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Future changes in snowpack will impact seasonal runoff and low flows in Czechia

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ABSTRACT

Study Region: Central Europe, Czechia.
Study Focus: Mountains are referred to as “water towers” because they substantially affect the hydrology of downstream areas. However, snow storages will decrease in the future due to the increase in air temperature which will affect streamflow regime. Main objectives of this study were 1) to simulate the future changes in snow for a large set of mountain catchments in Czechia, reflecting a wide range of climate projections and 2) to analyse how the snow changes will affect groundwater recharge, streamflow seasonality and low flows in the future.
New Hydrological Insights: The future hydrological projections showed a decrease in annual maximum SWE by 30 %–70 % in the study area until the end of the 21st century. Additionally, snowmelt was found to occur on average 3–4 weeks earlier. The results showed the large variability between individual climate chains and indicated that the increase in air temperature causing the decrease in snowfall might be partly compensated by the increase in winter precipitation. Changes in snowpack will cause the highest streamflow during melting season to occur one month earlier, in addition to lower spring runoff volumes due to lower snowmelt inputs. The future climate projections leading to overall dry conditions in summer are associated with both the lowest summer precipitation and seasonal snowpack. The expected lower snow storages might therefore contribute to more extreme low flow periods.

1. Introduction

Mountain snowpack is highly sensitive to air temperature which influences the phase of precipitation. As a result of increasing air temperature documented by observations, the snowfall fraction, the proportion of snowfall water equivalent to the total amount of precipitation over the given period, decreases (Knowles et al., 2006). This decrease leads to a generally decreasing snowpack, later snow accumulation and earlier snowmelt (Fyfe et al., 2017; Marty et al., 2017b). For example, Zeng et al. (2018) showed that in the most affected parts of the western United States, the annual maximum SWE decreased on average by 41 % and the snow season was shortened by 34 days. It is also predicted that the above changes will continue in the future, due to further changes in climate (Jenicek et al., 2018; Marty et al., 2017a; Musselman et al., 2017a).

Nevertheless, the response of snow to the increase in air temperature is often not straightforward, resulting in a large spatial variability between individual regions and elevations with different climate (Blahušiaková et al., 2020; Cooper et al., 2016). For

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example, the negative effect of increasing temperatures on snow storage may be partially compensated by an increase in precipitation, specifically by the increase in the number of extreme snowfall events (Lute et al., 2015; Marshall et al., 2020). Below-average snow conditions, defined as snow drought, can be caused by either high air temperatures or low winter precipitation or a combination of both (Cooper et al., 2016; Harpold et al., 2017). With predicted future warming, many regions might become more susceptible to snow droughts, especially those with winter air temperatures currently close to the freezing point (Dierauer et al., 2019).

It is well known that changes in climate indices, specifically snowfall and snow storages, significantly affect the seasonal water balance and rainfall-runoff process (Langhammer and Bernsteinová, 2020). Mountain areas behave as a complex system where the total catchment runoff is a mixture of different input components, namely rainfall, snowmelt, and potentially glacier melt. This mixing substantially affects the seasonal distribution of runoff during a water year. Many studies calculated the relative importance of snow in generating runoff compared to rainfall (Berghuijs et al., 2014; Jenicek and Ledvinka, 2020a; Meriö et al., 2019) resulting in the fact that a higher fraction of snow-generated runoff leaves the catchment than what would correspond to the snowfall fraction (Li et al., 2017). This might be explained, e.g., by the lower winter evaporation causing the higher winter runoff coefficients compared to the summer period, lower water demand by vegetation, and differences in soil moisture and runoff generation during snowmelt compared to rainfall (Barnhart et al., 2016). Nevertheless, the decreasing snowfall and thus lower runoff inputs from snowmelt during spring may be offset by increasing rain inputs to the total runoff during winter (Hammond and Kampf, 2020). This increase in winter runoff is predicted also for central Europe, including Czechia (Hanel et al., 2012; Lamačová et al., 2014; Zheng et al., 2021), although the water balance in this area is expected to change mostly in summer due to the increase in evapotranspiration which mostly contributes to changes in summer low flows (Lamačová et al., 2014; Vlach et al., 2020).

Snowmelt also strongly affects soil moisture and groundwater recharge and thus it influences the catchment storage, runoff seasonality and baseflow. Snowmelt may control relative partitioning of the water between evapotranspiration and groundwater recharge leading to higher runoff (Barnhart et al., 2016). Decreasing snowfall fraction and thus lower and earlier snowmelt causes decrease in early summer soil moisture (Potopová et al., 2016). Higher elevations areas play an important role in the catchment storage as they store a lot of water (Floriancic et al., 2018; Hood and Hayashi, 2015; Šípek et al., 2021; Staudinger et al., 2017) and thus support the streamflow at lower elevations which is important mainly during low flow periods (Cochand et al., 2019). The groundwater recharge in upper parts of the snow-dominated catchments appears decoupled from year-to-year variations in snow storages and thus those areas are often resilient to droughts (Carroll et al., 2019).

The ongoing changes in snow storages, snowmelt and changes in spring and summer runoff, including low flows are at the centre of recent research (Blahušiaková et al., 2020; Dierauer et al., 2018; Li et al., 2018; Van Loon et al., 2015; Vlach et al., 2020). Higher snowpack may lead to higher baseflow (Hammond et al., 2018) which influences summer low flows. Together with later melt-out for years with higher snowpack, the summer low flows may be higher than for years with lower snowpack. This is especially valid for the Mediterranean climate with uneven precipitation distribution between winter and summer (Dierauer et al., 2018; Godsey et al., 2014). Nevertheless, the positive effect of higher snowpack for summer low flows may be seen also for humid catchments with precipitation more equally distributed in the water year (Jenicek and Ledvinka, 2020a), although summer rainfall and evapotranspiration is certainly more important (Cooper et al., 2018; Floriancic et al., 2020; Šípek et al., 2021). With future climate changes, the positive effect of snow on summer runoff and low flows may decrease at higher elevations and completely disappear at middle and low elevations. Consequently, the low flows might drop to even lower values not only because of higher evapotranspiration and potential decrease in precipitation, but also due to lower seasonal snowpack.

Although many studies quantified the impact of climate changes on the water cycle, less attention has been paid to how future snow changes influence summer runoff and low flows considering a wide range of hydrological responses to different climate projections. Specifically, due to a complexity of the precipitation-runoff processes, hydrological responses of catchments to changing climate conditions may be more variable than the climatic changes themselves, especially in snow-dominated catchments and thus they may be surprising and unexpected. To investigate these hydrological responses may be specifically important for the region of central Europe because existing climate projections available for this region do not agree in the prediction of precipitation, simulating either its increase or decrease (Svoboda et al., 2016). Therefore, it is important to analyse whether the increase in air temperature causing the decrease in snowfall might be partly compensated by the potential increase in winter precipitation. This is also important since snow may represent an important factor of summer low flow severity at least for snow dominated catchments, although the snow is usually not a dominant factor in driving summer low flows in humid climates. Therefore, it is important to study whether the seasonal snow is able to partly compensate the low summer precipitation and thus contribute to higher low flows and how it will change in the future. All the above points guided the main objectives of this study which were 1) to simulate the future changes in snow storages for a large set of mountain catchments, representing different elevations, 2) to assess the mutual interplay of changing air temperature and precipitation for snow accumulation reflecting a wide range of climate projections and 3) to analyse how the changes in snow storages will affect groundwater recharge, streamflow seasonality and low flows in the future. We benefit from a multi-catchment level approach which enabled us to generalize the results for a broader region. In Czechia, there were only few studies addressing the topic of hydrological impact of future climate changes (Hanel et al., 2012; Lamačová et al., 2014). These studies were focused on overall changes in water balance (including winter runoff caused by snowmelt), but the study investigating the changing effect of snow on spring and summer low flows is still missing. With this study we follow up with our previous studies carried out in the same study area, which were focused on studying the inter-annual variability of the snow influence on summer low flows based on historical data (Jenicek and Ledvinka, 2020a) and the role of catchment storage in this process (Šípek et al., 2021).

2. Data and methods

2.1. Study area and data

We selected 59 catchments located at higher elevations in Czechia with a substantial influence of snow on streamflow (Fig. 1, Table S1, and kml file in Supplementary data section). The selected catchments have a minor influence of human activity on runoff. The selection covered most mountain ranges in Czechia showing the variety of natural conditions (Table 1). The mean catchment elevation ranges from 491 to 1297 m a.s.l and the mean slope ranges from 2.2 ° to 13.0 °. The annual maximum SWE ranges from 35 mm for the lowest catchments to 664 mm for the highest catchments, and the snow contributes from 16.6–41.3% to runoff. The catchment areas range from 1.8 km^2 to 383.4 km^2. The same catchment selection was used in a previous study (Jenicek and Ledvinka, 2020a).

Daily precipitation, daily mean air temperature and weekly SWE data were available from stations located within or nearby the individual catchments for the period 1980–2014 (except for three catchments with data that started from one to three years later). The data were obtained from the Czech Hydrometeorological Institute (CHMI). The temperature-based method defined by Oudin et al. (2005) was used for calculation of potential evapotranspiration (PET) for individual climate stations. The above data, together with daily runoff data for all outlet profiles were used to calibrate and evaluate the rainfall-runoff model (the nearest station with respective climate data was used for each catchment, see also kml file in Supplementary data section).

2.2. HBV model

To simulate the individual components of the water cycle and to bring the information from point climatological measurements to a catchment scale, we used a conceptual bucket-type HBV model (Lindström et al., 1997) in its implementation HBV-light (Seibert and Vis, 2012). The model consists of four routines, namely 1) the snow routine, which uses a degree-day approach including potential refreezing of meltwater calculation, 2) the soil routine with a calculation of groundwater recharge and actual evaporation (AET) simulated as functions of the actual soil moisture, 3) the response routine, which calculates the catchment runoff using two groundwater boxes and 4) the routing routine, which applies a triangular weighting function for runoff propagation to catchment outlet. Details on the model structure and routines used to calculate the individual runoff components can be found in several studies (Jenicek et al., 2018; Seibert and Vis, 2012).

The catchments were sub-divided into elevation zones by 100 m to distribute the information about air temperature, accumulated snowpack, snowmelt and related soil moisture and groundwater recharge according to elevation. The calibration of the HBV model was done for a previous study (Jenicek and Ledvinka, 2020a). We refer the reader to this study for more details regarding the calibration process and results. In general, the HBV model was calibrated for each catchment using a genetic algorithm procedure which is

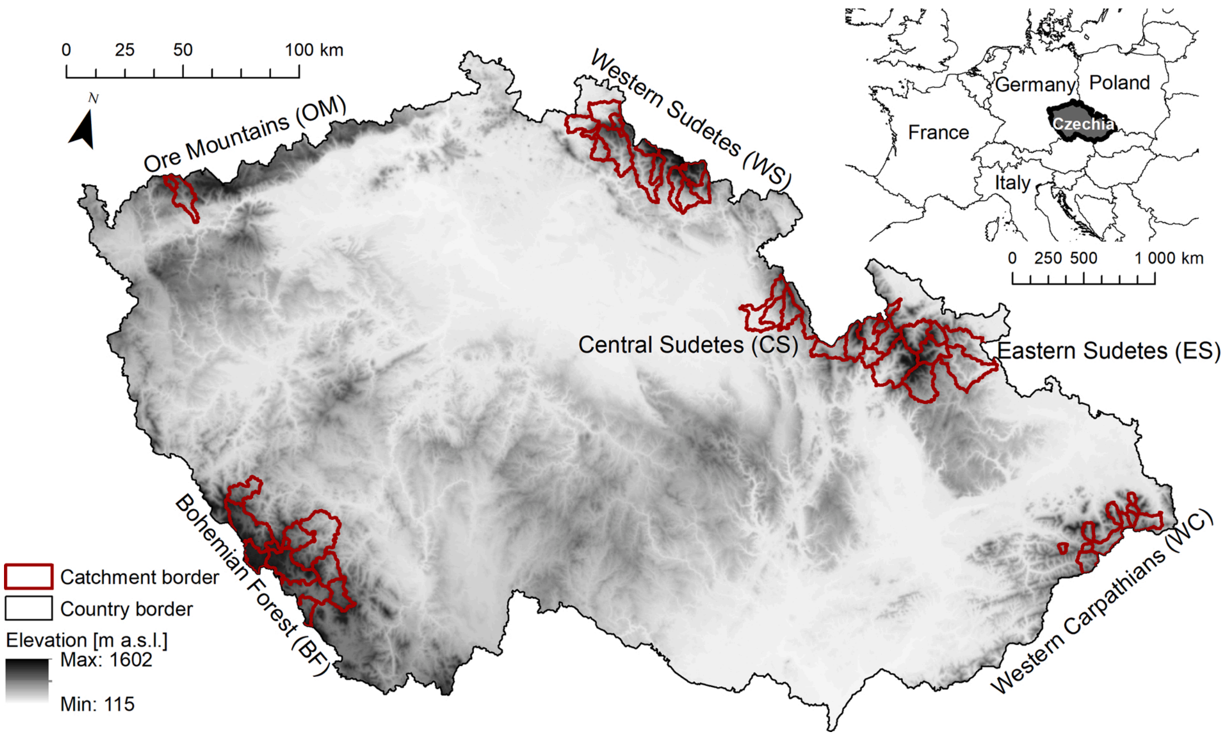

Fig. 1. Geographical location of the study catchments (Jenicek and Ledvinka, 2020a).

Table 1
Basic catchment characteristics of the study data set. Snowfall fraction, SWE_{max} and snowmelt contribution to runoff represent catchments means (1980–2014) resulting from model simulations.

Catchment characteristic	Mean	Min	Max
Area [km^2]	107.5	1.8	383.4
Mean elevation [m a.s.l]	790	491	1297
Mean slope [°]	6.3	2.2	13.0
Mean annual air temperature [°C]	5.7	2.2	8.4
Mean snowfall fraction [-]	0.20	0.12	0.38
Mean SWE_{max} [mm]	141	35	664
Mean snowmelt contribution [%]	26	17	41

implemented in the model (Seibert, 2000). The calibration period was set to 1980–1997 and the validation has been done using a period 1998–2014. The two periods included both wet and dry years and cold and warm years. The multi-variable model calibration was used using a combination of three criteria; 1) Nash-Sutcliffe efficiency for runoff using logarithmic values (Nash and Sutcliffe, 1970), 2) volume error for runoff and 3) model efficiency for SWE. The calibration of the model against both runoff and SWE enabled to better control the simulation of the SWE in individual catchments and improve the model's ability to simulate individual components of the water cycle. One hundred calibration runs resulting in one hundred parameter sets was performed to address the uncertainty of the model parameters. These one hundred sets served to create one hundred simulations. A median simulation was used for further analyses in this study. The results of the model calibration and validation are shown in the supplementary material which shows the values of objective functions for calibration and validation period for each catchment (Fig. S1) and the comparison of observed and simulated SWE (Fig. S2) and flow duration curves (Fig. S3) for six selected catchments, one per each mountain range. Additionally, Fig. S4 shows comparison of observed and simulated SWE and runoff for two selected hydrological years, first with extremely snow-rich conditions (2006), and second with extremely snow-poor conditions (2014). Resulting simulations are publicly available (Jenicek and Ledvinka, 2020b).

2.3. *Simulations of the future climate and runoff*

The impact of future changes in climate variables on snow storages and streamflow in the study catchments was assessed using climate simulations from four global circulation models (GCMs) dynamically downscaled by six regional climate models (RCMs). Three Representative Concentration Pathways (RCP 2.6, 4.5 and RCP 8.5) were chosen to capture the variability of the future climate evolution including the RCP 2.6 which limits global warming to between 1.5 °C and 2 °C, compared to the pre-industrial period which is reflected by the Paris Agreement (UNFCCC, 2015). Data were downloaded from the European Domain of the Coordinated Regional Downscaling Experiment (EURO−CORDEX; https://www.euro-cordex.net/). In total, 17 unique combinations (further referred as climate chains) of GCMs, RCMs and RCPs were used in the study for which the data from the period 2020–2099 were available (Table 2). The time series for individual basins were derived as weighted averages of RCM grid boxes intersecting a basin with weights proportional to the intersection area.

The precipitation and temperature data from climate models were bias corrected using a multivariate approach by Piani and Haerter (2012). In this approach, a correction of temperature is performed as the first step. Subsequently the pairs [corrected temperature, uncorrected precipitation] are split into several bins according to temperature quantiles and a correction of precipitation is performed separately within each bin. In this study, the number of bins was obtained following the Sturges's formula. The precipitation

Table 2
Climate chains used in the study. Last column indicates the boundary climate chains used for the more detail assessment.

Climate chain No.	Forcing GCM	RCM	RCP	Ranked as
1	CNRM-CM5	CCLM4−8-17	4.5	
2	CNRM-CM5	ALADIN53	4.5	
3	CNRM-CM5	RCA4	4.5	
4	CNRM-CM5	ALADIN53	8.5	
5	CNRM-CM5	RCA4	8.5	
6	EC-EARTH	RCA4	2.6	
7	EC-EARTH	CCLM4−8-17	4.5	
8	EC-EARTH	RCA4	4.5	
9	EC-EARTH	RACMO22E	4.5	medium snow-rich
10	EC-EARTH	HIRHAM5	4.5	
11	EC-EARTH	CCLM4−8-17	8.5	
12	EC-EARTH	RCA4	8.5	snow-poor; snow-poor and dry
13	EC-EARTH	HIRHAM5	8.5	
14	IPSL-CM5A-MR	WRF331F	4.5	
15	MPI-ESM-LR	CCLM4−8-17	4.5	snow-rich
16	MPI-ESM-LR	RCA4	4.5	snow-rich and dry
17	MPI-ESM-LR	RCA4	8.5	

thresholds were determined and applied specifically for each bin to prevent the drizzling effect. For both temperature and precipitation, a precise transfer functions given by the quantile-quantile plot were replaced by their linear fits in order to increase the robustness with respect to changing climate conditions. A linear approximation of the transfer functions is justified by the underlying probability distributions of both variables. In the case of temperature, the gaussian distribution is considered as an appropriate model, which implies that a required shift of the probability distribution can be achieved by a linear transformation of the dataset. The same linear nature can be found in the case of precipitation, where the gamma distribution is a generally acceptable model. It was shown by Hnilica and Pus (2013) that for precipitation data a precise quantile mapping transfer function can be well approximated by a linear function. An explanation based on a theoretical background can be found in Hnilica et al. (2017). The bias corrections were calibrated separately for the warm (May-Oct) and the cold (Nov-Mar) periods using the available overlap of the control RCM run (specific for each RCM) and the observed data from the period 1980–2014.

Simulations of the future hydrological behaviour of the study catchments as a response to changing climate were compared for three future periods 2020–2049, 2045–2074 and 2070–2099, relative to the reference period 1980–2005. In order to keep data consistent and comparable, for the reference period we used reanalyses of the individual climate chains which were corrected using the same procedure as for the future simulations described above. Finally, both the reference and future climate data were submitted to the calibrated HBV-light hydrological model which was used to simulate the individual water balance components in the study catchments.

2.4. Snow and streamflow signatures and results analysis

To analyse the potential influence of snow on catchment runoff, several snow, groundwater and streamflow signatures were calculated in this study (Table 3).

To better capture the variability of the hydrological responses to different climate inputs, we identified climate chains according to whether they lead to 1) snow-rich or snow-poor conditions, 2) wet or dry conditions and 3) precipitation-rich or precipitation-poor conditions. These conditions were assessed for the future period 2070−2099 using a combination of several characteristics. For identifying snow-rich/snow-poor climate chains, a combination of annual snowfall, SWE_{max} and DOY_{meltout} was used. For identifying wet and dry climate chains, Q_{JJA}, Q_{b} and D_{v} were used. Finally, summer (JJA) precipitation was used to identify the climate chains with the highest anomalies in summer precipitation. The above characteristics were used to rank the individual climate chains.

The climate chains ranked as the most snow-poor and snow-rich according to procedure described above were used for most of the analysis in this study. For determining the potential relationship between snow and summer low flows, the approach of selecting chains leading to snow-poor and snow-rich conditions cannot be used. This is because both climate chains are different in terms of summer precipitation, and thus the effect of snow storages on summer runoff cannot be separated from the effect of summer precipitation, which is more important. Therefore, we selected the two climate chains which both led to dry summer conditions (ranked as 1st and 3rd driest from all considered climate chains) caused by low summer precipitation (the selected climate chains were the two with the lowest summer precipitation). The first chain led to the most snow-poor conditions from all climate chains (further referred as snow-poor and dry) and the second selected chain led to the second most snow-rich conditions from all climate chains (further referred as snow-rich and dry). The selected climate chains are highlighted in Table 2. With this selection we wanted to analyse whether snow-rich conditions might lead to less dry summer conditions despite the fact that summer precipitation is comparable (and low) in both of the selected climate chains.

Table 3
Snow, groundwater and streamflow signatures used in analyses.

Signature	Description
Snowfall fraction (S_{f})	A fraction of annual snowfall to annual precipitation; single threshold temperature was calibrated individually for the specific catchment (values ranges from -1.6 °C to 1.1 °C)
Annual maximum of SWE (SWE_{max})	Annual SWE maximum calculated from February to April
Day of year of melt-out (DOY_{melt})	A day with the first occurrence of SWE below the threshold of 10 mm after the day with SWE_{max}
Snow cover duration (S_{dur})	Number of days with snow cover on ground with SWE higher than 5 mm
Seasonal groundwater recharge (G_{w})	The amount of water from the soil storage which contributes to the groundwater storage simulated by the HBV model (Seibert and Vis, 2012)
Snowmelt runoff (Q_{s})	Contribution of snowmelt runoff to total runoff simulated by the model using an “effect tracking” algorithm assuming complete mixing (Weiler et al., 2018)
Snowmelt runoff fraction (Q_{sf})	A fraction of the annual snowmelt runoff to the total annual runoff
Seasonal runoff (Q)	Winter (Dec-Feb; DJF), spring (Mar-May; MAM) and summer (Jun-Aug; JJA) runoff
Annual minimum runoff (Q_{min})	Minimum of 7-day moving average of the runoff during the warm period (May-Oct)
Base flow (Q_{b})	Summer (JJA) baseflow calculated by the HBV model as an outflow from the lower groundwater box which is a part of the model response routine controlled by a recession coefficient (Seibert and Vis, 2012)
Deficit volume (D_{v})	A water volume lacking in rivers for Jun-Aug below the defined threshold (90$^{\mathrm{th}}$ percentile of the flow duration curve). A variable level threshold method was used with different thresholds for individual months
Deficit volume duration (D_{vdur})	Number of days below runoff threshold (90$^{\mathrm{th}}$ percentile of the flow duration curve)

3. Results

3.1. Changes in snow signatures at different elevations and for different climate chains

Future simulations showed a considerable decrease in snowfall fraction for all catchments at all elevations (Fig. 2a) followed by a large decrease in SWE_{max} from 30 % to 70 % until the end of the 21st century with higher values below 1100 m a.s.l (Fig. 2b). In addition, the melt-out day occurred earlier by four weeks at the highest elevations and by 2–3 weeks at the lowest elevations (Fig. 2c). This also showed that the snow-covered season will be significantly shorter in the future (Fig. 2d). The bottom panels of the figure indicate that the shortening of the snow-covered season will be caused more by earlier melt-out rather than by later snow cover onset. Fig. 2 also shows that changes in SWE_{max} and DOY_{melt} are, in absolute terms, larger at higher elevations compared to lower elevations, suggesting larger changes in the seasonal water balance at these higher elevations. In contrast, snow changes are mostly lower for lower elevations due to overall low snow storages.

Due to the large variability in simulated snow storages for individual climate chains, we assessed changes in snow signatures for the future period 2070−2099 for three selected climate chains leading to the most snow-poor, mean, and snow-rich conditions (see methods). The results show that despite the large decrease in S_f, SWE_{max}, DOY_{melt} and S_{dur} was simulated for the mean conditions, the differences between the two border conditions are relatively large (Fig. 3). The relatively small decrease in all selected snow signatures was simulated for the climate chain leading to snow-rich conditions, which is also represented by the highest precipitation during the cold period (the precipitation increased by 10.2 % relative to the reference period; results not shown). Besides, the increase in air

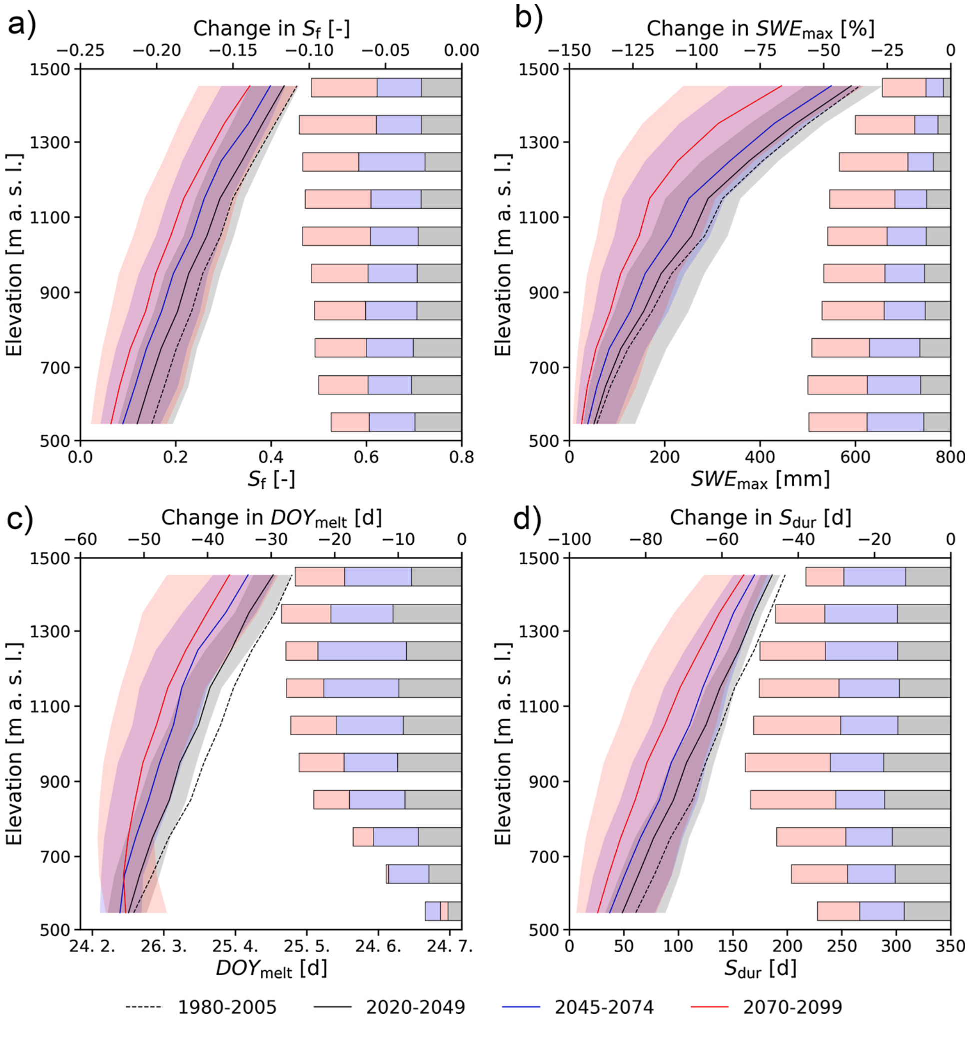

Fig. 2. Changes in (a) snowfall fraction (S_f), (b) annual maximum SWE (SWE_{max}), (c) day of year of melt-out (DOY_{melt}) and (d) snow cover duration (S_{dur}) at different elevations for the reference period and three future periods. Lines related to the bottom x-axis show the mean value from all climate chains with coloured area indicating climate chain variability. Bars related to top x-axis show relative differences from the reference period (a mean from all climate chains).

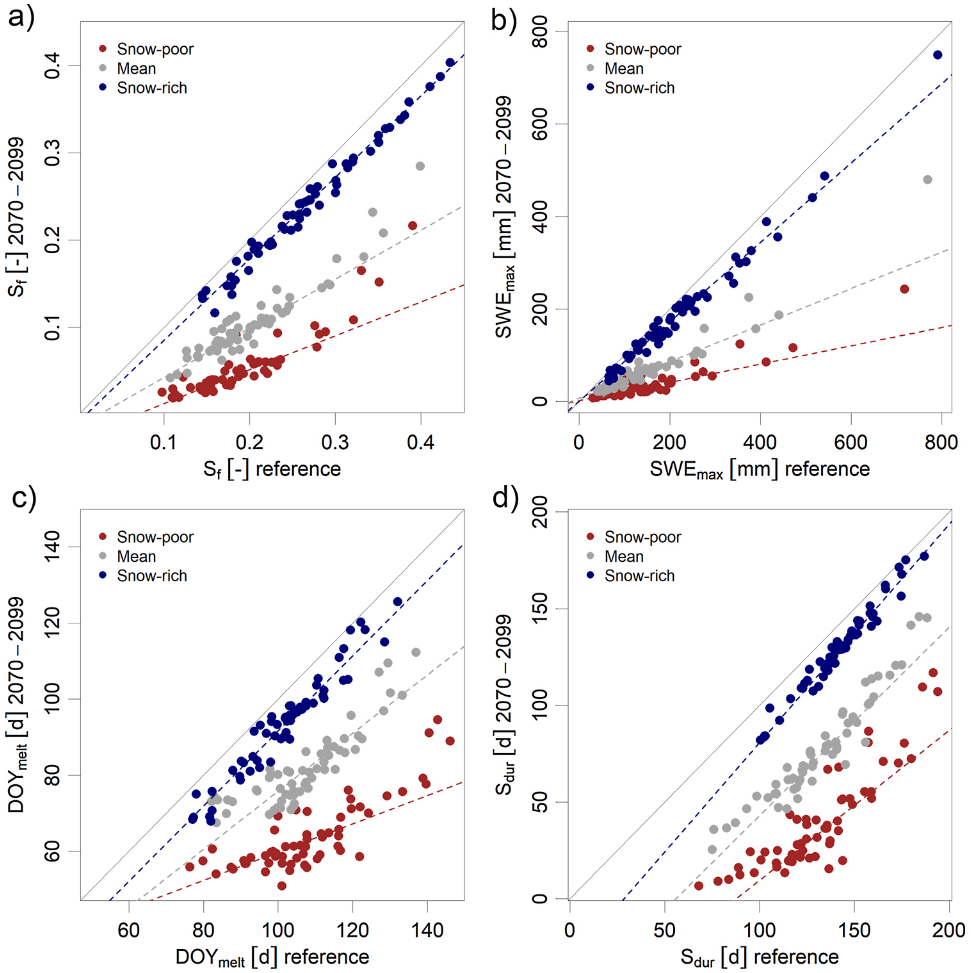

Fig. 3. (a) Changes in mean snowfall fraction (S_f), (b) mean annual SWE maximum (SWE_{max}), (c) mean DOY of melt-out (DOY_{melt}) and (4) mean snow cover duration (S_{dur}) in individual study catchments for climate chains leading to snow-poor (brown points) and snow-rich (blue points) conditions and for the mean conditions (grey points) for the future period 2070-2099 compared to the reference period. Dashed lines represent Theil-Sen regressions (for interpretation of the references to colour in the Figure Legend, the reader is referred to the web version of this article).

temperature for the cold period compared to the reference period is less dramatic compared to other climate chains (by 1.4 °C compared to the mean temperature increase by 2.8 °C). In contrast, the climate chain leading to snow-poor conditions showed a dramatic decrease in all selected snow signatures. This snow-poor climate chain had the largest increase in air temperature of 4.3 °C for the cold period, and one which led to the driest conditions with annual precipitation 8% lower compared to the climate chain mean. The above results indicated that the increase in air temperature causing the decrease in snowfall might be partly compensated by the increase in winter precipitation. Despite this potential compensation, none from the 17 climate chains led to a future increase in snow storages or later melt-out compared to the reference period.

The results in Fig. 3 also showed that the decrease in snow signatures is, in absolute terms, higher for catchments with higher S_f, SWE_{max} and later DOY_{melt} which is indicated by the increasing distance between brown and blue points from lower to higher values. Since the mentioned signatures are correlated with elevation (e.g., for S_f, the Pearson correlation coefficient is 0.53 with p value<0.001), the results suggested that that snow at highest elevations is more affected by the increase in air temperature in our study region.

For mitigation and adaptation strategies, it is important to look at how snow signatures will change in the future for different representative concentration pathways (RCP 2.6, 4.5 and 8.5) reflecting future evolution of greenhouse gases concentrations and their radiative forcing. The results of the most optimistic RCP 2.6 scenario showed the relatively smaller decrease in snow signatures compared to the moderate RCP 4.5 scenario (Fig. 4). The most pessimistic RCP 8.5 scenario showed significantly larger decreases in all selected snow signatures, compared to the reference period, due to the largest increase in air temperature (by 3.7 °C by the end of the 21st century compared to the reference period for the study catchments). For example, S_f will decrease on average by 0.13, it means by 60 %, for the RCP 8.5 for the period 2070−2099 (Fig. 4a). For RCPs 2.6 and 4.5, this average decrease is between 0.04−0.07, representing a decrease of 30–40 %. The simulated future decrease in S_f caused large decreases in SWE_{max} (of 29 % for the RCP 2.6 and 68

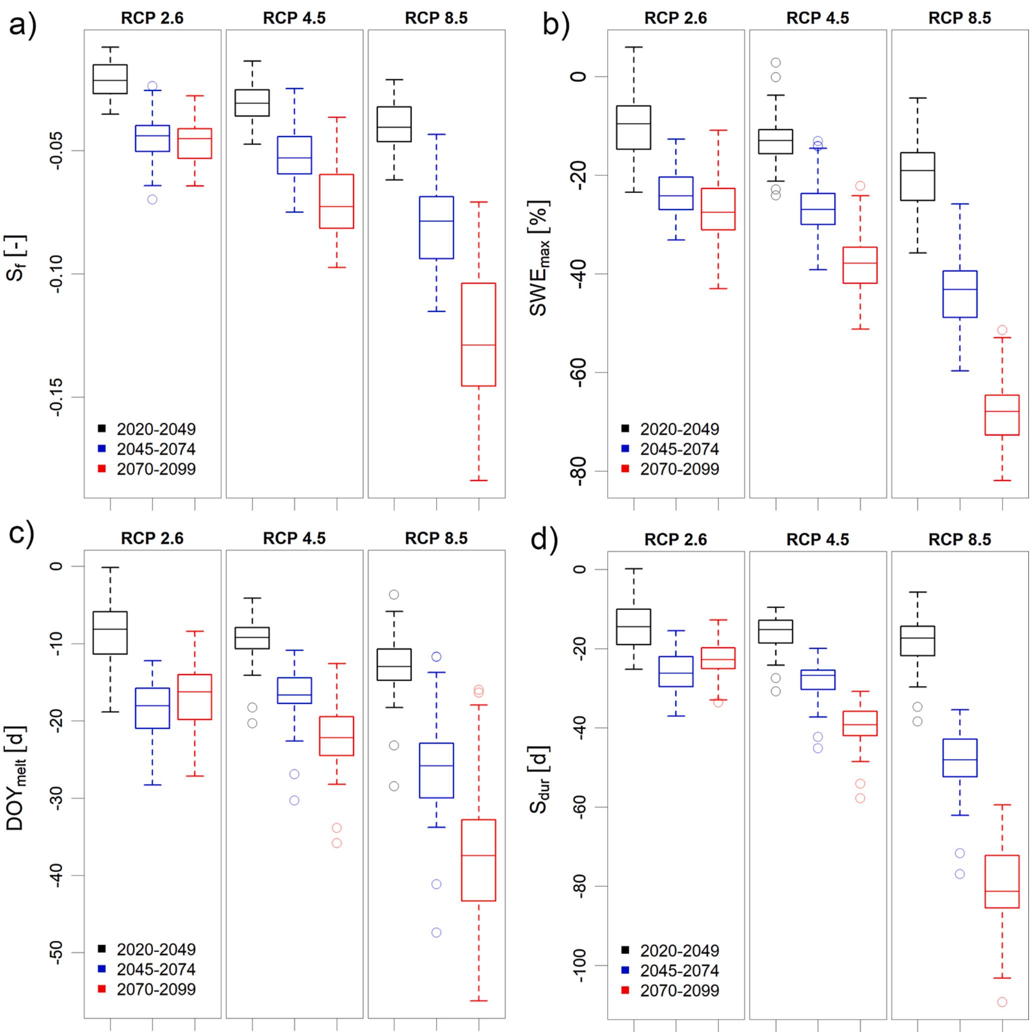

Fig. 4. (a) Change in S_f, (b) SWE_{max}, (c) DOY_{melt} and (d) S_{dur} in the study catchments for different RCPs (2.6, 4.5 and 8.5) for future periods: 2020-2049 (black boxes), 2045-2075 (blue boxes) and 2070-2099 (red boxes). Individual boxes represent the 25th and 75th percentiles (median as a thick line), whiskers represent 1.5 times the interquartile range.

% for RCP 8.5 for the period 2070–2099; Fig. 4b), earlier melt-out (on average 17 days earlier for the RCP 2.6, and 38 days earlier for the RCP 8.5; Fig. 4c) and shorter snow cover duration (by 22 and 82 days, respectively; Fig. 4d). These results clearly show that although the decrease in snow storages and the shortening of the snowmelt period was simulated for all RCPs, the RCP 2.6 predicted much lower decreases in snow compared to the most pessimistic RCP 8.5 scenario.

3.2. Changes in seasonal runoff distribution

The above results showed that the seasonal snowpack is highly sensitive to changes in climate indices. The simulations of the future runoff showed that the changes in snow storages significantly affected seasonal runoff distribution in individual catchments (Fig. 5). The figure shows the seasonal runoff and its relative change for six selected catchments (one catchment from each mountain range as defined in Fig. 1). In the reference period, the largest runoff occurred in April or May depending on the catchment elevation. For the future period 2070–2099, the period of highest streamflow will occur on average a month earlier following the earlier snowmelt, and the seasonal runoff volume is significantly lower in this period due to less water originating from snowmelt. Additionally, the model predicted an increase in winter runoff for the future period due to an increase in air temperature and thus more liquid rain and partial snowmelt during the winter season will occur. Consequently, the decrease in spring runoff will be compensated by the increase in winter runoff in most of the catchments. Nevertheless, this compensation is also caused overall by a slight increase in winter precipitation predicted by most of the climate chains. The changes in runoff in the period June-October are not as large as in the case of winter and spring runoff. In many catchments, June to October runoff will not change or slightly increase, mostly due to the small increase in precipitation predicted by most of the climate chains.

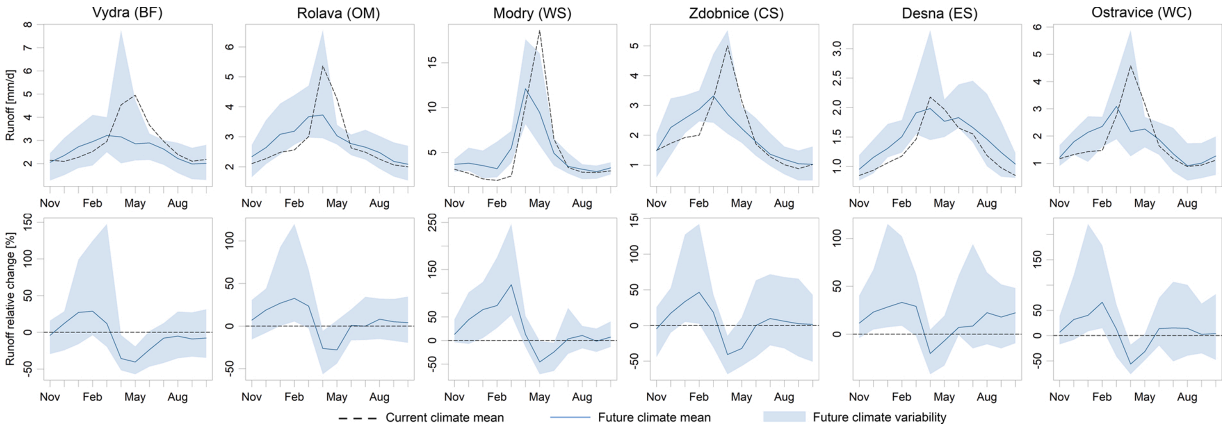

Fig. 5. Monthly runoff in six selected catchments for the reference period and for the future period 2070-2099 (top panels), and relative changes in monthly runoff for the future period 2070-2099 compared to the reference period (bottom panels). Black dashed line indicates reference period, blue line represents future period 2070-2099, light blue area indicates the range of individual future climate chains. Note the different scales used for the y-axis (for interpretation of the references to colour in the Figure Legend, the reader is referred to the web version of this article).

The results of comparison of the future snow runoff contribution (Q_{sf}) and consequent spring runoff for all climate chains sorted according to individual RCPs, showed overall a decrease in both runoff signatures (Fig. 6). For RCP 2.6, the decrease in Q_{sf} is lower compared to RCP 4.5. The RCP 8.5 showed much larger decreases in Q_{sf}, especially for middle and far future periods compared to the reference period. The reason for such a decrease lies in the decrease in snow storages shown in Fig. 4. Similar to the decrease in Q_{sf}, the March to May runoff decreased as well. The decrease in spring runoff is largest for RCP 8.5, but lower for RCPs 2.6 and 4.5, probably because the expected decrease due to lower snow storages was compensated by the small increase in total precipitation in most of the climate chains. Overall, future changes in snow storages will be the primary cause for changes in winter and spring runoff in most of catchments.

3.3. *The effect of snow on summer runoff and low flows*

As explained in the methods, the approach of selecting snow-poor and snow-rich climate chains cannot be used to show the effect of snow impact on summer streamflow and low flows since it is not possible to separate the effect of snow storages from the effect of summer precipitation. To answer the question of whether or not the snow is an important driver of summer runoff, we selected two climate chains which were ranked as the driest (see methods) and which have a similar amount of summer precipitation (the lowest summer precipitation from all climate chains), but the first chain led to snow-poor conditions and the second chain led to snow-rich

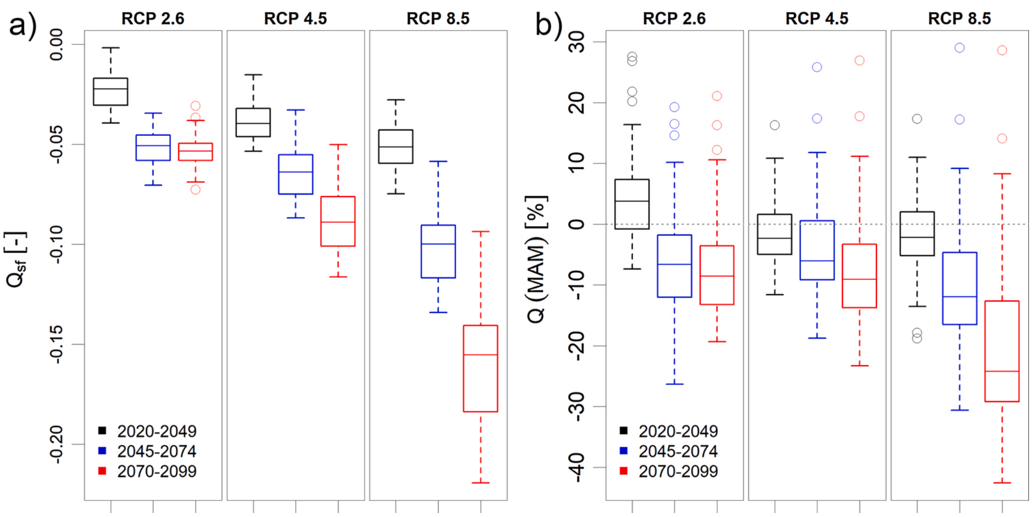

Fig. 6. (a) Change in Q_{sf} and (b) Q (MAM) in the study catchments for RCPs 2.6, 4.5 and 8.5 in future periods 2020-2049 (black boxes), 2045-2075 (blue boxes) and 2070-2099 (red boxes). Individual boxes represent the 25th and 75th percentiles (median as a thick line), whiskers represent 1.5 times the interquartile range (for interpretation of the references to colour in the Figure Legend, the reader is referred to the web version of this article).

conditions. The basic climate and snow characteristics of both selected climate chains are shown in Table 4.

The results displayed in Fig. 7 showed that the decrease in snow storages for the future period compared to the reference period caused lower snowmelt runoff volume and thus lower spring runoff (Fig. 7a-b). In addition, lower snow storages caused lower groundwater recharge during the cold period (Fig. 7c). This lower groundwater recharge is an important indicator of available water in the groundwater zone which contributes to the river streamflow during the summer period, especially during low flows.

The large difference in spring runoff for both selected climate chains were expected since the spring runoff is directly related to snowmelt. Nevertheless, differences can be seen also for summer runoff and low flow signatures (Fig. 7d-h), despite the fact that summer precipitation is almost the same in both climate chains. Summer runoff, summer baseflow and summer minimum runoff are on average lower for snow-poor and dry conditions compared to snow-rich and dry conditions, although the differences are rather small, and they change from catchment to catchment (Fig. 7d-f). Nevertheless, drier conditions for snow-poor and dry climate chain become more evident for higher elevation catchments where snow plays a more important role in generating the seasonal runoff. In these higher elevation catchments, the absolute decreases in snow storages are larger than in lower elevation catchments and thus the summer runoff in those catchments is more affected by changes in seasonal snowpack. This becomes clearer when looking on summer deficit volumes and their duration (Fig. 7g-h). Both characteristics are displayed as a difference between the 2070−2099 future period and the reference period. Both summer deficit volumes and their duration will significantly increase in the future, whereas the largest increase in deficit volumes (but not for their duration) seems to be associated with catchments with higher S_f.

To make the results of snow relevance for summer runoff more robust, we assessed all available climate chains to see whether there is a tendency of climate chains leading to relatively snow-rich conditions to generate higher summer runoff and low flows and vice versa. Fig. 8 showing the climate chain ranking suggests that snow is not important for mean summer (JJA) runoff volume (left panel). However, when deficit volumes are considered as an indicator of low flow extremity, the driest climate chains are associated with both the lowest summer precipitation and seasonal snowpack. It means that lower deficit volumes (= less serious droughts) were simulated for chains leading to more snow (light brown and blue colour in the bottom right quadrant compared to the bottom left quadrant). In other words, for the same low summer precipitation, the summer deficit volumes are lower for climate chains with higher snow storages. This indicated that snow can partly increase summer low flows, although the summer precipitation is clearly more important. The expected lower snow storages might therefore contribute to more extreme low flow periods. The same results as for deficit volumes were achieved also for the duration of the low flow periods (results not shown).

In addition to lowest low flows for both lowest precipitation and snow storages, Fig. 8 shows that the highest low flows (the lowest deficit volumes) were simulated for climate chains with mutual existence of high summer precipitation and snow storages (the darkest blue colour in the top right quadrant). In other words, the same amount of high summer precipitation led to higher low flows for climate chains, leading to more snow.

4. Discussion

4.1. *Changes in snow storages and their influence on seasonal runoff*

The results showed a considerable decrease in snow storages in the future in our study catchments. The snow decrease is expected at all elevations above 600 m a.s.l. Below this elevation threshold, the snow storages are low even in the current climate and thus changes are rather small. The results of this study support results achieved earlier in the study region which showed either past or predicted future changes in snow characteristics, snowmelt and low flows (Blahušiaková et al., 2020; Lamačová et al., 2014; Vlach et al., 2020). Our results also support results from other world regions with comparable climate (Marty et al., 2017b; Musselman et al., 2017b), although some of them showed smaller relative decreases in snow storages at the highest elevations not covered by our study selection (Fyfe et al., 2017; Jenicek et al., 2018).

For interpretation it is important to mention that all climate chains simulated decrease in snow storages, although in few of them, the decrease was very small. This means that no climate chain simulated higher snow storages compared to the reference period despite the fact that some of the chains predicts higher precipitation in the cold period. Therefore, it is clear that the increase in air temperature (predicted by all climate chains) is the major cause for the decrease in snow storages, although this decrease might be partly compensated by the increase in winter precipitation. The mutual interplay of both air temperature and precipitation in

Table 4
Selected climate and snow characteristics for 1) snow-poor and dry and 2) snow-rich and dry climate chains for the reference period and for the future period 2070-2099 for the study catchments.

Climate chain	Annual P [mm]	P cold period [mm]	P warm period [mm]	Annual T [°C]	T cold period [°C]	T warm period [°C]	S_f [-]	SWE_{max} [mm]	S_{dur} [d]
Snow poor and dry, reference	1130	518	612	5.5	−0.8	11.7	0.19	147	131
Snow-rich and dry, reference	1134	523	612	5.5	−0.7	11.6	0.21	176	131
Snow poor and dry, 2070−2099	1044	488	556	9.7	3.5	15.8	0.05	35	39
Snow-rich and dry, 2070−2099	1150	576	574	7.1	0.7	13.4	0.2	153	120

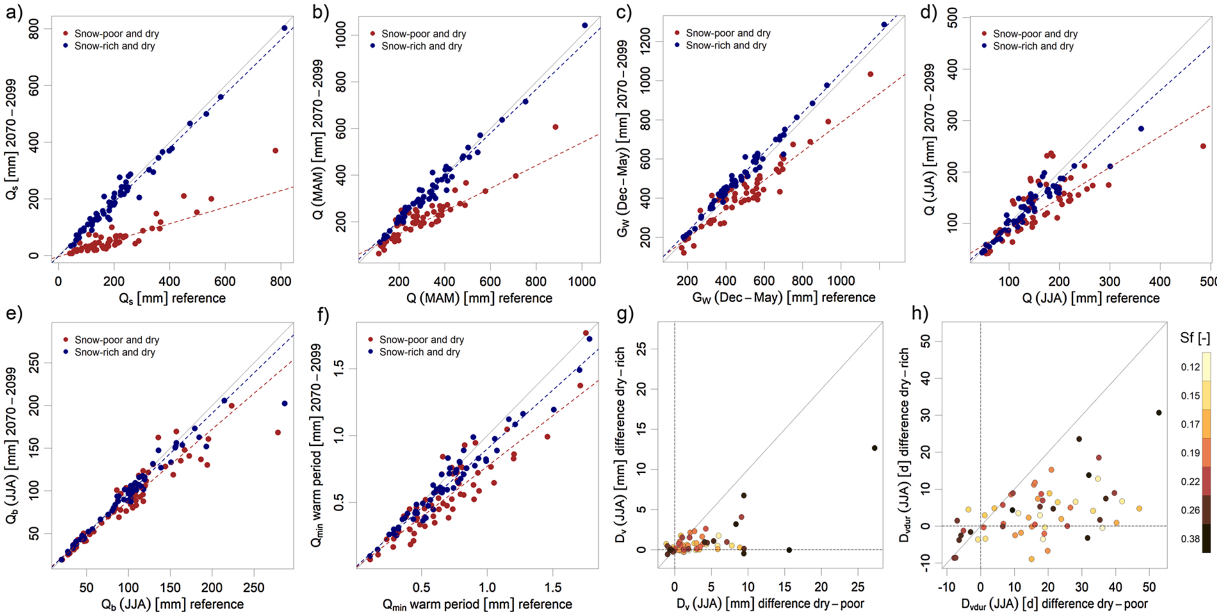

Fig. 7. Changes in selected runoff signatures for climate chains leading to 1) snow-poor and dry and 2) snow-rich and dry conditions in the study catchments for the future period 2070-2099 compared to the reference period; (a) Snowmelt runoff Q_s, (b) Spring runoff Q (MAM), (c) Groundwater recharge G_w, (d) Summer runoff Q (JJA), (e) Summer baseflow Q_b, (f) Minimum runoff Q_{min}, (g) Summer deficit volume D_v, and (h) Summer deficit volume duration (D_{vdur}). Dashed lines for (a) to (f) represent Theil-Sen regression lines. Colour scale for (g) and (h) used for snowfall fraction.

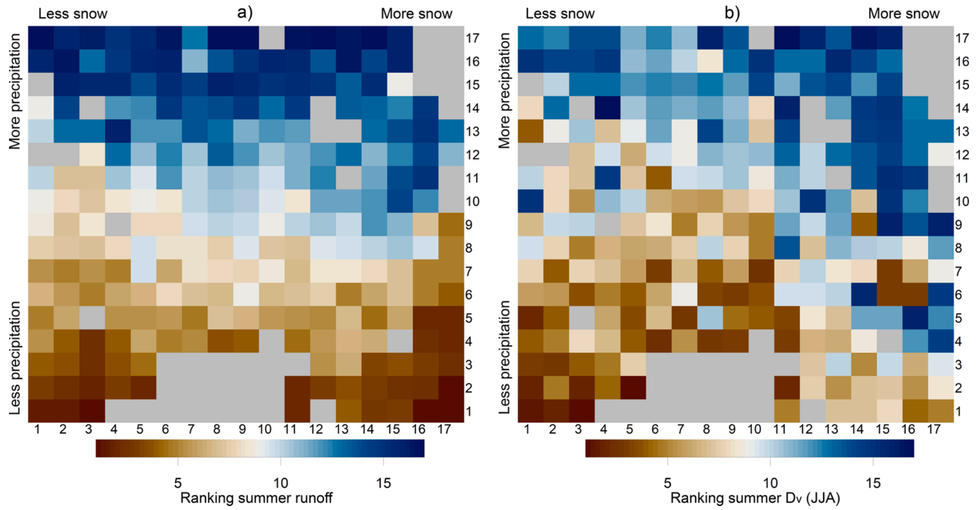

Fig. 8. Climate chain ranking for snow signatures (x-axis) against summer precipitation (y-axis) sorted from snow-poor or low precipitation (ranking 1), to snow-rich or high precipitation (ranking 17), for all study catchments for the period 2070-2099. Colour indicates ranking for summer runoff (panel a) and summer deficit volumes D_v (panel b). Ranking 1 means low summer runoff or high D_v, ranking 17 means high summer runoff or low D_v, respectively. Grey colour used for no combination.

influencing snow storage is also well documented by several studies which also showed that future changes in precipitation can be associated with changes in snowfall extremity (Lute et al., 2015; Marshall et al., 2020). Such changes in number and extremity of snowfall events, or amount of winter precipitation might also help to explain sometimes missing clear trends in selected snow signatures over the last few decades, specifically for snow cover onset (Klein et al., 2016).

The results also showed that the decrease in snow signatures is in absolute terms higher for higher elevations compared to lower elevations. This might also indicate that the overall water balance and the seasonal runoff distribution in higher elevation catchments will be affected more in the future than the water balance in lower elevation catchments. This might also affect downstream areas which strongly depend on water from higher elevation areas, especially during late spring and summer as shown by results from Swiss catchments (Brunner et al., 2019; Cochand et al., 2019).

Analysing the changes in winter runoff is important to assess the water amount generated from winter precipitation. This water is important for recharge and filling the groundwater storage. Overall, the results in this study showed that future changes in snow storages will be the primary cause for changes in winter and spring runoff in most of catchments. Additionally, climate chains leading to the lowest snow storages are also associated with lower amounts of winter precipitation (results not shown). This shows that future snow droughts will be associated with both increased in air temperature and reduced precipitation (Blahušiaková et al., 2020; Cooper et al., 2016; Dierauer et al., 2019).

Mountain regions are often referred to as "water towers", because they substantially affect hydrology of downstream areas. For example, mountains store a lot of water in the form of snow and ice which contributes to runoff in lower elevation areas, even during the warm season with overall lower water availability due to high evapotranspiration and vegetation demand. Therefore, any change in seasonal runoff means a potential problem for environmental systems (such as river ecology) as well as for many human activities, such as hydropower, irrigation, industry or drinking water supply (Brunner et al., 2019). The decrease in snowmelt runoff might be also important for reservoir management in providing enough water for human activities. Additionally, changes in the water cycle and the main driving forces causing low flows might also be challenging for drought forecasting systems (Livneh and Badger, 2020).

4.2. Influence of snow on summer low flows

Despite the fact that most of the climate chains led to a dramatic decrease in snow storages, the results did not clearly show whether the reduced snowpack will also lead to overall dry or wet summer conditions. Although the climate chain leading to snow-poor conditions led also to the driest conditions, the climate chain that led to snow-rich conditions was not ranked as one which significantly leads to overall wet conditions. In other words, climate chains which led to higher snow storages do not necessarily lead to overall wet conditions during warm periods.

However, the results showed that future snow-rich conditions are associated with higher low flows (and thus lower deficit volumes) compared to the snow-poor conditions with a similar amount of summer precipitation. Nevertheless, the summer precipitation is the dominant driver for summer low flows as also shown by other studies (Cooper et al., 2018; Floriancic et al., 2020). The expected lower snow storages might therefore contribute to more extreme low flow periods as indicated by several studies (Godsey et al., 2014; Jenicek and Ledvinka, 2020a; Šípek et al., 2021). Additionally, some climate chains which led to low snow storages simulated lower low flows despite the relatively high summer precipitation.

Due to the methodology used, the simulated decreases in streamflow signatures do not account for an increase in PET due to an increase in air temperature. This was motivated by the temperature-based equation for PET which was used to calculate the input data for the HBV model. It was shown that such equations often overestimate the effect of PET increase, which means that the PET is more sensitive to changes in air temperature than it is expected in reality (Kingston et al., 2009; Milly and Dunne, 2011; Oudin et al., 2005; Shaw and Riha, 2011). Besides a clear limitation of this approach, it enabled to separate the effect of snow from the effect of increasing PET on the low flows extremity as also used in previous studies (e.g. Jenicek et al., 2018). Consequently, changes presented in this study are caused only by changes in snow and related changes in soil storages and thus AET. Therefore, the results of this study cannot be fully interpreted as expected future behaviour of the catchments, but the study rather shows the sensitivity of the catchments to changes in air temperature, precipitation and related changes in snow storages and runoff. Due to expected PET increase in the future which might partly result in increased AET (at least for humid regions with $P >> PET$), the future decrease in summer streamflow (and lower low flows) might be even larger than simulated by the hydrological model in this study.

Unlike the PET, the simulated AET may change using our modelling approach due to the changes in soil moisture because the HBV model assumes an increasing AET with increasing soil moisture. Additionally, the model calculates AET only for days without snow cover on the ground. Consequently, the simulated AET increased in future scenarios in the cold period because the model simulated lower snow cover duration and thus more days when AET occurred. This increase in AET was partly compensated by the decrease in AET due to lower soil moisture in the summer, especially in lower elevation catchments with lower precipitation and higher PET (results not shown).

4.3. The role of different RCPs on snow and runoff variability

However, despite the clear tendency of snow signatures changes for the climate scenario mean, large variability exists between individual climate chains, which reflects the large variability in climate inputs for individual GCM, RCM and RCP combinations. Therefore, it is important to assess the snow differences simulated by different climate chains. For example, for mitigation and adaptation strategies, it is important to look at how the snow signatures will change in the future for different representative concentration pathways (RCP 2.6, 4.5 and 8.5) reflecting future evolution of greenhouse gas concentrations and their radiative forcing. The most optimistic RCP 2.6 predicts the increase in air temperature between 1.5 °C and 2 °C until the end of the 21st century compared to the pre-industrial period, which means an increase of around 1 °C, compared to the reference period. Therefore, the RCP 2.6 is the scenario which most matches with the goals of the Paris agreement in 2015 (UNFCCC, 2015). In contrast, the RCP 8.5 is from today's perspective the most pessimistic scenario with a temperature increase of 4–5 °C until the end of the 21st century compared to the pre-industrial period.

The comparison of the future snow evolution for individual climate chains according to RCPs showed an overall decrease in snow storages and the shortening of the snow-covered period. For the RCP 2.6, however, the changes were somewhat lower compared to most of the climate chains considering RCP 4.5 and considerably lower compared to chains assuming RCP 8.5. The reason for the smaller change for the RCP 2.6 is mostly related to the smaller increase in air temperature. Nevertheless, our results also showed the importance of mutual interplay of both temperature and precipitation changes to affect the future snowpack. For example, the climate chain leading to the most snow-rich conditions did not use RCP 2.6, but that used RCP 4.5, although the differences between both climate chains were rather small and may be explained by higher winter precipitation predicted by the specific RCP 4.5 climate chain. This particular example highlights the importance of using a variety of GCMs and RCMs combined with more RCPs for hydrological projections to obtain more robust results (Chegwidden et al., 2019).

The simulated future changes in snow storages were the primary cause for changes in winter and spring runoff simulated for most of the study catchments. Overall, the results highlighted the importance of the effort of individual countries to reduce greenhouse gases emissions in order to keep warming well below 2 °C compared to the pre-industrial period as defined in the Paris agreement. This is important for mitigation and adaptation strategies related to climate change impacts in mountain regions (Marty et al., 2017a).

4.4. Uncertainty in climate projections and modelling approaches

It is well known that raw climate model outputs cannot be used directly in climate change impact studies. The inherent biases in model data are introduced by a low spatial resolution of climate models which leads to a simplified representation of important physical processes (Chen et al., 2015). Therefore, several post-processing (bias correction) methods have been proposed. The corrections are frequently calibrated and applied separately for individual meteorological variables. As a result, the correlation structures of the corrected data are biased, although individual corrected variables correspond to the observed data in their statistical indicators. Several studies recently attempted to overcome this limitation and consider also a dependence structure between (or within) variables in addition to basic statistical properties (Meyer et al., 2019). Such an approach has been followed in this study, since the relation between precipitation and temperature obviously affects a hydrological response of a catchment.

Several complex multivariate correction methods can be found in the recent literature (e.g. Hnilica et al., 2017; Johnson and Sharma, 2012; Mehrotra and Sharma, 2016). Nevertheless, the increasing complexity of such methods and their unclear effect on the climate change signal have recently been the subject of serious criticism (Ehret et al., 2012; Maraun et al., 2017). Hnilica et al. (2019) pointed to the high sensitivity of complex correction methods to outlying values. For these reasons, the approach of Piani and Haerter (2012) used in our study represents a compromise between the overly complex multivariate methods and the standard single-variate approach (e.g. quantile mapping), which leaves the relations between variables entirely uncorrected.

Individual climate chains from the EURO−CORDEX experiment predict the amount of precipitation for the future differently, where some of them predict an increase in precipitation for central Europe while others predict a decrease. The corrected data for our study catchments showed that the variability between individual climate chains as well as between individual catchments is very high and ranges from -12 % (a relative decrease in precipitation compared to the current climate) to +28 % (a relative increase in precipitation). This variability, together with different downscaling methods may results in large differences in simulated runoff seasonality (Meyer et al., 2019). Our results clearly showed that despite the simulated future decrease in snow storages, the large variability exists between individual climate chains mainly due to the large variability in precipitation. This was also clearly reflected by the large variability in simulated seasonal runoff. Since the snow storages are strongly sensitive to changes in both air temperature and precipitation in the cold period, it is still a question as to which degree the predicted increase in air temperature will be compensated by the increase (or decrease) in precipitation since the future evolution of precipitation remains uncertain, at least for the region of central Europe. Nevertheless, despite this large variability in future precipitation, the snow storages decreased in all scenarios giving the high confidence of increasing air temperature as a dominant factor which controls the snow storages in our study catchments.

An important uncertainty of the hydrological model represents the input data used for model calibration and validation. In our study, the input climate data were formed by stational data located within or nearby the individual catchments. Since no gridded product of precipitation, air temperature and SWE is available for Czechia, the stational data represented the only possible input to the model for our study area. To enable correct spatial distribution of all climate and snow data within the catchment and to reduce the uncertainty in the model calibration, lapse rates (correction factors for elevation) for air temperature and precipitation were calibrated by the model separately for each catchment. Additionally, the uncertainty of the model calibration has been reduced using a multi-parameter approach which enabled calibration of the model not only against runoff, but also against SWE. This approach contributed to higher agreement of the observed and simulated values especially in higher elevation catchments with higher importance of the snowpack in the water cycle (Girons Lopez et al., 2020; Jenicek and Ledvinka, 2020a). The uncertainty in hydrological model simulations was further reduced using the 100 parameter sets resulting from 100 calibration runs for each catchment which contributed to more robust results. This way, the effect of equifinality (multiple parameter sets provide equally good or acceptable model outputs) was partly reduced. Although the results may be influenced by specific model choice, model structure and parameters uncertainty, several studies showed that the uncertainty mainly stems from the GCMs and RCMs and from the specific emission scenario (Addor et al., 2014; Chegwidden et al., 2019; Parajka et al., 2016). However, the model choice may by more important for future projections in higher elevation catchments with dominating snowmelt runoff regime (Addor et al., 2014).

5. Conclusions

Future simulations of the individual components of the water cycle showed a considerable decrease in snow-related variables for all study catchments in Czechia at all elevations. The HBV model simulated decrease in annual maximum SWE by 30 %–70 % until the end of the 21st century compared to current climate with higher values for elevations below 1100 m a.s.l. Additionally, the snow-covered season will be on average 40–60 days shorter in the future. The results indicated that the shortening of the snow-covered season will be caused more by earlier melt-out rather than by later snow onset (Fig. 3).

Nevertheless, the response of the snowpack to changes in climate indices predicted by individual climate chains showed a large variability with the mean decrease in maximum SWE by 13 % for the most snow-rich conditions and by 76 % for the most snow-poor conditions for the future period 2070−2099. The results also indicated that the increase in air temperature causing the decrease in snowfall might be partly offset by the increase in winter precipitation.

The simulations of the future runoff showed that the period of highest streamflow will occur on average a month earlier (following the earlier snowmelt), and the seasonal runoff volume will be significantly lower due to lower snowmelt inputs (Fig. 5). Additionally, the model predicted the increase in winter runoff for the future period due to the increase in air temperature and thus the shift from snowfall to rain. These changes may impose more pressure to create adaptation strategies for water reservoirs management to keep all reservoir functions, such as flood and drought protection, drinking water supply and hydropower.

The absolute decreases in snow storages in higher elevation catchments are larger compared to lower elevation catchments and thus the summer runoff in those catchments will be more affected by changes in seasonal snowpack. Consequently, the largest increase in deficit volumes seems to be associated with the highest elevation catchments (Fig. 7). This might have important implications for water availability in downstream areas.

The climate chains leading to overall dry conditions are associated with both lowest summer precipitation and seasonal snowpack (Fig. 8). It means that higher deficit volumes were simulated for chains with both low summer precipitation and low seasonal snowpack. This indicated that snow can partly increase summer low flows, although the role of summer precipitation is obviously more important. The expected lower snow storages might therefore contribute to more extreme summer low flow periods in the future next to increase in evapotranspiration and potential decrease in precipitation.

The comparison of the future snow changes for individual climate chains combining different RCPs showed an overall decrease in snow storages and significant changes in spring and summer runoff (Figs. 4 and 6). However, the RCP 2.6, showed significantly smaller changes compared to the RCP 4.5 and RCP 8.5 both in terms snow storages and seasonal runoff. In this respect, the results highlighted the importance of efforts to keep the warming well below 2 °C compared to the pre-industrial period as defined by the Paris agreement. This is important for mitigation and adaptation strategies related to climate change impacts in mountain regions.

Data availability

Meteorological and hydrological data for the calibration of the HBV model were obtained from the Czech Hydrometeorological Institute (contact person: Ondrej Ledvinka, ondrej.ledvinka@chmi.cz). EURO−CORDEX data were downloaded from https://www.euro-cordex.net/. The bias-corrected climate data as well as HBV model outputs for study catchments are available from the corresponding author upon request.

Author contribution

MJ initiated the study, developed the methodology and together with ON performed all analyses. JH and VS prepared future climate scenarios which further controlled the hydrological model. MJ prepared the manuscript with a contribution from all co-authors.

Declaration of Competing Interest

The authors declare that they have no conflict of interest.

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Appendix A. Supplementary data

Supplementary material related to this article can be found, in the online version, at doi:https://doi.org/10.1016/j.ejrh.2021.100899.

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