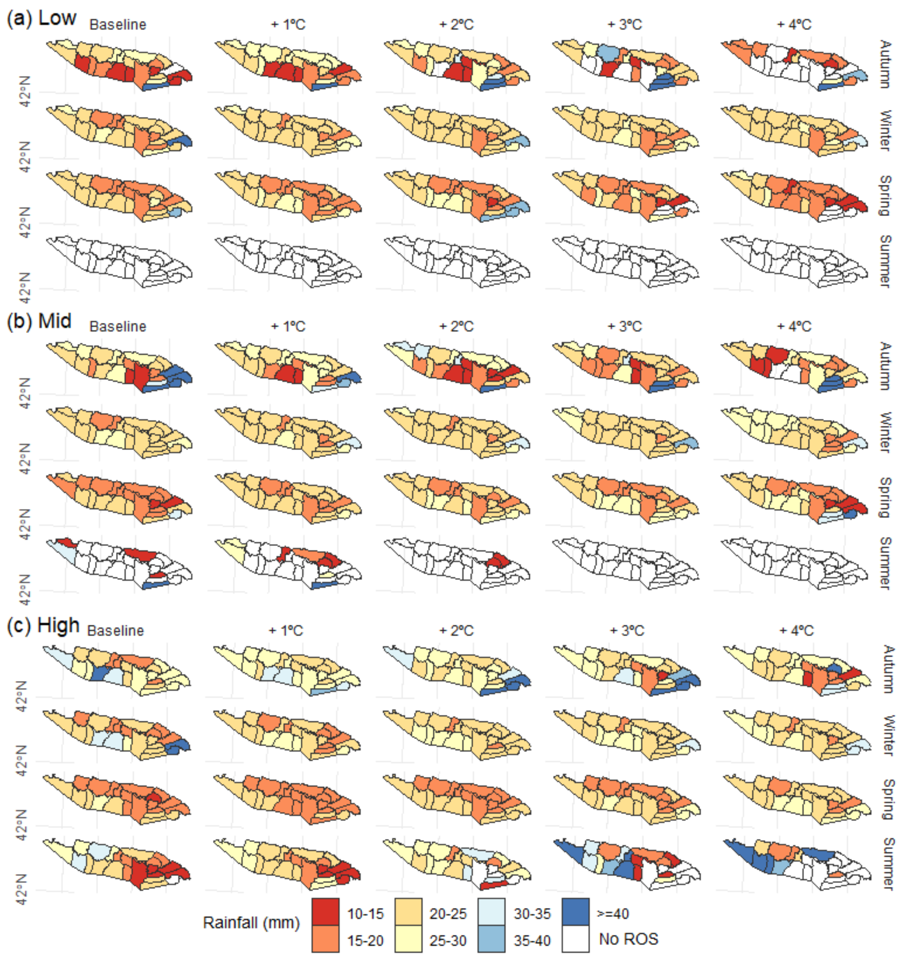

Figure 6. Average ROS rain (mm) for a season for (a) low, (b) mid and (c) high elevation. Data are shown for the baseline climate period and increment of temperature (left to right).

Data suggest that ROS exposure generally increases for all elevations and sectors during winter (except in SW for temperatures greater than 3°C). Nonetheless, remarckable spatial and seasonal differences are found. SE show the maximum values in autumn. On the contrary, small changes in frequency are detected in SW and NW, despite ROS rain is expected to increase (< 10mm/day). For the majority of sectors and elevations, ROS exposure generally increases in winter and spring. The minimum differences between sectors are dected in these seasons. In summer, ROS exposure tends to generally decrease for all elevations under severe warming due to snow cover depletion.

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

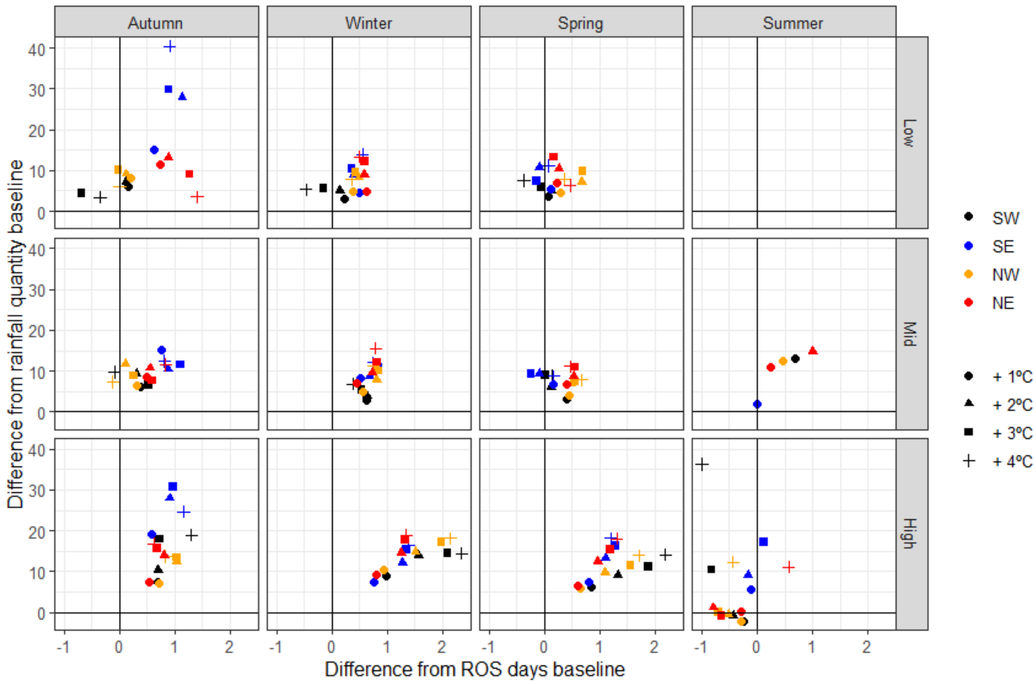

Figure 7. Average ROS exposure. Points are obtained by a scatterplot between ROS rain difference from baseline climate (y-axis) and ROS days difference from baseline climate (x-axis). Data is calculated by the average difference between (a) the baseline scenario and (b) the different perturbed scenarios, only for the massifs where ROS fr exists on (a) and (b). Data are shown for each season (columns), elevation (rows), sector (color) and increment of temperature (point shape).

4.4. ROS ablation

ROS ablation is presented at Figure 8 and 9. ROS ablation ranges from -10 cm/day in NW high elevation (summer) to - 5 cm/day in NE high elevation (winter). ROS ablation nearly doubles the average daily snow ablation for all days on a season (Figure S6). Comparison with the reference baseline period reveals contrasting ROS ablation changes depending on the season, elevation and sector. Overall ROS ablation progressively increases due to warming in coldest zones and months of the season. The largest ROS ablation increments are detected in autumn and winter. For the former, ROS ablation increases at a generally constant rate in SW (11 %) NE (19 %) and NW (4 % per ºC). For the latter, ROS ablation increases also in SW (11 %), NW (14 %) and NE (34 % per ºC). In detail, maximum ROS ablation due to warming is found for mid elevation during autumn (Figure 8). ROS ablation exhibit slow and no-changes in the warmest zone (SE), as well in the warmest months of the season, regardless the elevation band.

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

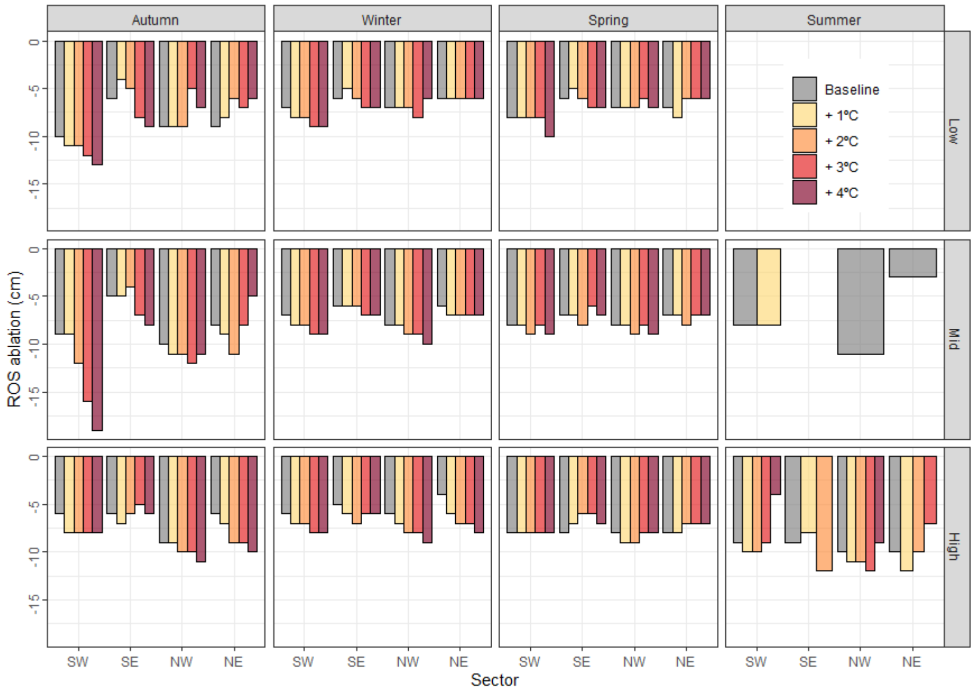

Figure 8. ROS ablation (y-axis) for baseline climate period and increment of temperature (colors), sector (x-axis), season (columns) and elevation (rows).

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

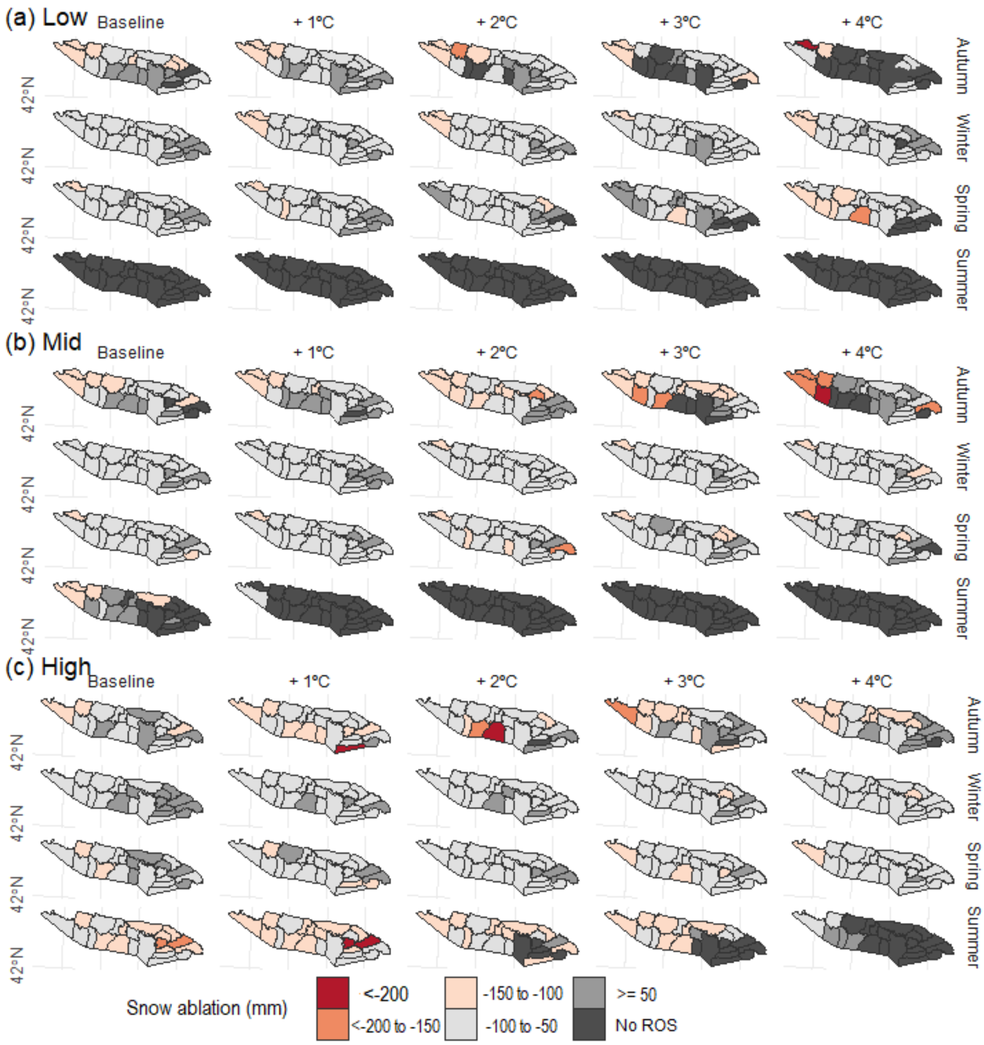

Figure 9. Average ROS ablation for a season for (a) low, (b) mid and (c) high elevation. Data are shown for the baseline climate period and increment of temperature (left to right).

5 Discussion

The Pyrenees experienced a statistically significant positive temperature trend since the 1980s (ca. + 0.2 ºC/decade) but no statistically significant precipitation trends are detected (OPCC, 2018) due to strong spatial (Vicente-Serrano et al., 2017), inter-annual and long-term variability of the latter (Peña-Angulo et al., 2021). Depending on the study period different snow trends were found. From ca. 1980 to 2010, non-statistically

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

significant snow days and snow accumulation positive trends were generally detected at > 1000 m (Buisan et al., 2016), 1800 m (Serrano-Notivoli et al., 2018), and > 2000 m (Bonsoms et al., 2021a). Long-term trends (1957 to 2017), however, reveal statistically-significant snow depth decreases at 2100 m, but large variability depending on the sector and the snow indicator (López-Moreno et al., 2020). Climate projections for the end of the 21st century suggest an increase of temperature (> 3ºC), together with low precipitation shifts (< 10%) from autumn to spring (Amblar-Francés et al., 2020). Within this climate context, ROS spatio-temporal patterns will likely change. In order to anticipate future scenarios, ROS sensitivity to warming was analyzed through three key indicators of frequency, rainfall intensity and snow ablation.

5.1 ROS spatial variability

The climatic setting of the Pyrenees as well as its relief configuration determines a remarkable spatial and temporal variability of ROS events. The contradiction between rainfall ratio increases and snowpack reductions, as well as the high spatial and monthly differences found, explain the complex ROS response to warming. HS decrease by 39 %, 37 % and 28 % per ºC, for low, mid and high elevations, respectively. Similarly, Sf decreases by 29 %, 22 %, and 12 % per ºC for low, mid, and high elevations, respectively, providing evidence of an elevation-dependent snow sensitivity to temperature change. HS and Sf maximum reductions are reached for 1ºC of warming, suggesting non-linear HS decreases, in accordance with previous snow sensitivity to climate change reported in central Pyrenees (López-Moreno et al., 2013). In detail, SW and NW annual ROS fr almost doubles (17 and 12 days/year, respectively) the one recorded in SE and NE (9 days/year, for both sectors). Maximum ROS fr for a season are found in SW and NW because of larger snow magnitudes in this sector (i.e., López-Moreno, 2005; López-Moreno et al, 2007; Navarro-Serrano et al., 2017; Bonsoms et al., 2021a). Thus, snow cover last longer until spring when minimum Sf values are found (Figure S1). This sector is the most exposed to SW and W air flows (negative NAO phases) (López-Moreno, 2005), which bring wet and mild conditions over the mountain range, leading to most ROS-related floods in the range (Morán-Tejeda et al., 2019). The generally ROS rain increase reported in this work (independently of the increment of temperature and elevation) is explained by the Sf reduction expected for all sectors (Figure 2). Maximum ROS rain is generally detected in spring (May), except in NE high elevation zones and SE (all elevations). In the latter sectors, ROS rain tends to dissapear in Octuber under large (> 2ºC) increments of temperature. The seasonal snow accumulation in NE and SE is lower-than-average due to the lower influence of Atlantic climate in these sectors of the range. Hence, large increments of warming decreases ROS fr due to snow cover depletion in early autumn and late spring (Figure S1). In addition, SE is closer to the 0ºC due to higher-than-average sublimation, latent and radiative heat fluxes (Bonsoms et al., 2022a) and for this reason in this sector each increment of temperature has larger effects on the Sf, HS and ROS fr reduction (Figure 2). High elevation show the largest variation over the baseline climate as well as ROS exposure because of the larger snowpack magnitude and duration compared to low and mid areas. Thus, high elevation snow duration last until spring and summer, when the largest shift from snowfall to rainfall is found. On the other hand, mid elevation shows the maximum ROS rain since the amount of moisture for condensation decreases while air

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

masses increase height (Roe and Baker, 2006). Furthermore, the largest ROS rain is detected in SE during autumn (Figure 6), because of the exposure of this region to Mediterranean low-pressure systems (negative WeMO phases), that usually trigger heavy rainfall events during this season (Lemus-Canovas et al., 2021).

5.2 ROS temporal evolution

Recent ROS trends in other mid-latitude areas are in accordance with ROS analysis presented here. Freudiger et al. (2013) analyzed the ROS trends (1950–2011 period) of the Rhine, Danube, Elbe, Weser, Oder, and Ems (Central Europe) basins. They found an overall ROS fr increaase during January and February (1990 to 2011 period), which is consistent with the ROS rain and frequency increase detected in winter for the Pyrenees for all elevations and increment of temperature. Similarly, in Sitter River (NE Switzerland), a ROS fr increase of around 40% (200%) at <1500 m (>2500 m) was detected between 1960 and 2015 (Beniston and Stoffel, 2016). During the last half of the 20^{th} century, ROS fr trends show an upward (downward) trend in high (low) elevation in western United-States (McCabe et al., 2007), as well as in southern British Columbia (Loukas et al., 2002) and at catchment scale in Oregon (United-States) (Surfleet and Tullos, 2013). Same ROS fr increases (decreases) has been detected from 1980 to 2010 in Norwegian high (low) elevated mountain zones (Pall et al., 2019). However, in contradiction with our results and previous studies, winter Northern-Hemisphere ROS fr trends (1979-2014 period) show no-clear trends (Cohen et al., 2015).

Results exposed in this work provide more evidence of ROS fr increases in high-elevation zones, as it has been suggested by climate projections and ROS sensitivity to temperature studies. ROS show an elevation-dependent pattern that was previously reported in the Swiss Alps (Morán-Tejeda et al., 2016). In Sitter River (NE Switzerland), an increase of 2 to 4 °C over the 1960 to 2015 period results in an increase of the ROS fr by around 50% at > 2500 m (Beniston and Stoffel, 2016). Likewise, 21st century high-emission scenarios (RCP8.5), suggest increases in ROS fr and intensity in Gletsch (Switzerland) high-elevation area; however, on climate projections for ROS definitions that include snow melting (Musselman et al., 2018), natural climate variability contributes to a large extend (70 %) of ROS variability (Schirmer et al., 2022). Li et al. (2019) analyzed the future ROS fr in the conterminous United-States and detected a nonlinear trend ROS due to warming, which is consistent with the different ROS rain and frequency responses depending on the increment of temperature detected in our work. Climate projections for the mid-end of the 21^{th} century projected positive ROS fr and rainfall trends in Western United-States and Canada (il Jeong and Sushama, 2018). Similarly, ROS fr will likely decrease (increase) in the warmest months of the season in low (high) elevation areas of western United-States (Musselman et al., 2018). The same is projected Norwegian mountains (Mooney and Li, 2021). López-Moreno et al. (2021) analyzed 40 worldwide basins ROS sensitivity to warming. In their study they found a decrease of ROS events in warm mountain areas. However, they detected ROS fr increases in cold-climate mountains where large snow accumulation is found despite warming. In accordance with our results, they identified large seasonal differences and ROS fr decreases in Mediterranean mountains due to snow cover depletion in the lasts months of the snow season.

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

5.3 ROS ablation

Warming increases ROS ablation from autumn to winter on deep snowpacks and in the coldest sectors of the range, due to higher energy for snow ablation and closer 0ºC isotherm conditions in a warmer than baseline climate. Nevertheless, data show low or decreases in ROS ablation in SE and spring, since the snowpack is already near to the isothermal conditions. These results go in line with results modelled for cold and warm Pyrenean sites (López-Moreno et al., 2013) as well as for different Northern-Hemisphere sites (Essery et al., 2020). ROS ablation indicator is also indirectly affected by the HS magnitude decreases (30 % per ºC; Figure 2), and therefore lower ROS ablation is direclty affected by lower HS magnitudes. Previous literature pointed out that warming have counter-intuitive effects on snow ablation patterns. Higher than average temperatures advance the peak HS date on average 5 days per ºC in mid and high elevations (Bonsoms et al., 2022b), triggering earlier snow ablation onsets, and therefore lower solar radiation fluxes (López-Moreno et al., 2013; Lundquist et al., 2013; Pomeroy et al., 2015; Musselman et al., 2017a; Sanmiguel-Vallelado et al., 2022), as well as earlier snow depletion before the maximum advection of heat fluxes into the snowpack (spring) (Bonsoms et al., 2022a). Slower snow melt rates in a warmer climate have been detected in Western United-States (Musselman et al., 2017), as well as the entire Northern-Hemisphere (Wu et al., 2018). Low or inexistent changes in snow ablation on warm and marginal snowpacks has been previously detected in the central Pyrenees (López-Moreno et al., 2013), in forest and open areas (Sanmiguel-Valellado et al., 2022), in the entire range (Bonsoms et al., 2022b), and other Iberian Peninsula Mountain ranges outside the Pyrenees (Alonso-González et al., 2020a).

ROS ablation is larger than the average snow ablation during a snow ablation day (Figure S6) due to higher SEB positive fluxes. Several works analyzed SEB changes on ROS events, and different SEB contributions has been found depending on the geographical area (Mazurkiewicz et al., 2008; Garvelmann et al., 2014b; Würzer et al., 2016; Corripio and López-Moreno, 2017; Li et al., 2019), ranging from net radiation in Pacific North West (Mazurkiewick et al., 2008) to LWin and turbulent heat fluxes in conterminous United-States mountain areas (Li et al., 2019) or the Swiss Alps (e.g., Würzer et al., 2016). In general, studies in mid-latitude mountain ranges have shown that turbulent heat fluxes contribute between 60 and 90 % of the energy available for snow ablation during ROS days (e.g., Marks et al., 1998; Garvelmann et al. 2014; Corripio and López-Moreno, 2017). In the central Pyrenees (> 2000 m) the meteorological analysis of a ROS event reveals that ROS ablation is larger than a normal ablation day because of the large advection of LWin and especially sensible heat fluxes (Corripio and López-Moreno, 2017). LWin increases due to the high cloud cover and warm air, as it is frequently observed during ROS episodes (Moore and Owens, 1984).

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

5.4 ROS socio-environmental impacts and hazards

Temperature-induced changes in the seasonal snowpack and during ROS days suggest several hydrological shifts including, but not limited to, earlier peak flows on the season (Surfleet and Tullos, 2013), rapid streamflow peaks during high precipitation events in frozen soils (Shanley and Chalmers, 1999), faster soil moisture depletion and lower river discharges in spring due to earlier snow melt in the season (Stewart, 2009). The shortening of the snow season due to warming reported in this work will potentially alter alpine phenological patterns (i.e., Wipf and Rixen, 2010) and expand forest cover (Szczypta et al., 2015). Although vegation branches intercepts a large amount of snowfall, intermediate and high vegetation shields short-wave radiation, reduces snow wind-transport and turbulent heat fluxes (López-Moreno and Latron, 2008; Sanmiguel-Valellado et al., 2022). Snow-forest interactions, their sensitivity to climate change as well as the ROS hydrological response within a changing landcape is far from understood across the range and should be the base of forecoming works.

The higher ROS exposure (Figure 7) will likely imply an increase of ROS-related hazards and impacts in the mountain ecosystem. Heavy ROS rain changes snow methamorphism on saturated snowpacks and leads to high-speed water percolation (Singh et al., 1997). The subsequent water refreezing changes the snowpack conditions and creates an ice-layer in the snowpack that can reach the surface (Rennert et al., 2009). ROS can cause plant damage (Bjerke et al., 2017) and the ice encapsulation of vegetation in tundra ecosystems can trigger severe wildlife impacts, such as vertebrate herbivores starvation (Hansen et al 2013), reindeer population mortality (Kohler and Aanes, 2004) and higher competition between species (Hansen et al 2014). Nevertheless, any study to the date analyzed ROS-related impacts in flora and fauna across Southern-European mountains. Snow albedo decay due positive heat fluxes and rainfall in ROS events (Corripio and López-Moreno, 2017), lead to faster snow ablation even on the next days (e.g., Singh et al. 1997). The combination of changes in internal snowpack processes, larger ROS rain, and more energy to ablate snow during spring could enhance snow runoff, especially during warm and wet snowpack conditions (Würzer et al., 2016). In snow-dominated regions ROS can lead to a specific type of avalanching (Conway and Raymond, 1993) and floods (Surfleet and Tullos, 2013). The latter are the most environmental damaging risk in Spain (Llasat et al., 2014) and around 50% of the flood in the Iberian Peninsula are due to ROS events (Morán-Tejeda et al., 2019). More than half of the historical (1940 to 2012) flood events in the Ésera river catchment (central Pyrenees) occurred during spring (Serrano-Notivoli et al., 2017), which coincides with the snow ablation season. ROS floods have also economic impacts. For instance, a ROS flood event that occurred on 13th June of 2013 in the Garonne River (Val d'Aran, central Pyrenees) cost approximately 20 million of euros to the public insurance (Llasat et al., 2014).

6 Conclusions

The expected decreases in Sf and HS due to climate warming will likely change ROS spatio-temporal patterns

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

across the Pyrenees. Therefore, a better understanding of ROS is required. This work analyzed the ROS sensitivity to warming by forcing a physically based snow model with perturbed reanalysis climate data (1980-2019 period) for low, mid and high elevation areas of the Pyrenees. ROS delta-change is evaluated by frequency, rainfall intensity and snow ablation during ROS days.

During the baseline climate period, annual ROS fr totals on average 10, 12 and 10 day/season for low, mid and high elevations. Higher-than-average annual ROS fr are found in mid elevation SW (17 days/year) and NW (12 days/year), which contrast with the minimums detected in SE (9 days/year). The different spatial and seasonal ROS response to warming suggest that contrasting and shifting trends could be expected in the future. Overall ROS fr decreases during summer in high elevation for > 1ºC. When temperature is progressively increased the greatest ROS fr increases are found for SW high elevation (around 1 day/month for + 1ºC). ROS fr is highlty sensitive to warming in the snow onset and offset months, when counterintuitive factors play a key role. On the one hand, maximum Sf decreases are modeled for spring, leading to rainfall increases; on the other hand, warming depletes the snowpack in the warmest and snow driest sectors of the range. Consequently, data suggest a general ROS fr decrease for the majority of the SE massifs, where the snowpack is near the isothermal conditions in the baseline climate period. Yet, during spring, the highest ROS fr increases are detected in SW and NW, since these sectors are less exposed to radiative and turbulent heat fluxes and record higher-than-average seasonal snow accumulations.

ROS rain generally increases due to warming, independently of the sector and elevation, being limited by the number of ROS days. The largest and constant increments are observed in spring, when ROS rain increases at a rate of 7, 6 and 3 % per ºC for low, mid and high, respectively. ROS rain increases are explained by Sf reductions, which decrease at a rate of 29 %, 22 %, and 12 % per ºC for low, mid, and high elevations, respectively. ROS rain maximum values are detected in SE (28 mm/day), especially in mid elevation during autumn (45 mm/day), since this sector is exposed to subtropical Mediterranean flows.

Finally, ROS ablation shows contrasting patterns depending on the season, sector and elevation. Generally, ROS ablation increases in cold snowpacks, such as those modeled in high elevation and during cold seasons (autumn and winter). Here, ROS ablation follows a constant ablation rate of around + 10% per ºC, due to higher-than-average positive sensible and LWin heat fluxes. However, in SE and low elevations, where marginal and isothermal snowpacks are found, no changes or decreases in ROS ablation are detected due to snowpack magnitude reductions in a warmer climate. Results demonstrate the high snow sensitivity to climate within a mid-latitude mountain range, and suggest significant changes with regards to water resources management. Relevant implications in the ecosystem and socio-economic activities associated with snow cover are anticipated.

https://doi.org/10.5194/egusphere-2023-178
Preprint. Discussion started: 20 March 2023

Data availability

FSM2 is an open access snow model (Essery, 2015) provided at https://github.com/RichardEssery/FSM2 (last access 15 January 2023). SAFRAN climate dataset (Vernay et al., 2022) is available by AERIS at https://www.aeris-data.fr/landing-page/?uuid=865730e8-edeb-4c6b-ae58-80f95166509b#v2020.2 (last access 16 December 2022). Data of this work is available upon request by the first author (josepbonsoms5@ub.edu).

Author contribution

J.B., J.I.L.M., and E.A.G. designed the work. J.B. analyzed the data and wrote the manuscript. J.B., J.I.L.M., E.A.G., C.D.B., and M.O. provided feedback and edited the manuscript. J.I.L.M., M.O. supervised the project and acquired funding.

Competing interests

The authors declare that they have no conflict of interest.

Acknowledgements

This work frames within the research topics examined by the research group "Antarctic, Artic, Alpine Environments-ANTALP" (2017-SGR-1102) funded by the Government of Catalonia, HIDROIBERNIEVE (CGL2017-82216-R) and MARGISNOW (PID2021-124220OB-100), from the Spanish Ministry of Science, Innovation and Universities. JB is supported by a pre–doctoral University Professor FPI grant (PRE2021-097046) funded by the Spanish Ministry of Science, Innovation and Universities.

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Preprint. Discussion started: 20 March 2023

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