Importance of snowmelt contribution to seasonal runoff and summer low flows in Czechia

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Abstract. The streamflow seasonality in mountain catchments is largely influenced by snow. However, a shift from snowfall to rain is expected in the future. Consequently, a decrease in snow storage and earlier snowmelt is predicted, which will cause changes in spring and summer runoff. The objectives of this study were to quantify 1) how inter-annual variations in snow storages affect spring and summer runoff, including summer low flows and 2) the importance of snowmelt in generating runoff compared to rainfall. The snow storage, groundwater recharge and streamflow were simulated for 59 mountain catchments in Czechia in the period 1980–2014 using a bucket-type catchment model. The model performance was evaluated against observed daily runoff and snow water equivalent. Hypothetical simulations were performed, which allowed us to analyse the effect of inter-annual variations in snow storage on seasonal runoff separately from other components of the water balance. The results showed that 17-42% (26% on average) of the total runoff in study catchments originates as snowmelt, despite the fact that only 12-37% (20% on average) of the precipitation falls as snow. This means that snow is more effective in generating catchment runoff compared to liquid precipitation. This was documented by modelling experiments which showed that total annual runoff and groundwater recharge decreases in the case of a precipitation shift from snow to rain. In general, snow-poor years are clearly characterized by a lower snowmelt runoff contribution compared to snow-rich years in the analysed period. Additionally, snowmelt started earlier in these snow-poor years and caused lower groundwater recharge. This also affected summer baseflow. For most of the catchments, the lowest summer baseflow was reached in years with both relatively low summer precipitation and snow storage. This showed that summer low flows (directly related to baseflow) in our study catchments are not only a function of low precipitation and high evapotranspiration, but they are significantly affected by previous winter snowpack. This effect might intensify the summer low flows in the future when generally less snow is expected.

1 Introduction

Mountain catchments are often influenced by snow, which largely affects seasonality in runoff. However, snow water
30 equivalent (SWE) has been decreasing in many mountains regions over the last decades and spring snowmelt occurs earlier in
the year (Beniston, 2012; Fyfe et al., 2017; Harpold et al., 2012; Klein et al., 2016; Marty et al., 2017b). This suggests that
snow and snowmelt dynamics respond to increase in air temperature and precipitation due to climate changes (Barnett et al.,
2005; Bavay et al., 2013; Jenicek et al., 2018; Marty et al., 2017a). The higher air temperature causes a shift from snowfall to
rain during winter resulting in lower snowfall fraction, a rate of snowfall water equivalent (hereafter referred to as snowfall)
35 to annual precipitation. Consequently, the amount of snow and peak SWE is affected as well (Berghuijs et al., 2014; Jenicek
et al., 2016). As a response to increasing temperature, the snowmelt starts earlier in the year which can result in slower
snowmelt rates due to lower available energy, such as solar radiation (Klein et al., 2016; Musselman et al., 2017). On the
contrary, water demand of vegetation is lower earlier in the year.

Changes in snow accumulations, snowmelt timing and snowmelt rates significantly affect groundwater recharge which is lower
40 and occurs earlier (Beaulieu et al., 2012; Foster et al., 2016). For the groundwater recharge, the topography is important as the
water is transported from steep terrain near mountain ridges to lower elevation (Carroll et al., 2019). Therefore, higher
elevations are important for catchment storage (Floriancic et al., 2018; Hood and Hayashi, 2015; Staudinger et al., 2017) and
important to stabilize streamflow at lower elevations especially during drought periods (Carroll et al., 2019; Cochand et al.,
2019). Higher snowpack generates higher groundwater flow driven by snowmelt rates and thus contributes more to streamflow
45 (Barnhart et al., 2016).

The decreasing snow storages and groundwater recharge in mountain regions further influence summer low flows (Dierauer
et al., 2018; Ledvinka, 2015; Li et al., 2018; Van Loon et al., 2015; Potopová et al., 2016). Higher snowpack and later snowmelt
contribute to summer baseflow and increase the period for which snowmelt contributes to streamflow (Hammond et al., 2018;
Langhammer et al., 2015). For catchments at higher elevations, this period may cover the whole summer (Godsey et al., 2014;
50 Jörg-Hess et al., 2014). Earlier snowmelt and melt-out in the future will further shorten this period (Etter et al., 2017; Jenicek
et al., 2018). For summer low flows in Europe, the liquid precipitation and evapotranspiration is usually more important than
previous winter snow storages (Floriancic et al., 2019), but low snowpack causes a decrease in summer low flows in the case
of simultaneously low summer precipitation (Jenicek et al., 2016).

Several earlier studies were focused on identification of physical mechanisms of how earlier or later snowmelt influences the
55 runoff generation (summarized in Barnhart et al., 2016). One mechanism is that due to higher air temperature during earlier
snowmelt, a greater proportion of snowmelt evaporates (Barnhart et al., 2016; Bosson et al., 2012). In contrast, earlier snowmelt
occurs in periods when vegetation is less active, thus uses less water. Therefore, more water flows into the stream. Another
effect is that slower snowmelt (associated with earlier snowmelt; Trujillo and Molotch, 2014) leads to lower streamflow
generation due to the fact that during later and faster snowmelt, soil moisture might be more often above its capacity leading
60 to higher streamflow. Additionally, higher snowmelt rates lead to higher baseflow (Barnhart et al., 2016). Therefore, it is

important to analyse inter-annual variability of snow, climate and streamflow characteristics in areas where snow is an important source for runoff generation.

The role of snow in seasonal catchment runoff, summer low flows and water supply is frequently quantified using several snow-related metrics, such as peak SWE or snowfall fraction (Curry and Zwiers, 2018; Hammond et al., 2018). When a
65 modelling approach is applied, the ratio of snowmelt runoff to total runoff is often used (Jenicek et al., 2018; Li et al., 2017; Stahl et al., 2016). Such approaches enable to quantify the importance of snow in spring and summer runoff and the "memory effect" of a specific catchment, i.e., how long snow affects runoff after snowmelt (Godsey et al., 2014; Jenicek et al., 2016). Several studies also showed that snow is more effective in generating the runoff compared to liquid precipitation. For example, in the western United States 53% of the total runoff originates from snowmelt, despite the fact that only 37% of the precipitation
70 falls as snow (Li et al., 2017). The mentioned discrepancy between snowfall amount and snowmelt runoff might have substantial impacts on seasonal distribution of the runoff and water supply in a warming climate, when more rainfall is expected compared to snowfall (Harpold et al., 2017; Safeeq et al., 2016).

Snowfall fraction might be an interesting indicator to what degree snow affects summer low flows. For example, Meriö et al. (2019) found a threshold of snowfall fraction of 0.35 in their study catchments in Finland. In catchments with snowfall fraction
75 above the threshold, the summer low flows were sensitive to inter-annual variations in snow storages. Below the threshold, summer precipitation and air temperature together with catchment characteristics were more important. Although specific thresholds might differ across world regions, the role of snow storages on summer low flows was shown in several studies from Europe and North America (Godsey et al., 2014; Jenicek et al., 2016).

The above studies show that changes in snow storages and their impact on runoff and groundwater recharge are in the centre
80 of the current research. However, there is still a need to investigate mutual interactions between individual characteristics, such as snow storages in snow-poor and snow-rich years and their interaction with groundwater recharge, snowmelt runoff, spring and summer rainfall, summer baseflow and low flows. The need to explain this variability and change of temporal and spatial processes also arose from the hydrology community initiative to identify major unsolved scientific problems in hydrology (Blöschl et al., 2019).

85 Therefore, the objectives of this study were to quantify 1) how inter-annual variations in snow storages affect spring and summer runoff, including summer low flows and 2) the importance of the snowmelt in generating the runoff compared to rainfall at different elevations. The focus on elevation is important since snow storage and its potential change is highly dependent on elevation. We quantified the influence of inter-annual variability in snow storages on summer runoff and low flows in 59 catchments in Czechia with significant snowmelt contribution to runoff. Focusing on non-alpine region of Central
90 Europe is important, as most of studies addressing this topic were conducted in alpine region at higher elevations and with partly different climate conditions. For snowmelt contribution, we used an "effect tracking" algorithm which is now accepted and used by the modelling community aiming to track the effect of individual water sources in the system to assess either the inter-annual variability of runoff components or their potential changes in the future (Weiler et al., 2018).

2 Data and methods

2.1 Study area and input data

95 The 59 mountain catchments with minor influence of human activity together with significant snow influence on runoff were selected (Fig. 1). An important criterion for selection was the availability of long-term time series of hydrological and meteorological observations (>35 years). With our selection, we covered most of the mountain regions of the Czechia, namely the Bohemian Forest (BF, 12 catchments), Ore Mts. (OM, 2 catchments), Western Sudetes (WS, 16 catchments), Central

100 Sudetes (CS, 5 catchments), Eastern Sudetes (ES, 13 catchments) and Western Carpathians (WC, 11 catchments).

The daily runoff data were available for all closure profiles of the catchments from the Czech Hydrometeorological Institute (CHMI). Similarly, daily precipitation, daily mean air temperature and weekly SWE data from selected climatological stations located within or nearby individual catchments were obtained from the CHMI. Stational data were further used in a hydrological model. All data were available for the period from 1980 to 2014, except three catchments, from which data started

105 in 1981 or 1982.

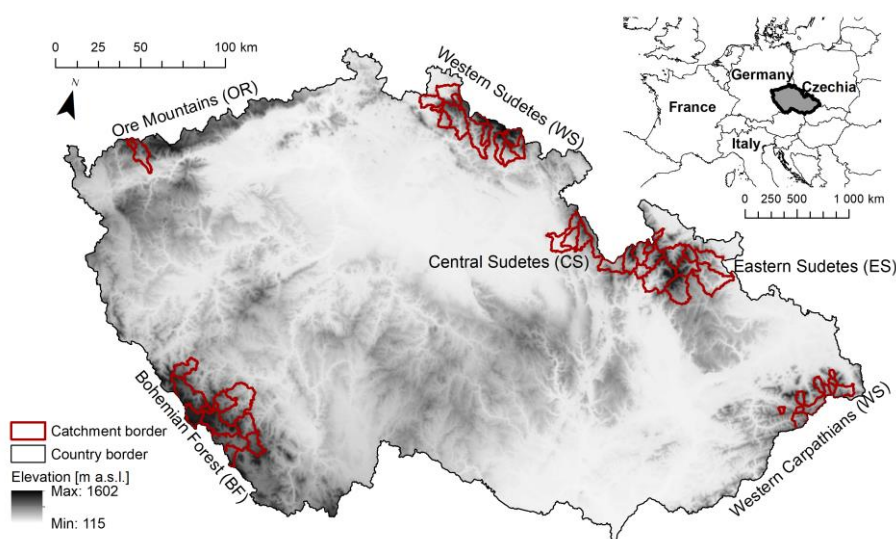



Figure 1: Location of the study catchments in Czechia.

2.2 HBV model

To assess the impact of inter-annual variability of snow on streamflow, we need to simulate individual components of the

110 rainfall-runoff process at a catchment level. For this, we used a bucket-type HBV model (Lindström et al., 1997) in its implementation, called HBV-light (Seibert and Vis, 2012). Details for model structure and routines are described in several studies (Jenicek et al., 2018; Seibert and Vis, 2012).

The study catchments were sub-divided into elevation zones by 100 m reflecting the changes of precipitation and temperature with elevation. The time series of daily precipitation, daily mean air temperature and monthly potential evapotranspiration

115 (P_{ET}) represent the main model inputs. The temperature-based method described by Oudin et al. (2005) was used for P_{ET} calculation.

The HBV model was calibrated for each catchment against observed runoff and SWE using a genetic calibration algorithm (Seibert, 2000). The integrated multi-variable model calibration approach was applied using a combination of three goodness-of-fit criteria (Table 1); 1) Model efficiency for runoff using logarithmic values (R_{runoff}) (Nash and Sutcliffe, 1970), 2) model

120 efficiency for SWE (R_{SWE}) and 3) volume error (R_{vol}). The resulting objective function ($R_{weighted}$) was calculated from these three criteria as a weighted average with a, b and c representing weights for each criterion. Different weights were tested to achieve the best possible performance of the model.

Table 1: Objective functions used for model calibration and validation

Objective function	Equation	Weights
Model efficiency for runoff	$R_{runoff} = 1 - \dfrac{\sum(\ln Q_{obs} - \ln Q_{sim})^2}{\sum(\ln Q_{obs} - \overline{\ln Q_{obs}})^2}$	60%
Model efficiency for SWE	$R_{SWE} = 1 - \dfrac{\sum(SWE_{obs} - SWE_{sim})^2}{\sum(SWE_{obs} - \overline{SWE_{obs}})^2}$	20%
Volume error	$R_{vol} = 1 - \dfrac{|\sum(Q_{obs} - Q_{sim})|}{\sum(Q_{obs})}$	20%
Weighted efficiency	$R_{weighted} = a{\cdot}R_{runoff} + b{\cdot}R_{SWE} + c{\cdot}R_{vol}$	n.a.

125 The observed time series were divided in two sub-series; first (1980-1997) was used for model calibration, second (1998-2014) for model validation. The two periods covered both cold and warm years and wet and dry years. Although mean annual air temperature increased by 0.7°C for the validation period, annual precipitation and peak SWE did not change significantly between both periods.

The parameter uncertainty was addressed by performing 100 calibration runs resulting in 100 parameter sets. These 100 sets

130 were further used to create 100 simulations. A median simulation was used for further analyses. This median simulation was derived in a way, that individual daily values for specific simulated variables (runoff, SWE, groundwater recharge etc.) were calculated as a median from all 100 respective values resulting from simulations. The similar procedures for model set-up and calibration was also used earlier in (Jenicek et al., 2018), although in different region.

2.3   Modelling experiments

135 To study the effect of inter-annual variations in snow storages on seasonal runoff characteristics (such as baseflow, deficit volumes and recharge) separately from other meteorological controls (mainly liquid precipitation and actual evapotranspiration, A_{ET}), the hypothetical simulations were performed. These simulations consist in changing the threshold temperature T_T (a parameter included in the HBV snow routine) which is used to differentiates between snow and rain. By changing the T_T, we can control the amount of accumulated snow and snowmelt timing, while other variables remain unaffected

140 (such as total amount of precipitation). Therefore, the T_T was progressively changed from -5°C to +5°C. Changes in this parameter influenced the simulated snowfall and thus SWE, snowmelt onset, melt rates and melt-out. The experiment was applied to the whole study period and thus capturing a variety of meteorological conditions. In this experiment, the snowfall correction factor (S_{FCF}) correcting solid precipitation for undercatch, was set to "1" (meaning no correction applied to original input precipitation data). Similarly, P_{ET} was not adjusted according to the inter-annual variations in air temperature (meaning

145 only input daily P_{ET} was used, calculated from long term data as described earlier). This enabled a separation of the effect of changing T_T on snow characteristics and seasonal runoff from other potential effects. With this procedure, we were able to attribute simulated runoff changes to changes in winter conditions.

2.4   Snow and streamflow signatures

We calculated several snow, groundwater and streamflow characteristics to analyse the impact of inter-annual variations in

150 snow accumulations on seasonal runoff. These characteristics were calculated from median simulations as described in the section 2.2.

Snow conditions for individual years of the study period were represented by February to May maximum SWE (SWE_{max}) which represents the late winter snow maximum. The snowfall fraction (S_f) describes the phase of precipitation. The snowfall fraction was calculated as the fraction of annual snowfall to annual precipitation using a single threshold temperature T_T. The

155 value of T_T resulted from calibration of the specific catchment.

The snow component of the streamflow (Q_s) was simulated by the HBV model. The method to track both rain and snow components, so called "effect tracking", is based on complete mixing of the two components in a "virtual mixing tank" (Stahl et al., 2016; Weiler et al., 2018). There are two assumptions for calculation; 1) the liquid water occurring in the snowpack is either considered as snow (in the case of melt of the snowpack) or rain (in the case of rain on snow), 2) when refreezing of the

160 liquid water in the snowpack is simulated, the source is considered as snow. Both daily and seasonal snow runoff (Q_s) and the fraction of snow runoff to annual runoff (Q_{sf}) was used in further analysis.

The groundwater recharge (G_w) was simulated by the HBV model. It represents the outflow from the soil box into groundwater boxes defined by the model (Seibert and Vis, 2012). From simulated time series, the fraction of winter (Dec-Feb, G_{w-DJF}) and spring (March-May, G_{w-MAM}) recharges to total annual recharge were calculated.

165 While the winter and spring recharge is a useful indicator to show how snow contributes to groundwater storage, the summer baseflow is related to the state of groundwater storage and thus represents both summer precipitation inputs and previous (spring) precipitation and snowmelt groundwater recharge. The summer (June to August, JJA) baseflow (Q_b) was calculated by the HBV model as an outflow from the lower groundwater box (S_{LZ}) which is a part of the model's response routine. The inflow into the S_{LZ} is controlled by percolation (parameter P_{ERC}, mm d^{-1}) and outflow is controlled by a recession coefficient

170 K_2 [d^{-1}] (Seibert and Vis, 2012).

The deficit volume (D_v) represents a water volume lacking in rivers below the defined threshold (Van Loon, 2015). Therefore, it is used as a measure to describe hydrological drought conditions. We used 90th percentile of the flow duration curve. We

also tested the 75th percentile without any major impact on the results. We used a variable level threshold method which uses different thresholds calculated separately for individual months in summer (June to August).

175 To show the inter-annual and seasonal differences in snow runoff in study catchments, most of the signatures were calculated for relatively snow-rich years and relatively snow-poor years. The snow-rich years were defined as years with annual SWE_{max} above the third quartile of the study period (1980-2014), while the snow-poor years represent years with annual SWE_{max} below the first quartile of the study period.

3 Results

3.1 Model calibration and validation

The results arising from model testing showed the overall good performance of the model both for the calibration and validation periods (Fig. 2). The model simulated correctly both runoff (R_{runoff}, R_{vol}) and SWE (R_{SWE}). The SWE was reproduced better at higher elevation catchments which contributed to mostly high agreement of observed and simulated runoff in terms of water balance and runoff seasonality. The calibration period shows slightly higher values of individual objective functions compared to those from the validation period. The results for the validation period represent more realistic model performance since calibration values represent rather the theoretically best possible model performance at the specific catchment.

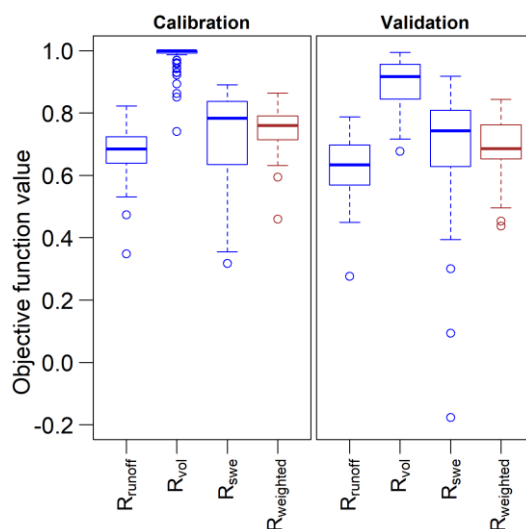



Figure 2: Results of model calibration and validation for study catchments. The objective function value for each catchment represents a median from 100 parameter sets. Boxes represent 25th and 75th percentile (with median as a thick line), whiskers 190 **represent 1.5 multiplier of interquartile range.**

3.2 Relative importance of snow on runoff during snow-poor and snow-rich years

The snowmelt is more effective in generating the runoff compared to liquid precipitation (Fig. 3). On average, 26% of the total runoff originated as snowmelt in our study catchments (17-42% for individual catchments for the study period), despite the

fact that only 20% (12-37%) of the precipitation occurred as snow (Fig. 3a). Additionally, catchment runoff coefficients (the

195 ratio between catchment precipitation and runoff) are higher for snowfall to snowmelt runoff than for rainfall to rainfall runoff

(results not shown). Both S_f and Q_{sf} are lower for snow-poor years (brown points) compared to snow-rich years (blue points).

It also seems that the difference between Q_{sf} and S_f was higher for snow-rich years (up to 12% for some catchments) and

increased with elevation for catchments where the snow is relatively more important in generating the runoff (Fig. 3b, p value

< 0.001). These results might have important implications for annual runoff volume in the future when the decrease in snowfall

200 fraction is expected.

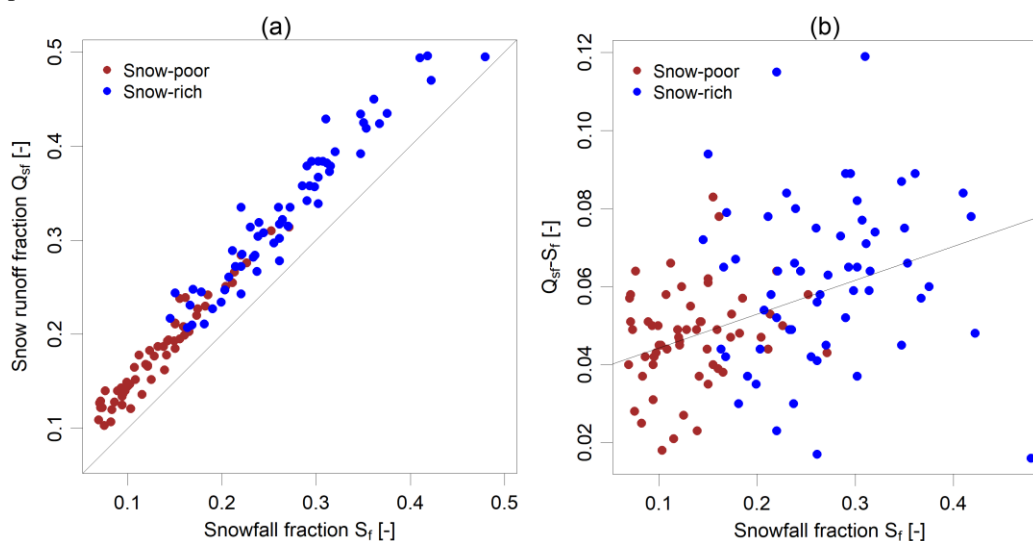



Figure 3: (a) A relation between snowfall fraction (S_f) and snow runoff fraction (Q_{sf}) for all study catchments. (b) A dependence of Q_{sf} and S_f difference on snowfall fraction. Brown points represent individual catchments during snow-poor winters; blue points represent catchments during snow-rich winters.

205 To show the seasonal distribution of snow runoff in study catchments, we calculated the relative contribution of snow runoff

on total runoff (Q_{sf}) as an average for each day of the year. Then we compared this relative contribution of snow runoff for

relatively snow-poor years and snow-rich years (Fig. 4a). The figure shows a significantly lower snowmelt contribution for

snow-poor years (up to 40%) in all catchments. The largest decrease in Q_{sf} occurred at the end of April/beginning of May in

the highest elevation catchments. This largest decrease in Q_{sf} is somewhat shifted towards March in lower elevation catchments

210 indicating that the snow runoff in lower elevation catchments occurs earlier in the year due to higher air temperatures and thus

earlier snowmelt onset.

In contrast, there was an increase in relative snow contribution to runoff during snow-poor years in winter, which indicates

that snow poor years were usually warmer compared to snow-rich years and thus the runoff increased due to more snowmelt

periods during winter (Fig. 4a). Additionally, the decrease in relative snow runoff contribution also occurred during summer

215 (June-August) which indicates that snow melted earlier in snow-poor years and less contributed to spring groundwater recharge

and thus summer runoff.

While Fig. 4a shows the snow contribution to total runoff (Q_{sf}), Fig. 4b shows monthly differences in simulated total runoff for snow-poor and snow-rich years for four selected catchments representing different geographical regions and elevations. As expected, March to May (sometimes even June) runoff was much lower for snow-poor years compared to snow-rich years due

220 to lower snow storages and thus snowmelt runoff. In contrast, winter runoff increased in snow-poor winters due to more rain than snowfall (lower S_f) and thus winter runoff occurred without delay. Summer runoff differed little between snow-rich and snow-poor years. Interestingly, total annual runoff in snow-poor years was often much lower compared to snow-rich years. This may be partly explained by higher effectiveness of snowfall to generate the runoff (Fig. 3) or by the fact that less snow in specific years was connected not only to warmer air temperatures, but also to the lack of precipitation (results not shown).

225 Therefore, this lack of winter precipitation caused the decrease in total annual runoff depth in our study catchments during snow-poor years.

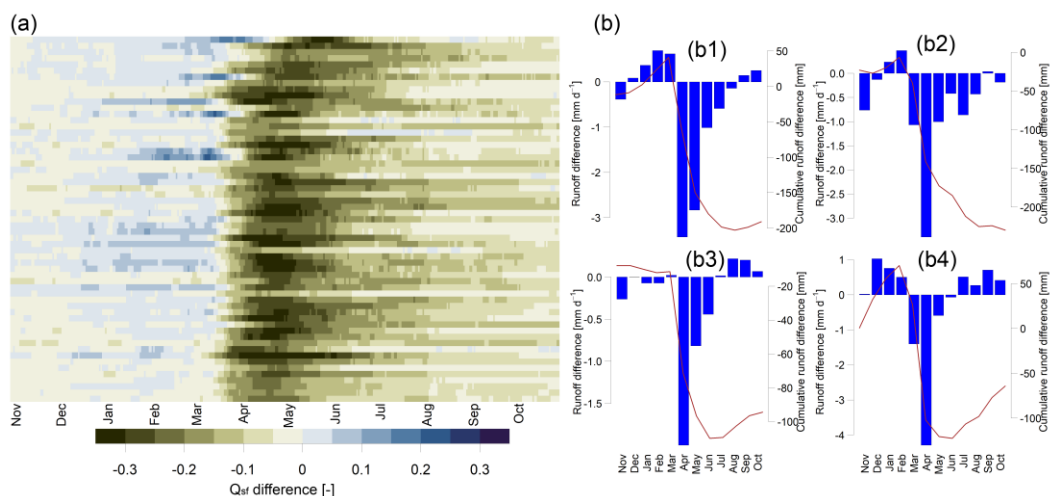



Figure 4: (a) Difference between mean daily Q_{sf} [-] for snow-poor and snow-rich years. Rows represent individual catchments (sorted from top to bottom according to elevation from highest to lowest), columns represent day of year. (b) Difference between mean

230 **monthly runoff (blue bars) and cumulative monthly differences in runoff (red line) for snow-poor and snow-rich years; (b1) Vydra (Bohemian Forest), (b2) Cerna Nisa (Western Sudetes), (b3) Desna (Eastern Sudetes), (b4) Ostravice (Western Carpathians).**

Lower snow storages in snow-poor years compared to snow-rich years led to lower snowmelt runoff contribution (Fig. 5a) and thus lower seasonal groundwater recharge. This seasonal recharge was expressed as a fraction of December to May recharge on total annual recharge (Fig. 5b). These recharge fractions are clearly lower for the entire cold season for snow-poor years in

235 57 out of 59 catchments (97%), with higher differences for catchments with higher S_f. However, recharge fractions were higher for the period from December to February (DJF) and lower from March to May (MAM) for snow-poor years (not shown) indicating that winter liquid precipitation during snow-poor years led to higher recharge during winter and thus partly compensated the lower recharge fraction during spring. The lower seasonal recharge fraction also caused lower annual recharge for snow-poor years compared to snow-rich years despite the fact that annual precipitation was almost the same for both groups

240 (results not shown). Lower annual groundwater recharges probably caused lower annual runoff in snow-poor years compared to snow-rich years (Fig. 5c). This partial result supports the results shown in Fig. 3.

Lower snow storages, snowmelt runoff and spring recharge also caused lower summer baseflow in 42 out of 59 catchments (71%), although only for 26 catchments was the difference higher than 5 mm (a sum for JJA period), affecting partly catchments with higher S_f (Fig. 5d). Additionally, summer baseflow strongly negatively correlates with summer deficit

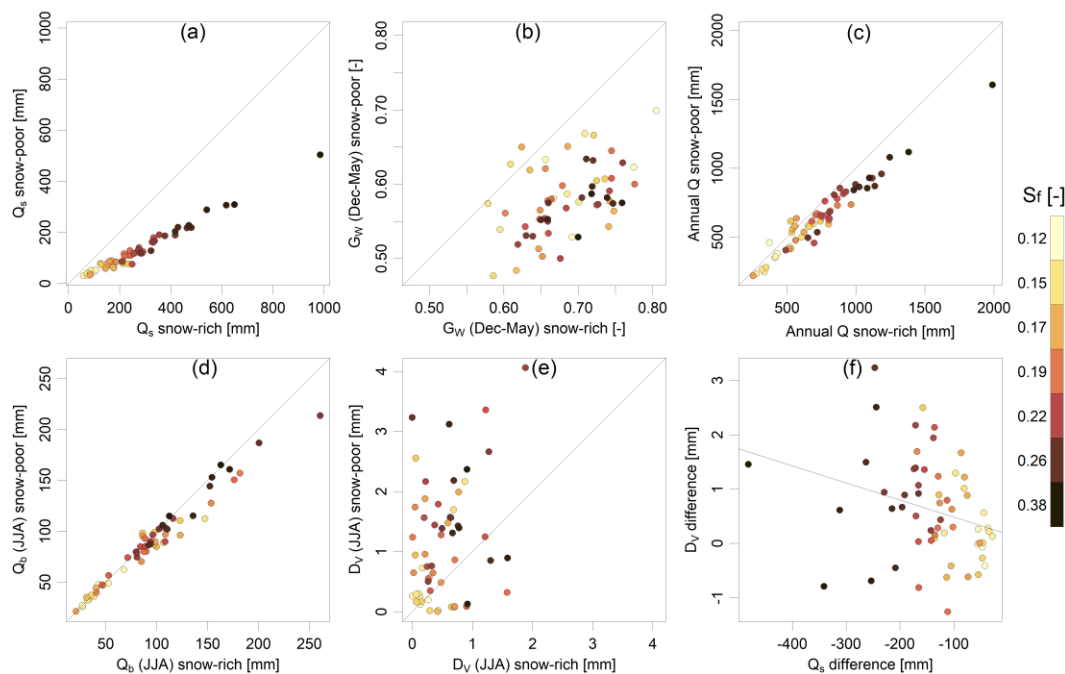
245   volumes (median Spearman rank correlation for all catchments reached the value of -0.8). It resulted in higher summer deficit volumes in snow-poor years compared to snow-rich years for 47 out of 59 catchments (80%, Fig. 5e). Additionally, the difference in deficit volumes between snow-poor and snow-rich years was negatively correlated (p value < 0.05) with difference in snowmelt runoff between snow-poor and snow-rich years with usually larger absolute difference for catchments with higher S_f (Fig. 5f).

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Figure 5: Difference in selected signatures for snow-rich and snow-poor years for study catchments; (a) Annual snowmelt runoff, (b) Seasonal recharge fractions (Dec-May), (c) Annual runoff, (d) Summer baseflow (JJA), (e) Deficit volumes (JJA), (f) Relation of deficit volumes (JJA) difference and snow runoff difference between snow-poor and snow-rich years. Colour scale used for snowfall fraction.

255   Figure 4 and Fig. 5 showed that less snow led to lower snowmelt contribution to total runoff and to lower winter and spring groundwater recharge. Besides, there was lower baseflow and higher deficit volumes in most catchments in snow-poor years. However, the figures do not provide us with information about how important snow storages are in influencing summer baseflow and deficit volumes compared to summer precipitation. Therefore, we analysed the relation between relative anomalies in summer (JJA) precipitation and relative anomalies in SWE_{max} compared to summer baseflow (Fig. 6). For this

260   analysis, four catchments were selected to show that summer baseflow is associated with both summer precipitation and annual SWE_{max} (Fig. 6). The lowest summer baseflow is associated with both the lowest summer precipitation and SWE_{max} (dark brown points are mostly located in the bottom left quadrants in individual panels of Fig. 6). Although, summer precipitation

seems to be more important for baseflow amount, Fig. 6 indicates that for some unit precipitation amount, the baseflow was lower for years with low annual SWE_{max}. Nevertheless, the described behaviour differs for individual catchments. For example,

265   the summer precipitation seems to be crucial for summer baseflow in the Ostravice catchment (Fig. 6d), while snow does not play an important role in influencing summer baseflow (the darkest brown points are located roughly equally in both bottom-left and bottom-right quadrants).
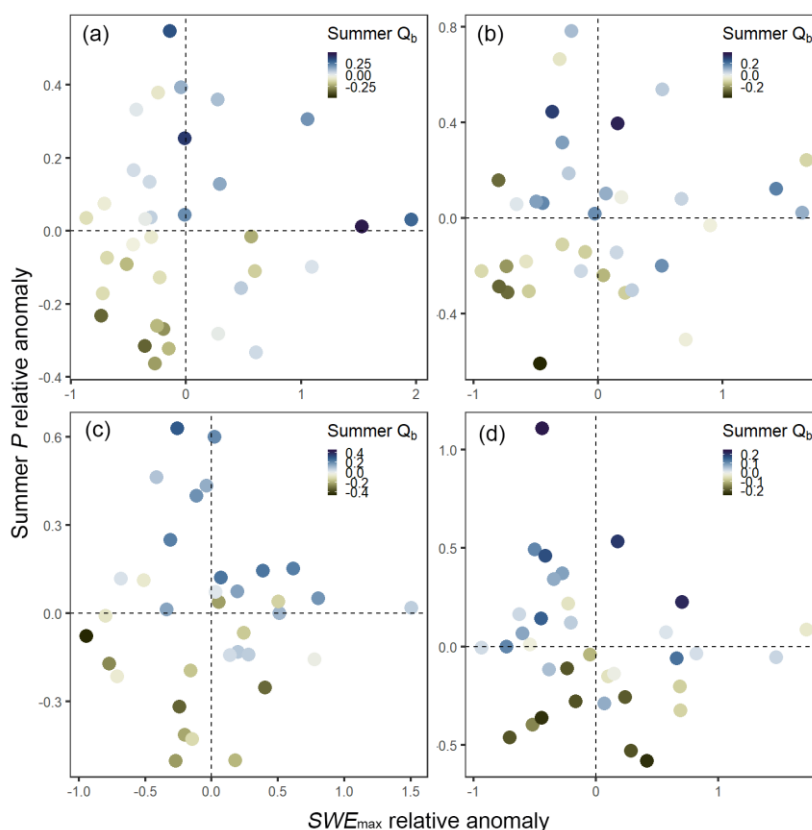
Figure 6: Dependence of summer baseflow (Q_b) relative anomalies on SWE_{max} and summer (JJA) precipitation relative anomalies at
270   **four selected catchments. (a) Vydra (Bohemian Forest), (b) Cerna Nisa (Western Sudetes), (c) Desna (Eastern Sudetes), (d) Ostravice (Western Carpathians)**

Similar mutual relations of SWE_{max}, summer precipitation and summer baseflow as shown in Fig. 6 were explored and generalized for all catchments (Fig. 7). Figure 7a shows that 1) below-average summer baseflow occurred for 58 from 59 catchments (98%) for below-average summer precipitation (points located left from x=0 line) and for 40 out of 59 catchments

275   (68%) for below-average SWE_{max} (points located below y=0 line) and 2) below-average summer precipitation generates lower baseflow than below-average SWE_{max} (points located above one-to-one line). This implies that snow is important in generating baseflow in our study catchments, but summer precipitation is more important. This is specifically valid for catchments in the Eastern Sudetes (ES) and Western Carpathians (WC) where summer precipitation seems to be the dominant driver for summer

baseflow generation (the darkest points in Fig. 7a are located above the $y=0$ line and thus show positive baseflow anomalies
280   for below-average SWE_{max}).

Figure 7b shows that baseflow is lowest for both below-average precipitation and below-average SWE_{max}. However, for 41 out
of 59 catchments (69%), the baseflow for years with below-average precipitation increased when there was simultaneously
above-average SWE_{max} (points are located above the one-to-one line). Besides, it shows that 13 out of 59 catchments (22%)
generated above-average baseflow despite below-average summer precipitation thanks to above-average SWE_{max} (points
285   located above $y=0$ line).
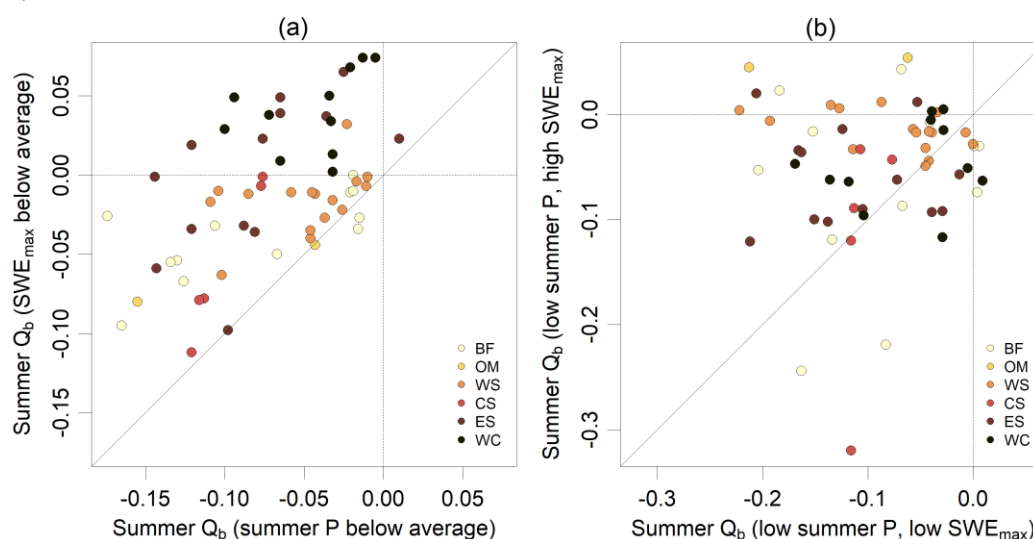
**Figure 7: (a) Relation of median summer baseflow for years with below-average summer precipitation and years with below-average
SWE_{max}. (b) Median summer baseflow for years with both below-average summer precipitation and below-average SWE_{max} related
to median summer baseflow for years with both below-average summer precipitation and above-average SWE_{max}. Points represent
290   individual catchments, colour represents region of the specific catchment (see the section 2.1 for region abbreviations).**

3.3   Modelling changes in runoff for snow-rain transition

The results presented in the section 3.1 indicated that less snow caused a decrease in spring recharge and thus a decrease in
summer baseflow and an increase in summer deficit volumes. However, the above approach did not allow to fully split the
effect of snow storages on summer runoff from the effects of other meteorological drivers, mainly summer precipitation and
295   A_{ET}. Therefore, we performed a simple modelling experiment simulating the transition of snowfall to rain while the total annual
precipitation and air temperature remained unchanged (see methods).

The model simulated lower snowfall and thus decrease in snow storages and shorter snow-covered season as a response to
threshold temperature T_T increase (result not shown). This snow storages decrease caused a decrease in both annual and
summer runoff for all 59 catchments (Fig. 8a, Fig. 8b). Simulated snow decrease resulted in lower groundwater recharge in
300   winter and spring for 56 out of 59 catchments (95%, Fig. 8c). Expectedly, the groundwater recharge increased in winter months
(Dec-Feb) thanks to earlier snowmelt and lower S_f (and thus rain infiltrated into the soil immediately). However, the decrease

in groundwater recharge in spring (March-May) was much larger than the increase in recharge in winter (not shown), resulting in an overall decrease in Dec-May recharge.

While the decrease in both summer and annual runoff provides the information about overall water availability regardless of
305 the seasonal distribution and extreme situations, the summer deficit volumes provide information about water availability during most critical situations, such as summer low flows. The modelling experiment showed an increasing trend in deficit volumes with decreasing S_f for most of the catchments (Fig. 8d). In contrast to the decrease in annual runoff with decreasing S_f, which was simulated for all study catchments, the increase in summer deficit volumes was simulated only for 34 out of 59 catchments (58%). The remaining catchments do not show clear tendency of deficit volumes or they behave in the opposite
310 direction, which indicates more complex behaviour of such catchments (e.g. due to specific catchment properties, such as geology or slope steepness). However, when looking at June to August deficit volumes separately, the deficit volumes increased for 80% of the catchments in June, for 52% in July, and for 55% in August (results not shown). This evolution suggested the decreasing role of snow in influencing deficit volumes in the analysed months. Similar results as for summer deficit volumes were also achieved for the number of days with deficit volumes, which also mostly increased when the snowfall
315 fraction decreased (results not shown).
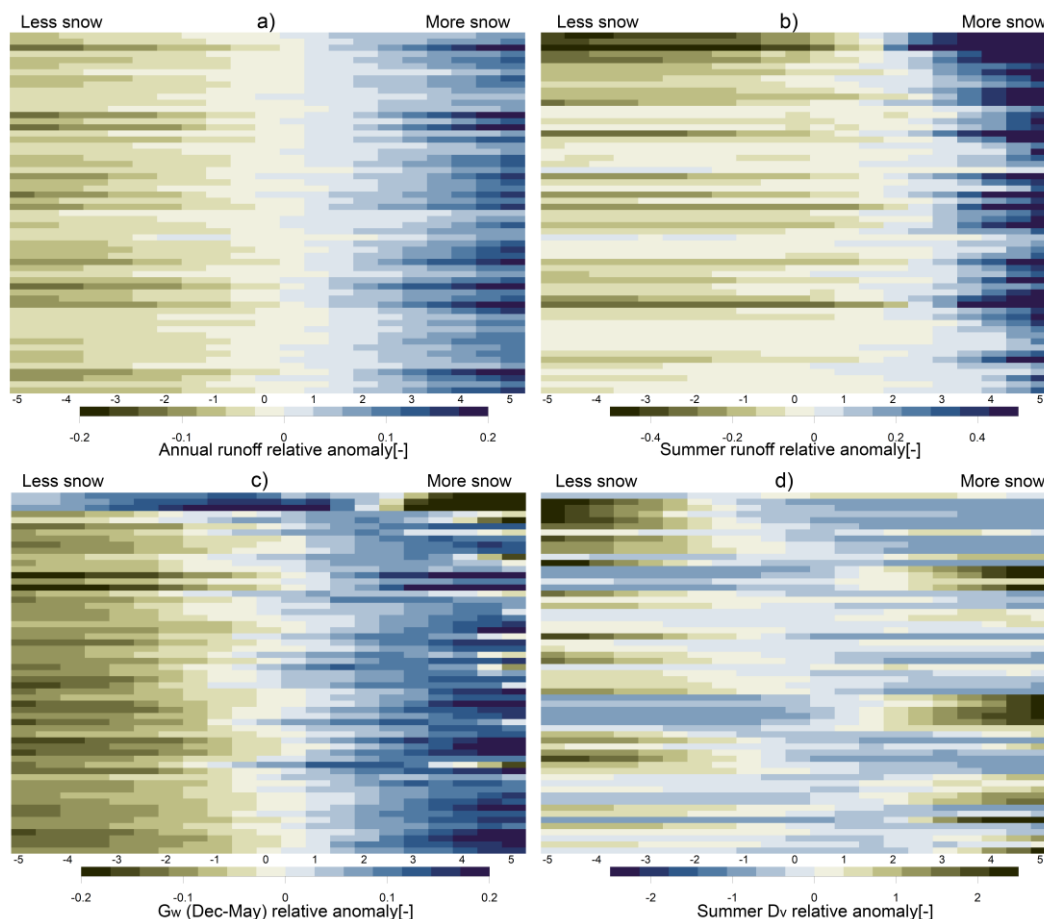
Figure 8: Relative change of selected signatures with increasing T_T for study catchments. (a) annual runoff, (b) summer (JJA) runoff, (c) groundwater recharge (Dec-May) and (d) summer (JJA) deficit volumes. Rows represent individual catchments (sorted from top to bottom according to elevation from highest to lowest), columns represent T_T used in model simulations. Colours show normalized values relative to their means (different scales used for individual panels).

4   Discussion

4.1   Data errors, model set up and parameters uncertainty

Since the presented results are based on HBV model runoff simulations, the uncertainty arising from the model parametrization needs to be addressed. This was done by 100 model calibration trials resulting in 100 parameter sets. This way the model generated more robust results. Additionally, a multi-variable approach was used for calibration to correctly simulate both SWE and runoff. This procedure led to better model performance especially in higher elevation catchments with higher snowpack as also shown in other studies (Etter et al., 2017; Jenicek et al., 2018; Seibert, 2000).

There are several issues related to model parametrization and calibration. Many of the model parameters might have an important effect on result interpretation. For example, P_{ET} calculated for a specific day did not account for actual air

330   temperature on that day, but it only reflected long-term mean of daily air temperature for the full study period (1980-2014). This affected simulated P_{ET} values which then did not reflect inter-annual variability due to inter-annual variability of air temperature. Nevertheless, by neglecting the potential feedbacks from P_{ET} inter-annual variability, this approach enabled a better separation of snow influence on summer low flows.

Modelling experiments also opened further questions related to model structure and parameterization, specifically how
335   individual model procedures and parameters represent the real natural processes. For example, a snowfall correction factor (S_{FCF}) is used in the model to correct solid precipitation for undercatch (e.g. due to wind). Nevertheless, it may also compensate some processes not explicitly included in the model, such as snow interception, sublimation and A_{ET} from snow cover. Similarly, threshold temperature T_T used in our modelling experiment to control snowfall fraction is also used to set the snowmelt onset and as a threshold temperature to distinguish whether the S_{FCF} will be applied or not. Therefore, S_{FCF} was set
340   to "1" in the modelling experiment to avoid modification of the solid precipitation amount throughout the experiment and thus entire water balance. The above-mentioned potential model artefacts might be important when using HBV or similar bucket-type models for impact studies, such as modelling the impact of climate change on catchment runoff.

4.2   Influence of snow storages on snowmelt runoff and groundwater recharge

The results showed that the fraction of runoff originating as snowmelt is by 2-12% higher than the fraction of precipitation
345   occurring as snow (higher values for snow-rich years and for catchments at higher elevation). This indicates that snowmelt is more effective in generating runoff compared to liquid precipitation. This interesting effect was also shown by Li et al. (2017) in the western United States using a similar modelling approach and it also supports results achieved by Berghuijs et al. (2014) who showed that higher snowfall fractions generates higher annual runoff in the western United States. The reason for such behaviour is probably due to a lower A_{ET} during winter and lower water demand by vegetation, both resulting in lower
350   precipitation losses. Besides, the A_{ET} is calculated only for a snow-free ground in the HBV model. Another reason might result from the fact that soil moisture is more often at its field capacity, thus the model also simulates higher runoff during snowmelt events (Barnhart et al., 2016; Li et al., 2017). In other words, snowmelt rates control the relative partitioning of snowmelt water between evapotranspiration and streamflow (Barnhart et al., 2016). The described catchment behaviour might have an important impact on runoff generation in the future, where the shift from snow to rain during winter is predicted due to the
355   increase in air temperature (Jenicek et al., 2018).

The results showed that contribution of snow to runoff is much lower for snow-poor than snow-rich years. The decrease was obvious also for the period from June to August indicating that snow also contributed to summer runoff for snow-rich years. The snow contribution to summer runoff including low flows was also shown by several other studies (Godsey et al., 2014; Jenicek et al., 2016; Stahl et al., 2016), although mostly for higher elevation catchments with later snowmelt compared to our
360   study catchments. The summer snow runoff originates from spring snowmelt, which propagated into deeper groundwater layers and contributed to runoff with delay. The snow runoff contribution was calculated using "effect-tracking" algorithm implemented in the HBV model using "virtual mixing tanks" with limited capacities. The method was tested within several

studies (Stahl et al., 2016; Weiler et al., 2018). The algorithm is a useful approach to assess changes in discharge components (Weiler et al., 2018).

365 The average contribution of groundwater to streamflow was from 22% to 84%, with higher values for catchments with generally higher snow storages. However, besides snow amounts, some regional differences were found. For example, catchments in the Bohemian Forest (rather flat catchments on metamorphic or granite rock with large portion of peatland) has larger groundwater contributions than catchments in the Western Carpathians (more steep catchments on flysch). Nevertheless, the potential influence of basin attributes on runoff generation needs to be further investigated. The relative fraction of
370 groundwater in streams is sensitive to inter-annual variation in snow storages as also shown by Carroll et al. (2019).

4.3 Influence of snow storages on summer baseflow and deficit volumes

Inter-annual variations in snow storages also affected summer baseflow, which is, additionally to spring snowmelt, related to summer precipitation and evapotranspiration. Our results showed that less snow led to lower snowmelt contribution to total runoff and to lower spring groundwater recharge with larger differences for catchments with higher S_f. These results correspond
375 to other studies (Carroll et al., 2019; Cochand et al., 2019; Meriö et al., 2019). However, it does not necessarily mean that summer baseflow and potentially low flows are lower as well since both the baseflow and low flows are more influenced by other water balance components, such as precipitation and evapotranspiration during spring and summer (Floriancic et al., 2019; Jenicek et al., 2016). Since the baseflow is a major runoff component during low flow periods, it can be used as an indicator showing the potential extremity of such periods. Our results showed that snow-rich years produced higher summer
380 baseflow in our study catchments. This indicated that spring snowmelt increased groundwater storage. Although the lower baseflow does not mean that potential low flow periods are more likely to occur, it indicates that if the low flow period occurs (e.g. due to lack of summer precipitation and/or high A_{ET}), the minimum streamflow might drop to lower values.

Although in most of catchments the summer precipitation was more important for summer baseflow, in some catchments with the highest snowfall fractions, the winter snowpack was probably of a similar importance as the summer precipitation. The
385 snow as the dominant mechanism controlling summer low flows was also proved by Meriö et al. (2019) for Finnish catchments with snowfall fractions higher than 0.35. Similar to the mentioned study, we also discovered some regional differences between our study catchments. For example, for catchments in the Eastern Sudetes and Western Carpathians, summer precipitation dominated the summer baseflow despite relatively high snow storages. This might indicate that those catchments had a shorter "memory effect" of snow to influence the runoff. Thus, the water exchange is faster in these catchments resulting in shorter
390 residence times. Our results did not provide possible reasons for such behaviour, but differences in climate regime (increasing continentality from west to east), geology (flysch in the Western Carpathians vs. metamorphic or granite rock in other regions) and morphology (higher slopes in the Western Carpathians) could provide some explanation as shown by several authors in other regions (Floriancic et al., 2018; Li et al., 2018; Staudinger and Seibert, 2014). However, more detailed research would be necessary.

395   Nevertheless, understanding how snow storages in snow-poor and snow-rich years influence summer baseflow and deficit volumes is always limited due to different meteorological conditions in individual years. For example, winter seasons in years with lower snow storages were also warmer and drier which could affect summer baseflow and low flows. However, despite these differences in winter precipitation and temperatures, summer precipitation, summer air temperature and summer A_{ET} were almost the same for both snow-poor and snow-rich years. This indicated that differences in summer runoff signatures can

400   be related to changes in snow storages and spring groundwater recharge.

4.4   Snow-rain transition and potential future impacts

The results related to catchment response in snow-rich and snow-poor years indicated important potential consequences for annual and seasonal runoff and deficit volumes, which might decrease (or increase, for deficit volumes) in the future when the decrease in snowfall fraction is expected. However, the hypothesis highlighting the importance of snowfall fraction on runoff

405   amount cannot be fully proven by splitting the study period into snow-poor and snow-rich years, due to the fact that snow-poor and snow-rich years differed not only in snow storages, but also in other meteorological signatures (as explained in the section 4.3). Therefore, this runoff volume decrease was proven by a modelling experiment simulating the progressive change from snowfall to rain leaving the total precipitation, air temperature and P_{ET} unchanged. This way, we were able to separate the effect of changing snow to rain from other water balance components. A similar approach, using the same model and

410   parameters (but for different purposes), was applied also by Jenicek et al. (2018).

The results of this hypothetical experiment showed an increase in both annual and summer (JJA) runoff for all 59 catchments. A closer look at the results suggests that the model simulated lower snowfall and thus snow storages that melted earlier. Due to more days without snow cover, A_{ET} increased (A_{ET} is calculated for days without snow cover in the model) which caused a decrease in total runoff. It means that increasing the contribution of liquid precipitation to total runoff cannot compensate the

415   lower contribution of solid precipitation in total runoff. Moreover, potential snow-rain transition in the future will be caused by an increase in air temperature, which was not changed in our modelling experiment. Therefore, the A_{ET} might be even higher due to a temperature increase (in the case of enough available water). Consequently, annual runoff will likely decrease even more than indicated by our experiment.

In contrast to the decreasing annual runoff due to decreasing snowfall fraction, the increase of summer deficit volumes was

420   simulated only for 58% of the study catchments. The remaining catchments did not show a clear trend or they behaved in the opposite direction. This suggested more complex behaviour of these catchments, which is probably caused by their location rather at lower elevations and thus lower snow storages. Therefore, other climatic variables and catchment properties, such as geology and related groundwater storages might be more important than snow storages. However, more research would be necessary to find a detailed explanation. Nevertheless, the results from modelling experiments are consistent with analyses

425   based on snow-poor and snow-rich years. Similar results as for summer deficit volumes were achieved also for the number of days with deficit volumes, which also increased when snowfall fraction decreased.

The decreasing annual runoff in the case of a precipitation shift from snow to rain showed that this shift in the precipitation phase will change the catchment behaviour such that less water might be available for summer runoff in the future. The lower annual runoff might be critical for water supply and water reservoir management (Brunner et al., 2019). For the seasonal water

430 balance, it will therefore be important to understand whether the future increase in winter precipitation predicted by climate models for the region of Central Europe can compensate the expected future reduction of the snowmelt component.

5   Conclusions

We found that 17-42% (26% on average) of the total runoff in our selected study catchments originates as snowmelt on average, despite the fact that only 12-37% (20% on average) of the precipitation falls as snow. It also seems that the difference is

435 increasing at higher elevations with higher relative importance of snow on runoff regime. This particular result proves that snow is more effective in generating catchment runoff compared to liquid precipitation. This might have an important impact on water availability in the case of a future decrease in snow.

The mentioned difference between snowfall fraction and snowmelt runoff fraction was also documented by modelling experiments which showed that total annual runoff decreased in the case of a precipitation shift from snow to rain even in the

440 case where the total amount of precipitation and P_{ET} remained unchanged. This might imply lower annual catchment runoff in the future when a precipitation shift from snow to rain due to air temperature increase is predicted by climate models.

In general, snow-poor years are clearly characterized with lower snow runoff contribution to total runoff compared to snow-rich years in the analysed period 1980-2014. Additionally, snowmelt started earlier in these snow-poor years and influenced the runoff for a shorter period compared to snow-rich years. Snow-poor years generated lower annual groundwater recharge

445 and annual runoff compared to snow-rich years despite similar annual precipitation and A_{ET}, which resulted in mostly higher deficit volumes.

Inter-annual variations in snow storages also affected summer baseflow, which is, besides snow, related to summer precipitation and evapotranspiration. For most of the catchments, the lowest summer baseflow was reached in years with both relatively low summer precipitation and snow storage. This showed that summer low flows (directly related to baseflow) are

450 not only the function of low summer precipitation, but they are significantly affected by previous winter snowpack. Although the summer precipitation is usually the most important climatic factor to control the summer low flow, the decrease in snow and earlier snowmelt might intensify the summer low flows in the future when generally less snow is expected.

Modelling experiments performed in this study using the HBV model opened further questions related to model structure and parameterization, specifically how individual model procedures and parameters represent the real natural processes. An

455 understanding of potential model artefacts might be important when using HBV or similar bucket-type models for impact studies, such as modelling the impact of climate change on catchment runoff.

Results indicated that future higher liquid precipitation will not compensate the lower solid precipitation which might result in lower annual runoff. The lower annual runoff might be critical for water supply and water reservoir management. For the

seasonal water balance, it will therefore be important whether the future increase in winter precipitation predicted by climate
460   models for the region of Central Europe can compensate for the expected future reduction of the snowmelt component.

6   Data availability

Meteorological and hydrological data to calibrate the HBV model were obtained from the Czech Hydrometeorological Institute
(contact person: Ondrej Ledvinka, ondrej.ledvinka@chmi.cz). The HBV model outputs are available from the first author upon
request.

465   ## 7   Author contribution

MJ initiated the study, developed the methodology performed all analyses. OL prepared input meteorological and hydrological
data used to calibrate the HBV model. MJ prepared the manuscript with a contribution from OL.

8   Competing interests

The authors declare that they have no conflict of interest.

470   ## 9   Acknowledgements

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